# EIGHTEENTH ANNUAL REPORT

OF THE

# UNITED STATES GEOLOGICAL SURVEY

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DIRECTOR

IN FIVE PARTS

PART II—PAPERS CHIEFLY OF A THEORETIC NATURE



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PART II.—PAPERS CHIEFLY OF A THEORETIC NATURE.

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# THE TRIASSIC FORMATION OF CONNECTICUT.

BY

WILLIAM M. DAVIS.

# CONTENTS.

# ILLUSTRATIONS.

# OUTLINE OF REPORT.

After a brief introductory sketch of the Triassic area of Connecticut and a review of the labors of geologists upon it, the history of the region is treated in three parts, entitled Deposition, Deformation, and Denudation. The floor upon which the Triassic strata rest is regarded as a rough peneplain, the result of the long-continued denudation of an earlier mountain range. A part of this peneplain being depressed to form a trough in which the Triassic strata were accumulated, the conditions prevailing during the period of deposition are examined and the sequence of strata thus formed is described. Especial interest attaches to the igneous rocks, which are found in three forms—dikes, occupying fissures; intrusive sills, driven in between the strata nearly parallel to the bedding planes; and extrusive flows, extensively poured out at three successive epochs. The evidence of extrusion and intrusion is presented in detail, with particular attention to the nature of the argument by which a conclusion is reached. Brief paragraphs are given to vulcanism and isostasy, the first advocating deep-seated forces resulting from the deformation of the earth's crust as the chief cause of the ascent of lavas toward or to the surface, the second doubting the sufficiency of crustal equilibrium to explain the facts here observed.

Part II, Deformation, includes warping, faulting, tilting, and uplifting. The argument leading to the demonstration of faults receives close attention, and the extrusive lava flows are shown to have a high value in this connection. The faults are described in detail under two classes: First, those which traverse the Triassic area; second, those which determine its margin, especially along the eastern side. The proof of the existence of faults is then reviewed, and the arrangement of two belts of black shale is introduced to give them final confirmation. The extrusive flows are manifestly of earlier date than the faulting, and the intrusive sills are likewise shown, in all probability, to have taken their places before the time of tilting and faulting. This division of the report closes with an explanation of the process by which the warped and faulted monoclinal structure is thought to have been produced.

Part III, Denudation, includes a consideration of the various forms that the region has had from the time when deposition was stopped by deformation to the present day. Particular attention is given to forms obtaining at three stages: First, the oblique ridges initiated by the monoclinal faulting; second, the peneplain produced at the close of the first cycle of erosion; third, the present form carved in the uplifted peneplain of the first cycle. The importance of composite topography, the product of two cycles of erosion separated by a movement of the land mass with respect to base-level, is thus emphasized. The different treatment of faults in geological and geographical problems is illustrated. Account is given of some of the notches in the trap ridges now or formerly occupied by transverse streams, and the origin of the existing drainage arrangement is explained. A brief statement concerning glacial deposits and recent changes of level closes this division of the report. A reconstruction of the original Triassic basin is then attempted. It will be noted that problems of paleontology and petrography are not included in the report.

Most readers may to advantage pass quickly over those sections which treat of repeated examples of similar structures. Thus, under "Dikes" the sections on Mount Carmel and Fairhaven will suffice for their needs. Under "Ridges of the western range" West Rock or Gaylord Mountain will serve as typical structures. For the Eastern Range the anterior sheet is well shown under Lamentation Mountain and at Tariffville; the main sheet, in Saltonstall Mountain or Quarry Ridge; the posterior sheet, back of Lamentation. The discussion of the date at which the sills were intruded is considered important. It is presented on pages 79, 80, and 81. In Part II the geometrical relations of faults and the illustrations of these relations in Lamentation and East Talcott blocks will be found suggestive. The summary of the evidence of faulting gives a measure of the validity of the hypothesis offered to account for the facts of observation. In Part III the discussion of the Cretaceous peneplain is considered important. The explanation of the arrangement of the existing streams may interest readers of a speculative turn of mind. The reconstruction of the entire Triassic structure from such of the remaining parts as are now in sight may recall attempts of a similar kind undertaken in the study of ancient organic forms.

# THE TRIASSIC FORMATION OF CONNECTICUT.

By William M. Davis.

## INTRODUCTION.

### THE CONNECTICUT VALLEY LOWLAND.

When the early colonists on Massachusetts Bay "became like a hive overstocked with bees, and many thought of swarming into new plantations," as Cotton Mather said at the time, some of the more adventurous pushed their way into the interior, and after crossing a rugged region of uplands and valleys came upon a fertile lowland through which ran the Connecticut River. Here the old settlements of Windsor, Hartford, and Wethersfield were founded. Beyond, the up lands rose even higher than before. Thus as early as 1637 the broad depression worn down on the weak Triassic sandstones between the resistant crystalline rocks of the uplands exerted a determining influence on the history of New England.

Even to this day the uplands repel rather than invite occupation. The isolated manufacturing villages in the upland valleys have, indeed, during this half of our century grown rapidly, but the population of the hill towns has now for many years been stationary or has turned to a decline. The broad lowland, on the other hand, becomes year by year more populous. Its farms prosper and its villages and cities flourish. The underlying cause of the strong difference in human opportunity is found here, as so often elsewhere, in the geological structure and geographical development of the region. The uplands consist of resistant schists and gneisses. These withstand the destructive attacks of the weather so well that they are only trenched by narrow valleys, and still retain over large areas a good measure of the height that was given them by the last general elevation of the region above sea level. The lowland is underlain by relatively weak sandstones and shales, and although this belt was at the time of general elevation lifted to as great an altitude as that given to the areas of harder rock on either side, the sandstones and shales have already wasted down to a lowland of moderate relief. The many manufacturing villages in the valleys of the uplands are shut in and separated by the hills on either side. Movement there is difficult except in the direction of the streams. On the lowland the surface is comparatively smooth. A great part of it is open to easy occupation, and movement is easy from place to place in almost any direction. Only here and there ridges of harder rocks rise over the low ground, and these are so often broken by deep notches that they serve rather as boundaries than barriers between the valley towns. Large villages and populous cities have grown upon the lowland; railroads traverse it lengthwise and crosswise; agriculture thrives, as well as manufactures. While the uplands—"the land of the lingering snow"—have a rigorous and forbidding climate, the lowland yields products, such as tobacco and peaches, that one does not ordinarily think of as associated with the farms of New England, but refers instead to a region of milder climate. The harvests of the lowland certainly can not compete with those of the western prairies, but the humbler forms of plant life find here so good a home that garden seeds are raised as an important article of trade. The hospitality of the lowland extends to its schools and colleges, which not only educate its own youth, but attract boys and girls from the uplands and young men and young women from all parts of the country. And all these unlikenesses of the lowland to the upland are because the rocks of the one have wasted away, while those of the other have, comparatively speaking, endured. Here is taught one of those simple and impressive lessons of the dependence of the manner of our life on the structure and sculpture of the land.

It is the purpose of this report to lead the reader toward an understanding of the existing geographical conditions of the Connecticut Valley through its geological history. The processes of the distant past must be restored from the records of their action still preserved. The whole must be inferred from its remaining parts. The succession of geological processes must be reconstructed in proper order by reasonable argument until they lead as accurately as possible to the evolution of the conditions now before us. It may thus be possible in the end to perceive how long and how unconscious the preparation has been for the geographical surroundings that control the population of this part of southern New England to-day.

## UPLAND AND LOWLAND.

The area to be here described has no popular name that is precisely applied. The lowland as a whole includes part of the valley of the Connecticut River; but that extensive trough has its beginning far north, between the marginal slopes of New Hampshire and Vermont, and, properly speaking, the river valley departs southeastward from the lowland where the Connecticut enters the eastern upland at Middletown. The southern part of the lowland does not fall within the basin of the Connecticut River, but is drained by several small streams that enter New Haven Harbor. It is only by a geological definition that the lowland can be satisfactorily bounded. As already stated, it includes a belt of country underlain by weak sandstones and shales and bounded by harder rocks on the east and west. The geological name of the sandstones can not advisedly be applied to the lowland for several reasons: First, because, in spite of their strong unlikeness to the rocks of the upland, they have not, as a group, received any popular name from the people of the region; further, no one of the geological names applied to the sandstones and shales is to-day generally adopted. "New Red sandstone" is an antiquated term. "Triassic" is by some regarded as of questionable propriety, because it implies too definite correlation of the valley formation with a certain division of the geological series elsewhere. "Newark," a term introduced expressly in order not to imply definite correlations for indefinite formations, is not yet sufficiently domesticated in Connecticut to serve satisfactorily as a name for the valley lowland or for the title of this report as a whole, however well it may guard against the expression of over confident knowledge. Moreover, these geological names apply to other areas of sandstones and shales, both to the northeast in Nova Scotia and to the southwest in New Jersey, Pennsylvania, and Virginia. In view of these various objections to geological terms and in lieu of any better name, it has come to be my habit to call the district the "Connecticut Valley Lowland," the proper name being taken from the river and the State with which the lowland is associated, and the compound topographic name being intended to separate this open belt of low ground from the narrow valleys elsewhere followed by the Connecticut River. The accompanying sketch map (fig. 1) will make this plain. Between the rugged uplands on the east and west lies the Valley Lowland, stretching from north to south. Its entire length is 95 miles. Its width varies from 5 miles near either end to 15 or 18 miles about the middle. Its area is about 1,000 square miles. Of this total, Massachusetts has approximately a third, and that portion is treated in a monograph by Prof. B. K. Emerson.[1] The remaining 600 or more square miles are in Connecticut, and this portion only is considered in the present report.[2]

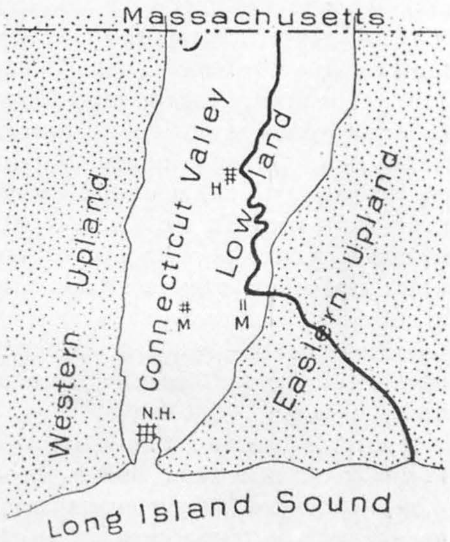


FIG. 1.—Uplands and Lowland of middle Connecticut.

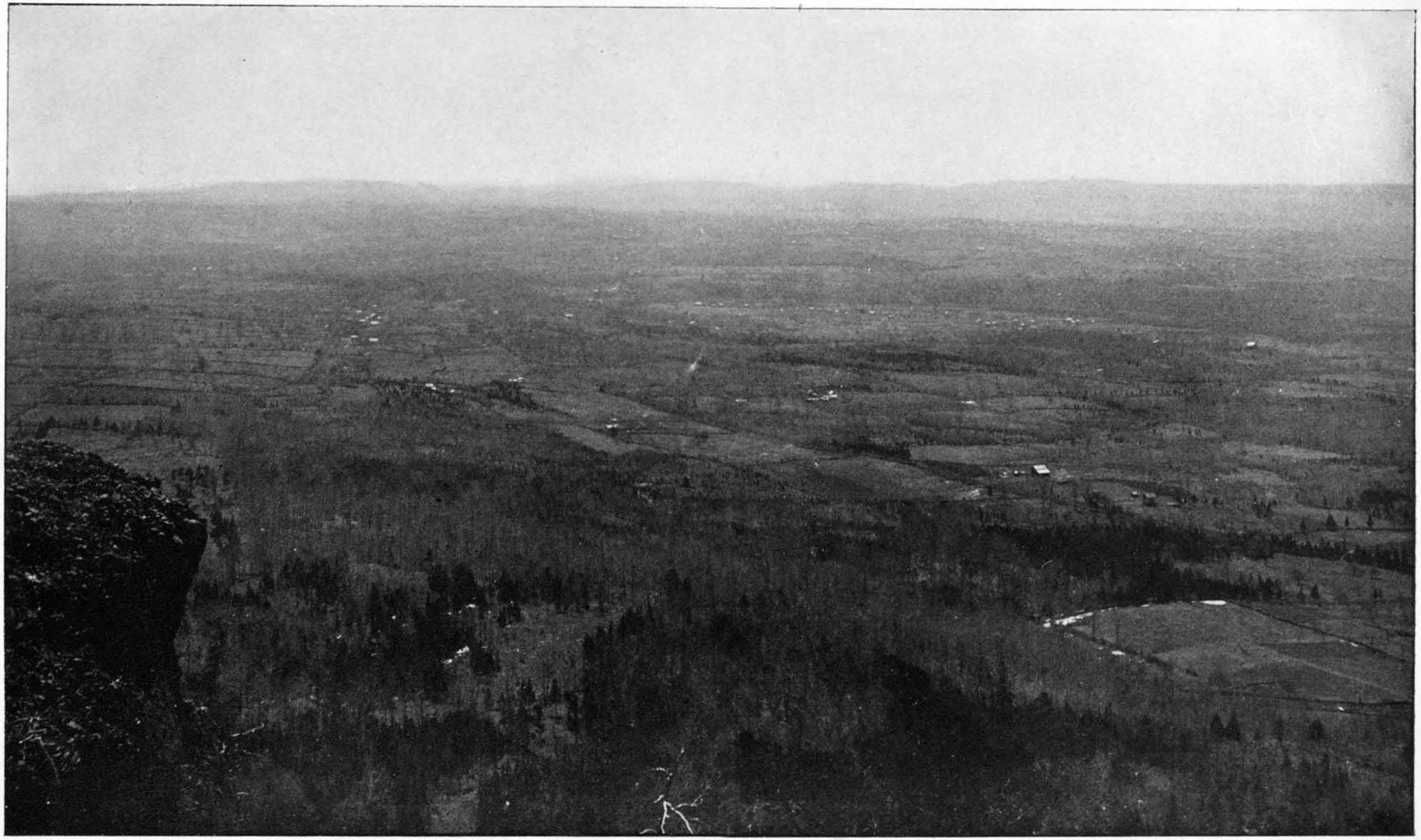
[1] The geology of old Hampshire County, Massachusetts, comprising Franklin, Hampshire, and Hampden counties: Mon. U. S. Geol. Survey, Vol. XXIX; in press.

[2] A small outlying area of sandstones and shales, with associated trap rocks, known as the Southbury area, lies about 12 miles west of the main area, south of middle Connecticut. This was described in the Seventh Annual Report of the Geological Survey.

The uplands that inclose the Valley Lowland on the east and west rise from their low southern margin on Long Island Sound at an average rate of about 20 feet in a mile, thus attaining elevations of 1,000 feet at the northern boundary of Connecticut; then, ascending more slowly, they rise to 1,200 or 1,400 feet at the northern border of Massachusetts. From any of the upland hills the view across the upland discloses a succession of other hills of roughly similar height, still bearing much wood on summits and flanks, thinly inhabited, and more or less deeply separated by valleys. But from a hill summit on the border of the uplands the lowland is seen to lie like a broad trough within well-defined inclosing walls. At the southern border of Connecticut the lowland dips under the waters of Long Island Sound, its middle part underlying New Haven Harbor. At the northern border the average elevation of the lowland floor is hardly 100 feet, and near the northern border of Massachusetts it is only about twice as high. Just as the upland is interrupted by the valleys that are sunk beneath its surface, so conversely the lowland is diversified by the notched ridges that often rise above its plain to the upland height. Unlike the lowland, which is underlain chiefly by sandstones and shales, these ridges are in most cases defined by the outcropping edges of inclined trap sheets or lava beds, trending about north and south, to which a peculiar interest attaches, not only on account of the period of ancient volcanic action that they record in the history of the region, but also because—as will appear in the sequel—it is chiefly by means of the lava beds that the structure of the lowland belt can be deciphered.

On examining the lowland the order and attitude of its bedded rocks first take our attention. When the sandstones and shales of the lowland are observed, their layers are found with few exceptions to dip under ground at a moderate angle of 15 or 20 degrees to the eastward. The lava beds have the same general attitude. It may be inferred from this monoclinal structure that the oldest members of the lowland series, the basal members of the whole succession of sedimentary and volcanic strata, must lie along the western border of the lowland. The youngest members may, on the other hand, be searched for along the eastern border of the lowland, and in a very general way one might expect to pass from the oldest to the youngest members of the series while crossing the valley from west to east. This rule will be later found to submit to many exceptions; yet it may serve for the present to introduce the natural order in which our studies will advance.

Beginning along the western border of the lowland, we must search for the undermost member of the bedded rocks, and discover at the same time the foundation on which it lies. On finding the foundation, we must make inquiry as to its nature and origin, and as to the change in conditions by which a beginning in the deposition of the red sandstone was introduced. The succession of the whole series of lowland



THE VALLEY LOWLAND AND THE WESTERN UPLAND, AS SEEN FROM WEST PEAK, HANGING HILLS.

The upland is the Cretaceous peneplain; the lowland, the Tertiary peneplain.

strata, aqueous and igneous, must then be made out, and the conditions and processes of their formation inferred; thus a general understanding of the period of deposition in the history of the valley will be acquired, and this will occupy the first division of our report. The geological date of the period of deposition must be settled entirely by paleontological studies, which are not here undertaken. The disturbed attitude in which the strata now stand and the character of the deforming forces which gave the present uplifted and tilted position to the entire mass must next be investigated, and this will constitute the second division of the report. The study of the denudation by which the manifest former extension of the inclined strata has been much reduced is finally to be undertaken, and here we shall hope to learn the manner in which the present form of the region has been carved; this forming the third and last division of the report. Briefly stated, these divisions may be grouped under three headings—deposition, deformation, and denudation. A short sketch of the history of geological exploration in the valley will detain us but little from beginning the geological exploration of its history.

## PREVIOUS STUDIES OF THE CONNECTICUT TRIASSIC.

The names of two American geologists, Percival and Dana, will always be closely connected with the study of the Triassic formation in Connecticut. James G. Percival was appointed State geologist in 1835, and during the five years following he accomplished a remarkable piece of geological exploration. His methods were those of an earlier stage of our science, but his accuracy of observation and his perseverance in tracing out the tangles of complicated structures have not been exceeded in later years. His report, published in 1842, is a painstaking record of facts observed in his repeated traverses of the State. His map of the trap ridges is remarkably accurate, and the description of them in the text is detailed in a high degree. Nearly every ridge is located and grouped according to a peculiar system of classification, based on relative positions. Percival recognized the constant relation of the subordinate trap ridges before and behind the larger main ridges, and his terms "anterior" and "posterior" are still in use, although with an enlarged meaning. But he does not seem to have had the least idea that the most of these ridges were outpoured lava beds; all are described as if they were dikes or intrusions. No suggestion is made as to the occurrence of faults in the Trias, although several fault breccias are described in his report under the name of "clay dikes."[1]

Prof. James D. Dana gave much attention to the southern portion of the Triassic belt in the neighborhood of New Haven, particularly to the trap ridges of that district, which he described or referred to in a number of essays. Accepting Percival's recoids and supplementing

[1] Report on the Geology of the State of Connecticut, by James G. Percival, New Haven, 1842, p. 320.

them by local study, he discussed also the theoretical explanation of the observed structures. It is with regret that during the progress of my own studies, as well as in preparing this final report upon them, I have been compelled to dissent from a number of his conclusions. Indeed, had it not been for the great weight of an opinion supported by his name, I should not, either in the field or in this record, have given so much attention to the minute peculiarities attending the contacts of the trap sheets with the overlying sedimentary strata—the so-called "upper contacts"—upon which depends the interpretation of the origin of the trap sheets, and through these the elucidation of the structure of the formation.

In the investigation and literature of the Massachusetts area the names of Hitchcock and Emerson are preeminent; the first in connection with his account of the Triassic formation in the Final Report on the Geology of Massachusetts (1841), and with a large number of independent essays in which the Triassic rocks are described; the second, from the extension of earlier views in the light of more modern methods.

Besides these leading investigators there are many others whose names come to mind in recalling the early studies that opened the way for later work. Beginning as early as 1810, Benjamin Silliman published a number of articles on the Triassic rocks, chiefly in the Journal which became famous under his name. One of these contains a suggestion which a number of years ago seemed to me to have a pointed bearing on more modern work. Referring to the structural relation of the igneous and stratified rocks, Silliman wrote: "I take the liberty to request that those who may have it in their power will make precise observations upon the appearances at the junctions . . . accompanied by drawings and specimens when convenient, and at least with accurate description. We might thus be in a condition to form a general opinion of the origin of our trap rocks."[1]

On entering the valley in 1877 and finding the relations of its traps and sandstones differently explained by various observers, I conceived a plan of work similar to that advocated by Silliman; and when his statement was found, several years later, it was adopted as a text in the introduction to one of my earliest papers on the region.

It was perhaps in response to Silliman's request for local observations that A. Smith and A. B. Chapin published brief essays in 1832 and 1835. Shepard's Report on the Geological Survey of Connecticut, published in 1837, is devoted chiefly to minerals and rocks, with little reference to their geological relations. Closely following the appearance of Percival's report, the Rogers brothers, and soon after them B. Silliman, jr., and J. D. Whelpley, discussed certain structural problems, especially the origin of the monoclinal attitude of the formation.

For a period of about thirty years little attention seems to have been given to structural studies in the Connecticut division of the Valley Lowland, although during this time great interest was aroused by the

[1] Am. Jour. Sci., 1st series, Vol. XVII, 1830, p. 131.

publications of Hitchcock and Deane on the footprints preserved in the sandstones in Massachusetts. In 1871 J. D. Dana published a detailed account of the Geology of the New Haven Region,[1] and shortly afterwards E. S. Dana and G. W. Hawes gave the first descriptions of the trap rocks in accordance with modern petrographic methods. In later years the elder Dana made a minute study of the "Four Rocks" of the New Haven district, J. H. Chapin described the Hanging Hills and other trap ridges around Meriden, Prof. W. North Rice gave an account of the contact of the anterior trap sheet with the overlying sandstones in the gorge of the Farmington River at Tariffville, and Dr. E. O. Hovey, then a student under Professor Dana, described several of the trap ridges east of New Haven.

A convenient means of referring to all these and to many other publications on this subject is found in Russell's compendious bibliography in his correlation paper on the Newark system.[2]

My own studies on this region began in a visit made to the Massachusetts division of the Valley Lowland with the Harvard summer school in geology in 1877, when the unsolved structural problems of the sandstones and lava beds excited my interest and invited further observation. During the summer of 1882 much of my college vacation was given to excursions over the Valley Lowland of Massachusetts and Connecticut, as well as in the Newark area of New York and New Jersey. A few years later several seasons of field work were spent in the Connecticut district for the United States Geological Survey, under the general direction at first of Mr. Raphael Pumpelly and later of Mr. G. K. Gilbert. About the same time close and repeated inspection of the Meriden district was made during successive sessions of the Harvard summer school in geology, which adopted this attractive locality as one of its training grounds in field study under my direction for several seasons. During this time I had as assistants for various brief periods Mr. C. L. Whittle, with whose aid the problem of sandstone and trap contacts was particularly studied; Dr. E. O. Hovey, who examined the structure of the New Haven district; Dr. H. B. Kümmel, who deciphered the details of faulting west and north of Hartford; Prof. W. North Rice, of Middletown, Connecticut, who surveyed his local district with the aid of several of his students (Messrs. Bryant, Graham, and E. L. Rice); Mr. S. Ward Loper, who collected numerous fossils from the beds of black shale; and Mr. L. S. Griswold, who examined the eastern and western borders of the district, traced various fault lines across the lowland belt, and finally revised the more difficult problems of earlier seasons. Assistance in field work was also given by Mr. J. H. Merrill in the Meriden district, Mr. L. J. Westgate around Middletown, and Mr. C. L. Rich in the detached Southbury basin of the Western Upland. While pursuing through the whole course of this

[1] Trans. Connecticut Acad. Arts Sci., Vol. II, 1871, pp. 45-112.

[2] Bull. U. S. Geol. Survey No. 85, 1892.

work a definite object—the solution of the structural problems of the region—the collaboration of these gentlemen has afforded a means of testing in the most thorough manner the various theories that have been suggested to account for the observed facts. As the conclusions reached concerning the relations of the traps to the sedimentaries and concerning the faults by which the formation is dislocated have been almost unanimously accepted by all these workers in the field, I have confidence that the presentation of these conclusions in the following report may be regarded as well based. Dr. Hovey is, I believe, disinclined to regard the trap sheets in all the eastern ridges as overflows preferring still to consider the sheets in the neighborhood of Blanford as intrusive; but with this exception all the observers who have been associated with me have come to the same opinions in the end.

It should be explicitly stated that two important problems have been omitted from the field studies and from the discussions of this report. The age of the formation has not been inquired into; the most recent examination of this vexed problem is to be found in Russell's correlation paper on the Newark system, above referred to. The petrographical composition of the rocks has not been studied as an object in itself, although some attention has been given to it, especially by Mr. Whittle.

The following essays have been published in the form of reports of progress on my own work:


Brief notice of observations on the Triassic trap rocks of Massachusetts, Connecticut, and New Jersey: Am. Jour. Sci., 3d series, Vol. XXIV, 1882, pp. 345–349.

The structural value of the trap ridges of the Connecticut Valley: Proc. Boston Soc. Nat. Hist., Vol. XXII, 1882, pp. 116–124.

On the relation of the Triassic traps and sandstones of the Eastern United States: Bull. Mus. Comp. Zool. Harvard Coll., Vol. VII, 1883, pp. 249–309.

Mechanical origin of the Triassic monoclinal in the Connecticut Valley: Proc. Am. Assoc. Adv. Sci., Vol. XXXV, 1886, pp. 224–227.

The structure of the Triassic formation of the Connecticut Valley: Am. Jour. Sci., 3d series, Vol. XXXVII, 1886, pp. 342–352.

The ash bed at Meriden and its structural relations: Trans. Meriden Sci. Assoc., Vol. III, 1888, pp. 23–30.

The structure of the Triassic formation in the Connecticut Valley: Seventh Ann. Rept. U. S. Geol. Survey, 1885–86 (1888), pp. 455–490.

Topographic development of the Triassic formation of the Connecticut Valley: Am. Jour. Sci., 3d series, Vol. XXXVII, 1889, pp. 423–434.

The faults in the Triassic formation near Meriden, Connecticut: Bull. Mus. Comp. Zool. Harvard Coll., Vol. XVI, 1889, pp. 61–87.

The intrusive and extrusive Triassic trap sheets of the Connecticut Valley (jointly with C. L. Whittle): Ibid., Vol. XVI, 1889, pp. 99–138.

Two belts of fossiliferous black shale in the Triassic formation of Connecticut (jointly with S. W. Loper): Bull. Geol. Soc. America, Vol. II, 1891, pp. 415–430.

The lost volcanoes of Connecticut: Pop. Sci. Monthly, December, 1891, pp. 221–235.

Eastern boundary of the Connecticut Triassic (jointly with L. S. Griswold): Bull. Geol. Soc. America, Vol. V, 1894, pp. 515–530.

The physical geography of southern New England: Nat. Geog. Monographs, Vol. I, 1895, pp. 269–304.

The quarries in the lava beds at Meriden, Connecticut: Am. Jour. Sci., Vol. I, 1896, pp. 1–13.

## PART I.—DEPOSITION.

### THE FLOOR OF OLDER ROCKS.

#### UNCONFORMABLE CONTACT OF TRIAS ON CRYSTALLINES.

The first problem here encountered concerns the relation of the sandstones and shales of the Valley Lowland to the foundation of older rocks on which they lie. Fortunately, the general eastward dip of the strata immediately suggests that the foundation should be looked for along the west side of the valley. But although the western border of the valley belt has been very carefully searched, and although at many places a close approach of the lower sandstones to the fundamental crystalline rocks on the descending slope of the Western Upland has been found, the two are seen in actual contact in only one place. This is in the ravine of Roaring Brook,[1] about 2 miles west of Southington—

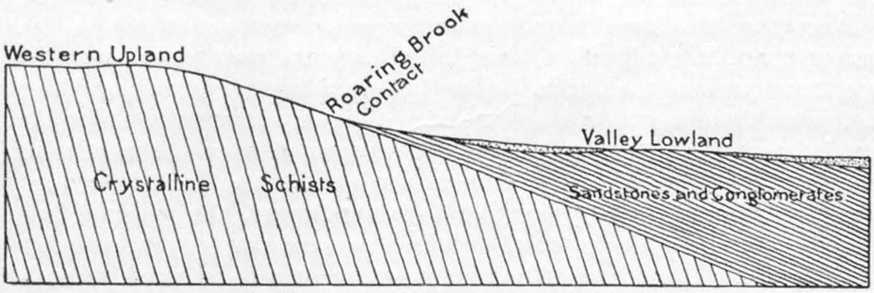



FIG. 2.—Basal contact at Roaring Brook, Southington.

a most interesting locality, perhaps known locally for some time past, but first brought to the attention of geologists by Mr. J. A. Merrill. The exposures are here so clear and fresh as to leave nothing to be desired on that score.

The crystalline schists, with strike N. 20° E. and dip 80° E., are cut by the brook to the depth of 5 or 10 feet beneath the somewhat irregular contact surface of the two formations, disclosing their usual foliated structure. The schists are clean and firm directly to the line of contact, which may be traced for a distance of nearly 100 feet in two unequal exposures, with minor vertical irregularities of 2 or 3 feet. Directly upon this uneven surface lie the sands and pebbles of the oldest known strata of the Triassic formation in Connecticut, fitting their curving

[1] This name is given to several other brooks that, like this one, hasten down the slope of the uplands.

layers perfectly to the inequalities of the floor. Pls. II and III illustrate parts of the contact. The floor is so minutely uneven and so free from all signs of displacement of the upper mass on the lower, that we may be sure that the strata of the pebbly sandstones were not brought upon the schists by any dislocation or overthrusting, but that the sandstones still occupy the position relative to the schists in which they were deposited. Not only so: the pebbly sandstones contain fragments of quartz and schist, some of which may be identified as corresponding to the crystalline rocks in place in their neighborhood; and this gives assurance that the sandstones were made from the ruins of the foundation rocks on which they lie.

The ravine of Roaring Brook is the only place in Connecticut where the sandstones have been found resting directly, undisturbed, on the schists; but the relative attitude of the two formations, where neighboring outcrops are exposed at many other points along the western border of the valley, is such as to indicate that the contact at Roaring Brook may be taken as typical of an unconformity that prevails for a number of miles north and south of its locality. This is particularly true for a part of the boundary, 10 or more miles in length, north and south of Southington. In Massachusetts several ravine contacts have been found by Emerson. They resemble that of Roaring Brook in essential particulars, and thus confirm the interpretation of its local features as the isolated exhibition of a widespread structural relation.

Accepting this conclusion, the linear contact of a few feet in the ravine must, in imagination, be repeated over and over again along the western margin of the lowland, so as to form a belt of contact; and this belt must then be widened by extension obliquely downward under the sandstones where they still remain, and upward over the bordering slope of the Western Uplands where the sandstones have been worn away. Although locally irregular in a small way, the widened belt thus reconstructed forms on the whole a comparatively even, undulating floor. No other form of basement surface can be inferred, when it is recalled that the contact line in the ravine descends eastward at about the angle of dip of the slanting sandstones, and that the western margin of the lowland is comparatively straight, without great indentations and projections along its course, although occasionally deflected where faults have been inferred (p. 123).

The restored floor should be pictured as a peneplain having moderate relief, perhaps several hundred feet in distances of fives or tens of miles; but there are no indications of sharp peaks above or of deep valleys below its general surface. Over the whole of the peneplain the features of the contact in Roaring Brook and in the Massachusetts ravines should be repeated. The truncated folia of the schists should be pictured as forming a generally even foundation, with only minor irregularities. The layers of the pebbly sandstones should everywhere be imagined as fitting closely down upon the slight inequalities of the

CONTACT OF BASAL CONGLOMERATE ON CRYSTALLINE SCHISTS, ROARING BROOK, SOUTHINGTON.

floor, in essentially unconformable attitude. Although now slanting to the east, it is manifest that this position had not been obtained when the sandstones were in process of accumulation. At that time the layers of sandstone and their floor were both essentially horizontal, as in fig. 3, and thus they should be considered through this part of the report.

The lesson of this extensive floor of schists buried under the sandstones is an impressive one, for it tells of a great interval of time between the making of the schists, from some unknown source of supply, and the making of the sandstones from the ruins of the schists. Let us attempt to measure this interval backward from the time when the first pebbly layer of the sandstones was deposited.

The sandstones lie on the truncated edges of the schists. The folia of the latter manifestly had a greater extension upward at some antecedent time, and were denuded to the surface which they now possess before the sandstones were deposited upon them. The time required

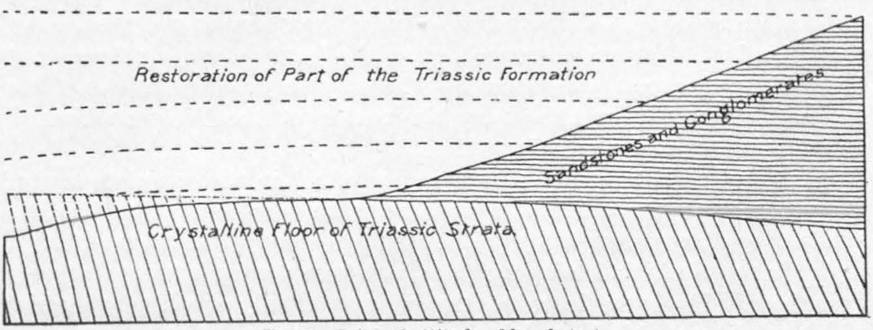


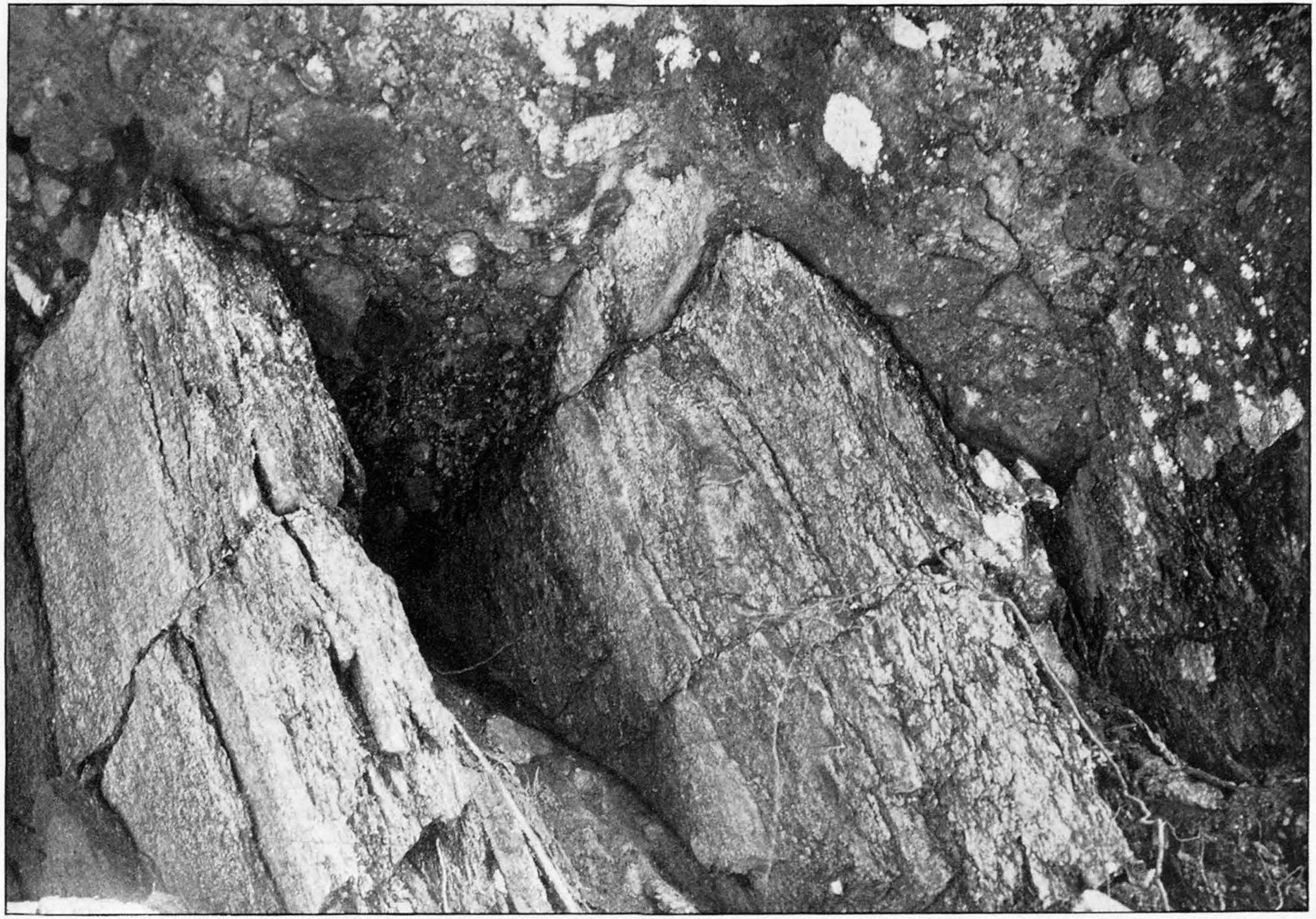

FIG. 3.—Original attitude of basal strata.

for this ancient denudation may be roughly inferred from its amount, and its amount may be roughly indicated in several ways. First, it is to be noted that the schists are metamorphosed rocks, whose mineral composition and arrangement are significantly altered from their original relations. All pertinent observations unite to prove that such alteration takes place only deep within the earth's crust, under conditions of pressure and temperature that are not naturally offered at the surface, but which may be more or less perfectly imitated in the artificial conditions of experiment. It must therefore be concluded that the schists here visible were once buried under an overlying mass of sufficient thickness to permit a physical and chemical rearrangement of their constituents. It can not be supposed that the overlying mass consisted of the Triassic sandstones, for the metamorphism of the schists was complete before the sandstones were deposited, as is proved by the occurrence of fragments of schist in the sandstone beds. It was the upward extension of the schists themselves, or of associated rocks now entirely removed, which once buried the portion of the schists here visible; and this upward extension must have been worn away by the slow processes of ordinary weathering and wasting before the pebbly Triassic sandstones were formed. A long interval must have elapsed while this denudation was in progress.

A better idea of the length of the interval and of the amount of erosion here inferred may be gathered when the general structure of the schists is examined and a reconstruction is attempted of the form that their total mass once possessed. The rocks of the Western Upland, to which the schists of Roaring Brook belong, have a structure of excessive deformation. The deformed area extends eastward through the Eastern Upland, and indeed embraces the larger part of southern and central if not of northern New England. The deformation has been more or less fully deciphered by the studies of Emerson, Wolff, Dale, and others, in Massachusetts, and interpreted as corresponding to the basal structure of a great mountain group, which must in its prime have risen high into the air. Just as we have lofty snow-clad mountains to-day in regions where the greater deformations of the earth's crust have occurred in comparatively recent geological time, so we must imagine noble mountains to have risen here above the area of the uplands in that ancient epoch when the local deforming forces put forth their greatest exertions on the now crystalline rocks. The removal of such mountains from the region of the Western Upland must have occupied a very long time.

This conclusion is still further supported when the completeness of the removal of the ancient mountains is understood. The Triassic sandstones do not seem to lie in deep valleys among ancient mountains, or to abut against their sloping sides, but to rest upon a comparatively even floor. The even floor does not consist of only one kind of schist which might possibly have been worn down low and flat sooner than its neighbors; various schists and gneisses in turn form a part of the basement surface, those of greater resistance to weathering being subdued about as completely as the rest. An even surface or peneplain of this kind, crossing deformed rocks indifferent to their structural attitudes, can have been produced only by the almost complete demolition of the ancient mountains, close down to the sea level of their time.

It must at first seem as if a broad conclusion were here erected upon a slender foundation; it might even be thought that alternative conclusions could be equally well supported on the same evidence; for example, that the beveled schists of the Southington belt represent merely the low flank of a neighboring mountainous area, or that they were worn down low as a result of local, but not of general, base-leveling. But it must be remembered that the beveled schists of Roaring Brook are not at the margin but near the axis of the ancient mountain system to which their folded structures belong; and, further, that the rocks here found so evenly beveled are not weak members of the ancient mountain frame, which might have been worn down low while lofty summits remained about them, but that they are as resistant as any



CONTACT OF BASAL CONGLOMERATE ON CRYSTALLINE SCHISTS, ROARING BROOK, SOUTHINGTON.

rocks in the region, as is fully proved by their resistance to Tertiary erosion (see p. 168). When these important relations of position and endurance are fully realized it will be seen that the beveling of the schists in the Southington belt can hardly be regarded as a local and exceptional result of pre-Triassic erosion, but that they may serve as representing the form to which a large surrounding region had been denuded; a form of moderate relief, with perhaps occasional low residual hills and broad open valleys, but greatly reduced in its old age from the strong relief of its maturity.

A vast length of time must have passed while these ancient mountains were being removed, grain by grain. A long time, in human estimation, is required to excavate the steep-sided valleys by which existing lofty mountain ranges are dissected. An interval, long even in comparison to this long time, would be required to wear down the peaks and ridges between the valleys till they were hardly higher than the valley floors. It was a time interval of this order of dimensions that must have been given to the preparation of the firm foundation on which the valley sandstones were afterwards spread. It was only after this long interval that the depression of part of the region allowed the accumulation there of the waste from elsewhere, and thus opened the Triassic chapter in its history.

The force of this may perhaps be better appreciated after considering a few other examples of buried mountains.

## GEOGRAPHICAL DEVELOPMENT OF THE CRYSTALLINE FLOOR.

Imagine a lofty mountain region, dissected by deep, steep-sided valleys. Let the region gradually subside so that the valleys are slowly converted into bays. The waste from the mountains then accumulates in the bays, slowly filling them with deposits of greater and greater thickness as the subsidence goes on. In time the material supplied by the progressive denudation of the mountain tops might fill the expanding bays, and at last the whole region would be leveled by combined cutting and filling. If opportunity were allowed for observation at any later date a formation of this kind would be recognized by the irregular margin of its deeper strata. Look, for example, at the case of the Adirondacks. These mountains consist of extremely ancient rocks, ancient even when compared to the rocks of our Western Upland. Their flanks are wrapt around by early Paleozoic strata, of which the oldest is the Potsdam sandstone. Pebbles of the Adirondack rocks occur in this sandstone, indicating the same relation between cover and basement that has been described for our Roaring Brook contact. Recent studies of the Potsdam sandstone trace it for some distance into the valleys between the Adirondack ridges. From this it is inferred that the sandstones occupy valleys that were excavated by pre-Potsdam erosion; that after the valleys had been excavated the rugged region subsided; that marine waters then invaded the valleys and received the waste from the inclosing slopes. The process of subsidence and bay-filling may have advanced so far that the mountains were finally lost, or at least reduced to moderate area and height; and the valleys of to-day, where the sandstones enter among the older rocks, are regarded as reexcavated along the lines of the very ancient valleys.[1] The St. Francis Mountains of southeastern Missouri repeat this illustration. A group of ancient mountains of considerable relief is here revealed by the modern denudation of the strata that for ages past have shrouded the mountain rocks from the attack of the atmosphere. The covering strata still lie essentially horizontal, as they must have lain when deposited; they are easily traced as occupying ancient valleys among knobs of porphyry and granite, for the valleys of to-day dissect both the old crystalline rocks and their sedimentary cover indifferently, and the boundary between the old crystalline rocks and the cover is extremely irregular.[2] A lesson of the same kind is taught in middle England, where the New Red sandstone—much less ancient than the sandstones around the Adirondacks and the St. Francis Mountains and roughly equivalent in age to the sandstones of the Connecticut Valley—occupies a great depression between the ancient mountains of Wales and of Lancashire and Derbyshire. The boundary line separating the deformed and denuded rocks of the ancient mountains from the lower strata of the newer deposits is extremely irregular, here entering an embayment among the older rocks, there rounding an ancient mountain spur, and again completely surrounding an isolated mountain group, as at Birmingham. There can be no question that in this case, as in the others, the sandstones were deposited around the irregular base of a subsiding mountainous region.

On the other hand, where the Cretaceous and Tertiary strata of our Atlantic and Gulf coastal plain lie on the older rocks of the Appalachian belt the basement surface is found, as a rule, to possess only moderate inequality; and except for the irregularity caused by modern erosion the inner boundary of the newer formations would be comparatively even. Here it must be inferred that the ancient Appalachians had been reduced to a comparatively even floor before the deposition of the coastal plain strata upon them. The same may be said of the relation of the Cretaceous strata of northern France, where they overlap the once mountainous district of the Ardennes. The basement surface of the Cretaceous formation is remarkably even not only where traced along its margin, but also where detected in the shafts of mines that are sunk through the Cretaceous to reach the contorted coal beds beneath. The great mountainous elevations appropriate to the deformed structure of the buried strata had been worn away before the deposition of the Cretaceous strata.

It is with the latter examples that the basement of the Triassic for-

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[1] Kemp, Bull. New York State Mus., Vol. III, 1895, pp. 330, 340.
[2] Missouri Geol. Survey, Vol. VIII, 1894 (1895), p. 103.

mation in Connecticut best conforms. No great indentations filled with the sandstones are found in the basement rocks of the Western Upland; the western border of the valley is a comparatively even line. The ancient mountains of the Western Upland must have been worn down to a peneplain, or at least reduced to hills of moderate elevation and gentle slope, at the time the accumulation of the sandstones began.

Particularly to the people of the Connecticut Valley Lowland would I commend the local and general evidence on which this conclusion is based. It is too generally a habit to think that while extraordinary examples of geological structure may be found in remote parts of the world they are not to be looked for at home, where familiarity with the hills and valleys brings them to be regarded as commonplace. Too strong a protest can not be entered against this habit of thought. Geological structures, simple and complicated, are strewn with great impartiality over all lands. The section shown in Roaring Brook and its great lesson of mountains made and wasted should become familiar to all the people round about it, and particularly to the teachers and scholars of the neighboring schools.

It is hazardous to say where the border of the continent lay during this ancient time of pre-Triassic denudation. Some of the waste from the ancient mountains may have been carried west and southwest to supply the great volume of sediments now seen in the Catskill plateau and in the mountains of Pennsylvania; some of the waste may have been carried southeast into the hypothetical Atlantic basin of pre-Triassic time; but the geology of to-day does not suffice to define with certainty the goal where the waste from the ancient mountains found a resting place.

In spite of the good reasons for believing that the Triassic strata rest on a generally even basement, it must not be concealed that the opportunity for error in this connection is at least two-fold. It might be suggested that the prevailingly direct course of the western boundary of the valley sandstones in the Southington district is due to its coincidence with a fracture, and that the even contact exposed at Roaring Brook is an exceptional instead of a prevailing structure. To a geologist not acquainted with the district the failure to discover other basement contacts might be construed as arguing against their general occurrence even along the Southington belt; but to one familiar with the prevalence of glacial drift by which actual contacts are so universally obscured, and with the appearance of unconformable superposition at nearly all points where the sandstones approach the schists and gneisses, this suggestion is not acceptable. True, there are certain short spaces along the western boundary of the sandstones that appear to be defined by fault lines, as will be stated in the second part of this paper (p. 123); but these spaces are relatively exceptional. A large part of the boundary is best interpreted as a line lying in the original contact surface.

Admitting that at least the Southington belt of the western boundary marks the original contact, it might still be urged that the present boundary is coincident with the straight flank or base of a mountain range that rose to considerable altitudes over the region of the Western Uplands at the time the sandstones were formed of its waste. This is rendered unlikely when it is perceived that the boundary does not follow any single line of composition or attitude among the ancient rocks. It is true, as will appear in the sequel, that during the time of Triassic deposition considerable movements of elevation and depression must have gone on hereabouts, and that the elevation might have been sufficient to have produced highlands of significant altitude if it were not counteracted by contemporaneous erosion; but this suggestion concerns a later chapter in the history of the region than the initial one now before us. At the beginning of Triassic deposition the region was probably an extensive lowland. If surrounded by hills, they were as exceptional as the monadnocks that surmount the uplands of southern New England to-day.

Here, as in so many other geological problems, we are tempted forward from the safety of known ground into the uncertain ground of speculation. That the Triassic strata rest unconformably upon older strata can not for a moment be questioned. This strong fact is plainly revealed by immediate observation. That a long period of time must have elapsed between the making of the underlying crystallines and the overlying sandstones in such a contact as that of Roaring Brook is a conclusion that stands in the front rank of geological certainties, even though it is only an inference of the unseen.[1] That the reduction of an ancient mountain range to a lowland of moderate relief can be safely inferred from the small contact at Roaring Brook, along with the general relations of the Trias to the schists, is undeniably a large speculative conclusion based on a small quantity of indisputable evidence; and yet it is, of all suggested conclusions, the most reasonable. Hence, taking all things into consideration, it appears most probable that the deformed crystalline schists rose in vigorous mountainous forms long before Triassic deposition began, and that a vast period of time was occupied with the denudation of these ancient mountains and the production of a peneplain of moderate relief in their place. Only after this long period of time was a certain portion of the peneplain depressed, and there the waste derived from the nondepressed portion was spread out, forming the Triassic strata whose remnants we see to-day occupying the length and breadth of the Valley Lowland. This is the larger consideration that comes to mind when pondering over the contact of the schists and sandstones in the shady gorge of Roaring Brook.

[1] The importance of this class of conclusions is ably discussed in an essay by Irving on the Classification of the early Cambrian formations: Seventh Ann. Rept. U. S. Geol. Survey, 1888, pp. 365–454.

## THE TRIASSIC STRATA.

### GENERAL SUCCESSION OF STRATA.

The contact of the sandstones and conglomerates on the schists at Roaring Brook, and the associated low-lying strata along the western border of the formation, have been dwelt on at length because of their extreme importance in the history of the Triassic formation. They illustrate the conditions existing at the beginning of the chapter of deposition. It is, however, very probable that these lowest visible strata are not absolutely the first of their kind; for the local depression of the oldland, by which the long-continued process of denudation was locally reversed to deposition, may have begun somewhat farther eastward than the present western border line of the Triassic trough. Only as depression extended could the Triassic strata cover a broader and broader belt of the oldland.

But accepting the strata now seen along the western border as essentially initial with respect to the great overlying mass, the sequence ol strata which make up the whole formation must be observed, and from this sequence must be inferred the conditions obtaining while the strata were laid down and the character of the adjacent land areas from which the material of the strata was derived. Here, as in the previous section, we begin with what are to-day accepted as incontrovertible facts of observation, and then by degrees enter upon more and more problematical hypotheses. It is in the measurement of the value of these hypotheses that the most critical judgment of the geologist is required, and it may be added that precisely here the differences of opinion among geologists are most marked. After these higher flights of geological theory, the possibility of settling down upon established conclusions might be regarded as hopeless if one contemplated only the present condition of the science; but when one reflects on its rapid emergence from the doubts of the past, faith arises in the certainties of the future.

Among the "incontrovertible facts" of to-day we may fairly include not only the existence of certain bedded sandstones and shales, as directly observed, but also the origin of these strata as ancient deposits supplied from a wasting land and laid down where the agencies of transportation could carry them no farther. Yet hardly more than a century ago this fundamental conclusion of modern geology was questioned by men of experience and reputation. It is the rapid emergence of the successful hypothesis of the last century from the region of debate into the domain of fact that must encourage every reflecting geologist to do his utmost to frame reasonable hypotheses for the explanation of problems still debatable, in the assurance that the coming century will make judicious selection among them, and that the fittest will survive.

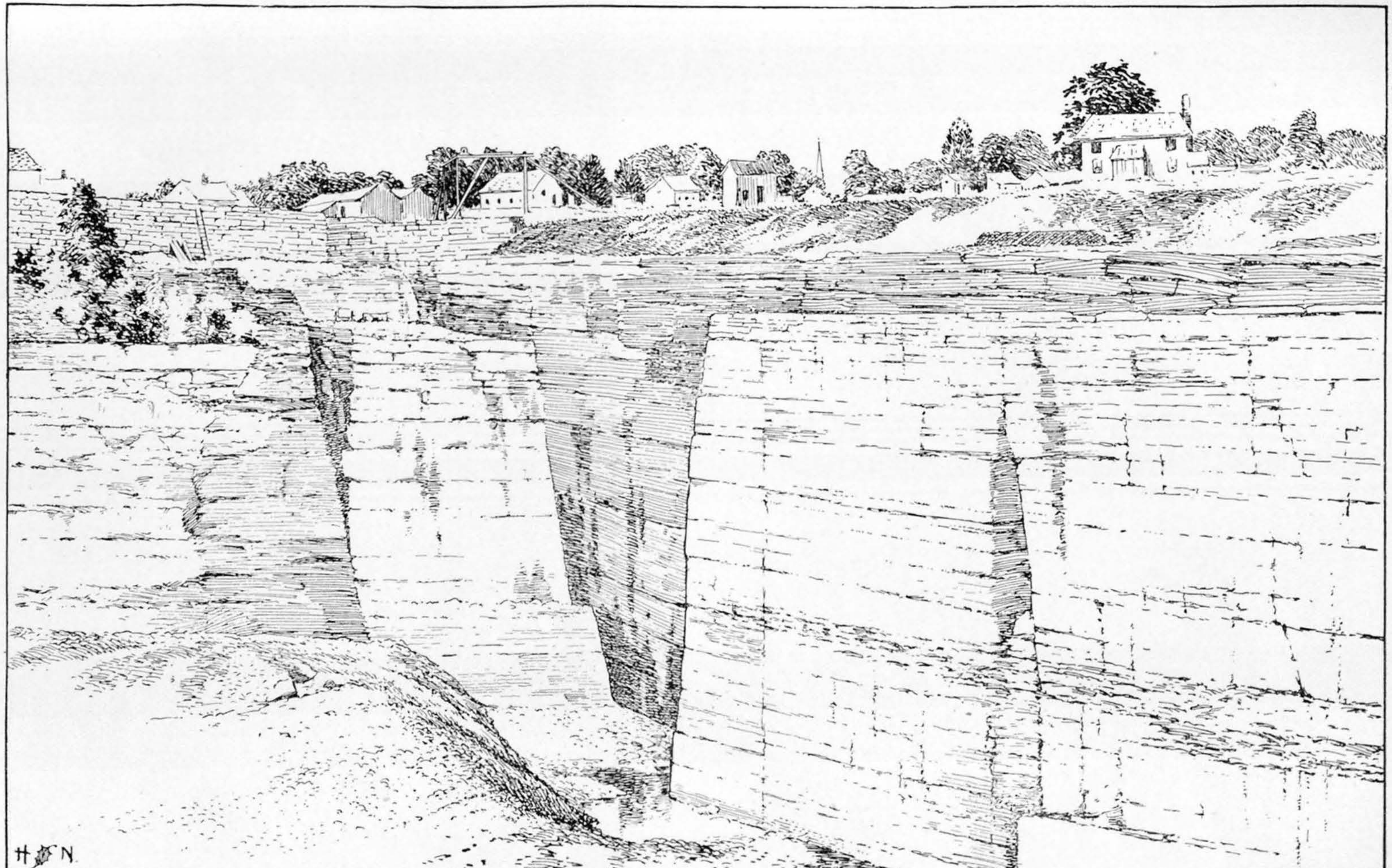
The general succession of the Triassic strata may be discovered by crossing the Valley Lowland from west to east, and thus passing over the successive edges of the tilted beds in ascending order; but, as will appear later, allowance must be made for the frequent repetition of the same beds by faulting. The great extent of drift surface makes it necessary to search carefully for outcrops, and to patch together the observations made on many transverse lines; but in the end a fairly complete series of strata is made out. Truly, if it were confessed that the sedimentary strata which constitute the chief bulk of the Triassic formation are exposed to sight in less than 1 per cent of the entire area of the valley, some readers might object that our pages are occupied too largely with imaginary quantities; but so fairly may the accidental small exposures in streams, quarries, and railroad cuts be taken as samples of the rocks elsewhere buried beneath the drift, and so consistently do they all lead to the conclusions here set forth, that common consent was long ago given to the occurrence of a general eastward monocline in which all the strata are combined; and it is felt that equally common acceptance awaits the statement of the series in which the order of the strata is established.

In a general way it may be said that the lower part of the formation, for a thickness of from 5,000 to 6,500 feet, consists of coarse sandstones with frequent conglomerates and occasional shales. These will all be called the "lower sandstones." Many of the basal layers contain fragments of feldspar, and in general the term "sandstone" can not be taken to imply a composition of siliceous grains only. The grains consist of the waste of granite and other crystalline rocks, and hence are called by Dana granitic sandstones. They are obviously composed of the waste from crystalline rocks similar to those on which they lie.

It is chiefly among the undermost strata, and generally not more than 200 or 300 feet above the base, that the trap sheets of the West Rock Range in the south and of the Barndoor Hills in the north are found. Their thickness may measure 500 or 600 feet, but in certain places, as about Cheshire, they are not so heavy. From the Hanging Hills southward, and somewhat below the middle of the lower sandstones, there are some sandstone layers of rather greater resistance than the rest, thus determining the location of Quinnipiac Ridge (p. 99), and separating the valleys of Mill and Quinnipiac rivers (p. 175).

Following these lower sandstones there are shaly sandstones and shales, with sheets of trap at the base, near the middle, and at the top, constituting the middle division of the formation. The lower trap sheet, called the anterior sheet for reasons that will appear further on, is generally about 250 feet thick, but it thins out and disappears before reaching the northern boundary of the State. Then come the anterior shales and shaly sandstones, 300 to 1,000 feet in thickness. The heaviest or main trap sheet is 400 or 500 feet thick. The overlying or posterior shales are commonly 1,200 feet thick, and are followed by the uppermost or posterior trap sheet, 100 to 150 feet in thickness.

An impure limestone frequently occurs just above the anterior trap



SANDSTONE QUARRIES, PORTLAND.

sheet, as is further referred to on page 189. Fossiliferous black shales occur among the red shales and shaly sandstones between the anterior and main and the main and posterior trap sheets; a special account of these shales is given on pages 137–140. The black shales are occasionally bituminous, and thin coaly seams have sometimes been found in them; but in spite of the hopes that these strata have locally aroused, there is no likelihood that workable coal seams are to be found.

Following this complicated middle division of the formation comes the upper division, consisting of at least 3,500 feet, and possibly a much greater thickness, of sandstones and shales, becoming locally conglomeratic near the eastern border of the lowland, where the uppermost beds are, as a rule, to be found. The most important sandstone quarries of the valley are to be found in this division, between Portland, Connecticut, and Longmeadow, Massachusetts. So large a share of the series, from bottom to top, consists of sandstones that this name is often used in a general sense, as applying to the sedimentary strata without regard to their particular composition.

All taken together, this great succession of strata, constituting the Triassic formation, must have originally extended over a significantly larger area than that which they now occupy; for their edges are manifestly eroded back from their original margin. An attempt will be made in Part III to reconstruct the original formation by restoring its lost parts, and thus to gain an estimate of the full size of the depression in which the formation was deposited.

## CONDITIONS OF DEPOSITION.

All these strata are at present inclined eastward at moderate angles. The science of the eighteenth century did not immediately accept such an inclined position as evidence of disturbance after essentially horizontal deposition. Indeed, it was stoutly argued by eminent geologists of the first half of our century that all the several belts of monoclinal Triassic strata on our Atlantic slope were examples of oblique deposition on a large scale, and that their inclined attitude could not be used as an argument for post-Triassic deformation; but to-day this view is excluded by a variety of arguments, and still more by a large assemblage of facts that had not been discovered when the theory of oblique deposition was advocated. The great extent of the several monoclines is an obstacle at the outset. It is difficult, if not impossible, to imagine any natural conditions in which the great series of Triassic strata could have been deposited as they now stand. They include not only coarse sandstones and conglomerates, whose deposition might perhaps have been locally oblique under fluviatile agencies, but also fine-grained shales, red, gray, and black, all indicating quiet deposition in slow-moving or standing water under comparatively uniform conditions as to depth. These are not to be thought of as having been laid down with their present slant of 10 or 15 degrees, which would have quickly extended each stratum from shallow into deep water.

The theory of oblique deposition would demand that all the sands and pebbles were washed into the ancient estuary from the western side; but along the eastern border there are numerous examples of coarse conglomerates dipping eastward, whose pebbles, cobbles, and bowlders are largely derived from the rocks of the Eastern Uplands. For example, south of Durham the material of certain conglomerates is in great part supplied from the greenish or gray gneisses of the hills next eastward. From Portland northward white quartzites and garnetiferous schists of the Eastern Uplands are often represented in the marginal conglomerates. Moreover, the coarsest conglomerates of the valley are not found at the western margin, but near the eastern margin, the very coarsest known in the State being at North Guilford, with strong dips toward their source, and therefore undeniably disturbed. Furthermore, as Hitchcock long ago pointed out, the numerous reptilian footprints, large and small, of which so large a collection is preserved in the museum at Amherst College, testify unanimously that the soft sands and muds over which the Triassic reptiles walked lay essentially horizontal when the prints were made, for the prints are square with the bedding, and not oblique, as they must have been had the reptiles walked along a sloping surface. It might be added that the tracks, when seen in the sandstone quarries, lead in all directions, and not prevailingly along the strike of the strata, as might be expected if the present dip of the beds indicated their original attitude.

The occasional changes of inclination from the prevalent eastward monocline to a local westward dip, and the occurrence of numerous faults, as demonstrated on a later page, absolutely contradict the theory of oblique deposition. The ripple marks, mud cracks, and raindrops, further mentioned below, all testify to the originally horizontal position of the strata that retain them. The minute deposits in the amygdular cavities of the lava beds, described beyond, give additional support, if desired, to the belief that the monoclinal structure results from uplift and tilting of the original formation. It should, however, be recognized that the theory of oblique deposition played a useful part during the exploration of the region. It was well set forth in the search for truth, but it failed to survive the encounter with the growing body of facts.

Leaving out of mind for the present all inquiry as to the process of deformation, and postponing also all inquiry as to the origin of the trap sheets, we must in imagination undo what has been done in the way of tilting, and restore the strata to the essentially level position that they had originally. This is an exercise in which geological students are perhaps less practiced than they should be. Fig. 2 (p. 19) having illustrated the existing condition of a characteristic section across the western border of the lowland, we have now in fig. 3 (p. 21) essentially the same structures, but revolved so as to bring the observed strata and their underground extension into a horizontal position, and at the same time extended to the left, or westward, so as to restore a portion of the formation that has been eroded away. Only a small part of the entire series of strata is here represented.

Certain essential elements of the problem in hand are thus brought more clearly to light. With the reversal of the monoclinal tilting the basement of schists is thrown from its present eastward inclination into a horizontal attitude, as in fig 3. It is prolonged eastward under the strata, descending a little, to suggest that the lowest strata now seen are not necessarily the earliest deposits of the formation. The basement is prolonged westward beyond the present edge of the upland and into the air of to-day, so as to restore a considerable extent of the peneplain of early Triassic time. The Triassic strata are reasonably carried westward toward their original but now lost and unknown boundary. An attempt will be made at the close of Part III to define in a rough way the dimensions of the entire formation; here it is desired only to direct attention to the successive layers, pondering a little while on each of them in order to bring out more clearly the appearance of the region while the strata were accumulating, and to intensify the conception of the period of deposition, over which the mind is apt to pass too rapidly unless intentionally detained.

The change from a long chapter of denudation to a long chapter of deposition is the first problem here encountered. Such a change is ordinarily explained as the result of submergence beneath the sea of a land area that had before stood above sea level. The problem might be more generally stated, in noncommittal form, by saying that conditions of denudation were in some way changed to conditions of deposition; the process of change and the new conditions thus introduced being then sought for, with special reference to such as will provide heavy stratified deposits similar to those found in the Connecticut Valley Lowland.

Marine submergence may certainly claim to be the generally accepted condition of deposition; but during the advance of this century lacustrine, fluviatile, glacial, æolian, and subaerial deposits have also been well recognized, and the mere occurrence of stratified deposits can no longer be taken to determine a period of transgression by the sea. The conditions and characteristics of the several kinds of deposits may therefore be briefly reviewed. It should be noted that in any case there is here involved the supposition that some district near that on which the deposits accumulated was exposed to more active denudation than before, for the supply of plentiful waste, often coarse in texture, from a land previously worn down to a lowland, calls for a more forcible attack of denuding agents than would be possible if it remained a lowland.

If the Triassic strata are of marine origin, a warping of the pre-Triassic peneplain must be postulated, whereby the Triassic area was depressed beneath sea level, while adjacent areas on the east and west were elevated so as to shed their waste actively into the submerged trough. But there is little or no direct evidence for marine deposition of the Connecticut Trias. There are no marine fossils yet found. The fish whose imprints occur plentifully in certain occasional strata of black shale are allied to fresh- or brackish-water forms. The prints of land plants and the tracks of land animals argue against the presence of the sea. The tidal currents that have been assumed to be necessary to carry the materials found in some of the coarser layers may be replaced by other agencies that can as well accomplish this result over the moderate distances here involved. If marine at all, the waters must have been littoral and shallow, and the bottom must have been frequently bared to the sun.

The pre-Triassic peneplain might have been warped by strong deforming forces so as to form a basin in which an extensive lake could have been held. But true lacustrine sediments are of fine and uniform texture over most of the lake bottom, being coarse only close to the shore line. Temporary lakes of shallow water might account for some of the beds of finer shales, but not for the great body of the Triassic formation.

The pre Triassic peneplain might have been warped so as to alter the action of the quiescent old rivers that had before flowed across it, yet not to drown or to pond them. Such a change would set the streams to eroding in their steepened courses, and to depositing where their load increased above their ability of transportation. As with marine or lacustrine deposits, the thickness of the strata thus produced would depend on the duration of the opportunity for their deposition. A progressive warping, always raising the eroded districts and depressing the area of deposition, would in any of these cases afford the condition for accumulating strata of great total thickness. The heavy accumulations of river-borne waste on the broad plains of California, of the Po, or of the Indo-Gangetic depression all agree in testifying that rivers may form extensive stratified deposits, and that the deposits may be fine as well as coarse. They are characteristically cross bedded and variable, and they may frequently contain rain-pitted or sun-cracked layers.

Postponing the consideration of glacial action to a later paragraph, there remain the wind-borne or æolian deposits, such as accumulate in and to leeward of arid regions, and the subaerial deposits that creep or wash forward from mountains on piedmont slopes. The former are found in great volume over the interior of China, as described by Richthofen, but as they are universally of very fine texture they can not be paralleled with our Triassic deposits. The latter are well known in the arid basin of Utah and Nevada and in the desert basin of Persia—coarse around the margin, fine and often saline about the center. They are prevailingly unfossiliferous, not only because they have few inhabitants, but also because whatever remains fall upon them are long exposed to the weather before being buried.

In contrast to marine deposits, Penck has suggested the name "continental" for deposits formed on land areas, whether in lakes, by rivers, by winds, under the creeping action of waste slopes, or under all these conditions combined. This term seems more applicable than any other to the Triassic deposits of Connecticut. It withdraws them from necessary association with a marine origin, for which there is no sufficient evidence, and at the same time it avoids what is to-day an impossible task—that of assigning a particular origin to one or another member of the formation. A continental origin of the formation would accord with Dana's conclusion that the Triassic beds "are either fresh-water or brackish-water deposits."[1] There may possibly be included an occasional marine deposit along the axis of the depressed trough, for at one time or another a faster movement of depression than usual may have outstripped deposition and thus caused submergence; but, in the absence of marine fossils, the burden of proof must lie on those who directly maintain the occurrence of marine deposits.

It requires a conscious effort to picture the geographical conditions that must have long prevailed within and around the depressed area in which the strata were accumulated. The bedding planes of the strata, revealed only in scanty exposures in which the Triassic strata are generally worn across their edges, must in imagination be transformed into broad floors of washed sands and pebbles, derived from a land area on the west or east, and gradually drifted from the margin toward the middle of the trough, where they accumulated. For every grain remaining in a sandstone bed thousands of grains must have gone past it, slowly moved by transporting agencies, slowly worn finer and finer. Every layer seen to-day is more a witness of transportation than of deposition. The sands were not washed directly to their place of settlement and there at once deposited; they were gradually moved along the water floor. The finest silts may have been actually carried in muddy lacustrine or estuarine waters, but they must have been many times laid down and taken up before finding a final resting place. The coarser beds are to-day generally found near the margin of the formation, on the east or west; but sometimes pebbles or cobbles up to 6 or 8 inches in diameter are found near the medial axis of the lowland, as north of Meriden. Such strata may be taken to indicate a more than ordinary activity of transporting forces in the middle of the depressed area, probably during a time of less rapid depression of the region than usual, and an encroachment of the coarser marginal deposits on the flatter surface of the finer sediments along the middle of the trough. On the other hand, fine-textured and evenly laminated shales sometimes occur close to the border of the lowland, as in the curves of Pond and Totoket mountains, north of Branford; and these must be taken to indicate periods when the tranquil middle waters reached over a broader area, probably because of an increase in the depth and breadth of sub-

[1] Manual of Geology, 1880, p. 409.

mergence, and possibly because of a less active supply of sediments from the adjoining lands.

Every pebble, every grain of sand, every particle of silt, is best understood as having been made in the manner of to-day—detached by weathering from the adjoining land surface and moved downhill in the creeping soil cap, carried down valleys by the wash of streams, or drifted from deltas into shallow lakes. There may have been storms and floods then as now, but there is no sufficient reason for supposing that Triassic time was significantly more unquiet than is the present. As the hills of New England are weathering and wasting before our eyes, as the streams are flowing down their valleys, so can we best picture them on the ancient New England of Triassic time. Only in this deliberate manner must the accumulation of one stratum on another be imagined. Even the coarse bowlders of the marginal conglomerates, such as occur inside of the Pond and Totoket crescents, must have long rolled along the water courses from their source, and must have witnessed the passage of a great volume of finer materials over their heads before they were finally buried.

Cross bedding and ripple marks are among the commonest of the detailed structures in the Triassic strata. Hence we are assured the sediments did not advance by a steady and rapid movement from margin to center, but by an intermittent migration, settling down for a time and perhaps buried to the depth of a foot or more, only to be uncovered and drifted along again in the general line of progress. Deposition, on the whole, prevailed; but at any one point the deposited material can have been only a small residue of the transported material. Ages and ages must have passed, every day of which had its deliberate dawn and close, every year of which must have shown only such minute changes as are now to be witnessed in the wearing down of uplands and in the filling of lowlands; yet in the end the Triassic strata grew to be two miles thick. No just appreciation of this great labor of wasting and building can be gained by rapidly walking across the valley of to-day. The slow changes of the past must be deliberately dwelt upon if they are to be clearly conceived.

## DEPRESSION AND DEPOSITION.

Perhaps the most notable feature of all this time is that, in spite of the great depth of the Triassic basin as measured in the total thickness of its strata, by far the greater number of layers give clear evidence of having been deposited by streams or in comparatively shallow water; thus repeating the geological paradox, encountered in many other regions, of a shallow basin always filling and yet never becoming filled. The coarse texture of the sandstones that make so large a share of the whole series of deposits, the cross bedding and ripple marks in the sandstones, the ripple marks, mud cracks, and rain prints in the finer layers, and the innumerable footprints of large and small reptiles in strata at various horizons in the formation are conclusive evidence that where any waters stood they were prevailingly shallow, and that their floor was occasionally laid bare. Yet in these shallow waters strata measuring thousands of feet in thickness were deposited.

The accumulation of a great series of strata under such conditions requires the supposition that a gradual depression of the trough must have been in progess during all the time of deposition. The trough could not have been depressed to its full depth before the deposits began, and then gradually filled to the surface. On the contrary, the depression of its floor must have gone on at about equal rate with the accumulation of sediment, for only in this way could fluviatile and subaerial conditions or shallow waters have been maintained all through the period of deposition. It must be carefully recognized that the depression thus inferred was of the most gradual kind. From the bare term "depression," one might at first gather an idea of subsidence at altogether too rapid a rate, picturing it as advancing visibly. On the contrary, could the process have been watched, it would have wrought no perceptible change from year to year; and if it had been followed from century to century there is sufficient reason for thinking that its effects would not have been more noticeable than are the changes of level on the coast of Scandinavia to-day.

It should not be supposed that there was a perfect uniformity in the rate of depression, and that any irregularity of depression was at once compensated by an equivalent irregularity of deposition; but only that, on the whole, the two processes went on together in a rather equable manner, and that at no time was there a deep submergence. The conglomerate beds north of Meriden are best interpreted as marking the advance of active streams to the middle of the trough and indicating a gain of deposition over depression. The shales near the margin in Branford indicate quieter conditions, caused by a gain of depression over deposition; but the great body of sandstones points to a well-preserved balance of the two counteracting processes. From the more permanent inclosing lands, whence the waste was persistently shed, the slopes should be pictured as descending into a shallow trough occupied by broad washes of sand and gravel, crossed by wandering streams. Shallow lakes may have now and then overflowed a middle strip or a greater part of the trough, and there finer sediments would gather. The lake floors were sometimes slowly shoaling, and then invaded by sands and pebbles from the shores; sometimes sinking to greater depth, and perhaps overflowed by a brackish estuary, but tranquil enough to tempt occupation by great numbers of fish, whose forms are still well preserved in the black shale beds near the middle horizons of the formation.

Nothing contributes so vividly to the reality of this picture as the footprints of the reptiles, crossing and recrossing on the sandstones of to-day, the sand flats of the earlier era. The skeletons of these ruling inhabitants are exceedingly rare, only one having yet been discovered in Connecticut; but this, accidentally found in the abutment of a bridge at Manchester, is so well preserved that it excites the hope of finding more.[1] The footprints are, on the contrary, very common in many of the sandy layers. The specimens in museum collections are generally from the Massachusetts division of the formation, where track-yielding quarries have been more actively worked; but judging from the number of prints found by Mr. Loper when making small openings in his exploration of the black shales (p. 138), I am inclined to think that the greater part of the valley is traversed by them, and that it is only from the accident of better exposure to-day along the banks of the Connecticut River at Turners Falls in Massachusetts, and from the activity with which the prints were collected in that district by Hitchcock and Deane, that they are commonly associated with the northern part of the Triassic trough. The prints are generally of even impression, indicating a leisurely stride, as little hurried as the processes of accumulation. All told, they would be counted by millions; and yet this should not be taken so much as indicating a dense reptilian population as a slow accumulation of strata, and perhaps an industrious search for food by the print makers. So with the fossil fish; they are not often found, because the black shales that contain them, very weak and easily eroded, are commonly worn down in valleys and covered with drift; but they must exist in myriads, for almost wherever opened the black shales bear their skeletons and scales on every few feet of surface. Here every print means an individual, while a hundred thousand tracks may have been made by a single reptile. Hence, in spite of the slower deposit of shales than of sandstones, the fish must have been more plentiful in the shale-forming waters than the reptiles were on the sand flats.

The tracks are commonly found in association with strata that are seamed with cracks, as if dried by exposure to air and sunshine, cracks of this kind being often formed at times of low water on the bed or flood plain of aggrading, wandering rivers. Marshes of reeds and ferns must have flourished on the lowlands, and forests of an antique flora must have grown on the surrounding hills, for many prints of land plants, large and small, are found in the shale beds. The forests and the marshes, the uplands and the streams of the trough, received the rain of passing showers, but the raindrops made a lasting record only now and then, when a fitting surface caught their dimpled marks and a gentle overflow quickly covered them with sediment. Not the showers themselves, but the chance of preserving their slight record, was the rarity; and so to-day, not the surface covered by these and other ancient records, but the chance of seeing them, is exceptional. When a stratum is stripped along a bedding plane, as happens now and then in the quarries, the surface should be in imagination restored to its initial horizontal position, extended far and wide over the floor of the

[1] Marsh, Notice of new American Dinosauria: Am. Jour. Sci., 3d series, Vol. XXXVII, 1889, p. 331.

trough, with more pebbles toward the nearer shore and more fine silt the other way, and thus pictured it should be deliberately contemplated. No hurried changes should be introduced, but, as waste weathers, creeps, and washes from the land on which we live to-day, as sediments now grow on flood plains and shallow water floors with the slow-coming waste from the lands, as sea floors subsiding to-day gradually bring about changes of depth and currents faintly noticed from century to century, so in the older times did wearing, washing, and sinking advance with the utmost deliberation. The Triassic scene would have appeared unchanged from day to day, and hardly changed from year to year; it shifted slowly through the countless centuries of filling and sinking. These centuries we are too prone to pass over unconsidered when crossing in an afternoon's stroll the worn edges of the uncounted leaves in this ancient book, whose pages, covered with their slowly written records, are here and there exposed in little patches to our view.

## ORIGIN OF THE TRIASSIC TROUGH.

The relation of the body of sediments in the Connecticut–Massachusetts trough to other bodies of its time can not now be closely determined, but it may be roughly inferred. Its area was certainly greater than the existing Triassic area. The sections given on Pl. XX suggest that the trough was at least 8 or 10 miles wider than the present Valley Lowland (p. 192.) The narrowing of the lowland at its southern end should not be taken as an immediate indication of an original narrowing of the Triassic trough in that direction. The vast time that has passed since these strata were deposited and upheaved may have witnessed great geographical changes, which our Eastern geologists have been apt to underestimate because of their greater attention to the ancient divisions of geological time and their little concern with Mesozoic history. Experience in the West has done much to correct this underestimate and to open possibilities of great changes in land and sea since the Triassic sandstones were formed. It is believed that the longer axis of the original trough coincided essentially with that of the valley lowland of to-day, and that this trend resulted from the control of the trough by Appalachian forces, as maintained by Dana, but the boundaries of the trough can not be closely defined.

The account already given of the basement on which the Triassic strata rest has indicated that it had been worn so low that no great additional amount of waste could be worn from it; and yet from that time onward it yielded great bodies of waste to the sinking trough. The depression by which part of the basement was aggraded in the expanding Triassic trough must therefore have been accompanied by a correlated elevation of the adjoining areas on the west and east, and it must have been from these rising lands that the waste was continually shed into the intermediate depression. Two suppositions may be made as to the character of these correlated elevations. The trough may have been bent down between two arched areas on either side, as in fig. 4, *A*; or the trough may have been faulted down between two uplifted blocks alongside of it, as in fig. 4, *B*. The first supposition finds a parallel in the Gulf of Bothnia, at the head of the Baltic, where the water appears to lie in a shallow downfold of a well-denuded oldland between upraised areas in Finland and Scandinavia. The second supposition finds a modern instance in the down-faulted trough, or graben, of the Upper Rhine between the uplifted blocks of the Vosges on the west and the Black Forest on the east. Both the Gulf of Bothnia and the valley of the Rhine are comparatively young forms; that is, the inequalities produced by their folding or faulting have not been nearly worn down or filled up; only a fair beginning is made toward that end by valley cutting in the uplifted parts and by the partial filling of the depressed spaces with land waste. Examples of further advance in the process

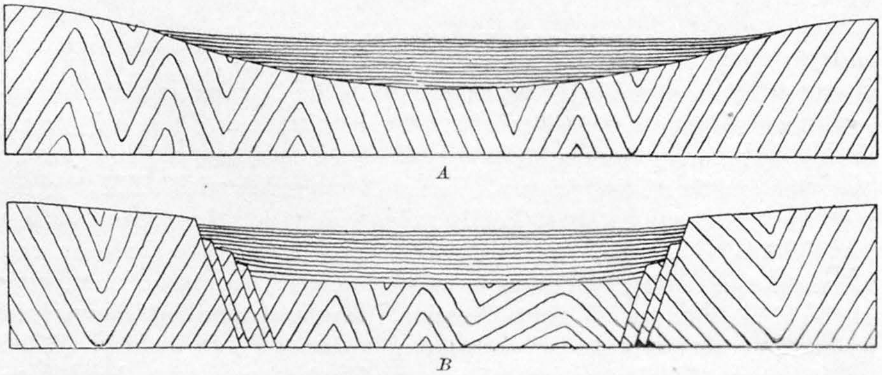


FIG. 4.—*A*, Triassic trough produced by warping; *B*, Triassic trough produced by down-faulting.

of filling may be found in the valley of California, and in any one of the aggraded depressions between the uplifted mountain blocks of Utah and Nevada. The first of these examples represents a down-bent trough into which heavy deposits of waste have been washed from the adjoining uplifted areas; the second imperfectly corresponds to the mature stage of a down-faulted graben.

The diagrams here presented illustrate several peculiarities by which these contrasted processes of trough making could be discriminated if opportunity were only allowed to apply the desired tests in the field; but so much of the Triassic structure has been lost by denudation since tilting and uplift, so few are the exposures of the Triassic strata, and so many significant parts still remain buried deeply, that no decisive choice can be made between the two plans. The lower strata of a trough down-folded without fractures should possess slight centripetal dips when the trough is filled, while the uppermost strata would lie almost horizontal. If the whole mass of deposits were afterwards deformed together, this inequality of dip would still be apparent. A general eastward tilting, for example, with more or less faulting, would result in moderate dips for the lower strata in the east and stronger dips on the west; but the upper strata would have dips of equal value all across the basin. In the Valley Lowland of to-day none of the upper strata are to be found on the western side of the trough, but the lower strata on the two sides, as far as they have been identified, exhibit essentially such inequality of dips as has just been deduced. The dip along the western border is usually 20 to 30 degrees; in the eastern half it is seldom more than 20 degrees, except near fault lines. So far as this goes, it is in favor of a down-folded trough.

A down-faulted trough would be filled with strata of more nearly horizontal attitude throughout. Toward the margin they might frequently overlap small chips of the uplifted mountain blocks on either side, and the dislocations that separated the chips would not of necessity pass through the strata by which they are covered. The marginal strata would be coarse conglomerates, supplied from the cliff faces of uplifted blocks, and the successive strata would abut rather abruptly against the bordering cliffs instead of overlapping a sloping land. In practice it might be difficult, unless exposures were very clear, to distinguish between this original relation of strata and cliffs and the forced contact produced by faulting after the strata had all been accumulated. Thus, to-day, the strata in our lowland seem to abut abruptly against the eastern side of the valley, but all indications point to this arrangement being dependent on a post-Triassic fault, rather than on a pre-Triassic cliff.

It does not seem advisable at present to make final choice between these alternatives; hence the sections of the entire formation presented in Part III must be interpreted as implying much less definite conditions than they actually represent.

The strength of the relief attained by the border lands during the depression of the Triassic trough is naturally a matter of much uncertainty; but except close to the margin, where low cliffs yielded bowlders, cobbles, and pebbles, it may be reasonably limited to a moderate measure. The prevailingly red color of the Triassic strata is best explained as a result of slow and deep weathering in a mild climate, rather than as a consequence of rapid wasting in the cold climate of high mountains.[1] While much of the waste that crept into the depressed trough and formed the lower few hundred feet of Triassic strata may have been weathered in pre-Triassic time, it does not seem possible that nearly all the great volume of Triassic deposits was thus supplied; but rather that, along with the slow deposition and depression of the strata in the trough, there progressed a slow upheaval and weathering of the adjacent lands, and that, as usual on waste-covered lands, the wash of waste from the surface went on no faster than the formation of new waste beneath.

[1]Russell, Subaerial decay of rocks: Bull. U. S. Geol. Survey No. 52, 1889; The Newark system: Bull. U. S. Geol. Survey No. 85, 1892, p. 52.

Contrasted to this supposition is one which attributes the heavier conglomerates of the formation to glacial action, advocated by several authors; but no direct evidence has been found in favor of this view. It is true that certain of the conglomerates are very coarse, bowlders of 2, 3, or even 4 feet in diameter having been seen near Branford; but in no case have these bowlders or any others shown facets or striæ, such as characterize modern glaciated bowlders, and such as are well preserved in the conglomerates of the Salt Range in northwestern India; in no case are the conglomerates unstratified, like glacial till; in no case have the finer strata been found to contain occasional scattered stones and bowlders, as if ice-borne, such as occur in the Simla slates of the Himalaya and in presumably equivalent strata in Kashmir.[1] Certain very coarse conglomerates in Massachusetts have been shown by Emerson to lie close to the ledges from which their bowlders were derived, and thus altogether to dispense with glacial action.[2] The heaviest bowlders in the Triassic conglomerates near Branford, Connecticut, are of trap rock; they seem to lie in the same horizon with a lava flow, and not half a mile from its edge (p. 72). Indeed, to interpret the Triassic conglomerates as records of glacial action practically postulates the impossibility of making conglomerates without glacial aid. In view of the well-known action of waves on rocky shores, and of the ability of streams to carry coarse waste down moderate slopes, it seems hardly advisable to infer glacial action from conglomerates that give no direct evidence of this exceptional process.

## IGNEOUS ROCKS.

### MANNER OF OCCURRENCE.

A peculiar interest is attached to the study of the igneous rocks that frequently rise in ridges over the Valley Lowland. They occur in various associations with the sedimentary rocks. They determine the chief topographic relief of the lowland. They have been diversely interpreted by different observers, and the final adoption of one or another opinion in regard to their origin involves careful observation and critical discussion. They are of the highest value in deciphering the singular dislocations by which the valley rocks are traversed. On all these accounts they have been examined with great care in all relations save one; namely, their petrographic character. This element of study has been reduced to its minimum, because it was not essential to the solution of the chief structural problems in hand, and because it was aside from my training. Described as dolerites or diabases by competent petrographers, the igneous rocks are here generally referred to by the old term, trap, long familiar in the Valley Lowland; or by the common

[1] Oldham, Geology of India, 1893, pp. 133, 135.

[2] Bull. Geol. Soc. America, Vol. II, 1891, p. 453. See, also, Russell, Bull. U. S. Geol. Survey No. 85, pp. 47-53.

term, lava, which has an advantage in certain cases from its suggestion of ordinary volcanic action.

The trap rocks occur in two contrasted forms—either as dikes transverse to the strata, or as sheets essentially parallel with the strata. The dikes are seldom of much topographic importance, the only prominent example of this kind being the stock-like mass of Mount Carmel. The sheets are much more distinct topographically, as they form many long and short ridges of peculiar arrangement. These ridges may be divided into two ranges, a western and an eastern, shown in fig. 5.

### THE WESTERN RANGE OF TRAP RIDGES.

The western range lies for the most part near the western border of the formation, East Rock,[1] in the suburbs of New Haven, and the hills of Granby and Suffield at the northern border of the State, being the most exceptional in this respect. Two subranges may be distinguished; a southern, south of Plantsville, and a northern, north of Unionville. They may be called the West Rock Range and the Barndoor Range, after their chief members. They are distinguished not only by position, but also by habits in the arrangement of their members, long ago noted by Percival. The successive members of the southern group of ridges are arranged in "advancing order;" that is, each more northern ridge stands somewhat to the west of its predecessor, their ends overlapping by a moderate distance. In the northern range this arrangement is reversed to the "receding order." As will appear below, many members of the northern range have a dike-like habit.

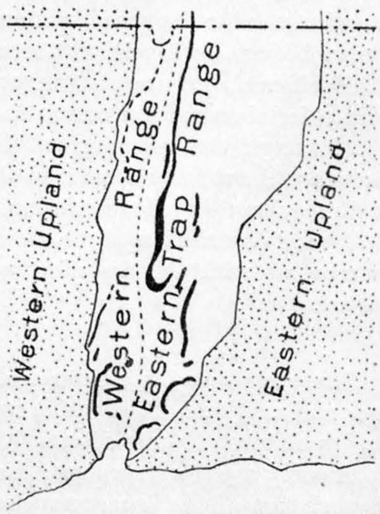


FIG. 5.—Plan of eastern and western trap ranges.

### THE EASTERN RANGE OF TRAP RIDGES.

The eastern range contains ridges of greatly diversified form, size, and arrangement, but throughout the whole series a succession of dominant members is noticeable, with respect to which certain subor-

[1] It may not at first sight appear proper to associate East Rock with the western ridges, where its name would suggest that it did not belong. East and West rocks are, indeed, well named for an observer offshore in Long Island Sound or on the ground between the two, but to an observer who is studying the general distribution of trap ridges in the Triassic lowland, East Rock stands in much closer relation with West Rock than with Saltonstall, Totoket, and other eastern ridges farther north. East Rock is therefore not included here under the group of eastern ridges. It should be clearly understood that the generalizations as to fact and explanation which are applied to the eastern ridges are not to be applied to East Rock.

dinate members are rather systematically placed on the west and east. To these minor and major ridges Percival applied the terms anterior, main, and posterior. As the anterior and posterior ridges generally follow the subdivisions of the main ridge, the arrangement of the latter only need be considered for the present.

The main ridges of the eastern range are disposed in two patterns. In the south two ridges are curved into irregular crescentic form, convex to the northwest, and with the horns of the crescents reaching close to the crystalline area on the east. This peculiar form early attracted attention as a notable peculiarity of the trap ridges; but it is exceptional rather than characteristic of the Connecticut area, unless interpreted in a very free manner, as will appear further on. The other main ridges, to the number of perhaps twenty, are arranged with their anterior and posterior subordinates in the same advancing and receding order that was noticed in the western range, the advancing order being displayed with exceptional strength in the neighborhood of Meriden and the retreating order less emphatically about Tariffville, while the change from one to the other takes place west of Hartford.

Although the meaning of these systematic and peculiar arrangements will not fully appear in this section of my report, it seems desirable to give this brief introductory account of them, in order that the following description of the various trap dikes and sheets may be better connected with the distribution of the topographical forms that they determine.

## DIKES.

There has been little difference of opinion as to the origin of the numerous and manifest dikes which in the southern part of the valley are seen to break more or less distinctly across the bedding of the sandstones in nearly vertical fissures; but some confusion has arisen in the description of the igneous rocks from the occasional application of the term *dike* to intercalated sheets. Thus Hitchcock's first section across the valley, published in 1818, represented the Mount Tom (Massachusetts) sheet as a vertical dike; but this misapprehension was soon corrected. Percival[1] called the posterior sheet of the Totoket Mountain a dike where it has a steep dip at its northeastern end, near Quonnipaug Pond, apparently for no reason but its strong inclination. At a much later date Emerson wrote of the "Deerfield [Massachusetts] dyke and its minerals,"[2] although he explicitly shows the "dyke" to have been "poured out on the lower sandstone and to have been covered by the upper sandstone."

To avoid misapprehension, the term dike will in this report be applied only to those masses of igneous rock that occupy fissures standing in a general way transverse to the bedding of the formation that incloses them, and possessing a considerable length compared to their breadth;

[1] Report on the Geology of the State of Connecticut, by James G. Percival, New Haven, 1842, p. 340.
[2] Am. Jour. Sci., 3d series, Vol. XXIV, 1882, pp. 195-196.

but in connection with these dikes there are local enlargements that resemble stocks or necks, and lateral branches that sometimes imitate the behavior of sheets.

Dikes are numerous from Cheshire and Wallingford southward. They have not been observed in Connecticut north of Meriden. They sometimes follow comparatively direct courses for considerable distances, their continuity being recognized by the low ridge determined by their outcrop, as in the long dikes north of Mount Carmel, and again in East Wallingford and Clintonville. One of the most peculiar features of the long dikes is their frequent upward termination under a cover of sandstone, as if they had never reached higher among the sandstones than they now stand. In other cases they are confusedly arranged in a congeries of irregular outcrops, as about Rabbit Rock, 5 miles northeast of New Haven. These irregular trap ledges are here classed with dikes, not because distinct lateral contacts have been found for every one of them—indeed, contacts are often hard to find hereabouts—but rather because they do not, like the sheets, follow close to the strike of the sandstones, and because they do not exhibit the unsymmetrical outcrop face and back slope that so strongly characterize the outcrops of the inclined sheets. Small sheets often accompany these dikes. Besides these larger exhibits, there are many exposures of small dikes in road and stream cuts, as above Wallingford, whose extension across country can not be traced because they do not keep their heads above the general drift cover.

The dikes are in nearly all cases of dense texture; but Hovey describes a dike in Fairhaven, 23 feet wide, in which a 9-inch band of amygdules runs parallel to the side 4 feet from the southeastern wall.[1] The date of the intrusion of the dikes with respect to the tilting and faulting of the sandstones will be discussed later (see p. 131). The following special accounts are given of the most notable dikes of the region.

*Pine and Mill rocks.*—These strong dikes are described in detail by Prof. J. D. Dana[2] in connection with the sheets of East and West rocks, with which they seem to be closely related. The chief mass of Pine Rock has a length of 2,000 feet and a thickness of about 300 feet; its trend is northeast by east and its dip 50° to 55° NW. Mill Rock is about 3,000 feet in length and nearly 200 feet wide; its trend is about east-southeast and its dip 72° N. About in a line with it, to the east, is another dike, somewhat narrower, but thickening at its eastern end, where it forms a blunt knob known as Whitneys Peak. These two dikes are separated by about 4,000 feet of sandstone; it is highly probable that the dike is continuous beneath the covered space and that the sandstone is only a cover, such as is frequently seen over the dikes east and west of Wallingford. All three dikes are of dense texture, not at all amygdaloidal. The dike rocks are regarded as true

[1] Am. Jour. Sci., 3d series, Vol. XXXVIII, 1889, p. 368.

[2] The four "Rocks" of the New Haven region: Am. Jour. Sci., 3d series, Vol. XLII, 1891, pp. 79-110.

dolerites by E. S. Dana, who finds them essentially free from chlorite and other evidence of interior alteration. They have an imperfect columnar structure roughly at right angles to their walls. The metamorphism produced by the dikes in the adjacent sandstone is of moderate amount, being chiefly limited to a consolidation of the strata for 10 or 12 feet on either side. The walls of the fissures are somewhat irregular, as might be expected from the absence of well-marked joints in the sandstones.

*Mount Carmel and the Blue Hills.*—A heavy mass of trap forms a group of irregular hills extending 3 miles from west to east between Mill and Quinnipiac rivers, and measuring nearly a mile in breadth. The western hills, of which Mount Carmel is the most prominent, exhibit a large area of trap, but in the Blue Hills, farther east, the trap is largely covered by a thin blanket of sandstones, in domelike forms, more or less disturbed and indurated, and suggesting by their prominence above the sandstones of the Valley Lowland that the trap is not far below the surface. This is confirmed by the occasional appearance of the trap on the summit or flanks of the sandstone domes. Where the trap is somewhat more continuously exposed its form resembles that of the sandstone domes, and suggests that it has been revealed by the stripping of the sandstone cover, with little erosion of the trap itself. The trap is dense throughout. Near the sandstone cover or margin it is finer grained; but where more deeply eroded it is of coarser texture, somewhat like the trap commonly occurring in the ridges. Side and top contacts are found where the trap breaks across the sandstone beds, but the sandstones as a whole maintain their usual eastward dip. Close to the contacts the sandstone is almost fused to the trap. Successive intrusions are indicated by the occasional occurrence of dikes intersecting the greater igneous masses.

The boldest member of the Mount Carmel group occurs north of Mount Carmel village and directly east of Mill River. It rises abruptly in a bold bluff above a long talus slope, and forms a landmark easily recognized from the north or south. The long Cheshire dike is directed toward this bluff, but is not seen to be connected with it. A dikelike extension of the bluff is traceable southwestward across the river to a low hill, where it broadens to greater size. This dike is well exposed in a cut north of Mount Carmel station, on the New Haven and Northampton Railroad. It is here seen to contain a number of large sandstone fragments, which are much indurated, and at some points are apparently fused so as to blend with the trap.

There can be little doubt that the Mount Carmel mass represents a heavy stock or volcanic neck.[1] It has every sign of intrusive origin and no indications whatever of extrusion at the horizon now visible. It is, however, entirely possible that while the eastern part of the stock

[1] This interpretation was suggested in my article on the lost volcanoes of Connecticut: Pop. Sci. Monthly, Dec., 1891. The occurrence of volcanic necks or plugs of moderate dimension has been assured in Massachusetts by Emerson's observations: Bull. Geol. Soc. America, Vol. II, 1891, p. 456.

CONTACT OF INTRUSIVE TRAP AND CONGLOMERATE, MOUNT CARMEL.

Trap on the right; conglomerate on the left.

failed to ascend through the sandstone cover, the western part may have risen to the surface of its time and there poured forth some of the extrusive lava flows now seen to the eastward. The intrusive sheets now outcropping west of Mount Carmel may have connections with it underground. The small disturbance of the sandstone monocline in the immediate neighborhood of this great mass of trap distinguishes the latter from laccolithic intrusions, and suggests very strongly that the trap gained its present position by melting its way rather than by pushing its way into the sandstones; yet at the contacts there is slight indication of fusion of the sandstones.

*Fairhaven dikes.*—A group of irregular dikes extends from the east shore of New Haven Bay, at Brightview, northeastward past Fairhaven nearly to Montowese. The dikes frequently form knobs or short ridges, as in Hemingway Mountain. The separate ridges are sometimes arranged in linear order, as in the ridges that run for several miles a little west of and parallel to the anterior ridge of Pond Mountain. The largest dikes measure 100 or 200 feet in thickness. Their walls are generally rather irregular. Side contacts are frequently found in road and railroad cuts, the best being in the cuts lately opened on the new location of the Shore Line Railroad south of Fairhaven. A moderate amount of induration is perceived in the adjoining sandstones and conglomerates; this has been most carefully observed by Hovey.[1] Some of the knobs suggest that the trap takes the form of stocks, as at Rabbit Rock, between Montowese and Totoket. It is highly probable that deeper erosion would connect many of the separate outcrops of to-day.

In the northeastern part of this group it is sometimes difficult to distinguish the dikes from the irregular ledges of the anterior trap sheet, which here has an uneven form, owing to its complicated structure of flow and ash (p. 63). There is much probability that some of the dikes or necks of this district mark the vents up through which the lavas and ashes of the anterior sheet hereabouts were extruded.

*Clintonville and East Wallingford dikes.*—These dikes are of much simpler arrangement than those of the preceding group. Their width varies from 50 to 200 feet. They give frequent exposures of side contacts, sometimes showing fine stringers branching into the sandstones and leaving no doubt of their intrusive character and generally vertical attitude. The adjoining sandstones are moderately indurated. A more peculiar feature is the occurrence of top contacts with the overlying sandstones, from which it is argued that the dikes at such points never rose higher than they now stand. Since the top contacts are found at various altitudes, it is further inferred that the discontinuity of the dikes as indicated on the map is only apparent, and that as erosion degrades the region to lower levels the dikes will become more and more continuous.

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[1]Am. Jour. Sci., 3d series, Vol. XXXVIII, 1889, p. 374. See, also, ibid., 4th series, Vol. III, 1897, pp. 289-292.

The mechanism of the intrusion of such dikes, narrowing toward the top and ending upward in rather blunt edges, is mysterious. They do not appear to occupy fissures that were gradually wedged open from beneath by the force of the intrusion. They seem to have eaten their way between the inclosing sandstones to a certain height, with no sign of a fissure appearing above their abrupt termination. The possibility that the top contacts were produced by lateral overthrusts, which might bring sandstones over the truncated portion of a once higher dike, is excluded by the narrowing of the dikes upward, by the absence of all signs of movement at the contact, and by the induration of the covering sandstones.

The northern end of the East Wallingford dike runs close to the ridge of the anterior trap sheet of Tremont, and the first is with difficulty distinguished from the second, yet there does not appear to have been any eruptive connection between the two. The anterior sheet here is comparatively thin, without lava blocks and ashes, such as might be expected on the line of emergence of a dike, and such as actually occur elsewhere in the same sheet. The close approach of the southern end of the same dike to the anterior ridge about Northford is more apparent than real, for, as will be shown later (see p. 97), there is good reason to believe that a fault passes between the two ledges, with upthrow of over 2,000 feet on the northwest. Farther south the outlying portions of the Collinsville dike approach the anterior ridge, after the habit of the northern members of the Fairhaven dikes.

*Wallingford and Cheshire dikes.*—The intersecting dikes north of the Mount Carmel group may be named after the villages near which they terminate on the east and north; but they may be more briefly referred to as the "crossed dikes." Their structure is, as a whole, very simple and regular; but on the east the Wallingford dike is seen to rise obliquely northward through the sandstones, and locally to assume the form of a nearly horizontal sheet capping a small hill. Elsewhere the lateral contacts indicate a steep attitude for the dike. Top contacts are occasionally noted, as in the dikes of the previous section. The point of intersection is covered with drift. The dislocations in these dikes are described on page 104.

A number of small dikes and intrusive sheets, too small for representation on the map, are found both east and west of Wallingford. They present no novel features, but confirm the intrusive character of the trap ledges hereabouts. A larger dike, forming a low ledge a third of a mile long, is seen west of Brooksvale.

*The Buttress dike.*—This is a unique feature of the region. As described by Percival[1] and Dana,[2] it traverses the ridge of West Rock—the chief instrusive sheet of the valley—and thus, like the dikes in the Mount Carmel trap, indicates at least two periods of

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[1] Geology of Connecticut, 1842, p. 399.
[2] Am. Jour. Sci., 3d series, Vol. XLII, 1891, p. 90.

action among the members of the trap series. The Buttress dike "intersects the West Rock ridge just below the margin of Wintergreen Lake, or about 1¼ miles north of the southern termination of the ridge and 4 miles from New Haven Bay. It descends the eastern slope of West Rock in an interrupted ridge, forms part of the southern bank of Wintergreen Lake, sinks to the level of the West Rock surface at the summit, but stands out like a buttress along the steep west front of the rock."[1] Its trend is east-northeast; dip, 65° N. Its columnar structure runs about transverse to the greater columns of the face of West Rock. The continuation of this dike southwestward into the neighboring crystalline area becomes more nearly vertical. The Buttress dike rock, as described by Hawes,[2] is somewhat unlike that of the other dikes, in containing sparsely scattered anorthite crystals, a fourth to a third of an inch long. It is thus possible to identify its passage into the West Rock ridge.

*Dikes of peculiar composition.*—A peculiar dike, first found by Mr. J. C. Graham, intersecting the posterior sandstones on the outlet of Beseck Lake, about a quarter of a mile west of the Air Line Railroad at Baileyville, has been specially described by Mr. L. S. Griswold.[3] It is dense, dark colored, almost black, and differs from the other dikes of the region in containing many and large phenocrysts of augite, with some hornblende and occasionally biotite, which occupy about a third of the total volume. It is interesting as being the first of a group of basic dikes found in the eastern United States having a geological age determinably later than the Carboniferous. It is locally noteworthy as occupying a higher Triassic horizon than any other well-determined dike in the Valley Lowland.

Another dike deserves mention from its exceptionally acid composition. It is described by Mr. E. O. Hovey as occurring in a new railroad cut in Fairhaven, east of New Haven, but its very small size makes it rather a petrographical specimen than a geological structure.[4]

Some questionable trap outcrops, associated with the posterior ridges of Totoket Mountain, but possibly of dike-like attitude, are described on page 73.

*Dikes in the crystalline areas.*—Percival's report[5] gives a rather detailed account of certain dikes that occur in the area of the crystalline rocks west and east of the Triassic belt. These have been examined at various points, but they have not been found to present features of special interest. They are not represented on the map that accompanies this report, as it was thought their study could be better under-

[1] Dana, loc. cit., p. 90.

[2] Am. Jour. Sci., 3d series, Vol. IX, 1875, p. 188.

[3] A basic dike in the Connecticut Triassic: Bull. Mus. Comp. Zool. Harvard Coll., Vol. XVI, 1893, pp. 239-242.

[4] A relatively acid dike in the Connecticut Triassic area: Am. Jour. Sci., 4th series, Vol. III, 1897, pp. 287-292.

[5] Geology of Connecticut, 1842, p. 412 et seq.

taken with that of the great body of crystalline schists in which they are comprised.

The dikes are of dense texture and steep attitude, like those occurring within the Triassic area. They are commonly from 10 to 50 or more feet wide, being thus large enough to determine small topographic reliefs. Their most significant feature is their occasional interruption or apparent displacement, and this, when further studied, may be found to bear on the extension of the faults that are here described within the Triassic area.

Although now disconnected with the Triassic formation, it is by no means impossible that at the time of intrusion of these dikes the sandstones extended far enough east and west of their present limits to cover the crystallines where the dikes rise; for the greatest distance from the Triassic margin to the dikes is seldom more than 5 miles, although one of the eastern dikes is now as much as 10 miles distant. The true-scale cross sections figured on Pl. XX makes a lateral extension of the original sandstone area for 3 to 6 miles on either side of the present margin highly probable. At the same time, our existing knowledge of the processes of vulcanism does not in the least demand that these dikes should have risen through the crystalline rocks only within the depressed Triassic trough.

## SHEETS.

### EARLY VIEWS AS TO THE ORIGIN OF THE TRAP SHEETS.

The various interbedded trap sheets have given rise to different opinions as to their origin. Most of the early observers were inclined to class them all together as intrusions; apparently not so much because decisive evidence of such origin had been discovered, but because it was conceived that such origin was the more appropriate or natural for inclined sheets of igneous rock. Percival, indeed, recognized that both underlying and overlying sandstone and shale are in direction "very generally adapted to that of the trap ridges they accompany."[1] He described the trap sheets as being very frequently "overlaid by the sandstone on their eastern, more gradual, declivity; thus apparently forming interstratified mass or inclined dikes."[2] Yet there is nothing in the context to warrant the inference that truly contemporaneous interstratification was meant. Such an origin does not appear to have been even considered, for the language of the State report repeatedly confirms the statement made near its beginning—that the trap rocks "have obviously the character of intrusive rocks, of igneous origin;"[3] but as early as 1833 Hitchcock found trap fragments in the sandstones overlying the posterior trap sheet back of Mount Tom, in Massachusetts, and reasonably concluded that at least some of the trap sheets in Massachusetts were

[1] Op. cit., pp. 319, 320.
[2] Ibid, p. 300.
[3] Ibid, p. 11.

overflows. This has been abundantly confirmed by Emerson's detailed work in recent years. Nevertheless, another interpretation was given by Dana to the facts noted by Hitchcock. The generation of steam by the intrusion of molten trap through moist strata was thought to be sufficient to account for what Hitchcock had regarded as evidence of extrusion; and all the sheets of the valley were explained as intrusions of later date than the deposition of the sandstones that inclose them, although not necessarily of later date than the latest sandstones. As this view was presented in successive editions of the Manual of Geology, as well as in various scientific essays, it naturally came to be much more generally quoted than the opinion of Hitchcock, which had no such means of obtaining currency. It is, therefore, hardly too much to say that when the Triassic area of Massachusetts and Connecticut was examined by geologists from elsewhere, it was generally entered with a prejudice in favor of the intrusive origin of the trap sheets.

At the same time the later studies of the Triassic area of New Jersey by the geological survey of that State gave countenance to conclusions similar to those advocated by Dana; for though the Newark or Watchung ridges have since then been shown by Darton to be extrusive flows,[1] their intrusive origin was always maintained by Cook in his official reports.

An important part of my work has therefore consisted of a search for facts that might be quoted as decisive in choosing between the contrasted theories of intrusion and extrusion. From the first it has been held possible that both intrusive sheets or sills and extrusive sheets or flows might occur in the same field, and that the only method of detecting their origin lay in the discovery of significant features in every separate sheet appropriate to the one origin or the other, but contrasted as between the two.

The method of distinguishing between originally intrusive sills and buried extrusive flows is not inherently difficult. Their characteristic features are well known and agreed upon. Opportunity for the observation of these critical features is fairly good, although naturally much less extensive than would be the case if the region were less covered with soil and vegetation. Glacial action has, on the whole, been favorable to observation; for, although the desired lines of contact are extensively covered by drift, the contacts themselves when discovered are often sharp and fresh in firm rock; all the loosened waste of preglacial time having been scraped away by the ice sheet. In this respect Connecticut offers a better field for observation than Pennsylvania, where old soils obscure contacts. The minute examination of the Trias, under Lyman's direction, for the Second Geological Survey of that State has disclosed only one upper contact of sedimentary beds on trap.[2]

[1] Bull. U. S. Geol. Survey No. 67, 1890, pp. 16-35.
[2] Final Report, Vol. III, 1895, p. 2022.

The essentially simple character of the distinguishing features being recognized, it is not too much to say that the divergent views concerning the origin of the trap sheeps are not to be ascribed to any inherent difficulty of the problem, but to the natural diversity of opinion that usually marks the developmental stages of a subject. If Connecticut were now entered for the first time by a geologist of to-day, it is not at all likely that he would hesitate in deciding on the origin of the trap sheets, or that his conclusions would be questioned by his successors. But, as a matter of fact, the study of geology in the eastern part of our country was undertaken when the science was young. We find one early observer[1] discussing the alternative of the aqueous or igneous origin of the trap in the Palisades of the Hudson, giving us one of the few examples in this country of the controversies that the Neptunists and the Plutonists waged so vigorously in Europe a hundred years ago. This stage being soon passed and the igneous origin of the trap being generally admitted, explanation of its different processes of origin had to be invented and the means of recognizing these processes developed. Most of the work from 1830 to 1870 was in this direction. It is probably fair to say that geological argument was in this period less critical than it has now become. Suggestion was then more readily expanded into proof than it could be now. At the beginning of the last quarter of the century the younger geologists, perhaps less influenced by current opinions than were the older geologists who had made them current, might have soon solved the problem of the Triassic traps had their attention been directed to it; but the new West absorbed their time and the Connecticut Valley was little studied.

Not until 1883 were the results of new observations published. In that year Emerson confirmed Hitchcock's belief in the case of the Deerfield (Massachusetts) trap sheet. In the same year my vacation excursions over the Trias of Massachusetts, Connecticut, and New Jersey led me to announce the occurrence of both intrusive sills and extrusive flows. Since then nearly all the newer observations have led to the same conclusion, as will appear in the account of specific localities.

### METHOD OF INVESTIGATION.

The method of discovering the origin of the trap sheets that stand essentially parallel to the bedding of the Triassic strata seems to me to afford good illustration of what has been called "the logical method of scientific investigation," as applied to geology. There is, on the one hand, the accumulation of facts of to-day by observation independent of theory, followed by the grouping together of similar occurrences and the announcement of the results in inductive generalizations. There is, on the other hand, the mental invention, on the basis of previously acquired pertinent knowledge, of appropriate theories which shall give explanation to the generalized results of observation by

[1] S. Akerly, Geology of the Hudson River, New York, 1820.

supplying the unseen facts of the past. Then comes the deduction of all the consequences that follow from each of the proposed theories, with experiments to aid deduction where necessary, and finally the comparison of the generalizations from observed facts with the consequences deduced from the several theories; and according to the measure of agreement found in this comparison the theories stand, waver, or fall. The processes of observation and invention are extremely unlike. They may advance together, but they should be held carefully apart. It is seldom that both processes are equally well performed by one person. Some excel in the observation of the facts of to-day, in which originality, ingenuity, and intuition have little share. Some excel in the invention of theories—that is, in the discovery of the roads that lead to the facts of the past. Conscious observation is here often replaced by unconscious cerebration, on which originality, ingenuity, and intuition so largely depend. Both observation and invention are again unlike the later processes of deduction and comparison, in which the logical and judicial faculties preponderate.

All these processes advance together, interlaced through and through. Invention, deduction, and comparison can not wait until all the facts are gathered. But in the end all the mental processes should be carefully separated and reviewed before a choice of conclusions is announced. The analysis of the entire method of investigation and the conscious application of its several parts to the solution of the problem in hand has been of great assistance, and, indeed, a delight, in the progress of my work.

When the results of an investigation can be stated numerically, it is frequently possible to assign limits to their probable error. It is not possible to gauge the accuracy of geological theories in so simple and definite a manner, but a measure of their accuracy may nevertheless be usefully indicated, this phase of the work touching on the "nature of proof," a general question which has interested various logicians. It may be here briefly said that in geological work the chances of error increase with inaccurate or incomplete observation and with faulty generalization on the one hand, and with irrelevant invention and illogical deduction of consequences on the other hand; and still again with loose comparison where prejudice or habit is allowed to influence judgment. While untested, a theory may be called an hypothesis, so as to warn even its inventor of its unknown value. The importance of entertaining several hypotheses at once, so as to avoid becoming too fond of any one alone, has been well stated by Chamberlin.[1] Hypotheses and conclusions should be held in an open hand and continually subjected to revision in the light of increasing experience. This principle has been practically inculcated in several essays by Gilbert.[2]

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[1] The method of multiple working hypotheses: Jour. Geol., Vol. VI, 1897, p. 837.

[2] The inculcation of scientific method by example: Am. Jour. Sci., 3d series, Vol. XXXI, 1886, pp. 284-299. Also, The origin of hypotheses: Science, Vol. III, 1896, pp. 1-13.

Successful hypotheses gradually acquire the position of accepted theories, and accepted theories in time become hardly distinguishable from first-hand facts. We no longer look upon the aqueous origin of bedded rocks or the organic origin of embedded fossils as matters of theory, although the history of geology shows that these "facts" of our generation were the doubtful fancies of earlier ones. Time is usually an important element in this kind of evolution, as in others; more than a human lifetime is often needed for it. But the investigator is impatient over so long a delay. He naturally wishes to discount the future of his hypotheses and learn at once the value that they will ultimately come to have. His wishes are often vain, yet they may sometimes be gratified. If the numerous and peculiar results of his generalized observations are successfully confronted by the highly varied and specialized consequences of his theory, such an agreement, both qualitative and quantitative, can hardly be referred to chance. If confused occurrences are reduced to order; if many apparently dissimilar facts are shown to be merely natural variations of a simple type; if recurrent and widespread phenomena are rationally explained, all by means of the insight gained through a theory, the theory is certainly to be commended. If after the theory is framed new classes of facts come to light, and the theory, reasonably extended, accounts for them, it is a good theory, for then it explains not only what it was made to explain, but something more. If the theory leads to consequences not matched by the facts previously observed, but perfectly matched by facts subsequently searched for and found under the guidance of the theory, it is excellent, for it endows its inventor with something of the power of predicting the unseen. If, in fine, all these recommendations are enjoyed by the same theory, its evolution toward the grade of certainty is greatly accelerated, and its conclusions may be regarded as stable enough to support other theories. It is manifest, however, that the higher the superstructure thus erected the more remote its upper story must stand from its real foundation on immediate observation. The geology of to-day is full of "facts" that were in the stage of contested hypotheses only a century or two ago. It deals so largely with the conditions of the remote past that it is necessarily a speculative science. But when attention is consciously aroused by an analysis of the different parts of an investigation, and of the dangers as well as of the safeguards of its different steps, the chances of error arising from careless or incomplete work on any of the parts is much reduced, and the chance of determining the correctness of the points at issue is greatly increased.

### DISTINCTIVE FEATURES OF SILLS AND FLOWS.

During the progress of an investigation of the kind here outlined, the external and internal—inductive and deductive—phases naturally advance together. Facts secured from observation and consequences deduced from theory continue to accumulate as long as the work is maintained. At its close, either the observations or the deductions may be presented first, some writers preferring one order, some the other. The deductions are here given precedence, because when they are first in mind the reader can more critically read over the record of observations which follows.

It is conceivable that a trap sheet, lying evenly between layers of stratified rock, might be a contemporaneous volcanic overflow, poured out upon the under layers and buried beneath the upper layers. It is also conceivable that such a sheet might have been intruded between the layers long after their deposition, and deep beneath their uppermost surface. In either case the stratified mass with its included trap sheet might be subsequently tilted, bent, or broken; the composite mass might be more or less deeply dissected and denuded, thus laying bare its internal structures.

If a trap sheet were an extrusive flow it might be accompanied by beds of volcanic ashes, bombs, and lava blocks, thrown some distance from the eruptive vent; the vent itself might be discovered in the form of a fissure, such as prevails in Iceland, where the lavas are poured rather than blown out, or in the form of a pipe, over which a cone might have been built by eruptive violence, now more or less deeply denuded to the form of a volcanic neck or butte. The dimensions of the flow would vary with the amount and rate of supply, with the fluidity of the lava, and with the form of the surface that received it. A pasty lava might form a thick local heap; a fluid lava might spread far and wide in a thin level-topped sheet, and its greatest thickness might be found far from its vent. The texture of the sheet might vary greatly, as does the lava from active volcanoes to-day; it might be dense, scoriaceous, ropy, clinkery, or vesicular; it might be of composite structure, as if consisting of two or more flows in rapid succession. Scoriaceous or ashy texture might be locally developed at points where the hot flow gave rise to explosions during its advance. The lava would be prevailingly of looser texture at the upper than at the under surface; but inasmuch as the under surface is often formed by the rolling down of the front surface, vesicles might be found at the bottom as well as at the top, though as a rule rare in the under part of the sheet. Invading a body of water, it might cause some disturbance by boiling the water. The under contact of lava flows and aqueous strata under water might thus be locally somewhat confused, although on the whole conformable.[1] Some signs of intermixture and baking of the under strata might be expected,[2] although, on account of the poor conductivity of lava, the indurating effects of this kind might be insignificant. The lava flow might be thick enough to fill an arm of the sea, and thus form a land surface; and while exposed to the air the lava

[1] The characteristics of a lava flow on a degrading land surface are not specially considered. In that case the under contact would probably be unconformable to the strata of the land mass. Moreover, a subaerial flow would be exposed to loss by denudation rather than to preservation by burial.

[2] See B. K. Emerson, Bull. Geol. Soc. America, Vol. II, 1891, p. 455.

would rust, weather, and wear. Only after depression and submergence would such a flow receive the sedimentary cover that would more immediately have begun to accumulate on a lava sheet that had not risen above the water surface.

The upper contact of the lava flow with the overlying strata would be unlike the lower in the want of disturbance by rolling and steaming and in the absence of induration by baking. However uneven, clinkery, and vesicular the lava surface, the sediments strewn over it would tend to sift into every accessible crevice, their success in this increasing with their fineness of texture, the deliberation of their supply, and the strength of currents or oscillations in the water. Where the water movements were active loose lava fragments might be rolled about and rounded before they were buried by the invading sediments. The more exposed mounds of the lava surface might be for a time washed bare; the depressions of the surface might at the same time be filled with the loose fragments from about them, intermixed with a greater or less proportion of terrigenous sediments. In time all the lava might be cloaked over, and then deposition would proceed as before the eruption, burying the flow to a less or greater depth.

Many a buried flow might long exist beneath the sediments in a depression, completely inaccessible; but if uplift should raise the depression above base-level, and allow dissection of the uplifted mass by the destructive forces of the atmosphere, the various significant structures above described would be brought to light. Even if greatly deformed, the essential features of the flow might be recognized, as in the porphyries of the central Alps; even if metamorphosed, as in South Mountain of Pennsylvania, they might not escape identification under the microscopic search of to-day.

If a trap sheet were an intrusive sill, it might spread for considerable distances evenly between the horizontal strata that it invaded, but the probability is great that here and there it would break across one stratum or another or send out branching dikes of greater or less size downward or upward. The sheet might gradually change its attitude, becoming more and more oblique to the strata, and thus take on the character of a dike. The dimensions of a horizontal sill are probably less limited by natural restrictions than by our ignorance of the precise process of intrusion and by the difficulty we find in imagining how the intrusion could be carried on. The sill must at some point or other be supplied by a neck or dike, ascending from beneath, and it is entirely conceivable that a sill might be associated with contemporaneous overflows on the surface of the strata above it. Its texture should be prevailingly dense throughout, owing to the great superincumbent pressure, which would tend to prevent the formation of cracks or gas bubbles. Fragments of the inclosing strata might be found within the sill, but no fragments of the sill could be found detached from the mass in the inclosing strata; unless, indeed, the intrusion were so violent as to shatter its walls, in which case a contact breccia might be formed. The inclosing strata might be baked at the upper as well as at the lower contact, and as all the heat of the intrusive lava must pass away by conduction through the adjoining rocks, a greater amount of induration might be expected both below and above the sill than at the lower contact of an extrusive flow. But induration will vary greatly with the composition and moisture of the adjacent beds, and in both cases it is important not to confuse induration by cementation long after the arrival of the lava with induration by heat at the time the lava arrived.

It is possible that many sills may exist beneath the volcanoes of to-day. The regularity of their form will depend largely on the manner in which the country rock yields to the insistence of the lava. All gradations may be imagined, from vertical necks and dikes to branching and irregular dikes, to thin sills, thick laccoliths, and massive intrusions. Here, again, it is probable that the variety of nature exceeds that of our imagination; but during this century knowledge of natural examples has grown rapidly, and the imagination needs to be appealed to less and less as it is replaced by observation. Deeply dissected regions have disclosed dikes and necks, sills and laccoliths, and massive intrusions in great variety.

In reviewing these sets of distinctive features it is worth while to notice that those of the second set are more deductive than those of the first. It is entirely through inference that we reach an opinion as to the structure of an intrusive sill and as to its relations to the adjacent rocks, for the process of intrusion has never been observed. However well we may be persuaded that a sill is, in a given instance, correctly explained as the product of an intrusion, our belief has been reached only through some process of reasoning essentially comparable to that here repeated. Inference may be strengthened by experiment, but very little imitative work of this sort has been done. In the case of extrusive flows many of the distinctive features are directly observable, and are thus in one sense safer than the deduced features appropriate to intrusions. Yet both sets of distinctive features are believed to be trustworthy. They are fully in accord with the extended observations of many persons in various parts of the world. They have been more or less independently reached and accepted by geologists in different surroundings, of different nationalities, and of different habits of thought. If observation of a group of trap sheets discovers a series of facts that corresponds with the peculiar and complicated distinctive features of sills or of flows, there will be as little reason for hesitating to regard the sheets as of one origin or the other as there is now in referring the cold trap rocks to an igneous origin.

The difficulty of making a continuous, linear statement of the problems of the Triassic formation in Connecticut has already been indicated. Actual progress through their advancing solution has been by no means orderly, for questions of deposition, dislocation, and denudation have been over and over again encountered in all possible sequences during the progress of the work; but it would profit no one if the attempt were made to repeat these mental oscillations. The linear sequence of statement is therefore followed as nearly as possible, but at this stage a serious interruption to orderly progress arises. The trap sheets whose origin is here considered now stand in by no means continuous lines of outcrop, but form many interrupted ridges. Yet in Part II it will be shown that these interruptions are of secondary importance, and that before they were formed the trap sheets had a much greater continuity than they now possess. Hence their dislocated fragments must now be treated as if joined together in their initial arrangement, although the arguments leading to this conclusion must be postponed to the second part of this report.

If each ridge, cut from its fellows by an oblique notch, were taken as the edge of an individual sheet, there would be about 130 sheets to be studied within the limits of the State. But it will appear later that nearly all of these ridges may be referred to five sheets, consisting of two chief intrusive sills (which may locally assume the attitude of dikes) and three extrusive flows; and that the apparently great number of separate sheets is really due to the dislocation and dissection of a much smaller number. This important result of the later part of the report is here anticipated for the greater simplicity it gives to the rest of the present part; and being provisionally admitted, we may go on to the following local descriptions, which present such facts as are significant with regard to the intrusive or extrusive origin of the trap sheets.[1]

## RIDGES OF THE WESTERN RANGE.

*East Rock and associated trap hills, New Haven.*—This is a short but heavy monoclinal sheet of trap, interrupted at two points, with bold outcrop showing strong columnar structure on the southwest. At its northern end the sheet assumes a dike-like position, with nearly as steep a face on the east as on the west, as it approaches Whitney Peak at the east end of the Mill Rock dike. The summit of East Rock, reached by excellent roads, forms part of a public park, whence one may obtain an extensive view, including the city and harbor of New Haven, the southern part of the Valley Lowland, with various trap ridges standing above its floor, and limited portions of the Eastern and Western uplands near the coast. Long Island appears on the southern horizon, beyond the Sound.

Several outcrops of sandstone beneath the strong cliff of East Rock are essentially parallel to its under surface. The chief alteration of the sandstones near the under contact is in changes of color. No close induration was noted in their generally rather loose texture. The trap is dense throughout, of coarser texture about the middle, but fine-

[1] Fuller details are given for a few localities in the paper on Intrusive and Extrusive Triassic Trap Sheets, mentioned on p. 18.

grained at the top and bottom. On the northeastern slope, within a few feet of the fine compact trap, there are outcrops of a dense indurated sandstone and conglomerate. No trap fragments are found among the pebbles. Farther down the slope the sandstone soon becomes loose-textured and fragile.

The detailed account of this rock given by Dana[1] explains it as a laccolithic expansion from a feeding dike, whose ascent is inferred from the steepness of the slope along the northeast side of the hill near its northern part. Indian Head, a dependence of East Rock on the south-east, is believed to have a similar structure. In this respect these two rocks may be compared to the Palisade sheet of New Jersey and New York, which Darton has shown to ascend obliquely from a dike along its lower or western side.[2] Snake Rock, separated from Indian Head by a patch of sandstone, seems to be a steeply inclined sheet of small dimensions.

Dana's interpretation of the attitude of the sheet with respect to the

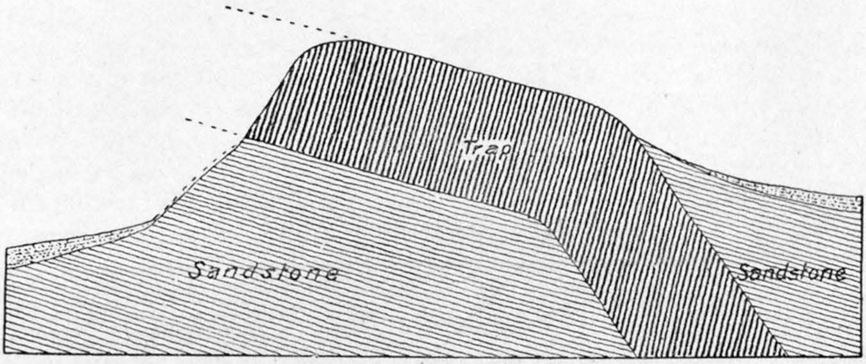



FIG. 6.—Dana's restoration of East Rock.

underlying sandstones, as shown in fig. 6, is based on two suppositions. First, that the upper surface of the rock has not been significantly changed by denudation from its original shape; and, second, that the upper and under surfaces are parallel. Another interpretation of the hidden structure is therefore given in fig. 7. The distinct separation of East Rock and Indian Head is explained by Dana as chiefly the result of stripping of the sandstone cover from two nearly independent laccolithic intrusions. The original continuity of the two trap masses and the origin of the notch between them by Tertiary erosion along the line of a possible fault by which they were dislocated was not considered. Such notches will later be shown to be common in other parts of the valleys (see p. 170), and of possible occurrence here.

*West Rock ridge.*—This is a long monoclinal sheet, with a bold bluff on the west and a long slope to the east, terminating at the south in a

[1] Am. Jour. Sci., 3d series, Vol. XLII, 1890, p. 99.
[2] Bull. U. S. Geol. Survey No. 67, pp. 51-52.

strong curve to the east; this part of the ridge being "West Rock." The sandstones beneath the trap are made of coarse granitic waste, with red and purplish shales. Close to the trap they are described by Dana as "hard baked." At the southern curve of the rock the under surface of the trap sheet obliquely intersects the bedding of the sandstones for about 200 feet, as shown in Pl. VI, this having been originally published by Dana, in his paper already cited. Here the columnar structure of the trap is superbly shown in a majestic bluff, 200 feet or more in height. The trap is very dense throughout, without amygdaloidal structure. Near the base of the bluff and near the upper surface of the sheet on the eastern slope the trap is finer grained than within the mass.

The eastern slope of the ridge, near its southern end, is covered for a short distance above the base by shaly sandstones, which are sometimes exposed in little gullies. The beds vary in color from gray to

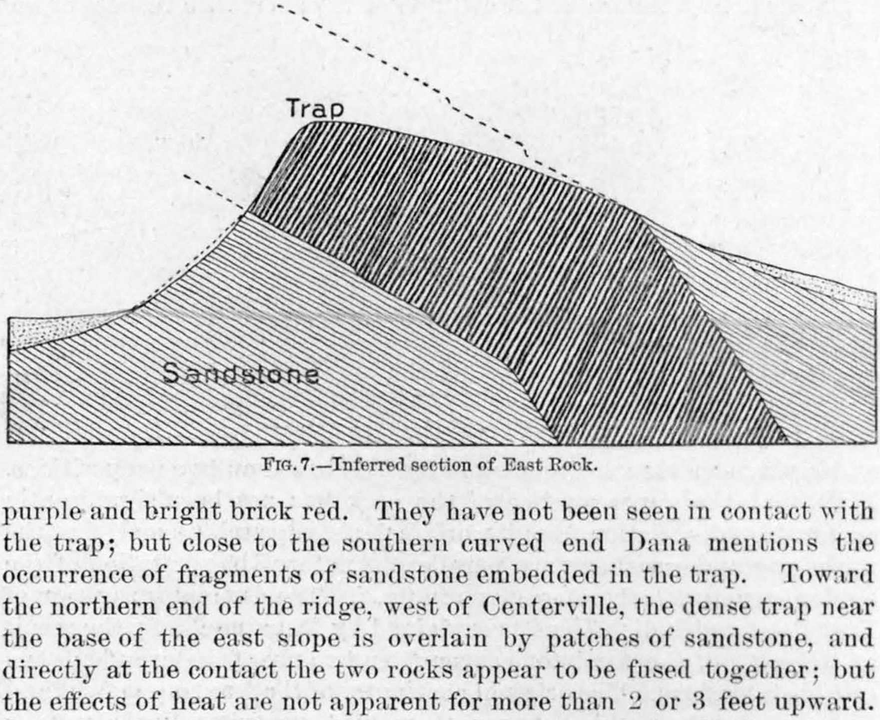
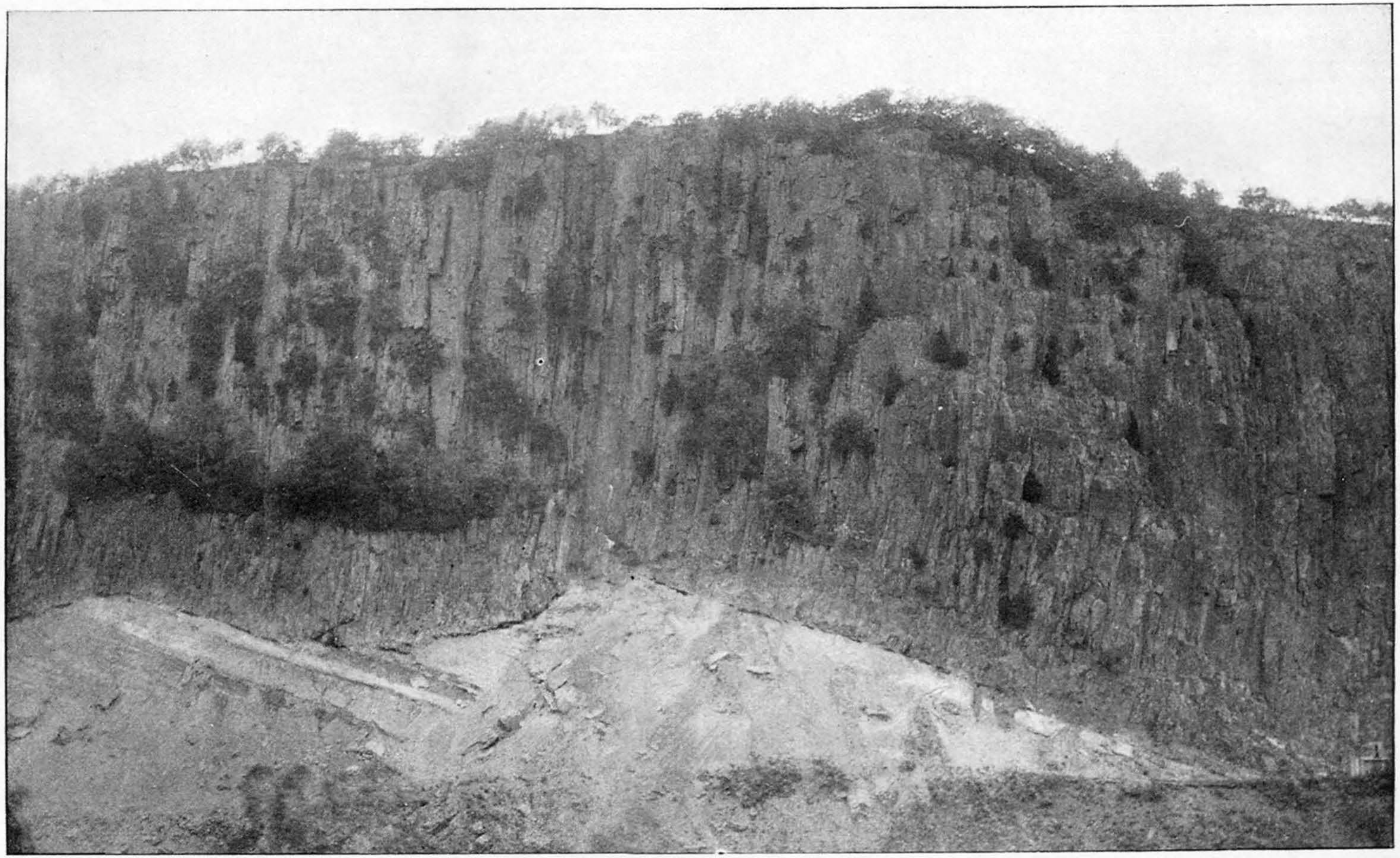



FIG. 7.—Inferred section of East Rock.

purple and bright brick red. They have not been seen in contact with the trap; but close to the southern curved end Dana mentions the occurrence of fragments of sandstone embedded in the trap. Toward the northern end of the ridge, west of Centerville, the dense trap near the base of the east slope is overlain by patches of sandstone, and directly at the contact the two rocks appear to be fused together; but the effects of heat are not apparent for more than 2 or 3 feet upward. The eastern slope of the ridge is sometimes covered with loose trap waste, such as commonly forms the talus at the western bluff of all the larger ridges. This may be the result of a somewhat steeper attitude of the trap sheet than of the sandstone strata, and may in so far be taken as an indication of the intrusive origin of the sheet.

The ending of the trap ridge at its southern hook appears to be the result of the termination there of its intrusion. If its ordinary thickness were maintained farther south, the ridge should not die out. It



SOUTH END OF WEST ROCK, WESTVILLE.

does not appear to be cut off by faulting, for in spite of the excellent opportunity for observation afforded in the bluff and quarries, no significant signs of movement are to be found. The great joint faces, by which the massive sheet is divided into irregular columns, still preserve all details of their original feather fractures, and are entirely free from slickensiding or breccia, such as are found plentifully in the trap sheet of the Meriden quarries (see p. 109). The gulleys that are found on the inside of the hooked end of the ridge, descending northeastward, have much the appearance of small ravines, such as are elsewhere worn down on lines of fault breccia (see p. 111), but here they seem rather to exhibit the form of the trap casting after its initial sandstone mold has been worn away. The outline of the north end of the ridge is also strongly indicative of irregular intrusion among the sandstones, for it is much more uneven than the strike of the neighboring strata, which seem to be cut across irregularly by the trap bluffs; but the termination of the ridge on the north appears to be due to faulting, as will be discussed on p. 132. The traps of East and West rocks have been petrographically studied by E. S. Dana and G. W. Hawes, who pronounce them dolerites, little affected by alteration subsequent to intrusion, and in this respect unlike the trap in several of the eastern ridges.

*Bethany Mountain.*—This name may be applied to a strong trap ridge, about a mile and a half in length, standing between two deep notches, which will later be referred to as the Bethany gaps. The trap here is essentially like that of West Rock ridge, but no critical contacts have been found. The small detached knobs of trap hereabouts will be referred to later (see p. 104).

*Gaylord Mountain.*—This name may be applied to the long ridge that forms the last member of the southern division of the western ridges. It includes Mount Sanford, a culminating summit over 900 feet in elevation. The ridge decreases in height north of Cheshire, and beyond its end there are only a few isolated trap ledges above and below Millvale. The under sandstones are not seen in contact with the trap; indeed, near the southern end of the ridge the trap seems to rest almost directly on the schists, as if it lay at the very base of the Triassic formation. Farther north a narrow valley separates the ridge from the Western Uplands, and here some significant thickness of underlying sandstones may be inferred. The general course of the ridge trends about parallel to the strike of the strata hereabouts, and this, together with the bold western face and the gentler eastern slope, warrants the belief that the trap has the form of a sheet rather than that of a dike; but the occasional occurrence of talus slopes on the eastern side of the ridge suggests that the sheet may lie at a slightly steeper angle than the dip of the adjoining strata.

The trap is dense throughout. The middle of the sheet exposed near the crest of the ridge is, like the trap from the New Haven localities described by Hawes, much less altered than the trap of the eastern ridges. Only two upper contacts with the overlying sandstones have been found. One is in the gorge just north of Mount Sanford; the other in the ravine of Roaring Brook, a mile beyond. At the first locality, alongside the stream near the head of the gorge, a conglomeratic sandstone is seen in immediate contact with dense trap, the two appearing to be welded together. No trap fragments are found among the conglomerate pebbles. Further mention of facts noted here is made on page 133.

Roaring Brook, in the town of Cheshire,[1] gathers a considerable volume of water on the uplands, west of Gaylord Mountain, and has cut a picturesque ravine down the eastern slope. This ravine affords an excellent contact of the trap with the overlying strata; the best known in the southern division of the western ridges, whose total length is 17 miles. It has therefore received careful examination. The trap is dense and without vesicles. At the contact it is fine grained and glassy; the flow of the melted trap is recorded in the arrangement of feldspar prisms parallel to the junction surface. This surface obliquely traverses several beds of conglomerate, sandstone, and shale, but its outline is not nearly so irregular as that of the upper surface of the eastern trap sheets, yet to be described. Slender glassy leaders branch from the upper surface of the sheet into the overlying sandstone. Some of these are mere threads; the largest observed begins with a width of 3 inches and ascends about 20 feet into the overlying strata. The finer-grained adjacent beds are indurated, but are not changed to the extent of developing new minerals; the coarser beds are very little altered, except immediately at the contact. No pebbles of trap are found in the conglomeratic layers, although here, as at East Rock, the ancient waters must have been active enough to have gathered trap fragments, if any such were to be found near at hand.

The farther northern extension of this division of the western range in several short ridges about Milldale has not afforded any contacts of trap and sandstones. The trap remains dense and nonvesicular, as already described.

*Barndoor Range.*—The many disconnected hills that form a notched and interrupted series in the northwestern part of the Triassic area constitute the northern division of the western range. As far as observed, they all consist of dense trap, without the vesicular structure that so constantly characterizes the upper surface of the sheets in the eastern range. The range is heavily wooded, and none of the ridges are large enough to gather streams on their uplands by which lateral ravines could be cut; hence no actual contacts of trap and bedded rocks have been discovered on either side of the ridges; but in a few places an apparently indurated conglomerate has been found near the trap. A notable feature of many of these ridges is the occurrence of a

[1] Not to be confused with Roaring Brook of Southington, where the basal members of the Triassic formation are seen, as described on p. 19.

UNDER CONTACT OF WEST ROCK WITH SANDSTONES, WESTVILLE.

steep talus-covered slope on the east as well as on the west side. In this they differ from the greater part of the southern division, and more emphatically from the whole of the eastern group of ridges. The absence of contrasted form between the eastern and western slopes is taken as indicating a steeper attitude for the trap masses of the Barndoor Range than prevails elsewhere, and as suggesting that they turn somewhat

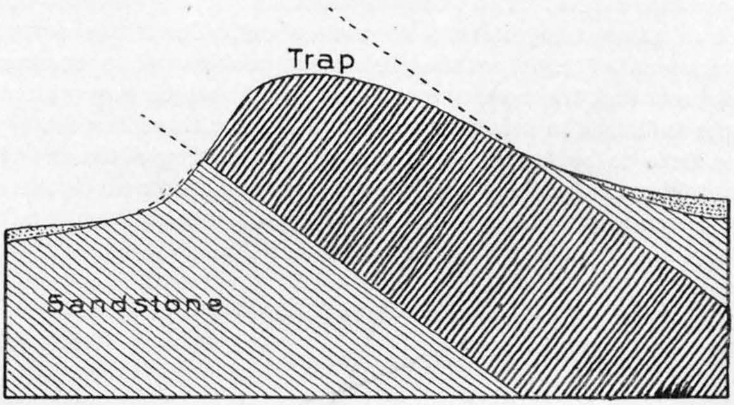



FIG. 8.—Section of oblique sill.

from the type of sheets to that of dikes, as indicated in the accompanying figures (figs. 8 and 9.) It is, indeed, somewhat questionable whether these ridges should be grouped with the sheets or the dikes; but as they are situated symmetrically with the undoubted sheets of the southern division, they are placed provisionally in the former class.

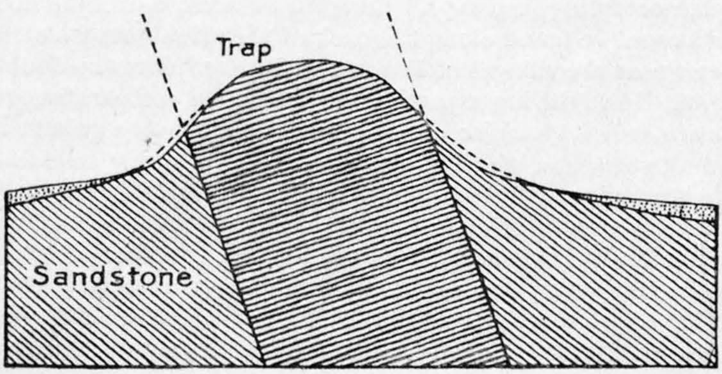



FIG. 9.—Section of slanting dike.

SUMMARY FOR THE WESTERN RANGE.

Although significant exposures are rare, the following characteristics prevail in the western range of ridges (including East Rock, New Haven).

The trap of these ridges is dense throughout. As a whole, the trap shows less hydration and alteration of its constituent minerals than occurs in the eastern ridges. The upper as well as the lower part of the trap masses is free from amygdules. Close to the lower and upper contacts the texture of the trap is finer grained. Although in a general way interbedded among the stratified rocks, oblique passage across the inclosing strata has been detected in the members of the southern division, and a distinctly dikelike habit prevails in the form of many of the Barndoor ridges. The overlying strata are cut by minute or small branch dikes or stringers from the main sheet. The strata do not contain fragments of trap; yet the presence of pebbles in the conglomerate layers shows that the movement of the waters in which they were deposited was sufficient to bring trap pebbles along with others, if any were at that time to be found. While hardly exhibiting distinct fusion at the contact or a significant metamorphism or baking near by, the overlying strata are prevailingly dense and firm close to the trap, and much more so than is generally the case elsewhere.

## RIDGES OF THE EASTERN RANGE.

### ANTERIOR RIDGES.

*Anterior of Saltonstall Mountain.*—The numerous outcrops along the back of the many divisions of the anterior ridges are always highly vesicular. This structural feature is so constant that it will not be separately noted in all the following accounts of special localities.

A cutting made for a roadway at the south end of the Saltonstall anterior, near Easthaven station, Shore Line Railroad, is described by Dana as possessing pipestem vesicles, sometimes 2 or 3 inches long, and often occurring in groups.[1] In spite of much searching no upper contacts could be found along the back of the ridge; but its composite character near the villages of Foxon and Totoket leaves no doubt as to its origin. Here the lower part of the sheet is of comparatively dense trap, more or less vesicular on its back; then follows a greater or less breadth of surface in which irregular knobs of a highly scoriaceous or ashlike trap appear, but of very variable texture. Crystalline pebbles occur in greater or less proportion through the ash; and there is sometimes the appearance of a stratum of sandstone or conglomerate separating the lower, denser lava from the higher ash bed. The irregularity of outcrops hereabouts is not thought to indicate a corresponding discontinuity of trap and ash, but to testify only to their irregular texture and variable resistance to erosion. The greater breadth of surface here occupied by the composite anterior sheet is due in part to its greater thickness and in part to a decrease of dip as compared to the southern part of the same sheet.

The close approach of numerous dikes and necks to the heavy and irregular ash beds of the Saltonstall anterior sheet has already been mentioned as suggesting that surface eruptions may have taken place

[1] Am. Jour. Sci., 3d series, Vol. XLII, 1891, p. 105.

in this locality at that time in the history of the formation represented by the basal members of the anterior shales.

*Anterior of Totoket Mountain.*—The anterior sheet opposite the southern part of Totoket Mountain repeats all the structural features just described. The lower part generally consists of dense trap followed by a more or less distinct separating belt of sandstones or conglomerates, which is in turn succeeded by irregular ledges and knobs of loose-textured ash carrying lava blocks and crystalline pebbles. The last division is very variable; sometimes exhibiting dense trap, as if locally replaced by small lava flows; sometimes with so largely increased a share of pebbles as to gain the appearance of an ordinary conglomerate. After passing a certain space where the anterior sheet is not seen (see p. 97), this composite and variable structure is repeated in the farthest northern outcrops of the Totoket anterior, beyond the northern hook of the mountain; but the anterior of Piscapaug Mountain next to the northwest seems to be a simple and thin lava sheet.

The considerable area occupied on the map by the northernmost part of this anterior sheet by no means represents a continuous exposure of trap rock on the ground, but there seems to be no good reason for subdividing it into trap and sandstone areas. It is best interpreted as an extended area of the lower, denser part of the anterior sheet, here spreading over a larger breadth than usual on account of a faint dip; but sandstone outcrops hereabouts are rare, and this interpretation is open to doubt.

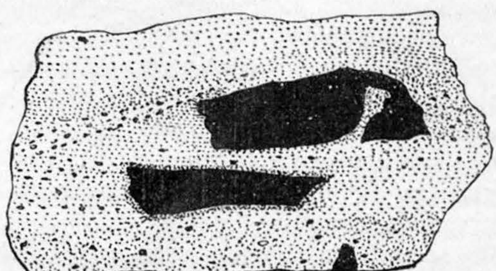


FIG. 10.—Fragments of trap in sandstone, Higby anterior; natural size.

*Anterior of Higby Mountain.*—The long anterior of Durham and Higby Mountain has afforded only two localities where the relation of the trap to the overlying sandstone is revealed. One of these points is a quarter mile south of Black Pond, East Meriden. Here a slight hollow on the back slope of the ridge contains numerous loose blocks of sandstone holding angular and subangular pieces of vesicular trap as in fig. 10.

The second locality is by the roadside nearly a mile south of the north end of the anterior ridge on its back slope. Trap fragments are here again contained in the overlying sandstone, and at both localities a close examination shows that minute clastic deposits occur in many of the vesicles of the trap. Still nearer the north end of the ridge, under its western face, a bed of volcanic "ashes" and trap blocks is found, similar to that in the anterior of Lamentation Mountain, described below.

*Anterior of Chauncy Peak.*—A mixture of "ash" and trap blocks like that just mentioned occurs here also at several points along the western base of the ridge. It was freshly exposed several years ago in the side of the Meriden-Westfield road. During the construction of a reservoir that now occupies the anterior valley between the main and the anterior ridges a great number of fragments of vesicular trap were seen embedded in the sandstone blocks that had been taken from the foundation of the dam.

*Anterior of Lamentation Mountain.*—A bluff of this ridge overlooking the road between Meriden and Berlin, about midway between these places, consists of "ashes" and trap blocks, to which the name "ash bed" has come to be locally applied. The occurrence has excited much interest in the neighboring towns since it was pointed out in 1887, and a well-worn path now leads up to it through the underbrush. The "ashes" consist of fine lapilli, in a bed that here has a thickness of about 30 feet. No stratification is perceptible in the deposit. The lapilli are recognized under the microscope as fragments of fine-grained trap, much altered; chlorite, quartz, and calcite being thus abundantly produced. No grains of waterworn quartz or other clastic material have been noted in the mass. Technically the deposit should be called tuff, if the origin later suggested for it is accepted. The deposit is repeated as far as 1¾ miles to the southeast, at the localities mentioned above. It soon disappears to the north, its place in the western bluff of the ridge being taken for a short distance by a bed of trap conglomerate.[1] Waterworn vesicular fragments of trap are here embedded in stratified sand.

The tuff contains numerous oval blocks of dense trap from 6 inches to 3 feet or more in diameter, giving the face of the bluff a curiously mottled appearance. The blocks are not in the least vesicular after the manner of ordinary volcanic bombs, but are of compact and uniform texture from center to surface. They have no definite arrangement, but are more plentiful near the base of the tuff than higher in the mass. One of them has embedded itself about 6 inches in the underlying sandstone, which is exposed for a short distance at the base of the tuff. The bedding of the sandstone bends downward under the embedded block in a suggestive manner.

The trap sheet above the tuff is more or less vesicular throughout, and particularly so at its upper surface along the back of the ridge. At one point, about opposite to the best exposure of the ash bed, a low knob of trap was blasted and thus shown to contain many layers, films, and strings of sandstone, which frequently penetrated fissures and cavities in the scoriaceous rock to a depth of 2 feet beneath its upper surface. In some cases the sand layers seemed to lie between or to surround isolated blocks of spongy trap. The overlying sandstone conforms closely to the minor irregularities of the trap surface; its color is darker than usual from the presence of trap grains, recogniza-

[1] See Percival, Geology of Connecticut, 1842, p. 365.

BLOCKS OF DENSE TRAP IN FINE TRAP BRECCIA, ANTERIOR RIDGE OF LAMENTATION MOUNTAIN.

ble under the microscope. A few feet higher in the sandstones there are two tuffaceous layers of a rusty brown color, each about an inch in thickness and a foot apart. They contain small waterworn fragments of trap, much weathered and mixed with quartz grains. The microscope shows the fragments to have a vesicular, porphyritic structure; alongside are grains of quartz, orthoclase, and muscovite, presumably derived, like the sandstone as a whole, from the bordering areas of ancient crystalline rocks. Altogether, this is a very instructive locality.

It may be at once stated that two special explanations have been offered for the structures here observed. One regards them as the product of some central vent, from which the anterior lava was emptied, and Mount Carmel has been suggested as possibly representing the lower part of the stock beneath such a vent; but the distance of 10 or 11 miles from Mount Carmel to the Lamentation anterior makes it improbable that blocks of lava 3 feet or more in diameter could have come from that source. The other and more acceptable suggestion, made by Emerson,[1] regards the lava blocks and ash bed as the result of local eruptions caused by the heat of the advancing lava flow itself.

*Anterior of Cat-hole Peaks.*—Two small openings near the base of the sheet, back of the Poorhouse barn 1½ miles northwest of Meriden, expose dense trap masses of extremely irregular form, like ropy flows of lava, the spaces between being filled with a loose material that may be called lapilli. Numerous pipe-stem cavities occur near the bounding surface of the ropy masses, standing at right angles to their curving surfaces.

*Anterior of South Mountain.*—Near the preceding locality, but separated from it by a fault of moderate throw, a small exposure of the sandstone immediately overlying the trap sheet afforded a fragment of waterworn vesicular trap, at a point 100 feet southwest of the Meriden Poorhouse. The western part of the anterior in the same block is cut by a road leading through Reservoir Notch, and shows the sheet to be highly vesicular for at least 10 or 15 feet beneath its upper surface.

*Anterior of Ragged Mountain.*—A few poor exposures on the back of the sheet along the roadside half a mile south of Shuttle Meadow reservoir reveal weathered fragments of vesicular trap in the overlying sandstone. Some of the vesicles in these fragments contain minute clastic deposits.

*Anterior of Rattlesnake Mountain.*—The northern of the two roads that cross the main trap ridge east of Farmington village leads to an instructive exposure of the overlying sandstone resting on the uneven surface of the anterior trap sheet; unfortunately, the outcrops are much weathered. The trap is very vesicular at the surface, where

[1] Bull. Geol. Soc. America, Vol. VIII (for 1896), 1897, pp. 63-65.

some of the amygdules consist of indurated bitumen. The overlying sandstones lie conformably on the uneven trap surface, the sand grains filling the open vesicles. The lower layers of the sandstone consist of an intimate mixture of trap fragments and sand grains. Some of the larger amygdules contain banded deposits, like those described under the Tariffville locality, below. About a mile southwest of this locality, on the west side of a small pond, sandstone is found in the irregular upper surface of the vesicular trap; but the exposure is much weathered.

*Anterior of Pinnacle Mountain.*—The face of a bluff in the anterior ridge, opposite a point midway between Simsbury and Weatogue, exhibits the slaglike, ropy flow structure, such as was described in the anterior of Cat-hole Peaks, but without the loose material (lapilli?) between the separate parts. Numerous pipestem amygdules are found here.

*Anterior ridge at Tariffville.*—This interesting locality was first discovered and described by Prof. W. North Rice.[1] It is one of the most instructive in the valley. The Farmington River here cuts across both the anterior and main ridges. A little south of the river a railroad cut has been opened through the anterior trap sheet. Most of the anterior trap is thus seen to be of dense texture, but near the eastern side of the ridge—the top of the sheet—the texture becomes distinctly vesicular. There the sheet is covered by a thin bed of tuffaceous material, which locally passes along the strike into a bed of trappy sandstone. Above this is a second sheet of trap, of moderate thickness, but no contact with the overlying sandstone is here to be seen. In the more compact trap immediately below the tuffaceous layer, pipestem cavities containing spike amygdules are common, always standing normal to the neighboring surface of the trap sheet. They are several inches in length and commonly a quarter of an inch in diameter. Banded amygdules of oval form are occasionally found, the bands standing parallel to the bedding of the adjacent sandstone. Under the microscope the lower part of the amygdules is found to consist of granular calcite and secondary quartz, stained with iron; the banding appears to be due to variations in the supply of ferric oxide while the vesicle was filling. The upper part usually consists of composite calcite crystals, free from iron. It may be at once argued here that the parallelism of the bands in these amygdules to the monoclinal beds of the Triassic formation shows that, whatever origin is to be attributed to the trap sheet, it must have gained its present position with respect to the sandstones before the latter were tilted. This minute structural detail repeats the occurrence noted by the author in several sheets of amygdaloid in Brighton, near Boston, some years ago.[2]

Certain vesicles close to the surface of the sheet contain grains of

[1] Am. Jour. Sci., 3d series, Vol. XXXII, 1886, pp. 430-433.

[2] Proc. Boston Soc. Nat. Hist., Vol. XX, 1880, p. 426.

clastic quartz and orthoclase along with the calcite filling, the grains being arranged with their major axes parallel to the bedding of the sandstone, from which the above argument may be repeated. This arrangement will be noted for a number of localities to be described on the main sheet. The overlying sandstone contains many water-worn fragments of vesicular trap. The trap fragments in the tuff bed resemble the lapilli of the ash bed in the Lamentation anterior.

Descending from the railroad cut to the river and crossing to the farther bank by the road bridge, one finds a good exposure of sandstone containing trap fragments and lying on the back of the trap sheet.

*Anterior of Peak Mountain.*—The northern end of the anterior ridge lies west of the highest summit of Peak Mountain and a little south of the abandoned copper mine that is locally famous under the name of "Newgate Prison." About 1½ miles farther north two small knolls of hard sandstone are found, capped with trap, in the position that the anterior sheet would occupy if continued. Near the under contact some of the trap exhibits a ropy flow structure; some of it consists of angular blocks, but all are surrounded by a thin layer of sandstone. The local occurrence of these knolls and the peculiar relation of the trap to the sandstone are best interpreted as marking the outermost tongues of a stagnating lava flow, whose ropy and broken upper surface was rolled forward and downward onto the floor of the depression as they slowly advanced.

MAIN RIDGES.

The greater thickness of the main sheet—its measure frequently reaching 400 or 500 feet—renders it much more important topographically than the thinner anterior and posterior sheets. It consists in greatest part of dense and rather coarse-grained trap. It is always vesicular near the upper surface, but the vesicular portion of the sheet is much less important in proportion to the whole than it was in the anterior sheet; yet the absolute thickness of this portion is not so dissimilar in the two as would be inferred from Percival's descriptions. As described by E. S. Dana and Hawes, the trap in several of the eastern ridges is much altered as compared to that of the western ridges, so that it belongs with the diabases rather than with the dolerites. The studies by C. L. Whittle have extended the application of this conclusion.[1]

*Saltonstall or Pond Mountain.*—This southernmost member of the long series of main eastern ridges presents a strongly crescentic outline. It is extremely scoriaceous on the back slope in its descent to Saltonstall Pond. The notch near its southern end, deepened by the cut of the Shore Line Railroad, exposes the trap resting conformably on the under sandstone. The sandstone is very little altered, but at a depth of 3 inches beneath the trap it contains minute and isolated fragments

[1] Bull. Mus. Comp. Zool. Harvard Coll., Vol. XVI, 1889, pp. 99-138.

of trap, implying that when these layers were deposited a neighboring mass of trap supplied fragments to be mixed with the grains of sand. The base of the trap for several feet upward is decidedly close grained and somewhat vesicular. It is more or less fractured, the fractures being filled with subsequent infiltrations of secondary quartz and calcite. This breccialike structure corresponds, on a smaller scale, with that observed at the base of the Totoket anterior, near its northern end.

Approaching the upper surface at the base of its eastern slope, the dense and coarser texture of the middle becomes finer and finer and at the same time more vesicular. Near the east end of the northern hook of the ridge, there is fortunately a small exposure of sandstone in a stream bed directly overlying the trap sheet. Good contact specimens were obtained here, and as they have been interpreted by E. O. Hovey as indicating induration from heat, a very careful examination of them has been made in field relations, in hand specimens, and under the microscope. The sandstone lies conformably upon the trap as far as the few feet exposed suffice to indicate the relative attitude of the two rocks. Although the sandstone is unquestionably indurated, ordinary observation does not suffice to decide whether this results from cementation or from baking. When a section across the contact is examined under the microscope the sandstone is seen to occupy all the inequalities of the trap surface, the lamination of its minute stratification conforming to the general contours of the trap. Fragments of vesicular trap not infrequently occur in the sandstone. Vesicles, more or less open upward, are filled with stratified clastic deposits, and are connected with the overlying sandstone by narrow necks. These clastic grains are usually of the most enduring minerals of the ancient crystalline rocks, namely, quartz, feldspar, hornblende, and muscovite, cemented by granular calcite and stained by ferric oxide. Minute fragments of vesicular trap are found, along with other clastic grains, in the fillings of the open vesicles, these being the most interesting of their constituents. The first grains deposited in the vesicles are arranged with their longer axes roughly parallel to the floor of the cavity, but as the center filled faster than the sides, the grains later deposited assume a general parallelism to one another. It is especially notable that the plane of deposition then becomes accordant in various independent vesicles, and that it stands parallel to the stratification of the sandstone, which here has a steep southward dip. Although the hardness of the sandstone is distinctly greater than usual, there is no appearance of fusion or of alteration of the mineral grains; the hardness is entirely due to infiltration of cementing calcite and iron oxide.

*Totoket Mountain.*—Like the preceding, this member of the series of ridges formed by the main trap sheet is strongly curved. The only localities on this mountain that need be mentioned in this connection are at upper contacts of the trap with the sandstones, found in stream beds near North Bradford village and near Quonnipaug Pond, on the inside of the southern and northern hooks in which the mountain terminates. The upper surface of the trap at both of these localities is very vesicular and irregular. The lamination of the sandstone conforms to these irregularities. There is intimate mixture of rounded trap grains and sand close to the contact; occasional trap fragments are found for a few feet above the contact at the northern locality. Grains of sand and trap fill open vesicles, repeating the minute and peculiar structural features of the Saltonstall contact. The sandstone in the northern hook is decidedly indurated near the contact, but it exhibits no signs of baking, fusing, or other alteration due to heat. Its hardness seems to result solely from cementation, as in the previous case.

A ravine in the northern bluff of Totoket discloses the even contact of the trap on the sandstone beneath it, but without significant features. Few contacts of this kind are here referred to; they were not often carefully searched for, seldom being of discriminative value.

*Higby Mountain.*—The long ridge known in its successive parts as Tremont and Higby Mountain has yielded only two exposures of upper contacts. Both of these are on the back of Higby Mountain, on Fall Brook, about a quarter of a mile upstream from the southern of the two roads that run from Westfield toward Meriden. All the features of adjustment of the sandstone to the trap, such as mixture of trap fragments with the sand grains and filling of open vesicles with clastic fragments, are here repeated. Some of the rounded trap fragments occur in the sandstone layers even 5 feet above the general surface of the trap. Downstream from these contacts many stones in the bed of the brook contain fragments of vesicular trap. It was by following these up to their source that the contacts were found.

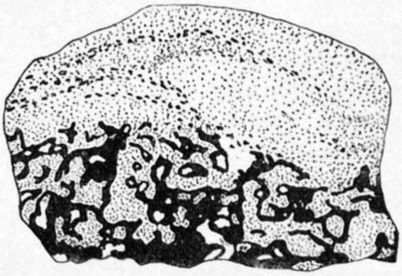


FIG. 11.—Scoriaceous trap and sandstone, Lamentation Mountain; natural size.

*Chauncy Peak.*—The channel of a wet-weather stream leading down the back of this portion of the main sheet passes gradually from dense to vesicular trap, and then to a mixture of trap fragments and sandstone. The rock surface is glaciated and comparatively little weathered, but the area open to observation is small. No special study of it has been made, as it appeared to repeat very closely the features better shown on the back slopes of Higby and Lamentation mountains.

*Lamentation Mountain.*—At the northern end of this ridge Spruce Brook cuts a little trench, passing from the trap to the covering sandstone. The walls of the trench exhibit vesicular trap of very irregular texture. Intimate and complicated mixture of sand and trap occurs in the upper portion of the sheet, as in fig. 11. The upper portion is highly vesicular, with unevenly rolling surface. Sand and trap grains fill open vesicles near the surface, and occupy interstices among the trap fragments for 2 or 3 feet downward. The stratification of the sandstone is conformable to the trap, the minute layers in the vesicles having the same dip as the larger beds. Isolated waterworn fragments of vesicular trap occur in the sandstone for 2 or 3 feet above the trap sheet.

*Quarry Ridge.*—The small, easternmost trap ridge of the Hanging Hills group has been deeply quarried for railroad ballast and road metal at and near its southern end, in Meriden, and now presents an excellent dissection of a complex trap sheet. It is one of the most important exposures in the region. The trap is separated into a lower and an upper sheet by a divisional surface, slanting with the general monocline of the region. The lower sheet is dense and of bluish color in its deeper exposure, but becomes red within 4 or 5 feet of the top; here it contains numerous amygdules of chlorite, which give it a mottled appearance and cause it to simulate an altered sandstone, for which it has indeed been mistaken. Even 10 feet or more beneath the upper surface the rock is greatly altered, the glassy base being often devitrified, and calcite being so plentiful that dilute acid causes effervescence. The amygdules, frequently of regular form, are at this greater depth nearly all due to replacement. Toward the surface the texture steadily grows finer and the cavities due to gas expansion become more numerous. Five feet below the contact the cavities occupy about one-fourth of the volume; a foot from the surface their proportion has increased to fully two-thirds. The origin of these vesicles by gas expansion is shown by the tangential arrangement of the feldspar crystals around the cavities, conforming even to their minor irregularities.

The red color near the surface is due to the formation of ferric sesquioxide. Hawes has pointed out[1] that alteration of iron-bearing minerals within a rock mass not exposed to the weather is from one protoxide to another; but if exposed to the weather, it is from the protoxide to the sesquioxide. It is the former change that generally characterizes the alteration of the trap in the eastern ranges, as contrasted with the unchanged trap of the western ranges. The occurrence of reddened trap, such as is found here, even at a depth of 30 or 40 feet beneath the surface of the overlying dense steel-green trap of the upper sheet, is therefore to be ascribed to some particular cause, not operating generally.

The upper surface of the under sheet has been occasionally stripped bare in process of quarrying. It then appears to be of undulating form, each convex wave being from 2 to 4 feet across and from 3 to 6 inches in relief. The resemblance of this surface to the surface of ropy lava flows on Vesuvius or Kilauea is very striking.

A little clastic material occurs between the two trap sheets, but it is recognizable only by aid of the microscope, as the sheets are generally

[1] Am. Jour. Sci., 3d series, Vol. IX, 1875, p. 190.

UPPER SURFACE OF LOWER LAVA FLOW, LANE'S QUARRY, MERIDEN.

The lower flow is vesicular near its upper surface; the upper flow is dense at its base

closely in contact. At certain points they are so closely welded that hand specimens may include fragments of both. The clastic material consists of rudely stratified grains of quartz and orthoclase, with angular fragments of red trap, like that in the upper part of the under sheet, the whole cemented by quartz and calcite.

At one point in the northern quarry a deposit resembling the ash of the Lamentation anterior bed has been found by J. B. Woodworth.

The upper sheet is bluish-green on fresh surfaces, but weathered to reddish-brown on joint faces. It is dense throughout the quarries, and is much less altered than the lower sheet. It falls off abruptly on the eastern side, for reasons to be explained on a later page; but by following the Quarry Ridge northeastward half a mile or more, to a point where its eastern side descends normally to the low ground, the trap surface near the base of the ridge is found to be vesicular, as is usual with all the eastern members of the series.

It may be at once stated that this composite structure is interpreted as a double lava flow, the second flow being spread out on the weathered surface of the first.

*South Mountain.*—A bench rises on the back of this mountain (shown

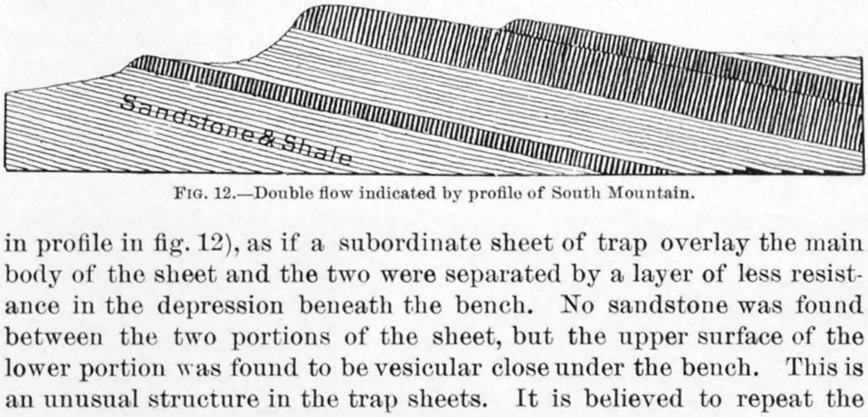



FIG. 12.—Double flow indicated by profile of South Mountain.

in profile in fig. 12), as if a subordinate sheet of trap overlay the main body of the sheet and the two were separated by a layer of less resistance in the depression beneath the bench. No sandstone was found between the two portions of the sheet, but the upper surface of the lower portion was found to be vesicular close under the bench. This is an unusual structure in the trap sheets. It is believed to repeat the composite structure exposed in the Quarry Ridge. The front view of South Mountain (Pl. X) gives only the ordinary aspect of a bold bluff, without showing the bench on its back.

*West Peak.*—The easternmost trap outcrops that are believed to belong to the main sheet in this block appear to the east of a road leading north from the Reservoir Notch, and are separated from the main ridge by a shallow longitudinal valley. Although no critical contacts have been found, it is thought probable that these eastern outcrops may here indicate the same upper flow already seen in the Meriden quarries and inferred on the back of South Mountain. Similar supplementary ridges are found on the back slope of the main sheet in the New Britain district.

*East Talcott Mountain.*—The duplication of the main trap sheet in

Talcott Mountain by a longitudinal fault will be duly described in Part II. Where the West Hartford–Bloomfield town boundary crosses the back slope of the eastern main sheet Tumbledown Brook has cut a shallow gorge. Amygdaloidal trap is seen in place at the water's edge, and for several feet upward on the banks there are distinctly bedded layers containing fine trap waste, dipping 8° or 10° eastward. A few rods farther downstream there is a thin-bedded gray shale, followed by heavier bedded red shale.

No significant features were found in the trap sheet of the main ridges farther north, although the trap is always vesicular if seen at points near the base of the eastern slope.

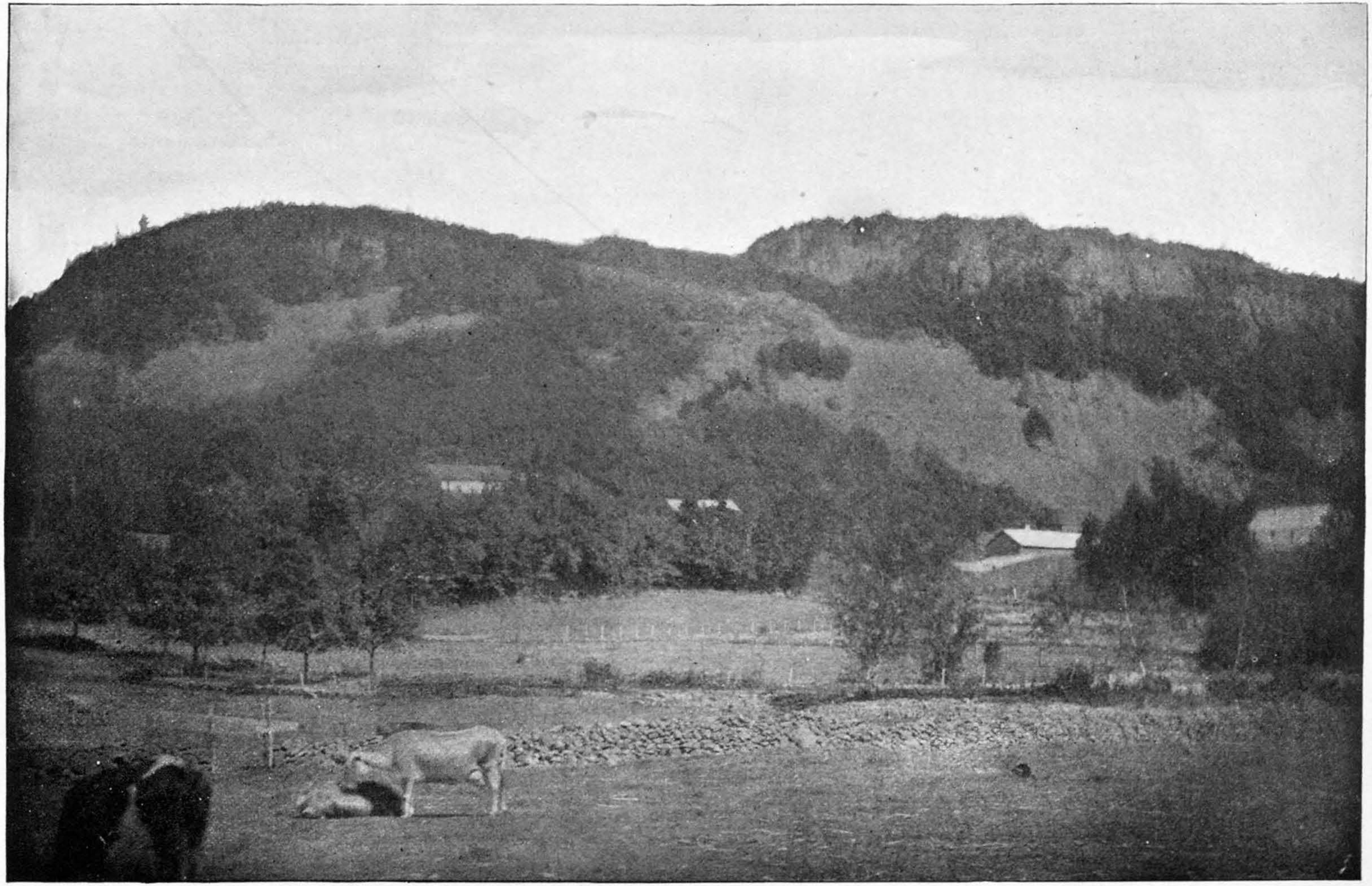
### POSTERIOR RIDGES.

*Posteriors of Saltonstall Mountain.*—Two low ridges lie within or posterior to the crescentic curve of this member of the main sheet near Branford. The first is a long, narrow ridge, sympathetic in its curvature with the line of the main sheet. The base of the sheet in this ridge is slightly amygdaloidal; the middle, dense. Near its northern end, where it is crossed by the road from Branford to Totoket village, the overlying sandstones are found on the roadside in close contact with the trap. The upper portion of the sheet is here very vesicular, but not locally close grained at contact with the sandstone. Sand grains and trap fragments are mixed in the contact layer, and sand grains occur in open vesicles. Occasional waterworn fragments of trap are found in the sandstone a foot or more above the contact.

Ledges of very coarse trap conglomerate, with bowlders occasionally 2 or 3 feet in diameter, occur to the eastward of the north end of this trap sheet, and are believed to have been originally continuous with the trap sheet. They now seem to be faulted away from it, but this can not be clearly demonstrated.

The second posterior ridge of Saltonstall Mountain rises on the northwest of Branford village, close along the boundary between the Triassic and crystalline rocks. It will be later correlated with the first posterior, of which it is believed to be a repetition by faulting. Here only its primary features will be discussed.

The base of the sheet is brecciated for 2 or 3 feet upward, the fissures being filled with sandstone, as if the two kinds of rock had been confusedly mixed. The texture is generally dense, except near the upper surface, where it becomes highly cellular. The overlying sandstone, exposed on the south side of a small pond, contains numerous fragments of vesicular trap, but the exposures are much weathered, and on the whole are unsatisfactory. A trench was dug here, but no unweathered rock was reached. The disturbed attitude of the trap in this ridge, whereby a small portion of it, on the southwest side of a crossroad, seems to be overturned past the vertical, so that the under sandstones appear to lie on the trap, will be referred to later. (See p. 127.)



SOUTH MOUNTAIN, HANGING HILLS, MERIDEN; LOOKING NORTHWEST.

*Posterior of Totoket Mountain.*—The two portions of the posterior within the curve of this mountain exhibit the usual vesicular upper surface, and are everywhere accordant with the variable strike and dip of the conglomeratic sandstones that frequently outcrop near them. A small trench was opened on the back of the northern ridge, half a mile west of the south end of Quonnipaug Pond, revealing vesicular trap overlain by red shaly sandstone, in which fragments of trap were numerous.

Each of these curved posterior ridges is attended near its northern end by isolated ledges of trap. It is uncertain whether these ledges represent faulted fragments of the posterior or independent dikelike intrusions. The absence of dikes elsewhere at so high a horizon in the formation, the close proximity of the normal posterior sheet, and the occurrence of certain irregular basal structures at the under surface of one of these ledges (near a milldam southwest of Quonnipaug Pond), similar to the basal structure of the eastern posterior sheet in the Pond Mountain crescent, would all support the former view; but there is no independent field evidence indicating the occurrence of the inferred faults.

*Posterior of Piscapaug Mountain.*—The eastern member of the northern posterior ridges of this mountain has been quarried alongside the Durham road, exposing a loose-textured, vesicular trap, much decomposed and overlain by conglomerate. The contact was too much weathered for critical observation. A mile south of Durham Center, near the eastern boundary of the lowland, there are ledges of conglomerate containing trap fragments. These can not be certainly correlated with any one of the trap sheets hereabouts, but they are plausibly associated with the posterior ridges that approach on the southwest.

*Posterior of Tremont Mountain.*—The cut of the Air Line Railroad through the posterior ridge near Middlefield discloses an irregular vesicular structure at the base of the trap sheet, where it conformably overlies red shale. The shales are confusedly mixed through the lower part of the trap for from 3 to 15 feet from the base.

*Posterior of Higby Mountain.*—This is a continuation of the ridge mentioned in the preceding paragraph, 4 miles farther north. It is here crossed by a stream, and Rock Falls station of the Air Line Railroad is near by. Sandstone is found lying on vesicular trap at the south side of a mill pond alongside the railroad. The uppermost vesicles in the trap are often filled with clastic grains, connecting with the overlying sandstone by narrow necks. Grains and fragments of trap occur plentifully for a foot or more upward in the sandstone above the contact. A beautifully waterworn pebble of vesicular trap was found embedded in the sandstone several feet above the contact. Drift bowlders in the railroad cut near by show a mixture of trap and sandstone.

*Posterior of Chauncy Peak.*—The southern end of this ridge is cut by the Meriden, Waterbury and Connecticut River Railroad, a quarter of a mile east of Highland station. When still freshly opened the rock showed a remarkable mixture of angular and vesicular trap fragments in micaceous sandstone, the layers of the latter wrapping around the former. All the phenomena already described regarding the occurrence of clastic grains in open vesicles are here repeated. In some of the vesicles the filling is incomplete, as in fig. 13, the upper part being then occupied with calcite, and the line of division across the vesicle running parallel to the monoclinal structure of the region. A number of specimens were secured exhibiting this peculiar feature very clearly. It was first noticed at this locality, and afterwards found to characterize many others.

*Berlin posterior of Cedar Mountain.*[1]—The sharply curved hook of the posterior ridge near Berlin village exhibits a considerable area of vesicular structure on its back slope, but no overlying sandstone can be found. The underlying sandstone is beautifully conformable to the curve and dip of the trap ridge; it was early perceived to be significant of the relations between the igneous and aqueous strata. Percival makes distinct record of this feature.[2]

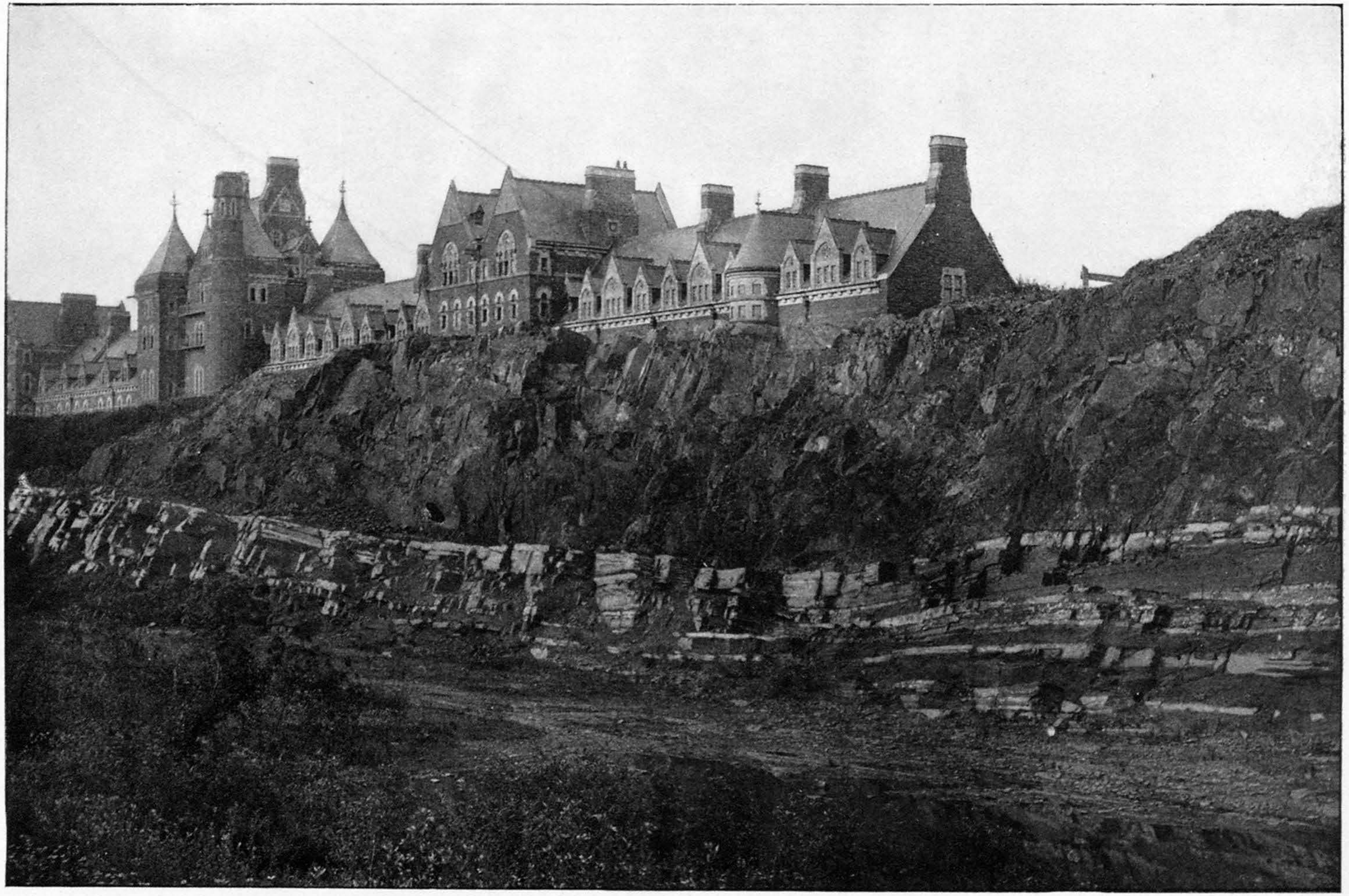
FIG. 13.—Trap vesicle containing clastic deposit, posterior of Chauncy Peak; magnified 4 diameters.

*South Glastonbury trap conglomerate.*—A trap conglomerate is found on a hillside near the crystalline border, east of South Glastonbury, on the east side of the Connecticut River. It contains plentiful fragments of dense and vesicular trap, along with pebbles and cobbles from the crystallines. This is correlated with the horizon of the posterior trap sheet, from which it is now separated by one of the chief faults of the valley lowland (see p. 103).

*Rocky Hill posterior of Cedar Mountain.*—A railroad cut on the back of the posterior ridge in the town of Rocky Hill, but near the village of South Weathersfield, disclosed sandstone containing fragments of vesicular trap similar to that overlying the anterior sheet at Tariffville or the main sheet of Lamentation Mountain.

*Hartford posterior of Cedar Mountain.*—At the north end of this posterior, on the grounds of Trinity College, Hartford, the vesicular upper surface of the sheet is well exposed, with intermixed portions of sandstone. The under surface of the sheet, resting evenly on the sandstones,

[1] This part of the main sheet was called Newington Mountain by Percival.
[2] Geology of Connecticut, 1842, p. 358.



TRAP ON SANDSTONES, CITY QUARRY, HARTFORD.

is well exposed in an extensive quarry (Pl. XI). It was after observation of this contact that the elder Silliman wrote the statement quoted on page 16.

*Posterior of Ragged Mountain, New Britain.*—The overlapping ends of two small trap ridges on the northeastern border of New Britain are regarded as the slightly faulted portions of the single posterior sheet belonging with the main sheet of Ragged Mountain.[1] A small stream runs between them. The eastern ridge is quarried, exposing the base of the sheet lying evenly on the sandstones. The trap is here generally dense, but incloses local amygdaloidal areas. The upper surface, seen at the eastern base of the western ridge, close to the stream, exhibits the usual vesicular structure of all these sheets, and is covered by a mixture of trap fragments in sandstone of the ordinary kind, but much weathered and not easily accessible.

*Posterior ridge of Farmington Mountain.*—East of Hartford Reservoir No. 4, and about 3 miles north of New Britain, a posterior trap ridge is crossed by a small stream. Here the upper surface of the vesicular trap is exposed, with a cover of tuffaceous deposit, all much weathered. No sandstone or shale was visible.

*Posterior sheet at Farmington River Gap.*—Close on the south of the river, nearly a mile north of North Bloomfield, the overlying sandstone approaches the vesicular trap within a few feet. An opening was made here and the irregular contact disclosed, but it was so much weathered that no good specimens could be secured. The sandstone appeared to fill inequalities in the trap surface. About a mile south of this point a road lies along a shallow trough on the crest of the low posterior ridge, as if the trap sheet consisted of two thin flows, the upper surface of the under flow by the roadside being vesicular, while the low bluff face of the upper flow is dense.

### SUMMARY FOR THE EASTERN RIDGES.

Certain features are found constantly to prevail in the trap sheets of the eastern ridges wherever observation is permitted free opportunity; other features are of inconstant occurrence. The trap of these ridges is prevailingly more altered than that of the western ridges. The sheets show no tendency to cut across the beds of the inclosing sedimentary rocks, but are, on the contrary, surprisingly accordant with all changes in dip and strike of the sandstones. The trap is more altered here than in the western range. The upper surface of the sheets is invariably vesicular; pipe-stem cavities are often noted near and normal to the upper surface. The bedding of the overlying sandstone is minutely adjusted to the inequalities of the upper surface of the trap, even occupying open vesicles; fragments of trap, generally vesicular and more or less waterworn, occur in the sandstones just over the trap. These are all features of constant occurrence.

[1] This main trap ridge ends in a strong bluff at its south end, called High Rock in Percival's report.

The under surface of the sheets is sometimes smoothly accordant with the under sandstone; sometimes it exhibits a peculiar intermixture of sandstone and trap, as if the two had been kneaded together. The lower part of the sheet is generally dense, but sometimes moderately vesicular. The trap is sometimes replaced by lava blocks and ashes, either at the bottom or at the top of the sheet. In two cases a sheet appears to be of composite structure, as if consisting of two flows, each of which has a vesicular structure at the upper surface. The adjustment of the overlying sandstone to the uneven trap surface and the mixture of trap fragments in the sandstone are subject to innumerable variations in quantity and quality. The overlying sandstone is sometimes indurated, but in no case does it give any indication of baking. These are features of inconstant occurrence.

### CONCLUSION AS TO ORIGIN OF TRAP SHEETS.

Recalling the method of investigation described above, the problem of the trap sheets may now be reviewed in accordance with its requirements. If only the crystalline texture of the trap rocks were observed, they would simply be regarded as of igneous origin, without specification as to their relation to the adjoining rocks. When the linear arrangement of the trap ridges was noticed, the trap being seen to rise to greater height than the sandstone on either side, it was natural to regard the trap as a dike rock. When observers in certain districts noted the interbedded attitude of the sheets and the indurated condition of the overlying sandstones these sheets were naturally interpreted as intrusions, and the same interpretation was applied to various other sheets of similar form. On the other hand, when observers in other districts discovered the mixture of trap fragments in the sandstone overlying the sheets, it was natural to regard the traps as overflows, and to extend this conclusion to the apparently similar sheets in all parts of the valley. Every advance of this kind resulted from new observations of significant facts. The fuller determinations now possible have been reached in precisely the same way; hence it is at present inadmissible to group all the sheets together as of one origin. The sheets of the western ridges are indisputably unlike those of the eastern ridges. The group of intrusive sills is decisively separated from the group of extrusive flows.

This decision does not rest on the permissive evidence of a few characteristics, occasionally discovered, but on the compulsory evidence of numerous characteristics, repeatedly observed, this being especially true for the extrusive flows. In their case, particularly, the correspondence of consequences deduced from theory with facts observed in the field is so peculiar, so complicated, and so plentiful that the conclusion as to their origin may be regarded as demonstrated with all the certainty that should be attached to geological reasoning.

Not only do the two groups of sheets fall naturally into the classes of sills and flows, but by the presence of the two classes in a single field each illuminates the other by the light of contrast. Thus it happens that although the contacts and other critical structures have been seen only on outcrops of very small extent, the conclusions to which they lead may be justly applied to the trap sheets over the whole extent of the valley lowland. On later pages these conclusions will frequently be utilized, the sheets of the western ridges being referred to as sills and those of the eastern ridges as flows or lava beds.

### CORRELATION OF DIKES, SILLS, AND FLOWS.

The understanding that has now been reached as to the manner of origin of the various ledges and ridges of trap rock permits an important advance in the historical account of the valley. To the description of the processes of transportation and sedimentation already given, there must now be added the picture of volcanic eruptions and lava flows, with their usual accompaniment of underground intrusion. Just as the small exposure of the Triassic foundation in Roaring Brook of Southington must be expanded to represent a broad pre-Triassic land surface, just as the local exposures of cross-bedded and ripple marked strata must be carried over scores and scores of miles in reconstructing their original extension, so the scanty outcrops of the upper surface of the lava beds must be magnified till they floor over the greater part of the ancient depression; here smooth and ropy, there loose and clinkery; at one place a firm lava bed, at another largely made of loose ashes. The creeping advance of the flows from some unknown source or sources should be deliberately pictured—a slow process that probably occupied months or years for each sheet, yet rapid in comparison to the accumulation of sediments; sometimes a quick advance, sometimes a tumult of shattered ashes and lava blocks. Three distinct episodes of this character interrupted the simpler processes of denudation, transportation, and deposition, by which the greater part of the Triassic deposits were formed. Near the middle of the period of deposition the placidity of the region was disturbed by the spreading of a lava sheet—the anterior—up and down the trough for some 45 miles inland from the coast, nearly to the Massachusetts border of to-day. Its thickness is generally about 250 feet. Its breadth can not be measured, because its western part is worn away and its eastern part is buried. The vent from which it was extruded can not be identified with any certainty, for at no place is a dike or neck seen to connect distinctly with the flow. The nearest approach to such connection is west of Reeds Gap, where the East Wallingford dike and the anterior ridge of the Higby Range approach within about a quarter of a mile of each other. The appearance of ash beds and lava blocks in the southern and central parts of the anterior imply explosive eruption at no great distance, but it would be gratuitous to assert on this account that the eruption took place at any definite place. It is, on the other hand, entirely within reason to picture the growth of a volcanic cone of greater or less size at some point in the ancient trough, while the lava flow crept far and wide over its smooth floor. The cone may be buried under the upper sandstones in the eastern part of the trough, or lost by erosion in the uplifted western part of the trough. If in the former, its position is entirely unknown. If in the latter, it must have risen above some of the existing dikes, and of these it should be remembered that those of Mount Carmel are the largest of all that stand not remote from the ash beds of the anterior sheet. It has, however, been suggested that the ash beds of the anterior sheet were produced by local explosive action caused by the advance of the hot lava through water sheets in the trough, and in such case their distribution would have little or no relation to the site of the lava vent.

After the extrusion of the anterior lava bed, volcanic action became extinct, and placid conditions reigned again for a long time. Reptiles once more stalked over the sand flats. Fish became very abundant while the anterior black shales were accumulating 50 or 100 feet above the lava bed. Any volcanic cone that may have been built at the time of the lava flow must have been deeply dissected during the deposition of several hundred or a thousand feet of sandstones and shales before the greater outpouring of the main sheet took place. This was truly a gigantic lava flow. Its thickness is frequently 400 or 500 feet. Curiously enough, it gives no signs of explosive eruption in the way of ash beds and lava blocks at any point in Connecticut, except in the most insignificant way in the Meriden quarry. It spread much farther northward than the anterior, now reaching certainly to Mount Holyoke, and perhaps almost to the northern end of the sandstone trough in Massachusetts. Its length was thus at least 70, and possibly 95, miles. It approaches close to the eastern border of the sandstone at several points, and as it holds a good measure of thickness in both its eastern and western exposures, it must have originally had a decidedly greater breadth than now. Hence its average breadth may be reasonably estimated at 10 miles. Its area may therefore have been about 700 square miles and its volume 70 cubic miles.

So heavy a flow must have formed land surface for some time after its eruption. But the depression of the region continued, and sediments were spread over the lava bed once more. Here we must imagine those minor processes by which the loose lava clinkers were washed about, their edges rounded, and their grains mixed with the sand and slowly sifted into the open vesicles or into the crevices among the clinkers, until at last all was sealed over, and again the more placid processes of distribution and sedimentation of waste from the crystallines went on undisturbed. It was during this quiet interval that the posterior black shales, with their countless fish skeletons, were deposited. Only once more did the volcanic forces break forth. Then the posterior lava bed was spread out, in extent intermediate to the other two, as it reaches from near the coast almost to Amherst, about 70 miles inland, in

Massachusetts. Its thickness is commonly 100 or 150 feet. After it was sealed over there is no indication of any revival of volcanic activity in Connecticut, unless in the basic dike of Baileyville (p. 47). It is interesting to note that trap-bearing conglomerates occur in apparent continuation of the posterior flow at three points near the eastern border of the lowland—north of Branford in the Pond Mountain crescent, south of Durham Center, and in South Glastonbury. This suggests that the lava flow was spread out chiefly in the lower medial part of the trough, and that its fragments, along with the waste from the bordering uplands, were washed into the marginal belts.

The southern division of the western ridges is 15 miles in length. The northern division, more interrupted, measures about 17 miles to the State boundary. Their observable volume is therefore much less than that of the main flow.

### DATE OF INTRUSION OF SILLS.

It is manifest that the date of the intrusion of the interbedded or dikelike sills of the western ridges can not be immediately correlated either with the date of deposition of any definite stratum of the sandstones and shales or with the date of eruption of any one of the lava flows. They must have been driven into place after a considerable thickness of sandstone had been accumulated above the low horizons that they invaded, but there is no direct reason for supposing that their intrusion did not take place until after the tilting of the sandstone into its present monoclinal position, although this view has often been held. Intrusions among horizontal strata are by no means unknown. The eruption of the extrusive sheets is so closely associated with the deposition of the sandstones that it is permissible to associate the intrusions with a similar date. This is confirmed by finding that the intrusive sheets have suffered dislocation of a kind very much like that which has effected the extrusive sheets, as will appear in Part II (p. 133).

By anticipating the conclusion there reached, that the sills were in all probability intruded horizontally before the tilting of the formation into its monoclinal position, it will be perceived that it is not necessary to refer the sills to a feeding dike on the east of their present outcrops. It is true that, in the case of East Rock, Dana has inferred a supply from a steep fissure on the northeastern side of the rock, but the evidence of this is gathered only from the steeper slope of this side near its northern end, and not from unquestionable exposures of a supply dike.[1] It is also true that in the Palisade trap sheet, so similar to the West Rock ridge, Darton has found more definite evidence of supply from a dike near the western base—this corresponding to the eastern base in the Connecticut monocline.[2] But if the observed structures generally prevailing in the West Rock ridge are drawn in true

[1] Am. Jour. Sci., 3d series, Vol. XLII, 1891, pp. 94–95.
[2] Bull. U. S. Geol. Survey No. 67, 1890, p. 37.

section, and in the attitude that obtained before tilting, it is manifest that the argument for the ascent of the trap, roughly between the bedding planes of the sandstones, has little or no force. The sills did not *ascend* between the layers; they spread laterally from some unknown source, wedging their way between the layers as they advanced, and thus uplifting all the overlying mass. The present intrusive ridges need not have any definite relation in direction or position to the fissures from which they were supplied. The Mount Carmel dikes and stocks appear to be the fissures that are most available as supply vents for the sills of the West Rock ridge and its northern neighbors; but the possibility of the supply coming from the dikes which now outcrop in the crystalline upland on the west can hardly be excluded by anything now known.

The oblique passage of the West Rock sill across the sandstone strata near its southern end has been described and illustrated. Such a structure does not seem to be easily accounted for if the trap entered from the east, for in that case the observed structures would require the advancing lava to break its way downward into successively deeper and deeper layers as it moved forward; but if it departed from the guidance of a single bedding plane it would most probably break upward, not downward. If, on the other hand, an intrusion from the north is assumed, the gradual upturning of the hooked southern end of West Rock is reasonably accounted for by the falling in of the roof layers near the end of the uplifted covering mass, much in the manner advocated by H. D. Rogers.[1] Pine and Mill rocks probably occupy transverse fissures that were broken across the sandstones at the time of the intrusion of the West Rock sill, and near its rapidly thinning southern end. East Rock and its dependencies appear to be intrusions at higher levels, but closely associated with the transverse dikes. The oblique attitude of the northern members of the southern intrusive range in Cheshire, where they apparently creep to higher and higher horizons, repeats more deliberately the sudden upturning of the intrusives at the south end of West Rock.

### HORIZONS TRAVERSED BY DIKES.

The dikes are peculiar in being limited to the crystallines and to the under and lower division of the sandstones, with the exception of the basic dike in Baileyville, which cuts the posterior shales, and which by its position and composition proclaims itself to be the single member of another family of intrusions, and with the further exception of the very questionable dikes near the posterior ridges of the Totoket crescent. Their not infrequent upper termination beneath a sandstone cover has been mentioned. The dikes, like the sills, have suffered dislocations similar to those suffered by the flows (p. 132); hence, on all

[1] Am. Jour. Sci., 1st series, Vol. XLV, 1843, p. 333.

accounts, it seems most probable that all the igneous rocks, dikes, sills, and flows were introduced into the body of the sedimentaries during the filling of the subsiding trough, or at least before tilting and faulting had begun.

## VULCANISM.

Without entering far into the great problem of vulcanism, there are certain considerations bearing on it that result from our field study and that deserve brief mention. There is an intimate association of trap dikes and sheets with the time and place of Triassic deposition. The association in place, not only in Connecticut and Massachusetts but in the other Triassic areas of the Atlantic slope, has often been remarked, but the association in time has been less generally recognized, chiefly because many writers have not felt the force of the observations that support the theory of the contemporaneous origin of the trap. That the extrusive flows of the eastern ridges are contemporaneous does not seem to me longer open to doubt. That the intrusive sills of the western ridges are also essentially contemporaneous—that is, that they

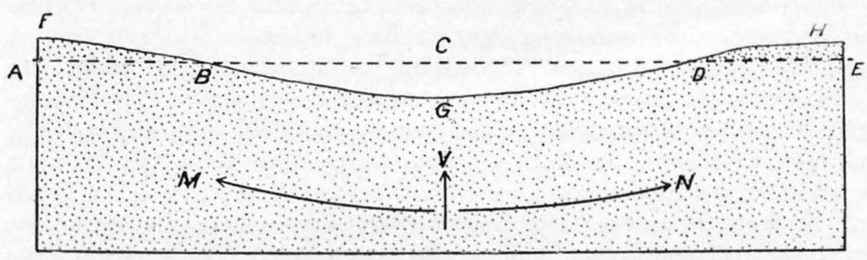



FIG. 14.—Relation of Triassic warping to vulcanism.

were intruded before the period of deposition was closed by the beginning of deformation—will be argued to a very probable conclusion later (see p. 134). There thus appears to be an intimate connection between depression of the surface and eruption of the lavas from deep-seated sources, and it therefore seems reasonable to regard the first as causally related to the second. That such a relation is permissible may be seen from the accompanying figure. If a region (shown in vertical cross section in fig. 14) is warped from a profile ABCDE to a profile FBGDH, there must be relative upward movement on either side and downward movement in the middle, and this is probably accompanied by a compensating lateral flow of the deep-seated yielding parts of the earth's structure away from the middle toward either side, as indicated by the arrows M N. It is conceivable that a perfect compensation might be thus effected, but it is also conceivable that the compensation by deep-seated lateral flow might be imperfect, and that the deficiency might be made up by vertical escape through fractures, thus producing eruptions of lavas in the trough of depression. Although volcanic action is often associated with elevation of the region where it occurs, this association with depression also seems to deserve consideration in the general discussion of vulcanism.

A corollary of the supposition that connects eruption with deformation is that the expansive force of steam and other vapors may be sometimes or generally dispensed with as an essential cause of the ascent of the lavas. This corollary finds direct support in the great sills of the western ridges, as well as in the huge laccoliths of the Rocky Mountain region. These great intrusive masses have certainly been driven upward or forward by some great force, but they present no indication whatever that the expansion of gases has taken any part in the labor of driving them to their present position. Taken along with the plateau-like extrusions of Iceland, this tends to the conclusion that deformation may very generally be the first cause of intrusions and extrusions, and that the gaseous explosions by which many cinder cones are built may often be hardly more than accidents of the surface, as advocated by Prestwich.

## ISOSTASY.

The doctrine of isostasy has been held to account for such conditions as prevailed during the period of Triassic deposition; for all through that time the trough was slowly sinking, while it received an increasing load of detritus from adjoining regions, and the latter were, by reasonable inference, slowly rising as they were denuded. But in this case, as in many others, it is hardly logical to say more than that the facts permit but do not compel a belief in the doctrine. As in all examples of supposed isostasy, an external initial cause is necessary to set the processes of denudation and deposition at work. In our case, some local down-bending must have preceded the beginning of deposition and thus have served as its cause rather than as its effect. Similarly, some other process must have put a stop to the continuance of depression and deposition. How largely the isostatic forces alone acted between the beginning and end of deposition, or how small a proportion they may have borne to greater forces of other kinds, is to-day a question for speculation rather than for argument. As the problem stands in an undetermined position, little is gained by assuming definite values for any of the factors involved in it. It is not at present advisable to go further than to conclude that for a long time deposition about balanced depression, and that over the lands to the east and west erosion may have roughly compensated for elevation.

## RELATION OF DEPOSITION AND DEFORMATION.

It is important to clear the mind of a false idea that grows unconsciously concerning the relation of deposition and deformation in such a problem as this. From the diverse nature of the two processes, and from the habit of considering them in separate chapters for convenience of discussion, it comes to be natural to conceive of deposition ceasing of itself; and then to imagine, after a greater or less interval of inaction, an independent beginning of deformation. The two processes are commonly considered separately, as if deposition had been at work up to a certain hour of the geological clock, and deformation began to work when a certain later hour struck, perhaps after a noon interval of rest.

While this is conceivable, it does not seem at all probable. It is more likely that deposition was, throughout, the creature of deformation. Deposition might go on as long as the warping of the region depressed the trough area and elevated adjoining areas, but it was in all probability stopped when the trough was elevated, with more or less disturbance, above sea level. This conclusion seems to be of general application. Instead of saying, as is commonly the practice, that after the Paleozoic strata of the Appalachians had been deposited they were deformed in making the Alleghenies, it seems better to say that the deposition of the Paleozoic strata continued until it was stopped by the deformation which produced the Alleghenies. So, in Connecticut, it is eminently probable that the down-bending or breaking of the Triassic trough was only an early stage of the disturbance which tilted and faulted the Triassic strata, and however large the type that stand at the head of Part II to separate it, for the convenience of the reader, from Part I, these visible signs should be interpreted only as hyphens, by which successive phases of the subject are blended, and not as barriers by which they are held apart.

## PART II.—DEFORMATION.

### CHANGES FROM ORIGINAL ATTITUDE.

The existing attitude of the Triassic formation in Connecticut differs in three ways from the inferred original attitude. The formation as a whole stands higher than when it was deposited, for much of it that was accumulated at a low level has now been worn away from abvoe sea level. It has been upraised unevenly, more on the west than on the east, for as a whole it dips eastward. It has been both warped and faulted, for its different parts are more or less discordant with one another. These three classes of movements will now be considered in reverse order. Evidence of the warping and faulting will be found by immediate observation; the generally eastward dip will be explained as a result of the peculiar mechanism by which the faulting was produced; and the general upward movement of the whole region, by which its present altitude was gained, will be deciphered in the third part of this report, from a consideration of the process and progress of denudation.

It is in connection with the faults by which the formation is repeatedly broken that we find the chief structural problem of the region, a problem whose solution reaches further than at first might be anticipated, for it indicates the existence of dislocations in the underlying and adjoining crystallines where they might not otherwise have been expected. But associated with these strong and numerous dislocations there are occasional warpings of the strata. Deformations of this class are not called "folds," because the dips are, as a rule, of moderate amount, and still more because it is not desired to imply that lateral compression, acting to produce folds after the manner ordinarily attributed to such a force, has had anything to do with the disturbance of the region.

It is possible that the bending and breaking of the strata advanced, in general, together; but, on the other hand, it might be inferred from the nature of the processes that the bending probably began before the breaking; and there are on the ground indications that some of the warps were completed before the faults began. For example, in association with the posterior trap sheet in the hills southwest of Durham Meadows there is a local flat-warped trough, whose ill-defined axis runs somewhat west of north. It is crossed by the fault that separates Paug and Durham mountains—the Paug Notch fault—and the two parts into which the trough is thus broken are now out of line, as was discovered by L. S. Griswold, the southern standing about half a mile east of the northern. It seems hardly likely that the two parts of the trough could have been warped independently after the faulting; a flat warping previous to the faulting is much more probable.

In connection with the other warped structures, the evidence as to the relative date of bending and breaking is less definite. All that can be said is that the various fault lines pursue their courses, as far as traced, without systematic regard to the flat warpings, and hence that the breaks were probably subsequent to the bends; but on this point no positive conclusion is announced.

## WARPS.

The most manifest warps are found along the southeastern border of the formation, in the towns of Branford, North Branford, and Durham. Here the two irregular crescents of Saltonstall and Totoket mountains, and of the associated anterior and posterior trap ridges, with the intervening conglomerates, sandstones, and shales, are all conformably warped into half-boat-like basins. If the southeastern half of the boats ever existed, it has been removed by the great marginal fault that here bounds the formation. The strata, aqueous and igneous, in the northwestern half, dip not only from the side toward the keel, according to the general monoclinal dip of the region, but also from bow and stern toward the middle, appropriately to the peculiar structure here disclosed. Here, as so often elsewhere, the early interpretation of the trap sheets as dikes or intrusions prevented the discovery of the real structure of the district. The curved edges of the warped and denuded lava flows were taken to indicate the form of the fissures in which they had risen through the crystalline basis of the Trias.[1] They were introduced as illustrations on a small scale of the supposed habit of eruptions to follow curved fissures, elsewhere thought to be manifested in a larger way in the arrangement of mountain ranges and island trends.[2] H. D. Rogers said of these curved ridges: "The sandstone being disrupted in a plane parallel to the dip, the beds on the upper side of the sloping dike will be lifted off from those on which they reposed, and in this tilting of the beds there will arise toward the extremity of the fissure seams or transverse cracks extending in the direction of the dip."[3] This is quoted with acceptance by Hovey.[4] However applicable this view may be to the hooked end of West Rock (p. 59), it can hardly explain the curved edges of the extrusive flows.

On deciding that the three trap sheets here included are all contemporaneous lava beds, conformable with the sandstones, they must be regarded as entirely passive, so far as their present attitude is concerned. Their edges give no indication whatever of the form or position of the fissure through which their material rose from its deep source. Along with the earlier and later stratified deposits, below and

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[1] Whelpley, Proc. Assoc. Am. Geol. Nat., 1845, p. 64.
[2] Dana, Am. Jour. Sci., 2d series, Vol. III, 1847, pp. 391–392.
[3] Am. Jour. Sci., 1st series, Vol. XIV, 1843, p. 334.
[4] Am. Jour. Sci., 3d series, Vol. XXXVIII, 1889, p. 381.

above the lava beds, they were bent from an original horizontal attitude into their present warped structure, the structural sympathy of the diverse members of the series being a notable feature of the district.

Even the massive main trap flow, a seemingly inflexible sheet, curves submissively around the ends of the crescents, appropriately changing its dip and strike along with the sandstones and conglomerates below and above. This is markedly the case in the greater crescent of Totoket Mountain, where the short hooks at either end turn round so far that their inner slopes look obliquely back toward the longer medial portion of the ridge. Rather more than half of the boat structure is manifested in this fine example. Inside of the northern hook the posterior sheet and the conglomerates underlying it are strongly warped, their strike being even a little south of east, with steep southward dip. It was undoubtedly the steep inclination of this part of the posterior where it descends to the roadside by Lake Quonnipaug that gave Percival the idea that it was a dike.[1]

The separation of the two crescents of Totoket and Saltonstall mountains results from the greater denudation of the uplifted part of the once continuous main sheet that arched over the space between them. The posterior ridges within the crescents are separated by a greater distance because their once continuous sheet was originally arched up to a greater height, and is now worn down to a greater depth; but in both cases there is every indication that the lava flows once stretched over the space where they are now absent, just as they still stretch under the floor of the crescents where they are now buried. It is to be noted that the manner of study ordinarily applied to deformed and denuded aqueous beds is here held to be perfectly applicable to these deformed and denuded igneous beds.

The structure at the points of separation of the main or posterior crescents would warrant the use of such a descriptive phrase as "transverse anticline, with an east-plunging axis," but as this might give implication of a force of compression acting along the axis of the crescents it will not be here employed. The greater uplift at the point of separation is preferably ascribed to irregular warping, with which term no definite quality of deforming processes is associated, and for which at present no special explanation can be assured.

North of Totoket two other crescentic structures may be recognized, but of increasing dimensions and greatly complicated by faults. The first may be called the Middletown crescent. It is best indicated by the course of the posterior ridge, which turns distinctly eastward at either end of its curve. The second crescent begins with the Hanging Hills at Meriden and ends with the Mount Holyoke Range in Massachusetts. This may be called the Springfield crescent. Just as in the other example, but to a less degree, the strike and dip of the aqueous as well as of the igneous beds turn from their ordinary directions at

[1] Geology of Connecticut, 1842, p. 340.

either end of these large crescents, so as to accord somewhat with the pattern so well exhibited in Totoket crescent; but the dislocations by faulting are here so great that the warped pattern is not at first recognizable.

One of the most peculiar warped structures in the whole valley is that by which the main sheet is brought up in Cedar Mountain, southwest of Hartford, and by which the overlying posterior ridge is so notably deflected eastward out of its ordinary course. This upheaval might in one sense be regarded as comparable with that between the Pond and Totoket crescents, inasmuch as it terminates the Middletown crescent on the north, but it is irregular in being quite unrelated to the great Springfield crescent. It is remarkable in not being continued westward. On account of the numerous faults in this district, by which the three trap flows are so irregularly repeated, the structure hereabouts was not deciphered till nearly all the problems of the valley were solved. Then the ingenious suggestion of Prof. W. N. Rice, that Cedar Mountain must be a repetition of the main sheet locally uplifted, was happily confirmed by the discovery of the posterior black shales north of Rocky Hill village in appropriate position with respect to the underlying and overlying lava flows, and containing appropriate fossils (see p. 139). At no other point in the valley is there a more beautiful concentration of various lines of evidence upon an unexpected conclusion.

If the outcrops of sandstones and shales were more plentiful they would furnish sufficient evidence of warping, independent of the interpretation of the trap sheets as flows or sills; and wherever the sedimentary strata appear they confirm the inferences derived from the more persistent outcrops of the anterior, main, and posterior flows; but without the latter it would be very difficult to discover the structures that have just been explained.

## FAULTS.

An observer who should traverse the Connecticut Valley from west to east would soon notice the prevailing eastward dip of the strata, and would therefore naturally assume that he could, in advancing eastward, pass over their natural succession, from the lowest at the western margin of the valley to the highest—or at least to the highest of those still remaining—at the eastern side of the valley; and by a simple calculation he might then obtain the total thickness of the formation. This supposition was the first one entertained by early observers. It is expressed in their descriptions and diagrams; for example, those prepared by Smith in 1832 and by Hitchcock in 1833; the latter probably having given foundation for the more explicit diagram in Le Conte's Elements of Geology,[1] which entails an enormous denudation along the western side of the formation, and an even more extravagant wasting of the crystallines adjoining.

[1] First edition, 1878, p. 440.

The vast dimensions of the series thus measured gave rise to the suspicion that its thickness might be only apparent, and that longitudinal faults most probably occur, whereby a moderate series of strata, several times repeated, would suffice to cover the valley floor. The sudden termination of the monocline against the crystallines on the eastern side of the valley might also reasonably have suggested faulting; but little attention seems to have been given to this aspect of the problem. It is here discussed on page 122, under the heading "Marginal faults." The faults that traverse the Triassic lowland will be first considered.

Before 1880 no specific location or measurement of any faults was accomplished. Since then many fault lines have been well identified and their throws measured in Massachusetts [1] as well as in Connecticut. This has been done in the manner customary among stratigraphic geologists. The location of the fault planes was not defined by direct observation, but by determining points at which certain members in the normal sequence of stratification began to be repeated.

In my own work the first suggestion of faulting came in the neighborhood of Meriden. It was there noticed that a series of beds, consisting of conglomeratic sandstones at the base, an amygdaloidal trap sheet of moderate thickness, a series of sandy shales with an impure limestone among their lower members, a heavy sheet of trap several hundred feet in thickness, more sandy shales, another thin sheet of trap, and then an indefinite thickness of shales and sandstones, was repeated several times in passing from west to east over the Hanging Hills, Lamentation Mountain, and Higby Mountain. Faults were naturally suspected in the valleys between these ridges, but they were not at first demonstrated and located. As now stated, it seems surprising that so ordinary a supposition should not have been made at an earlier date. The delay in its announcement is explained better by the prevalence at that time of the belief in the intrusive origin of all the trap sheets in the Meriden district than by the failure to recognize the repetition of the sedimentary series.

It is truly curious that the clear announcement in Percival's report of the constant relation of the three trap ridges, anterior, main, and posterior, always associated in this order wherever they occurred, should not long ago have suggested that their repetition was due to faulting; but it must be remembered that they were generally looked upon as intrusions, or even as dikes, and that methods of study that might be applicable to normal aqueous strata in the Appalachians of Pennsylvania, for example, were not considered appropriate in a region where the stronger ridges were formed of igneous rocks. It was, indeed, distinctly against the custom of the time to apply ordinary stratigraphic methods to the study of structures in which igneous rocks were so prominent; but it gradually came to be perceived that

[1] See B. K. Emerson, Am. Jour. Sci., 3d series, Vol. XXIV, 1882, p. 195, and Bull. Geol. Soc. America, Vol. II, 1891, p. 455.

such methods were entirely justifiable whenever the igneous rocks were contemporaneous lava beds, and it is no exaggeration to say that upon the recognition and introduction of this simple principle the interpretation of the structures of the region has turned. It was manifestly necessary to determine, in the first place, the character of the trap sheets—not their composition and their place in the classification of igneous rocks, but their structural relations and their place in the Triassic formation; hence the careful search along the numerous trap ridges to discover the significant upper contacts, as already detailed in previous pages. The direct and accordant testimony of the vesicular upper surfaces, of waterworn trap fragments in the overlying sandstone, of infiltrated sands and muds in the loose texture of the trap surface, led so inevitably to the conclusion that the sheets of the eastern ridges were extrusive, and the conclusion was so strongly fortified by the contrasted testimony of the dense upper surfaces of the sheets of the western ridges, that belief was demanded rather than allowed. The use of the trap ridges as locating the more resistant members of a great series of bedded rocks thus came to be regarded as entirely legitimate; and the success with which this method of study has led to the explanation of many more facts than it was first expected to explain justifies it in the fullest manner.

Although the trap ridges here referred to overlap for moderate distances in a very systematic manner, as was clearly stated by Percival, it was not at first perceived that this might be due to their truncation by oblique faults. This significant discovery was not made until it was noted that the anterior ridge southwest of Shuttle Meadow, near New Britain, was dislocated and its two parts overlapped to a small extent by faulting. It was then undertaken to locate and trace out the fault line between the Hanging Hills and Lamentation Mountain. The systematic oblique truncation of one member of the series after the other, aqueous as well as igneous, was thus first clearly perceived, and the essential clew to the structure of the valley was discovered. In looking back to these successive steps of interpretation, the advance through them seems remarkably slow. It is now possible to lead a party of observers to the more significant localities in a day's excursion, and then, from some eminence, such as Chauncy Peak, to expose clearly in half an hour what has required the intermittent search of a number of years gradually to bring forward from the unknown.

### GEOMETRICAL RELATIONS OF FAULTS.

The demonstration of the existence of faults involves essentially the same method of investigation as that already detailed in the case of the trap sheets. The repeated occurrence of a certain succession of bedded sandstones and lavas was perceived, and two alternative explanations were presented: that these repetitions represent actual imitative repetitions in deposition, or that they are outcrops of a single sequence of deposition, now repeated by faulting. The first supposition becomes untenable when the number of repeated sequences is found to be much larger than three, as at first observed, and when a peculiar and highly specialized arrangement of outcrops is found to accompany the areas of repeated sequences. The second supposition is forced upon belief when it is discovered that all of its appropriate consequences are to be found in the field, and that the geometrical deduction of these consequences indoors often enables the observer to anticipate observation outdoors. In order to appreciate this it is important that the geometry of the structural problem here involved should be clearly set forth, for only then will the consequences that necessarily follow hypothetical faulting stand out, ready to be confronted by the facts of observation.

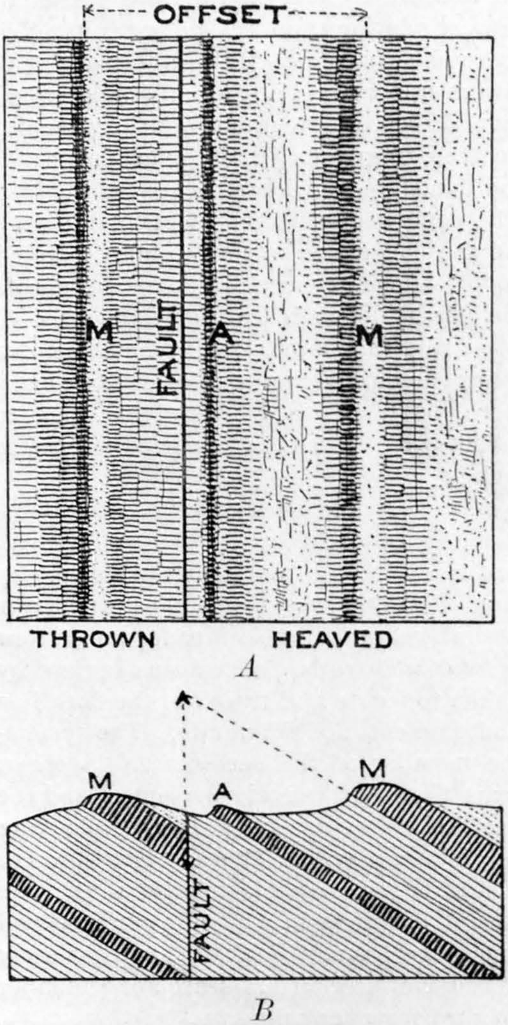


FIG. 15.—*A*, plan of strike fault, heaved on east; *B*, section of fault, heaved on east.

The dip of the monocline being eastward, let the fault line be at first assumed to be north and south, parallel to the strike of the strata, the hade vertical, and the heave or upthrow toward the dip. The effect of this kind of displacement will be, after deep erosion, to repeat a certain sequence of strata, their number depending on the heave and the dip, and their visibility depending on their hardness, as in fig. 15, *A* and *B*. Corresponding members in the two sequences will be set apart or offset by a distance that equals a simple function of the heave and the dip.

Now let the fault line have any oblique position. If heave and dip remain unchanged, the offset is also unchanged; but, instead of an indefinite continuation of each sequence along the strike, the successive members are now systematically cut off, the ends of corresponding members overlapping by a distance that is a simple function of the offset and the angle between the strike of the strata and the trend of the fault. According as the fault trends to right or left of the strike, the offset and overlap will be arranged in what Percival called "advancing" or "receding" order, as in figs. 16 and 17. If an observer cross a district thus faulted, in the direction of the dip, he will find some strata to be repeated, and the breadth of the repeated strata will measure the offset.

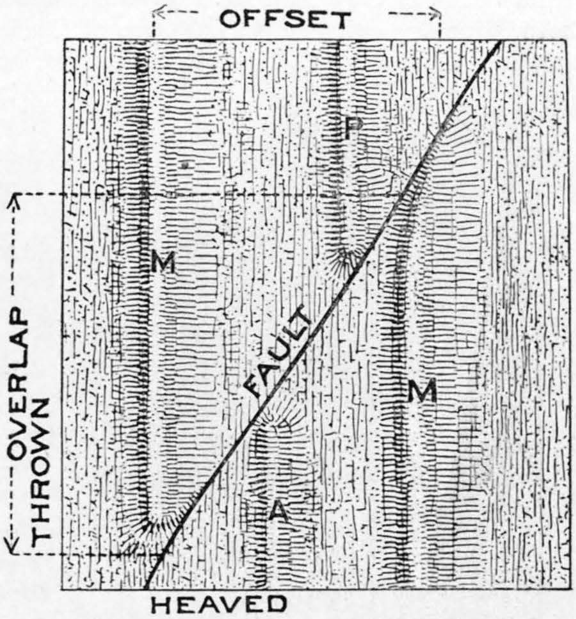

FIG. 16.—Plan of NE.-SW. fault, heaved on east.

Another series of cases may be considered in which the heave is on the side of the fault opposite to the dip. These are illustrated in figs. 18, 19, 20. When the fault is parallel to the strike, certain strata will fail to reach the surface; their breadth may be called the loss of the fault. When the fault is oblique, full sequences of strata will appear on either side, one being again offset with respect to the other, and the offset equaling the loss as just defined; but instead of an overlap, as in the previous examples, there will be a certain space between the ends of corresponding members in which they fail to appear or lapse.

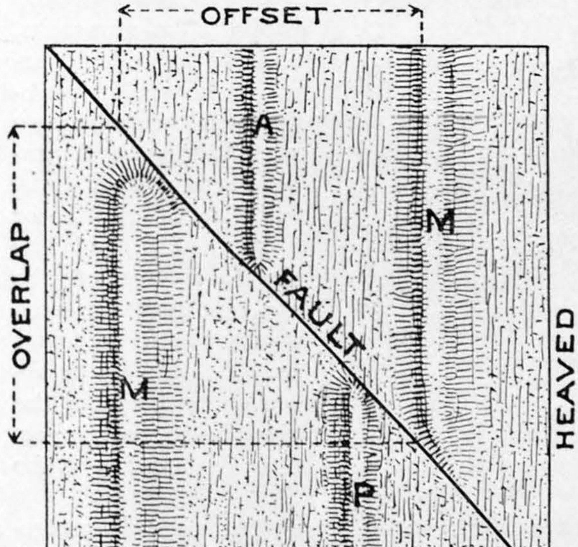

FIG. 17.—Plan of NW.-SE. fault, heaved on east.

If an observer cross the fault line in any of these latter cases, some strata will be missing, instead of repeated as in the previous examples; the missing breadth being a measure of the loss. Curvature of the fault line will leave offset or loss unchanged, but will alter overlap or lapse; variation of heave will alter both the offset and the overlap, or loss and lapse.

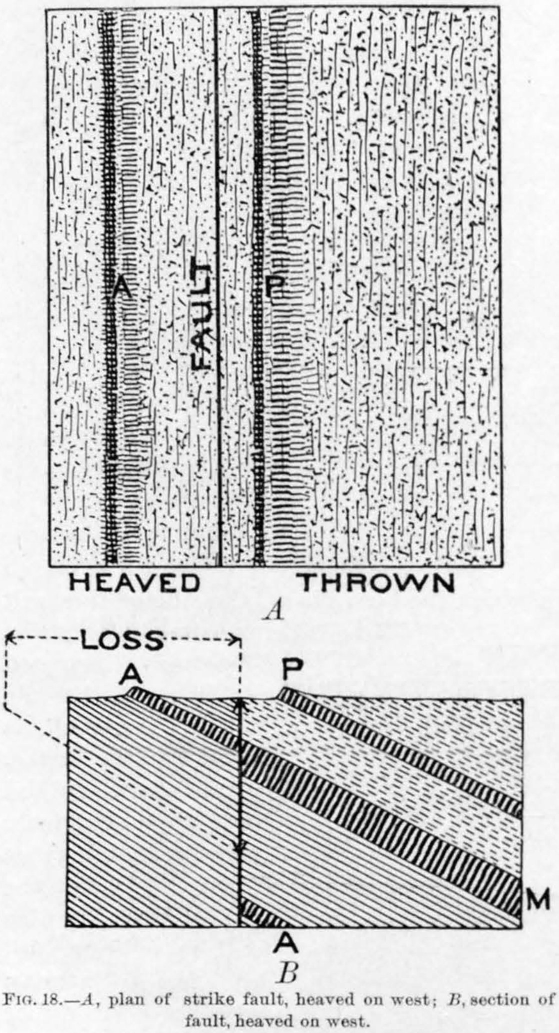


FIG. 18.—*A*, plan of strike fault, heaved on west; *B*, section of fault, heaved on west.

In case a vertical dike occurred in the monocline, its outcrop after peneplanation would not be displaced by a vertical fault on a vertical fault plane. If the line of fault movement made a small angle with the plane of the dike, the outcrop of the dike would be slightly displaced, but much less than the outcrops of the inclined members of the monocline.

If the faults should prove to be numerous, and sympathetically arranged in parallel or subparallel courses, they would divide the formation into a number of "blocks" of greater or less breadth, and the true sequence of the aqueous and igneous strata would be found only so long as the observer kept within the limits of a single block. This is manifestly a principle of great importance, not only in the interpretation but also in the exploration of the region.

Besides these consequences, in the way of purely geometrical arrangement, as seen in horizontal plan, there are others of importance that affect the shape of the ridges formed by resistant strata. In fig. 16 (p. 91), for example, the northern end of the ridge on the heaved side of the fault will slowly decrease in height as its determining stratum is obliquely truncated, and it will finally dwindle away in an inconspicuous trail of fading ledges; but the southern end of the ridge on the thrown side of the fault will terminate abruptly in a bold bluff, the result of weathering on the acute point of its determining stratum. This control exerted by faults on form is more fully discussed on page 169.

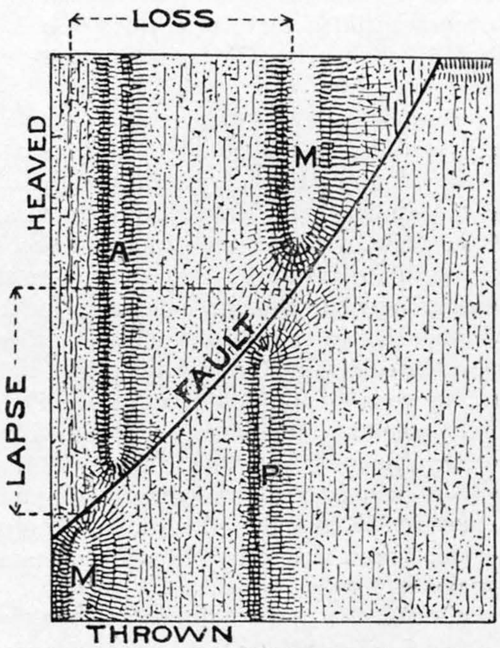


FIG. 19.—Plan of NE.-SW. fault, heaved on west.

Certain special structural features may also be expected along the line of the fault. The adjacent strata may be more or less dragged, bent, and broken by the movement. The fault may vary from an almost smooth joint plane to a broad, irregular fracture, filled with breccia, more or less cemented. Subordinate branching of lateral fractures may be associated with a master fault. The fault faces may be more or less striated or polished. The attack of erosion upon a belt of fault breccia may often reduce it to a notch or valley, this being particularly important where a fault of small throw crosses a trap ridge.

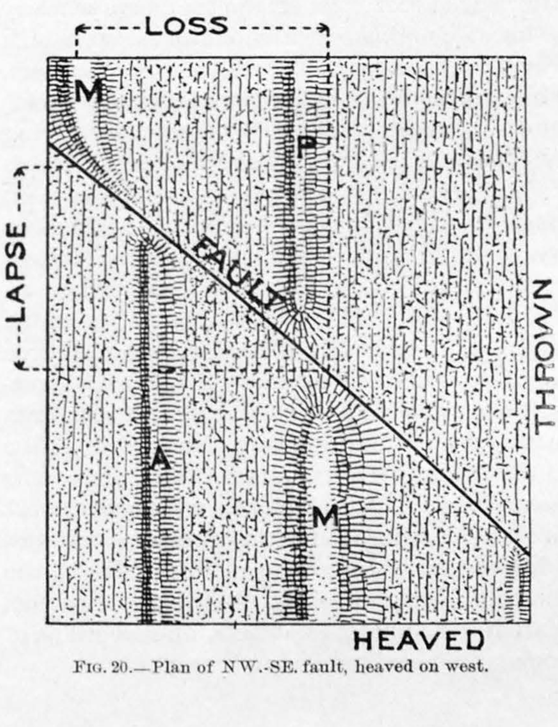


FIG. 20.—Plan of NW.-SE. fault, heaved on west.

In a region of numerous outcrops or of readily identifiable strata there would be no need of so deliberate an examination of an elementary structural problem of this kind; but when the surface of the country is largely covered by drift, when the sedimentary strata are identifiable only in a most general way, and when the identification of the trap sheets depends on so unessential a feature as thickness, it is proper that close scrutiny should be given to all aspects of the problem in order to insure its correct solution.

After the valley has been explored, or a geological map of the region inspected, there can be no hesitation in deciding that the hypothesis of frequent repetition by faulting is vastly superior to the hypothesis of the independent production of imitative sequences. Under the latter view even the repetitions of the sedimentary strata would be extraordinarily improbable, and when the igneous rocks are added repetition by imitation becomes simply incredible. It is here immaterial whether the trap sheets of the eastern ridges are intrusive or extrusive. In either case they form systematic members in sequences of aqueous and igneous strata, and it is inconceivable that so many systematic arrangements could have been gained independently in the various series, either by sills or by flows. All the apparently independent sequences must really be dissevered portions of an originally single structure. Rational explanation must give place to the most gratuitous assumptions if any other view is maintained.

### TEST OF HYPOTHESIS OF FAULTING.

Abandoning the supposition of imitative sequences, we may test the hypothesis of repetition by faulting by its ability to explain all the structures of the valley—except the warps already described—on the postulate that the entire formation consists of the members already stated, namely, a mass of lower conglomeratic sandstones, with sills or dikelike intrusions near the base in the northern and southern areas; then the anterior trap sheet and shale; the main trap sheet; and the posterior shales and trap sheet; followed by a great overlying series of shales and sandstones. The sedimentary members of this series are so open to variation, especially on approaching the eastern border of the formation, that reliance can be placed upon them as guides to faulting only over short distances. The only proved exceptions to this statement are the anterior limestone and the anterior and posterior black shales, with their assemblage of fossil fishes, to which special reference will be made below (p. 137). It is chiefly to the three lava beds of the eastern ridges that we must look as guides to the structure of the region. They are singularly constant over the whole area studied, the anterior sheet alone thinning out and disappearing near the northern boundary of the State. They now stand up above the generally weak sandstones and shales, and above the widespread drift cover, except for drumlins that sometimes bury the anterior and posterior ridges. The trap ridges thus form conspicuous guides to the position of certain horizons standing about midway between the bottom and top of the whole formation. The notches, offsets, and overlaps of the ridges are so many proclamations of breaks and faults.

Although this would be measurably true even if the eastern ridges were determined by a group of three intrusive sills, it becomes necessarily true if they are formed on extrusive flows. It was when this principle was perceived that a special effort was made to establish the true nature of the trap sheets beyond all question. In the belief that this has been done, and that the extrusive nature of the sheets in the eastern ridges is thoroughly demonstrated, they have been used in deciphering the stratigraphy of the region as confidently as the Medina and Pocono sandstones have been used in untangling the stratigraphy of middle Pennsylvania.

In the following detailed record of observations the most convenient method of testing the theory of faulting will be to accept its truth for the time and to employ its language, finally reviewing the whole accumulation of facts to determine the nature of the proof that they afford. In the account of the faults that have been thus identified, and of the numerous blocks into which they divide the formation, mention can be made only of the chief occurrences. Each block will be named after the ridge formed on its portion of the main trap sheet; but, unfortunately, it has been necessary in several cases to invent names for these ridges in default of names locally applied to them. The breadth of the blocks and the throw of the bounding faults are measured in the belt of the eastern ridges, unless otherwise stated. The strata recognized in each block are briefly described, and the fault limiting it on the north or west is traced. Particular attention is called to the Lamentation block, which exhibits the structural features produced by faulting more distinctly and over a larger area than any other. The minor faults, that cause notches in the main sheet and slight offsets in the anterior and posterior ridges, are located as well as possible on the map of the region, but are not in all cases described in the text. They are of doubtful location in the New Britain region. The marginal faults, continuous along the eastern and intermittent along the western boundary of the formation, will be specially described on pages 122–131. Frequent reference to the map, Pl. XIX (in pocket at end of volume), will be necessary in following the statements of the text.

TABLE OF FAULTED BLOCKS.

The following tabular statement, based on profiles constructed by Mr. Griswold, may serve as an index to the detailed description of the faulted blocks in the succeeding sections. It begins at the south. Many of the measures are only rough average values of indefinite or variable dimensions. A minus sign before the throw indicates uplift on the north or west. The fault values are measured on the line of the main ridge unless otherwise stated.

*Faulted blocks described in succeeding sections.*

| Block. | Length, main sheet bluff. | Breadth of block. | Throw of north bounding fault. |
|---|---|---|---|
| | *Miles.* | *Miles.* | *Feet.* |
| Totoket:[1] | | | |
| Saltonstall crescent axis | 5 | 4–4½ | —2,500 |
| Totoket crescent axis | 7¼ | | |
| Paug | ¾ | ⅝ | + 500 |
| Tremont | 2½ | 2 | ........... |
| Higby | 6 | 4½ | +1,300 |
| Lamentation[2] | 3¼ | At West. Upland, 1¼ | +2,000 |
| | | At Lamentation, 2 | +3,000 |
| | | At East. Upland, 2¾ | ........... |
| Cedar[3] | 3 | ........... | +4,000 |
| Hanging Hills | 3, N—S. | 2¾ | ........... |
| The subdivisions are— | | | |
| Quarry | ........... | ¼ | ........... |
| Cat-hole | ........... | ½ | + 300 |
| South | ........... | ½ | + 300 |
| West Peak | ........... | 1½ | + 700 |
| Short | ½ | ½ | — 200 |
| Ragged | 1½ | 1½ | + 600 |
| Farmington | 6-7 | 3 | —1,500 |
| Bradley | 1¾ | The throw here given decreases rapidly to the northeast and north. | |
| Farmington | 5 | | |
| East Talcott | 6 | ¾ | +1,200 |
| West Talcott | 6 | (?) | ........... |
| Simsbury | 4½ | 2-2½ | 170 |
| Granby | 3 | 1½ | 250 |
| Suffield | 6¼ | 2¼ | ........... |

[1] The bounding fault on the southeast of this block is the marginal fault of the valley lowland; its throw here is 6,000 or 7,000 feet, or more.

[2] The Chauncy Peak fragment of this block is three-fourths mile wide; its fault is + 400 feet.

[3] This anticlinal uplift is 6 miles wide at its western faulted end, as measured along the outcrop of the main sheet, or 11½ miles, as measured between the ends of its posterior ridges. At its eastern end the main sheet is not seen; the posterior ridges are 1½ miles apart. Its greatest throw at the west end is + 400 feet.

## DESCRIPTIONS OF FAULTED BLOCKS.

*Totoket block.*—A large block of the formation, including the crescentic ridges of Saltonstall and Totoket mountains, is cut off by an oblique fault trending northeast from Northford to the crystalline boundary in Durham, and thus is divided from the adjoining narrow Paug block.

When the southwest parts of the main trap sheet in Totoket Mountain and its anterior ridge are compared with the corresponding members in the Paug block, it is seen that the latter recede eastward from the strike of the former, and that there is a lapse of about 2 miles between the anterior ridges in the two blocks. The uplift of the fault must therefore in this case be on the northwest, with a movement of about 2,500 feet in the vicinity of Northford. But the somewhat irregular warping of the members of the Totoket crescent carries the northern part of the main sheet east of the fault line, and brings up the anterior sheet north of the crescent, where it forms an ash ridge and a considerable area of irregular trap hills in the northeast corner of the block. A local increase of dip to over 30° is found in some sandstone outcrops near Northford, as if produced by the drag of the fault.

The southwest prolongation of the fault line carries it to the neighborhood of East Rock, New Haven. Here the notches between East Rock, Indian Head, and Snake Rock, and the apparently offset arrangement of these ridges, are strongly suggestive of faulting rather than of accidental erosion or of irregular intrusion (see p. 133). It is therefore provisionally suggested that the Totoket-Paug fault line may pass near these smaller fractures. The absence of identifiable sequences of strata in the intermediate districts and the great extent of drift and marsh make it seem hopeless to settle this question definitely. A fault was traversed by the new tunnel of the Shore Line Railroad in Fairhaven; the sandstones were much broken and slickensided, and the tunnel, therefore, had to be lined with heavy brickwork.[1]

If the above-suggested course for the limiting fault be accepted, the Totoket block, including both the crescentic ridges, possesses representatives of a great part of the Triassic series: the lower conglomeratic sandstones, frequently cut by dikes east of Fairhaven; the anterior lava bed, scoriaceous and ashy for much of its length and thickness; the anterior sandstones and shales, with an impure limestone near their base and highly fossiliferous black shales at a little higher horizon both northwest and northeast of the northern hook of Totoket; the heavy main lava bed expressing in its two crescentic ridges the curious warped attitude into which all the members of the series have been thrown; the posterior sandstones and shales, often containing coarse conglomerates and twice exposing the belt of posterior black shales containing fossils; the posterior lava beds; and a moderate thickness of the upper sandstones, often conglomeratic. It is noticeable that in this block the anterior lava bed and its associated sedimentaries reach the eastern boundary of the formation, this exceptional arrangement being due to the equally exceptional warping which characterizes the block. Not before reaching the northern end of the great Springfield crescent, near Amherst, Massachusetts, is this arrangement repeated.

Several small faults have been noted in the Totoket and Saltonstall Mountain areas. One of these produces a displacement of a few feet in the under surface of the main sheet on the northern hook of Totoket Mountain, and erosion has here made a small notch in the face of the bluff. Accordant displacements in the main and posterior sheets east

[1] E. O. Hovey, Am. Jour. Sci., 4th series, Vol. III, 1897, p. 289.

of the Branford-Totoket road in the northeast part of the Saltonstall Mountain crescent are referred to a fault of 50 or more feet throw, trending north-northeast. Several notches in the southern part of Saltonstall Mountain are ascribed to erosion along fault lines, the northernmost of these being used by the Shore Line Railroad and the next one by the highway from New Haven to Branford. Although the fault lines are not known to be parallel, the heave in these several small faults is on the east or southeast, and the advancing order of offset and overlap thus produced contributes to the curvature of the south end of the Saltonstall Mountain crescent. The underlying sandstone is seen to be dislocated with the trap, and weathered breccias are seen in some of the notches. It should be stated that Hovey ascribes the irregular form of Pond Mountain hereabouts, with its notches and the breccias in the notches, to "irregularities in the original fissure rather than to faulting after the trap was in place;"[1] but the explanation by faulting is the only one admissible if the extrusive origin of the main sheet is accepted; and it is confirmed by the occurrence of breccia in some of the notches. It is believed that the frequent dislocations in this locality have so greatly weakened the anterior sheet as to allow it to be worn down and covered by drift. It is not to be found in East Haven south of the railroad, where the ridge ends in a distinct bluff.

*Piscapaug block.*—The main lava bed in this narrow block forms Piscapaug (commonly abbreviated to Paug) Mountain, of irregular outline, accompanied on southwest and northeast by the anterior and posterior lava beds, the latter being irregularly warped and broken. The structure here is much more uncertain than the definite coloring of the map would indicate. The posterior fossiliferous shales are found between the main and posterior lava beds. The southwestern extension of the block can not be defined. Where the north-bounding fault cuts the posterior lava bed it is broken into several knobs, as if the fault were separated into several lines of fracture.

*Tremont block.*—The deep oblique notch holding Piscapaug Pond separates Tremont Mountain from Piscapaug Mountain, the former following the latter in advancing order with distinct overlap. On the north, Reeds Gap, utilized by the Air Line Railroad, separates Tremont Mountain from Higby Mountain. Here the offset is again in advancing order, but the overlap is hardly apparent, and a nearly square cross-fault would therefore be inferred. This inference is confirmed by the position of the notches in the anterior and posterior ridges. Beyond these points it has not been possible to trace the north-bounding fault line of this block. There would appear to belong in this block, however, the lower sandstones and conglomerates, with many dikes from Clintonville to East Wallingford; the anterior, main, and posterior lava beds, with accompanying sandstones and shales; and a considerable thickness of upper sandstone, containing some fossiliferous shales

[1] Am. Jour. Sci., 3d series, Vol. XXXVIII, 1889, p. 378.

northeast of Round Hill and becoming conglomeratic near the crystalline border. North and northeast of Durham a broad, flat, anticlinal dome replaces the usual monocline; its northwestern dips may be associated with the north-bounding fault of this block.

The anterior ridge disappears beneath the drift before the southern side of the block is reached, but it can not be said that the anterior lava bed is there wanting. A mile west of Piscapaug Pond the anterior limestone was quarried early in the century, according to Percival.[1] A subanterior conglomerate begins to be conspicuous in this block, and may be thence traced more or less continuously to the Lamentation block, north of Meriden. It is of interest as occurring close along the medial axis of the Valley Lowland, and presumably removed by at least from 6 to 10 miles from the shore line of its time. When this is considered in connection with the fine black anterior shales that approach so close to the crystalline border at the north end of the Totoket crescent, it becomes manifest that however potent the distance from shore line may have been in determining texture of deposits, other potent controls also had a share in deciding the character of materials that were strewn over the floor of the ancient estuary.

A fault of slight throw is thought to divide this block into nearly equal parts. It is peculiar in producing a slight recession of the main sheet and advance of the posterior, as if the fault movement changed from heave to throw in a distance of about 2 miles. It is further noteworthy that the space between the main and posterior ridge widens southward on the southern side of this block, this being fairly explained as the result of a local southward decrease of dip observed in the belt occupied by the posterior shales and sandstones. Several small faults are indicated by ravines that cut the back of the main ridge, descending obliquely to the northeast instead of directly eastward, as they would if guided by the slope of the sheet alone.

*Higby block.*—This large block is ill defined on the south, except by the three notches in the anterior, main, and posterior ridges at Reeds Gap, already mentioned; but on the north its bounding fault is thought to be traceable intermittently for 35 miles, passing all across the valley lowland on a strongly oblique course and extending into the crystallines in the northeast and southwest, as will be described in a later section (p. 102). It thus includes a full section of the entire formation.

Beginning on the southwest, the long-continued West Rock ridge is taken as belonging with the slightly broken Higby Mountain. Between the two are the lower sandstones, with the massive stocks and dikes of Mount Carmel and the many smaller dikes west of Wallingford. Somewhat below the middle of this great series of strata there appear to be some layers that are a little more resistant than the others, thus determining the flat-topped sandstone ridge known as Quinnipiac Ridge, that separates the valleys of Mill and Quinnipiac rivers. The

[1] Geology of Connecticut, 1842, p. 363.

ridge may be traced from the northern side of the block, about $1\frac{1}{2}$ miles northwest of Wallingford Center, along a south-by-west course, narrowing as it goes, but with little interruption, to New Haven. It does not seem to be displaced by the fault from Reeds Gap; hence there is no need of separating the Tremont and Higby blocks in this part of the formation.

The great breadth of country covered by the lower sandstones of the Higby block is less abnormal than it appears at first sight; for, in the first place, the dips over most of the area are gentle, commonly about 15°; and, in the second place, the strike of the lower strata is more oblique to the sides of the block than is the case with the middle division of the formation, the former being somewhat warped with respect to the latter. A not immoderate thickness of lower sandstones is thus deployed over a large area of the lowland. Then come the three lava beds, with the anterior and posterior sandstones and shales, the fossiliferous beds being very appropriately found in the anterior and posterior valleys, close to the north side of the fault. A great thickness of upper sandstone follows, with conglomerates near the eastern boundary. The peculiar trap conglomerate of South Glastonbury, presumably equivalent with the posterior lava bed of the Cedar Mountain-Rocky Hill warp, belongs in the Higby block.

Three small faults are indicated by notches in the anterior ridge of Higby Mountain, a well-defined advancing offset and overlap occurring with the northernmost. Only the middle one of these faults can be traced over the main ridge, through Black Pond Gap, where a dislocation is clearly indicated by the strong bluff of trap on the northwest side of the gap road, trending obliquely northeast. If the line thus defined is projected about 3 miles farther it leads to a well-marked oblique valley through the posterior ridge. The throw of the fault must be small, as it nowhere produces a distinct offset. An indentation in the northern end of Higby Mountain, opposite Highland station, is probably due to a fault, as otherwise no reasonable cause could be suggested for its erosion. The gradual dying out of the main ridge, as the fault cuts obliquely across it, is an excellent example of the control that the unseen displacement has on the termination of the ridges formed on the massive lava bed of the main sheet. The greater breadth of the posterior shales and lava beds north of Middlefield is due to a decrease of eastward dip.

*Lamentation block.*—This well-defined block with its submember, the Chauncy block, is in many respects the most remarkable example of faulted structure in the valley. The north-bounding fault, by which it is strongly separated from the Hanging Hills block, is traceable for 40 miles, even a greater distance than the south-bounding fault, already mentioned with the Higby block. The strongly oblique course of the fault lines and the strong movement upon them has produced striking offsets and overlaps, as well as numerous local variations of dip, and "dikes" or belts of breccia. The offset on the northwest brings the lower sandstones of the Lamentation block against the posterior and upper shales and sandstones of the Hanging Hills block, and thus determines an open belt of low ground a mile wide between the trap ridges on the east and west. Along the pass thus defined the chief railroad of the valley finds easy passage from the great lowland area of lower sandstones between Meriden and New Haven to the great lowland area of upper sandstones between Hartford and Springfield; and long ago this passage was pointed out by Percival as "determined by the geological arrangement of the country."[1]

The members of the Triassic formation represented in this block are the lower sandstones, with the intrusive sheet of Bethany Mountain and the crossed dikes of southern Cheshire, the sandstones being distinctly conglomeratic in their basal layers near the crystallines and in the upper layers near the anterior trap ridge; the three lava beds and their intervening shales and sandstones; a considerable area of upper sandstones and shales, complicated by the reappearance of the posterior shales and lava beds due to faulting and warping, as is more fully explained on page 107, and a series of sandstones and conglomerates from Glastonbury to Talcottville, of problematic position in the series, but more likely belonging near the bottom than near the top of the formation, as will be shown later (p. 130). The anterior shales contain a limestone near the anterior lava bed, mentioned by Percival,[2] but not now visible, and also some fossiliferous black shales, disclosed by trenching across their expected position (p. 138). The direct distance from the anterior ridge, northeast of Meriden, obliquely across the lower sandstones to the crystalline border beyond Bethany Mountain, is 13 miles, but the transverse measure of the lower sandstones square across their average strike is only 4½ miles. With an average dip of about 15°, the thickness of this division of the formation would be 5,000 or 6,000 feet in this block.

The two trap ridges east of East Berlin are assumed to be repetitions of the posterior lava bed by branch faults, but there is no complete proof that this assumption is correct. The chief reason for its acceptance lies in the great improbability that an additional lava flow should have been poured out in this district long before the time of faulting, yet having its edges prophetically agreeing with the margins of the block as determined by long subsequent dislocation. The discovery of the belt of fossiliferous black shale that ordinarily lies about a quarter of a mile west of the posterior ridge would be a desirable confirmation of this view. As in so many other cases, the faults that are believed to have produced these repetitions are first drawn so as to cut the ends of the ridges, but their location is generally confirmed by the occurrence of fault breccias or distorted dip at various points along their lines.

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[1] Geology of Connecticut, 1842, p. 486.
[2] Ibid., p. 365.

The double repetition of the posterior lava bed on either side of the arched warp in Rocky Hill is more satisfactorily determined. The identity of these outcrops was first suggested by their relation to the heavy trap sheet, presumably the main sheet, of the Cedar Mountain triangular block; it was afterwards practically demonstrated through the discovery by Mr. Rice of the posterior belt of fossiliferous shales in a position appropriate to the small trap ridge near Rocky Hill village. Additional outcrops of these shales would be very welcome witnesses to the scheme of structure here advocated; but, in spite of close and persevering search, they have not been found.

The following notes illustrate the character of the evidence by which the long-continued bounding faults of this block are defined. Beginning on the southeast side of the block, Bethany Mountain is seen to be distinctly offset in advancing order from the northernmost irregular portion of West Rock ridge, and half a mile to the southwest there is a corresponding offset in the margin of the crystalline uplands. Joints occur in increasing numbers in the trap on either side of the gap, as if the sheet had been locally disturbed. The crystalline schists, whose trend hereabouts is generally regular, are locally deformed, brecciated, and slickensided along the projected trend of the fault to the southwest. To the northeast a well-defined break occurs near the southern end of the Cheshire dike, producing an offset of about 50 yards in a gap trending N. 40° E., in which a trap breccia is found. The path of the fault across the line of the Wallingford dike is not indicated by any systematic offset, but it must be remembered that a nearly vertical fault need not offset a nearly vertical dike. Quinnipiac Ridge appears to gain greatly in breadth west of Wallingford, but this is thought to be only the effect of several advancing offsets caused by successive minor faults into which the bounding fault appears to be divided in this district. It should be understood, however, that the fault lines are not identified as such; that Quinnipiac Ridge is very generally drift-covered; and that the existence and arrangement of the faults are inferred entirely from the general form of the ridge. Within an area of heavy drift on monotonous sandstones, beyond Yalesville and southeast of Meriden, no decisive evidence of faulting is found; but at several points there are oblique northeast-southwest valleys, interrupting the low sandstone ridges, and these are most likely due to erosion in fault breccias.

Then come the well-defined offsets and overlaps of the recognizable members of the middle division, indicating a heave of at least 1,300 feet, and producing the bold south bluff of Chauncy Peak, in contrast to the long north trail from Higby Mountain. Near the end of the latter, where the trap is crossed by Fall Brook, the trap is much brecciated, increasingly so toward the oblique fault line, which is here well defined. A generalized section of this locality is given in fig. 21, looking north. A band of breccia in the trap is weathered out as a small chasm trending northeast, transverse to the stream near its chief fall. The very face of the fault is disclosed at the side of a pool where the stream flows off the trap; a few layers of shaly sandstone lie against the trap, with a strong reverse dip to the northwest. The heavy trap sheet, which comes for a number of miles from the south in strength enough to make a mountain, here disappears, and the line of its strike runs through meadows. A little farther on the sandstones have reverse dips of from 5° to 40° NW., and bands of breccia are seen trending N. 45° to 55° E. in the Meriden, Waterbury and Cromwell Railroad cut a short distance west of Westfield station; these give every reason for thinking that similar disturbances would be found if similar exposures were made at any other point along the fault line.

The southern termination of one of the East Berlin supernumerary posterior ridges falls on the same line almost opposite the northern point of the trailing main sheet of Higby Mountain. The eastern end

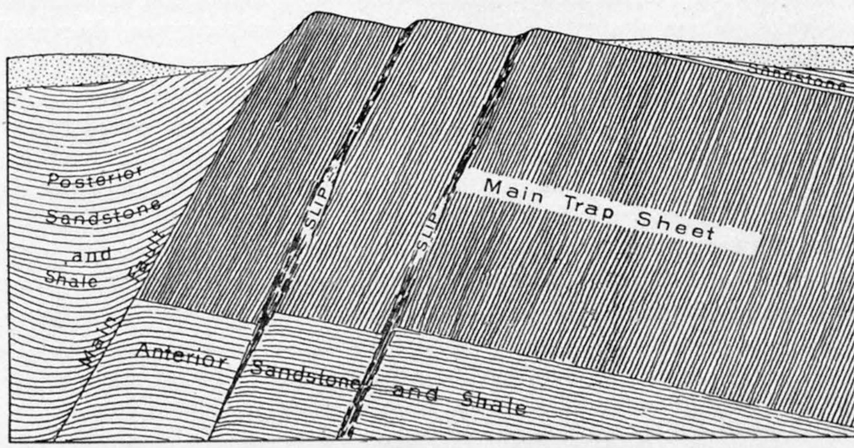


FIG. 21.—Diagram section of north-bounding fault, Higby block.

of the posterior ridge on the south side of the Cedar Mountain anticline defines the extension of the fault line close to the Connecticut River. Across the river, in South Gastonbury, the trap conglomerate is dislocated from its probable equivalent, the Rocky Hill posterior ridge. Then comes the sudden change in the course of the lowland margin, turning at this corner from a northward course, that has been maintained from near Middletown, to a northeastern course, now continued for 8 miles to a well-defined corner near the beautiful village of South Manchester, where the northward course is resumed. No sign of the contact of sandstones and crystallines is seen in this distance of 8 miles; indeed, only one actual contact has been found along the eastern margin, and that by trenching; but a line drawn so as to separate the outcrops and forms appropriate to the upland and lowland rocks accords closely with the prolongation of the fault line already traced. Following the same direction beyond the South Manchester corner into the crystalline area, one finds an oblique valley along its course near the border of the upland, and 3 miles beyond is the deep depression of Bolton Notch. Here the cut of the New York and New England Railroad discloses numerous fault breccias of small movement, with a northeast trend, suggesting that breccias of greater movement may be worn deeper in the notch and buried in drift. Still farther northeast the depressions of the Willimantic reservoir and of several villages on either side of Tolland fall into the same line. Their value as evidence of faulting can not be measured until the structure of the crystalline upland is deciphered.

It is true that drift obscures by far the greatest part of this fault line, and that in certain parts of its length no direct indication of faulting is found, while in certain other parts evidence of complicated faulting is discovered, as near Westfield; hence the proof of faulting is by no means continuously and uniformly maintained. But when all the facts are considered, it is difficult to resist the belief that a strong dislocation has taken place along the line pointed out by so many peculiar and accordant witnesses.

The northwestern side of the long Lamentation block is defined with a similar measure of certainty. For much of its length the chief fracture seems to be accompanied by a number of parallel subordinate fractures, chipping the sides of the adjacent blocks. It begins with the strong advancing offset and overlap of the south end of Gaylord Mountain from the north end of Bethany Mountain, accompanied by a corresponding offset in the margin of the crystalline area, indicating movement of nearly 2,000 feet. In the south end of Gaylord Mountain the trap is much more jointed than usual. Subordinate faults a quarter of a mile southeast and northwest are indicated by oblique notches in the ridges of the intrusive sheet, both showing a slight advancing order of displacement. The prolongation of the fault line into the crystalline area leads to local deformation, breccias, and slickensides in the schists, as on the other side of the block. Passing the drift-clogged valley of Mill River, the west end of the Wallingford dike and the west end of the Cheshire dike fall closely on the line drawn from the Bethany break in the intrusive sheet to the great offset and overlap between Lamentation Mountain and the Hanging Hills group. Recalling that these cross dikes give frequent indication of top contacts, showing that their exposed ledges are near their original summits in the Lamentation blocks, one perceives it is natural that they should not be visible in the next block to the northwest, which is less uplifted by about 2,000 feet. On approaching the main fault line, the Cheshire dike is frequently notched and slightly offset. The Wallingford dike is once notched, but not offset. In this respect each dike behaves consistently on the two sides of the Lamentation block, thus permitting the belief that the fault movements ran in the plane of the Wallingford dike, but a little out of the plane of the Cheshire dike. Near the

SMALL FAULTS IN SHALY SANDSTONES, SOUTH MERIDEN.

northern end of the latter a number of openings were made in former years for barite and copper. As far as they are now accessible they appear to be associated with the compound fractures of the main fault. Quinnipiac Ridge is distinctly cut off by the fault line, and its continuation in the next block to the north has a strong advancing offset. Near South Meriden a fine exhibition of disturbed and reversed dips is found on both sides of Quinnipiac River, west of Hanover Pond. The bluff north of the river and west of the railroad in the next block gives persistent exposure of sandstone ledges on the east slope of Quinnipiac Ridge, outcropping eastward and dipping 10 or more degrees westward for about half a mile along their strike, thus indicating an uplift on the east, already known by the advancing order of the trap and sandstone ridges. Near the eastern limit of their outcrops a roadside exposure exhibits the small faults shown in Plate XII. The minor dislocations of the Quarry Ridge in Meriden will be described in the next section. The fault line is here indicated by a narrow valley, close to which the sandstone or conglomerate ridges are truncated on either side. One of the conglomerate ridges would, if continued along its strike, run directly into the trap sheet of Quarry Ridge.[1] The dip of the strata in this ridge decreases almost to horizontality close to the valley, and the joints are here seamed with barite. The edge of the trap ridge where closest to the little valley is crushed to a coarse breccia, now deeply weathered. On either side of the valley the strike of the ledges northwest of Meriden differs by nearly a right angle, owing to the warp in the Hanging Hills block, described below. Farther northeast, in the belt of low ground where the lower sandstones and conglomerates of the Lamentation block stand opposite the posterior shales of the Hanging Hills block, a number of low ridges formed by conglomerates may be traced one by one northward to their endings, each more easterly member of the series extending farther northward, as an oblique fault would require. North of the invisible fracture the weak posterior shales are seldom visible, the surface being worn down low and covered with drift gravels and sands or with the waters of Beaver Pond. The posterior trap ridge of the first subblock of the Hanging Hills group fails to appear in this low ground, probably on account of weakening by many faults. (See p. 173.)

The northern end of the Lamentation anterior is slightly notched and displaced in advancing order where the Berlin road crosses it. The southern part of the posterior ridge of the Cedar Mountain block stands opposite the anterior valley of Lamentation Mountain, but its precise ending is buried under heavy gravels. Farther on, the bold face of the heavy main sheet of Lamentation Mountain obliquely retreats, obeying the command of the invisible fracture as submissively as the little conglomerate ledges just mentioned. Its trailing end may be followed north of a crossroad and down Spruce Brook. As the fault line is

[1] See Am. Jour. Sci., 4th series, Vol. I, 1896, p. 8.

approached the overlying shales are dragged to a northwest dip, and finally end in a violently crushed and dislocated ledge, beyond which is an open meadow with no outcrop, thus closely repeating the features of the northern ending of Higby Mountain. A little farther on, the projected course of the fault leads along the base of the slope between the higher ground of the Lamentation block and the meadows to the northwest; and here is found a local anticline in the red shales with westward dip toward the fault line, a feature noted by Percival,[1] but finding no explanation until it is brought into line with the various other disturbances of this long-continued fracture. Next follow some isolated ledges of trap, much jointed, without definite monoclinal expression, and unconnected with any of the adjacent trap ridges. These are thought to be chips from the posterior trap sheet, dragged about 400 feet downward in this compound comminuted fracture, and now happily exposed in the present state of erosion, as illustrated in fig. 22. The northern end of the normal posterior ridge is soon reached near East Berlin, two small chips being neatly detached from it, with small advancing offsets and overlaps. The supernumerary posterior ridges of this

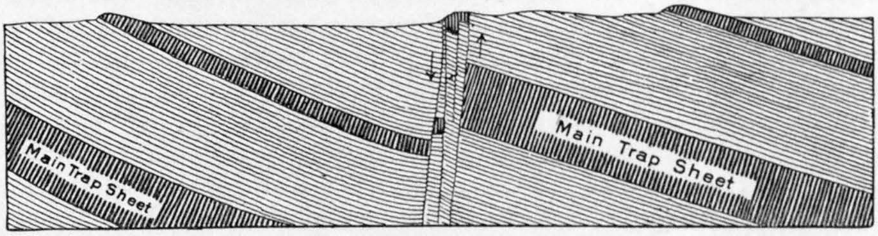

FIG. 22.—Fragments of posterior trap sheet in north-bounding fault of Lamentation block.

district are hardly reached by the main fault, but rather by the medial fault of the Lamentation block, of which more beyond (p. 107).

Nothing could be more satisfactory in the way of geometric evidence of a fault than the broadening of the space between the posterior ridges of the Cedar Mountain anticline as it enters the uplifted Lamentation block, and nothing more unsatisfactory than the 12 miles of drift-covered surface that then intervenes before the Vernon corner of the eastern crystallines is reached. Through all this blank area, where the rare outcrops of normal-dipping sandstone are always to one or the other side of the projected line, the fault is nevertheless believed to pass, for the oblique boundary between sandstone lowland and crystalline uplands stands just in line with it, and is remarkably well defined by sharp corners at Vernon and Rockville, before and after which the general north-south course of the boundary is resumed. This is almost as demonstrative of a fault as the broadening of the Newington anticline. Beyond Rockville the fault line has been probably located by local contortions, breccias, slickensides, and alteration in the crystallines, for about 8 miles, nearly to West Stafford.

[1] Op. cit., p. 357.

Viewing the north-bounding fault as a whole, it may be regarded as somewhat better established than the south-bounding fault of the Lamentation block. The evidence is of essentially the same kind along the two lines, but is of greater quantity on the northern than on the southern side. The essential parallelism of the lines to each other and to the various fault lines that have been traced for shorter distances between many adjacent blocks deserves to be given a high value in the demonstration of their occurrence. The repetition of similar offsets in the western crystalline border on each of these long lines is highly suggestive; but the explanation of the retreating offset in both cases on the eastern border is postponed until it may be considered in connection with the eastern marginal faults (p. 129).

The medial fault of the Lamentation block sets off Chauncy Peak from the southern end of Lamentation Mountain, the throw of the fault here being about 400 feet, producing a very distinct offset and overlap of normal arrangement. The fault is first recognized in an oblique valley in the lower sandstones and conglomerates northeast of Meriden; then in the displacement of the anterior ridge, a small chip being dislocated from the Lamentation anterior on the northwest side of the fracture. As an "example for practice," there is hardly anything in the valley better than this. The anterior of Chauncy Peak terminates directly south of the main sheet in Lamentation Mountain. The trend of the fault line, as thus located, carries it through the deep and narrow rocky notch between Lamentation and Chauncy to a displacement on their back slopes. The further course of the fault is for a time obscured by heavy drift; but a distinct dislocation in the posterior ridge, again with normal offset and overlap, falls in range with the previously located points. Farther northeast the fault is not recognized.

*Cedar Mountain block.*—This abnormal member of the series is wedged in between the Lamentation and Hanging Hills blocks, which would otherwise follow in regular succession all along their length. The southern point of the wedged block is indicated by the Berlin posterior ridge north of Lamentation Mountain. Its western margin must lie between this posterior and that of the Cat-hole member in the Hanging Hills block, and farther north it must lie between the base of the main sheet in Cedar Mountain and the small posterior ridge west of Newington that probably belongs in the West Peak member of the Hanging Hills block. Judging by the moderate height of the main sheet in Cedar Mountain and the ridges that continue it southward, the fault should run near its western base, as otherwise these ridges should be as high as those elsewhere formed by the main sheet (see p. 173); but, except by this argument, no precise position for this bounding fault can be assigned. Its necessary indication by a line on the map is one of the many cases where graphic representation is more definite than the actual state of the problem warrants. Minor faults are indicated by notches in the main ridge. Farther north the main fault must run along the oblique northern termination of Cedar Mountain and of the posterior ridge on which Trinity College stands in Hartford. In a stream bed just north of the end of the main sheet the fault is revealed in a belt of breccia, one of Percival's "clay dikes."[1] At other points between the ends of the main and posterior ridges, and along the curved prolongation of this line into Hartford, just west of the Capitol, the fault is indicated by reversed dips, even to 30° or 45° W., in the sandstones. The reversed or westward dips of 10° or 15°, 10 miles north at Hayden, and thence west toward Farmington River, may be attributed to an extension of the same fault.

The manner in which the Cedar Mountain ridges of the main lava beds are flanked on the north and south by the converging posterior ridges of Hartford and Berlin clearly indicates that this block is an arched warp, with the axis of the arch pitching eastward toward Rocky Hill, where the same pair of posterior ridges is somewhat more uplifted in the Lamentation block, so as to delay their convergence. The main trap sheet in the Cedar Mountain ridges, as well as the Berlin member of the posterior ridges, is dislocated by several small faults whose extension can not be traced. The southernmost member of the Berlin posterior ridges terminates at the northern end in a well-defined hook, around which a series of sandstone layers curve in a perfectly conformable manner, noted by Percival,[2] but then not explained. To day, the hook must be explained as the manifest sign of a warp or drag of the aqueous and igneous strata near the minor fault on its eastern side. In spite of the extremely vesicular and scoriaceous upper part of the lava bed on the inner slope of this hook, no actual contact with the overlying strata has been found (p. 74).

The posterior ridge between Hartford and Griswoldsville does not appear to be faulted. An indefinite series of upper sandstones and shales follows in the northeast, extending across the Connecticut River into East Hartford. The continuation of this block into Manchester and beyond will be discussed in connection with the eastern boundary of the formation (p. 130).

*Hanging Hills block.*—The fine group of the Hanging Hills northwest of Meriden is one of the most picturesque districts within the valley lowland. Although subdivided by several minor faults, the narrow members or slabs into which it is thus broken may all be grouped under one block. The notable change in the course of the strong bluffs formed by the main sheet, so that they trend about east and west in the district northwest of Meriden, is due chiefly to a corresponding change of strike, in which the whole series of layers hereabouts shares; but it is secondarily due to the greater retreat of the main sheet in the narrower eastern slabs of the great Hanging Hills block, as is further explained on page 173. The greater part of Gaylord Mountain,

[1] Geology of Connecticut, 1842, p. 387.
[2] Ibid., p. 358.

including the high summit of Mount Sanford, is the basal intrusive member of this block. Only one short dike running east and west at Brooksvale is here found in the lower sandstones. A similar freedom from dikes continues through all the rest of the lower division of the formation to its northern end in Massachusetts, thus strongly contrasting with the frequency of dikes in the lower sandstone of Cheshire, Wallingford, and East Haven.

The igneous and aqueous members of the series in the district of the Hanging Hills do not call for special remark as to their structural features, unless it may be on account of the very infrequent exposure here of critical strata and contacts. The subdivision of the block into many subordinate blocks or slabs is of more interest. First is the Quarry block, a narrow slice in which the main sheet is the only lava bed identified; the anterior and posterior seemed to be too much splintered to stand up in ridges. The main sheet, however, is of much interest from its dissection in large quarries, where its composite structure as a double lava flow, described on page 70, and its numerous minor faults are well shown. The minor faults are roughly parallel with the great marginal fault, already detailed in connection with the description of the Lamentation block; they are marked with a breccia of angular trap fragments, often slickensided, the interstices being filled with infiltrated sandstone débris, adjusted to the trap fragments. Secondary faulting of the breccia is sometimes seen, as if the movement had been long continued. A well-defined depression parallel to the general course of the system of faults, and more or less interrupted by local splinters or ledges of trap, separates the Quarry Ridge from the next stronger mass of the main sheet. Half a mile or more southwest of the Quarry Ridge the sandstone ledges have a strike about west-northwest.

The strong bluffs that inclose the narrow gap known as "Cat-hole" on the east are the southern ends of the main trap sheet in what may be called the Cat-hole block. A medial fault is indicated by the division of the trap into two ridges separated by a narrow, rocky valley. The western face of the ridge is more variable in strength of outcrop than is usual, from which it is inferred that there is some variation in the heave of the west-bounding fault. The anterior and posterior ridges are both identified in this block, and both are notched by the medial fault recognized in the main sheet. The anterior shales here exposed are peculiarly bright red. These, with the anterior trap sheet and the adjacent lower sandstones, have a northwest strike. For a mile or more southwest of the anterior ridge the course of the west-bounding fault may be traced by the oblique truncation of several sandstone ridges, and then by the course of a valley or depression between low hills. These minor features, as well as the dislocation of the several stronger ridges from the corresponding members of the next adjoining block, are well viewed from a hill that stands about half a mile northwest of the Meriden station of the Meriden, Waterbury and Cromwell Railroad.

The outcropping bluff of the main sheet in the next subordinate block of the Hanging Hills forms South or Notch Mountain, which is much higher, though of smaller area, than the Cat-hole ridges. The anterior trap sheet forms a well-defined ridge beneath the main bluff, with strike about northwest; the posterior sheet forms a long ridge, partly drift buried, extending northward to Berlin Junction. The gradual decrease in height of the main sheet, as it is cut back by the oblique bounding fault on the northwest, is an excellent illustration of this structural feature, so often repeated and so demonstrative when its meaning is once perceived. It may be the southwestward extension of this fault that causes the oblique gap in the intrusive ridge of Gaylord Mountain just north of Mount Sanford, where the trap is much jointed and broken; the less conspicuous oblique notches in Gaylord Mountain farther south near its termination may be associated with the smaller faults that bound Cat-hole block.

The West Peak main sheet rises to the highest summit of the trap ridges in Connecticut. It has several notable features in addition to its large size. Its back slope is surmounted near the base by a supplementary ridge, apparently the topographic expression of an upper lava bed, separated from the main bed by weaker members; the latter, however, do not outcrop in the wooded valley beneath the supplementary ridge. The change in the front of the bluff from facing south to facing west is well exhibited in this part of the main sheet, as the bluff may be traced continuously around the culminating point of West Peak. The anterior shales, anterior trap, and adjacent lower sandstones curve conformably with the main sheet, and it is believed that this curvature is essentially the result of a warp in the structure of these massive beds at this point, the southwestern corner of the great Springfield crescent. The homology of this curvature to the crescents of Pond and Totoket mountains has already been remarked. Quinnipiac Ridge broadens north of the Quinnipiac gorge, as if somewhat affected by the warp; but it disappears a little north of the West Peak corner, as if its determining strata had weakened or thinned out. It is curious to notice that in this locality the more resistant strata by which the ridge is determined stand near the top of the lower sandstones, while in the New Haven district they are below its middle. The possible bearing of this on glacial erosion is mentioned on page 179.

Part of the posterior lava bed assigned to the West Peak block extends north of Berlin Junction, being slightly dislocated from the posterior of South Mountain in advancing order. The gap between the two is occupied by the Sebethe River, in whose bed there is a fortunate exposure of the fault breccia, with strike N. 40° E. and hade about vertical, the breccia being about 2 feet thick and composed of sandstone and shale fragments, much crushed and slickensided. The locally reversed dip of the strata, as in fig. 23, *A*, indicates that the heave of the fault was, as usual, on the southeast side of the fracture. This breccia, so manifestly connected with the fault that defines the gap in the posterior ridge, is one of Percival's "clay dikes."[1] Among the posterior shales is a group of strong-colored red beds, not infrequently making good outcrops in this block, but less serviceable in determining fault lines than it was hoped they might be.

A close sympathy with the general system of fault lines is shown by several oblique gullies that trench the back slope of the West Peak main sheet, and lead, when prolonged, to recesses in its bold bluff. These are taken to indicate lines of weakness caused by slight dislocations in an otherwise nearly homogeneous mass, and hence are selected as guides for denuding forces in preference to the lines of slope ordinarily followed in the dissection of monoclinal ridges. In one case a narrow band of breccia, about a foot wide, and trending N. 50° E., essentially parallel to the course of the gullies, was observed in a ditch that led the streams from the mountain into the reservoir between West Peak and Notch Mountain. It is probably a southwestern continuation of some of these minor faults that produces the bands of breccia encountered in the isolated trap ridges north of Cheshire in mining copper. They show a small amount of "copper stain," along with cementing calcite and barite.

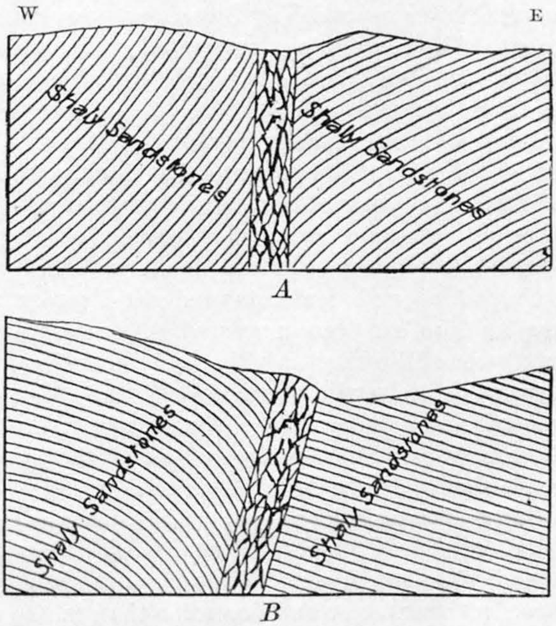


FIG. 23.—*A*, breccia and dragged strata caused by fault with heave on east; *B*, same with heave on west. Scale, about 15 feet to the inch.

The notch in which the reservoir stands between West Peak and Notch Mountain has a course that is in striking contrast to that of the gullies of the back slope and to the course of the master faults. At the south end it occupies the fault gap between the bluffs of the adjacent mountains; but while the fault line between the two blocks then runs east of the reservoir about N. 40° E., as is well proved by the relative attitude of the two parts of the dislocated main lava bed, the notch pursues a course leading about N. 10° E., and is undoubtedly

[1] Geology of Connecticut, 1842, p. 378.

eroded for the greater part of its length in the main sheet of West Peak on the west of the fault line. An explanation of this divergence will be offered in Part III.

The posterior ridge of West Peak is strongly subdivided, as if by one of the minor faults that notches the bluff of the main sheet and the anterior ridge, the greater displacement of the posterior probably being due chiefly to the change in strike. The fault crosses the anterior nearly square, and the posterior very obliquely, but it is also probable that the fault has increased in throw to the northeast. It must, however, be avowed that the posterior ridges from here northward for several miles beyond New Britain are very complicated in their arrangement. The correlations suggested on the map are the best that many trials can invent, but it is highly probable that the actual dislocations are much more complicated than those here described. A fault breccia of trap and sandstone is seen at the dam of a burnt mill in the village of Kensington. It is about 4 feet thick, with strike N. 40° E. and hade 15° from the vertical toward the northwest. Slickensides on the fault face indicate a nearly vertical movement on the fault plane. Certain subordinate outcrops of trap, in all probability portions of the posterior lava bed, occur hereabouts in slight disorder, for which no satisfactory explanation has been suggested. A trap ridge of considerable size rises just west of Newington village, departing most conspicuously from the system that elsewhere prevails. It may be merely suggested that it is most likely a dislocation of the posterior sheet, and that its occurrence here serves as a sort of forerunner of the greater uplift by which the main sheet rises in Cedar Mountain, a little farther east.

*Short Mountain block.*—Short Mountain is the main sheet of a narrow block or slab in which the usual middle members are represented. Its members follow those of West Peak in slightly advancing order, better seen in the anterior and posterior than in the main ridges. In the gap between the anterior ridges there was formerly a "paint mine." Old waste heaps lie about the shaft, exhibiting a quantity of breccia of dense and vesicular trap, cemented chiefly by barite. As in the Cheshire paint mines, the occurrence of these exceptional structures precisely on lines that are otherwise known to be followed by faults must be regarded as a crucial test of the theory of faulting. The bounding fault on the northwest, separating Short Mountain from the Ragged Mountain block, is one of the exceptional dislocations that produce a retreating order, distinctly shown in the anterior ridge, less distinctly in the main ridge, but recognizable again in the posterior ridge also, as it is interpreted on the map; hence this fault must have its heave on the northwest instead of on the southeast, as is the general rule. It is interesting to find that the permissible path of this fault leads us to a sandstone breccia striking northeast, one of Percival's "clay dikes,"[1] in the bed of a stream a mile south of New Britain Center,

[1] Op. cit., p. 320.

and at the eastern base of a great drumlin, and that the drag of the adjoining beds of shaly sandstones confirms the exceptional movement on the fault line, the dip being locally increased to 30° SE. on the northwest side of the fault, as indicated in fig. 23, *B* (p. 111), in contrast to fig. 23, *A*. Examples of reversed drag are rarely known in the valley; hence the occurrence of this one on the line of a fault otherwise indicated as heaved on the northwest is of considerable value. Farther northeast the fault has not been traced.

*Ragged Mountain block.*—From the broad notch north of Short Mountain to the oblique valley now flooded by Shuttle Meadow reservoir the main sheet stands out in strong bluffs, forming Ragged Mountain. The oblique truncation of the trap sheet on the southeast and northwest and its oblique bisection by the valley of Panther Swamp, now converted into a reservoir, testify clearly to the general course of the bounding and medial faults. The bluff turns from the western face of the mountain around its southern point, where its cliff face is very bold, and continues for a distance around to the southeast border, thus repeating the form that is so conspicuous in Chauncy Peak. Northward the bold western bluff gradually falls off in height and trails away, as the bounding fault cuts obliquely across it. The long northeastern trail of the trap sheet is cut squarely across by a broad valley a mile southwest of New Britain, a continuation of the broad and deep valley by which the main sheet of the Bradley and Farmington Mountain block is transected west of New Britain. This aberrant feature will be considered in Part III.

The fault lines indicated by the form of the main sheet are fully confirmed by the dislocations of the anterior ridge, and are permitted, if not demanded, by the arrangement of the posterior ridges, as they have been mapped after very careful search. As already stated, the anterior and posterior ridges exhibit the retreating order, with lapse of occurrence along the southeast border of the block, better than the main sheet. They are both dislocated in the line of the Panther Swamp bisection of the block. The duplication of the small ridges in the northeast part of the city of New Britain is ascribed to branches of this fracture, whose movement has here become of much greater value than in the main and anterior ridges; but it should not be concealed that the interpretation of these posterior traps on the map is somewhat doubtful, owing to the heavy drift that frequently covers their low ridges. It is probable that the actual structure is more complicated than the map would imply. The stratified members of the series are seldom of critical value in this block; but as far as their outcrops are traceable, they are terminated in a manner perfectly conformable to the scheme of faulting inferred from the trap ridges.

*Farmington block.*—A great mass of the main sheet extending from the bluffs of Bradley Mountain, north of Shuttle Meadow reservoir, 8 miles to the end of the long trailing ridge northeast of Farmington, must be included in a single block, in spite of the deep valley by which it is cut square across between New Britain and Plainville, this aberrant feature seeming to be entirely independent of faulting, and finding an explanation in Part III, along with its homologue, already mentioned, between Notch Mountain and West Peak. To the south of the transverse valley the main sheet forms Bradley Mountain; to the north the greater part of the ridge is known as Farmington Mountain, but its high northwestern knob forms Rattlesnake Mountain, separated from Farmington Mountain by a narrow ravine and belonging to the next block to the north. The various middle members of the series are fairly well shown, although the desired outcrops of the fossiliferous shales have not been found.

The bounding fault on the southeast produces a well-defined advancing offset and overlap in the trap ridges, the measure of overlap increasing greatly in passing from the anterior to the posterior ridge, and thus indicating a curvature of the fault line from a more transverse toward a more longitudinal course, as well as a probable increase in fault movement on the northeast. The curvature of the fault line will be recalled as of importance when tracing the course of other faults farther north. The bluff of the southern part of the main sheet in Bradley Mountain is somewhat irregular, indicating subordinate faults, one of which is perceptible in causing a small notch and advancing overlap in the anterior ridge at the Southington reservoir. Various oblique depressions and gulleys in the back of the main sheet suggest the occurrence of oblique dislocations of small throw.

The bounding fault on the northwest is proved to be of reversed movement by several peculiar features, best considered in connection with the next block. The long trail in which the main sheet of Farmington Mountain terminates northward is remarkably well developed. When followed northward the mountain ridge turns to the north-northeast back of Farmington, and gradually decreases in height until it disappears under the drift in low ground near the Hartford road, north of Burnt Hill. It thus repeats the habit of the main sheet in so many other blocks, especially in West Peak, South, Lamentation, and Higby mountains. In strong contrast with the long trail of Farmington Mountain, slowly descending to the north, is the bold bluff from which Bradley Mountain looks south over Shuttle Meadow reservoir. Here the outcropping base of the main sheet curves around from its usual exposure and forms a cliff to the southeast, gradually decreasing in height as it turns to the northeast. All this is perfectly conformable to the pattern deduced from the geometrical discussion of faults on page 89. It is only a repetition of what has been repeatedly seen at the southern end of the main sheet ridges in various other blocks, as Ragged, Short, and Notch mountains, Cat-hole and Chauncy peaks, and Tremont Mountain. When these consistent and orderly features of form are appreciated it will be easier to overlook the deep transverse valley in the Farmington block, and to regard Bradley and Farmington mountains as dissected parts of an unfaulted portion of the main sheet.

*East Talcott block.*—The eastern of the two trap ridges that form Talcott Mountain gives name to this block. For simplicity of statement, the account of Rattlesnake Mountain and its anterior will be postponed until the larger part of the block, farther north, is explained.

The west face of East Talcott Mountain stands with a retreating offset of about 2 miles from that of Farmington Mountain, and there is a lapse of about 5 miles between the south end of East Talcott Mountain and the north end of the corresponding bluff of Farmington Mountain. Similarly, the anterior of East Talcott Mountain retreats with an eastward offset of about 1½ miles from the anterior of Farmington Mountain, and there is a lapse of about 3 miles between the south end of the East Talcott anterior and the north end of the Farmington anterior (the anterior of Rattlesnake Mountain not being now considered). These retreating offsets and lapses of outcrop agree in demanding a reversed movement on the fault plane by which the two blocks are separated; and this is confirmed by the peculiar relations of the posterior members. The posterior ridge of Farmington block may be continued northward as far as Bloomfield through a series of interrupted ridges, which may be called the Farmington-Simsbury posterior; although often concealed, the course of the ridges runs steadily without any of the confusing offsets or overlaps that have been so conspicuous in the New Britain district. It is therefore impossible that the strong fault by which the anterior and main sheets are caused to lapse and offset over distances to be measured in miles, should intersect the posterior ridge; the fault line must curve to the west of the ridge, along the belt of low ground between the ridge and the base of East Talcott Mountain. For part of this distance the fault probably runs close to the outcrop of the Farmington-Simsbury posterior, for the ridge is discontinuous, as if the sheet were here and there splintered and shattered.

The curvature of the fault line to a more northward course is, however, precisely what might have been anticipated from the curved course already described for the Shuttle Meadow fault. The location of the fault between the long Farmington-Simsbury posterior and the main sheet of East Talcott Mountain is further confirmed by the very narrow space here allowed for the posterior shales, although the dip maintains its usual value of about 15° E. In the New Britain district the shales generally occupy a belt at least 2 miles wide, while here the width is reduced to about two-thirds of a mile; but this is precisely the necessary result of a fault parallel to the strike of the strata and with heave on the west side. The greater part of the posterior shales thus lapse hereabouts; and the posterior trap sheet of the East Talcott block is not to be seen. A curving fault, at first trending northeast and then north, with strong heave on the west, is thus fully established.

Inquiry is naturally awakened as to the further course of the curved fault. If it continues to turn to the left, it must again intersect the main and the anterior sheets, once more dislocating their ridges, with lapse and offset appropriate to its movement. Precisely such an arrangement is found at Weatogue Gap, between Bloomfield and Weatogue; but the dislocation is small, the lapse and offset of the main sheet not being over 2,000 and 600 feet, respectively, and before going much farther the fault must die out, for the anterior ridge is not broken.

It should be noted that, as the fault runs northwest, the offset that it causes is in advancing order westward; but as the heave is still on the west side, there is a lapse as before, although its measure is small.

A curious and unexpected consequence of the curved course of the fault is that Simsbury Mountain, next farther north, must lie in the same block with Farmington Mountain; and if the system of curvature be continued to the other faults, other northern and southern members of the main trap ridge north and south of Talcott Mountain would likewise fall in pairs within single blocks, but unhappily no other fault to the east of this one has been traced from the New Britain district past the latitude of Hartford.

Returning now to Rattlesnake Mountain, south of Farmington, it is found to be separated from the main trap sheet of Farmington Mountain by an oblique ravine, with trap bluffs on the west as well as on the east, in the path of the curved fault above described. The greater height of the Rattlesnake summit is probably due to the greater altitude in which erosion found the main sheet on the west side of the fault line; not that the movement of the fault directly increased the height of the mountain, but that it placed the trap at a greater height to be acted on by erosion.

The anterior of Rattlesnake Mountain is separated from that of Farmington Mountain a mile north of the Plainville-New Britain gap by a small retreating offset, with brief lapse of outcrop. The Rattlesnake anterior ridge thence continues north and northeastward until it almost abuts against the foot of the main trap bluff in the Farmington block back of Farmington village, causing a nearly complete lapse of the anterior shales. This must mean that the heave of the fault increases to the northeast. A little farther on the anterior sheet itself lapses, thus indicating a still greater increase of heave. But shortly afterward, passing obliquely northeastward, and hence somewhat down the dip of the monocline, the anterior reappears, this being the portion of the ridge that fronts the south end of East Talcott Mountain mentioned above.

The exceptional arrangement of the anterior and main trap ridges in this block gives a more severe test to the theory of faulting than it encounters anywhere else in the valley, and at the same time leads to results that are as unexpected as they are satisfactory. It is safe to say that so long as the trap sheets are regarded as intrusive sills the irregularities here met would be explained as the result of unsystematic intrusion, and the peculiar order of faulting would remain unsuspected. But when varied and consistent evidence proves the sheets to be extrusive flows, originally of great continuity for many miles north and south, their interruptions or lapses to-day call for some systematic explanation, such as is given above, consistent with that applied successfully elsewhere in the valley. No refuge is then needed in disorderly intrusion, for the features of this exceptional district fall into perfect accord with the regional system of faults of the entire formation. The explanation offered thus not only demands acceptance for itself, but permits, nay, compels, fuller belief in the interpretation given by the same general scheme of explanation elsewhere in the valley.

*West Talcott block.*—This block contains the higher of the two parallel ridges to which the name Talcott Mountain is ordinarily given. The mountain is normal on its western slope, the strong trap bluff descending to an elevated and narrow anterior valley, followed by a high anterior trap ridge; these special forms being apparently due to a decrease of thickness and a local increase of dip in the anterior shales and sandstones. But the similarity of the terminal bluffs at the north and south ends of the main ridge, the overlapping ends of the anterior ridge, and the abrupt descent of the main ridge eastward down a bluff-and-talus slope into the narrow valley that separates it from East Talcott Mountain, are three associated features not repeated elsewhere. Exceptional as they may be, they are the necessary results of another curved fault, the Talcott fault, parallel to the one just described, but with uplift on the east in the fashion generally prevailing in the region. The East Talcott block may now be described as a narrow, concavo-convex slab. The northeast course of the Talcott fault at the south end of the mountain gives offset and overlap corresponding to those so frequently seen farther south; the northwest course of the fault at the north end of the mountain changes the overlap from southward to northward, and reverses the offset to the retreating order of arrangement. Here the offset is about 3,200 feet, and the throw 1,125 feet. The gradual curvature of the fault into a course parallel to the strike of the monocline causes a parallel duplication of the main sheet not elsewhere to be found. The effect of this arrangement of the fault line on the east-facing cliff and talus by which West Talcott Mountain is cut off on its eastern slope, as well as on the moderate height of East Talcott Mountain, will be further considered in Part III. Several minor faults are implied by oblique notches and gullies in the trap sheet, as indicated on the map.

The advancing order of the ridges in the southern division of the intrusive sheets and the retreating order of the northern division make one of the most significant contrasts of the two divisions. The meaning of the contrast is evidently to be found in the direction of the fault lines. The advancing order is maintained by certain subordinate faults somewhat north of the middle of Talcott Mountain, but beyond this point the arrangement is in retreating order in the ridges of the intrusive sill, as well as in those of the extrusive sheets. It is surely most suggestive of a common control that these peculiarities of arrangement should agree in the two groups of ridges, one formed on an intrusive and the other on an extrusive sheet of trap.

It will be noted that only in the Lamentation block have the bounding faults all been traced across the Valley Lowland. Elsewhere the fault lines are well defined only as they cross the middle members of the formation. Although the northern intrusive sheet is undoubtedly faulted, the correlation of individual fault lines in the middle and lower members of the formation can seldom be safely accomplished in the northwestern part of the Valley Lowland. It can, however, hardly be doubted that the faults continue through the monotonous and generally covered lower sandstones and conglomerates, and that this part of the formation is broken into blocks probably of concavo-convex form about as completely as the rest. The space occupied by the lower sandstone west of Talcott Mountain should not be regarded as a single block, but rather as consisting of several curved blocks accordant with those containing East Talcott and West Talcott mountains. A faulted structure probably prevails also in the great area of sandstones, largely drift covered, north and east of Hartford. It is highly probable that if the strata there occurring could be as well traced as those of the middle division they would be found dislocated in some similarly complicated fashion, and the present simplicity of that part of the map would be changed to a complexity more accordant with that of the rest of the Valley Lowland.

*Simsbury block.*—North of Talcott Mountain the main trap range is nowhere so strongly dislocated as it is farther south, and perhaps for this reason its different parts have, as a rule, ill-defined local names. The names here used for its chief divisions are taken from the adjoining villages.

Simsbury Mountain, extending from Weatogue Gap to the notch of Farmington River at Tariffville, may be subdivided into three roughly equal parts by minor oblique faults, indicated by gullies, notches, and offsets in the main ridge and by slight interruptions in the anterior and posterior ridges. The middle portion of the main ridge is called Pinnacle Mountain. The breaks in the subordinate ridges can not, however, in all cases be well correlated with those in the main ridge. The best-proved continuity is along the pair of faults that separate the second and third subdivisions of the main sheet, 2 miles east of Simsbury village. Here two faults, about 500 feet apart, isolate a narrow splinter of the main sheet and little knobs of the anterior and posterior sheets; their combined throw is 160 feet. The oblique course of many gullies in the main sheet is highly characteristic hereabouts, and deserves attentive recognition on the map. A subanterior sandstone makes a distinct bench here and farther north, beneath the anterior trap ridge.

The termination of the main sheet in the Simsbury block in Round Hill at Farmington River Gap is again defined by a pair of faults, with combined throw of 170 feet or more, slicing off a knob of the anterior and a narrow slab of the main sheet, about 60 feet across, but apparently diverging southeastward, so as to cut off a wider slice of the posterior ridge. The view of these faults from the Bartlett tower on the north summit of the main sheet near Tariffville is very instructive; various details of surface expression, so faint as to elude all but the most minute survey, fall into accord when the scheme of oblique faulting is applied. The ridges end abruptly on the north, while those of the next block begin gradually; this expressive feature being frequently repeated from here northward.

The Simsbury block is, as already stated, faintly separated from the East Talcott block on the south, the offset of the main sheet measuring perhaps 600 or 700 feet, and the line of the fault being located by a well-defined oblique gap with advancing displacement; but, as has been stated, the fault disappears before reaching the anterior sheet, and the displacement here seems to be accomplished by a slight warp, indicated in a deflection in the course of the trap ridge. The bounding fault on the east side of the block and the subordinate faults within the block can not be traced east of the posterior ridge, although if the sandstones were laid bare of their drift cover it is highly probable that these faults could be followed along curved courses until they met some of the faults of the New Britain region.

Following the block to the northwest, one sees that it merges with the East Talcott block, but the somewhat broader block thus formed is well separated from the West Talcott block by the fault that produces the offset and overlap at the deep-worn entrance of Weatogue Gap. It seems probable that this fault may cross the northern intrusive sheet in the neighborhood of the Barndoor Hills, for here occurs the most distinct retreating offset of that range, as if to be correlated with the strong retreating offset between East Talcott and West Talcott mountains; but the fault between the two Barndoor Hills, as locally determined, is deflected to the west of south, and would not reach the Weatogue Gap unless strongly curved. The subordinate faults of Simsbury Mountain may be correlated with the notches in the intrusive range southwest of Granby village, and the north-bounding fault may be plausibly carried through the gap north-northwest of the village. It is, however, quite possible that if the sandstone floor were visible a somewhat different arrangement of fault lines would be discovered; for it must be remembered that the intrusive sheets may themselves have broken obliquely and intermittently through the sandstone, and that the arrangement of their ridges, here interpreted as the result of faulting, may be to a greater or less degree the result of original intrusion.

Although possible, this is not thought to be probable; for in the few places where outcrops permit observation, breccias have been found in the gaps and notches of the intrusive ridges, thus indicating that faults have been controlling factors in locating the gaps here as well as in the once undeniably continuous sheets of the middle lava flows.

*Granby block.*—The separation of this block from its neighbor on the north is not demanded by the small amount of dislocation between them, but it is convenient to give the two blocks separate names on account of the absence of the anterior sheet in the northern of the two. Like the main ridge of the Simsbury block, that of the Granby block is subdivided by minor faults, the first two being identifiable in their oblique extensions a little east of south across the posterior ridge, but becoming doubtful when projected in the other direction on account of the weakness of the anterior ridge near the termination of its lava sheet; while the third fault dislocates the anterior but not the posterior ridge. The north-bounding fault defines East Granby notch. The interest in this block centers in the peculiar dislocations of its northern end, ingeniously explained by Mr. Kümmel as the result of an exceptional northeast-southwest fault ($F_4$, fig. 24) through Vineyard Gap, which is supposed to cross three northwest-southeast faults, $F_1$, $F_2$, $F_3$. The northernmost of the latter is the bounding fault of the Granby block, with a throw of 200 feet or more, against which the subanterior sandstones, the anterior trap, the main trap, the posterior sandstones, and the posterior trap all end in abrupt bluffs of appropriate form, beginning again, after a slight retreating offset and overlap, in the next block. The adjacent north-

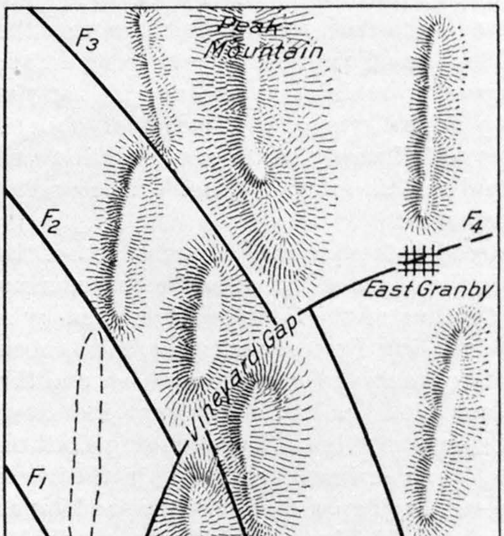

FIG. 24.—Abnormal fault in Granby block, causing Vineyard Gap.

west-southeast faults cut off two slabs from Hatchet Hill, the chief main ridge of the Granby block. The irregular dislocations of the main sheet, as well as the peculiar forms of bluff and trail in the main ridge of each slab, may be fully accounted for by the faults as here figured, but the behavior of the subordinate ridges causes some trouble. Two portions of the anterior are missing, as indicated by dotted lines in the figure. This may be partly the result of low erosion near so many fault lines, and partly the effect of burial under drift, for a large drumlin occurs on part of the area of the southern missing member. The absence of a posterior ridge in the village of East Granby may be accounted for as the result of weakness due to its intersection here by one of the subordinate faults of the Suffield block; and its continuity farther south, where it should be dislocated, may be explained by supposing the fault, which breaks both the anterior and the main ridges, to die out before reaching the posterior. The difficulty in this case does not lie in any inherent improbability in the suggested arrangement of fault lines, but only in the absence of outcrops, by which the suggestions might be proved. The scarcity of outcrops is particularly to be deplored, for here the explanation offered is of an exceptional character, not supported by repetition of its essential features elsewhere, and hence demanding much fuller local proof than is needed to confirm the existence of any one of the numerous faults farther south, that so plainly belong to a single system, where the evidence in one example supplements that in another. Manituck Hills, two large members of the intrusive range, probably belong in this block.

*Suffield block.*—The only members continuously recognizable in this block are the main and the posterior trap sheets. The southern part of the main ridge includes Peak Mountain, or Copper Hill. The anterior sheet weakens and disappears, except for two small ledges near the southern side of the block, at the old copper mine, known as Newgate Prison, and the outcrops of the posterior shales and sandstones are much interrupted. Apart from the peculiar dislocations already described under the Granby block, there are only two special features to be noted here. First, the broadening of the main ridge in Bald Mountain, near the middle of the block, apparently the result of two very oblique minor faults of 150 and 500 feet throw, the course of the faults being indicated by ravines, trending nearly north and south. The narrow block of trap between these fault ravines forms Bald Mountain. It has a bold bluff on the north, while the ridge to the east gradually trails away southward. Bluffs and talus slopes occur on both sides of the fault ravines. It is to these faults that we may ascribe the weakening of the posterior ridge in the village of East Granby, as above stated. Second, near the village of West Suffield the posterior ridge offsets and overlaps in such manner as to indicate three parallel faults, strongly oblique and close together, of moderate displacement. The effect of these faults on the main sheet may be perceptible in the low gap of Rising's notch, just north of the State boundary, but it is not certainly identified.

### MOVEMENT ON BEDDING PLANES.

It is not uncommon to find the bedding planes of shales more or less slickensided, with the striations about in the direction of the general dip of the monocline. Sometimes a very perfect mirror surface is thus produced. The sandstones are much less affected in this way. The trap sheets show practically no signs of such disturbance. The amount of movement thus indicated does not seem to be large, and it does not appear to have been attended by any other consequences than these subordinate results of frictional slipping. In this conclusion it is necessary to differ from that reached by Dana, who thought that the bedding-plane movements produced overthrusts of sufficient magnitude to pack the strata together in a mass of greater thickness than it originally possessed. He wrote: "The abraded surfaces of the beds, extensively exhibited in some regions, indicate that there was a vast amount of intestinal movement as well as ordinary faulting. The sandstone should therefore have acquired its greatest thickness, from piling on itself, on the side of the area in the direction of movement; that is, on the west side in the Connecticut Valley."[1]

### MARGINAL FAULTS.

The faults thus far described traverse the Triassic formation obliquely with varying throw. They are best defined where they cross the trap ridges, but in a few cases they have been plausibly traced through the sandstones to the border of the formation on the east and west, where they determine the present boundary between the Triassic lowland and the crystalline upland for a greater or less distance, as indicated by full lines in fig. 25. This is best seen in the two northeast stretches of the eastern boundary, from South Glastonbury to South Manchester, and from Vernon to Rockville, already mentioned in connection with the Lamentation block. On the western border the effect is much less marked, and is seen distinctly only for short distances on the sides of the Lamentation block, southwest of the two notches that limit Bethany Mountain, and at certain points farther north. Near Bethany Mountain the same two faults that deflect the eastern boundary for so great a distance cause only slight offsets and overlaps in the western boundary. The reason for this change is probably to be found chiefly in a diminution of the fault movement, but it may also depend on conditions discussed on page 130.

We have now to inquire whether other parts of the Triassic boundary may not be determined by faults. It should be recalled in this connection that the present boundary stands in an indefinite relation to the original boundary of the formation. Since deposition was interrupted

[1] Manual of Geology, 1896, p. 800.

by uplift, there has been ample time for the denudation of great bodies of crystallines, as well as of less resistant sandstones, as will appear more fully in Part III. The hypothesis of a widespread Triassic cover, stretching far across the crystalline areas, is not to be excluded because of any difficulty in disposing of its lost parts; for in Jurassic, Cretaceous, and Tertiary time good-sized mountain ranges might have been laid low. The objection to a very broad original area of the Trias lies first in the absence of direct evidence indicating the existence of such a cover, and more pointedly in the occurrence of coarse conglomerates along the present border, with pebbles that are frequently identifiable in the neighboring rocks of the crystalline uplands. It would appear from this that the Triassic shore line was not very distant from the present border. An original breadth 10 or 20 miles greater than the present breadth is the utmost that known facts demand.

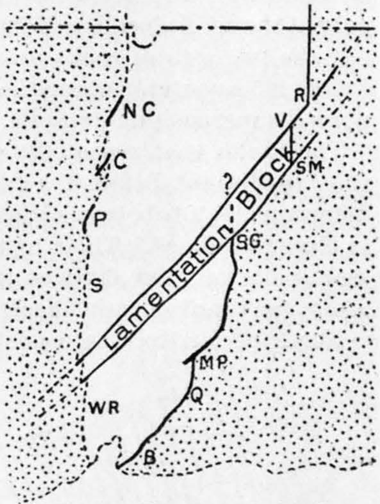


FIG. 25.—Plan of marginal faults.

The present inquiry is to determine whether the existing boundary of the formation marks the present limit of partial stripping of the formation from its foundation, or whether it marks the line of various faults which have brought the sandstones and crystallines together on either side of a steep-dipping fracture, as in fig. 26.

*Western boundary*.—The western boundary, as a whole, seems to be

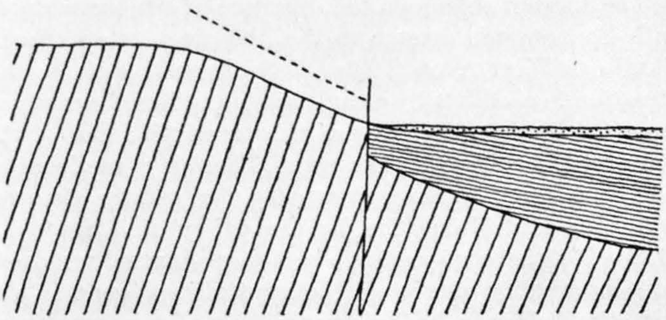

FIG. 26.—Western boundary, determined by faulting.

independent of faults. The long, straight stretch of the Southington district (S, fig. 25), where the basal sandstones and conglomerates appear to lie normally upon the crystalline slope of the Western Uplands, indicates merely a slow stripping of the weaker formation from the resistant floor on which it lies. This simple relation seems to prevail for about 20 miles, from Bristol to Bethany Mountain; then it is briefly interrupted by the offsets above mentioned, only to be reestablished parallel to West Rock ridge, and thence southward to the seashore, a distance of a little over 10 miles. It should be remembered, however, that in the total distance of 30 odd miles thus included the basal beds of the Trias are actually seen in contact with the crystallines only at the single locality of (Southington) Roaring Brook. Everywhere else the two sets of rocks are separated by the universal drift cover; but as far as outcrops near the boundary bear testimony, the relation of normal superposition prevails.

The faults that repeatedly dislocate the extrusive flows seem, as a rule, to die out before reaching the western border. The five well-determined faults between West Peak and Farmington are not traceable to the crystallines. Their extinction recalls the disappearance of the strong Farmington fault at its north end in Weatogue Gap, where, after breaking the main sheet, it fails to dislocate the anterior.

North of Bristol there are several localities where the boundary

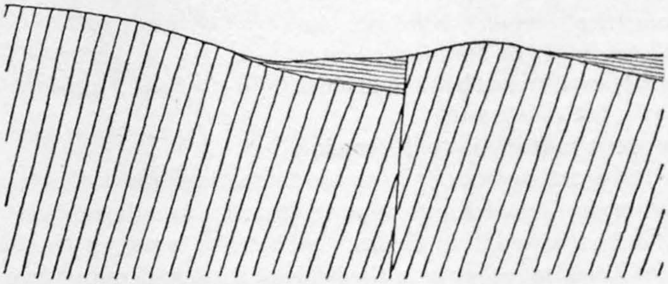

FIG. 27.—Outlying area of sandstone, western boundary.

appears to coincide with fault lines rather than with the normal contact surface. The distinct change in the direction of the boundary from the north-south Southington stretch to the oblique north-northeast trend from Polkville to West Avon (P, fig. 25) is suggestive of the truncation of the formation hereabouts by a fault with uplift on the west; and this is countenanced by the parallelism of this trend with that of the faults in the Farmington ridges farther east. There is some local evidence of faulting, but there is no opportunity for certain determination. Between Canton and Unionville, in the valley of (Canton) Roaring Brook (C, fig. 25), there appears to be a down-faulted wedge of sandstones, unconformably overlying the crystallines on the west, and apparently cut off from a hill of crystallines on the east by a normal fault, as in fig. 27; but outcrops are so few and far between that this can be only a tentative explanation. The long intrusive ridge east of North Canton is separated from the crystallines by a deep, narrow ravine, with walls of trap and schist, and it is believed that this is determined by a fault of moderate downthrow on the east (NC, fig. 25). Farther north, toward Granby, there are several short stretches where the crystalline border does not stand parallel to the strike of the sandstones and conglomerates, and where faults are therefore inferred; but in these cases the faults must have upthrow on the west. From Granby northward there is no indication of faulting along the boundary.

Taken altogether, the faults on the western border do not seem to define more than 12 or 15 miles of distance, or about a quarter of its length. They are discontinuous, of small throw, with uplift indifferently on the east and west. Although somewhat irregular in the Canton-Granby district, these western marginal faults may be plausibly associated with the great system of curved oblique faults that prevail within the Triassic area, as already described.

*Eastern boundary.*—In striking contrast with the western boundary, the eastern border is determined by strong faults through its entire length. These faults have two dominant trends, north-south and (about) northeast-southwest. South of Middletown the two trends generally merge on curved courses; north of that place they are separated by rather well-defined angles or "corners" of the Eastern Upland.

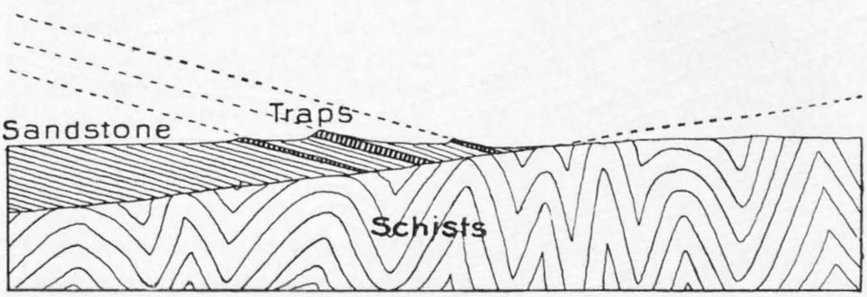


FIG. 28.—Eastern boundary, as overlap.

The chief reason for believing in the existence of these eastern marginal faults is found in the abrupt termination of a heavy monocline of sandstones, with associated conglomerates, shales, and lava beds, at the eastern border. Walking southeastward across the strike of the Saltonstall or Totoket crescents at their middle, or east-southeastward from the anterior ridge of Higby Mountain, the observer passes over a continuous series of strata, which may begin with the sandstones and conglomerates under the anterior lava bed, and pass regularly through the normal succession into the upper sandstones above the posterior lava bed, the thickness involved varying from 6,000 to 7,000 feet at least. Suddenly and with comparatively trifling disorder near the margin the sandstones are replaced by the crystallines, and the lowland changes to the upland. It may at first sight seem possible to explain this by supposing the crystalline floor to descend very gently under the Triassic strata, whose upper beds overlap by increasing distance to the eastward, as shown in fig. 28; but this possibility is excluded when it is perceived that all horizons from subanterior to superposterior strata abut on the boundary in one place or another, in consequence of the warping in the crescentic areas. If the Trias lay unfaulted on the crystalline floor the observed warping of the strata would require a strong curvature of the boundary, and the crystallines would advance toward the Triassic area where the strata are warped upward. No such systematic curvature is to be found. The boundary line runs straight across the strongly warped arch where the crescents of Saltonstall and Totoket mountains are separated, and its general trend is well preserved southwestward from the end of the main sheet in Branford across the lower sandstones to New Haven Bay, and again northward from Lake Quonnipaug across the posterior, main, and anterior lava beds, and the associated sedimentaries on the north end of Totoket. It is, therefore, not conceivable that the surface of separation between the Trias and the crystallines hereabouts should stand

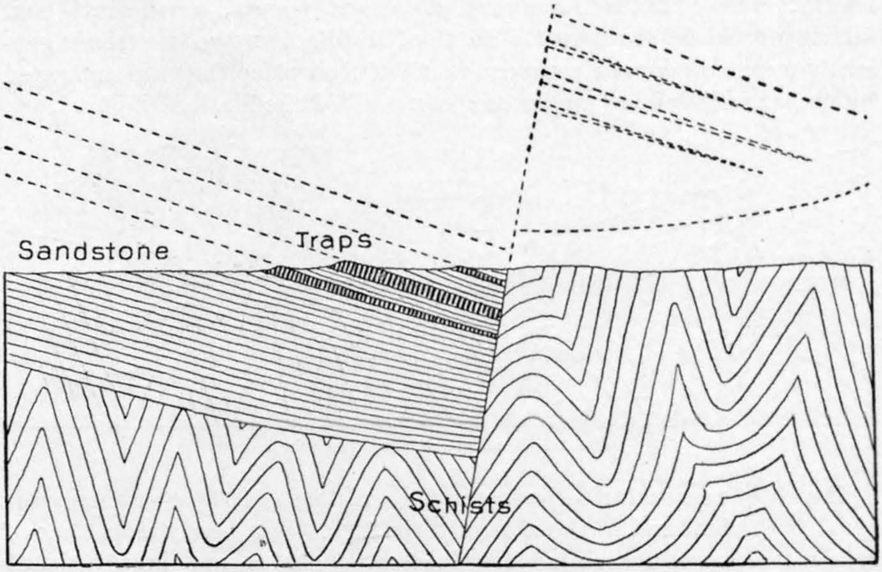


FIG. 29.—Eastern boundary, as determined by a fault.

otherwise than nearly vertical, and in such a position it must be interpreted as the plane of a great fault. It is perhaps imaginable that such a fault might have been developed during the period of deposition, its downthrow being greater in the middle of the crescents than at their ends; but while geometrically possible, this is geologically improbable in a high degree. The surprising accordance of the various members of the Triassic series involved within the crescents with those found farther north is not compatible with deposition in various local basins, but proclaims broad and even conditions of accumulation. Not only are the lava beds discovered hereabouts with much regularity, but the anterior and posterior shales, with their fossils, are found hardly a mile from the crystalline border, the anterior shales under the northern curve of Totoket, and the posterior shales near the southwestern end of Pond Mountain. This can mean only that we have here to deal with a great series of even deposits, afterwards warped and faulted. The eastern boundary in this locality must be a fault line, and in the middle of the crescents the throw must be at least 6,000 or 7,000 feet, as in fig. 29.

Still further evidence of the marginal fault is found in the trough of low ground that is traceable almost continuously along the boundary line, this feature being better defined here than farther north. At no place do any members of the Totoket block come in contact with the crystallines; there is always a drift-covered depression between the two.

Considering the magnitude of the fault, there is singularly little disturbance in the adjacent rocks, the only important instance being in the Saltonstall Mountain crescent. Here Percival's "second posterior" trap ridge,[1] near the village of Branford (B, fig. 25), is interpreted by me as a dragged portion of the normal posterior. The reasons for this interpretation are as follows:

The extrusive nature of the trap has already been shown (p. 72). It is in a remarkable degree improbable that here only in the whole valley a second posterior lava bed was poured out and preserved. The upper horizons are abundantly represented farther north without a sign of trap. The ridge has not the northwest outcrop face and southeast back slope elsewhere so generally found; it faces southeast and slopes northwest. Sandstones and conglomerates are seen beneath its face with accordant northwest dip of moderate measure. Breccias of trap and sandstone, as in fig. 30, are disclosed in a transverse road cut, showing that the mass is disturbed by minor faults. Putting all this together, the most probable explanation for the ridge makes it a large fragment of the posterior, dragged upward by the movement of the great marginal fault which passes close along its southeastern base, and at the same time tilted from the usual southeast dip of this district to a northwest dip. A diagram of the inferred relation is given in fig. 31.[2]

FIG. 30.—Breccia in posterior ridge, near Branford.

There is a small part of this peculiar ridge that has given rise to diverse interpretations. This is a ledge of firm trap just south of the cross road; it seems to dip steep to the southeast, and it is (apparently) overlain conformably by a conglomerate. This view was taken by Hovey.[3]

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[1] Geology of Connecticut, 1842, p. 325.

[2] The repetition of the posterior trap as here explained depends chiefly on faulting, and this seems preferable to repetition by warping, as figured in the Seventh Annual Report of the U. S. Geological Survey, p. 480; for, as Hovey has shown, the necessary warping is not permitted by the observed dips.

[3] Am. Jour. Sci., 3d series, Vol. XXXVIII, 1889, p. 372.

The opinion here preferred is that this is only a local overturning of a small block of trap with its under conglomerate. The trap being shown to be extrusive by evidence of larger areas near by, the dense trap here seen must be referred to the bottom of the sheet, where such a texture is normally found.

From New Haven Bay to Quonnipaug Lake (Q, fig. 25), the marginal fault may be treated as belonging with the series of oblique faults by which the Trias is broken into blocks. As such, it stands roughly parallel to the fault on the other side of Totoket block, which has been described as running from about East Rock through the valley between Totoket and Paug mountains. Accepting this suggestion, we are led to look for the continuation of the marginal fault northeast from Lake Quonnipaug through the crystallines; but a careful search through the wooded hills and valleys failed to discover decisive evidence of it. There is truly a certain alignment of valleys through Rockland, past Candlewood Hill, and on to the Connecticut, a little above Higganum, all trending fairly well in prolongation of the marginal fault; but the

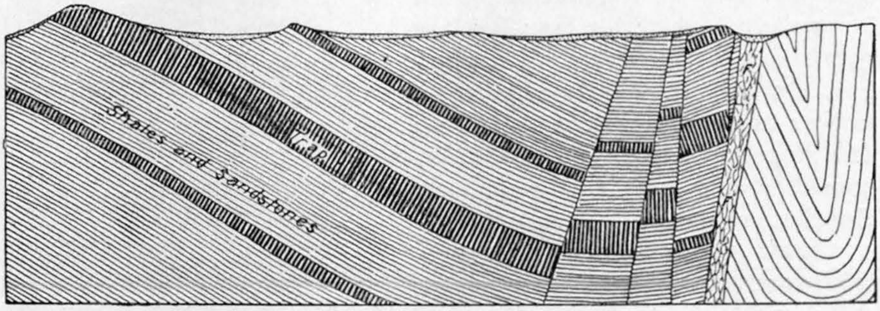


FIG. 31.—Diagram section of marginal fault, Saltonstall crescent.

settlement of the matter must be left to future work on the crystallines themselves. If it should appear that the oblique fault is thus prolonged, the short northward marginal stretch from Lake Quonnipaug to the corner of the upland near Mount Pisgah should be set apart as belonging with the north-south group of faults.

The long curve of the eastern border from Mount Pisgah to South Glastonbury is not marked by a persistent depression; indeed, the outcrops of sandstone are often found upon slopes that lead directly to the upland hilltops. Yet a fault is confidently believed to run along this line, for, as in the previous examples, the sandstone monocline is abruptly cut off at the boundary. Moreover, there are in this stretch of the border many local irregularities in the dip of the sandstones and conglomerates such as have been found nowhere else in the valley lowland. There are also certain characteristic features in the crystalline rocks near the border line, especially north of the Connecticut, that are not noticed in the body of the upland. The greater development of chlorite along the border is a marked occurrence. Microscopic examination shows that the crystals of the different minerals in the schists and gneisses are commonly broken; in many places near the border the simple gneissic or schistose structure has been lost, and the rock appears to be twisted or broken. This is the more noticeable because the belts of the various crystalline rocks trend northeast away from the marginal faults hereabouts, and thus the change in the character of various belts may be observed as they approach the marginal fault. The bordering parts of the crystallines is sometimes a loose breccia, as in the quarry for road metal at South Glastonbury; sometimes it is recemented into an exceedingly tough rock by quartz or chlorite, or by barite. Although to-day a firm rock, none of its fragments are found in the neighboring Triassic conglomerates; hence it is thought to be of later origin than the conglomerates. Taken altogether, these marginal features of the crystallines are strongly confirmatory of the existence of a post-Triassic marginal fault.[1] Where the Connecticut flows into its gorge in the Eastern Upland below Middletown, the uplift of the fault is estimated at more than 7,000 feet. The measure is somewhat uncertain; first, because of the uncertainty as to the thickness of the Triassic series here, and second, because of our ignorance as to the height at which the Triassic floor once stood above the present crystalline upland on the east of the fault.

At the Mount Pisgah corner the curved marginal fault appears to be the continuation of the fault between the Totoket and Paug blocks, described on page 96. Likewise, at the South Glastonbury corner the marginal fault may continue its northward course across the long oblique fault that separates the Higby and Lamentation blocks (p. 103), thus leading it directly into the great drift-covered Triassic area of Glastonbury and East Hartford, and thence indefinitely beyond. There is, unhappily, no means of deciding this possibility, but we believe the witnesses in the case are silent because they are all gagged by drift, and not at all because they have nothing to say. If, as is probable, the fault continues into the drift-covered area, there are some interesting consequences of its displacement that deserve brief consideration.

*Problem of the South Manchester corner.*—Let it be assumed that the north-south marginal fault runs indefinitely northward, maintaining the throw that it has at the South Glastonbury corner (S G, fig. 32). Inasmuch as the north-south and the oblique faults both appear to have steep fracture planes, and as the movement on these planes may be, with small probable error, taken as parallel to their line of intersection, neither fault is represented as displacing the other in the following diagram. While the marginal north-south fault follows the Triassic border south of the corner, its throw, $t$, is equal to the sum of three quantities: the depth, $x'$, of the Triassic floor beneath the lowland surface west of the fault line at B, plus the height, $h$, of the upland above the lowland, plus the unknown height, $y$, of the lost Triassic

[1] See L. S. Griswold, Bull. Geol. Soc. America, Vol. V, 1893, p. 524.

floor above the upland next east of the fault at D; or $t = x' + h + y$. A reasonable value for $x'$ is 4,000 or 5,000 feet.

The oblique fault has its downthrow on the northwest, with a movement of $f$ feet. All horizons in the Lamentation block must stand $f$ feet lower than they stand in the Higby block. The Triassic floor, for example, must lie $f$ feet deeper at A than at B; similarly, the floor must lie $f$ feet lower at C than at D. At C the lowland is occupied by the Trias, of depth $x'''$, while at D the upland is occupied by the crystallines, above which the Trias once stood at the height $y$. It follows from this that $f = x''' + h + y$, or $x''' = f - h - y$. The general difference, $h$, between lowland and upland hereabouts is roughly 400 feet. The throw of the oblique fault is probably 2,000 or 2,500 feet. Adopting the smaller value, we have $x''' = 2{,}000 - 400 - y$. Hence, even if $y$ be zero, the depth of the sandstones at C is at least 1,600 feet. The crystalline floor would here be found in bored wells of no great depth. Further, if this be the true state of the case, the sandstones about C can not belong to the upper division of the formation, like those about A, but must be the *lower sandstones*, which are elsewhere found only west of the trap ridges. It is unfortunate that for the present this conclusion is reached only by geometrical argument from a postulate as to the probable path of the north-south marginal fault; but the conclusion is so surprising and so peculiar that it seems to deserve statement in order to serve as a test of the postulate on which it rests, when opportunity arrives.[1]

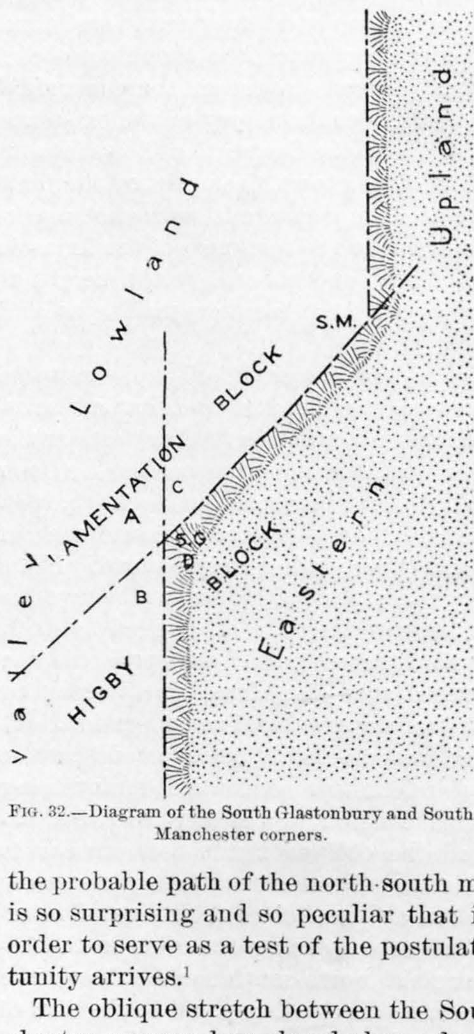


FIG. 32.—Diagram of the South Glastonbury and South Manchester corners.

The oblique stretch between the South Glastonbury and South Manchester corners has already been described (p. 103). Next comes the north-south border between the South Manchester and Vernon corners. Here, in addition to the general evidence of faulting from the termination of the Triassic monocline—evidence of comparatively small value

[1] This argument was first worked out by L. S. Griswold, Bull. Geol. Soc. America, Vol. V, 1893, p. 528.

because the outcrops of sandstone on the lowland are few and far between—there was by good fortune an opportunity of finding the actual contact fault plane by a small amount of digging near a paper mill at Highland Park. The contact surface of the Trias and crystallines was found to have a dip of 55° to the west, with strike north and south. The conglomerate and sandstone layers dipped 45° W., as if showing a strong drag from the usual eastward dip of the Triassic monocline. Near the contact the crystallines were much shattered and decomposed. The field relations left no doubt that the contact marked the plane of the marginal fault.

The throw of this fault is not directly measurable, for no values can be given to the depth of the Triassic floor under the lowland west of the fault or of the height of the floor above the upland east of the fault. By means of a series of not unreasonable postulates and geometrical constructions, similar to those employed about the South Glastonbury corner, some interesting results may be obtained; but so long as the postulates have no sufficient proof in observed facts, it does not seem necessary to rehearse them here.

### RELATIVE AGE OF TRAP SHEETS AND FAULTS.

In the early period of more speculative study of the Triassic formation, when all the trap sheets were commonly regarded as intrusions, it was supposed that the date of their intrusion was later than that of the tilting of the monocline. The faults, whose existence was at that time merely based on the assumption that so great a monocline could not be an unrepeated series of strata, were associated in date with the time of tilting; and thus the intrusion of the trap sheets was held to be later than the tilting and faulting.

When the extrusive origin of certain of the trap sheets was recognized, those sheets were seen to be of much earlier date than the time of faulting; but doubt remained as to the date of the intrusions. The prevalent opinion was still distinctly in favor of assigning them a date later than the tilting, and hence presumably later than the faulting. This was explicitly argued for the Palisade sheet in New Jersey by Cook and others, as well as for the sills of East and West rocks near New Haven, by Dana. Hovey stated in his earlier paper that the "true dikes do not appear to have been faulted, and therefore must have been formed since the tilting of the strata"[1]; he afterwards found that one of the largest of the Fairhaven dikes is slickensided, and recognized that "some differential movements in the region have acted on the trap as well as on the sandstone."[2] There is, indeed, much to be said in favor of the view that the sills and dikes, as well as the flows, are of earlier origin than the disturbance by which the faulted monocline was produced.

There is, in the first place, no inherent probability that the sills must

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[1] Am. Jour. Sci., 3d series, Vol. XXXVIII, 1889, p. 382.

[2] Am. Jour. Sci., 4th series, Vol. III, 1897, p. 289.

have been intruded after, rather than before, the tilting of the strata that include them. Sills of considerable extent and of nearly horizontal attitude in nearly horizontal rocks are known elsewhere, as in the famous "whin sill" of Yorkshire, England, or in a well-defined example on the walls of Purgatory River Canyon, Colorado.[1] It might, indeed, be argued that as the intrusive and extrusive sheets are all so much alike in composition, the burden of proof rests on those who would maintain a considerable diversity of age for them. But all this leads to no definite conclusion. The only means yet invented of dating the intrusions are by the relative arrangement of the several ridges formed on the sills and on the flows, and still more positively by the breccias found in the notches of the intrusive ridges where they are crossed by the oblique fault lines of the lowland system.

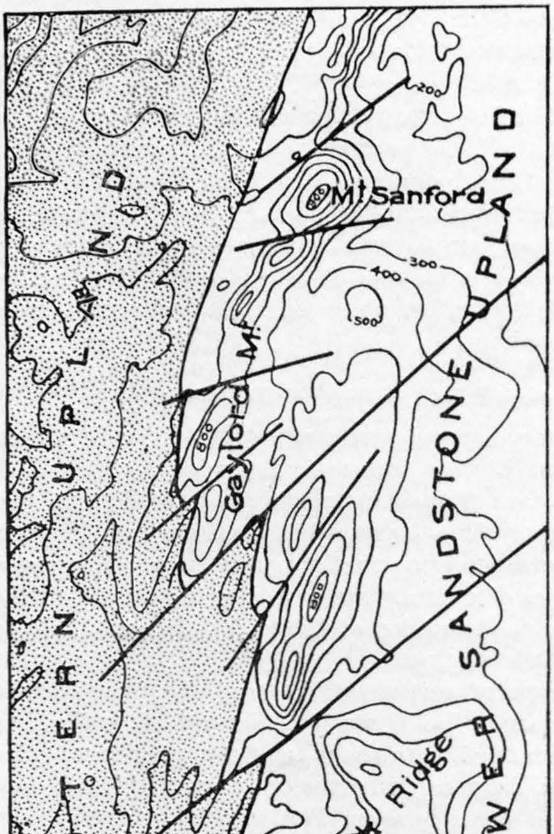


FIG. 33.—Displacements of trap ridges near southwest end of Lamentation block.

The general arrangement of the ridges on sills and flows is strikingly similar. South of Hartford both are prevailingly placed in advancing order; north of that city, in retreating order. The faults bounding the Lamentation block in particular have been shown to dominate the arrangement of the western as well as that of the eastern ridges. Bethany Mountain, the intrusive member of the Lamentation block, has every appearance of being a faulted member of the sill, elsewhere continued in the West Rock ridge on the south and in Gaylord Mountain ridge on the north. The several oblique notches hereabouts (fig. 33), with their advancing offsets and overlaps,

[1] Russell, Jour. Geol., Vol. IV, 1896, p. 178.

accord perfectly with the effects of the northeast-southwest system of faults, as determined elsewhere. The crystallines (hydromica-schists) enter the angle west of Bethany Mountain and south of the terminal bluff of the Gaylord range, and are obliquely cut off across the trend of their foliation by a narrow valley, located on the path of the presumed fault line.

East Rock is separated from Indian Head by what may be called a double-notched valley—that is, the gap between the two rocks is not a simple V-shaped notch, but, as described by Dana,[1] has a low ridge of trap in its middle, so that its shape somewhat resembles a W. The direction of the little ravines on either side of the medial ridge is northeast.

The northern division of the western trap ridges exhibits the retreating order of the northern extrusive ridges, as if both arrangements were controlled by a common cause. Local study of the Barndoor Gap convinces the observer of its dependence on faulting, although he may not be able to lead others to his conclusion, as the evidence is general rather than specific.

It might be contended that an intrusive sill guided by some easily split stratum would follow the lead of that stratum even after it had been faulted, and thus the sill would come to have an apparently dislocated arrangement, even though the intrusion was later than the faulting. While this may be possible, it does not seem to be very probable. It is more reasonable to suppose that an originally continuous sill, roughly accordant with the bedding, was dislocated along with the flows and the rest of the formation, and that the present advancing and retreating offsets and overlaps are thus to be explained in the western as well as in the eastern ridges.

This conclusion is well supported by the bands of breccia or thoroughly jointed structure that are occasionally found in the notches of the western ridges, on the line of faults that come obliquely across the lowland from the eastern ridges. The trap is much jointed in the Bethany notches. Trap breccia is distinctly seen in the stream that follows the notch just north of Mount Sanford, and here the sandstone on the northwest may be seen abutting against the dense trap on the southeast of a vertical plane of separation which strikes in accord with the local fault system. A few signs of brecciation have been noted in some of the gaps of the Barndoor Range, where, as above stated, the arrangement of the adjacent ridges has every appearance of dislocation by faulting. In the fourth notch west of Barndoor Gap the trap is finely splintered, as if near a fault line. Belts of breccia and slickensided joints are seen in a trap ridge farther south, where cut by the Central New England and Western Railroad. Their bearing varies from northeast to east-northeast. Although not accordant with the system of curving faults shown in the extrusive ridges,

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[1] Am. Jour. Sci., 3d Series, Vol. XLII, 1891, p. 97.

these breccias and slickensides clearly prove that some movement has taken place since the sill was intruded. The next small trap ridge on the south is displaced from the larger one in which the railroad cut is made precisely as if dislocated by a northeast fault of moderate throw. The same is true of a second small ridge a quarter of a mile farther south. The indications of faulting in the dikes north of Mount Carmel have been stated on page 104.

When all these significant facts are considered along with the peculiar sympathy in the order of arrangement between the eastern and western ridges, it seems conclusive that the dikes and sills, as well as the sheets, were present before the formation was disturbed, and that all suffered tilting and faulting together. Looking at the formation as a whole, the sills as well as the flows may be called contemporaneous, because they were intruded before the beginning of the deformation that put a stop to deposition. It need not be argued from this that the sills of East and West rocks ever formed a continuous sheet, or that the various members of the Barndoor Range ever formed a single sill close to the bedding planes of the sandstones, but it may be argued that before the time of dislocation the sills were much more nearly continuous sheets than they are now.

### SUFFICIENCY OF THE EVIDENCE OF FAULTING.

In Part I the system of faulting now described was provisionally accepted, and the whole discussion under which the sheets of the eastern ridges were shown to be fragments of only three extrusive flows was founded on this acceptance. At the beginning of this chapter it was stated that the theory of faulting would be tested by accepting its results and employing its language, while reviewing the distribution of the various members of the formation. This has now been done, and the value of the theory, particularly with regard to the oblique faults within the Triassic area, must be estimated. Its merits and deficiencies may be briefly summarized.

The theory of faulting leads through logical geometrical deduction to a variety of peculiar and highly specialized consequences. When these are confronted with the facts of observation the two groups are found to agree in a remarkable manner. It is well known that the verity of a theory is tested by agreement of this sort; and further, that the degree of verity, if such an expression may be used, is measured not only by the accuracy of agreement, but by the complication and specialization of the factors in the groups of deduced consequences and observed facts, respectively. In the problem under discussion it may be fairly claimed that the complication and specialization of both groups of factors are so great that their agreement by chance is out of the question. The agreement that they show can mean only that the theory from which the complicated consequences are deduced is a correct one.

The theory of faulting reduces confusion to order. This must be sufficiently clear from the special accounts given above of the various trap ridges. The possibility of reducing the labyrinth of trap ridges in the district bounded by Hartford, Middletown, Meriden, and New Britain to a reasonable geometrical system is alone a strong argument in favor of the scheme by which the reduction is accomplished. I can recall very distinctly the meaningless confusion of the trap ridges between Meriden and Westfield when first seen, over ten years ago, before any suspicion had arisen that the anterior, main, and posterior ridges, as described by Percival, were perhaps dislocated repetitions of a single series of lava beds, and that their terminations might be controlled by a system of oblique faults. The turbidity of the view, in regard both to structure and to topography, as seen on that day's walk, was clarified only as the systematic peculiarities of arrangement were perceived and a scheme to account for them was invented.

The theory by which confusion is reduced to order has the great merit of calling for only one essential element in itself. The repeated operation of a single simple process is all that is needed. With every repetition the signs by which it is known are more easily and certainly recognized. In this respect the theory of faulting resembles the theory of extrusion for the trap sheets of the eastern ridges.

The theory of faulting reduces complexity to simplicity. This alone may perhaps not be regarded as competent testimony in its favor, yet it has a certain value which deserves mention. It is not to be denied that the processes of nature could, in their marvelous variety of action, produce successive intrusions and eruptions as great in number and as peculiar in distribution as the lava sills and flows that may be counted in the separate trap ridges of the Valley Lowland; yet no one can avoid leaning in belief toward the simpler process of origin, on finding that it is permitted by the theory of faulting. When, further, it is found that the assumption of many independent intrusions and eruptions in an unfaulted monocline involves the accidental origin of repeated systematic arrangements of outcrop, while the theory of two sills and three flows in a faulted monocline gives immediate explanation for these peculiar arrangements, choice is soon made between the two.

The theory of faulting correlates many facts, previously noted, but not seen to have any bearing upon one another. Most notable among these is the arrangement of the ridges in what Percival empirically called "advancing order" and "retreating order," now seen to be a natural consequence of faulting and erosion. The form of the ridge endings is most peculiar. South of Hartford the south end of a ridge usually ends in a bluff, while the north end falls off in a long, slowly descending trail. North of Hartford this arrangement is reversed. This is, like the advancing and retreating order of dislocation, a natural consequence of faulting and erosion, as will be more fully stated in Part III. The systematically truncated ridges of conglomerate or sandstone associated with the truncated trap ridges are inexplicable on any other ground than faulting. Finally, there are the many points of detail, such as the oblique gullies on the back of the trap ridges, the variations of dip in the sandstones along the supposed fault lines, and the occasional exposures of "clay dikes," or fault breccias. A theory that gives rational correlation to so many facts must command confidence.

The theory of faulting has the further recommendation of explaining much more than it was intended to explain, of giving rational meaning to many facts not known when it was first conceived. There was at first no suspicion that its application would be so general, that it would so effectually account for all the ridges from Branford to Granby as the outcropping edges of only three lava beds. Indeed, for a number of years the ridges of the Cedar Mountain block were thought to call for one or more supplementary trap sheets; their final reduction to orderly members of the general series was a most gratifying warrant of the theory, lacking only the addition of the anterior shales and trap west of Cedar Mountain—carefully searched for, but not found—to make it perfect. Likewise, there was at first no thought that the fault lines by which the Lamentation block is locally bounded between Meriden and Westfield would account for various breccias and disturbed dips, and for the Bethany notches in the intrusive ridge, all afterwards found to lie on the extension of the fault lines. When these aberrant phenomena were found to fall in line with the ridge endings of the extrusive flows, assurance was made doubly sure.

The theory of faulting has even another and a final recommendation: it supplies, to a limited degree, the power of prediction. A fault once well defined by several ridge endings serves to foretell the endings of subordinate ridges before they are seen. This test has been frequently applied in the field, and with remarkable success. Critical points are thus quickly searched for, located, and examined. If the drift did not so generally limit the opportunity for observation, predictions might be made and verified in much greater detail. The theory of faulting, far from asking for support from its advocates, becomes their guide.

The deficiencies of the theory lie rather in the difficulties of application than in any vagueness of its deductions. Close and continuous observation of outcrops being impossible, the faults as drawn upon the map are probably too few and too simple. Many small faults are lost to sight. The larger faults, represented on the map by single lines, are probably compound, comminuted fractures. The branching faults that seem well proved in a few places suggest the occurrence of similar complications on a smaller scale elsewhere. The large district northeast of Hartford is doubtless faulted thoroughly, but it must remain a blank on the map. In the New Britain-Meriden district, and in a smaller area about East Berlin, doubt must still attend the interpretation recorded on the map; it may claim to be reasonable, and in most cases probable, but it does not reach the grade of certainty. Over the formation as a whole, if the drift were stripped off and the rocks laid bare, the faulting would probably be found much more extensive and intricate than it is represented on the map; but the main faults by which the larger blocks are separated are thought to be correctly located.

It may seem to the reader that undue attention has been given to the demonstration of geological structures so commonplace as faults; but it must be remembered, after all that has been said about them, that they are still imaginary, as far as direct observation goes. It is true that at a few points trifling exposures of breccia or slickensided surfaces are found; but these alone would by no means suffice to prove the existence of an extensive system of fractures, many miles in length and thousands of feet in depth. They are not known by observation; they are entirely the creation of inference and argument. They can therefore be assured only when the inference is legitimate and the argument logical. The erection of these creatures of the imagination to equal rank with observed facts is not a light undertaking.

THE ANTERIOR AND POSTERIOR FOSSILIFEROUS BLACK SHALES.

The occurrence of black shales containing fossil fish and impressions of plants at various points in the Valley Lowland has long been known. While heretofore of value in determining the geological age of the formation as a whole, their value in this report lies chiefly in the confirmation that they give to the theory of faulting. They are therefore now described in supplement to the preceding paragraphs regarding the value of the theory of faulting in leading to correlations and predictions.

When the dislocation of the various blocks into which the Triassic formation is divided was perceived, it became plain that certain of the beds of black shale might prove to belong to a single horizon. If so, their correlation might well be claimed as testimony in favor of a theory that was framed without a thought of its bearing on these beds. Furthermore, it was soon perceived that if the beds of shale were thus correlated, and if, as in the case of lava beds, various scattered localities of black shale were shown to belong to only a few horizons, it would then be possible to trace the approximate positions of the shale horizons in the various faulted blocks, and thus perhaps find new localities at designated points, or at least within designated limits. If in a formation so generally unfossiliferous as the Connecticut Trias such a search proved successful, it might be claimed as a thorough test of the prophetic powers of the theory of faulting, after which no further tests need be asked for. It is therefore chiefly with regard to their correlation and prediction that the fossiliferous shales have been studied. At the same time, a fine collection of plants and fishes has been brought together and deposited in the National Museum at Washington, awaiting special study.

On reviewing the recorded localities at which the black shales were known to have yielded fossils, it was found that with a few exceptions all of them fell either into the anterior shales 50 or 100 feet above the anterior trap sheet, or into the posterior shales about a third of the way down from the posterior to the main trap sheet. A simpler or more satisfactory correlation could not have been hoped for. Thus encouraged, a systematic search for new localities was made at the proper horizons for a considerable distance along the anterior and posterior valleys in the larger blocks, from Totoket Mountain northward to the gorge of the Farmington River at Tariffville. Mr. S. Ward Loper was engaged as special assistant in this work on account of his extensive acquaintance with the previously known fossil localities. The drift cover proved a serious obstacle to success, but did not entirely defeat the search. After repeated failure to find outcrops in the long anterior valley from Tremont to the north end of Higby Mountain, the shales holding a few fish scales were discovered in a little ravine near the north-bounding fault, and on making an opening here, some of the characteristic fossils of the anterior shales were speedily found. Associated with them was a bed of sandstone containing plentiful large reptilian foot prints. A second attempt was made in the anterior valley of Lamentation block, and as no natural outcrops were seen here, it was decided to open a short trench as near the proper horizon as could be determined. A less likely place, as far as local signs are concerned, for this sort of "fishing" could hardly be imagined, but the shales and their fossils were nevertheless unearthed not 20 feet from where the opening was begun. Similar black shales were found at three other points in the anterior valleys farther north—one near the outlet of Southington reservoir in Bradley Mountain block, another on the northwest slope of the same mountain, and a third near the base of the anterior shales on the west slope of Rattlesnake Mountain; but in all these localities the expense required to make sufficient trenches for thorough testing was more than could be justified, and the search there was given up before finding any fossil fish.

Search in the posterior shales was hardly less successful. During the progress of field work in the various topographic quadrangles my associates were instructed to keep a close watch along the horizons where the black shales might be expected, and to examine all outcrops carefully, with a view to opening the beds if they seemed promising. Two most fortunate discoveries of the posterior shales were made in this way. Mr. E. O. Hovey found the shales with fish-prints near the south end of Saltonstall Pond, about 100 feet beneath the posterior trap sheet of Pond Mountain. Good specimens of fish and plants were afterwards taken out from a trench opened here by Mr. Loper. Later, Mr. E. L. Rice found black shales containing fish prints north of Rocky Hill village. When further opened by Mr. Loper the beds yielded the *Ischypterus gigas*, known in the normal posterior shales elsewhere, and were therefore held to be of the posterior horizon.

Both of these discoveries of the posterior shales are of special value. The first is important because the sequence of main trap sheet, posterior sandstones and shales with fossils, and posterior trap sheet in the Pond Mountain district is thus shown to be stratigraphically equivalent to that of the same series of aqueous and igneous beds in the Higby block, where the posterior black shales with fossils have long been known in a stream bed about half a mile northwest of the village of Westfield, and where the trap sheets are more generally admitted to be of extrusive origin, and to be the equivalents of the main and posterior trap sheets in the various other blocks near by. An extrusive origin should therefore be assigned to the main and posterior sheets in Pond Mountain, even if no other evidence on the point were to be found. The reason that cumulative evidence is desirable here is that these trap sheets of the Pond Mountain crescent have, in spite of their very vesicular upper surface, with the detached trap fragments in the overlying sandstone (pp. 67, 72), still been regarded as intrusive sills by some observers.[1]

The posterior black shales near Rocky Hill are important in affording acceptable evidence of the identity of the neighboring trap sheet with the normal posterior sheet of other localities, and thus resolving a doubt which, as has already been said, was long felt as to the structure of that particular region. As now represented on the map, with colors to designate the various divisions of the formation, and fault lines distinctly drawn to bound the various blocks, no reason for doubt may appear; but the case is very different on the ground, where the trap sheets differ only in thickness, where the anterior, posterior, and upper sandstones and shales are seldom distinguishable, where the fossiliferous beds are very rarely found, where the fault lines are invisible, and where heavy drift covers a large part of the surface. Under such conditions the discovery of the black shales, with the distinguishing fossil of the posterior horizon, in such position as to fall in precisely with the scheme that had been tentatively held before, was an unexpected piece of good fortune. It served finally to confirm the belief as to the Cedar Mountain anticline, first advocated by Professor Rice, which had become stronger and stronger as field work advanced over the debatable area.

Black shales without fossils, as far as examined, were found at Gilletts Mills, in the posterior valley of Talcott Mountain, and again in North Bloomfield, about a hundred feet under the posterior trap a short distance south of Farmington River, here bearing many plant remains.

Several other localities of black shales at various horizons have not yet been correlated.[2] It is very desirable that a careful watch should be kept by interested observers on the ground for any chance openings that may in future disclose new localities of the fossil-bearing shales.

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[1] E. O. Hovey, Am. Jour. Sci., 3d series, Vol. XXXVIII, 1889, p. 378.
[2] S. W. Loper, Bull. Geol. Soc. America, Vol. II, 1891, p. 427.

New horizons may yet be discovered, and it is very probable that the two horizons here described in the anterior and posterior shales will each be found to consist of several closely associated beds.

### ORIGIN OF THE FAULTED MONOCLINE.

The specific understanding now gained of the attitude into which the formation has been thrown by tilting and faulting makes it possible to reach a correspondingly specific determination of the character of the deforming forces. A plausible solution of this problem was announced in 1886, and an account of it was given in the Seventh Annual Report of the Survey. Since then the faulted structure has been traced over a larger area; the suspected turn from the southwest-northeast trend of the fault lines in the southern district to the southeast-northwest trend in the northern district has been fully confirmed; the extension of the faulting over the drift-covered sandstone area in the northeast, and to a less extent into the adjacent crystalline uplands on the east and west, has been shown to be highly probable. But all this addition of well-proved fact and well-supported probability has not significantly modified the explanation previously announced. The argument leading to the explanation having been fully presented before, it need only be outlined here.

The hypothesis that the present monoclinal attitude of the formation represents the original attitude of deposition[1] need now hardly be discussed. It was suggested when few of the more detailed facts of structure now known had been discovered, and it disappears entirely before the array of structural details concerning the conglomerates along the eastern border, the extrusive origin of the lava flows, and the various features of warping and faulting.

The supposition that the monoclinal attitude resulted from the intrusion of the trap[2] is also only of historical interest. It would not be maintained to-day by its advocates, had they lived to trace the upper contacts of the passive lava flows and the oblique and curved paths of the fault lines. In recent years Hovey suggests that "since the transverse anticlinal (between Pond and Totoket mountains) can not owe its existence to the general tilting process, it may be assigned to another one, viz, dikes having a general WNW. course, most of which have not yet been exposed."[3] This suggestion is not accepted, because none of the visible dikes in the valley are accompanied by significant deformation of the adjoining strata. The heavy sills of the western ridges, the large dikes north of New Haven, and the great intrusive mass of Mount Carmel produce no noteworthy disturbance in the general monocline.

[1] H. D. Rogers, Third Ann. Rept. Geol. Survey Pennsylvania, 1839; Geol. New Jersey, 1840, p. 166; and other authors.

[2] This view was held by Percival and several other early writers.

[3] Am. Jour. Sci. 3d series, Vol. XXXVIII, 1889, p. 380.

Those who advocated a simple monoclinal tilting of the whole formation did so before good evidence of faulting transverse and parallel to the bedding had been found. The facts known to-day would naturally lead them to modify their first supposition.

Whatever the true explanation of the deforming forces is, it must be so framed as to account for some very explicit peculiarities of warping, faulting, and tilting. It must be competent to warp the strata into crescentic curves of various dimensions, and this without the exertion upon them of strong lateral compression, of which there are no indications in cleavage and minor folds. After or in the latter part of the warping process the deforming forces must break the whole formation into great blocks of variable breadth and length, sometimes straight, sometimes curved, here nearly parallel-sided, there wedge-shaped. The blocks must then be tilted eastward and faulted, with uplift generally on the east side of the fracture. The faults must affect not only the Triassic strata, but must penetrate eastward, westward, and downward into the crystalline floor on which the strata rest. That the crystalline floor should be affected with the rest of the formation has probably been an implicit condition of all theories of deformation, but something is gained by raising it to the rank of explicit statement and recognizing that the sandstones, however great their thickness may seem when we pace across their strike, are really only a part of the deformed mass, probably only a small part. The value of the sandstones is indeed not so much as a measure of the mass that was deformed, but as an index of the amount of movement that a great mass of other rocks suffered along with them.

It seems advisable to attach considerable importance to this explicit expansion of the problem of deformation, so that it shall concern not only the Triassic strata, but the crystalline schists as well. This is the more necessary because of the tendency when studying the schists of the uplands to regard their existing attitude as having been assumed at an ancient date and to overlook the probability that their position was significantly affected by any disturbance so late as in post-Triassic time. The same tendency may be seen in the study of the Paleozoic strata of Pennsylvania. It is demonstrable, in the neighborhood of the tilted Triassic strata of that State, that the dip of the Cambrian beds in pre-Triassic time differed by 15 or 20 degrees from their dip in post-Triassic time; yet it is seldom that this angle has been subtracted from the present dip in order to learn what must be accounted for by pre-Triassic deformation.

In our own region, when a liberal measure is given to the downward extension of the fault planes, and a juster idea is thus gained of the great slabs of crystalline rocks involved in the problem of deformation, the attention may be more easily diverted from the superficial to the deeper requirements of the problem. A fuller measure is thus gained of the facts to be explained, and naturally a safer approach may thereafter be made toward their true explanation. Treated as in fig. 34, the faulted monocline is seen from too near by, and its greater dimensions are not easily realized. Treated as in fig. 35, the relation of the surface parts to the whole is more easily perceived. It may thus be seen that a horizontal force, acting eastward near the surface with relation to the deeper mass and embracing a great thickness of fractured or

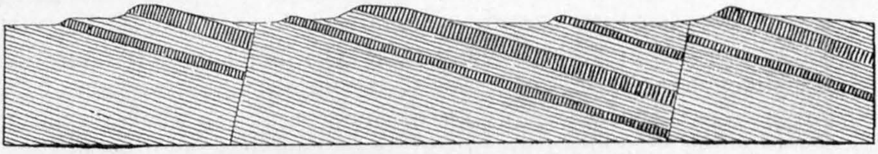

FIG. 34.—Section of superficial parts of faulted blocks.

fracturable earth's crust, would push one slab on another in such a manner as to produce a faulted monoclinal surface where a peneplain existed before, and that a series of horizontal strata resting on the peneplain must adjust themselves to the disturbance when it comes. The slipping of slab on slab would produce faults with uplift prevailingly on the east. The unequal yielding of the different parts of a slab

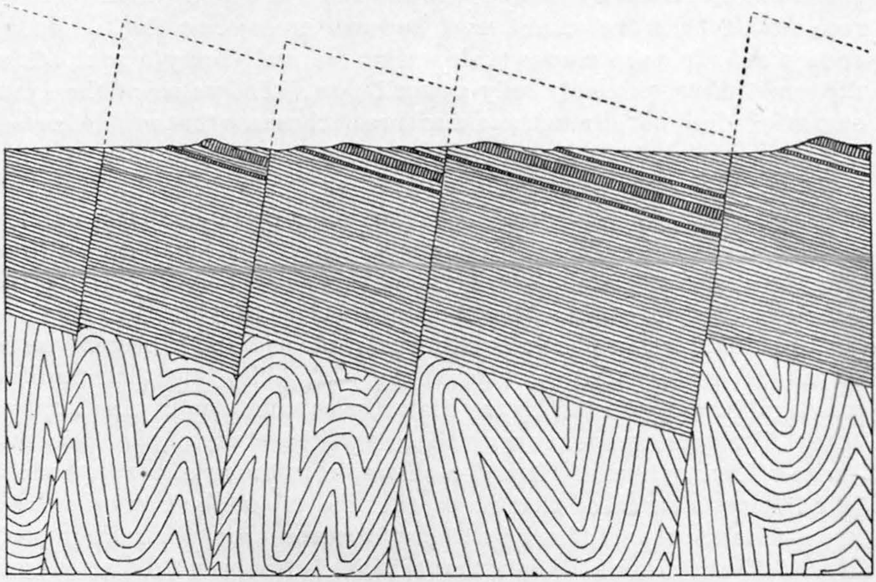

FIG. 35.—Deep section of faulted blocks of Triassic strata and underlying schists.

would cause more or less warping at its surface and in the overlying layers, and in this general way the crescentic structure of Pond and Totoket mountains may be explained in connection with the monoclinal faulting. Irregularity of deep structure might result in branching and curving faults, or it might produce occasional uplifts on the west of the fault line.[1] The local shearing forces produced in the faulted blocks

[1] Seventh Ann. Rept. U. S. Geol. Survey, 1888, p. 489.

would naturally cause slickensides on the dip lines of the weaker strata. To a limited extent a correspondence might be expected between the structure of the schists and the fractures by which they are broken into slabs, and it is thought that the bounding fractures of the Triassic blocks may be thus related to deep underground structures; but it is felt that more importance was at first attached to this point than it deserves. The correspondence is strongly suggested by the parallelism of the strike of the schists with the trend of the faults in the southwest part of our area, but it may well be that there is not so much correspondence between the dip of the schists and the dip of the faults. Until the crystallines of the uplands are attentively studied this element of the problem remains indefinite. It does not seem advisable to pursue speculation further in this direction at present. It is sufficient to have brought the faulted monocline within the accomplishment of horizontal east-west (but not necessarily from east to west) shearing forces, and to indicate that the fracturing of the blocks in the Triassic cover is plausibly connected with the faulting of the crystalline foundation. The forces by which the shearing was done appear to have acted in general in the same direction as those by which the trough was made, and all seem to have been of Appalachian habit.

A few words may be given to the eastern marginal faults. If the north-south faults are not continued beyond the short stretches where they form the Triassic boundary, and if the region be then viewed as a whole, the marginal faults of the eastern boundary may be regarded as a large-scale system of northeast-southwest joint planes, with short connecting fractures, according to the scheme advocated for smaller joint structures by Woodworth.[1] Thus treated, they somewhat resemble the system of faults that determines the eastern margin of the central plateau of France, where it is suddenly cut off by the descent to the valley of the Rhone.

[1] Proc. Boston Soc. Nat. Hist., Vol. XXVII, 1896, p. 182.

## PART III.—DENUDATION.

### GENERAL PRINCIPLES OF LAND SCULPTURE.

The implied separation of natural processes by the division of this report into distinct parts must be even more carefully guarded against in passing from deformation to denudation than in the earlier passage from deposition to deformation. It is true that in a general way the greatest deformation occurred after the deposition of the formation had been practically completed and before the great work of denudation had been far advanced. But, on the other hand, slight marks of both deformation and denudation are recognizable during the first chapter of deposition; denudation undoubtedly made some progress during the movements described in the second part on deformation; and one of the most important and far-reaching lessons of the region is found in the occurrence of a general upheaval—a gentle deformation—late in the final chapter of denudation, whereby the weakened forces of destruction were awakened into the cycle of renewed activity in which we now find them. Indeed, it is quite possible that deposition—not Triassic, but Cretaceous—occurred over at least a part of our area during the pause of denudation before the last general uplift. From beginning to end the contrasted processes are interlaced. They are here untangled and presented separately only in order more easily to follow the threads of description and argument.

During the early years of geological study little value was given to the ordinary processes of weathering and washing in the sculpture of land forms. Processes of uplift, deformation, and fracture were regarded as rapid, and it was thought that their almost unaltered effects were to be seen in mountains and valleys. The mighty currents of the sea, and not the smaller streams of the land, were regarded as erosive agents where erosion was so manifest that it could not be overlooked. By very gradual degrees a greater and greater value has come to be given to the slow but long-acting processes of weathering and washing, and a slower rate of action is given to processes of deformation. Thus guided, land forms are, as a rule, seen to be carved, not built. A noticeable phase of the advance in this belief was the introduction of the idea that the shore waves of the sea were the most effective agency for the greater tasks of denudation, and that broadly abraded surfaces were best explained as "plains of marine denudation," as they were called by Ramsay in 1847. It was admitted at this time that subaerial processes might carve valleys, but the wholesale reduction of a land mass to a surface of moderate relief was reserved for marine processes.

It was chiefly among American geologists that there afterwards arose a confident belief in the competency of subaerial processes to wear down any land mass to a peneplain of moderate relief; and although this seems to be nothing more than a legitimate extension of the general principles of denudation, it has as yet found less favor with many European geologists.

The processes of denudation are all dependent on antecedent processes of uplift or deformation, by which a land mass is raised above base-level and placed within reach of the destructive attack of weather and water. In contrast to earlier views, it has later been discovered that while slight movements are attended by earthquake shocks, thousands of such movements are required to produce great deformation and uplift; moreover, long intervals of relative repose occur between the times of these comparatively slight disturbances; and, furthermore, considerable changes of elevation probably go on without any violent disturbance whatever. Deformation is, therefore, much less rapid in comparison with denudation than was formerly supposed. A curious result of these discoveries is a contradiction of the early belief that rivers always run away from regions of uplift, and a very general acceptance of the principle that in certain cases rivers are competent to maintain their course in spite of an uplift, cutting their channels downward nearly as fast as the land beneath them rises. Rivers of this kind are called antecedent. It is nevertheless highly probable that only large and favorably situated rivers can maintain antecedent courses across strong uplifts, and that most of the greater uplifts of the world have determined the courses of many of the streams by which they are at first drained, such streams being called consequent. However this may be, it is not to be doubted that most uplifts are rapid enough to initiate land forms faster than the general process of denudation on the interstream surfaces can wear them down. Even where antecedent rivers occur, the land surface between them probably attains a form largely determined by processes of deformation and is influenced but little by denudation during the advance of these processes. Admitting freely that some denudation must occur during uplift, it nevertheless appears fair, in practically all cases, to estimate the period of uplift as of much less duration than the succeeding period of complete denudation. If the long, unmeasured period of time that is required to reduce an uplifted land mass to a smooth plain of denudation be called a geographical cycle, then it may be confidently said that the duration of the period of uplift or deformation is relatively short when compared to the entire cycle. Initial forms, slightly worn, characterize interstream surfaces during the earlier stages of the cycle; the later stages witness the development of a succession of changes, ending in the reduction of all forms to a surface of faint relief close to sea level.

Let it be conceived for the moment that after the initial uplifts no movements of the land mass occur during the cycle of denudation. It may then be proved by a great variety of evidence that the forms produced by the processes of denudation are remarkably systematic in their succession, certain forms being associated with early stages, others with later stages, and still others with the ultimate stages of the destructive processes; and this statement holds true whether the agency of destruction is wholly subaerial or wholly marine, or a reasonable combination of both. It has thus come to be convenient to subdivide a cycle of denudation into infantile, youthful, mature, and senile stages, each one merging into its successor. The infantile stage is very brief. The stage of absolute old age would be everlasting but for the intervention of new disturbances.

Leaving antecedent rivers out of consideration for the moment, it may be shown that infancy is characterized chiefly by slightly modified initial forms with consequent drainage, the stream channels being little incised beneath the surface. Youth witnesses the deeper dissection of the surface by the streams, and the production of narrow, steep-sided consequent valleys, all running down the slopes of the initial surface. Lakes that might have occupied initial basins are largely obliterated by the detritus that is actively washed from the consequent valleys. Youth advances toward maturity when the chief consequent streams have cut down their channels to such moderate declivity that their ability to do work is just equal to the work they have to do; or, in briefer terms, when the streams have graded their courses. This result is accomplished by degrading their channels in the uplands and aggrading depressed areas or basins. As the graded slope is approached the stream is acted on by a smaller component of gravity than before. Any obstacle in its path is, therefore, more effective in deflecting it, and it thus begins to swing irregularly from side to side, first undercutting one valley slope and then the other, and thus widening the valley floor. Thus flood plains are initiated, and after this stage is reached they are characteristic features of land surfaces. When the beautiful equilibrium involved in the graded slope is once gained, it is always preserved by means of very slight and slow changes through the rest of the cycle, as long as the land mass is undisturbed.

Youth advances into maturity as the smaller side streams grow headward from the consequent streams and dissect the uplands, increasing the variety of surface forms to a maximum. Where the side streams are located by what may be called accidental controls, indifferent to the form of the initial surface and independent of internal structural guidance, they may be called insequent branches of consequent streams. Where they are subject to guidance by headward erosion along weak structures, laid bare on the valley walls of the consequent streams, they may be called subsequent branches, here following the descriptive term used by Jukes in his important essay on the origin of certain river valleys in the south of Ireland.[1] Subsequent streams are distinguished

[1] Quart. Jour. Geol. Soc. London, Vol. XVIII, 1862, p. 378.

from consequents and insequents without much difficulty, inasmuch as the first follow the lead of the initial slopes and the second are unrelated to weak internal structures. With the headward growth of the insequent and subsequent branches of the consequent streams the initial interstream surface is more and more dissected, and from having been imperfectly apportioned among the various river basins in infancy it is more and more sharply and definitely divided and subdivided among them as maturity is reached. Not only so; it frequently happens, and especially in regions of disordered structure, that the subsequent streams grow so far and so rapidly along the weaker structures as to tap the smaller consequent streams and divert their upper courses to the basins of the master consequents, leaving the beheaded lower courses of the smaller consequents to survive as best they may with diminished volume. An adjustment of streams to structures is thus brought about—an adjustment of the most wonderful kind, not to be accomplished in any other way than through the most deliberate and thorough search for the best lines of drainage by the processes of subaerial denudation. Although the principles of uniformitarianism, liberally interpreted, do not need more proof than has been given by the school of Hutton, Playfair, and Lyell, it should be noted that the occurrence of well-adjusted drainage systems affords them a confirmation of the most unequivocal kind.

In infancy and youth the streams are for the most part engaged in cutting down their channels, because the initial slopes offered for their guidance generally give them a greater competence than they need in the performance of their work. But, as a result of intrenching their channels, their lower and middle courses become flatter and their ability to carry their load of land waste is lessened. At the same time the area of valley-side slopes is expanded, and the load which comes chiefly from these slopes is therefore increased. Thus the two variable quantities, ability to do work and work to be done, approach and reach equality; this stage in the larger consequent streams serving, as has already been said, to mark the entrance into maturity. Now the side streams, insequent and subsequent, are growing rapidly. As they dissect the interstream surfaces the area of steep valley sides is even more expanded, and the load delivered to the larger streams grows far beyond its earlier measure. The graded master streams are therefore often required, on approaching maturity, to increase their ability to do work to equality with the increased work that they have to do, and they obey this requirement most ingeniously by continually laying down some of the excessive load and in this way steepening their slope and thereby at once increasing their ability and decreasing their work so that the relation of equality between the variables is always maintained after once being established. When this process is coupled with the lateral swinging already mentioned, the development of alluvial flood plains becomes distinct; and it is thought that many flood plains are in great part thus determined.

Maturity is the period of fullest development of functions. The drainage basin of a river is then most perfectly divided and subdivided. The courses of the streams have come to be well adjusted to the structures that they are working upon. The rainfall is most perfectly shed from the dissected surface to the streams, seldom delayed in lakes, seldom impetuously hurrying in falls. Although the valley sides are less steep than in youth, their area is so greatly enlarged that the rate at which land waste is shed to the streams is at its maximum. The streams are nevertheless everywhere able to handle the land waste that is washed into them; there is an active transportation down the valleys to the sea. But, as further time passes, the activities of maturity are tamed down to the placidity of old age. The mountains and hills are slowly subdued, their slopes becoming flatter and flatter, and their waste creeping down the slopes more and more slowly. With loss of height, rainfall normally decreases; but with less rainfall and less slope the stream heads shrink a little, as the smaller twigs fall from an old tree.

In only one feature is greater perfection found in old age than in maturity—the extension of the graded condition from the stream channels, where it is first developed, to the general interstream surface, over the whole of which it finally spreads. In order to make this more apparent, a close analogy may be drawn between streams of water in channels and streams of waste on unchanneled slopes. The upper course of every water stream is a waste stream; one grades into the other. Water streams always contain some waste, and both water and waste vary in volume with changes of weather, season, and climate. Waste streams often contain water, the proportion of the two constituents varying as before. Both are guided by a component of gravity; but while one runs the other creeps. Both attain a graded condition by the establishment of an equality between ability to do work and work to be done. Consequent water streams generally establish a graded slope by entrenching their channels, because their initial slope is commonly so steep that their youthful ability is in excess of their work. Consequent waste streams, on the other hand, generally establish a graded condition by calling for a reduction in the rate of supply and a refinement in the texture of the load to be carried, because their initial slopes are seldom steep enough to enable them to carry forward all the waste that is provided for them by the weather. The early products of weathering remain on the surface, hardly moved; the further attack of the weather then deepens the layer of loosened material—the "discrete" of Gilbert; at the same time the texture of the surface waste is refined by the continued attack of the weather on it, and the rate of underweathering is decreased by partial protection from the atmosphere. With the decrease in the rate of production of waste, and with the refining of the surface texture, the work to be done falls to equality with the working ability of the waste stream.

Then the graded condition is reached, and this is the ordinary condition of soil-covered surfaces.

Some consequent water streams, however, find the initial land slopes very faint, or even reversed, as in initial lake basins. Here they are unable to carry forward the load that is brought to them, and they must increase their loitering velocity by aggrading their floor to a steeper slope. Conversely, some consequent waste streams find the initial slopes very steep, as on initial cliffs of displacement; and here a grade must be established by the same process as that ordinarily employed by young water streams. The cliff must be weathered back to a less declivity, so that the removal of detached waste shall be no faster than its supply. But this is a very slow process; it is not accomplished nearly so soon as the grading of the water streams.

As the result of the action of water and waste streams, the whole area of an initial land surface gradually assumes a new form, which may in general terms be called a sequential form. The development of sequential forms by consequent waste streams is generally very slow, but their development on the valley sides of consequent water streams is rapid. On these sequential side slopes both insequent and subsequent waste streams are to be found; indeed, the water branches of consequent streams already described always begin as waste streams. As a rule the sequential waste streams on the sides of consequent valleys at first have an ability of transportation much in excess of the supply of load, hence bare rock faces are commonly found along the sides of narrow valleys. But as the valleys widen and their slopes flatten a greater and greater part of the lateral waste streams is graded.

The analogy to be drawn between streams of waste and streams of water may be extended to various details. The long waste-covered slopes that prevail on many of our western mountains are closely analogous to the flood plains of our great rivers. Truly the two differ in various aspects, the slope of one being faint, of the other steep; the texture of the waste in a flood-plained river being fine, on a mountain side coarse; the one being associated with a large river, the other only with wet-weather wash; yet both are forms assumed by the waste of the land on the way to the sea, and their likenesses are as important as their differences. The bare ledges from which the even waste slopes descend are like the ungraded upper courses of streams; both water and waste here plunge in torrents or rock falls, and the forces of transportation are able to carry much more load than is given to them. The ledges that interrupt the grade of waste slopes are like falls in the course of young streams, where ability is locally increased by increase of declivity; this in turn resulting from the quicker degradation of some weaker structure farther downstream. Waste falls are therefore much like waterfalls, even though intermittent instead of relatively continuous in behavior, and truly as much less picturesque in the landscape as they are less manifest in action.

During early infancy much the greater part of a land surface is occupied by the ungraded consequent waste streams which modify its surface very slowly. The lesser part of the surface is trenched by consequent water streams. As time passes a greater and greater part of the initial surface is reduced to a sequential surface. Where the water streams act on the sequential surface they reduce it to a grade in the late youthful stage of the cycle of denudation if they are large, in the late mature stage of the cycle if they are of moderate size. Where only waste streams act they do not succeed in grading the whole surface until a very late mature or early senile stage, for not until then are all the ledges cloaked over with waste and the whole surface reduced to smoothly modulated profiles, the forces of transportation being everywhere competent to perform the work offered to them. A landscape of this kind is found in certain parts of the driftless area of Wisconsin, and although the mental effect produced by it is altogether unlike the sensation caused by vigorous, maturely dissected mountains, the landscape is nevertheless extremely pleasing, and all the more so when it is recognized that the flowing profiles everywhere apparent are the sign of the perfection of drainage processes, in the waste streams on the slopes as well as in the water streams in the valleys. A completely graded surface of this kind, where the ability to move the waste down the slopes is everywhere just equal to the work required in moving it, is another witness to the verity of uniformitarianism, and not a less trustworthy witness than is found in a system of well-adjusted water streams.

The account thus far given of the processes and forms appropriate to various stages of a geographical cycle has been stated in a general manner. It postulates an ordinary condition of climate, neither glacial nor arid. It does not specify the structure of the land mass, although upon that will depend the detail and arrangement of form at every stage of development. It omits all consideration of movements of the land mass during the progress of the cycle, by which its orderly advance from infancy to senility might be interrupted. All this generality is intentional, for it is only after reaching the most general view of the cycle and its stages that the variations from it can be fully appreciated. No land mass has been observed to pass through its cycle of changes. It is only by putting together examples of many different land masses in earlier and later stages of denudation that their succession can be determined, and even then many gaps in the series must be filled in by reasonable deduction. But on the whole the general sequence of processes and forms through the normal cycle appears to be well established, and its verity is attested by a great array of actual occurrences.

In order to illustrate the elasticity with which the general scheme of the geographical cycle accommodates itself to special cases, variations in structure and in climate may be briefly considered. Variations of structure call forth appropriate variations of initial and sequential form and drainage. In regions of horizontal strata subsequent streams are wanting, and the rearrangement of drainage lines by capture and diversion is relatively unimportant. The master consequents have in maturity a great number of insequent branches, dividing again and again, and the waste streams, as well as the smaller water streams, have numerous falls over the benches or cliffs that contour around the digitate hill spurs between the divaricating water ways. In regions of tilted or folded structures the master streams develop numerous subsequent branches along the weaker structures in youth and maturity, and the harder structures survive in ridges. Rearrangements of drainage by capture and diversion of consequent headwaters along subsequent courses are extremely common; hence maturity witnesses extensive adjustments of streams to structures. Waterfalls are relatively rare. Waste falls are common. They are not systematically arranged on contouring cliffs, but follow the disordered outcrops of the harder strata wherever they lead. In maturity they are chiefly head-waste falls on the ledges near ridge crests, being in this respect comparable to headwater torrents in the uppermost courses of water streams.

In an ordinary climate with sufficient rainfall vegetation covers the greater part of the interstream surfaces, diminishing the number of water streamlets and increasing the area of unchanneled slopes during the varied dissection of the uplands. In a subarid region, where vegetation is scanty, the number of stream channels is greatly enlarged, in spite of the small measure of rainfall; nearly the whole surface is carved by little water ways, and the proportion of washed waste to creeping waste is greatly increased. Thus the difference between the flowing profiles of the driftless area of Wisconsin and the angular forms of the "bad lands" on the Western Plains may be in great part accounted for. Their chief contrast is found in the greater competence of water-stream carving on the bare surface of the latter, and in the relatively greater competence of the waste-stream creeping under the grassy cover of the former. In one the divides are almost sharp, as is appropriate under the action of running water;[1] in the other the divides are broadly convex, as is equally appropriate under the control of creeping waste.[2] In regions of greater aridity than the Western Plains the wind becomes the chief means of transportation. Under a cold and damp climate ice sheets are the chief agency of movement; and each of these produces its appropriate variations from the sequential forms of the normal cycle.

No class of interruptions in the normal advance of the cycle is more important than that which results from land movements of any kind. Movements of land mass with respect to base-level may happen at any time during the advance of a cycle. They may be of any degree of

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[1] Gilbert, Geology of the Henry Mountains, p. 116.

[2] See note by author on Convex profiles of bad-land divides: Science, Oct. 28, 1892.

complication from simple regional uplifts to the greatest deformations of mountain ranges. They may be long separated in time, as in the more stable parts of the earth's crust, or they may follow at short intervals, as in uneasy and growing mountain ranges, different regions showing a great diversity in this respect, as instanced by the stability of the oldlands of Wisconsin and Minnesota from Cambrian times to the present, in contrast to the uneasiness of the Appalachian belt during the same long portion of geological history. In no part of this attractive field of study is the importance of the deductive method greater than here. The ideal advance through a normal uninterrupted cycle of denudation must first be deliberately investigated. Equal attention must then be given to variations from the normal cycle, for only in this way can the elements of the scheme hope to match the variety of nature.

The simplest interruptions in the normal cycle are caused by uniform, regional movements of elevation or depression. A region that has reached a certain stage of denudation while standing in a certain attitude with regard to base-level may be raised or lowered, and its further development then advances with respect to a new base-level. Graded streams and graded waste slopes are particularly sensitive to all such changes. Many examples of composite topography resulting from essentially uniform vertical movements of land masses are now familiar. Gentle deformation is more probable than uniform uplift or depression; certain streams are then accelerated, others are retarded; and if the warping occurs in late maturity or old age, many rearrangements of drainage may be expected to follow, as has been shown by Campbell.[1]

Strong and relatively rapid deformation by folding and faulting may occur. Then few streams besides the largest ones can survive in their former courses; and here we find the appropriate place for the consideration of antecedent rivers. Most of the smaller streams will be extinguished or shifted by strong deformation and replaced by new consequent streams. Many mountain regions offer opportunity for study of river rearrangements by strong interruptions of this kind; few better than the Rocky Mountains of Montana, where the form and drainage established in one cycle of mountain development have been greatly affected by the strong warping that produced the basins and inclosing ranges of late Tertiary and Pleistocene time.

The reason for introducing so long a departure from our local problem at this part of the report is the desirability of emphasizing the principle that only by thus generalizing all pertinent knowledge can appropriate explanation be found for the existing forms of the Triassic belt. The case is essentially similar to that by which the origin of the lava sheets, or the existence and arrangement of the fault lines, has been determined. If existing forms are to be explained by denudation, the general principles governing land sculpture must be studied out as

[1] Drainage modifications and their interpretation: Jour. Geol., Vol. IV, 1896, pp. 567-581, 657-678.

carefully as the land forms are observed; and only by correspondence between observed fact and deduced consequence of theory can a safe conclusion be reached. The competence of subaerial denudation to produce a peneplain may be doubted by some to-day, just as its competence to carve valleys was doubted a century or two ago. It is only by comparing the deduced features of a peneplain with the observed features of a land surface that the doubts of the case can be settled.

## CYCLES OF DENUDATION.

There are three stages in the geographical development of the Triassic valley lowland that deserve special attention. The first includes the form initiated by deformation and more or less modified by contemporaneous erosion. The second is the peneplain to which the initial form was reduced by long-continued denudation, this being inferred from the remnants of the peneplain still preserved in the Eastern and Western uplands. The third is the form of to-day, carved in the peneplain after a gentle slanting uplift to about its present altitude. Shortly before the existing form was assumed came the glacial invasion, commonly recognized in New England, and associated therewith are certain changes of level, especially significant near the shore line; but these are minor episodes compared to the long and almost complete cycle of erosion by which the general peneplain was made, or even to the partial cycle during which the valley lowland was worn down between the uplands.

For ease of description and explanation, the geological dates of these critical stages may be given at once. The topography produced by the monoclinal tilting and faulting may be roughly dated as falling at the initiation of Jurassic time; Jurassic being used here not to indicate a very definite epoch, but rather to imply a succession after Triassic deposition. Jurassic time included a cycle of denudation that sufficed to reduce the forms initiated by tilting and faulting to a peneplain; and as the latter is found to be associated with Cretaceous strata, it will be spoken of as the Cretaceous peneplain. Its preparation appears, however, to be chiefly the work of Jurassic time, and a record of this division of the geological scale on our Atlantic slope is more likely to be found in the work of obliteration here manifested than in the usual record of deposition. The uplift and the carving of the peneplain into the form of to-day were accomplished in Tertiary time; hence the Valley Lowland will be spoken of as the work of the Tertiary cycle of denudation.

## THE INITIAL FORM OF MONOCLINAL FAULTING.

In order to emphasize the progress of denudation during the production of the faulted monocline, the topographic form attained at the close of the period of faulting and tilting is discussed in this rather than in the preceding part of the report. If the Triassic formation consisted of resistant lava beds throughout, and if there were any reason for thinking that the climate of the region at the time of deformation was arid, then it might be argued that denudation made insigificant progress during deformation, and that the initial form of the region could be reproduced by a purely geometrical reconstruction of the faulted blocks. The warrant for this is found in the remarkable region of faulted lava blocks in southern Oregon, as described by Russell.[1] In certain parts of that remarkable district "the orographic blocks of dark volcanic rock are literally tossed about like the cakes in an ice floe, their upturned edges forming bold palisades that render the region all but impassable. . . . Many of these fragments measure a mile or so on their edges, and are tilted in various directions, leaving narrow rugged valleys between their upturned margins. The diverse tilting and the numerous fault scarps that rise without system into naked precipices combine to make this a region of the roughest and wildest description." Some of the displacements measure 5,000 or 6,000 feet. The fault scarp overlooking Warner Valley is about 2,000 feet high. The orographic blocks may be traced 10, 20, 50 miles; they are long and narrow, sometimes in a ratio of 50 to 1. Some of the fault scarps have been carved more than others; that forming the eastern face of the Warner Mountains has deep recesses between outstanding pinnacles, with extensive alluvial fans outspreading from the mouths of the ravines. Other scarps are slightly affected by the attack of the weather, preserving their even, fractured face, and having a relatively small volume of detritus accumulated at their base. Many of these blocks thus represent hardly any advance from a purely initial topographic form. Some of the faults are, in their later movements, so recent as to displace the alluvial fans that spread out from the ravines and the Pleistocene lake beds that occupy the depressions.

## INITIAL DRAINAGE OF THE FAULTED MONOCLINE.

It is not, however, to be expected that the faulted Triassic sandstone and pebble beds in Connecticut were so little modified by erosion at the time when deformation was essentially completed as the tilted lava beds of Oregon now are. The Triassic strata are even now comparatively weak, and then must have been weaker. The climate was at that time probably wetter than now, for, in Jurassic time, the Great Plains of the West lay beneath a long and broad arm of the sea that stretched from the Gulf of Mexico toward the Arctic Ocean, and the Appalachians were probably not a lofty range of mountains. Yet it is believed that even under these unfavorable conditions the form initiated by faulting must have for a time dominated the relief of the Triassic belt and of its bordering area. The uplifted edges of the tilted blocks, ravined and dissected even while rising, must have determined the site

[1] A geological reconnaissance in southern Oregon: Fourth Ann. Rept. U. S. Geol. Survey, 1884, pp. 433–464.

of initial ridges. Short consequent streams must have run down the back slope, and shorter streams must have gnawed into the faulted face of each block. Much of the drainage must have been discharged along the fault lines between the ridges, for the lines of relative depression along the down-faulted side of each block must have determined the course of many longitudinal consequent streams. Lakes may have for a time occupied the lowest depressions between the tilted blocks in the Hartford area and along the eastern border, but it is quite possible that the long and narrow basins may have been filled as fast as they were depressed by the waste brought down by streams, aggraded alluvial plains thus taking the place of bodies of standing water. Omitting for the moment all consideration of antecedent streams, the longitudinal consequents should have overflowed eastward from the basins of depression in the direction of the general tilting, and thus escaped toward the sea along various unknown paths. Each warped basin, from the small southern crescent of Pond Mountain to the great Springfield crescent (p. 86), may be supposed to have its own eastward discharge; and it is possible that for this reason the Connecticut leaves the Triassic lowland as it approaches the southern end of the Springfield crescent at Middletown. This scheme of drainage might be called the early Jurassic consequent drainage. It is roughly shown in fig. 36.

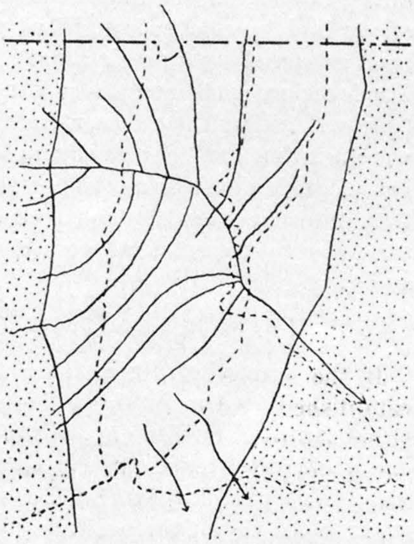


FIG. 36.—Diagram of consequent Jurassic drainage.

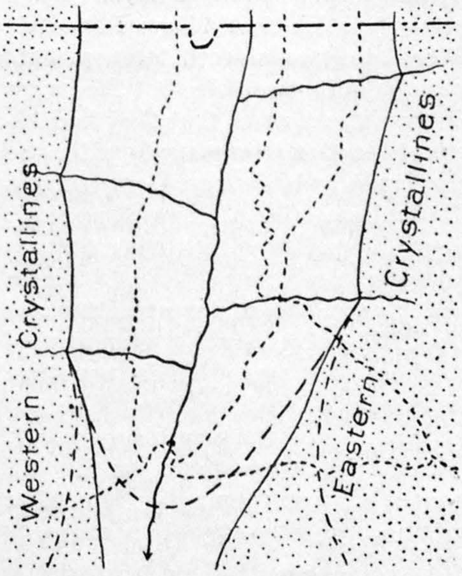


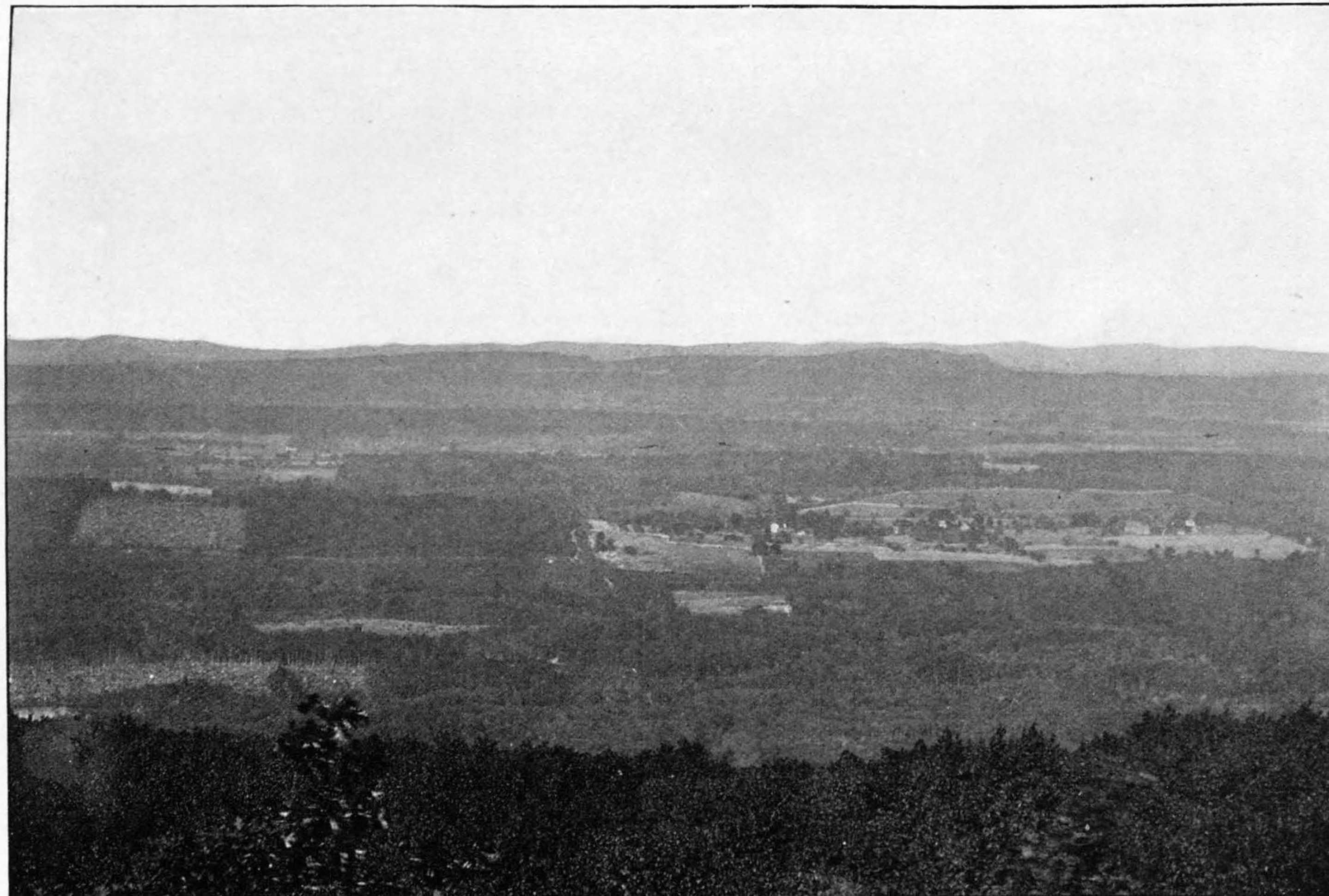

FIG. 37.—Diagram of antecedent Triassic drainage.

On the other hand, a certain share of drainage may have been retained by antecedent streams, that is, by such streams as held their previous courses in spite of the monoclinal faulting. These might be called the persistent Triassic streams. They are sketched in fig. 37.

It may be plausibly supposed that in the later stages of the chapter of deposition numerous streams entered the Triassic estuary in centripetal arrangement from the land areas on the east and west; and if the Triassic strata had been evenly uplifted the centripetal streams might have been gathered by a single axial river, presumably flowing southward. The location of the mouths of the chief centripetal streams might now be identified by the coarser texture of the marginal sediments, if it were not that drift covers so much of the lowland. The stronger members of such a system might persist during the monoclinal faulting, and thus determine a scheme of antecedent drainage, gathering numerous subordinate streams that were developed consequent on the faulting and tilting.

### CHANGES FROM INITIAL TO EXISTING DRAINAGE.

If the antecedent Triassic or the consequent Jurassic rivers, once established, had maintained the same pattern from the time of monoclinal faulting until to-day, nothing would be easier than to determine the origin of the existing river systems of the region by comparing them with the very unlike patterns of antecedent and consequent arrangement inferred to exist at the beginning. But in the long time since the monoclinal tilting there have been many opportunities for the new arrangement of stream courses; hence it is not to be expected that the existing stream pattern should closely resemble the pattern that prevailed when the carving of the monoclinal blocks began. There must have been, in the first place, many spontaneous rearrangements from the initial courses, due to the development of subsequent streams and the migration of divides during the Jurassic cycle of erosion that followed the first period of disturbance. These will be briefly considered later (p. 162). During the phase of peneplanation it is quite possible that the larger rivers may have wandered laterally from the courses that they had followed previously. It is eminently possible that after peneplanation the southern part of the State was submerged and blanketed over with a series of Cretaceous sediments, and that these, when afterwards uplifted, formed a smooth coastal plain, extinguishing all the preexisting streams beneath it and bearing on its surface an entirely new series of streams; the latter would then be superposed upon the under structures as the coastal-plain cover was worn off. Manifestly such streams could show no essential resemblance to those that prevailed in the same region at the time of monoclinal faulting. This part of the problem is discussed on page 165. After peneplanation and the possible episode of the coastal plain, the region was uplifted with a slant to the south. All the streams, whether on the never-covered or on the uncovered surface of the unlifted peneplain, then had a new opportunity for spontaneous rearrangement during the dissection of the uplifted peneplain. A few examples of changes thus produced are described on page 175. Finally, the changes determined by glacial action, directly or indirectly, are considered on page 181. All these possible changes, caused by two



THE VALLEY LOWLAND, AS SEEN FROM THE WESTERN UPLAND.

series of spontaneous rearrangements, one period of wandering on a peneplain, one probable opportunity for superposition, and at least one disturbance by glacial action, must be allowed for before the existing rivers can be compared with their ancestors, and the complexity of the problem thus becomes so great that it is hopeless at present to attempt a positive solution of it; but a comparison of theoretical initial drainage and existing river systems will be briefly made (p. 184).

## THE CRETACEOUS PENEPLAIN.

The erosion initiated by the faulting and uplift is believed to have progressed so far during a period roughly measured by Jurassic time, and before any later movement of significant amount occurred, as to have swept away all the initial relief and reduced the region to a low peneplain. The evidence of this is found less in the Triassic area than over the adjoining uplands.

A view of the Eastern and Western Uplands from any commanding summit of the trap ridges, such as West Peak, near Meriden, or Mount Tom, in Massachusetts, discloses the even sky line to which their rolling hills so regularly ascend. The same persistence of accordant altitudes over large areas may be seen from any of the eminences that surmount the general level of the uplands by a small amount, such as Great Hill, near the village of Cobalt, north of the Connecticut gorge in the Eastern Uplands. The upland has a gentle southward slant, so that it becomes a lowland near the coast line, and its further continuation would pass beneath sea level. The steep-sided narrow valleys of the uplands are sunk beneath the general summit level thus marked. Before the excavations of these valleys the uplands must have been much more continuous than now; they must, indeed, have constituted a gently rolling region of moderate relief—a peneplain.

It is one of the most important principles of land sculpture that a peneplain of this kind in a region of greatly deformed crystalline rocks can be explained only by supposing long-continued and widespread erosion, either by land or by sea forces, which reduced all structures, crystalline and Triassic indifferently, to a surface of small relief, everywhere close to the controlling base-level of the time. In no other way can the even beveling of the surface across the disordered rocks be accounted for; in no other way can the complete lack of sympathy between structure and form find rational explanation.

In such case the peneplain must have stood close to sea level, and only at some later time could its present altitude have been gained. Hence the first movement of post-Triassic elevation in connection with the monoclinal faulting could not have lifted the region so high as it now stands. The present elevation of the region must have been gained after peneplanation, by a process of general uplifting without significant deformation. Between the first period of elevation, with faulting and tilting, and the second period of elevation, characterized by a broad bodily uplift, there must have elapsed a time of long-continued denudation.

Although this principle is now generally recognized and applied in many researches of the last two decades, it is curious on reading over older geological reports and treatises to see how generally it has been overlooked. Percival recognized the plateau-like form of the uplands, but his explanation of it was characteristic of time when denudation had not gained its present rank of equality in the long run with deformation. He wrote:

The eastern and western Primary may both be regarded as extensive plateaus, usually terminating abruptly toward the larger Secondary basin [the Triassic Valley Lowland of this report], but sinking more gradually toward the Sound on the south. These plateaus present, when viewed from an elevated point on their surface, the appearance of a general level with a rolling or undulating outline, over which the view often extends to a very great distance, interrupted only by isolated summits or ridges, usually of small extent. . . . The western Primary . . . forms within the limits of this State a wide plateau, such as has been described, generally very abrupt, both west and east, toward that valley [the Berkshire limestone valley] andthe larger Secondary basin [the Triassic Valley Lowland], and of so uniform an elevation that from many points little elevated above its surface the view extends across its entire width and to a great distance north and south, presenting throughout its whole extent no extensive range of mountains, but only a series of more or less detached and isolated elevations. . . . The eastern Primary, viewed

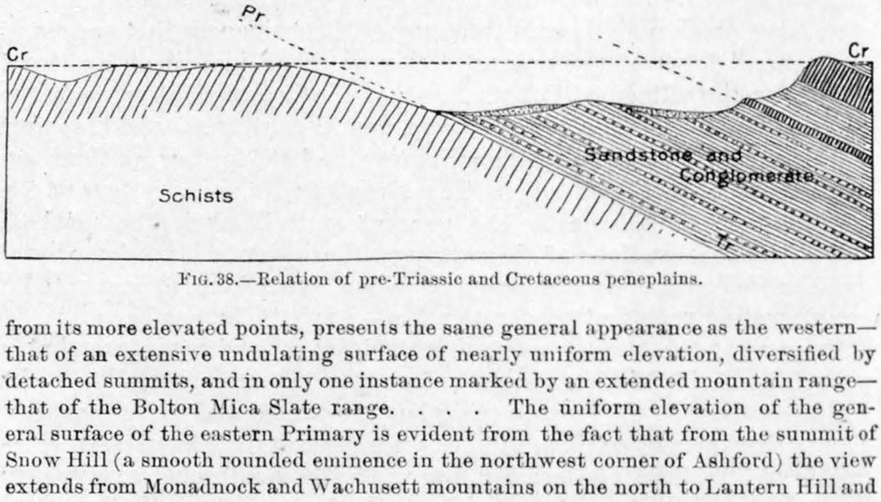



FIG. 38.—Relation of pre-Triassic and Cretaceous peneplains.

from its more elevated points, presents the same general appearance as the western—that of an extensive undulating surface of nearly uniform elevation, diversified by detached summits, and in only one instance marked by an extended mountain range—that of the Bolton Mica Slate range. . . . The uniform elevation of the general surface of the eastern Primary is evident from the fact that from the summit of Snow Hill (a smooth rounded eminence in the northwest corner of Ashford) the view extends from Monadnock and Wachusett mountains on the north to Lantern Hill and the East Haddam Hills on the south. . . . These plateaus are also intersected by valleys and basins, which serve to mark the arrangement of their surface even more definitely than the elevations. This arrangement will be found to correspond very exactly with that of the geological formations, indicating that it was caused essentially by the original form of the surface of these formations and not by any subsequent denudation.[1]

The last sentence is particularly striking in illustration of the little value formerly attached to the process of denudation in shaping the earth's surface.

The discussion of the floor on which the Triassic formation rests, in Part I (p. 23), might be repeated here, in so far as it touches upon the reduction of a region of strong relief to a gently rolling lowland;

[1] Geology of Connecticut, 1842, pp. 477-480.

but the previous discussion was concerned with the reduction of the great folds of the crystalline schists from their initial mountainous elevations to a pre-Triassic peneplain, while the present section considers the reduction of such elevation as was given to the crystalline areas by the warping, tilting, and faulting of Triassic and immediately following time to a post-Triassic peneplain. The relation of these two peneplains is indicated in fig. 38. Of the first, PrTr, a great part is lost in the air, a considerable part is buried under the Triassic strata, a small part is revealed with moderate change in the slope where the Western Upland descends to the Valley Lowland in the even Southington belt of the western border. Of the second peneplain, CrCr, a large part remains, with moderate change, in the rolling surface of the uplands; an unknown part is lost on the south, where the slanting upland descends to a lowland and passes beneath the sea level; a certain part is lost where the valleys of the uplands are now excavated, and an important part is lost where the Triassic Valley Lowland has been etched out.

### RELATION OF TRAP RIDGES TO THE PENEPLAIN.

It is only in the even crest line of some of the trap ridges, such as Totoket Mountain, closely accordant in altitude with that of the neighboring Eastern Upland, that the peneplain is recognizable within the Triassic area. Yet it can not be doubted that the even surface of denudation represented in the gently rolling uplands was once continuous across the region of the present Valley Lowland. The process of peneplanation is not local but far-reaching. If it finds application in the Eastern and Western uplands, it surely must be applicable over the intermediate area, and the more so when it is remembered that the Triassic strata are, with the exception of the trap rocks, notably weaker than the general body of the crystalline schists. The Triassic belt must have been more than a peneplain; it must have been a true plain of denudation during the later phases of the Jurassic cycle of denudation, relieved only here and there by low swells of stony soil marking the outcrops of the trap sheets and dikes. To this ultimate form must the initial faulted blocks have been at last reduced, passing on the way through a whole series of intermediate forms. During so long a cycle of denudation, acting upon a mass of greatly varying hardness, many spontaneous adjustments of streams to structures must have taken place, and at the close of the cycle it is not to be conceived that any but the larger streams could have preserved the courses that they had possessed, either through antecedent or consequent origin, at its beginning.

The reconstructed sections of Pl. XX clearly indicate that the trap sheets as well as sandstones must have suffered much denudation in the reduction of the original faulted monocline to a peneplain. This can admit of no doubt when the correspondence of the trap sheets, intrusive as well as extrusive, in adjacent blocks is once perceived. It is, moreover, entirely by such denudation, causing a retreat of the outcrop in the direction of the monoclinal dip, that the extrusive trap ridges are now so systematically offset from one another, and a similar process controls at least to a considerable extent the relative location of the intrusive ridges. The offsets of the ridges has therefore nothing to do with lateral faulting; this is excluded by the nearly vertical position of the slickensides occasionally revealed on the fault planes, as in the Meriden quarries. The trap ridges of the Lamentation block afford excellent illustration of this conclusion. Standing on the bold bluff at the south end of Chauncy Peak, or on some of the higher points along the arching crest line of Lamentation Mountain itself, one may, in imagination, reconstruct the trap sheet obliquely upward along the plane of dip into the air, or obliquely downward underground with equal confidence, as in fig. 39; but in either direction it must be extended only as far as the fault planes that bound the block. No definite limit may now be set to the oblique extension of the main sheet, or of its aqueous and igneous fellows, *within* the Lamentation block, but there is every probability that it is here to be measured in

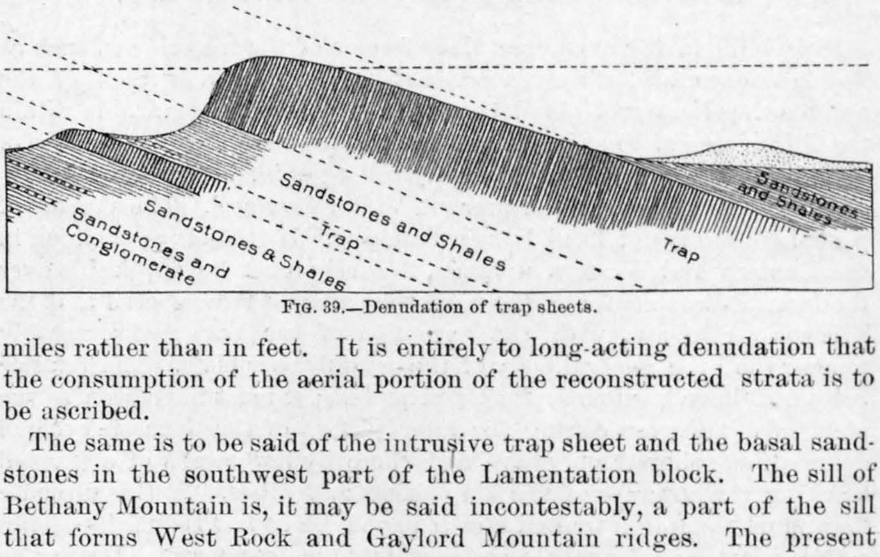
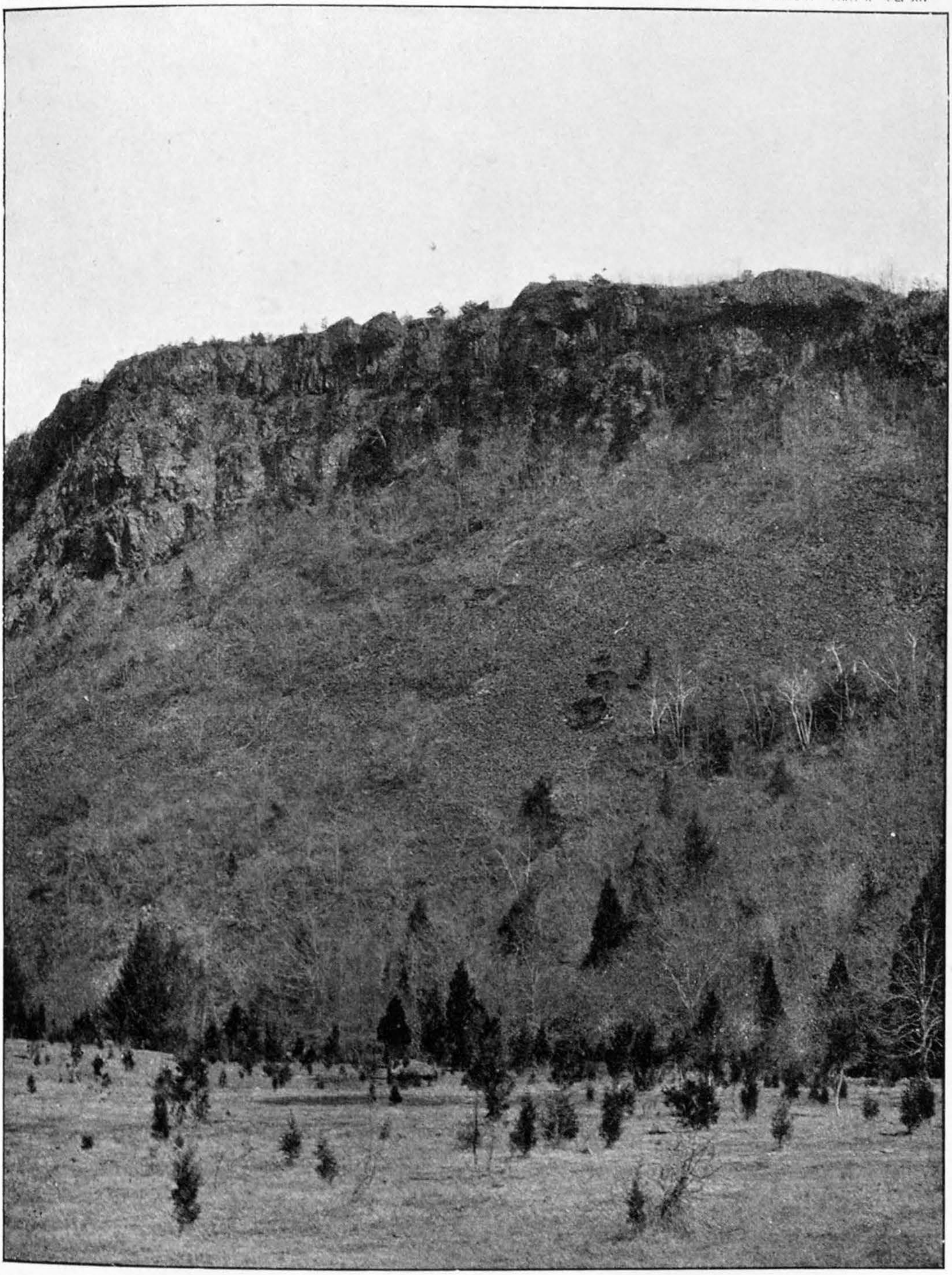


FIG. 39.—Denudation of trap sheets.

miles rather than in feet. It is entirely to long-acting denudation that the consumption of the aerial portion of the reconstructed strata is to be ascribed.

The same is to be said of the intrusive trap sheet and the basal sandstones in the southwest part of the Lamentation block. The sill of Bethany Mountain is, it may be said incontestably, a part of the sill that forms West Rock and Gaylord Mountain ridges. The present form of the ridges can be explained only by the retreat of the sill in Bethany Mountain from the strike line of the sill in Gaylord ridge, and by the retreat of the sill in West Rock ridge from the strike line of Bethany Mountain. These are minimum measures, for the western bluff of Gaylord ridge is as bold as the bluffs in the other ridges. Bearing in mind that all the evidence thus far found points to the intrusion of this great sill while the strata that inclose it still lay essentially horizontal, and remembering that horizontal intrusion must have been about as easy in one direction as another, and, finally, recalling that the most probable source of the sill is in the region of the Mount Carmel dikes, the observer will see that the present ridge crest can not stand in any definite relation to the original limits of the sill,



CLIFF AND TALUS OF HIGBY MOUNTAIN, EAST MERIDEN.

and that here, as in the flows, the ridge crests practically mark the edges down to which the unlimited sill was worn during the long cycle of denudation that produced the Cretaceous peneplain. In view of all this, it seems necessary to call upon extensive denudation to produce the ancient (Cretaceous) topography of the Triassic area, as well as to produce the even uplands of the crystalline areas.

On the nearly even surface of the peneplain all the monoclinal layers of the successive faulted blocks would outcrop in systematic pattern. In the absence of any drift cover at that time the strata would be well defined by their surface soils, and the deciphering of the structure might have been easier then than now. But excepting the few beveled crest lines on the trap ridges, the peneplain is now extinct in the Triassic area. The low-hanging clouds at upland height in winter storms are to-day its most material image above the Valley Lowland. If the invisible faults described in Part II make a strong draft on the imagination, the vanished peneplain that once roofed over the Triassic area is even more difficult to realize. Yet it must once have stretched evenly across the region of the Valley Lowland at the height of the uplands on either side; and all the Valley Lowland, so important in every economic relation, is the work of a later chapter of denudation.

### SUBAERIAL ORIGIN OF THE PENEPLAIN.

About the middle of this century the peneplain of the uplands would have been explained, if considered at all, as the work of marine erosion when the land stood lower than now. During the last quarter of the century an opinion has been gaining ground, especially in this country, that subaerial forces are more competent than marine forces to produce extensive peneplains. It is a difficult matter to make certain choice between the two processes; and such tests as have been devised for the solution of the problem are not applicable to our district in the present state of knowledge. But on tracing the peneplain of the uplands west and southwest, through the Highlands of the Hudson, in New York State, into the Highlands of northern New Jersey, it is there found, by means of arguments based on the arrangement of the streams, that subaerial rather than marine erosion gives best explanation to the facts,[1] and in lack of other argument the same conclusion will be held applicable in Connecticut.

In order to avoid misunderstanding, it is advisable to state that the present shore line of Connecticut does not stand in any definite relation to the shore line of Jurassic peneplanation, and that it is still more remotely connected with the shore line of the pre-Triassic peneplain discussed on page 25. When the upland sky line of to-day is traced southward it may be followed till it gently dips under the

[1] Geographical development of northern New Jersey: Proc. Boston Soc. Nat. Hist., Vol. XXIV, 1889, p. 374. Rivers of northern New Jersey: Nat. Geog. Mag., Vol. II, 1890, pp. 1–30. Plains of marine and subaerial denudation: Bull. Geol. Soc. America, Vol. VII, 1896, pp. 377–398.

waters of the Sound; and, although at certain points there is an appearance of a local and slight increase of slope on approaching the water's edge, it has not yet been shown that this slope marks the southward limit of the peneplain. To all appearances the peneplain of the uplands may continue as an inclined plane many miles into the continental shelf of our Atlantic border, there buried under marine deposits of late Mesozoic and Tertiary date. The shore line of the cycle of peneplanation may have lain many miles off at sea from the existing shore line.

### ADJUSTMENT OF STREAMS TO STRUCTURES.

During the long process of subaerial peneplanation there was excellent opportunity for the development of subsequent streams along the strike of the relatively weak strata and for the diversion of the upper courses of transverse streams by the growing subsequent streams. The great depth of erosion from the initial surface of the faulted monocline to the level of the peneplain, and the strong contrast between the resistant trap sheets and the weak sandstones and shales, would favor rearrangements of stream courses to nearly as extensive a degree as has prevailed on the strong sandstones and the weak shales and limestones of the middle Susquehanna district in Pennsylvania. The outcrops of the trap sheets would generally come to be divides, whatever their relation to initial drainage, and only the stronger streams could maintain their courses across these iron-like rock ribs. The outcrops of the weaker members of the lower and upper sandstones would come to be occupied by extensive longitudinal valleys, and streams running along the strike of these strata must have been the order of the day, as they are in middle Pennsylvania and eastern Tennessee; but such streams could have had no representatives among either the antecedent or the consequent streams of the initial stage of the Jurassic cycle. The greater part of the adjustments of this kind would have been accomplished in the earlier stages of the cycle; the later stages would have witnessed little more than the consummation of the process as the trap divides were slowly forced to retreat down the dip of their sheets.

It should be for these reasons expected that on the essential completion of peneplanation all the area of the lower sandstones should have become tributary to a few large rivers; that all the area of the upper sandstones should likewise have a relatively unified drainage system; that only a few of the stronger transverse streams would cut across the trap ridges by which the upper and lower sandstones are divided; and that only such longitudinal consequent streams as ran along thoroughly shattered fault lines could still maintain their courses through the low notches between the trap outcrops. This arrangement is indicated in fig. 40. The basal sandstones between the crystalline floor and the intrusive sill are hardly thick or continuous enough to determine the development of extensive subsequent streams, but to a certain extent such streams might be expected to follow their strike. No such systematic statement can be made for the uplands until their structure is worked out.

The explanation of the adjusted drainage here given is foreshadowed in one of Dana's essays. He suggested that the drainage of the undisturbed Triassic surface might have been southward to the Sound at New Haven, as represented in fig. 37 (p. 155), but the sandstone was intersected by dikes of trap, and "as erosion went forward, the trap dikes became the trap ridges of the country, standing, with some of the inclosing indurated sandstones, as high barriers between different parts of the sandstone area. * * * These trap barriers naturally determined to a large extent the drainage lines of the area. * * * South of Hartford * * * a succession of ridges * * * crowd toward the river, and, consequently, the Connecticut, after passing Hartford, loses its westing; and then at Middletown * * * turns abruptly out of its old valley through an opening heading southeast that then offered no doubt an unobstructed way to the Sound. Thus the eastern part of the valley lost the Connecticut river. This event in New England history occurred before the Cretaceous period."[1]

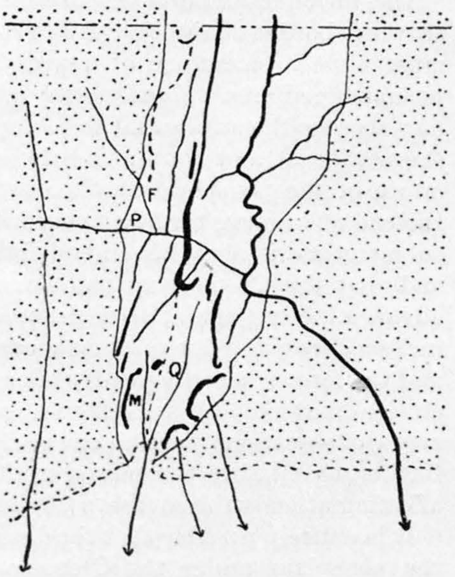


FIG. 40.—Diagram of drainage on the Cretaceous peneplain.

The larger streams may have wandered in somewhat meandering courses on the floor of the peneplain, after the habit of good-sized rivers in their old age. If antecedent or consequent, they might thus depart to one side or the other of the courses which they had held since the beginning of the cycle; if subsequent, they might stray away from the weak stratum under whose guidance they had been developed. It is, therefore, not to be expected that when these streams come to be revived by the elevation of the peneplain to its present upland altitude they should adhere very closely to the courses in which they first learned to run.

## GEOLOGICAL DATE OF THE PENEPLAIN.

The geological date of the general completion of peneplanation is of interest as marking the separation of the two chapters of denudation;

[1] The overflows of the flooded Connecticut: Am. Jour. Sci., 3d series, Vol. X, 1875, p. 500.

but here, again, no definite statement can be given on local evidence. It is true that the Cretaceous strata of Long Island seem to lie, in a general way, on the southward submarine prolongation of the peneplain that descends from the slanting uplands of southern New England; but the intervention of Long Island Sound and the considerable deformation of the Cretaceous beds observed at various points on Long Island render their relation to the peneplain obscure and doubtful. In New Jersey, Pennsylvania, and farther southwest there is better opportunity for determination. There the Potomac and Cretaceous strata manifestly lie on the peneplain in the attitude that is inferred to obtain underneath Long Island; and hence it may reasonably be said that the peneplain was essentially completed before these later Mesozoic strata were deposited; that is, in early Cretaceous or in late Jurassic time.

The unconformable superposition of the Cretaceous strata on the seaward border of the peneplain shows that after peneplanation there must have been a time of depression and partial submergence, the unsubmerged area on the north and northwest furnishing the waste now seen in the strata that were deposited on the submerged area to the southeast and south. By those who do not accept the subaerial origin of the peneplain another interpretation might be given here. Instead of arguing that Cretaceous deposition followed depression and submergence, it might be contended that the sea cut its way into the land and that the Cretaceous strata are the products of its destructive advance. Here, again, no decision can be reached in Connecticut; but in New Jersey the few relevant facts are against the latter supposition, and the former is consequently adopted. We are thus led to the conclusion that not only was there a long pause between the two movements of elevation by which our region gained its present altitude, but, further, that before the second uplift there was a moderate depression affecting at least the southern border of the peneplain.

It is entirely possible that the depression, submergence, and burial of the peneplain under the Cretaceous strata extended much farther inland than the present border of the Cretaceous strata in Long Island, for the erosive forces of post-Cretaceous time are competent to dispose of a great volume of weak strata, such as occur in the later Mesozoic formations on the Atlantic slope. The Cretaceous cover may very probably have reached 20 or 30 miles inland from the present coast line, and even farther than this up the courses of the main rivers. If so, the present arrangement of the rivers in the southern part of Connecticut can be explained by superposition, as was first suggested by Tarr. It is in this possible condition that we encounter the resumption of the process of deposition over a part of the Triassic area referred to on page 144, and the alternative explanation of some of the river courses mentioned on page 156.

## POSSIBLE SUPERPOSITION OF CERTAIN STREAMS.

As far inland as the Cretaceous cover reached, the pre-Cretaceous streams of the peneplain would be more or less completely extinguished. When post-Cretaceous elevation raised the landward part of the Cretaceous peneplain into an upland and converted the adjacent sea floor into a sloping coastal plain, the streams from the upland would be extended across the plain, guided in their new courses by its slope toward the new shore line, as in fig. 41. The usual sequence of processes normally obtaining on an oldland and a coastal plain elevated together from a former lower stand would then take place. As the streams incised their valleys through the coastal plain and discovered

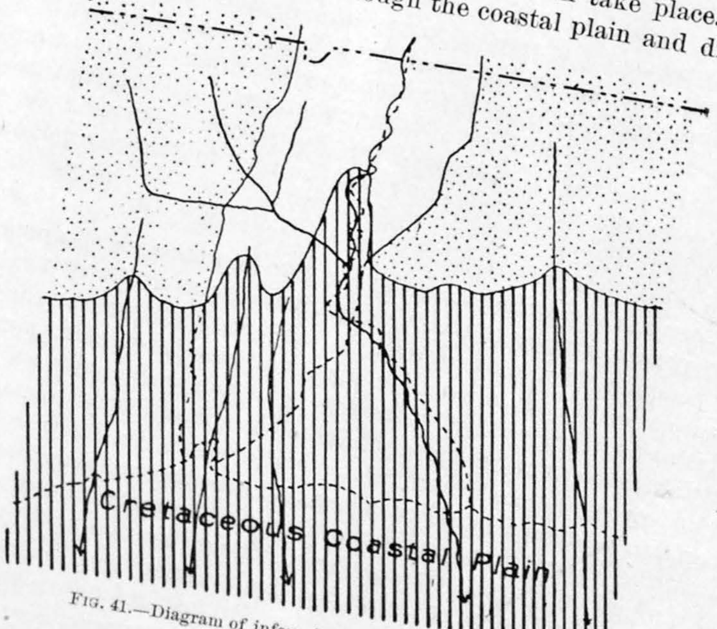



FIG. 41.—Diagram of inferred Cretaceous coastal plain.

the buried part of the oldland beneath, they would flow in paths entirely independent of the oldland structure and indifferent to the stream courses that obtained in the later stages of peneplanation before submergence, and they would thus take on the characteristics of superposed streams. As the covering strata gradually wasted away the stripped belt of the oldland would be characterized by these superposed streams, except in so far as new adjustments to oldland structures might be gained, particularly by the smaller streams. The Connecticut below Middletown may thus be an example of a superposed course, and not of a course consequent upon initial warping, as suggested on page 155. The lower course of the Housatonic may have a similar origin.[1] The same explanation may be given to several small

[1] H. B. Kümmel, Some rivers of Connecticut: Jour. Geol., Vol. I, 1893, pp. 371–393.

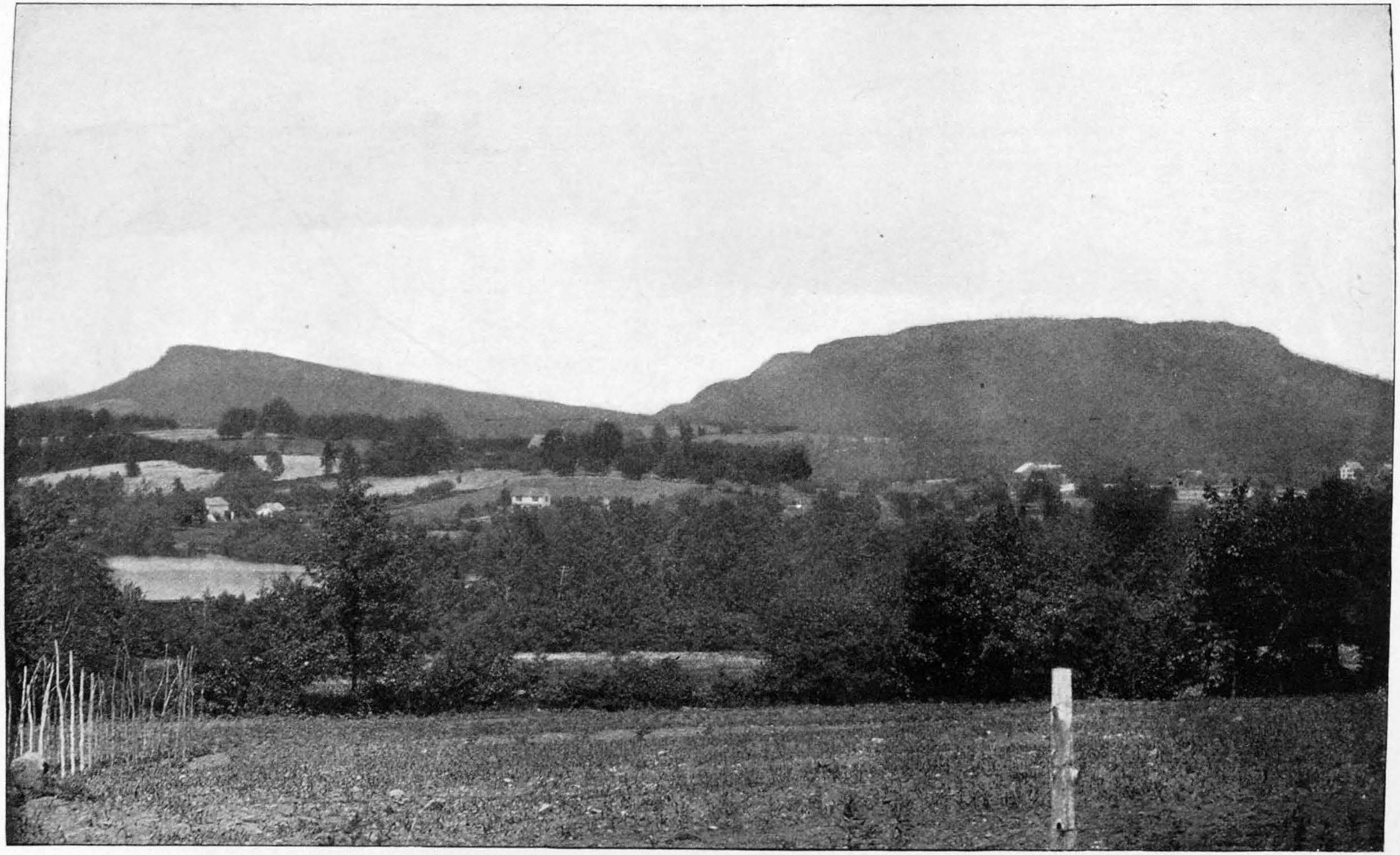
streams which drain the Totoket and Saltonstall crescents southward across the eastern crystallines to the Sound.

It should be noted that the area of the Western Uplands which drains into the Triassic lowland becomes narrower to the south. This is not consistent with the persistence of antecedent (Triassic) drainage, but accords very well with the results of adjustment to belts of weaker crystallines trending about north and south, or with the consequences of superposition. Again, it is significant that within the limits of Connecticut the eastern upland sheds but little water to the Valley Lowland. This is not consistent with the persistence of a Triassic drainage, in which numerous strong streams must have, in all probability, entered the estuary from the east; it is much more consistent with the persistence of consequent (Jurassic) streams, which gained an eastward course on the broken surface of the tilted monocline. With the present small understanding of the upland structures no very definite statement can be made on this aspect of the drainage problem. All that can be done here is to indicate the character of the alternative explanations for some of the river courses, thus blocking out the work for future explanation.

## UNEVEN UPLIFT OF THE PENEPLAIN.

As soon as the rolling uplands are seen to be remnants of a peneplain, either of subaerial or of marine denudation, the uplift of the region since peneplanation becomes evident. A peneplain must at the time of its completion stand close to the sea level, and its streams must have grades so flat that no further trenching of the surface is possible. A peneplain whose surface is now several hundred feet above sea level and is deeply dissected by numerous valleys is manifestly an uplifted peneplain; and such is the case here. Therefore, as has already been stated, the present position of the Triassic formation is not the result of a single initial uplift; it has been gained by two uplifts, and the dates of the uplifts were separated by a period long enough for the base-leveling of the forms initiated by the first before the second took place. Still further subdivision of the total movement may in time be recognized. It is interesting to notice that here, as in many other similar cases, the second uplift is not recognized from geological but from geographical evidence, and that in this way geography is making partial payment of the great debt that it owes to its older sister science.

The greater height of the uplands in the northern part of the State than near the coast, and the greater depth of the chief valleys in the interior than near their mouths, suggest that the peneplain, with the Cretaceous cover of indefinite width along its southern margin, was elevated more in the interior than near the coast, so that it slanted southward. This is confirmed by tracing the uplands and valleys into Massachusetts, at whose northern border the uplands reach altitudes of 1,400 to 1,600 feet, while Deerfield Valley is 1,000 feet deep beneath



LAMENTATION MOUNTAIN AND CHAUNCY PEAK; LOOKING NORTH.

A fault runs northeast between the two mountains, with uplift on the right.

the Western Upland. It is evident from this that the southward slope of the uplands during the excavation of the valleys in the Tertiary cycle of erosion is decidedly greater than the faint slope of the peneplain at the time of its completion, before its flat-graded streams were disturbed by depression or elevation. It is further evident that, although the existing altitude of the land has been affected by the ups and downs of the Glacial period, it nevertheless is not far different from that which obtained for a long preglacial time, the chief difference

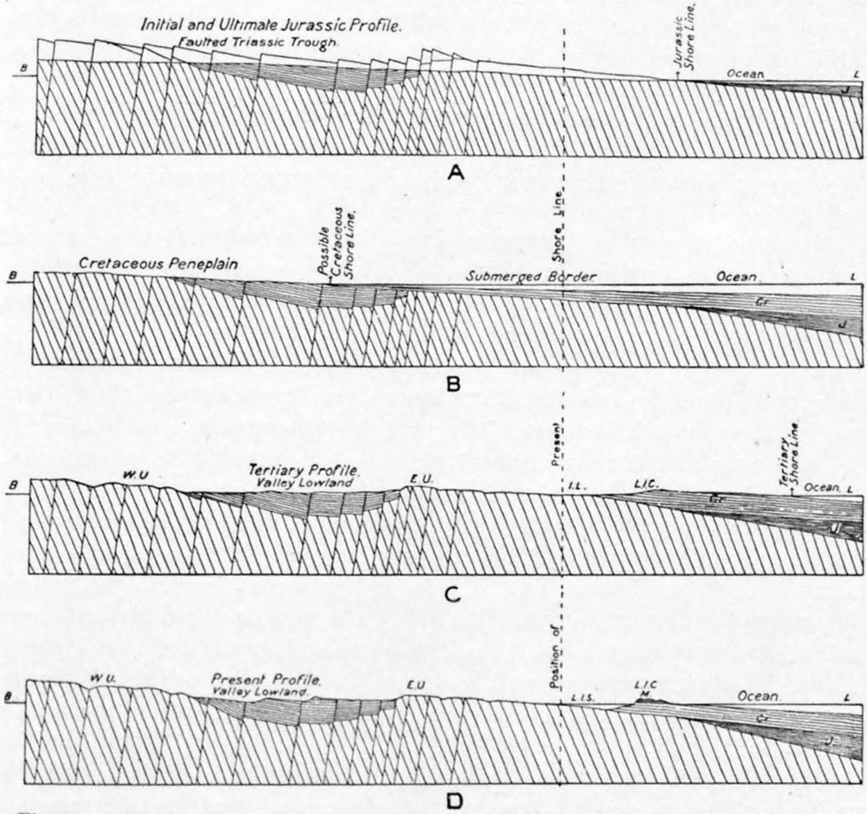



Fig. 42.—*A*, Position of shore line at close of Jurassic denudation; *B*, Position of shore line during Cretaceous submergence; *C*, Position of shore line during Tertiary dissection; *D*, Present position of shore line.

being found in a moderate depression, causing a slight submergence of the lower valley courses, to which further reference is made on page 183.

The accompanying diagrams, fig. 42, illustrate in rough fashion the successive attitudes of the region at certain critical stages in its geographical development. All the profiles are drawn from northwest to southeast, so as to cross the Triassic belt obliquely and extend some distance into the Atlantic of to-day. The vertical scale is necessarily exaggerated. The first profile, *A*, exhibits the initial form of the Jurassic cycle (unshaded) and the peneplain formed at the end of the Jurassic cycle; the shore line is placed some distance south of the shore line of to-day. The second profile, *B*, illustrates the Cretaceous submergence and deposition, with the shore line standing inland. The third, *C*, shows the gently slanting uplift of the peneplain from whose back the Cretaceous cover has been generally stripped almost as far as the inface of the Long Island cuesta, while the Valley Lowland is etched out on the Triassic strata, as explained below. The fourth, *D*, exhibits the summation of the oscillations that are commonly associated with the Glacial period, resulting in the slight depression by which the inner lowland of the Cretaceous coastal plain has been drowned to form Long Island Sound, the cuesta of the island now being surmounted by morainic hills. No attempt is made to show the deformation known to exist in the Cretaceous strata of the cuesta.

## TERTIARY DISSECTION OF THE UPLIFTED PENEPLAIN.

It was in the gently slanting upland that the ordinary processes of land sculpture, under the guidance of the revived streams of the uplifted peneplain, carved out the general features of the existing topography. The resistant crystalline rocks of the uplands have been trenched by valleys of moderate width, while the relatively weak Triassic sandstones and shales have been broadly consumed, thus illustrating the great variations in the rate of topographical development or the length of the geographical cycle in resistant and in weak rocks. The trenches of the uplands are, as a rule, rather steep-sided valleys, one or two miles wide, and in the interior four or five hundred feet deep; but already the greater part of the Triassic area is reduced to a new peneplain of denudation, a peneplain of the second generation.[1] Important as the Tertiary denudation is in determining the distribution of upland and lowland, whereby the distribution of population is so largely controlled, its amount is small compared to that of the Jurassic cycle or compared to the volume of sandstones still remaining below the lowland of to-day.

It is only the heavier sheets of trap that still stand boldly above the valley lowland at present, forming the eastern and western trap ranges, to which reference has been so often made on earlier pages. All these ridges must therefore be regarded simply as the residuals of the second cycle of erosion in the Triassic area, after the general peneplanation of the whole region in the first cycle. In no case does a trap ridge owe its present height above the sandstone lowland to a local uplift; for

[1] When viewed from the summit of West Peak, it is noticeable that the peneplain of the Western Upland is remarkably well preserved west of Southington, and from this justification may be found for the belief that the pre-Triassic denudation of that district, as illustrated in fig. 3, page 21, represents not merely a local, but a widespread denudation, producing a peneplain. For, inasmuch as the present upland thereabouts is still closely accordant with the altitude up to which it was raised at the beginning of the Tertiary cycle of denudation, it follows that the rocks of the district must be among the most resistant of the region, and as such they could not well have been locally worn down to the pre-Triassic floor, unless all their fellows were similarly worn down. See pp. 22, 23.

RESERVOIR IN ANTERIOR VALLEY OF CHAUNCY PEAK.

The talus slope of the main trap sheet descends on the left. The back slope of the anterior ridge appears on the right of the reservoir.

trap and sandstone were uplifted equally. It is entirely to the strong resistance of the trap against weathering that the present eminence of its ridges must be ascribed. In a few places the ridges reach above the level of the adjacent uplands, as in East Rock, certain parts of the West Rock ridge, Mount Carmel, and West Peak, and these eminences of to-day must have formed low mounds above the peneplain of early Cretaceous time. The even crest lines of Pond and Totoket mountains and of Talcott Mountain farther north are so near to or so little above the level of the adjoining uplands that they may be taken as remnants of the peneplain, as yet little affected by Tertiary erosion. Elsewhere the intrusive sill and the main sheet of the extrusive flows seem to be worn somewhat lower than the old peneplain surface, and the anterior and posterior flows are always worn down to ridges of moderate height. Within the limits of a single block each member of the series gives an appropriate relief where it is not worn down so low as to be smothered under the drift. The surface thus comes to have a distinctly monoclinal expression, the steeper bluffs of the higher ridges facing westward with bold outcropping cliff of ungraded rock and rocky talus of graded waste, while the back slope falls off gently eastward. This is manifest not only in the strong ridges of the main extrusive sheet and in the more evenly interbedded parts of the intrusive sill, but also in the subordinate ridges of the anterior and posterior lava beds, and, in a small way, in numerous ledges formed on relatively resistant sandstone and conglomerate layers; but, on the other hand, it often happens that the alternations of stronger and weaker sandstones in the valley lowland are so gradual that ungraded ledges are rare and monoclinal expression seldom appears in their rounded swells.

## EFFECT OF FAULTS ON THE TRAP RIDGES.

The effects of the fault system on the form of the trap ridges may be treated under four headings: First, in defining their termination; second, in controlling their terminal form; third, in causing occasional cliffs and talus slopes on the eastern as well as on the western slopes; fourth, in sometimes reducing the usual height of the main sheet.

The trap ridges necessarily end where the trap sheets are cut off; hence, a ridge is never continued across the fault that bounds its block. This has been abundantly illustrated in the descriptions of the blocks themselves. At first meaningless or mysterious, it comes to be regarded as one of the most impressive features of the region, nowhere better shown than in the Lamentation block, where the heavy main sheet is so far out of line from its neighbors. The same rule, of course, applies to the minor trap sheets, and to the little conglomerate ridges as well; but the drumlins, eskers, and sand plains cross the fault lines indifferently, much to the annoyance of the observer who is trying to follow the path of a displacement.

Where the displacement of a fault is small, so that the dislocated parts of the main sheet are not altogether thrown out of line, it might be expected that the ridge formed upon them would cross the fault with little interruption; but such is not the case unless the fault is of insignificant movement. It is surprising to learn from the depth and breadth of the notches, opened obliquely across the main sheet on the line of small faults, how greatly the trap is weakened on the fracture planes. For example, the valley of Shuttle Meadow, southwest of New Britain, is prolonged as a well-defined oblique notch between the two masses of the main sheet in Ragged and Bradley (Farmington) blocks. So between South Mountain and Cat-hole Peaks, or between Lamentation Mountain and Chauncy Peak, the fault fracture suffices without the aid of weak underlying shales to determine well-defined oblique ravines from one side of the trap ridge to the other. Reservoirs for water supply are often constructed in these ravines. The ravines can not be ascribed to local weakness or irregularity in the original structure of the trap sheet, for the main flow bears every indication of general uniformity over large distances. In the same way the slight reentrant angles, often found in the cliffs of the main ridges, may fairly be associated with fault lines, as on the south bluffs of West Peak, South Mountain, and elsewhere.

The same is believed to be true of the reentrants in the face of the intrusive sill in the western ridges. Wintergreen Notch, for example, a mile north of the south end of West Rock ridge, is not cut square to the trend of the ridge, but enters somewhat obliquely, thus falling into general accord with the northeast-southwest fault system; and the crest lines of the ridge on either side of the notch are a little displaced in advancing order, as Dana has shown. Here form and structure are regarded as mutually explanatory; the notch reveals a small fault, and the fault determines the notch. It is therefore concluded that the narrow belts of breccia in the smaller faults, and the increase in the number of joint planes near the fissures are competent to determine the location of reentrants, notches, and gullies; and that the large faults are not clean-cut fissures, but are rather belts of shattered rock, accompanied by minor fractures, thus determining the erosion of oblique ravines and valleys for a certain distance on either side. This is confirmed by the observed structures in the Quarry Ridge, north of Meriden. Here the trap has been much fractured and brecciated at the base of the ridge near the little valley that follows the great fault, and several minor belts of breccia are noted in the quarry (p. 109). It is probable that similar structures occur elsewhere, although the opportunity of observing them is seldom offered.

The well-defined effects of the larger faults on the larger topographic features warrants the inference that smaller faults would produce similar but smaller effects. Reversing this rule, it may be safely inferred that the little notches and oblique gullies, by which the front and back slope of the main trap ridge is often indented, mark smaller faults

WEST PEAK, HANGING HILLS, AND ITS ANTERIOR RIDGE; LOOKING SOUTHEAST.

wherever they occur. The fact that the gullies do not run down the back slope in the direction of its dip shows that they are not to be explained by ordinary erosion on a massive structure. The occurrence of faults, inferred from the notches and gullies in the trap ridge, as already considered under the description of the blocks into which the formation is divided (p. 136), is thus justified.

If the bounding fault of a block were squarely transverse to the strike of the monocline, the ends of a trap ridge would be symmetrically rounded off as the general denudation of the surrounding sandstones advanced, as in fig. 43. No such ridge is known in the valley. But if the faults run obliquely to the strike of the strata, the ridge

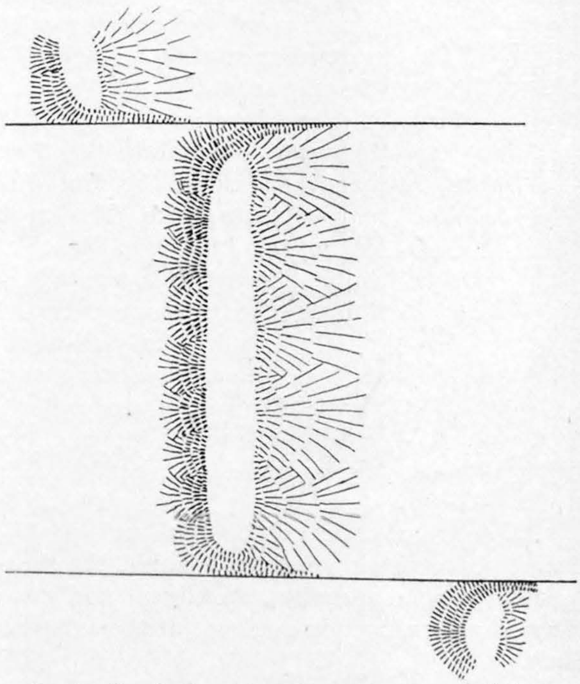


FIG. 43.—Termination of trap ridges at transverse fault lines.

endings would be unsymmetrically denuded, a slowly descending trail of trap, GJF, fig. 44, being developed at the obtuse angle, GEF, between strike and fault line, and a sharply rounded bluff, HKC, at the acute angle, HBC. This contrast of form has been repeatedly mentioned in describing the various blocks. As far as opportunity is given in the small faults of retreating displacement north of Talcott Mountain, the sharply rounded bluffs face northward, and the slowly descending trails point southward, this being the reverse of the arrangement occurring so distinctly in the district south of Hartford, where the advancing displacement prevails. West and East Talcott mountains are alone in possessing sharply rounded bluffs at both ends, and these are the only ridges that have the acute angle between strike and fault at both ends, on account of being cut by curving faults (see pp. 115, 117).

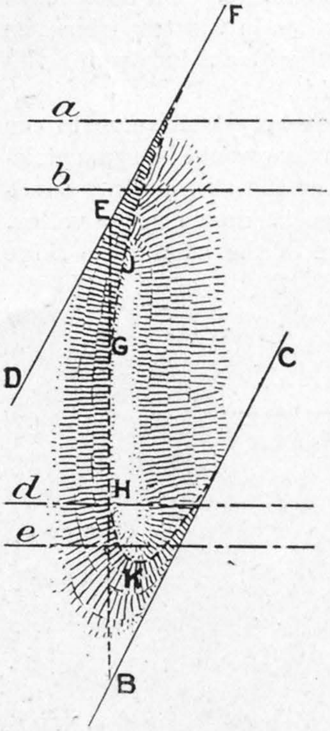



FIG. 44.—Termination of trap ridge at oblique fault lines.

The cliff and talus slopes normally occur on the outcrop or western side of a trap ridge, but it sometimes happens that similar forms are determined by the basset edge on the eastern side of a trap ridge, where it is retreating from a fault near its eastern base, as in the case here figured. The opportunity for such a profile is rare, because it depends on the fortuitous combination of three independent lines in particular relation, namely, the dip line of the trap sheet, TT; the fault line, FF; and the base-level of to-day, VL, fig. 45; and in order to maintain this form along the back of a ridge, the fault line must run parallel to its strike. The only conspicuous case of this kind is in West Talcott Mountain. The entire eastern side of this mountain is of exceptional form. It does not descend, as do nearly all the other ridges in the valley, by dipping under the posterior shales on a moderate slope, but instead it plunges down in cliff and talus of moderate height, but of as pronounced form as any in the valley. Although so exceptional, it is entirely explained by faulting (p. 117). Similar forms, but of much less extent, are found behind the sharply rounded bluffs at the acute angle end of many trap ridges; and conversely these have been many times employed in evidence of the existence and direction of a fault. Good examples are found in the cliffs of Bradley Mountain overlooking Shuttle Meadow Notch; in the cliffs and talus of South (Notch) Mountain, descending into Cat-hole Notch; in the cliffs at several appropriate positions in the Bethany Mountain district, and at many other points, as illustrated in sections *d* or *e*, fig. 44.

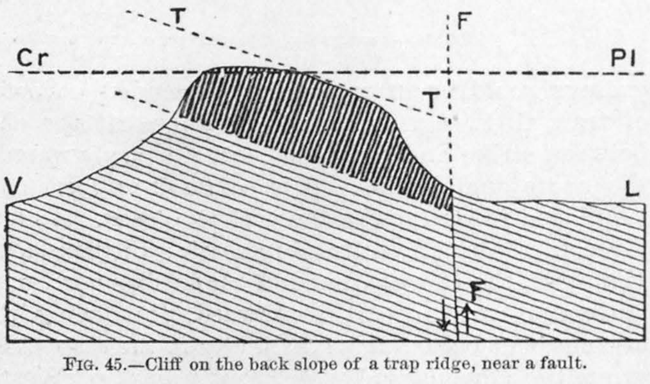
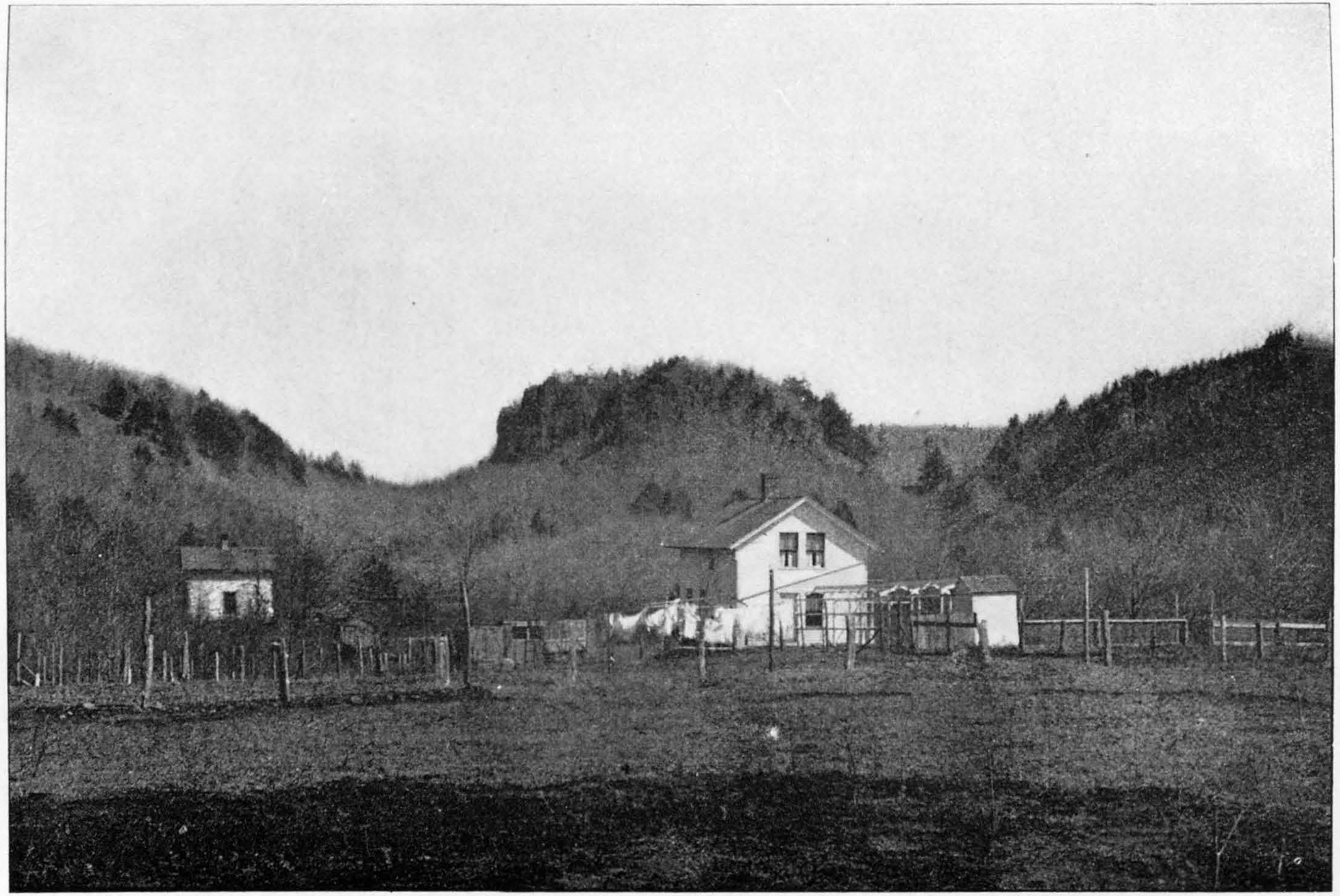



FIG. 45.—Cliff on the back slope of a trap ridge, near a fault.



NOTCHES WORN ON FAULT LINES, CAT-HOLE MOUNTAIN

Faults may determine a reduction in height of the main trap sheet ridge in two ways, which may be best appreciated in contrast to the normal ridge form, fig. 39. There no fault occurs for some distance on the outcrop side of the normal ridge face. But if, as in fig. 46, the fault cuts the trap sheet so as to rob it of some of the normal ridge face, then the remaining ridge must be lower than usual. This is well exemplified in East Talcott Mountain, which, as a rule, fails to reach the height of West Talcott Mountain; and better still in Cedar Mountain, which is much lower than the ridges of the main sheet in its neighborhood. The same thing is seen in the long Cat-hole Ridge, where an abnormal warping of the main sheet brings its strike for a mile or two about parallel to the fault that is inferred to run along its western base. These are all analogous to the loss of height in the slowly descending trail of the ordinary ridges, as in sections *a*, *b*, fig. 44.

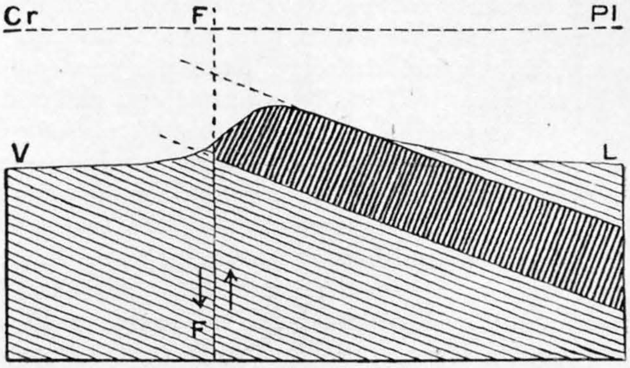


FIG. 46.—Loss of height of trap ridge near a fault.

The second way in which faults may decrease the height of a trap ridge is by occurring close together, so as to cut narrow slabs, instead of broad blocks, between them. Here the curves of the slowly descending trail, *c b a*, fig. 47, and the sharply rounded bluff, *f b d*, meet at *b* in the middle of the slab, and prevent the ascent of the trap sheet to the usual height. Examples of this kind appear in Short Mountain, and again in the Cat-hole and Quarry submembers of the Hanging Hills blocks; and in the latter the faults seem to be so numerous and so close together that the anterior and posterior trap sheets are weathered down low enough to be concealed by drift. It is by this excessive retreat of the main sheet in these two narrow slabs that the great turn in the cliff face of the Hanging Hills is in part accomplished (see p. 110). The aid of a transverse fault, invoked by some observers to account for this turn, does not seem to be necessary, nor is there any independent indication of its occurrence.

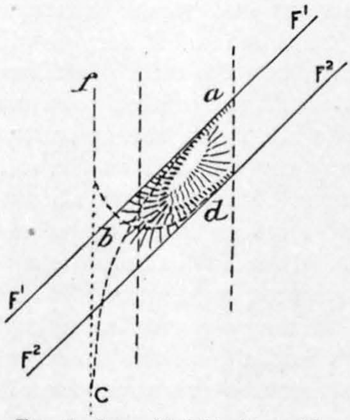


FIG. 47.—Loss of height of trap ridge between two faults.

## GEOLOGICAL AND GEOGRAPHICAL EFFECTS OF FAULTS.

Several of the preceding sections give good illustration of the difference between the geological and geographical effects of faults, to which sufficient attention is not always paid. Geologically, a fault is a displacement of one mass with respect to another; the value of the displacement remains fixed unless some later disturbance changes it. Hence, it may be said, once a fault, always a fault, to the geologist. Geographically, a fault dislocates parts of a previously continuous surface. The two parts are initially separated by a cliff of displacement; but as denudation progresses on the dislocated mass the cliff face is dissected by ravines and forced to retreat until it vanishes, and the heaved area is worn down lower and lower until, when the peneplain stage is reached, no difference of height remains on the two sides of the fault lines. The fault has vanished, as far as topographic relief is concerned, and appears only as a sudden change of soils when the fault line is passed. During this cycle of denudation the geographical treatment of the fault requires the consideration of four factors: the general structure of the region, the form that the region had before faulting, the amount of displacement along the fault line, and the time passed or amount of denudation accomplished since faulting. The frequent omission of some of these factors in published accounts of faulted regions makes it difficult to interpret their topography.

A further complication ensues if a regional uplift takes place after peneplanation, for then the fault may again come to have much topographical importance, provided that resistant and weak structures have been brought into juxtaposition on either side of it, for one will waste slower than the other, and thus gain a relative relief; but here the inequality of altitude has nothing to do with the original measure of dislocation on the fault plane; it depends simply on the amount of regional elevation by which denudation was revived and on the contrast of resistance in the adjoining structures. Moreover, the abruptness in the form of the revived fault bluff gives no indication of the date of faulting, as it did in the first cycle. It indicates only the contrast of resistances in the adjoining structures as denuded in the second cycle. Finally, a revived fault bluff may separate an upland on the downthrown side of a fault from a lowland on the upheaved side of the fault, if resistant and weak structures are appropriately arranged; thus giving an appearance of reversed movement. During the second cycle the geographical treatment of the fault should therefore include, besides the general structure of the region, a statement of the form reached when the preceding cycle was interrupted by uplift or other disturbance, the character and the amount of the uplift or disturbance, the structure on the two sides of the fault, and the time passed or amount of denudation accomplished since then.

## FURTHER ADJUSTMENT OF STREAMS TO STRUCTURES.

The system of adjusted streams, as deduced on page 162, would still prevail on that part of the peneplain never covered by Cretaceous strata; and in the Tertiary cycle of erosion the adjustments would be maintained or extended. To-day, in the late maturity of this cycle (as far as the Triassic area is concerned), the trap ranges might be expected to form a more perfect divide between the river systems of the lower and upper sandstones than they did at the beginning of the cycle.

In the southern district, where the streams might have been superposed from the Cretaceous cover, discordances with structure might still occur to-day, in so far as they have not been lost through new adjustments. But it is to be noticed that the general southerly course of the trap ridges is about parallel to the expected course of the superposed streams, and hence that few straddling streams need be looked for from this cause. It is believed that the main valley of the Quinnipiac from Meriden southward and the Mill River Valley from some distance north of its present head are either of purely subsequent adjusted origin or are the valleys of superposed streams somewhat modified by Tertiary adjustment. The Quinnipiac drains the upper division of the lower sandstones; Mill River drains the lower division. Only in crossing the Mount Carmel and Mill Rock dikes does the latter river work to disadvantage. Between the two rivers rises the long, round Quinnipiac Ridge of sandstone, of which some account was given in describing the Higby, Lamentation, and Hanging Hills blocks (pp. 99, 102, 110).

Any streams that crossed the main trap range independent of fault lines at the beginning of the Tertiary cycle but were diverted to other courses before the present stage of the cycle should leave a record of their transverse paths in the form of wind gaps, or of large water gaps now followed by disproportionately small streams. In this respect the adjustments of the Tertiary cycle are more discoverable than those of the Jurassic cycle, in which the wind gaps were obliterated in the peneplanation of the ridges. It must be for the latter reason that the unfaulted crests of Pond and Totoket mountains are so even now. There is every probability that some consequent streams must have drained into the warped crescents of these mountains in the earlier stages of the Jurassic cycle, and such streams may have then cut deep notches in the main sheet before they were beheaded by a subsequent stream that must have grown northward on the lower sandstones. No record of this process of Jurassic adjustment can be brought down to us in the form of the ridge crests, for they were practically base-leveled after the notches were cut; the record is to be found only in the arrangement of the streams. If it were not for the chances of superposition, the ancestor of the Quinnipiac in the Jurassic cycle might with much confidence be regarded as the subsequent stream that was chiefly responsible for the beheading of the consequent streams of the Totoket and Pond Mountain warps.

The adjustment of the Tertiary cycle may be associated with a prevailing northward migration of divides, on account of the greater northern uplift of the peneplain, in accordance with the principles of river adjustment announced by Campbell.[1] The large size attained by Mill River when, as described below, it probably drained all the western part of the Valley Lowland from the Massachusetts line to the Sound is thus accounted for. The stream captures by which this remarkable growth was accomplished involve the description of the two wind gaps here following.

Cooks Gap is a deep and broad cut, square across the main trap ridge west of New Britain, and to all appearance independent of any fault line. The anterior trap sheet is deeply worn in its path. The main trap stands up in full strength on either side, and at some points is seen on the floor of the cut. The mountains on the south and north, Bradley and Farmington, are therefore to be regarded as transected parts of the main sheet in a single block, and were thus described on page 113. No other example so distinct as this is to be found in the State, and no cause can be assigned for such a gap but a large transverse stream formerly running through it. On extending the course of the valley eastward it seems to turn somewhat to the right after passing the Shuttle Meadow fault line, and it then cuts across the northeastern trail of the main trap sheet in the Ragged Mountain block, for here again the broadly open depression is inclosed by trap on either side, although not so distinctly as in Cooks Gap. It is still entirely independent of any known fault line. The little stream now draining the eastern part of the depression runs eastward down the Sebethe to the Connecticut above Middletown. It is therefore concluded that a considerable river, draining a large area of the western crystalline upland and of the lower sandstones, persisted in this transverse course until the present cycle was well entered upon, and that it was then diverted southward by the lower-lying headwater of some stream flowing southward along sandstones only to New Haven Harbor. The most plausible origin for such a transverse river is that it was an antecedent or a consequent stream whose volume had been much increased by the capture of the headwaters of adjacent streams west of the trap range during the Jurassic cycle, but this is admittedly pushing inquiry far into hypothesis.

The Reservoir Notch in the Hanging Hills is peculiar in departing from the fault line by which it was at first located. The fault line bears N. 40° E.; the notch, N. 10° E. The notch and the fault coincide on the line of the south-facing cliffs of the Hanging Hills, but diverge steadily northward until they are nearly half a mile apart at the base of the back slope of the main sheet. The trap bordering the northern

[1] Drainage modifications and their interpretation: Jour. Geol., Vol. IV, 1896, pp. 567-587, 657-678.

end of the reservoir on the east side therefore belongs in the West Peak block, not in the South Mountain block. The fault line here lies farther east than the notch, and is traceable by a bluff of trap where the main sheet in South Mountain is lifted higher than the same sheet in the West Peak block.

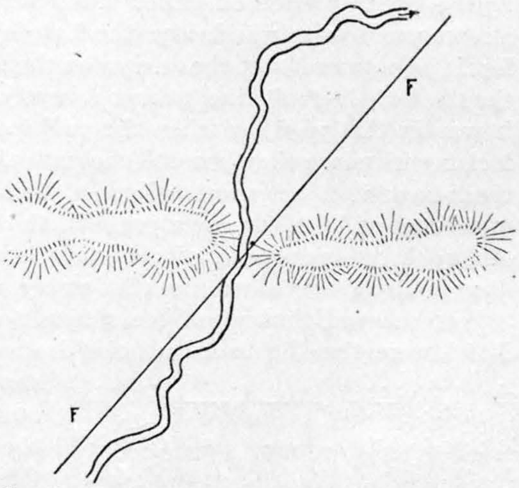


FIG. 48.—Stream wandering from fault line on Cretaceous peneplain.

As in the previous case, no adequate explanation for such a notch is found except in the erosive power of a considerable transverse stream whose headwaters have been diverted to some new drainage system. Here the most plausible origin for the stream would make it consequent on the fault line, initially flowing northeastward in the direction in which the trough line between the adjoining faulted blocks descended. The loss of the last of its headwaters south of the trap sheet must have occurred when its channel through the Hanging Hills main ridge had been cut down to within 300 feet of present sea level, and hence must be ascribed to the Tertiary cycle; but it is highly probable that the initial drainage area of the stream had been much reduced in the Jurassic cycle. The local departure of the notch from the fault line may be thus explained:

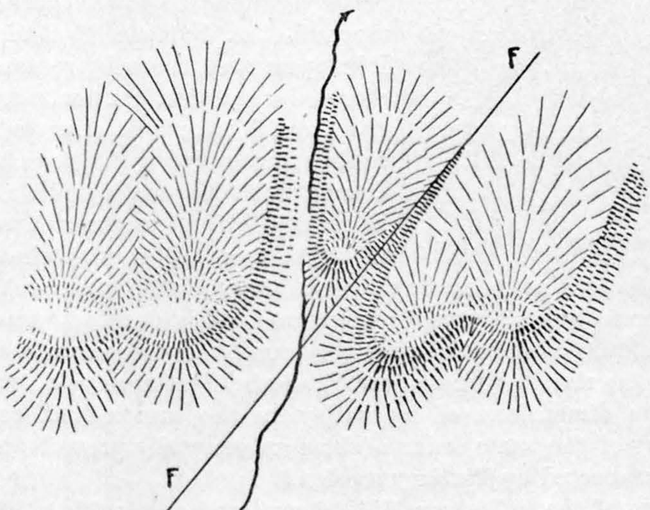


FIG. 49.—Former water gap, now wind gap, cut in trap ridge to one side of fault line.

On the Cretaceous peneplain the trap sheet was reduced to a low ridge, as in fig. 48, of much less breadth than is to-day occupied by the bluff, crest, and back slope of the main sheet. The stream at that time crossed the low ridge at the fault line, and was there held in place by the low trap walls on either side; but on the south and north the stream was free to wander on the base-leveled plain of weaker rocks. Let it be assumed that the stream on the north of the gap wandered to the west of the fault line, taking a nearly northward course, and that it was in this position when the uplift of the region occurred. On incising its new valley beneath the path that it happened to occupy at the time of uplift, it necessarily lost the aid of the fault line for much of its way and had to cut a gorge in the solid trap. Being thus seriously delayed in deepening its course, its headwaters were captured and diverted by the reenforced south-flowing streams, as in fig. 49.

It is believed that when the streams of Cooks Gap and Reservoir Notch were thus beheaded all the drainage of the lower sandstones ran southward to the Sound at New Haven, and that only the upper sandstones (within this State) drained into the Connecticut River. This condition is illustrated in fig. 50.

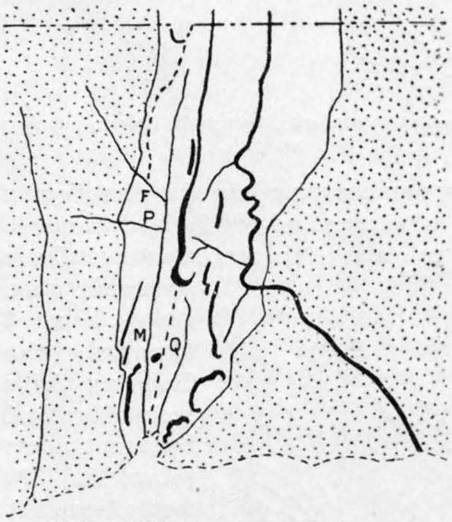


FIG. 50.—Diagram of preglacial drainage.

THE LOWER CONNECTICUT.

In whatever way the Connecticut gained its lower course through the Eastern Upland, it has persisted there by reason of its large size. A smaller river would have been diverted to the much easier path between Lamentation Mountain and the Hanging Hills, or perhaps through one of the deep notches on either side of Paug Mountain; but there is no indication that such a course was ever followed by the Connecticut in this or in any previous cycle of erosion. Certain it is that the present lower course has been followed by the river at least through all the Tertiary cycle, for only thus can the dimensions of its valley in the crystallines be explained, it being of the same order of form as generally prevails in the uplands now dissected by Tertiary erosion.

Standing on one of the hills near Middletown, and overlooking the river to the north and east, one may see on the left the broad open Valley Lowland, and on the right the narrow valley entering the eastern upland. Here is illustrated one of the most important principles of land sculpture—the relation of an inner lowland to the deep and narrow transverse valley by which it is drained through the inclosing upland, a principle concerning which there is a very general popular misapprehension. It is nearly always tacitly assumed that the upland

and lowland had their present forms before the outlet valley was cut down, and that the outlet valley is the result of the overflow of a lake that once occupied the lowland, or of a "convulsion of nature" that disrupted the upland. There is no evidence whatever in favor of these suppositions, and there is full array of evidence against them. The only explanation permissible for an outlet valley of this kind makes the inner lowland and the outlet valley of the same date. The deepening of the inner lowland proceeds at the same rate as the trenching of the narrow outlet valley, but while the latter still remains relatively narrow, because it is incised in resistant rocks, the former is already broadly opened, because its region is occupied by weak rocks. The marvelous consideration here is that every grain of the upper sandstones, and of such portion of the lower sandstones as were for a part of the Tertiary cycle drained by the streams of Cooks Gap and Reservoir Notch, was carried away through the narrow outlet valley of the Connecticut. Although impressive, this arrangement is not rare. The relation of inner lowland, inclosing upland, and transverse river outlet is repeated many times on our Atlantic slope, the finest example being seen along the course of the Hudson River.

## GLACIAL MODIFICATIONS OF FORM AND DRAINAGE.

No critical measure has been made of the amount of topographic change produced in our district by glacial action. In New Jersey, where many structural features of crystalline upland, sandstone lowland, and trap ridges are repeated, there is no striking difference of form inside and outside of the terminal moraine of the later glacial invasion which crosses that district transversely, and from this it may be fairly argued that glacial action caused no great modification of the forms produced in Connecticut by the erosion of the Tertiary cycle. Although necessarily indefinite, this is a great advance from that early stage of inquiry when no dates were assignable for the origin of the general upland and lowland forms, and when it was thought admissible to attribute most of the excavation of the valley lowland to glacial erosion. Although glacial action extended southward beyond the present shore line of Connecticut, there are no large topographic forms that must be necessarily ascribed to glacial erosion. Yet it must be recognized that ledges were somewhat worn down and that weak valley floors were somewhat deepened by the ice sheet. It may be that the oblique departure of Quinnipiac Ridge from the middle horizon of the lower sandstones (p. 110) is due to glacial gouging on either side, while the north end of the ridge is a trail left in the lee of West Peak.

In New Jersey and Pennsylvania, beyond the limits of glaciation, the cliffs are dull and the slopes beneath them are of gentle declivity, generally covered with trees. In Connecticut many of the trap cliffs are very steep and the talus slopes are often bare and treeless. It seems as if the trap ridges had resisted the destructive pressure of the ice, which therefore made reprisal on the weak underlying strata, dragging away the preglacial talus and scraping out the shales to so steep a slope that, when the ice melted away, a fresh supply of talus was quickly furnished from the unduly steep cliff above. To-day the talus just below an occasional tree is of a different color from that near by, this being due to a greater growth of lichens below the trunk that stems the slow flood. Hence, it may be inferred that the talus is, as a whole, in relatively rapid motion, having as yet hardly reestablished a graded slope.

The most significant changes that may be with certainty associated with glacial action are all constructive, in the form of broad sheets of till veneering the lowland surface, thicker deposits of till producing drumlins, kames, and sand plains formed near the ice margin during its retreat, and extensive valley deposits of clays and sands that are commonly associated with a lower stand of the land in the later stages of the Glacial period. All of these deposits are of subordinate value when compared to the trap ridges and to the valleys of the uplands.

Glacial till is widespread over lowland and upland, but calls for no mention here except where it forms large drumlins, such as occur in and south of New Britain, southeast of Meriden, and around Durham. Many of these are typical examples of this interesting style of glacial action. As already mentioned, these are the only considerable eminences of the Valley Lowland that pay no attention to the fault lines. Strong kames and associated sand plains occur in the depression between Lamentation Mountain and the Hanging Hills, making it somewhat uncertain whether the existing divide that here separates northward and southward streams had the same position as now in preglacial time. Temporary stream or river channels were taken in the later stages of the Glacial period. One of these is followed by the Consolidated Railroad just north of Berlin Junction. Others are indicated by the relation of terraces and divides, as further described in connection with Mill River, below. Clays and sands cover a large area of the lowlands, as east and west of Windsor and Hartford, north and south of Farmington, and about Wallingford. Here it is eminently possible that a stripping of the drift would reveal shallow and broad basins eroded in the underlying strata, and that the existing stream courses depart rather widely from those of preglacial time. Thus the Scantic, Podunk, and Hockanum rivers, branches of the Connecticut on the east above Hartford, seldom reveal the rocky floor. They present many features of young streams with fertile flood plains incising their valleys beneath the infertile drift plain. The flood plain of the Connecticut is a meadow of great agricultural value, although in danger from overflow. Stony Brook, draining the town of Suffield, cuts down into sandstones for much of its course, being manifestly superposed upon them through the drift cover. None of these cases, however, involve serious departures from the general adjustment of drainage systems to lower and upper sandstones that is inferred to have existed before the glacial accident. The following exceptional examples are therefore worthy of special attention.

### THE FARMINGTON AND QUINNIPIAC RIVERS.

It is believed that these two streams, F and Q, fig. 51, depart distinctly from the adjustments of their preglacial courses, as has been suggested by Kümmel.[1] The Farmington and the Pequabuck, issuing from the western upland, probably represent the upper waters of the river that cut Cooks Gap. By spontaneous adjustment late in the Tertiary cycle their waters appear to have been diverted near Plainville by a stream that carried them southward through Mill River Valley to New Haven Harbor. Now, the Pequabuck, issuing from the uplands as before and flowing directly toward Cooks Gap for part of the distance across the lower sandstones, turns north at Plainville, and soon meeting the Farmington the united waters flow northward nearly 20 miles on a very flat grade before turning eastward and crossing through the trap range at Tariffville. There is no reasonable explanation for this singular behavior by antecedent, consequent, or adjusted origin. Diversion by glacial drift gives the only rational cause that has been yet suggested for the abnormal northward course, and to appreciate the suggestions made by Kümmel in this direction, it is desirable to examine first the notch in the main trap ridge at Tariffville, and then the divide south of Plainville, by which the Pequabuck is separated to-day from the south-flowing streams.

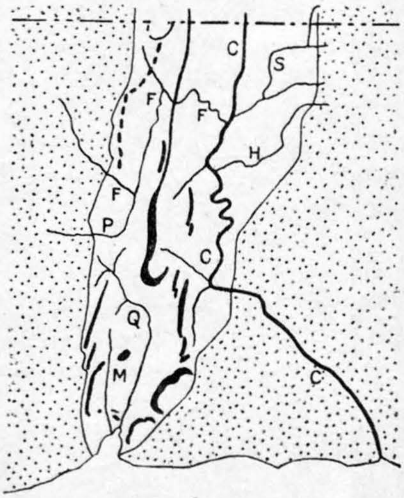


FIG. 51.—Diagram of existing drainage.

The Tariffville Gap is a narrow gorge incised in a broad sag of the ridge crest. The sag is explained by Kümmel as the remnant of the cut of a transverse stream from west to east in the early part of the Tertiary cycle; its association with and departure from the Tariffville fault line, separating Simsbury and Granby mountains, recall the case of Reservoir Notch, as explained on page 177. The floor of the sag, before the gorge was cut in, is estimated to have stood at an elevation of about 190 feet above present sea level. The gorge is so narrow and steep sided that no road follows the stream bank through it, but road and railway are perched well up on its slope in rock-cut niches. The gorge

[1] Some rivers of Connecticut, by H. B. Kümmel: Jour. Geol. Vol. I, 1893, pp. 371-393.

is from 80 to 100 feet deep and is free from drift. There can be little doubt that the gorge is a postglacial stream cut, the work of waters that were turned across the sag as the lowest available escape after their former paths had been somewhere blockaded with drift.

The most important part of the inferred drift blockade seems to be the divide below Plainville, at a level essentially equal to that of the estimated floor of the Tariffville sag in the trap ridge. The terraces thereabouts are somewhat higher than the sag, rising at least 200 feet above tide. It is therefore concluded that the preglacial southward course of the Farmington-Pequabuck along the strike of the lower sandstones was obstructed by drift below Plainville, and that the waters to the north were ponded high enough to overflow at Tariffville, cutting the gorge in the sag of the trap ridge, and continuing eastward to the Connecticut. It is thought that a large part of this barrier may have been the delta-like deposits of the rivers from the upland. The likeness of this case to that of the Deerfield River in Massachusetts has been pointed out by Kümmel.

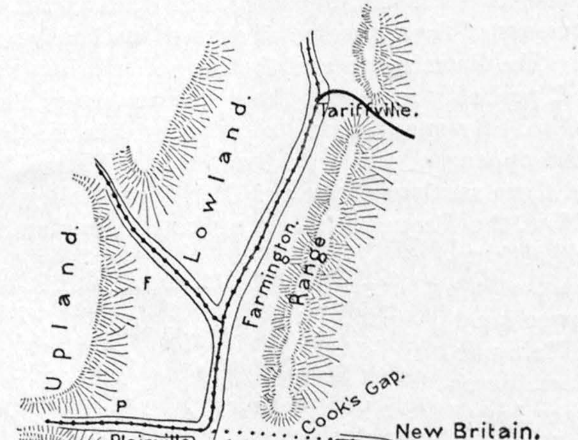



FIG. 52.—Diagram of Quinnipiac and Mill River headwaters.

The Quinnipiac (Q, fig. 52) now receives its headwaters from the ponds in the western part of Cooks Gap. It should, as a subsequent stream, flow southward through Cheshire village to Mill River, and thence to New Haven Bay, such being the probable course of the chief preglacial river of this district, then including the Pequabuck and Farmington rivers in its headwaters, as in fig. 50. The eastern boundary of this subsequent valley rises 100 or 200 feet above its floor in

Quinnipiac Ridge, a long, rounded swell of sandstone, capped with drift, extending southward from the angle of the Hanging Hills, past the Mount Carmel group of dikes, to the city of New Haven. This ridge is a continuous divide, except where crossed by the Quinnipiac in a narrow valley, 100 feet deep, at South Meriden. The sag in the ridge in which this transverse gorge was started has a height of 200 feet; the terraces on the divide between the Quinnipiac and Mill rivers, in Cheshire village, are a little above this level. It is therefore concluded that here, as below Plainville, the preglacial water way was divided by a drift barrier, a single longitudinal subsequent stream of preglacial time being now divided into three, the northernmost part flowing northward and crossing from the lower to the upper sandstones through the gap at Tariffville, the middle part escaping across Quinnipiac Ridge at South Meriden, and only the doubly beheaded trunk stream continuing in its former course directly to New Haven Harbor.

Much of this explanation accords with the interpretation given by Dana. He describes the terraces of the lower (western) sandstone valley from Massachusetts far southward, their grade being very gentle north of the point where the Pequabuck and Farmington emerge from the western upland, but relatively steep farther south; their composition being of fine deposits where they are flat, but of "cobblestone coarseness" where the grade is steep. A continuous discharge of glacial waters is thus argued along the western part of the lowland. As the Quinnipiac below Meriden has low-level terraces of fine materials, it is inferred that the gorge by which it comes through the sandstone ridge on the west was obstructed by ice, or perhaps not then formed.[1]

## GLACIAL AND RECENT OSCILLATIONS OF LEVEL.

The drowning of many valleys in their lower courses along the Connecticut shore line, as well as the submergence of the lowland inside of the Cretaceous cuesta of Long Island, leads to the conclusion that the land hereabouts stands somewhat lower than the attitude which it maintained during at least the later stages of the Tertiary cycle. How many oscillations have occurred between then and now, and how much unequal movement or warping results from the integration of these oscillations, can not at present be definitely stated. Indeed, for our purposes, the only important consequence of the relative depression of the coastal area is the loss of the southern termination of the Triassic basin. Judging by the convergence of the eastern and western boundary on either side of New Haven Harbor, the crystallines would meet south of the Trias about 5 or 6 miles south of New Haven center. A slightly higher stand of the land would reveal the completed border, not of the original Triassic basin, but of such low part of the basin as stood beneath

[1] On the western discharge of the flooded Connecticut, or that through the Farmington Valley to New Haven Bay: Am. Jour. Sci., 3d series, Vol. XXV, 1883, pp. 440-448.

the Cretaceous peneplain. The present relative attitude of land and water submerges the lower end of the reexcavated basin, forming New Haven Harbor, and if it were not for the great volume of the Wallingford belt of clays and sands that occupy the floor of the Quinnipiac Valley, that harbor would reach much farther inland than it does now.

## REVIEW OF ORIGIN OF DRAINAGE.

The remarkably successful manner in which the ridge of the main trap sheet acts as a divide (see fig. 51) demonstrates that the drainage of the region is, as a whole, neither antecedent nor consequent, but of subsequent origin and well adjusted to the structure. This adjustment can have been gained only by the growth of streams along the strike of the weaker strata. So complete is the obliteration of the possible antecedent or consequent streams of the initial stages of the Jurassic cycle that it is impossible now to decide whether one class or the other prevailed in that remote epoch. The lower Connecticut may be located in the eastern upland in consequence of the warp in the south end of the great Springfield crescent (p. 155), or it may have been placed there by superposition from the possible Cretaceous cover of the coastal border (p. 165). Cooks Gap may have been cut by a strengthened antecedent stream, but there are so few examples of this class that generalization and explanation are difficult and doubtful. The Farmington, at Tariffville, is certainly a new settler, and can not be regarded as related in any definite way to an initial stream. Excepting for the glacial accidents, the most striking feature of the drainage systems of the district is certainly the extraordinary accord between divides and hard structures and between valleys and soft structures, without regard to Triassic or Jurassic drainage or topography.

The remarkable degree of adjustment of streams to structures that must have prevailed in immediately preglacial times is an example of a general principle that is believed to be of wide application. Denudation of a region in one cycle allows a considerable, but seldom a nearly complete, obliteration of antecedent and consequent streams by the development of subsequent streams. Denudation of a region in two cycles, of which the first reached an advanced stage, gives much better opportunity for such obliteration, for the subsequent streams produced during the first cycle carry on active depredations through the early stages of the second cycle, and thus accomplish great results. It is therefore fair, in general, to say that in a region where subsequent streams prevail denudation has been carried on through more than one cycle, and the Triassic area of Connecticut is no exception to this rule.

## THE GEOLOGICAL MAP OF THE TRIASSIC AREA.

In the preparation of a geological map of a glaciated area, where a definite color must be assigned to every part of the surface, there is, as a rule, much difficulty found in tracing the limits of the weaker subdivisions, which are so generally sheeted over with drift. This difficulty is fully exemplified in the Triassic lowland of Connecticut, where the sandstones and shales outcrop in relatively small and scattered ledges, and only the trap sheets form continuous exposures. It is therefore advisable to forestall criticism of the map that accompanies this report by explicitly announcing that many of its lines are necessarily inferential. All that can be said of them is that they have been located only after careful search for all guiding outcrops and after much consideration of all admissible interpretations of underground structures. The work of compilation was chiefly performed by Mr. L. S. Griswold. The following notes concerning the several subdivisions of the formation indicated by colors will aid the reader to judge intelligently of the measure of accuracy reached.

The boundary lines where the Trias is adjoined by the crystalline rocks on the east and west are occasionally defined by neighboring outcrops, but they are much more largely guided by the form of the surface. When outcrops are seen, it is found that the crystallines occupy the slopes of the upland, descending nearly to the lowland level. Occasional outcrops of the sedimentaries are found at or near the base of the slopes; at one point on the western boundary (Roaring Brook, Cheshire, p. 19) the sandstones are seen actually lying on their crystalline foundation; at one point on the eastern boundary (Highland Park, p. 131) the sandstones are found abutting against the crystallines, from which they are separated by a fault. Crystalline ledges are generally common near the boundary, although they may be covered with drift for good parts of a mile at a time, but the sandstones are often out of sight for a mile or more from the margin of the lowland. Hence, as a rule, topographic form has been the chief guide in locating the boundary, yet it is thought that no significant error has been thus introduced, although the line may be at some points out of place by several hundred feet.

The intrusive trap sheets near the western border of the lowland are generally well exposed over the crests of the ridges that they form. When they outcrop in strong west-facing bluffs, their under contact can be closely located by the decrease of slope beneath it; but the upper contact or eastern limit of the trap is seldom well defined. The overlying sandstone rarely appears close to it; hence the boundary is, as a rule, somewhat arbitrarily drawn near the eastern base of the ridge, below the lowest outcrop of trap.

The main extrusive trap sheet is probably better bounded than any other subdivision of the formation. Its under contact with the sandstone follows very regularly along the base of the outcropping bluff; its upper contact, when found, lies with almost equal regularity near the base of the eastern slope of the ridge. The adjacent sandstones are, however, very generally concealed—those beneath by trap talus, those above by drift. The greatest doubt here concerns the upper contact from the West Peak to the Farmington blocks, where the breadth assigned to the main sheet is greater than usual in order to include some outlying trap ledges. It is possible that these belong to a local supernumerary flow, separated from the main sheet by a thin sedimentary deposit (p. 71). The occurrence of depressions between the main sheet and the outlying trap ledges lends some support to this view (p. 110), but in the absence of direct proof the whole area is colored as if the trap were continuous.

The anterior and posterior extrusive sheets are of variable definition. The former generally makes a stronger and more continuous ridge than the latter, but at the northern boundary of the State the anterior sheet is wanting (p. 121); in the Tremont block the anterior ridge seems to be weak, as it is much obscured by drift; and the same is true at the south end of the Saltonstall crescent (p. 98). It is therefore omitted at these places for a certain distance, and the under and anterior sandstones are brought together on the map.

A low ridge generally marks the outcrop of the posterior sheet, and for the most part it may be regarded as well defined; but in the New Britain district the map must be accused of much uncertainty. Several redundant ridges occur thereabouts, and it is only by a remarkable system of faulting that a single posterior sheet can be repeated frequently enough to account for all the outcrops that are credited to it. The discrimination of the anterior and posterior sheets where they are confusedly arranged in the district north of Totoket and east of Paug Mountains is admittedly doubtful, yet as far as existing outcrops go the interpretation offered on the map is thought to be consistent and on the whole probably correct.

The dikes in the southern part of the lowland are drawn with somewhat greater continuity than their actual outcrops warrant; and this can not be so well defended as in the case of the interbedded sheets, where continuity is an almost essential characteristic. It is possible that a sandstone cover still remains over certain parts of the dikes (see p. 45) that are mapped as showing persistent outcrops. A number of small dikes are necessarily omitted, especially in the Fairhaven district.

The several subdivisions of the sandstones, shales, and conglomerates—the under, anterior, posterior, and upper sandstones, so called—are identified almost wholly by their relation to the extrusive trap sheets. The colors on the map by which these subdivisions are distinguished are therefore not, as a rule, warranted by direct and independent observation, although they are demanded by the interpretation given to the structure of the entire region. For example, a certain trap ridge near Berlin is regarded, for reasons that appear sufficient, as an outcrop of the posterior trap sheet. Then the underlying beds are called posterior sandstones, and the overlying beds, upper sandstones; and their areas are thus appropriately colored on the map. The general consistence of the results gained by this method is the chief warrant of its correctness. The occasional discovery of the anterior and posterior fossiliferous shales is, as far as they go, a still more striking proof of the correctness of the method; and inasmuch as the localities of these shales are distributed in a most haphazard order, as far as the general outline of the region is concerned, their confirmation of the method adopted may be regarded as applicable over a much larger area than that of their actual exposure.

It is manifest that the stratigraphic interpretation of the bedded series would be much more conclusively established if every subdivision of the formation had its own characteristic composition or fossils; but, with the exception of the anterior and posterior fossiliferous shales, it seems impossible to identify any of the subdivisions of the bedded rocks by their individual features. The posterior subdivision, for example, commonly consists of fine sandstones and shales; but it includes a coarse conglomerate inside of the Totoket crescent. The under sandstones, next beneath the anterior trap sheet, are generally weak and relatively fine grained in the northern part of the area, but east of Meriden they are frequently coarse conglomerates, making strong ridges. The coarse, roughly bedded sandstones, so common at various horizons, have no subdivisions that can serve as stratigraphic guides. The drift-covered area north and east of Hartford, with no trap ridges and rare sandstone outcrops, can not be subdivided. It is only when these various difficulties of identification are realized that the importance of the three extrusive sheets as horizon markers can be appreciated.

The fault line that follows the eastern boundary of the formation is thought to be as well established and as accurately located as the boundary itself. The occasional faults indicated on the western boundary (p. 123) are much more problematic. The oblique faults by which the region is traversed are, as a rule, well determined in the neighborhood of the three extrusive trap sheets, and again in certain notches by which the intrusive sheets are dislocated, but away from the trap ridges the fault lines are, with the exception of a few points, entirely inferential. Like the colors assigned to the drift-covered subdivisions of the sandstones, the fault lines are for the greater part of their lengths drawn so as to give rational correlation to many well-determined facts, rather than to indicate the results of immediate observation. It may be thought hazardous to introduce the fault lines where they are not seen, but, in the light of what is seen elsewhere, it would be much more hazardous to leave them out, and the latter hazard is taken only where the extension of a fault line would not lead it to new confirmation. The termination of a fault line on the map can not therefore be taken as proof that the fault has actually ceased, but only that means of tracing it are wanting.

Accepting the existence of three, and only three, extrusive trap sheets near the middle of the formation, we must admit the occurrence of numerous faults, but the precise location and termination of these faults is often a matter of serious doubt. The difficulty here is one that is common to all graphic representations of somewhat indeterminate phenomena. The existence of the faults is proved; the precise location of their entire length is a matter of doubt, but the map makes their location as distinct as their occurrence. The gravest chance of error under this head is certainly among the posterior ridges in the New Britain district, where it should be clearly understood that the faults are, with hardly an exception, matters of interpretation, not of observation. Yet even here the occurrence of a few "clay dikes" or fault breccias (see pp. 111, 112), revealed by the chance cutting of streams, confirms the occurrence of faults as demanded by the truncation of the trap ridges, and, like the rare discoveries of the fossiliferous black shales, these occasional confirmations deserve an application far outrunning their actual area of occurrence. They are, like a few samples of manufactured goods taken from the open market, fair tests of their many untested fellows.

About New Britain, as elsewhere, the interpretation of the faulted blocks and chips has been guided by the belief that only three extrusive sheets occur, and hence that, as many ridges as there may be in that district, they must all be explained by repetitive faulting. The main and anterior ridges offer no serious difficulty, although the lapse of the anterior near Farmington was for some time a mystery; but the posteriors are very troublesome. For example, the low trap ridge west of Newington is taken to be a supernumerary outcrop of the posterior sheet in the northern of the two West Peak blocks. The posterior and upper sandstones are therefore appropriately colored in on either side of the ridge, and a fault line is introduced to separate this threefold group of colors from the same group nearer the main sheet. For all this graphic definition there is no warrant but the two trap ridges. The fault line here mentioned is absolutely unseen. The sandstones give no sufficient indication of the subdivision to which they belong. Yet it is seriously believed that the acceptance of the second posterior ridge as the outcrop of a second and independent posterior trap sheet would lead to much greater difficulties than its explanation as a repetition of the first posterior sheet by faulting. In the northern part of the Farmington block there is another supernumerary posterior ridge so curiously out of place that no attempt is made to explain it. The repeated posteriors of Lamentation and Chauncy blocks are open to various interpretations; the one given on the map seems simpler and more accordant with adjacent structures than any other yet suggested. The errors involved in drawing these hypothetical fault lines are chiefly such as are inherent in the graphic method, which requires the definite location of a fault line, when it is desired merely to express in a most general way that a fault probably occurs somewhere between two trap ridges.

Complicated as the color pattern is among the numerous faults of the New Britain district, there can be little question that it is much less complicated than the reality. There must be many minor branch faults in the sandstones of which the trap ridges give no indication. If the drift could be stripped away, and the subdivisions of the formation accurately discriminated, a much greater disorder would probably be discovered than is here indicated.

## RECONSTRUCTION OF THE ORIGINAL FORMATION.

Geology being a study of the earth in relation to time, it follows that the name Triassic formation is properly applied only to the entire product of the accumulative processes that were in operation in Triassic time, or during the accumulative portion of that time, in any region. What we see of the Triassic formation in Connecticut to-day is but a fraction of the original accumulation. Hence it is desirable to reconstruct as well as possible the whole from the remaining parts. This task is similar to that so well accomplished by the paleontologist who reconstructs an entire organism from its fossil fragments. He may make mistakes in his reconstruction, but such mistakes are small compared to the error of regarding the fossil fragments as the entire organism. So with the geologist. He is certainly liable to err in his efforts to reconstruct an entire formation, the completed product of a certain period of accumulation, but his errors are much less serious than the capital mistake of assuming that the existing remnant of a formation is the formation. It is not from the remnant, but from the entire formation, that the history of its formative period must be written.

These remarks are peculiarly applicable to the case in hand. The rock structures commonly known as the Triassic formation of Connecticut are manifestly but a part of that formation. The part that is now found should, when historically considered, be associated almost as closely with the late Triassic or early Jurassic deformation, with the long Jura-Cretaceous cycle of peneplanation, and with the incomplete Tertiary cycle of lowland excavation as with the Triassic period of accumulation; for the part, as we see it, can only be understood when deformation and denudation are taken account of along with deposition. The real operations of Triassic time can not be understood until a reconstruction of the entire product of that period of accumulation is accomplished.

In order to follow out this idea, four true-scale cross profiles, traversing the valley lowland from about west to east, were constructed by Mr. Griswold from the data presented on the geological map. The underground structures were then drawn in beneath the profiles, according to the observed and inferred dips. The faults were prolonged downward at the average of the hades observed in the few places where the fault walls are exhibited. The under sandstones were drawn in beneath all the blocks until their thickness about equaled that found west of the anterior trap ridge, and a comparatively even crystalline floor was outlined for their foundation. Overlying structures were next restored by the addition of the missing strata, until something more than the total observed thickness of the formation was reached in every block. The excess of the reconstructed upper surface of the formation over the highest observed member of the sandstones was demanded by the fact that denudation has removed at least 500 feet of sandstones above the highest horizon now seen.

On account of the prevailingly greater dip on the western than on the eastern side of the lowland, it has been concluded, as stated on page 38, that a continual movement of subsidence occurred during the period of Triassic deposition, resulting in the production of a trough-like depression, always filling with sediments nearly as fast as it deepened. Thus interpreted, the successive strata must be drawn of decreasing thickness from the middle of the trough toward either side, gradually wedging out at the original margin of the trough. It is highly probable, as stated on page 33, that the margins of the Triassic sediments advanced and receded by moderate measures during the period of depression and deposition, and hence that successive members of the formation wedged out and overlapped somewhat irregularly; but no attempt is made to illustrate this probable structure in the sections. The rate at which the formation thins out laterally, and hence the distance to its original margins, can not well be determined; indeed, this element in the problem of reconstruction is in greater doubt than any other. The sections do not indicate any faulting during the time of deposition; but it is not desired to exclude the the possibility of such movements.

The diagrams thus constructed exhibit the completed section of each block in its faulted position. After cutting the sections apart on the fault lines and then "unfaulting" and "untilting" the several blocks, a little accommodation sufficed to make the upper surfaces of all level and continuous and to bring the corresponding strata all in line. A fair approximation is thus made to a cross section of the original Triassic formation in its completed form. One of these four sections, crossing the lowland from a little north of Bristol to South Glastonbury, and passing Rattlesnake and Cedar mountains on the way, is reproduced on Pl. XX, in both its faulted and its restored position.

The darker tints represent the structures as they are believed to exist underground to-day. The lighter tints are the restored parts, above the present surface of the ground. The very slight difference between the profile and the sea-level line illustrates how small a share of the total formation is involved in the relief of the land to-day.

Six faults near the left end of the section are introduced without any field evidence for their occurrence, in order to diminish both the thickness of the under sandstones and the amount of denudation over the Western Uplands in the Jura-Cretaceous cycle. In defense of this apparently arbitrary construction, it may be urged that the occurrence of faults about as here represented is much more probable than their absence, in view of the frequency of faults farther east where they can be detected. It may be added that the character of the original formation would not be seriously altered by their omission.

There are various elements of uncertainty in work of this kind. Only the actual outcrops are within the reach of direct observation. The extension of the small observed area of the various sedimentary members of the formation over the rest of the drift-covered lowland, the identification of the repeated outcrops of certain members, and the limitation of the several members by fault lines, are all matters of interpretation. The interpretation is well supported in certain districts, but it is open to considerable error in others. The chance of error increases as soon as the attempt is made to extend observed structures underground, and still more when the same attempt is made with respect to overground structures. The dips of the strata are, however, generally so regular that it is thought no serious error is thus introduced. The hade of the faults is observed in but few cases, and the underground prolongation of the fault lines must remain greatly in doubt. The crystalline floor is drawn throughout as a peneplain; the evidence for this conclusion having been fully discussed on pages 20–26; but it is, of course, possible that buried monadnocks may occur here or there, singly or in groups, beneath the Triassic strata.

Yet when all these many chances of error are recognized, it is believed that the reconstructed sections are of serviceable approach to the truth. Their error is chiefly that of overdefinition such as is essential in graphic presentation; but, to counterbalance this disadvantage, they serve a valuable purpose in compelling the student of the region to give definite consideration to an important part of its geological history that might otherwise be neglected.

Certain interesting results follow from the sections as now drawn. The interpretation of the existing structures, as parts of a repeatedly faulted mass, is found to be consistent with a reasonable process of origin in a broad trough that was filled at about the same rate as that of its subsidence. It appears further that no great original extension of the formation is necessary on the east or west of its present limits. Indeed, a narrowing even within the limits given on the sections would be accomplished if the hypothesis of marginal faulting during accumulation were adopted. It is not the intention, in either text or diagrams, to rule out that hypothesis, but as it is less consistent with the observed facts than the hypothesis of troughing, it is not graphically represented.

The existing remainder appears to be about half the volume of the original formation, as reconstructed. If limited by marginal faults, instead of gradually wedging out, the existing remainder would be about three-quarters of the original volume. If extended far over the crystallines on the east and west, the proportion would be correspondingly reduced. Most of the denudation of the original mass was accomplished in the Jura-Cretaceous cycle of peneplanation; less than a tenth of the denuded mass was carried away in the Tertiary cycle. The existing volume, as determined from the several sections, is roughly 720 cubic miles. The original volume, as drawn in the reconstructed sections, is therefore about 1,400 cubic miles; it would be reduced to about 1,000 cubic miles if limited by marginal faults; it would be increased to much greater measure if extended far over the crystallines. These figures may be regarded as giving approximations to the order of the quantities involved, but they make no pretence to precision, inasmuch as they are based on cumulative hypotheses, as has been stated sufficiently above.

There is a natural hesitation felt against accepting a conclusion that involves so great a denudation as is required by the several restored sections, but when this hesitation is confronted with the reasonable line of investigation that leads to the conclusion, it can not be well justified. It is the order of the world that one period witnesses the denudation of the deposits formed in previous periods. The Triassic strata were but the ruins of the laborious accumulations of earlier times. As such, a part of their great original volume may have been called on for contributions to the Jurassic and Cretaceous formations that presumably constitute a considerable part of the Atlantic coastal shelf of the United States. There can be no doubt as to the sufficiency of time and process to accomplish vast works of denudation in the Jurassic and Cretaceous periods. Indeed, when the restoration of the original formation from the observed structures calls for the extension of strata deep underground and high overhead, there is no legitimate ground for objecting to one of these hypothetical constructions more than to the other. The first view of the formation as restored in section may suggest distrust, but a consideration of the whole problem that is epitomized in it will, it is confidently believed, lead to its acceptance as a reasonable conclusion to all that precedes.

# GEOLOGY OF THE EDWARDS PLATEAU AND RIO GRANDE PLAIN ADJACENT TO AUSTIN AND SAN ANTONIO, TEXAS, WITH REFERENCE TO THE OCCURRENCE OF UNDERGROUND WATERS.

BY

ROBERT T. HILL AND T. WAYLAND VAUGHAN.

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# ILLUSTRATIONS.

# GEOLOGY OF THE EDWARDS PLATEAU AND RIO GRANDE PLAIN ADJACENT TO AUSTIN AND SAN ANTONIO, TEXAS.

By Robert T. Hill and T. Wayland Vaughan.

## INTRODUCTION.

The artesian systems of the eastern half of Texas are numerous and underlie several large areas. The productive areas extend from near Red River, Denton County, to Del Rio, on the Rio Grande, and from near the center of the State to the coastal islands of the Gulf. Collectively they comprise a district 450 miles in length and averaging 300 miles in width. It is doubtful if there has been anywhere a more remarkable development of artesian wells in the last ten years than in this region. At numerous places copious flows of water have been obtained, and districts which a few years ago had only inadequate supplies now possess artesian wells furnishing water in large quantities to cities, ranches, and farms, improving the hygienic conditions, and yielding water for the stock raising, agricultural, manufacturing, and transportation industries.

The artesian wells of the eastern half of Texas belong to several distinct systems, the term "system" including all wells having their source in the same set of rock sheets or strata. It would require a large volume to describe each of these various artesian systems. In the Cretaceous formations alone there are no fewer than five, and two of these—the Travis Peak, or Waco, and the Edwards[1]—receive consideration in this paper. It is especially proposed to explain as well as possible the principles of the supply of the artesian belt supplied by the Edwards system and the probability of success or failure of wells sunk in different parts of its area.

Within the last two decades numerous artesian wells have been drilled in the vicinity of San Antonio, Texas, and for many years the source of the water has been a matter of perplexity. Unfortunately it has not been within our power to secure accurate logs of these wells, accompanied by specimens, or even to ascertain with accuracy the number and location of the wells, or to collect other essential data of a statistical

[1]The terms "Travis Peak" and "Edwards" are formation names used in this paper in place of "Trinity Sands" and "Caprina limestone," hitherto employed. The reasons for the employment of these new names are given on later pages (pp. 216, 227).

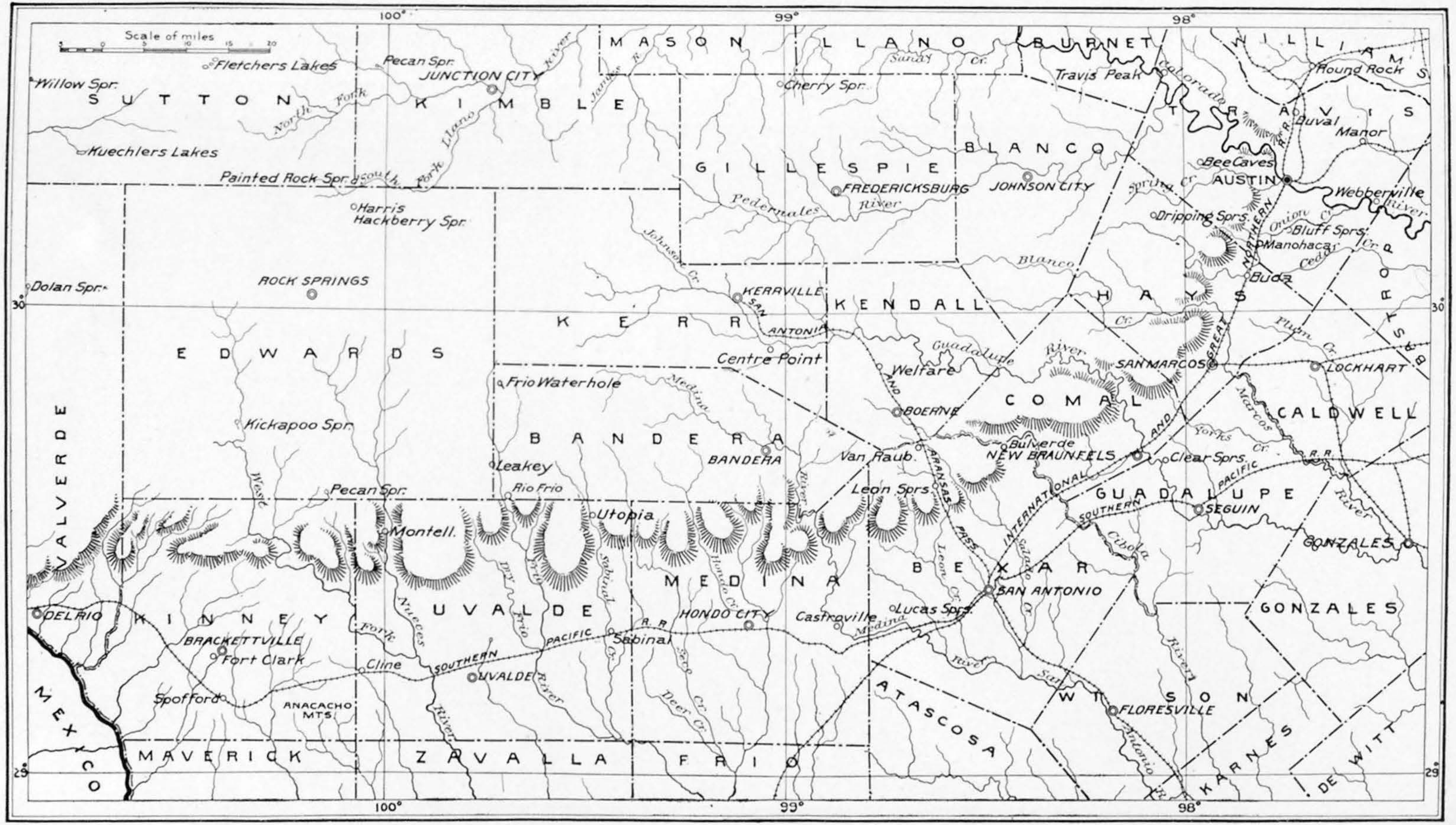
nature, such as the Government has deemed expedient to collect in other artesian regions. Such facts as are herein presented have been collected incidentally by the writers while making private reconnaissances or official geologic surveys in the region, and through correspondence with citizens. Accordingly, this report is not to be considered as a statistical or engineering paper, but rather one which deals with the geologic side of the underground water question.

The source of the San Antonio water supply has been ascertained by detailed studies of the structure and outcrop of the water-bearing beds in the adjacent regions. A detailed map of the Austin quadrangle was made, and the thickness, sequence, paleontology, mineral composition, and water capacity of every portion of the Cretaceous section along the Colorado River were ascertained. A reconnaissance was next made southwestward from Austin via Fredericksburg, Kerrville, and the headwaters of the Frio and Nueces to Fort Clark, Texas, for the purpose of studying minutely the variation of the rock sheets outcropping in that direction, which, owing to the direction of their dip, must necessarily underlie the city of San Antonio. At the southeastern end of this line detailed studies and mapping were resumed upon the Nueces, Brackett, and Uvalde quadrangles. By this method a base line of geologic sections, so to speak, was established along the strike of the outcrop of the Lower Cretaceous formations, the sections at the extremities of which were determined with the greatest accuracy possible, while check sections have been made at numerous intervening points along the line.

The result of this work was the discovery that, while these well waters come from the same series of beds that supply the artesian wells of the Waco, Fort Worth, and Dallas regions north of the Colorado, their occurrence presents some important differences of detail. Instead of having their immediate source in beds of porous sands, like the wells about Waco, they are derived largely from the Edwards limestone, hitherto supposed to be one of the most impervious formations of the whole Cretaceous section.

It became apparent that this hitherto unappreciated water-bearing formation had great possibilities for supplying with either flowing or nonflowing wells a large area of country lying between Austin and San Antonio, extending west of the San Antonio River along the northern margin of the Rio Grande Plain toward the Pecos Piver, and even comprising the extensive summit region of the Edwards Plateau.

There are also many remarkable springs in that portion of the Rio Grande Plains and Edwards Plateau lying between the Colorado and Pecos rivers, which will be discussed in this paper. These belong to two distinct classes, each characterizing one of the geographic subdivisions mentioned, and each exhibiting a method of escape of the underground water of the region. One line of these springs follows approximately the margin of the Rio Grande Plain, close to the line of




MAP SHOWING THE RELATIONS OF THE EDWARDS PLATEAU, THE BALCONES SCARP LINE, AND THE RIO GRANDE PLAIN.

The hachured line is the Balcones escarpment; the Edwards Plateau is to the north of it, and the Rio Grande Plain is to the south.

the railroad from Austin to Del Rio. The other class embraces the springs of the canyons of the Edwards Plateau.

In order to understand the occurrences of water of the San Antonio system, it is necessary to carry in mind:

(1) The geography of the region.

(2) The simple laws of the occurrence and distribution of underground water.

(3) The composition, variation, and arrangement of the rocks underlying the region and affecting the distribution of water.

The general map (Pl. XXI) will illustrate the geographic relations of

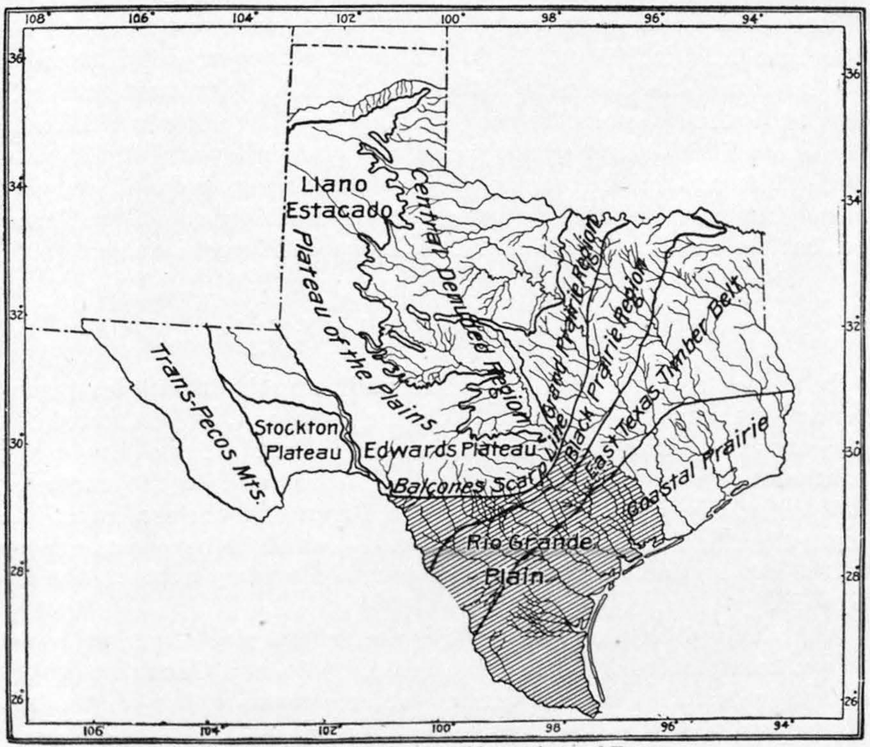



FIG. 53.—Key map of topographic provinces of Texas.

the Edwards Plateau to the adjacent and lower Rio Grande Plain. The local sections and descriptions of the formations given in the geologic portion of this paper (pp. 215–260) can be referred to for all necessary geologic data, such as structure and variation in thickness.

## GEOGRAPHY OF THE REGION.

The traveler by rail from Austin via San Antonio to Del Rio, on the Rio Grande, sees from the car window two conspicuous regions, each having its peculiar geographic features. On the left is a gently undulating plain whose margin the railroad follows. This is here termed the Rio Grande Plain.[1] On the right is constantly visible a line of low, circular, flat-topped hills, the Balcones scarp line, which represents the jagged southeastward front of a higher region which has been called the Edwards Plateau.[2] The Rio Grande Plain, the Edwards Plateau, and the Balcones scarp line are the chief geographic features of the region. Broadly considered, they are a lowland plain inclining gently southeastward to the Gulf of Mexico, an upland plain rising gradually toward the northwest, and a rugged zone of separation which includes a quick ascent from plain to plain.

The Atlantic and Gulf coasts of the United States from the Hudson to the Rio Grande are margined by a broad lowland called the Coastal Plain. The portion of it lying farthest to the southwest is called the Rio Grande Plain. One of the more important geographic divisions of the interior of the continent is the Great Plains. Its most southerly division is the Edwards Plateau. Farther north the Coastal Plain and the Great Plains lie far apart, the Mississippi Valley and the Appalachian belt and other geographic provinces being included between them; but southward they converge, finally meeting in southern Texas, so that the Rio Grande Plain and the Edwards Plateau lie side by side.

## RIO GRANDE PLAIN.

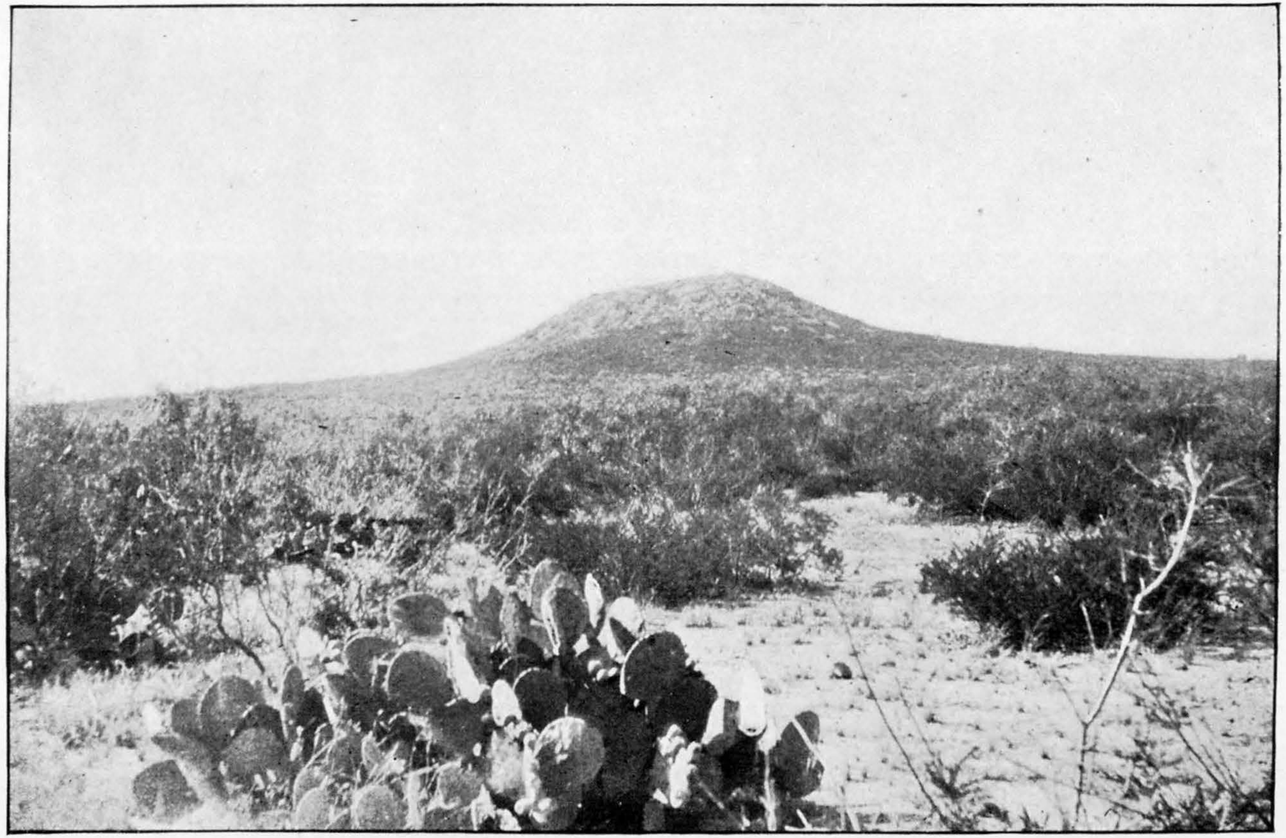
In shape this plain is an irregular quadrilateral with angles turned toward the four cardinal points. On the northwest, as just stated, the Balcones scarp line separates it from the Edwards Plateau. On the southeast it is bounded by the Gulf of Mexico, and on the northeast it is arbitrarily limited by the Colorado River. On the southwest it is limited by the folded mountains of Mexico, which lie beyond the valley of the Rio Grande. It also sends a tongue for many miles up the Rio Grande.

The plain's drainage, which follows the general slope from northwest to southeast, includes the Nueces, San Antonio, and Guadalupe rivers and their branches, besides various minor streams which join the Rio Grande and Colorado or enter the Gulf direct.

As compared with the adjacent plateau and mountain regions, its characteristic topographic feature is a low relief, but its surface is broken by occasional undulations and in places by hills of considerable height. Some of these are the low scarp lines of drainage valleys; others, like the Dos Hermanos Hills of Webb County, are buttes capped by limestone; others, like Pilot Knob, in Travis County, are old volcanic necks; Pinto, Las Moras, Turkey, and Elm mountains, in Kinney County, are buttes composed below of sedimentary rocks and above of caps of igneous rock; Sulphur Peak and Fort Inge, in Uvalde

[1] The indentation of this plain up the Rio Grande has been called the "Rio Grande Embayment" by Hill: Bull. Geol. Soc. America, Vol. III, 1891, p. 93.

[2] Hill, ibid., p. 90.



SULPHUR PEAK, UVALDE COUNTY, TEXAS; A BASALT NECK ON THE RIO GRANDE PLAIN.

County, are masses of solid basalt (see Pl. XXII); the Anacacho Hills, extending east and west in southern Kinney County and constituting the most rugose part of the plain, are of still another type, consisting of a monoclinal plateau, or cuesta, sloping southward and presenting a steep scarp to the north. (See geologic profile, fig. 66, p. 260.)

The eastern part of the plain belongs climatically to the humid and subhumid regions; the western part, to the arid. The fertile "black lands" occupy large areas as far southwest as San Antonio; but continuous cultivation is limited by increasing aridity west of Bexar County. The plain as a whole is mostly grazing land, and supports one of the greatest stock-raising industries of the United States. San Antonio is its commercial center.

In passing westward from comparative humidity to aridity, there is a corresponding decrease in the depth and richness of the residual soils. Soils derived from certain geologic formations which in the Black Prairie[1] and eastern Texas timber regions are fertile, here become progressively more sterile and barren, and are impregnated or coated with a peculiar calcareous incrustation known in the Southwest as "tierra blanca" or "tepetate."

Much of the region is covered by the peculiar flora known as chaparral, consisting of a thorny growth of many species of scrubby acacias—mesquite, guaxillo, and huisatche—between which is an undergrowth of cactus, especially the large Opuntia known as nopal. For this reason the region is sometimes called the "chaparral country." There are also stretches of open prairies covered by nutritious grasses.

It is not within the province of the present paper to deal with the plain in its entirety, but only with a strip along its western margin not exceeding 30 miles wide, a belt underlain by the Cretaceous formations.

## BALCONES SCARP LINE.

The Balcones scarp line is the frayed and ragged coastward border of the Edwards Plateau. From the more open and level lower country it appears as a sharp line of timber-covered hills, and these are universally called "mountains" by the people of the region.[2] It commences near the northern line of Travis County and continues a little south of west, through Travis, Hays, Comal, Bexar, Medina, Uvalde, and Kinney, to Valverde County, where it meets the Rio Grande. Near Austin its highest summits are about 400 feet above the margin of the lower plain; in Uvalde County, nearly 1,000 feet.

[1]The "Black Prairie" region is that underlain by the soft Upper Cretaceous chalky limestones and marly clays, where there is considerable rainfall. It comprises the richest agricultural lands of the State. The cities of Austin and Dallas are situated near its western margin; Terrell and Corsicana are near its eastern margin.

[2]A singular fact in the progressive cartography of our country is that notwithstanding the conspicuousness of the Balcones scarp line as a topographic feature and that it was shown upon all the mother maps of the region, such as those made by Bartless, Roemer, and J. De Cordovas (J. H. Colton & Co., New York), it has been omitted from the more recent maps.

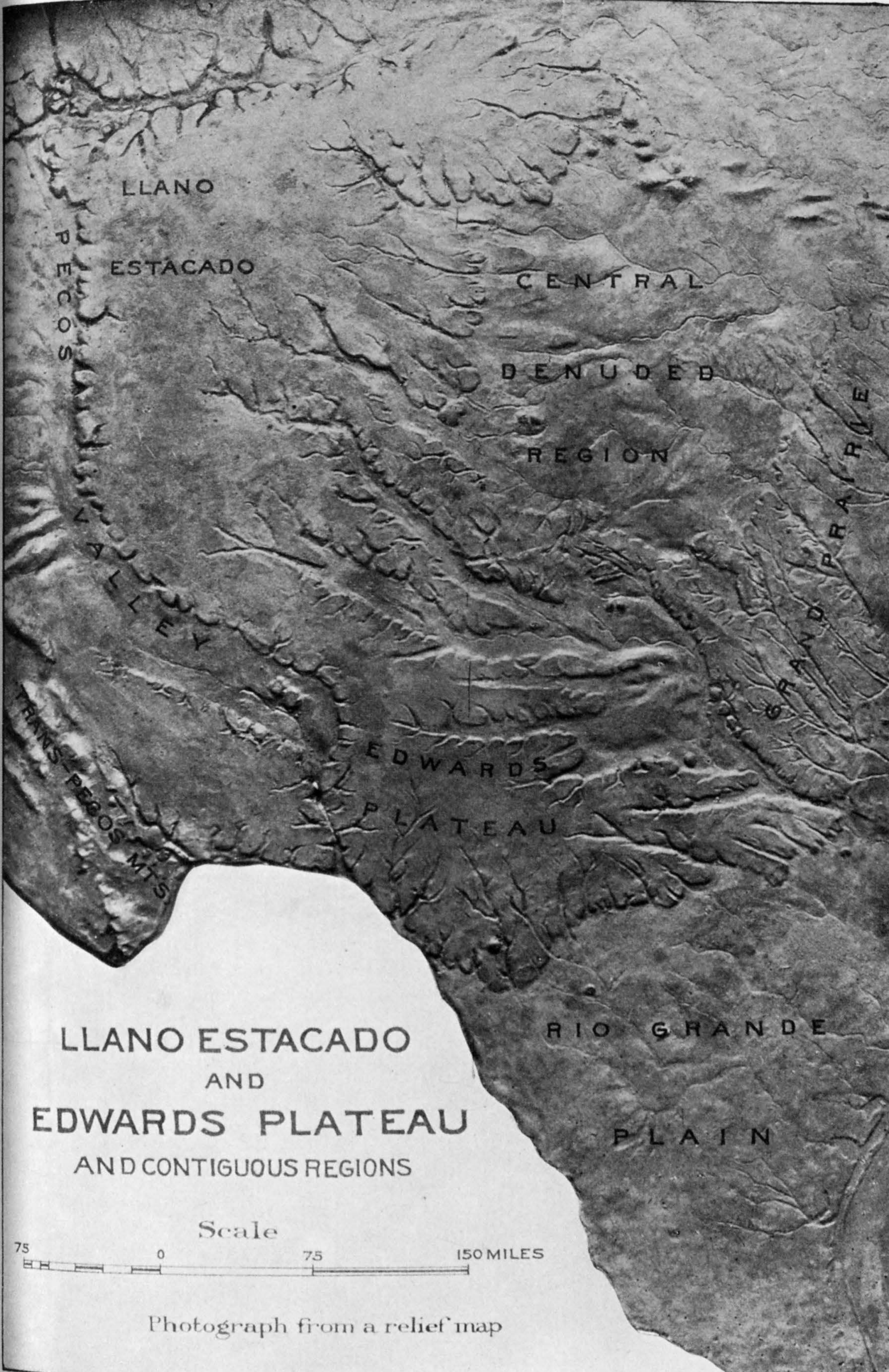
The drainage flows directly across the scarp line and has cut great canyons backward into the Edwards Plateau. The depth and precipitous character of these increase in the streams successively encountered as one goes westward. The portions of these streams lying within the area of the plateau before they cross the fault line have cut their channels approximately down to the level of the Rio Grande Plain. Their bocas[1] in many places cut the scarp line into a series of tongue-like salients projecting toward the plain, and the line is further diversified by plateau remnants in the form of outlying buttes. The position of the scarp is determined by a complex dislocation of the rocks, the Balcones fault, which will be described in a subsequent section (p. 258).

## EDWARDS PLATEAU.

The Llano Estacado and the Edwards Plateau together constitute in Texas the Plateau of the Plains. This lies within the area inclosed by the Canadian on the north, the Pecos River on the west, the Balcones escarpment on the south and southeast, and an irregular line of scarps along the headwaters of the eastward-flowing drainage of the Colorado, Brazos, and Red rivers of Texas. The general outline of this area is shown in the accompanying photograph of a model, Pl. XXIII. It is over 500 miles in length and in places 280 miles broad. It is a vast quadrangular mesa, surrounded on all sides by descending escarpments. In its horizontal geologic structure and its relations to the surrounding region it may be broadly compared to a book laid upon a table and very slightly tilted or raised at one end. Its component strata are almost as parallel and regular as the leaves of the volume. While its central portion still presents a general level, its borders are cut by headwater erosion into a fringe of projecting drainage divides, accompanied by many remnantal buttes and mesas, showing the great erosion by which the plateau has been and is being gradually etched away. Its eastern margin in particular has been greatly reduced by this process.

The Llano Estacado and Edwards Plateau merge into each other along the central portion of the summit of the greater plateau, and there is no known line of demarcation between them. There is a great difference between the characters of their surface, their soils, and the underlying geologic strata, which collectively gives to each of the two regions a distinct character. The soil and underlying rocks of the Llano Estacado consist of unindurated loams, marls, and sands. Hardly a stone of building size can be found in its broad extent. The Edwards Plateau is in part a rough limestone country, resembling in some respects the western margin of the limestone country of the Grand

[1] Boca, a Spanish topographic term, used to indicate the mouth of a canyon valley where it debouches on a plain. Example, Boca del Agua, Jamaica, corrupted by the English into "Bog Walk."




LLANO ESTACADO
AND
EDWARDS PLATEAU
AND CONTIGUOUS REGIONS

Photograph from a relief map

Prairie region (see Pl. XXIII), with which its rocks are geologically identical. The hard Cretaceous rocks of the plateau partially underlie the Tertiary formations of the Llano Estacado, and it is probable that the latter have been eroded from the plateau.

It is in the latitude of Upton and Midland counties, or a little north of the thirty-first parallel, that the hard rocks of the Edwards Plateau become covered by the marls and sands of the Llano Estacado, and it is here that the line between them is provisionally drawn while awaiting further exploration. In these counties the narrowest width of the whole plateau is found.

The Edwards Plateau occupies nearly the whole of the counties of Crockett, Valverde, Edwards, Sutton, Schleicher, Kimble, Kerr, Bandera, Gillespie, Kendall, and Blanco, and about one-half of the counties of Crane, Upton, Tom Green, Irion, Concho, Menard, Travis, Hays, Comal, Bexar, Medina, Uvalde, and Kinney. The northeastern and western boundaries are cliffs due to erosion. The northeastern overlooks a broad denuded area traversed by the Colorado River and its tributaries. The western margins the Pecos Valley, which at the north is an open plain and at the south a canyon. Beyond the Pecos are folded mountains, but in the region of the canyon these are separated from the river by a broad plateau similar in type to the Edwards.

The southeastern boundary is also a cliff, the Balcones scarp, but, as already stated, was primarily determined by dislocation rather than erosion. The main water-parting lies near the western edge, and the greater part of the plateau is drained by streams running eastward and southeastward. These have cut deep canyons, dividing its marginal portions into long, narrow tongues. The drainage westward to the Pecos has accomplished only a moderate amount of erosion, so that the western cliff is comparatively simple in contour.

The part of the plateau which has been most thoroughly studied and with which this report is more especially concerned is the eastern and southern, the portion adjoining the Colorado Valley and the Rio Grande Plain. The characteristics of the plateau are most strongly impressed on the observer who enters it from the Rio Grande Plain, for in crossing the Balcones line he experiences a sudden and complete change of scenery, with accompanying changes in floral, geologic, and cultural conditions. Instead of long, wide sweeps of prairie, void of sharp relief, he finds a region of steep canyons and sloping hillsides. The monotony of deep and dusty soils is replaced by alternate outcrops of cream-colored rocks and marls, occurring in long, continuous, and horizontal lines of stratification. A deciduous flora suddenly appears in the canyon valleys, replacing the semiarid chaparral. Rivers of flowing water, fringed by forests, replace the dry and stony stream ways of the plain, the mere trace of which is often lost in times of drought, and they now become fixed features of the landscape, boxed in with steep-walled canyons. Rugged evergreen hills succeed the long stretches of low undulating land with the yellow-brown adobe soils. Some of the beds of stratification composing the canyon walls are barren of foliage; others are occupied by the dark evergreen shrubs, juniper and Sophora, extending like garlands around the brows of the circular hills. With the increasing altitude the air becomes more exhilarating and the heat of the day is tempered by cooler nights.

The Edwards Plateau, like the whole of the great plateau of which it is a part, presents three simple topographic elements—(1) the flat-topped summits of the decaying plateau, (2) the breaks[1] or slopes of its crenulated borders and canyoned valleys, (3) the stream ways. In local parlance the main summit region is called the "divide," the marginal breaks and disconnected mesas the "mountains," and the stream ways are mostly "rivers."

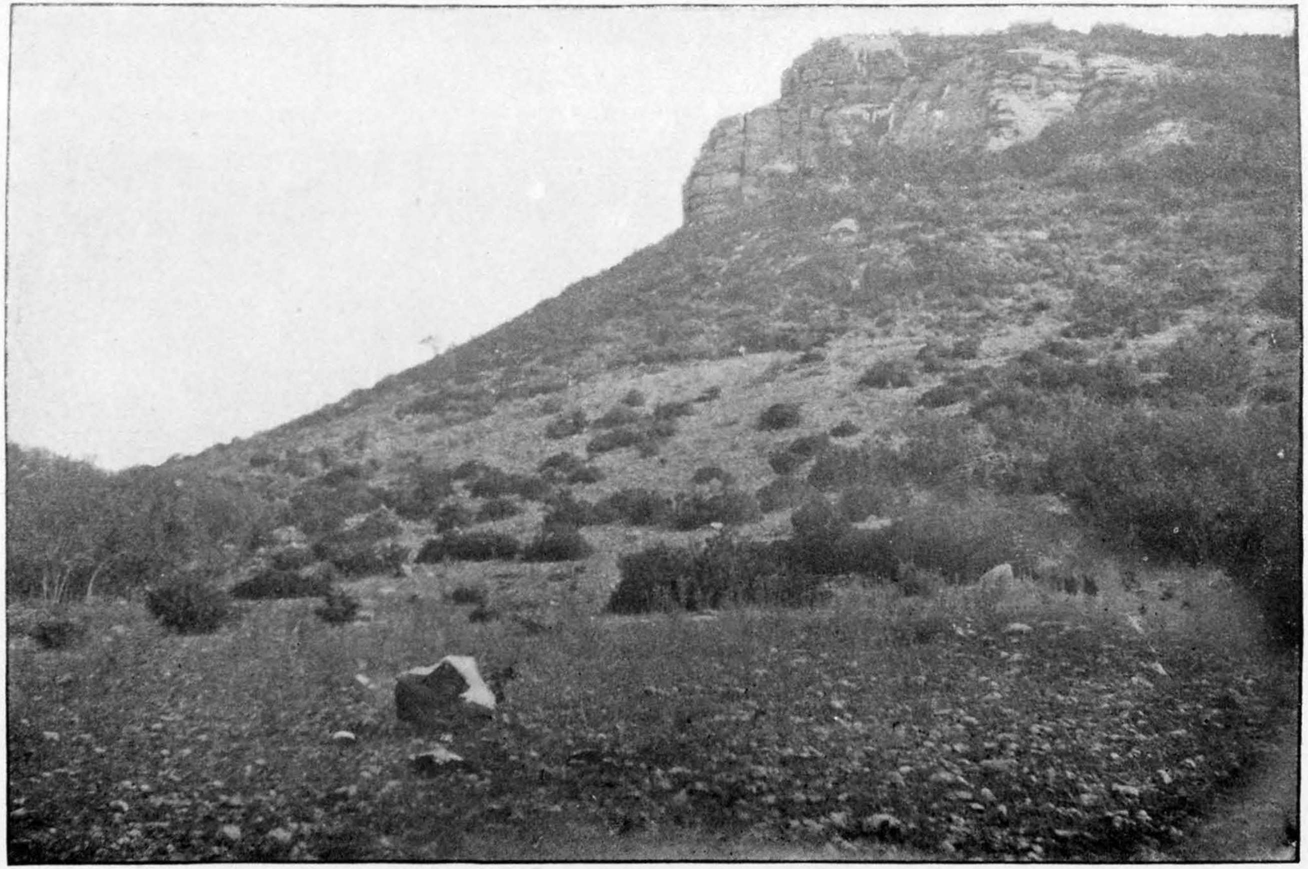
The summit of the plateau is reached by ascending the long canyons of the streams and passing out upon it through their "draws" or caletas.[2] Like that of the Llano Estacado, it is flat and void of constant-running streams. There are a few shallow, pondlike depressions or "sinks," which occasionally contain water immediately after rainfall. In general it is covered with a thick growth of nutritious grass and is without forest. Here and there, however, may be seen thick patches of scrub live oak, known as "shin oaks," growing in dense patches called "shinneries." For miles and miles the level, grass-covered plain stretches before the eye like a great sea, the view broken only at long intervals by the tall shaft of a ranchman's windmill, rising like the sail of a lonely vessel on the level sea. This summit region has an altitude of about 2,400 feet in the northern edge of Edwards County, gradually rising, with an ascent of about 4½ feet per mile, to the west and northwest and falling at the same rate toward the east.

In the dissected border of the plateau are hundreds of remnantal buttes or hills, like Mount Bonnel and Lone Tree Mountain, near Austin, and Round Mountain, in northwestern Uvalde County (Pl. XXIV). Standing upon one of these hills, one can see that its horizontal layers of rock once extended across the valleys to the opposing bluffs and summits and realize that the valleys have been cut out of these once continuous rock sheets. The major stream ways have eroded their wide valleys below the summit level, so that the surface of the old plateau is preserved only in the divides.

Ultimately the headwaters of the streams will all meet, and the general summit level will be cut into innumerable buttes and mesas, like

[1] This expressive term, denoting the cliff margin of an upland plain where it breaks away to a lower plain or valley, has not yet found its way into the dictionaries, but is in common use in many districts of the far West.

[2] From the Spanish. This is a useful word for the ultimate and smallest headwater ramification of a lateral stream. It is synonymous with the term "draw," used in the middle plains region of the United States, the "coulée" of Montana, and "drain" as used in Colorado. In this paper the local topographic nomenclature of Spanish-America will frequently be used. This has been set forth in a paper entitled Descriptive topographic terms of Spanish-America: Nat. Geog. Mag., September, 1896, Vol. VII.



ROUND MOUNTAIN, UVALDE COUNTY, TEXAS.

Type of canyon wall of the Edwards Plateau with talus slope at base.

those now found along its border. These in turn will be planed down to lower and lower levels. By this process many layers of rock have been stripped from the plateau in the past, as others will be in the future.

The present plateau summit exists because of the superior hardness and the capacity for resisting erosion of the rock sheet that caps it. This cap rock is the Edwards limestone.[1] It not only everywhere constitutes the summits, but gives character to all the surrounding scarps and hills. The beds which originally overlay it were comparatively soft and yielding, and have almost completely disappeared.

The streams of the plateau have important bearing upon the question of underground water, and as they constitute a group of rivers characteristic of a large region in Texas, they are well worth the serious attention of those interested in the economic and geographic problems discussed in this paper.

The Rio Frio will be described as a type of the rivers flowing through the southern margin of the plateau. From where it passes out of the plateau onto the Rio Grande Plain, just north of the Southern Pacific Railroad, for fully 50 miles toward its mouth its bed is a shallow, usually waterless stream way, meandering across the gravel-covered country. Flowing water is usually found in this portion of its

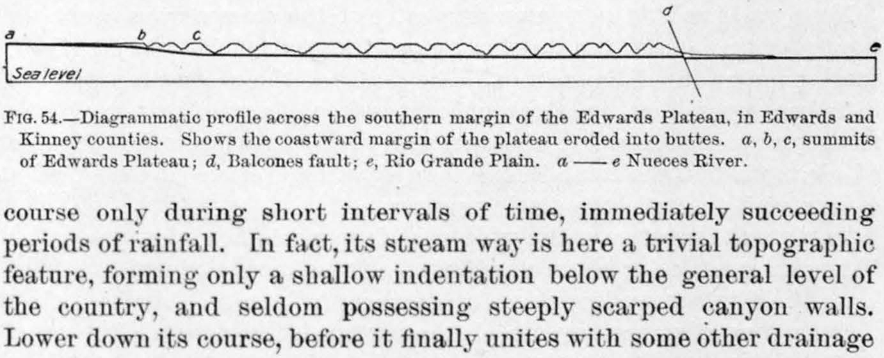



FIG. 54.—Diagrammatic profile across the southern margin of the Edwards Plateau, in Edwards and Kinney counties. Shows the coastward margin of the plateau eroded into buttes. *a*, *b*, *c*, summits of Edwards Plateau; *d*, Balcones fault; *e*, Rio Grande Plain. *a* —— *e* Nueces River.

course only during short intervals of time, immediately succeeding periods of rainfall. In fact, its stream way is here a trivial topographic feature, forming only a shallow indentation below the general level of the country, and seldom possessing steeply scarped canyon walls. Lower down its course, before it finally unites with some other drainage way, living water may reappear, but not within the region treated of in this paper.

North of the railroad the stream emerges from a boca (like that seen in Pl. XXV) of the Balcones escarpment, and still farther northward it traverses a canyon of the plateau. In this canyon the stream way is in the bottom of a flat valley filled with ancient gravel deposits and bordered on each side by high bluffs of stratified limestone. In its lower portion the valley averages 2 miles in width, and contains some fairly good agricultural lands. Entering the canyon, the road follows the dry, stony stream bed a short distance before flowing water is encountered. Farther up the stream running water again ceases and the dry bed again appears. In this manner the continuity of the river is interrupted and its bed is broken up into alternations of dry and

[1] The geologic formations here mentioned are described more fully on page 227.

watered segments. In the flowing segments are clear and swift bodies of water, supplying as much as 1,000 gallons per minute and competent to irrigate many acres of land.

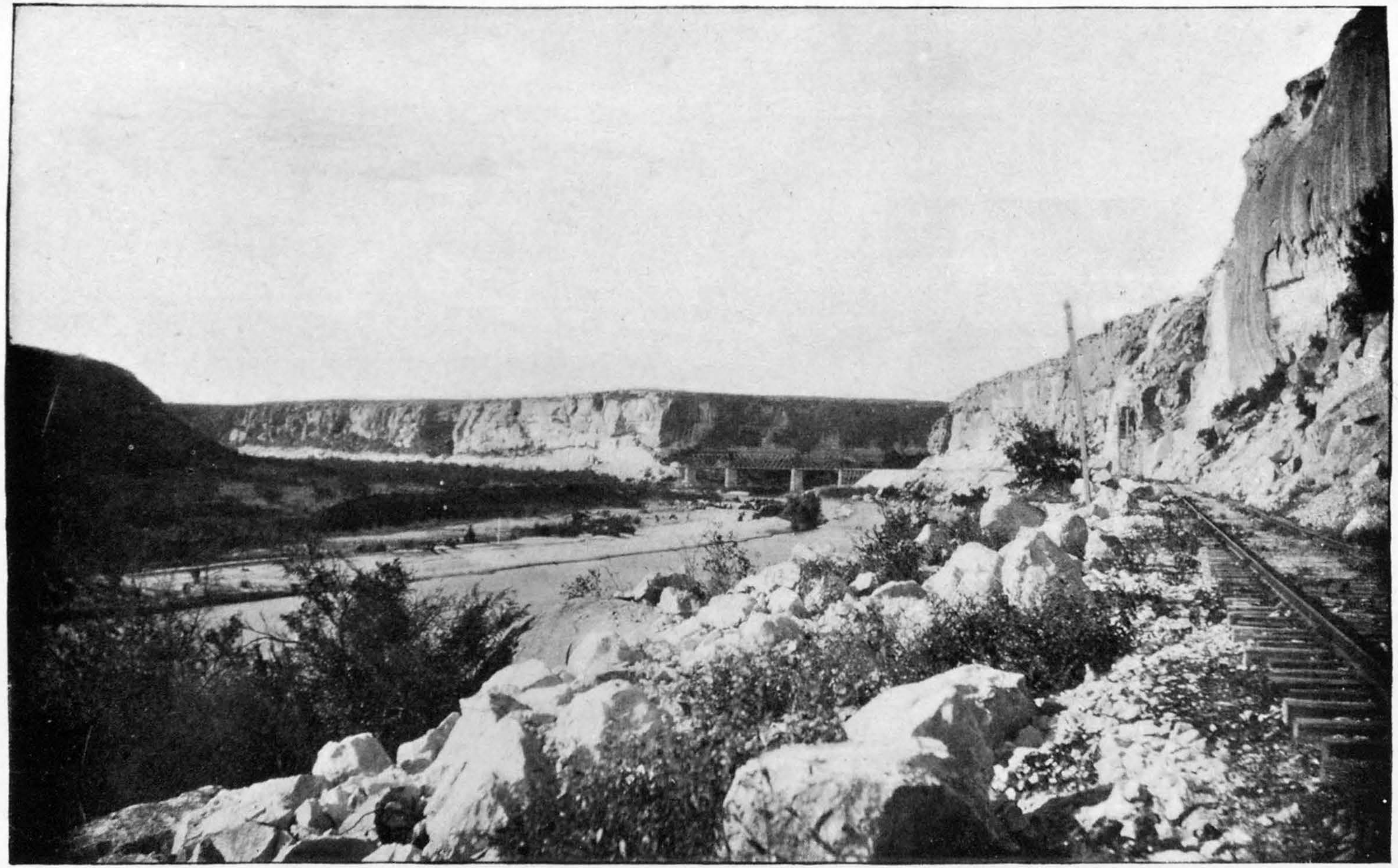
Wide, flat-bottomed canyons of this type continue from the Balcones scarp line via Van Pelts and Leakey to the "water hole" near the head of the West Fork, and the number of springs and quantity of living water progressively increase in that direction. Above the headwater "hole" a different type of canyon sets in and permanent water ceases. The fall or gradient of the meandering stream way in the canyon is about 23 feet per mile in the Nueces, and is probably the same in the Frio. The fall of the bottom of the canyon, without considering the stream meanderings, is about 60 feet per mile.

The scarps which make the outer borders of these flat-bottomed valleys are nearly always steep, usually consisting of alternations of bluffs, benches, and slopes corresponding with the horizontal lines of stratification and giving to the landscape a terraced topography. Between the widely separated escarpment walls are the broad, level second bottoms, standing some 50 feet above the level of the present stream ways. These bottoms consist of an alluvial deposit (the Uvalde formation) and record a peculiar event in the history of the region—an event to be discussed in subsequent pages.

Nearly all the living or perennial water of the stream ways occurs in the flat-bottomed canyons, and they are the seat of the chief agricultural population of the region. The beds of their stream ways are usually composed either of the smooth surface of some horizontal limestone stratum, over which the water may flow for a great distance, or of clean-washed flints and limestone bowlders, often bleached to a chalky white color in the glaring sunshine.

In places the waters make wide and deep pools having the peculiar light sea-green color characteristic of all the spring rivers breaking from the Cretaceous limestone of southwest Texas. Elsewhere they flow for long distances over a single stratum of horizontal limestone. In the latter case the water spreads out in a thin sheet only an inch or two in depth and sometimes 100 feet wide.

It is almost impossible for the traveler who has seen the continuation of this same stream in the dry region of the Rio Grande Plain to recognize it in the beautiful flowing river now before him. Forests of ash, pecan, and elm fill the valley, while gigantic cypresses border the water. If he should chance upon one of these water holes without having traced the continuity of the stream course, he would believe that he stood upon the banks of a large and continuously flowing river. He would soon find, however, that after flowing a short distance the water would disappear, either by absorption into the bed of the gravel-filled stream way or through fissures in the solid underbed. These running water holes are constant, and do not depend upon the local rains, but are supplied by perennial springs draining the rocks underlying the plateau.



RIO GRANDE CANYON AND BOCA OF PECOS CANYON.

The bluff in the foreground is covered with a tufaceous incrustation—tepetate.

Above the springs at the head of the flat-bottomed valleys the limestone bluffs become more vertical and close in toward each other, changing the cross profile of the canyon from U-shape to V-shape. In this, which may be called the cataract portion of the stream way, the gradient is much steeper, ranging from 100 to 300 feet per mile. Here the stream way carries no water except in time of flood, but then it is a series of cataracts leaping from one stratification plane down to another. The cataract portion of the stream way is usually through the upper part of the Edwards limestone, the rock which makes the scarp rock or mesa edge of the plateau. The flat summit, or "divide," is finally reached after climbing the rough and rocky slopes; and here the drainage is represented only by a few long, shallow caletas.

In résumé, each of these stream courses may be said to present four well-marked aspects: (1) the summit caletas, or draws; (2) the cataract portion, where they plunge from the summit through V-shaped canyons (tijeras)[1] to the flat-bottomed canyons; (3) the flat-bottomed canyons (plazas), and (4) the continuation of the dry stream bed across the lower plain.

The caletas and upper canyons are usually dry and waterless arroyos except in time of storm. The flat-bottomed canyons contain permanent pools of flowing water, fed by springs, and on the lower plain the running water disappears entirely or for a considerable distance.

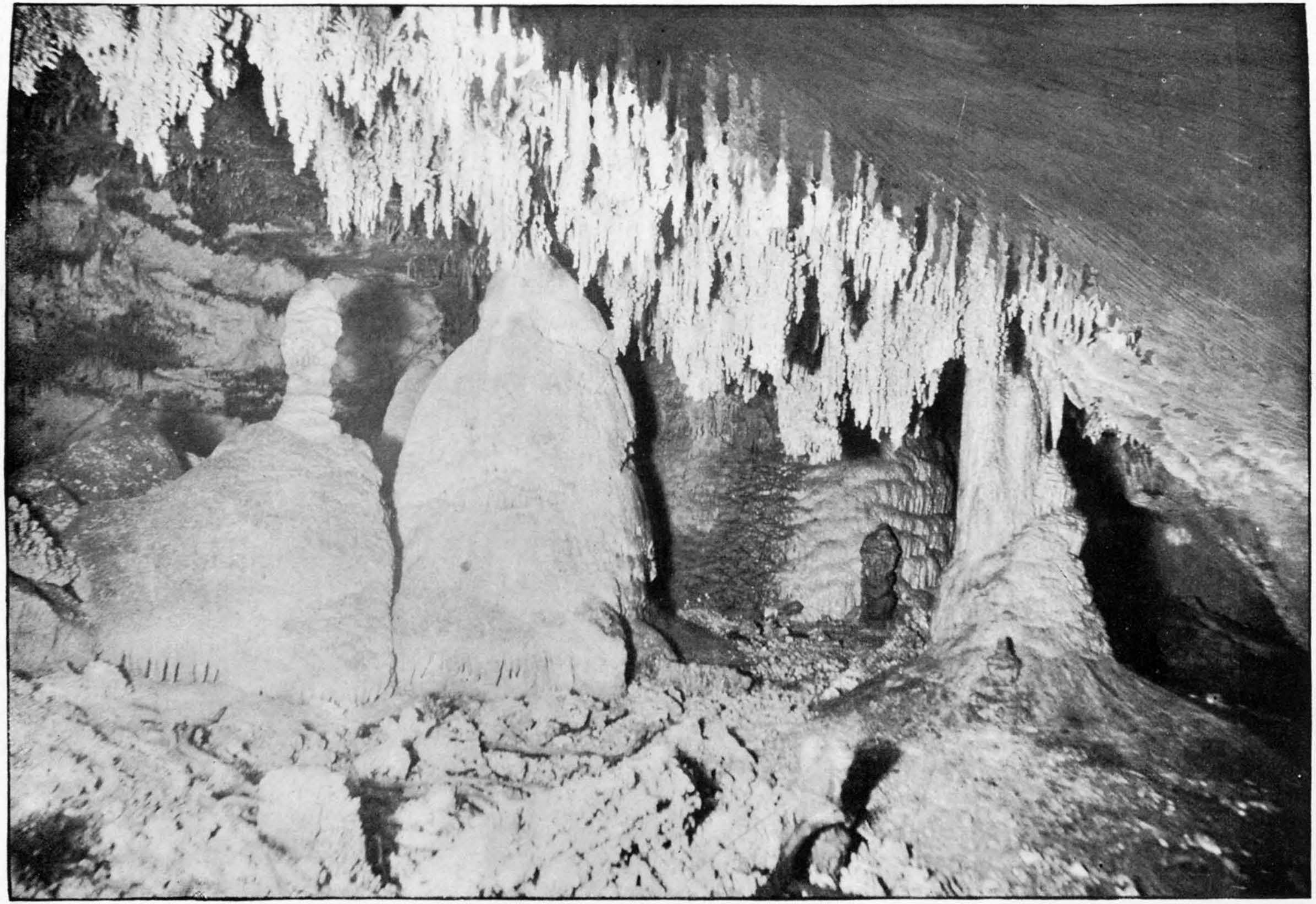
The chief laterals of these major canyons of the plateau are also peculiar and consist entirely of dry arroyos, carving the marginal canyon walls into thousands of entrant and reentrant curves, producing circular, flat-topped hills, with slopes composed of alternations of small scarps and minor slopes. Toward their heads these laterals sometimes expand into amphitheatral basins, from which minor laterals ramify in palmate arrangement. These basins are covered with the wash soil of the adjacent hillsides, and are locally known as "grass valleys." So evenly is the débris scattered over them that it is often difficult to trace the stream way. Water is found in these laterals for only a few minutes or hours after each ordinary storm, and is evaporated or imbibed before reaching the main stream, but in time of sudden and very heavy cloudburst rainfall they carry gigantic and dangerous torrents. Each torrent moves the "wash" of the land a step nearer the major stream ways, and rolls the gravel in the beds of the latter onward toward the lower country of the Rio Grande Plain. Thus it is that an intermittent stream of gravel is and has been for a long time gradually flowing away from the Edwards Plateau and spreading over the Rio Grande Plain.

There are many interesting caverns in the Edwards Plateau, and inasmuch as their occurrence, together with the general question of limestone solution, has great bearing upon the distribution of underground water, it is essential that they be briefly mentioned. They are

[1] A V-shaped canyon, from the Spanish for *scissors*.

of three general types: (1) small cavities within individual limestone strata, giving them what is locally termed a honeycombed structure; (2) open caverns occurring in certain bluff faces along the stream valleys; (3) underground caverns of vast extent dissolved out of many strata. One of the latter class occurs in Edwards County, and may be taken as a type (Pl. XXVI). It is situated just west of the McKenzie trail, about 6 miles northwest of Hillcoat's ranch. The entrance is near the summit of an oval, conical butte. The recesses, apparently undermining the whole of the hill, are elongated chambers having cross sections shaped like Norman arches. The total depth from the entrance to the bottom, as far as explored by the writers, is over 140 feet. The many chambers are lined with stalactites and stalagmites of great beauty and variety of form, and they are nearly dry, only a little water being found at the lowest depths. The method of rock solution here shown is especially interesting to students of underground water, as it gives an insight into the related problems discussed in later pages of this paper. Other caverns of a similar nature exist in nearly all the counties embraced within the area of the plateau, especially in Hays, Blanco, Medina, and Bexar counties.

The flora of the Edwards Plateau presents many peculiar variations from that of the adjacent regions, especially the Rio Grande Plain. It shows three distinct phases, viz: the phase of the stream bottom, that of the breaks, and that of the summit. The low, alluvium-filled valleys of the rivers present conditions of loose soil and constant moisture favorable to the growth of trees; hence narrow ribbons of forest are found along the streams and extending up into the semiarid region, far west of the limits of the upland forests of the humid region and dissociated from them. These embrace many species, such as the elm, chestnut oak, walnut, sycamore, cypress, live oak, and pecan. The live oaks and pecans attain great size and beauty. The occurrence of the cypress is a peculiar anomaly. This tree, which ordinarily grows only in the swamps and bayous of the low subcoastal regions, attains an enormous size at the edge of the deeper holes near the heads of permanent water of the Pedernales, Blanco, San Marcos, Guadalupe, Cypress, Onion Creek, and other streams. These localities are at altitudes from 1,000 to 1,750 feet above the sea, hundreds of miles west of the great cypress swamps of the eastern tier of Texan counties, with which they have no possible continuity. We have not noticed the cypress, however, in the Nueces, which flows from the still more arid western edge of the south side of the plateau, although the other trees usually accompanying it grow at Kickapoo water hole as large and as luxuriantly as elsewhere. These ribbons of valley forest near the headwaters of the Nueces and Frio terminate with the southern edge of the Edwards Plateau, practically ceasing at the margin of the Rio Grande Plain, and have no connections with any other forest region whatever. Accompanying them are numerous ferns, maidenhair and



HILLCOAT CAVE, EDWARDS COUNTY, TEXAS.

other delicate species, all plants requiring continuous moisture, while even the familiar mullein and Jamestown weed occur. This flora of the valleys is of interest, inasmuch as it is a modified representative of that of the great Atlantic timber belt, occurring as an isolated outlier in the semiarid region, preserved and nurtured in these valleys by the presence (due to geologic causes) of water and soil.

While the valleys support this modified flora of the humid region, the rocky slopes of the breaks between the streams and the summit present another group of vegetation—shrubby trees which prefer the crevices of rocks, or small scrubby plants which develop large, strong, and hearty roots, out of proportion to the size of the growth above ground. These plants are found along the ledges of limestone wherever their roots can find a hold, or on the almost soilless outcrop of the interstratified chalky marls. Among them are dwarf oaks, the piñon Sophora, and mountain juniper, which seem to prefer to follow the ledges of loosely jointed rock, besides many coriaceous perennials, including the agarita (Berberis)[1] and the eastern yucca. There are also many small species of Compositæ, Liliaceæ, the wild poppy (*Argemone mexicana*), and other plants growing on these slopes.

West of the Frio, in the breaks of the south end of the Edwards Plateau, where the rocks are hotter and more arid than to the eastward, the remarkable and unique resurrection flora of the limestone mountains of Mexico is found. This is characterized by plants growing upon the hot and sterile rocks, and adapted to irregular rainfall, long drought, and scarcity of soil by their thick, coriaceous parts, which ordinarily look dry and dead, but which rapidly unfold and revive after a rain, taking advantage of every drop of rainfall in order to store sufficient moisture to enable them to survive the long periods of intervening drought. Even the ferns, mosses, selaginellas, and kindred plants, which ordinarily constitute our ideals of delicate and tender herbage, have in this region a thick, leathery texture. On the almost barren surface of these rocks grows the melon-shaped, edible cactus—the devil's pincushion.

Here also, for the first time in proceeding westward across Texas or northward across the Rio Grande Plain from Mexico, one meets the peculiar plants (agaves and yuccas) which become so marked a feature farther westward, the serrated sotol, with its flower stalk rising to a height of 15 feet; the dagger-like lechuguilla or ixtle plant, and several species of yucca not seen farther eastward.

The flora of the summit of the plateau is radically different from that of the breaks or of the valleys. All trace of shrubs or trees disappears, save here and there a patch of shin oaks and dwarf evergreens, and in time of verdure the eye beholds apparently a never-ending sea of grass,

[1] In the lower slopes, around the headwaters of the Frio (Frio water hole), the writers last year discovered a large area of the edible *Berberis swaseyi* Buckley, a species which Coulter says in his Flora of West Texas is known only from the canyons of the Pedernalis. The fruit of this plant is a large edible berry well worthy of cultivation. We have never seen it elsewhere.

through which appear many bright-colored flowers. This is the southern end of the flora of the Great Plains region, which continues far northward.

Although there is much agriculture in the wide, fertile, plaza canyon valleys indenting the plateau, especially in Blanco, Gillespie, Comal, and Kendall counties, the slopes and summits constituting the larger part of the area are not adapted to agriculture, owing to the rocky character of the soil, the semiaridity of the country, and the impossibility of irrigation. So little are they fitted for agriculture that the extent of the summit of the Edwards Plateau can almost be traced upon the map by the scarcity of post-offices and other evidences of population. They constitute, however, good grazing country and support many large sheep and cattle ranches.

## GENERAL PRINCIPLES OF ARTESIAN WATERS.

The rocks of the earth form a system of natural works by which water is collected, stored, and distributed. They constitute the basins, reservoirs, conduits, and other portions of the plant for retaining and distributing underground water.

The details of natural waterworks differ in different places in the same way that the details of artificial systems differ in different plants, but the distribution in both is governed by the common principles of hydrostatics. The efficiency of a natural system is determined by the texture of the rocks and the geologic structure of the region, so that an understanding of the availability of underground water in any region necessitates a knowledge of the elementary geology of that region.

It is neither convenient nor advisable here to discuss minutely the source, storage, and distribution of underground water, but in order that the subject may be understood, we shall give a brief explanation of the elementary principles governing its occurrence.

(1) The primary source of all underground water is the rainfall.

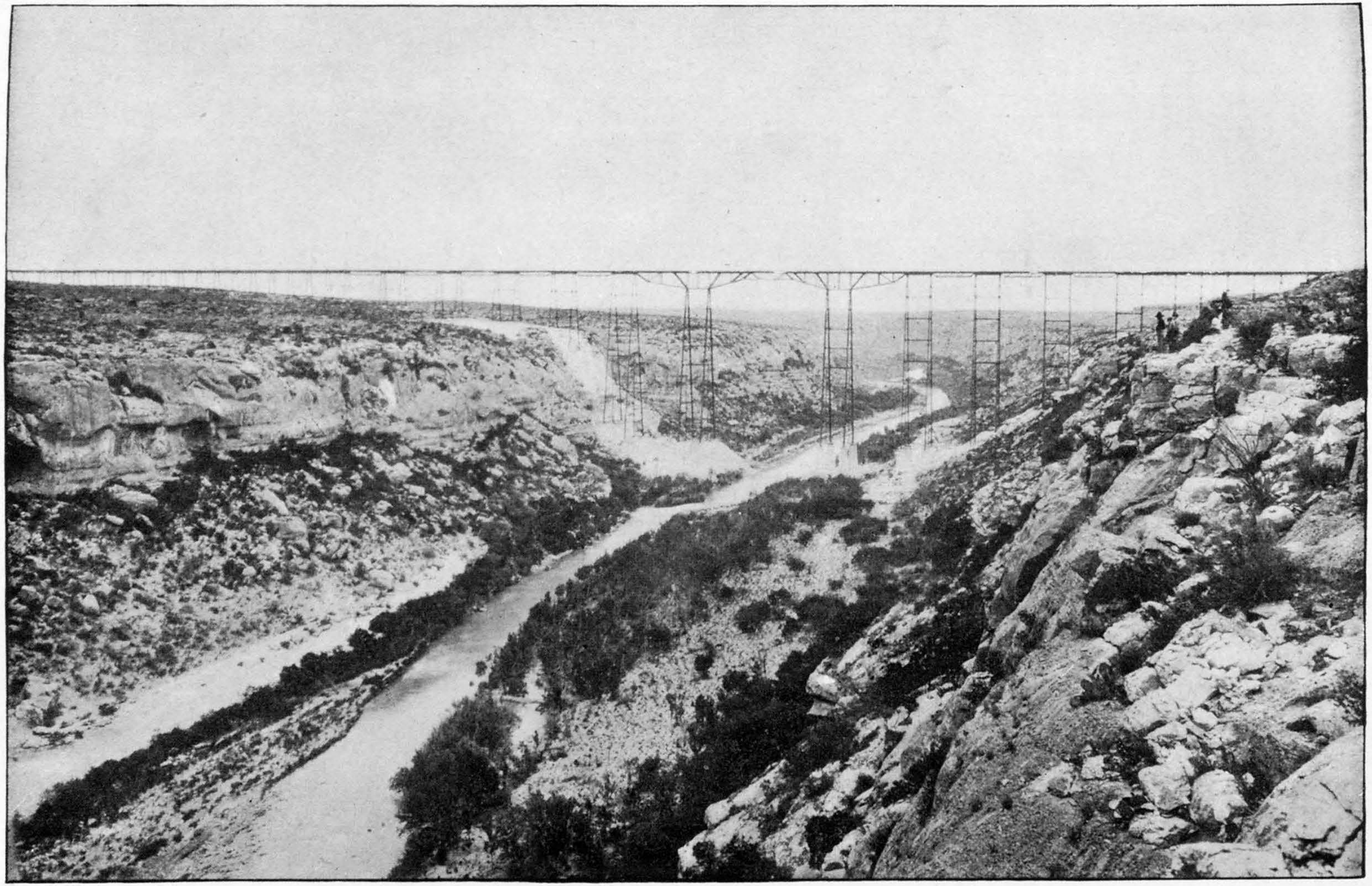
(2) Rocks imbibe water. Imbibition may take place by percolation and by absorption. By percolation is meant the process by which water proceeds through cavities, cracks, fissures, or other breaks in the continuity of the underlying rocks. Absorption takes place when water enters the small interstices of the rock.

(3) Water can flow in rocks, and the rapidity of this flow is known as the capacity for transmission.

(4) Different kinds of rock have different capacities for imbibing and transmitting water. These capacities are not proportional to one another, but vary independently, according to the kind of rock.

(5) All water entering the earth tends to gravitate downward along lines of least resistance (easiest transmission).

(6) The only ordinary agency by which rock sheets may be naturally drained of their water is gravity. Gravity drainage may be of two kinds, direct or artesian. (*a*) Direct drainage is that by which water



EDWARDS PLATEAU AND PECOS CANYON AT THE SOUTHERN PACIFIC RAILROAD BRIDGE.

The canyon has been cut into the Edwards limestone.

escapes at the bottom of a hydrostatic column. If sheets, beds, or masses of rock containing water be cut into by a well-hole, a ditch, erosion, or other means, to a lower level than that at which the water entered, all the water contained in the rocks above the level of the incision will be drained from the rocks by the simple process of seepage, just as water escapes from the bottom of a moist sponge or from the faucet at the bottom of a barrel. This is the method by which water occurs in ordinary nonartesian or dug wells, and in seepage springs so frequently found along the low banks of rivers, valleys, or other natural incisions into the strata. Modern agricultural field subdrainage depends upon the same principle. (*b*) Artesian drainage is that in which the water rises above or escapes from a level higher than the bottom of the hydrostatic column. The simplest illustration of this principle is the equilibrium maintained by water in the two limbs of a U-shaped tube. Water from high reservoirs carried by pipes to the upper stories of buildings also illustrates this principle. Artesian water will not rise higher than its head or source, and its maximum pressure varies with the difference between the altitude of the head and that of the outlet.

Water which falls upon the surface of the earth as rain is disposed of by surface run-off, by evaporation, and by absorption into the underlying rocks. The water forming the run-off passes over the surface to form streams and lakes; that which is evaporated passes again into the atmosphere; the remainder sinks below the line of evaporation into the rock mass beneath, supplying wells and springs. The water of the last class forms the subject of this paper.

## CAPACITY OF ROCKS FOR ABSORBING MOISTURE.

Rocks of open texture, such as loose sands and sandstones, gravels, and chalk, have a spongelike capacity for imbibing water. Water poured upon sand will quickly disappear by imbibition. If we wish to filter water, we run it through beds of sand or gravel. Bricks are sprinkled before they are put into buildings, and they absorb from 20 to 60 per cent of their weight of water. On the other hand, if one wishes to shed water or otherwise prevent its percolation, one constructs roofs of tile, makes tables of marble, and builds tanks and cisterns of cement or clay. Few stop to consider, when thus using rocks, that they are making practical application of the broad principles which control the occurrence of underground water. By careful and accurate experiments, such as anyone can make, the capacities of all the known rocks for the imbibition and transmission of water have been determined, and it is shown that sandstone or chalk will absorb many times as much water as slate, marble, or granite. These facts can be observed after every rain. If the rain falls upon loose, sandy soils, like those of the two Cross Timber regions of Texas, it quickly disappears by absorption; if, on the other hand, it falls on the clay soils of the prairies, it stands for some time in pools, such as those called "hog wallows" in Texas.

The capacity of rocks for the transmission of water is entirely different from their capacity for imbibition. If one could construct of sand, clay, slate, granite, chalk, and close-textured limestone filtering vessels of equal capacity, and then fill them with water, one would find diverse results, illustrating the capacity of these rocks for transmission of water. Water would pass so slowly through the close-textured limestone, slate, and granite that the quantity filtered would be practically imperceptible. At first the sand and chalk would drink in the water equally fast, but after complete saturation it would require longer time for the water to percolate through the chalk than through the sand.

The distribution of underground water is dependent upon the arrangement of rocks in sheets or strata. By a simple arrangement of the porous sands between impervious materials, nature has constructed reservoirs and conduits for the retention and distribution of water, and it is by the character of the arrangement of the rock sheets that the negative or positive conditions for the procurement of artesian waters are determined.

A stratum usually consists of two related parts—the outcrop and the embed. That portion exposed at the surface of the earth is the *outcrop*, and that portion which is concealed underground beneath and between the other rocks may be termed the *embed*. The outcrop of a water-bearing stratum constitutes its main catchment or receiving area, and the embed constitutes the storage reservoir.

If the water-bearing rock sheet is inclosed between impervious beds, and inclines beneath the surface, the water will be conducted to a lower level than the outcrop and will remain stored in the earth under hydrostatic pressure until an outlet is provided for it. A water-impregnated stratum embedded in this manner is an artesian reservoir. The water-bearing strata, together with the impervious strata beneath and above them, constitute an artesian system.

When water is conducted downward to a lower level by an embedded stratum, it acquires a tendency, due to hydrostatic pressure, to rise higher than the overlying retaining bed. When the beds overlying such a water-bearing stratum are penetrated by an opening, artificial or natural, the contained water will rise through the overlying bed. Such earth waters which rise under the influence of hydrostatic pressure through an opening to a higher level are known as *artesian* waters. Hence an artesian well may be defined as one in which the water rises by means of hydrostatic pressure above the top of the embed. If artesian waters rise to the surface, they are known as flowing artesian wells; if the water fails to rise to the surface in a well, such a well is known as a nonflowing artesian well. The height to which the water will rise depends upon several conditions. If the water-bearing stratum is embedded and incised at only one place down the dip from its surface outcrop, the water at the place of incision will rise to the same level as the line of complete saturation of the bed. This level is somewhat lower than that of the lowest surface outcrop. If there are several incisions, the question is not entirely one of static equilibrium. For instance, if water is naturally received by the stratum in one part of its outcrop and discharged from another lower part, the conditions in the intervening area are those of dynamic rather than static equilibrium, and the water will rise in a well to a level intermediate between the receiving and discharging levels.

The artesian water-bearing strata of the State east of the Pecos River are composed mostly of extensive sheets of sands, clays, and limestones, succeeding one another in orderly arrangement, except along the Balcones zone of faulting, and in general having a gentle inclination toward the sea, so that in traveling northwestward, although constantly ascending in altitude, one encounters the outcropping edges of rock sheets of lower and lower stratigraphic position. This produces the simple arrangement of a tilted plain built up of a series of alternately impervious and pervious layers. The rain falling upon the outcropping edges of the latter, sinks into the embed, and by gravity is conducted seaward down the plane of its inclination to lower levels beneath the surface. Each different stratum, including any particular water-bearing stratum, becomes embedded deeper and deeper to the southeastward of the point where it outcrops at the surface. This structure is very simple and its detail can be traced out, measured, and mapped as accurately as that of the successive layers of stone in a building. To do this is the work of geologists, and it is not a matter of speculation and hypothesis, but simply the application of specially acquired knowledge, similar to that which one must possess to be proficient in any profession.

The foregoing principles are all applicable in explaining the numerous artesian wells and springs of the Rio Grande Plain and the headwater springs of the Edwards Plateau, but the explanation requires also an understanding of the order and arrangement of the strata, and we therefore proceed to consider the geology of the region.

## GEOLOGY OF THE REGION.

The rocks of this region consist of various beds of marl, clay, limestone, sandstone, etc., a few igneous rocks (proportionally a very small part of the whole), and some alluvial or surface deposits derived by erosion from the older formations. With the exception of the igneous and alluvial rocks, the strata are all composed of material which was laid down beneath the surface of the ocean, and embedded in them are found many marine fossils, the remains of animals which inhabited the waters of the old ocean. These fossils are of the greatest value in determining the geologic position of the beds containing them, and therefore we give from each of the principal beds a few illustrations of the most important and most abundant forms, by the aid of which the layman can readily recognize the geologic horizons. (Pls. LI–LXIV.)

The nomenclature used in describing these rock sheets is as follows: The geologic unit is the *bed* or *stratum*. Series of lithologically and paleontologically related strata are grouped into *formations*, as the "Glen Rose formation." When the strata are all of one kind of rock the name of the rock is sometimes substituted for the word formation, thus: "Edwards limestone" or "Edwards formation." Groups of this class may be embraced in still larger groups called *divisions*, as the "Fredericksburg division," and these may be grouped into *series*, as the "Comanche series," and these compose the still larger geologic groups, such as the Cretaceous or the Eocene.

As the work of geologic investigation has progressed new facts have made necessary the revision of classification, and new names have from time to time been substituted for those first employed. In the classification herein presented the previous nomenclature has been refined by substituting appropriate geographic names (mononyms where possible) for all formations and abandoning the use of paleontologic and mineralogic names. Thus the terms "Dinosaur sands," "Caprina limestone," "*Exogyra texana* beds," "*Exogyra arietina* clays," "Fish beds," *Exogyra ponderosa* marls," "Hippurites limestone," etc., have all been replaced by appropriate geographic names. Even some geographic names originated in earlier writings have been abandoned because found to duplicate names previously used for different formations elsewhere.

The sequence and systematic classification of the formations of the region under discussion are presented in the table below, and in the following pages the several formations will be individually described. They will be taken up in the order in which they were deposited, the oldest and lowest first.

TABLE OF GEOLOGIC FORMATIONS OF THE REGION UNDER DISCUSSION.

RECENT.

Wash deposits of the hillsides, stream-bed material, etc.

PLEISTOCENE.

Onion Creek marl, Leona formation, and other terrace deposits.

PLIOCENE.

Uvalde formation.

EOCENE.

CRETACEOUS.

*Gulf series.*

| Formation | Division |
|---|---|
| Webberville and Eagle Pass formations | Montana division. |
| Taylor and Anacacho formations | Montana division. |
| Austin chalk | Colorado division. |
| Eagle Ford shales | Colorado division. |

*Comanche series.*

| Formation | Division |
|---|---|
| Shoal Creek limestone | Washita division. |
| Del Rio clay | Washita division. |
| Fort Worth limestone | Washita division. |
| Edwards limestone | Fredericksburg division. |
| Comanche Peak limestone | Fredericksburg division. |
| Walnut formation | Fredericksburg division. |
| Glen Rose formation | Trinity division. |
| Travis Peak and allied formations | Trinity division. |

## PALEOZOIC FORMATIONS.

Below the Cretaceous rocks which constitute the surface of the Edwards Plateau and Rio Grande Plain lies a foundation of the older and different rocks of the Paleozoic system. These are exposed by erosion only along the foot of the northern scarp of the plateau in Burnet, Blanco, Gillespie, and Mason counties. So far as known, no drill has as yet penetrated them beneath the Rio Grande Plain, and as they are so deeply buried they need little consideration in the present discussion. In the Edwards Plateau, however, the well drill has reached the Paleozoic rocks beneath the Cretaceous in several places, notably at Kerrville and at Morris ranch, in Gillespie County. From data derived from these drillings we are led to believe that the horizontal Cretaceous beds of the Edwards Plateau lie upon an uneven floor of the older rocks. The probable relation of these Paleozoic rocks to the overlying Cretaceous is shown in the figure of the wells at Kerrville (fig. 68, p. 270). One hundred feet of Carboniferous rocks form the base of the section exposed at the mouth of Hickory Creek on the Colorado River (see p. 220).

## THE CRETACEOUS.

The Cretaceous formations of Texas are by far the most important in the State, in both areal extent and economic value. They are sea-made rocks, and rest unconformably upon the older Paleozoic rocks, which formed the ocean floor when their deposition began. They constitute almost the entire surface of the Edwards Plateau and much of that of the Rio Grande Plain. Their southern margin is covered by the Eocene overlap, and they are overlain in places by superficial deposits of gravel of Neocene and Pleistocene age. They contain all the artesian water described in this paper, and for this reason it is important that their sequence and occurrence be well understood. They likewise supply the most valuable building material—stone, lime, and cement—and some of them contain oil and gas.

They are mostly limestones and clays, sometimes containing slight admixtures of very fine silica, but there are also great beds of sand and sandy mixtures, occurring principally at the base, middle, and top of the group. The limestones are of many kinds, predominantly light-colored, more or less chalky in texture and composition, and rarely of the hard, ringing, close-textured character, such as is usually met in American Paleozoic limestones. Some of them are very porous, even cavernous, in texture; others are more massive; some have slight admixtures of magnesia, clay, or silica; others are almost pure carbonate of lime. In some instances they are soft, often marly; in others, quite indurated. The clays are likewise very chalky, so that on exposure to air they readily crumble and weather into soils. The many beds of clay and limestone resemble one another so much that it is difficult for one not accustomed to making geologic observations to discriminate them. The geologist who has studied them carefully can readily recognize them, not only by slight lithologic differences, but by the peculiar fossils which characterize them.

The many formations which make up the Cretaceous system are classified into two great series, the lower of which has been termed the Comanche and the upper the Gulf series. They have also in various parts of Texas been classified into smaller groups called divisions. As most of the formations are of limited extent as compared to the whole system, these divisions are of importance for the indication of the chronologic relations of formations occurring in different regions, and they have therefore been indicated in the table on page 216, but they are not essential to the discussion of the economic questions of this paper.

### COMANCHE SERIES (LOWER CRETACEOUS).

In southern Texas this series includes the greater part of the Cretaceous system, and its total thickness is somewhat more than 1,500 feet. At the base the beds are of feebly coherent sand, then come marls and limestones in alternation, and then a heavy body of limestone, followed by other marls and shales.

#### THE BASEMENT BEDS.

The basement beds of the Comanche differ from the overlying formations of the series by the fact that they are usually composed of sands instead of calcareous marls and limestones. These are more or less fine grained and slightly compact, accompanied in places by small pebble conglomerate, and locally varying in composition according to the material of the adjacent rocks from which they were derived. Thin beds and laminæ of clay occur in the sands, and the residual surface usually has a reddish color. Furthermore, the outcrop of these beds usually bears a growth of timber such as is found in the region known as the Upper Cross Timbers. While the basal beds of the local Cretaceous section, wherever exposed, are usually of this sandy nature, these beds are not everywhere of exactly synchronous deposition, for they represent the littoral of a sea which was progressively transgressing northwestward across an uneven land during the whole Comanche epoch.

So far as the region considered in this paper is concerned, the lower beds and their relations to the overlying and underlying rocks are exposed chiefly in the slopes of the Colorado drainage in western Travis, eastern Burnet, northern Hays, Gillespie, and other counties along the northern edge of the Edwards Plateau, and especially along a line between the town of Burnet and Travis Peak post-office, via Smithwick

Mill. There is no doubt, however, that these formations occur embedded beneath most of the Edwards Plateau and Rio Grande Plain, and that if a drill could penetrate far enough it would reach them anywhere beneath these regions. They exhibit such variation from place to place that the precise correlation of different occurrences is difficult, and it has seemed best to give separate names to the rocks at the two most important localities.

### TRAVIS PEAK FORMATION.

These beds are especially well displayed on both slopes of the valley of the Colorado River, between the mouths of Sycamore and Cypress creeks, in Burnet and Travis counties. The name they bear was given them because they are well exposed in the vicinity of Travis Peak post-office.[1]

While they are arenaceous in composition and porous in texture, like the basement beds of the Comanche in general, they differ considerably from the allied beds to the northward. They consist of conglomerate, composed of coarse rounded pebbles of Silurian and Carboniferous limestones, granite, Llano schists, quartz derived from the adjacent Paleozoic rocks, beds of finely cross-bedded pack sand, white siliceous shell breccia resembling the Florida coquina, and some clay.

At the base is usually conglomerate. Succeeding this is coarse, angular, cross-bedded sand, which becomes more finely triturated until it reaches the condition known in Texas as "pack sand"—i. e., a very fine-grained, loosely consolidated sand, cemented by carbonate of lime. In the sands are occasional patches of red and greenish-white clays, resembling very much the characteristic colors of the Potomac beds of the Atlantic coast, and they are sometimes accompanied by lignite and fossil bones.

The following section[2] by Mr. J. A. Taff will give an idea of the sequence and composition of the formation as exposed in the valley of the Colorado, in the locality between Travis Peak post-office and Smithwick Mill, Burnet County:[3]

SECTION NO. 1.—*Hickory Creek section of the Travis Peak formation, beginning at the top of the divide between Hickory and Cow creeks and continuing to the Colorado River level at the mouth of Hickory Creek, Burnet County.* (Fig. 56 B, p. 223.)

| | Feet. |
|---|---|
| 13. Bands of conglomeratic and calcareous sandstone, alternating with beds of arenaceous limestone, the arenaceous limestone predominating | 40 |
| 12. Marly magnesian limestone | 40 |
| 11. Calcareous sand at base, grading upward to a siliceous limestone at the top, barren of fossils | 55 |
| 10. Yellow calcareous sand, stratified | 15 |

[1] See Annual Report Geol. Survey Texas, Austin, 1890, p. 118.

[2] All sections given in this paper are described from the top down and are numbered from the bottom up.

[3] Third Annual Report Geol. Survey Texas, Austin, 1892, p. 295.

| | Feet. |
|---|---|
| 9. Conglomerate similar in character to No. 2, with the exception that the pebbles are smaller and more worn, grading into sand below and into calcareous sand above | 25 |
| 8. Red sand, unconsolidated | 3 |
| 7. Friable yellow sand | 5 |
| 6. Cross-bedded shell breccia, containing many small rounded grains and pebbles of quartz, flint, and granite sand. Fossils: Trigonia and small bivalves, and *Ammonites justinæ* | 7 |
| 5. Ostrea beds, magnesian lime cement, fossils en masse | 3 |
| 4. Brecciated grit, composed of worn fragments of oyster shells and shells of other mollusca, with sand and fine pebbles, stratified in false beds | 5 |
| 3. Bands of friable bluish shale and calcareous sand, stratified. Fragments of oyster shells are common in the calcareous sandstone | 15 |
| 2. Basal conglomerate of pebbles of limestone, quartz, chert, granite, and schist, well rounded, in a cement of ferruginous yellow and red gritty sand. Some of the pebbles at the base are from 4 to 6 inches in diameter. They decrease in size, however, upward from the base, until we obtain a false-bedded calcareous shell grit at the top | 50 |
| Total thickness of Travis Peak beds | 263 |
| 1. Laminated, flaggy, carboniferous sandstones and friable light-blue clay of the Carboniferous (Coal Measures) formation, from the Colorado River level upward to the base of the Trinity conglomerate, the laminated sandstones containing prints of ferns; nearly | 100 |
| Total thickness of section | 363 |

The sandstone contains grains of silica from the size of a pea to the most minute particles, and small subangular fragments of clay in the cement of lime.

Fossils occur in these beds as low down as the contact conglomerates, but they are neither plentiful nor distinct. The upper or coquina-like beds are full of casts and molds, among which are undetermined species of Trigonia, Pholadomya, and Cyrena, and *Ammonites justinæ*.

In these beds also appears the first of the several oyster agglomerates of the Comanche series. This is composed of a solidified mass of large oyster shells, forming a stratum 7 or 8 feet in thickness, just below the junction of Post Oak and Cow creeks.

Accompanying the oyster breccia another noteworthy feature of the Trinity division appears—i. e., an excess of epsom salts, or magnesium sulphate. The oyster-shell bed effloresces into a powdered earthy substance accompanied by the epsom salts. Magnesian and pyritiferous layers occur in other horizons higher in the division, and their presence is no doubt in part the cause of the mineral character of some of the artesian waters, especially those wells which are not drilled into the basement sands below these layers.[1]

At the top of the sandy beds in the Colorado section a yellow, arenaceous, fossiliferous limestone appears. This marks the first or lowest appearance of the peculiar fossils Monopleura (Caprotina) and Requi-

[1] See artesian water discussion, p. 300.

enia, and indicates the beginning of the conditions which finally produced the Glen Rose formation.

The Travis Peak formation records a subsidence of the land which was taking place during its deposition. As the waters deepened the deposits changed from coarser to finer material, becoming more comminuted and calcareous at the top of the beds, until the sand grains are so fine as to be almost imperceptible to the eye, the whole mass becoming quite chalky and "magnesian" in appearance.

### GILLESPIE FORMATION.

In Gillespie County are other exposures of the basement beds of the Cretaceous, resting upon the Paleozoic. These appear in the valley of Barron Creek at Fredericksburg and along the Pedernalis below that town. They are composed of sandy grits and clays of a more brilliant vermilion hue than is ordinarily met with in similar deposits elsewhere. As shown in fig. 55, and in fig. 74, p. 314, they rest on the Paleozoic and grade up into the upper beds of the Glen Rose formation. These might possibly be correlated with the basal sands of the Travis Peak formation, but they are more probably the stratigraphic equivalent of the lower portion of the Glen Rose formation where it rests on the Travis Peak.

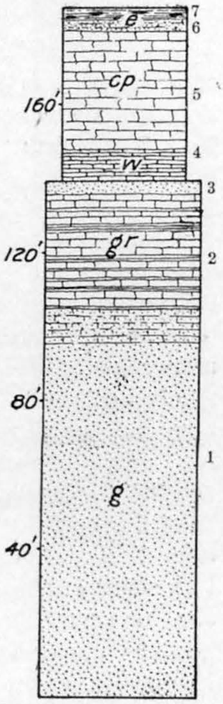



FIG. 55.—Section at Fredericksburg. *e*, Edwards; *cp*, Comanche Peak; *w*, Walnut; *gr*, Glen Rose; *g*, Gillespie.

SECTION NO. 2.—*A butte a few miles north of Fredericksburg.* (Fig. 55.)

| | Feet. |
|---|---|
| Edwards: | |
| 7. Limestone with flint | 3 |
| 6. Fine calcareous sands | 10 |
| Comanche Peak and Walnut: | |
| 5. Limestone with *Exogyra texana*, *Gryphæa marcoui*, etc | 40 |
| 4. Yellow calcareous clay with many *Exogyra texana*. | 10 |
| Glen Rose: | |
| 3. Soft yellow sandstone | 5 |
| 2. Limestone, sandy at base | 50 |
| Gillespie: | |
| 1. Cross-bedded vermilion-red sands | 120 |

### GLEN ROSE FORMATION.

This consists largely of beds of flaggy argillaceous or massive chalky limestones, alternating with thin strata of marly clay, white and yellow in color. From this alternation of limestone and clay the beds were provisionally called the "Alternating beds" when first differentiated by the senior author.

The indurated beds consist of white and yellow limestones of either brecciated, crystalline, arenaceous, magnesian, or chalky structure. They are of varying thickness and of great uniformity in extent. They are separated by softer, unconsolidated, slightly argillaceous marls of oolitic structure, sometimes white and sometimes yellow.

The Travis Peak formation grades upward into the Glen Rose without break in the sedimentation, as can be seen in the high bluffs of Cow Creek, immediately below Mr. Hensel's house at Travis Peak post-office, in the western part of Travis County.

The basement beds of the Comanche series have been described as being predominantly of an arenaceous character. The Glen Rose formation may be distinguished as being essentially calcareous. Each of the subdivisions, however, is accompanied by the material of the other as accessory. Thus the calcareous Glen Rose formation is slightly arenaceous at its base, the arenaceous material being siliceous grains so triturated that they are reduced to an almost impalpable powder. This gradually diminishes as we ascend in the beds, and the lime and clay proportionately increase.

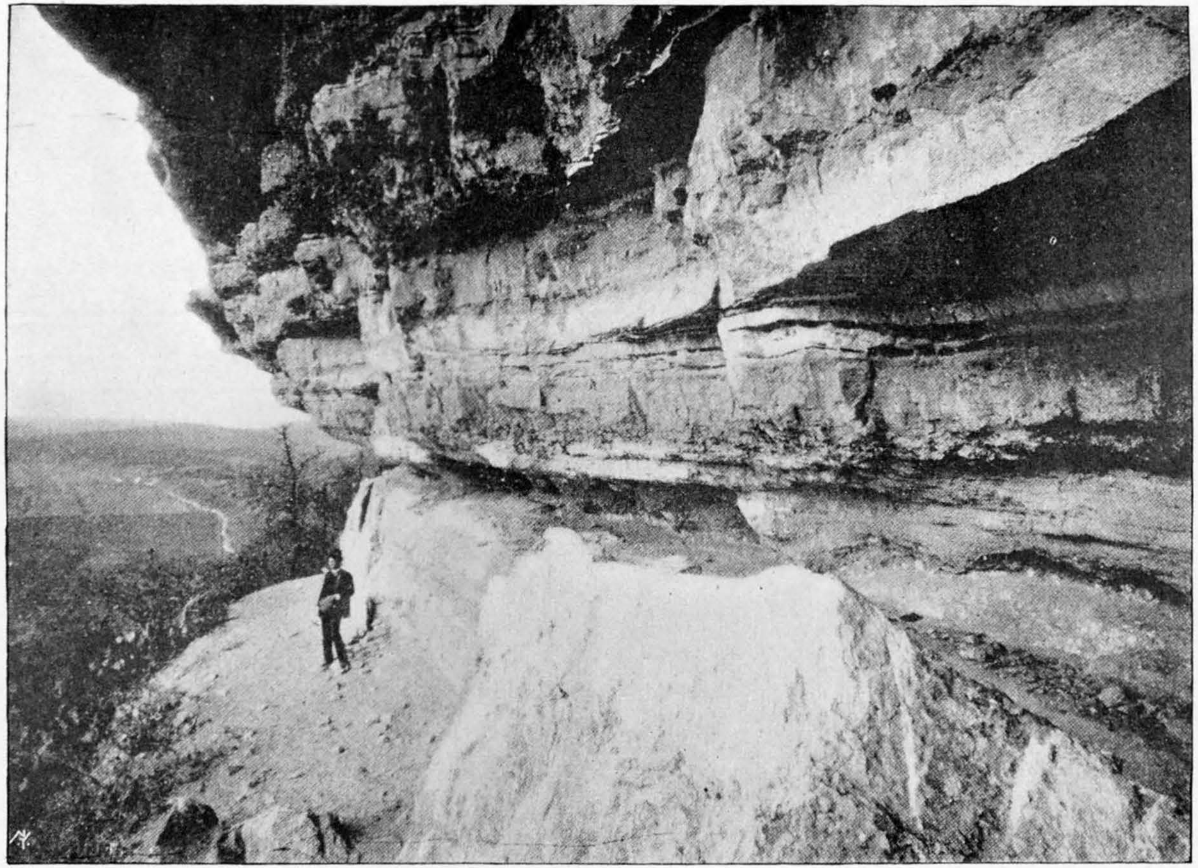
The lowest Glen Rose beds are marked by the appearance of strata of homogeneous texture, such as "magnesian"[1] marls and hard layers in which the fossil Requienia[2] occurs. The name "Caprotina Horizon No. 1" has been applied to these beds, because in the earlier geologic literature the fossils now called Requienia were termed Caprotina. The top of the Glen Rose formation is just below a bed of yellow marl which is persistent over a great area in central Texas, and is the culminating horizon of the oyster *Exogyra texana*, after which this bed has been called.

In nearly all complete sections the Glen Rose formation shows three marked subdivisions. The lower and upper thirds are composed of thinly bedded and alternating marl and flags, usually weathering into terraced slopes; the middle third is made up of thicker and more massive beds which constitute bluffs. Some of the beds near the base of the thicker layers are quite chalky in texture and carry many peculiar fossils, especially noteworthy being a large foraminifer (*Orbitolina texana*), besides many large casts of mollusks. The lower portion of the formation carries much fine arenaceous material accessory to the calcareous material and its indurated and unindurated beds do not occur in such uniform alternations as do those of the upper third. For instance, there will be 10 or 12 feet of soft, friable material and then a thin layer of less than a foot of indurated stone. In weathering this results in wide terraces with steep slopes.

The yellow magnesian strata also increase in thickness in ascending series, and become very conspicuous in the middle portion, often being from 5 to 15 feet in thickness, as seen in the bluffs of Mount Bonnel, near Austin. These magnesian limestones are soft and of a cream or brownish-yellow color, and alternate with strata of marls similarly con-

[1] The term magnesian has long been applied to certain yellow strata in these beds. Whether they are or are not magnesian in composition we can not state positively.

[2] A species of this genus of shells from the Edwards limestone is figured on Pl. LIV, fig. 1 *a*, 1 *b*.



UPPER GLEN ROSE BEDS OF MOUNT BONNEL; BLUFF OF COLORADO RIVER.

stituted, and are sometimes accompanied by pockets or nodules of calcite, aragonite, strontianite, celestite, and epsomite.

The upper third of the formation, as seen at the top of Mount Bonnel, presents alternations of friable marls and hard limestone strata. The limestone strata usually average less than a foot in thickness. These alternations occur with great regularity and persistence. Clay is the chief accessory of the calcareous beds. The marls are soft and laminated and are composed largely of minute shell fragments, giving the beds a distinctly granular, oolitic character. They have little clay and imbibe the moisture very freely.

While possessing no great agricultural possibilities, the basal or alternating beds are capable of producing valuable building material, among which are building stones. Some of these have rich "magnesian" buff-yellow colors, while the limestones often resemble the stones of Caen, France, which are imported into this country. Some of the beds are also valuable for the manufacture of hydraulic cements, although at present they are not utilized. These rocks also contain undeveloped beds of epsom salts, strontianite, and other materials.

The alternations of horizontal beds of soft marls and hard limestones above described produce the bench and terrace topography of the slopes of many of the canyons and along the margin of the Edwards Plateau from the East Fork of the Nueces to the Colorado, where the streams have cut downward through

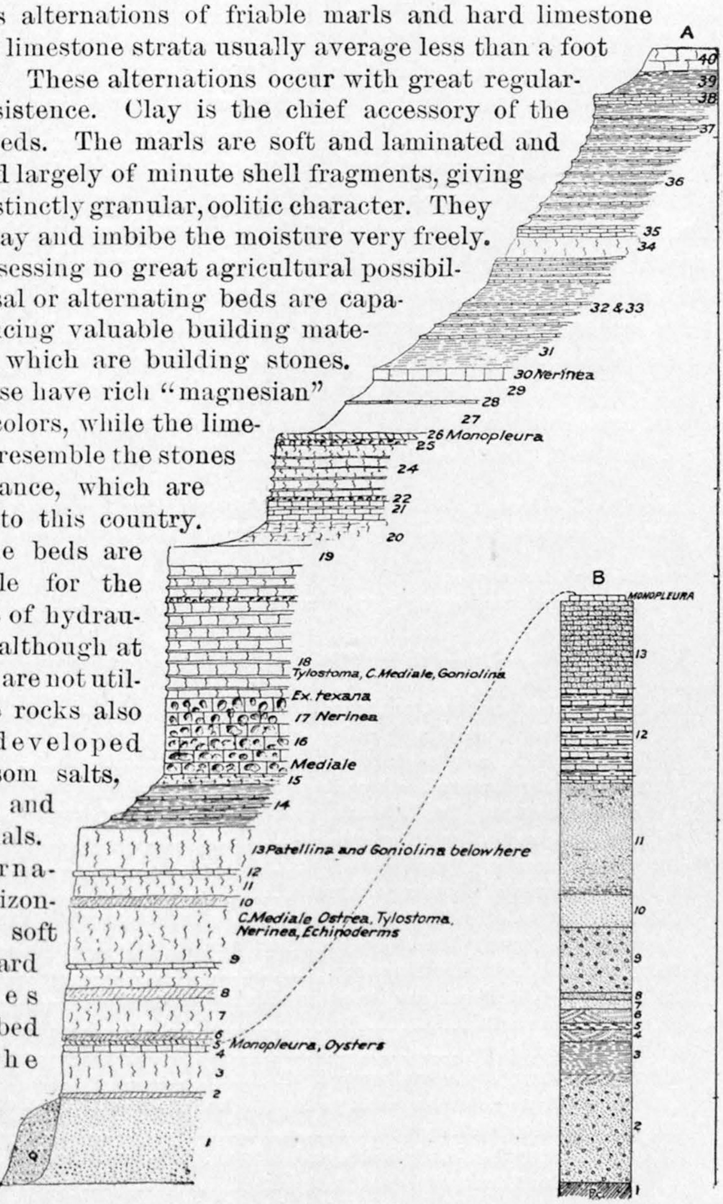



FIG. 56.—Sections at Lohmann's Crossing and Hickory Creek, Colorado River. A, section from top of high hill south of Round Mountain, Travis County, northward to Lohmann's Crossing; B, section of Travis Peak beds at the mouth of Hickory Creek; Q, river terrace; Pc, Carboniferous. For numbers, see Section No. 3, pp. 224-225.

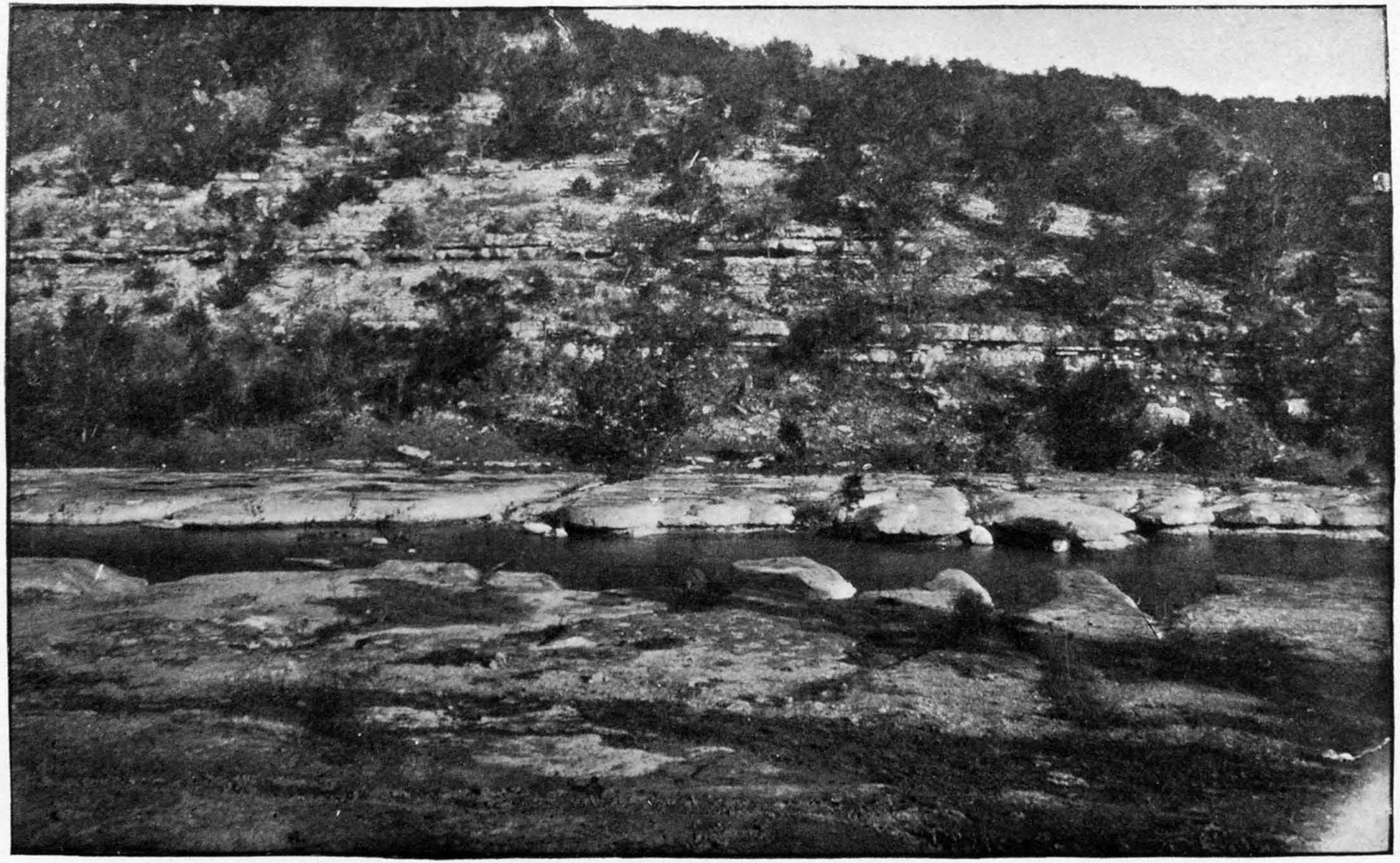
the Edwards limestone. Of this character are the beautiful terraces and bluffs of the Colorado, as seen from Mount Bonnel westward to the Burnet County line, as well as those of the Pedernales, Guadalupe, Comal, Medina, Hondo, Frio, and all the numerous streams indenting the southern margin of the plateau. In addition to the localities already mentioned the beds are exposed west of San Antonio along the line of railway between Aue and Boerne and in the valley of the Guadalupe above Kerrville as far as Vix. They are also exposed in the lower slopes of the valley of the East Nueces and its tributaries below Vance to a point a few miles south of Montell. The channel of the West Nueces has barely cut down to their top and exposes them at only two places along the stream bed—at Kickapoo springs and in the north bend of the river in northern Kinney County.

The accompanying detailed section of the entire thickness of the beds of the bluffs of the south side of the Colorado in the vicinity of Round Mountain, Travis County, is typical of these beds. It coincides almost exactly with Mr. Taff's[1] Sandy Creek section, previously measured, on the opposite side of the river.

SECTION NO. 3.—*From top of high hill south of Round Mountain, east of road from Bee Caves to Lohmann's Crossing of the Colorado River, to Lohmann's Crossing.* (Fig. 56).

| | Feet. |
|---|---|
| Comanche Peak limestone: | |
| 40. Limestone breaking easily; some firm slabs at top | 5 |
| Walnut formation: | |
| 39. Clays with large *Exogyra texana;* forms a shelf | 10 |
| Glen Rose formation: | |
| 38. Shaly limestone; not very fossiliferous | 10 |
| 37. Alternating harder and softer strata of limestone; some thin slabs about base; not fossiliferous | 15 |
| 36. Alternating hard and soft yellowish limestone; not very fossiliferous | 35 |
| 35. Shaly limestone, fossiliferous, contains a few individuals of *Cardium mediale* and a few other species | 4 |
| 34. White limestone; breaks easily | 15 |
| 33. Marly material, forming a terrace | 10 |
| 32. Alternations of soft argillaceous or marly limestone with harder thin layers of purer limestone (four hard and three soft layers) | 30 |
| 31. Slope and shelf; fossils at top | 15 |
| 30. Hard nodular limestone; contains Nerinea fragments | 5 |
| 29. Slope and shelf | 14 |
| 28. Thin, hard ledge | 1 |
| 27. Slope; very gentle—rather a shelf | 15 |
| 26. Bed of Monopleura in hard, yellowish limestone—thin, a foot or two. | |
| 25. Hard perforated limestone | 2 |
| 24. Alternating thin hard layers and soft thick layers; the thin layers 6 inches to 1 foot, the soft 3 to 4 feet | 20 |
| 23. Soft, chalky, argillaceous stuff, only a few feet. | |
| 22. Ledge of hard yellowish perforated limestone, 2 feet; hard ledges of limestone, 8 feet | 10 |
| 21. Small hard ledge, 1 or 2 feet. | |

[1] Mr. Taff's section measured 447 feet, or 8 feet less than ours. Geol. Survey Texas, Third Annual Report, 1891, Austin, 1892, pp. 298-299.



BULL CREEK, TRAVIS COUNTY, TEXAS.

The bluffs are of Glen Rose beds. The water in the creek is derived from gravity springs.

| | Feet. |
|---|---|
| Glen Rose formation—Continued. | |
| 20. Soft argillaceous limestone, marly; forms a slope | 10 |
| 19. Shelf above, ledge below, rises | 10 |
| 18. Soft chalky (argillaceous) limestone with *Exogyra texana* at base, with harder layers that form shelves—eleven hard ledges. Twenty feet from the top of these beds the hard ledge is honeycombed by solution, and is arenaceous. In the lower 20 feet numerous fossils occur. *Tylostoma pedernalis, Cardium mediale,* "Goniolina," etc.; also horizon of *E. texana*. Thickness of series | 60 |
| 17. Hard ledges of honeycombed (perforated) limestone. The limestone, hard, yellowish, contains many poorly preserved calcitized fossil shells, largely the remains of Nerinea | 30 |
| 16. Hard ledge of limestone; many *Cardium mediale* | 5 |
| 15. Soft, argillaceous, chalky limestone | 5 |
| 14. Ledges, 6 inches to 1 foot thick, with soft, shaly layers between | 20 |
| 13. Soft limestone | 20 |
| 12. Hard ledge | 2 |
| 11. Soft, chalky, argillaceous layer | 10 |
| 10. Ledge of hard brownish or yellowish limestone, containing embedded sand grains | 5 |
| 9. Soft, chalky, argillaceous limestone, with an occasional hard ledge. Hard ledge 2 feet thick 15 feet above base. In the upper part of this marly bed fossils are very abundant. *Cardium mediale, Tylostoma pedernalis,* many echinoderms, *Pseudodiadema texana,* Nerinea, Ostrea, etc | 35 |
| 8. Ledge of hard yellowish limestone | 5 |
| 7. Slope, underlain by soft chalky limestone | 25 |
| 6. Arenaceous ledge, a few feet. | |
| 5. Soft ledge with many Monopleura, a few feet. | |
| Travis Peak formation: | |
| 4. Rather hard ledge, with poorly preserved fossils; appear to be oysters. | 2 |
| 3. Soft chalky limestone | 20 |
| 2. Ledge of yellowish limestone, 2 feet, and 40 feet of the section covered by river alluvium | 42 |
| 1. Yellowish calcareous sandstone at river level, thickness not obtainable. | |

*Totals of above section.*

| | |
|---|---|
| Comanche Peak and Walnut formations (in part) | 15 |
| Glen Rose formation (entire) about | 455 |
| Travis Peak formation (in part) | 64 |
| Total of section, about | 534 |

The rocks of the middle of the Glen Rose formation are the oldest exposed in the Nueces Valley. The thickness of the formation in this region, estimated from studies in the vicinity of Kerrville, is approximately 500 feet, which if true would indicate a uniform thickness along the entire line of strike across the region treated in this paper. This uniformity of thickness is not maintained along the line of dip, however, as will presently be shown. Taff's measurement of the Travis Peak formation makes it about 213 feet in thickness at its outcrop in the Colorado Valley.

It is difficult to determine the thickness of the embedded portions of these strata, especially the Travis Peak. The careful measurements of the surface outcrop do not coincide with the artesian-well borings made to the east, the latter showing a much greater thickness. For instance, the borings of the San Marcos artesian well (see p. 287) show a thickness of 904 feet of the Glen Rose and Travis Peak penetrated by the drill at that place. At Austin also the well records indicate that at least 1,215 feet of the two formations are penetrated. Neither of these wells has as yet reached the bottom of the Travis Peak. These data indicate that the aggregate thickness of the formations increases rapidly coastward away from the line of outcrop. The beds were deposited on the eastern slope of an old, subsiding, preexisting Paleozoic upland, against which the formations of the Cretaceous in general and of the lower beds in particular were cumulatively deposited until it was buried, and toward which the calcareous beds changed into an arenaceous character.

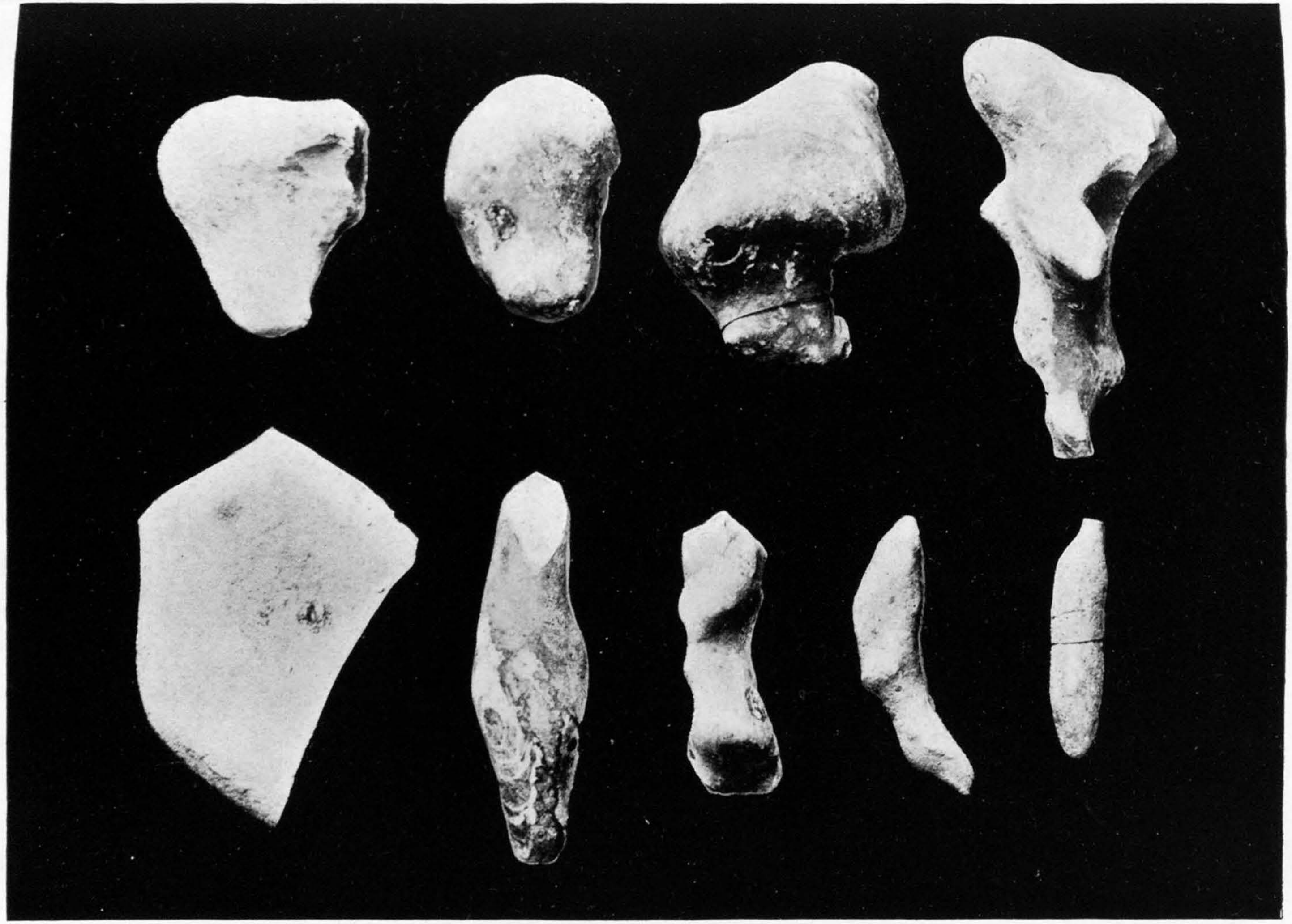
### WALNUT FORMATION.

If the classification of the Cretaceous formations had originally been made in the region here discussed, the Walnut formation would probably not have been given separate status, for, though well developed farther north, it is here thin and unimportant. In its typical development along the Brazos River it consists of laminated clays alternating with limestone flags, and both clays and flags are accompanied by great quantities of the two peculiar species of oyster, *Exogyra texana* and *Gryphæa marcoui*[1] sp. nov. H. & V., Pl. LIII. Along the Colorado River at the northern limit of the Edwards Plateau the formation is only 10 or 12 feet thick, consisting of thin, friable, yellow, arenaceous marls, in which are great numbers of *Exogyra texana*. It determines a distinct bench near the summit of the high mesas west of Austin. Southwestward, toward the Nueces, it becomes less and less distinguishable, until at that stream there is only a foot or two of yellow clay, accompanied by the characteristic fossils.

### COMANCHE PEAK LIMESTONE.

This is a persistent bed of white chalky limestone, presenting a shattered reticulated appearance on weathering. It is partly characterized by an abundant fossil fauna containing a large number of *Exogyra texana*, which is especially abundant in its basal portion. It is from 40 to 50 feet thick, thinning toward the Rio Grande. Although it is insignificant as regards thickness, and lithologically might be considered the base of the Edwards limestone, it is one of the most persistent paleontologic horizons of the Texas Cretaceous section.

[1] *Gryphæa pitcheri* (in part) of previous writings; described in Bull. U. S. Geol. Survey No. 151.



FLINTS OF EDWARDS LIMESTONE.

One-half natural size.

EDWARDS LIMESTONE.[1]

This formation is the most conspicuous and extensive in the Texas-Mexican region. It is composed mostly of limestone, but there are some marly layers. It shows slight variation in color, composition, texture, and mode of weathering. In general the beds are whitish, although layers of buff, cream, yellow, or dull gray are frequent. These colors depend much upon weathering. In composition most of the beds are as nearly pure carbonate of lime as can be found in nature, but some have small admixtures of silica, epsomite, chloride of sodium, and perhaps other salts as yet undetermined. Clay is absent except as a minor constituent in the few marly layers. Iron is sparingly present as pyrites, and is revealed by the red color of the clay that weathers out of a few beds. Exceedingly fine siliceous particles occur in the so-called "magnesian beds"—light-brown porous beds which appear southward from Comanche County—but no pebble, bowlder, lignite, or other undoubted piece of land-derived débris has ever been found.

The limestones vary in degree of induration from hard, ringing, durable strata to soft pulverulent chalk that crumbles in the fingers and resembles very much the prepared article of commerce. Some of the beds are coarsely crystalline, with calcitized fossils, and are susceptible of high polish. The beds also vary in texture. Some of them are quite porous and pervious, while others are close grained and impervious. Some are homogeneous throughout; others have hard and soft spots, the latter dissolving by the percolation of underground water and constituting what is popularly termed "honeycombed" rocks. The harder spots in some cases seem to be in process of induration, suggesting a step in the formation of flints. The holes in the honeycombed layers often represent what were once spots containing soluble salts of iron and other accessory minerals.

South of the Paluxy River the formation can always be distinguished by the immense quantity of flint nodules which are embedded in and between the limestones and which lie scattered over the surface everywhere. These are of many shapes (Pl. XXX); some are fusiform, like elongated roots; others are knotty, like warty potatoes; others are parts of extensive sheets or very flat lenses. They vary in size from that of a hen's egg to a foot or more in diameter. They also vary greatly in color. Upon fresh fracture some are almost jet black; others light blue, gray, or opalescent; still others are delicate pink in color. There is some evidence that each particular kind occupies a definite horizon, but we are not prepared to state this as a positive fact.

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[1] The geographic name Edwards is here substituted for the "Caprina limestone" of Shumard, the latter being abandoned because it is a paleontologic term; also for the term "Barton Creek limestone" of Hill, abandoned by him and revived by Cragin. The term "Barton Creek" is objectionable, and was abandoned by its author because it was not a good locality name and because it is a two-word name. The name Barton has been applied to a division of the English Eocene for many years, but we doubt if, in the absence of any defined code of geologic nomenclature, such an objection may be considered valid.

In most cases the Edwards limestone may also readily be distinguished by the peculiar aberrant mollusks of the genera Monopleura, Requienia, and Radiolites[1]—bivalve fossils which have cornucopiate form, suggesting a resemblance in shape to the horns of cows, goats, and sheep.

The formation is stratified into a succession of massive beds accompanied by very few flaggy and marly layers. Some of the strata are harder than others and project beyond the softer layers in the profile of the hills as overhanging shelves; others are soft and erode very rapidly. South of the Colorado, where the Walnut formation becomes insignificant, the Edwards limestone is almost inseparable from the underlying Comanche Peak, since both are composed chiefly of carbonate of lime. The Comanche Peak strata are less consolidated, and, as they are somewhat argillaceous, possess a more marly texture than the Edwards limestone, which is usually a firm, white, ringing limestone of great hardness and durability; so that the Edwards weathers into cliffs, while the Comanche Peak is wrought into lower-lying slopes; but in most cases reliance must be placed upon paleontologic determinations to distinguish the two formations.[2]

Neither is the Edwards limestone always sharply defined from the overlying Fort Worth, except by paleontologic criteria. It is true that the Fort Worth limestone is slightly more arenaceous, but the differences are so slight that their detection requires the trained eye of the geologist. As the upper limestone is less than 75 feet in thickness, the layman or well-driller unversed in paleontology can nearly always be sure that any rock occurring 75 feet below the Del Rio clays belongs to the Edwards formation.

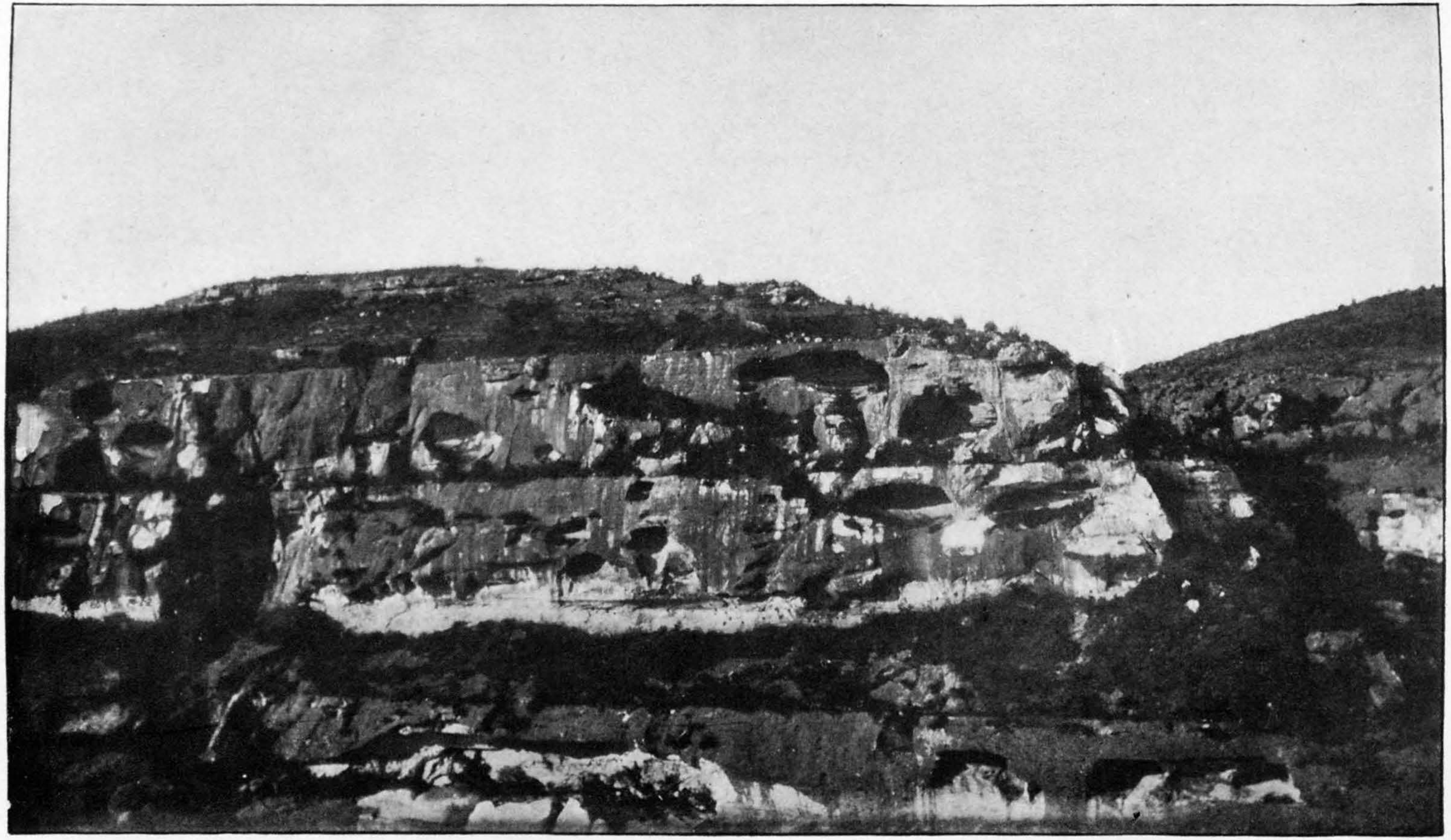
The Edwards limestone, being more purely calcareous than any other of the Comanche series, probably corresponds to the deepest and most extensive submergence of the Comanche epoch. It is true that in the Glen Rose formation occasional thin beds of chalk are met with, and that some of these are composed almost entirely of foraminifera, but such chalks usually contain a considerable percentage of clay, recognized as an offshore deposit.

Occasional bands of soft brownish-yellow stone are intercalated with the limestone. These bands are popularly called "magnesian," and are composed largely of an exceedingly fine-grained siliceous element like tripoli. As these beds often contain flints, the silex may be of organic origin.

Topographically, the Edwards limestone is one of the most important formations in Texas. In fact, it is the determining factor in the topography of the whole of the Edwards Plateau and Grand Prairie regions.

[1] See Plates LIV and LV, of typical fossils.

[2] It is doubtful if the interior of the Edwards limestone is always as hard as the surface outcrop, for many of the chalky rocks of Texas harden or set on surface exposure.



BLUFF CAVERNS IN EDWARDS LIMESTONE, RIO GRANDE.

Its hardness being superior to that of the overlying and underlying beds, its consequent resistance to erosion has preserved it as the capstone of the innumerable round "mountains" (buttes) and mesas of the State and of the extensive Edwards Plateau and Grand Prairie regions. Not only are most of the buttes and mesas capped by it, but these are accompanied by scarps overlooking the lower-lying valley prairies which follow the stream. The walls of the canyons which many of the streams have cut are also composed largely of the Edwards limestone, especially the higher and headwater portions of those rising in the Edwards Plateau. To its hardness is also largely due the topography of the limestone mountains of Mexico.

It shows many types of weathering. Some of the strata make bold cliffs nearly 50 feet in height, the faces of which, although apparently of homogeneous texture, weather into small open caverns (Pls. XXXI and XXXII). This weathering sometimes brings out a thinly laminated structure associated with white efflorescence. The bottoms of caverns of this character are filled with a layer of white, pulverulent earth. The residual products of other massive ledges weathering into caverns are vermilion-colored clays, in which are beautiful fossils composed entirely of crystallized calcite.

The hard limestones weather into vertical, square-cut bluffs, while the soft and more homogeneous beds of marly or chalky texture form slopes. Where these hard and soft beds occur in alternation there is a corresponding alternation of scarps and slopes in the topographic profile. Some of the beds of homogeneous texture having great thickness weather into pyramidal hills, as seen along the monoclinal fold at the south edge in the northern part of Kinney County (Pl. XXXIII).

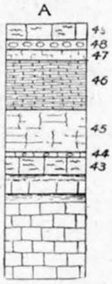
Near the summit of the Edwards Plateau, where the flaggy layers prevail, the slopes of the stream ways are gentle and are characterized by low, vertical steps, from 2 inches to 2 feet high, over which the stream descends from one rock layer to another. This slope is usually interrupted by vertical bluffs, composed of thick strata, which constitute a cornice in the profile of the canyon of the plateau. These slopes and scarps alternate until the base of the Edwards limestone is reached, beneath which the Comanche Peak bed weathers out in concave profile.

Where the Edwards formation forms extensive stretches of level country, such as that between Manchaca and Oak Hill in Travis County, in the western part of Williamson County, and the summit of the plateau, and the surface stratum is of homogeneous texture, it weathers into millions of miniature ridges, crests, and drainage lines, illustrating the whole process of erosion and mountain carving. These minutely eroded limestone surfaces are technically known as "karrenfelder" (see Pl. L, p. 318), and they are formed by the solvent effect of the rainfall upon the sun-heated limestone surfaces. The crevices in these level areas of Edwards limestone country are usually grass-covered, with occasional patches of scrub oak. The surface is very rocky, the karrenfelder protruding in jagged points through the rich but scanty soils. Sometimes residual flints occur in such immense quantities over these surfaces that one is apt to mistake them for a water-rolled gravel formation.

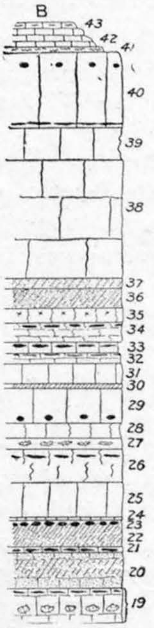
In Kinney County the high hills north of Fort Clark, with the exception of Las Moras Mountain, are made up largely of the Edwards limestone. The beautiful canyon of the Pecos, near its mouth, as shown in Pl. XXVII, is one of the finest examples of its outcrop.

A small, narrow belt also outcrops on the downthrown side of the Balcones fault. In Travis County an area of it occurs, as shown on the map (see Pl. XLVI, p. 282), along a belt lying between the International and Great Northern Railway and a line drawn through Oatmanville (Oak Hill) and Mount Bonnel, and the belt may be seen at many other places as far west as the Frio. The Aransas Pass Railway crosses it just south of Leon Springs.

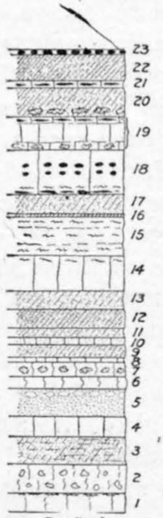
It is well displayed in the banks on the south side of the Colorado in the western part of the city of Austin, between McDonald's brick yard and the city dam. Owing to faulting, this section is somewhat complicated and not continuously exposed at any single locality. The details as made out at three localities at Austin are shown in the accompanying sections and figures (fig. 57). It will be noticed that in the lower portion of the Bee Creek section, which is still above the base of the whole of the formation, arenaceous marls and limestones are quite numerous. These play an important part in the artesian conditions from the Colorado southwestward.



Barton Creek.



DeepEddy Bluff.

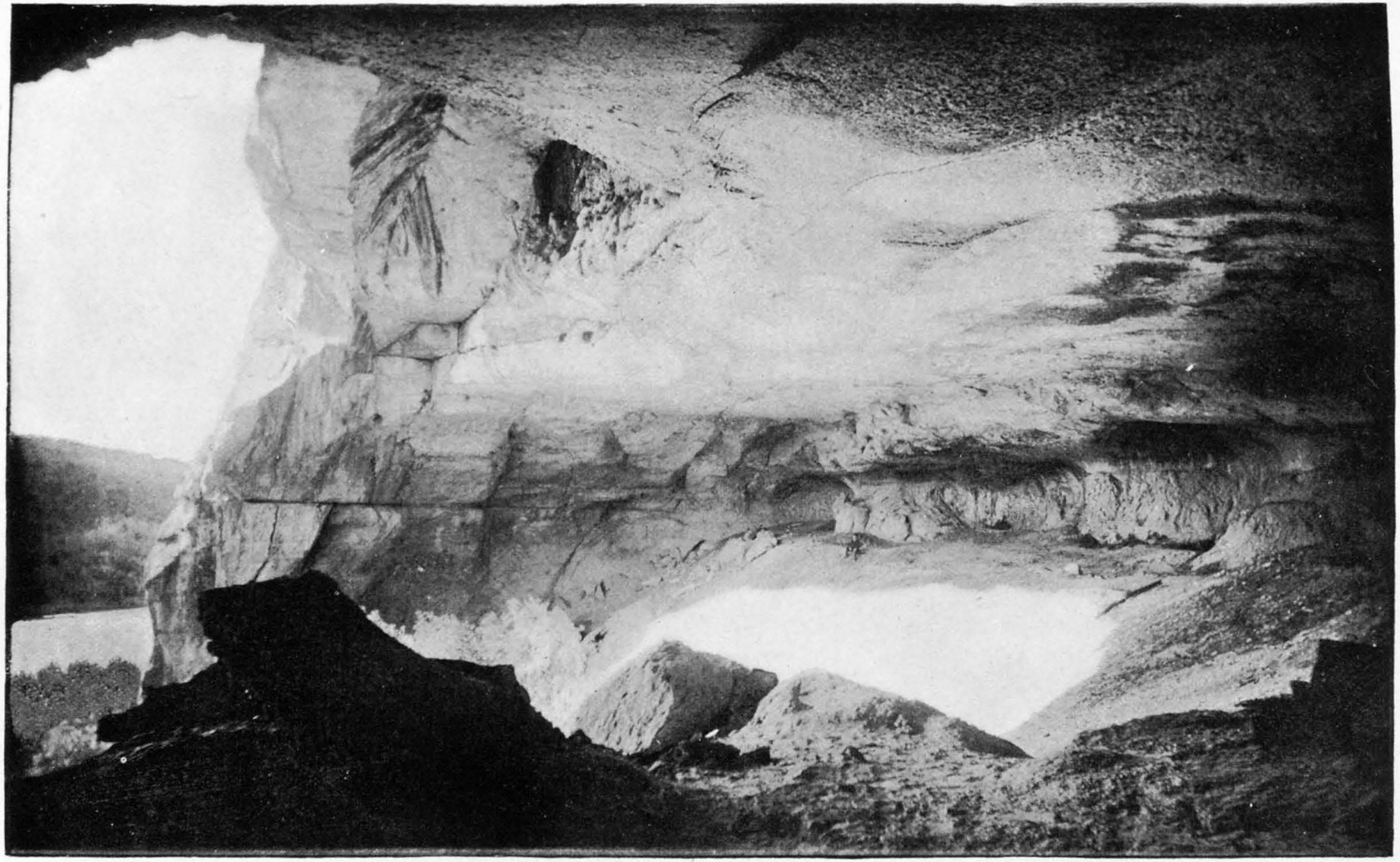


Bee Creek.

FIG. 57.—Sections of Edwards limestone, near Austin. For explanation of numbers, see Sections Nos. 4–6, pp. 231–233.

The accompanying sections, Nos. 4, 5, and 6 (see fig. 57, A, B, C), represent the entire thickness of the Edwards limestone exposed on the downthrown side of the fault in the bluffs of the Colorado between Austin and the river level at the mouth of Bee Creek.

The base of the beds is concealed, lying probably less than 100 feet below No. 1 of section C, but can be seen on the upthrown side of the fault, capping the remnants of the plateau.



BLUFF CAVERN IN EDWARDS LIMESTONE, RIO GRANDE; INTERIOR VIEW.

SECTION NO. 4.—*Bluff on Barton Creek, about 1 mile above Barton Spring, Travis County, Texas.* (Fig. 57, A.)

| | Ft. | In. |
|---|---|---|
| Fort Worth limestone:[1] | | |
| 5. Grayish limestone, irregular fracture, with *Alectryonia carinata* and *Gryphæa washitaensis* | 1 | 0 |
| 4. Yellow or reddish calcareous shale | 4 | 4 |
| 3. Alternating layers of hard and soft limestone with *Alectryonia carinata, Gryphæa washitaensis, Exogyra americana,* etc | 18 | 0 |
| 2. Hard grayish limestone | 33 | 0 |
| 1. Soft chalky limestone, with a saline taste | 13 | 0 |
| Total thickness of Fort Worth limestone | 69 | 4 |
| Edwards limestone: | | |
| 49. Nodular limestone full of Requienia | 3 | 0 |
| 48. Nodular limestone, nodules as large as one's head | 2 | 0 |
| 47. Hard chalky limestone | 3 | 0 |
| 46. Thinly laminated limestone (the so-called "lithographic flags") | 8 | 9 |
| 45. White, sublaminated, chalky limestone. The lower part of Nos. 45 and 46 contain many fossils, *Exogyra texana, Pholadomya knowltoni,* etc. | 8 | 5 |
| 44. Nodular limestone, no Requienia | 1 | 0 |
| 43.[2] Nodular limestone with many Requienia (second Requienia bed) | 3 | 9 |
| *e.* Laminated limestone | 1 | 0 |
| *d.* A series of hard limestone ledges (eight in number) separated by the thinly laminated layers. There are some flints, about as large as a man's fist—Radiolites and *Ostrea? munsoni* | 45 | 8 |
| *c.* Flaggy layer with discoidal flints | 2 | 4 |
| 29 *b.* Hard limestone, forming a shelf along this portion of Barton Creek and its bottom at the bridge below, eroded into deep pot holes. The lower 2 feet of this layer contains very large blue flints, often 1 foot across. Some of them are oval, others flattened out and very irregular in outline. The upper part of bed contains small flints | 12 | 8 |
| *a.* Limestone ledges with some flattened flints. All of the flints in this section belong to the blue variety | 11 | 0 |
| Base of *a* is Barton Creek bed. | | |
| Total thickness of strata in bluff | 171 | 11 |

SECTION NO. 5.—*Deep Eddy Bluff, south of the Colorado River, west of Austin.* (Fig. 57, B.)

| | Ft. | In. |
|---|---|---|
| 43. Nodular limestone with Requienia at top (the second Requienia bed of the Barton Creek section). | | |
| 42. Limestone ledges | 5 | 0 |
| 41. Limestone ledges containing Requienia. The three layers above described form a slope to the top of the hill (or bluff) above the face proper of the bluff | 1 | 0 |
| 40. Ledge of hard limestone, 10 inches above basal sheet flint. The upper part of the ledge contains rather small nodular flints | 15 | 0 |
| 39. Limestone weathering out and giving rise to a good deal of red clay, apparently representing the zone of calcitized fossils found in the high bluff above McGill's Ford | 6 | 6 |

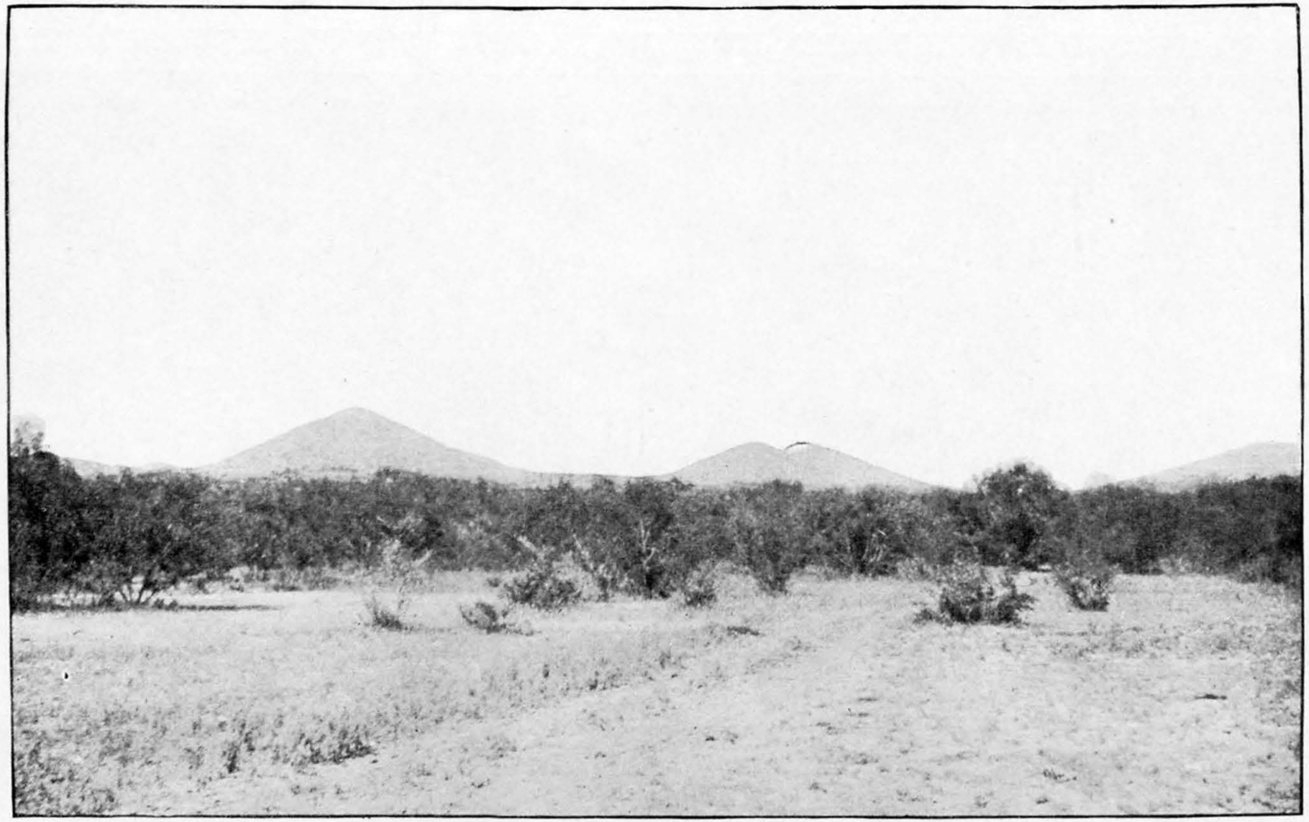
[1] The Fort Worth limestone is assumed to be 70 feet thick at Austin. Future study may modify this estimate. The uppermost layers, characterized by *Kingena wacoensis*, are missing in the section.

[2] The beds in the Barton Creek section below 43 can not be correlated layer for layer with the Deep Eddy Bluff section; therefore numbers are not used in the description of the former section for beds below the one numbered 43 (except in one case, 29). The numbers in the different sections indicate equivalence.

| | Ft. | In. |
|---|---|---|
| 38. Massive thick ledges of limestone, detail not exposed | 23 | 8 |
| 37. Soft, white, arenaceous limestone | 2 | 2 |
| 36. Soft arenaceous limestone | 3 | 10 |
| 35. Ledge of limestone, rather soft, emitting odor of petroleum | 2 | 10 |
| 34. Chalky limestone forming little caves, composed of a good many small ledges; discoidal flints at top | 4 | 0 |
| 33. Hard limestone, emitting odor of petreloum under blows of hammer. Texture of limestone rather mealy. Nodular flints, occasional discoidal flints in top | 2 | 1 |
| 32. Two thin ledges of limestone; layer of sheet flint in top | 1 | 2 |
| 31. Ledge of thick massive limestone | 5 | 0 |
| 30. Hard yellowish limestone | 2 | 0 |
| 29. Hard, thick, massive ledge of siliceous limestone, ringing under blows of hammer. At the base there is a layer about 9 inches thick of opalescent, pinkish or brownish flint. Apparently the limestone is being converted into flint by replacement, and the process has not yet been completed | 6 | 0 |
| 28. Soft chalky limestone, dissolving and forming small caves | 3 | 0 |
| 27. Soft chalky limestone with very large (may be 1 foot long), irregularly shaped blue flints at top | 2 | 3 |
| 26. White chalky limestone, apparently siliceous; zone of flint near top. The flints blue, discoidal, and tending to form sheet | 6 | 8 |
| 25. Massive ledge of hard bluish limestone | 7 | 0 |
| 24. Very hard limestone | 0 | 6 |
| 23. A layer of enormous blue flints, in some places over 1 foot thick | 1 | 0 |
| 22. Thick massive ledge of limestone, rather soft, yellow in color, and slightly arenaceous | 5 | 5 |
| 21. Ledge of hard yellowish limestone with a zone of flints tending to form a sheet at base | 1 | 4 |
| 20. Soft, white, slightly arenaceous limestone, composed of thin ledges; upper 2 feet, middle 4 feet, lower 1 foot | 7 | 0 |
| 19. Soft yellowish or whitish limestone with layer of flattish, bluish flints forming a sheet at top. This is really three ledges: upper ledge, with flints at top, 2 feet; middle, containing concretions of calcite in lower part, 4 feet; lower ledge, exposed at low water, 1 foot | 7 | 0 |
| Total, Deep Eddy section | 121 | 5 |

SECTION NO. 6.—*Bluff at the mouth of Bee Creek.* (Fig. 57, C.)

| | Ft. | In. |
|---|---|---|
| Limestone slope, detail not exposed | 11 | 0 |
| 23. Layer of enormous blue flints | 1 | 0 |
| 22. Arenaceous limestone | 5 | 5 |
| 21. Hard yellowish limestone with sheet flint at base | 1 | 4 |
| 20. Yellowish, rather hard limestone, somewhat siliceous; thin band of chalky limestone at top; calcite concretions near base | 6 | 0 |
| 19. Sheet flint at top (sheet flint at top of lowest ledge of Deep Eddy Bluff); three ledges of limestone: upper, 1 foot; middle, 2 feet 6 inches; lower (containing calcite concretions), 3 feet | 6 | 6 |
| 18. Sandy limestone, with two zones of nodular flint near middle; sheet flint at base; mass of Requienia just above the sheet flint | 10 | 0 |
| 17. Soft, yellow, calcareous sandstone, a part of the preceding ledge, about | 3 | 0 |
| 16. Yellow cherty limestone, about | 0 | 6 |
| 15. Three or four ledges of rather soft, whitish or yellowish limestone; the upper ledge containing a great mass of Requienia, the others fewer | 8 | 1 |



PYRAMIDAL BUTTES OF EDWARDS LIMESTONE, KINNEY COUNTY, TEXAS.

| | Ft. | In. |
|---|---|---|
| 14. Solid white limestone, granular, not very hard; contains a great many Requienia near top | 6 | 11 |
| 13. Yellow arenaceous limestone | 4 | 0 |
| 12. Blotched arenaceous limestone | 3 | 8 |
| 11. Soft, yellow, arenaceous limestone | 2 | 0 |
| 10. Hard, yellowish, granular limestone, with shell fragments, gray on fresh exposure | 1 | 6 |
| 9. Soft, yellow, arenaceous limestone or calcareous sandstone | 2 | 4 |
| 8. Ledge of nonindurated granular limestone, with indurated blotches, which are structureless and flinty looking | 1 | 0 |
| 7. Ledge of white, rather soft limestone, with many very irregularly shaped flints in a zone about the middle of the ledge. The flints are mostly small, bluish in color, and do not show concentric banding; about | 3 | 0 |
| 6. Ledge of white, rather soft limestone; no flints; a few fragmentary fossils | 2 | 6 |
| 5. A soft arenaceous ledge. The lower 1 foot 10 inches is a subledge. In the upper part (near top) are concretionary bodies that in their form resemble flints, but are not flints in texture. These bodies are hard, apparently siliceous, and contain white blotches, some of which appear to be of foraminiferal origin | 5 | 5 |
| 4. Hard limestone, whitish or bluish, without flint; not fossiliferous | 4 | 0 |
| 3. Arenaceous limestone, has a tendency to lamination, but in the ledge the laminated character is not always evident. The upper part of the ledge by solution becomes porous. The rock has a considerable absorbent power for water, and has a dark (wet) appearance, due to water contained | 5 | 9 |
| 2. Thick ledge of white limestone, not very hard, oxidizing yellow from contained iron. Contains a large number of irregularly shaped flint nodules. These may be as much as 1 foot long, but usually are rather small—3 or 4 inches in length. They are bluish in color and have a concentrically grained structure, resembling the graining of pine wood. Their long axes are not always parallel to the bedding planes of the limestone, an important exception to the usual position of the flints relative to the stratification of the limestone | 5 | 9 |
| 1. Ledge of yellowish or whitish limestone, without flints; in a thin layer about 6 inches thick at the top of this ledge there is an enormous number of *Requienia texana* | 4 | 0 |
| Total, Bee Creek section | 104 | 8 |

The total thickness of the Edwards limestone exposed in the vicinity of Austin, as determined from the foregoing sections, is as follows:

| | Ft. | In. |
|---|---|---|
| Bluff on Barton Creek, beds 49 to 43 | 29 | 11 |
| Deep Eddy Bluff, beds 42 to 24 | 99 | 8 |
| Bluff at mouth of Bee Creek, beds 23 to 1 | 104 | 8 |
| Grand total | 234 | 3 |

In the Nueces section the upper 100 feet consist of flaggy layers of hard white limestone devoid of flints. Below this are ledges of yellowish limestone of considerable thickness, marked by numerous black flints. The central portion is of white limestone of homogeneous texture, in which large caverns occur. The lower portions consist of thick and thin ledges and flags containing considerable numbers of flint nodules or strata of flint and many honeycombed layers.

*Overlay.*—On the summit of the plateau in Edwards County (altitude 2,375 feet) are occasional low, mound-shaped hills of clay marl containing thin bands of limestones with *Terebratula* (*Kingena*) *wacoensis* and *Exogyra arietina*, estimated not to exceed 50 feet in thickness. These beds belong to the lower formations of the Washita division.

SECTION NO. 7.—*Canyons of the Nueces, Edwards County, Texas.*

| | Feet. |
|---|---|
| Edwards limestone: | |
| 9. Thin flags of ringing white limestone, forming great slabs upon the summits of the plateau; very fossiliferous and containing in the Hackberry Creek section large Nerinæa near their base. These limestones weather into gentle caletas | 78 |
| 8. Limestone characterized by an abundance of black flints, often of great size and occurring in nearly all the beds. The limestone beds are thicker and more massive than those of the above division, and usually constitute the first escarpment met in descending from the plateau or the highest escarpment in rising from the canyons. This division has many fossil Requienia in it and also a large and undescribed pecten, which is often preserved in the flints. About 50 feet from its base in both the Hackberry Creek and Frio sections there is a conspicuous band containing numerous radiolites. The thickness of these beds in the Frio and Hackberry Creek sections measures about. | 170 |
| 7. Thin-bedded layers of yellow, arenaceous, marly material alternating with thin limestone flags. Some of these layers are honeycombed and present numerous fucoidal masses. The intervening layers of hard limestone are very persistent. From the softer beds of this division great volumes of water break out at the headwater hole of the west branch of the Frio in Edwards County, and create a beautiful stream which is described elsewhere. The headwater springs of the East Fork of the Nueces also break out at exactly the same geologic horizon. Some of the deep wells on the summit of the plateau at Rock Springs and Anderson's ranch also may derive their water from these beds, which are the highest water-bearing strata of the Edwards beds. About | 20 |
| 6. White limestone in massive, thick strata, some of which attain 10 to 15 feet in thickness, without laminated structure; contains a few small flints of cylindrical outline and pinkish color. This limestone weathers into underground caverns of the type of Hillcoat Cave on the Nueces quadrangle, elsewhere described. About | 105 |
| 5. Hard, rugged limestone in thick layers, with opalescent flints. Limited at its base by certain fossiliferous zones containing many Requienia. Forms the second scarp above the Nueces River bottom | 20 |
| 4. Thin, flaggy layers of limestone alternating with calcareous layers of shale which are very bituminous in places, and contain fossil Gryphæa. Approximately | 60 |
| 3. A thick band or collection of bands of rough, cavernous limestone weathering into what is known as the honeycomb structure; the cavities large and lined with red iron colors. The zone is made up largely of fossil Requienia. This band forms the conspicuous scarp rock of the Nueces Valley in the vicinity of Black water hole. Where the streams have cut into the lower Glen Rose beds it makes the summit of the hills, such as the first scarp of Round Mountain. It is undoubtedly water-bearing, and many springs break out from its base along the rivers | 35 |

| Edwards limestone—Continued. | Feet. |
|---|---|
| 2. Coarsely bedded limestones, with a few flints and fossil Requienia..... | 100 |
| 1. Cavernous honeycombed limestone with many black flints and Requienia. Forms lowest scarp rock of the West Fork section (36 feet), with thinner limestone beds at base (16 feet)........................ | 50 |
| Total ........................................................ | 638 |

The foregoing section is a composite of several small sections and is only approximate, having been measured by aneroid barometer, but it will suffice to give a general idea of the Edwards formation in the Nueces canyons. It will be noticed that the total thickness is more than double that of the Colorado section.

## FORT WORTH LIMESTONE.

This formation consists of a group of impure white limestones, regularly banded, and alternating with layers of marly clay. (See upper part of fig. 57, p. 230, and lower part of fig. 58, p. 236.) Before exposure they are dull blue in color, but when weathered they are white or yellowish. The lower portion of the section as exposed at Austin contains thicker and more massive beds than the upper. They are paleontologically characterized by *Epiaster elegans*, *Ammonites* (*Schlœnbachia*) *leonensis*, *Gryphæa washitaensis*, *Exogyra americana*, *Kingena wacoensis*, etc. These fossils occur throughout in definite zones and associations, and some of the strata are composed almost entirely of them.

Above the more massive lower layers is an agglomerate of *Gryphæa washitaensis*. Associated with this is found an oyster, *Alectryonia carinata*, a familiar European form, occurring only at this horizon in the Austin section.

At the top is a stratum of massive limestone less than 3 feet thick, consisting of a homogeneous calcareous matrix thickly studded with *Kingena wacoensis*.[1] (See Pls. LVI and LVII for figures of most of these fossils.)

The formation does not exceed 75 feet in thickness in the Colorado River section. Southwestward, toward Brackett, it becomes less and less distinguishable from the underlying Edwards limestone and is recognizable at only a few places. So far as the question of artesian waters is concerned, it would be better to consider it as the upper 70 feet of the Edwards limestone rather than as an independent formation.

As a rule these beds outcrop only immediately along the western margin of the Rio Grande Plain, at the foot of the Balcones escarpment, and occupy an exceedingly narrow belt from Austin to Del Rio.

[1]Paleontologically, the term Fort Worth limestone should be limited to the lowest member of this Austin section or to the thicker beds carrying *Epiaster elegans* and *Schlœnbachia leonensis*, for the few feet of upper marls are probably the southern attenuation of the Denison formation. The lowest formation of the Washita division (Preston) has not been found at Austin, or to the southward, although its equivalent has been reported by Taff at Georgetown, Williamson County, and is probably represented by a few feet of limestone below the Fort Worth just south of Round Rock in the same county. For definitions of Denison and Preston formations consult Bull. Geol. Soc. America, March 1894, Vol. V, pp. 303, 324-332.

Hitherto it has been supposed that these beds did not occur on the summit of the Edwards Plateau, but our observations in 1895 showed that on the highest summit of the plateau in Edwards County small areas are preserved, as can be seen at Anderson's ranch, between the headwaters of the Guadalupe and the Frio rivers, on the Kerrville and Rock Springs road.

From the paleontologic notes of Dr. G. G. Shumard,[1] who made an expedition across the region before the present geologic classification was made, we are led to believe that fragments of the beds are quite extensively preserved on the western side of the plateau, between Fort Clark and the Pecos. There can be little doubt that the whole plateau was once capped by these beds, and probably higher beds that have since been almost entirely removed by erosion.

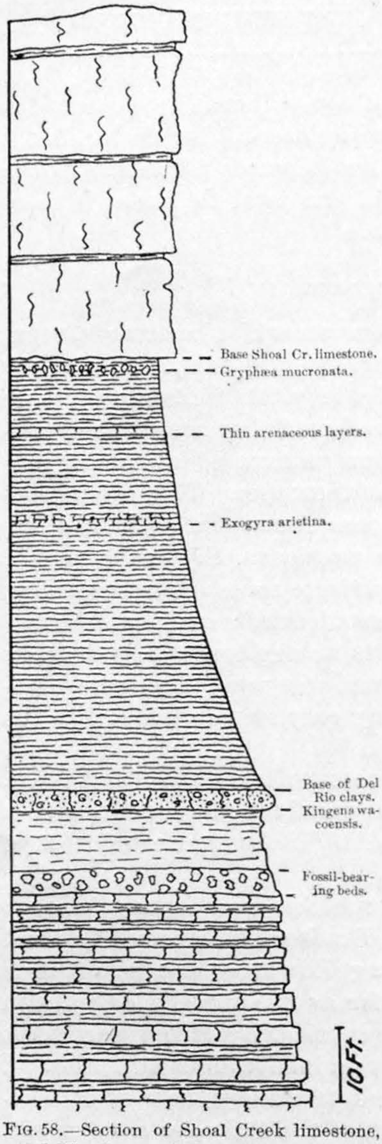


FIG. 58.—Section of Shoal Creek limestone, Del Rio clays, and Fort Worth limestone, at Austin.

#### DEL RIO CLAYS.[2]

These are peculiar greenish-blue laminated clays which weather dull brown or yellow and form a very black soil. They are some 80 feet thick at Austin, where they have their typical occurrence in Shoal Creek and at Fish Pond Bluff, at the mouth of Barton Creek. They outcrop immediately beneath the Shoal Creek limestone, and rest upon the Fort Worth limestone, the uppermost band of which is characterized by the occurrence of *Kingena wacoensis*. They are an especially important landmark in the geologic column, marking a break in a monotonous sequence of limestone beds, and possessing lithologic and paleontologic characters which render them easily recognizable. They can always be identified by means of a peculiar fossil, *Exogyra arietina*, a little oyster shown on

[1] A Partial Report on the Geology of Western Texas, etc., during the Years 1855–1856, by Prof. G. G. Shumard, Austin, 1886, pp. 69–77.

[2] The geographic name Del Rio clays is here substituted for the paleontologic designation *Exogyra arietina* clays of previous writings.

Pl. LVIII, figs. 2a–2e. This occurs in the greatest abundance, weathering out by the thousand in a state of perfect preservation. Attached to these shells, especially the umbonal region, are small cubes of iron pyrites. Upon decomposition this coats the shells with thin layers of brown hematite, and converts the lime into numerous crystals of fibrous selenite, which are intercalated in the seams adjacent to the shell horizons. In places the shells are cemented into thin layers of indurated argillaceous limestone, making persistent bands in the middle of the clay bed.

Above the zone of *Exogyra arietina* the clays are somewhat barren of fossils until near their summit, where they become slightly arenaceous and contain impure limestone slabs bearing other fossils, some of which also occur in the upper layers of the Fort Worth limestone.[1] Among these fossils is a gryphæate oyster, *G. mucronata*[2] of Gabb.

The Del Rio clays occur as occasional patches on the Edwards Plateau, in central Edwards County, between the headwaters of the Frio, Nueces, and Llano rivers. They were noted on the road from Dieter's ranch to Rock Springs. According to the paleontologic notes of Dr. G. G. Shumard previously quoted, there are also areas on the plateau between Fort Clark and the Pecos River.

South of the Colorado River these clays appear at various places along the interior margin of the plain adjacent to the Balcones fault, in Hays, Comal, Bexar, Uvalde, Kinney, and Valverde counties, at least as far west as Del Rio, 200 miles southwest of Austin, near which place they cross the Rio Grande in a bed of slightly increased thickness.

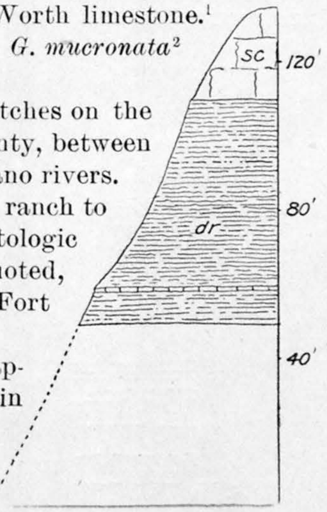



Fig. 59.—Section of Del Rio clays near Weymüller's ranch, Kinney County. *sc*, Shoal Creek limestone; *dr*, Del Rio clays. Base concealed.

### SHOAL CREEK LIMESTONE.

This has its characteristic exposure along the Colorado, in the steep scarps of Shoal Creek in the city of Austin, and in the bluffs on the south side of the river, where Bouldin Creek enters the valley, at the crossing of the International Railroad and the Oatmanville road. It forms precipitous cliffs, with toppling projections, owing to its jointed structure.

Its outcrop oxidizes to a slightly darker color than the limestones

[1] The senior author in a previous paper has said that "this fauna is the upward limit of the grand fauna of the Washita Division, the subfaunas of which show connection by a few common binding species," and that, on the other hand, not a single species passes upward from the *Exogyra arietina* beds into the Shoal Creek (Vola) limestone, thus showing between these beds a life break as marked as is the lithologic change. Later studies have shown a few connecting species between the Shoal Creek and Del Rio formations.

[2] This has been called "*Gryphæa pitcheri*" by Roemer, and is the "*G. navia*," or "*G. pitcheri* var. *navia*," in part, of the writings of Shumard, White, and of all the reports of the Texas survey.

already described. On fracture it is light yellow, with blotches or spots of pale pink, as if it had been subjected to fire. In places it is very hard, but in general is of varying texture, usually lumpy; in some spots it is efflorescent and decays into a soft, pulverulent material with slightly saline taste.

The minute red and pink blotches are peculiar to this limestone and have given to it the local name of "Burnt limestone." Microscopic study has revealed the fact that the rock is made up largely of foraminifera, filled and coated with a mineral, which in all probability is glauconite. Exteriorly the limestone presents no appearance indicating that it contains foraminiferal remains, but, so far as examined, it is more largely composed of them than any rock of the whole series. In one thin section Rotalia, Textularia, Globigerina, and fragments of three or four other genera of Foraminifera have been recognized.

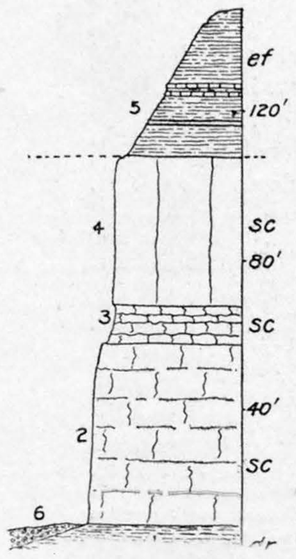



FIG. 60.—Section of Shoal Creek limestone, Kinney County. 1 (*dr*), Del Rio clays; 2, 3, 4 (*sc*), Shoal Creek limestone; 5 (*ef*), Eagle Ford shales; 6, river alluvium.

The outcrop of this formation is proportionately very limited, being better displayed at Austin than at any other locality. It can be found in the bluffs just below the Balcones scarp line near by many of the streams, such as Bear Creek at Manchaca; Onion Creek at Buda; the San Marcos at San Marcos Springs; the Frio, and along the West Fork of the Nueces, between Turkey Mountain and its mouth.

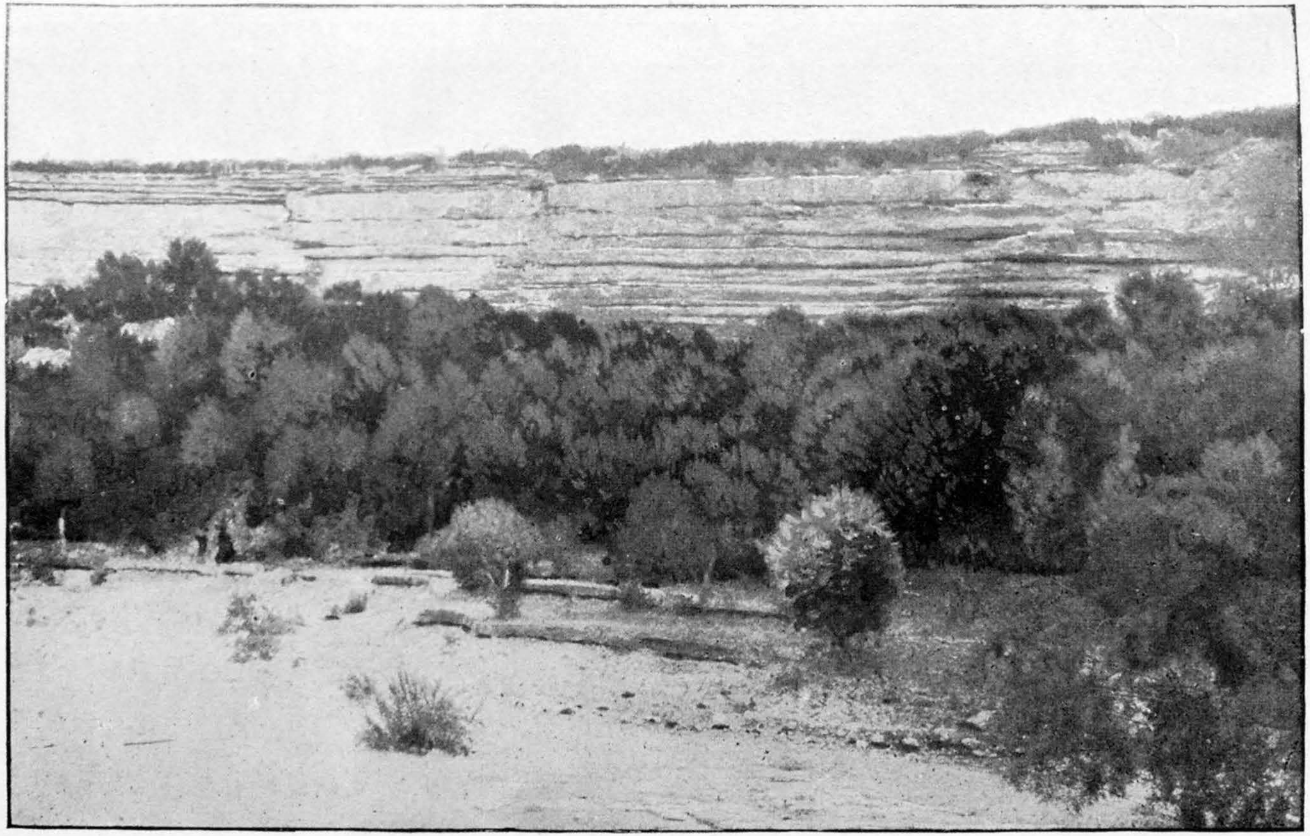
In the Austin section the Shoal Creek limestone rests without apparent gradation upon the Del Rio clays, indicating a rapid physical change in sedimentation, but in the Uvalde country there is intergradation.

## GULF SERIES (UPPER CRETACEOUS).

The Upper Cretaceous rocks in the portion of Texas under discussion in this paper are found only in the Rio Grande Plain. They are composed mostly of calcareous clays or other soft and unindurated strata. Owing to the generally soft and friable character of the material, the surface composed of it is naturally rather level or gently undulating.

The rocks of this series contain but little water, and their principal function in relation to the water question is that they serve as a cover for the water-bearing rocks proper. It is essential, however, to have some knowledge of them to be able to estimate the depth of the water-bearing beds beneath the surface.[1]

[1] The relations of the Upper Cretaceous formations of southwestern Texas to those of other regions are discussed in a paper, nearly ready for publication, by T. Wayland Vaughan, entitled "Reconnaissance in the Rio Grande Coal Fields of Texas," intended for the Bulletin series of the Geological Survey.



SHOAL CREEK LIMESTONE, WEST FORK OF THE NUECES RIVER, KINNEY COUNTY, TEXAS.

The beds at the top of the bluff are the Eagle Ford shales.

### EAGLE FORD SHALES.

This formation consists of laminated clays, shales, and impure limestones, usually blue or black when unweathered, but becoming light yellow and white on exposure. Both the shales and clays are distinguished from the other Cretaceous rocks, especially those immediately above and below them, by their laminated character. The beds usually contain remains of fishes, such as scales, teeth, and small bones, and a few mollusks, principally inocerami. (Pl. LX.)

At Austin the shales are exposed in the Sixth Ward and along the breaks of Shoal Creek, especially where Pecan street crosses it. They are finely displayed also in Bouldin Creek, on the south side of the river. At San Antonio they outcrop near the cement works and furnish the material from which Portland cement is made. They are also greatly developed around Fort Clark and at other points in Kinney and Valverde counties, the town of Brackett being situated upon them. In the last-mentioned area the formation consists of thick, flaggy, hard, or chalky limestones with interbedded or interlaminated marly layers.

The Eagle Ford shales are rarely over 50 feet in thickness anywhere in the vicinity of Austin. Near Brackett they are fully 250 feet thick.

### AUSTIN CHALK.

This formation consists of impure white chalky limestone with a conchoidal fracture, and is usually free from grit. It is so soft on fresh exposure that it is easily cut with edged tools. In places massive beds are interstratified with very chalky marls. Under the microscope the chalk exhibits a few calcite crystals, particles of amorphous calcite, and a great number of the shells of foraminifera and other minute organisms. The air-dried, indurated surfaces are white, but the saturated and unoxidized rock below the surface has a bluish color. The rock usually weathers in large conchoidal flakes.

In composition it varies from 85 to 94 per cent of calcium carbonate, the residue consisting of magnesia, silica, and a small percentage of ferric oxide.

It is easily distinguishable by its characteristic fossils, but superficially resembles closely some of the beds of the Comanche series. The Comanche limestones are usually harder and more crystalline, but this distinction can not always be made.

The Austin chalk is of great uniformity, presenting few local variations. Its thickness is difficult to determine, but averages about 500 feet in other parts of the State. The Manor well shows the thickness at Austin to be 410 feet, and from San Antonio westward to the Rio Grande it is probably 500 feet.[1] It is a most important bench mark or datum plane in the determination of the depth of underground waters, and will be frequently referred to in the economic discussions in this

[1] Fifteen hundred feet has been given as the thickness of this formation along the Rio Grande (Bull. Geol. Soc. America, Vol. III, 1892, p. 229), but this is probably excessive.

paper. The outcrop follows the northwestern margin of the Rio Grande Plain from Austin, via New Braunfels and San Antonio, to Fort Clark, crossing the Rio Grande between Del Rio and Eagle Pass. The greater portion of the city of Austin east of Shoal Creek is situated upon this rock, and the foundations of the State capitol are built upon it.

### TAYLOR FORMATION.[1]

The Austin chalk is overlain by a deposit of calcareous clays locally known as "joint clays,"[2] estimated to be about 540 feet in thickness in the Colorado River section. When fresh, these beds are fine grained, tough, unctuous, blue clays. They are apparently unlaminated until exposed to weathering, when their laminated character is developed. Their accessory constituent is lime in a chalky condition. Upon atmospheric exposure their color, owing to oxidation of the contained iron, changes to a dull yellow.

Because of their rapid surface disintegration the character of the unaltered beds is seldom seen, except when fresh material is brought up by the well-digger or exposed in freshly cut ravines or creeks. At the Blue Bluffs of the Colorado River, 6 miles east of Austin, there is a good fresh exposure.

At the top, as seen at various places in Travis County east of Austin, they grade into the marls of the Webberville formation. Their middle portion apparently contains no well-preserved fossils, but impressions are abundant in places. In the base of the beds *Exogyra ponderosa*, a large, heavy oyster, is abundant. (Pl. LXIII, fig. 3*a*, 3*b*.)

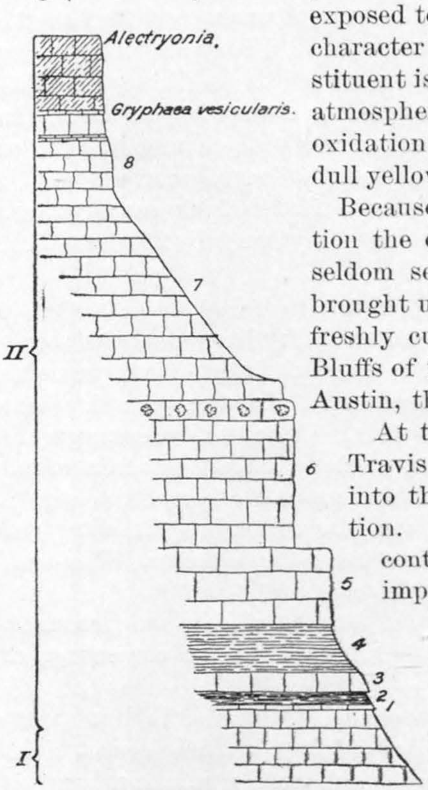


FIG. 61.—Section of Anacacho formation, Anacacho Mountains, Kinney County.

### ANACACHO FORMATION.

In Uvalde and Kinney counties, in the stratigraphic position occupied to the eastward by the Taylor marls, is a series of hard yellow and white limestones with interbedded marls and occasional sandstone ledges, for which the local name Anacacho formation is proposed, after the locality of their characteristic occurrence, the Anacacho Mountains of Kinney County, which are capped by this formation.

The following is a section at the east end of the Anacacho Mountains.

[1] *Exogyra ponderosa* marls of the earlier literature.

[2] This name is applied to massive clays, because of their tendency to break into blocks.

It gives a good idea of the constitution of the beds. Fig. 61 represents in a diagrammatic manner the general character of the beds, showing those that form slopes and those that because of their greater hardness produce escarpments.

SECTION NO. 8.—*Section of Anacacho Mountain, Kinney County.*

| | Feet. |
|---|---|
| Anacacho formation: | |
| 8. Scarp-making rock, forming the top of the hill. It is a hard, yellow subcrystalline limestone. In the top a species of Alectryonia was found. About 30 feet below the top great numbers of *Gryphæa vesicularis* occur, firmly embedded | 60 |
| 7. Softer limestone: *b*, Soft, yellow, marly limestone, containing a large species of Cardium, 50 feet; *a*, Soft, white, chalky limestone, containing a species of Turritella with three prominent revolving striæ on each whorl (*T. trilira* Con.?), 30 feet; total | 80 |
| 6. Hard limestone ledges. The upper 30 feet is brownish and contains great numbers of *Exogyra ponderosa* firmly embedded near the top. The next lower 20 feet is a yellowish granular limestone with glauconitics pecks.[1] This bed forms a platform on the east end of the Anacacho Mountains | 70 |
| 5. Ledges of yellow, ferruginous, not very hard, subcrystalline limestone, forming the lower scarp on the east end of the hill | 30 |
| 4. Slope composed of marly limestone in the upper part, the lower portion yellow marls with fragments of a very large and coarsely corrugated Inoceramus | 20 |
| 3. Yellowish limestone, weathering into nodular chunks, iron-stained along the weathering cracks. Contains some poorly preserved fossils—Trigonia (?), Mactra, and a finely ribbed Lima | 10 |
| 2. Soft material containing fragments of a very large Inoceramus | 5 |
| 1. Thin, very hard, brown, siliceous ledge | 3 or 4 |
| Total Anacacho beds | 279 |
| Austin chalk, at base of above: | |
| 3. Hard brownish limestone, containing many *Gryphæa aucella* | 5 |
| 2. Hard chalky limestone | 10 |
| 1. Unexposed—to bottom of arroyo | 20 |
| Total Austin chalk exposed | 35 |
| Total Anacacho beds | 279 |
| Total here exposed | 314 |

The measurements were made with an aneroid barometer and must be regarded as only approximately correct.

The deposit of asphalt at the Lithocarbon Rubber Company's mine is in the Anacacho limestone. The asphalt-bearing horizon corresponds with the upper part of No. 6, above.

### WEBBERVILLE AND EAGLE PASS FORMATIONS.

The highest beds of the Cretaceous outcrop along the eastern margin of the Black Prairie region and of the northern portion of the Rio Grande Plain. In the Texas region they present two distinct lithologic

[1]These beds contain a peculiar form of Rudistes not hitherto found in the Upper Cretaceous of this country. It will be studied by Mr. Stanton of this Survey.

facies, though the beds are probably synchronous. In northern Texas and as far south as the Colorado River the general aspect is that of the Upper Cretaceous formations of the Atlantic and Gulf States from New Jersey into Texas, which consist of sands, marls, dark clays, or other impure sediments, usually characterized by the appearance in some form or other of particles of the mineral glauconite (greensand) in the various beds.[1] From San Antonio westward to El Paso the beds of the uppermost Cretaceous consist of yellow clays, impure ferruginous limestones, and beds of lignite, and have a general increase in thickness, showing a transition to the facies of the synchronous beds of the Rocky Mountain region.

Along the Colorado River, below the mouth of Onion Creek, the Taylor formation grades upward into glauconitic marls with beds of impure limestone and black clays. The local term Webberville has been used[2] for these. They are exposed at only a few localities in eastern Travis County.

In the First Annual Report of the Texas Geological Survey, page 20, the characteristic outcrops of Webberville beds are well described, but the formation is erroneously referred by Penrose to the Wills Point beds of the Basal Tertiary, as follows:

> On the Colorado River it is seen outcropping at a point 16 miles by river below Austin, and 1 mile below the mouth of Onion Creek, in a bluff some 40 feet high and a mile long. Also at Webberville, on the line between Travis and Bastrop counties, where it is seen in a low bluff just above the water's edge. This is a much darker and more massive clay than that seen in most other outcrops. In the bluff 16 miles below Austin are found a few fragments of fossils, but they are all so broken as to make their determination very doubtful.

The investigations of this Survey in 1894 have shown them to contain fossils characteristic of the upper division of the Upper Cretaceous, identified by Mr. T. W. Stanton, as follows: *Anomia conradi*, *Leda protexta*, *Corbula crassiplica*, *Drillia ? distans*, *Sphenodiscus lenticularis*.

Although probably not the highest beds of the Cretaceous system, the Webberville beds are the highest Cretaceous exposures seen along the Colorado River, for below Webberville they are overlain by the basal division of the Eocene Tertiary.

[1] In order to distinguish the allied facies, the senior author has in previous papers spoken of the northernmost of these areas, presenting the Atlantic States facies as the glauconitic division and those of the Rio Grande and trans-Pecos region as the Montana division, the latter name having been previously used for the allied formations in the Rocky Mountain region. The term "Glauconitic" has always been considered unsatisfactory, inasmuch as it is not a geographic word. The word Montana can not be used with certitude for the beds of the northern area until they are proved to be identical on paleontologic grounds. The beds of the two regions are quite different in lithologic aspect, and they must be distinguished until they are proved to be the same. The term "Ripley" has been used as a generic one for beds which we know belong to this division in Texas, but without specific definition. Since the original use of the word Ripley in Mississippi was restricted to some of the beds of the many in that State composing the equivalent of this division as a whole, it is hardly appropriate to apply that name to the entire uppermost division in Texas.

[2] Preliminary check list of Cretaceous fossils of Texas, by Robert T. Hill: Bull. 4, Geol. Survey Texas, Austin, 1889, p. xxx. Also, First Ann. Rept. Geol. Survey Texas, p. 115.

From the Colorado River southwestward through San Antonio and onward to the Nueces the beds of this upper division appear only occasionally along the watercourses, where erosion has cut through the overlying Neocene and Pleistocene deposits.

Near San Antonio, as shown by the Terrell artesian-well borings, they are represented by a great thickness of lignitic joint clays, estimated at 600 feet.

In the Rio Grande Plain, southwest of San Antonio, and in the foothills of the Santa Rosa Mountains of Mexico, they attain great thickness, but differ entirely in detail from the north Texas extension, lithologically resembling very much the Fox Hills and Laramie beds of the Rocky Mountain region. They are well displayed from 12 miles above Eagle Pass to the Webb County line in Texas, and in Mexico south of the Sabinas River, north of the Santa Rosa Mountains, where they consist mostly of glauconitic sands, alternating with limestones, clays, and beds of lignitic coal. Near Eagle Pass they are probably 2,000 feet thick, and the whole formation may be even greater. They abound in fossil wood and bones, and are of great economic value as coal producers, the mines at San Felipe, Eagle Pass, and Sabinas being located in them. Fig. 62 illustrates the general character of the beds at Eagle Pass.


Fig. 62.—Section (in well) of 1,140 feet of the Eagle Pass beds. Vertical scale: 1 in. = 272 ft.

### THE EOCENE.

These beds overlie the highest Cretaceous, and consist of ferruginous sand, clays, and some impure marls. Lignite beds are very abundant. They mark the eastern and southern border of the portion of the Rio Grande Plain described in this paper, and the discussion of them does not properly fall within its province. A line from Littig through Lytton Springs, Lytle, and the southeast corner of Maverick County approximately marks their interior border. The artesian wells deriving water from these formations, such as those at Carizzo Springs, will be treated in a separate paper.

### ALLUVIAL DEPOSITS OF THE NEOCENE, PLEISTOCENE, AND RECENT EPOCHS.

These formations comprise the products of upland degradation, and have been laid down by streams at local base-levels. They are composed entirely of the débris of the Cretaceous uplands, except immediately along the Rio Grande and Colorado, where other material is mixed with such débris, the lithologic aspect of the beds remaining, however, practically unchanged.[1] As they are the chief source of the surface wells of the region, as well as of many important springs, they will be briefly described. The final classification of these deposits is not complete, but for convenience they will be temporarily grouped as follows:

The Uvalde formation ("Upland gravel"), supposedly of Pliocene age.
The terraces of the Colorado River.
The terraces of the Rio Grande.
The Onion Creek marl and allied deposits.
The terraces of the Nueces and Leona rivers.
The wash.
The fry-pan deposits.
The tepetate and "tierra blanca."

UVALDE FORMATION.

In the Rio Grande Plain lying off the foot of the Balcones escarpment, from San Gabriel to Devils River, and extending coastward many miles, there is a remarkable geologic formation or series of formations to which the name Uvalde has been given.[2] It consists of a vast deposit of gravel, composed almost entirely of rolled flint pebbles, with occasional pieces of limestone, partially embedded in a matrix of chalky marl and clays. Most of these materials have been derived from the decay of the Edwards limestone of the plateau, and spread like a mantle over the lower plain. In places where the marly material predominates over the pebbles, as on the high divide between the Colorado River and Onion Creek, southeast of Austin, east of San Antonio, and elsewhere, the formation weathers into dense black soils bearing great resemblance to the residual soils of the Upper Cretaceous beds, upon which the Uvalde formation rests. No fossils have as yet been found in it. It caps the higher divides in the Rio Grande Plain, and constitutes the highest terrace level in the canyon valleys of the plateau. North of the Colorado River it caps the hills west of Manor, having an altitude of 750 feet. South of the river it covers much of the high divide—the St. Elmo Plateau—between it and Onion Creek. The divide of the Blanco, Onion Creek, and Colorado drainages in northern Hays and Southern Travis counties is also made up largely of it. Similar remnants are found in many places southward toward

[1] None of the superficial and gravel deposits of the Texas region are of glacial origin, as people frequently suppose them to be. There is no reason to modify the opinion expressed by Dr. Roemer many years ago: "At the close of the remarks on the formations of the diluvium in Texas, it may be proper to point out that no trace is found in Texas of any drift blocks or gravel deposits of northern origin, which, of course, is quite in harmony with the well-known distribution of northern drift in the basin of the Mississippi, in which the erratic blocks, so common in the vicinity of the Great Lakes, nowhere extend as far as the Ohio, thus proving that the conclusion gained in Europe, that the erratic phenomenon is of northern origin, applies also to the American continent." (Kreidebildungen von Texas, 1852, p. 4.)

[2] Am. Geologist, June, 1891, p. 368.

San Antonio. The heights surrounding the latter city are largely capped by it. An instance is the Fort Sam Houston hill.

This flint-gravel formation is found along the lines of the Southern Pacific from Seguin to Del Rio, constituting, with few exceptions, the divides between the streams, and attaining a considerable thickness. Usually it extends up to the very foot of the Balcones hills, but occasionally, as in Uvalde and Kinney counties, its border lies farther south. A similar gravel constitutes the level prairie upon which the town of Spofford, in Kinney County, is situated, and extends northward to the foot of the highlands. This is 40 feet or more in thickness, and contains locally derived igneous as well as Cretaceous material.

Southwest of Uvalde, on the divide between the Leona and Nueces rivers, along the Eagle Pass and Carrizo Springs roads, the matrix of the formation is a fine silt that makes a stiff, black soil when wet. In the higher places and around the edges of the waterways in the wide, flat caletas the gravel is exposed. The gravel was at first embedded in the fine silt and covered by it. In high places the silt has been washed off the gravel and has accumulated in the caletas, the gravels having been too heavy to be moved by the sluggish waters. Along the stream ways in the draws the water has swept the silt away, leaving the gravel exposed in the lower part of the banks, while the silt forms the upper part.

Along the borders of the Rio Grande the surface of the formation, in spite of the incision of the drainage, still preserves well its original plain character. The altitude of the formation is higher in its western than in its eastern portion, its surface sloping both east and south. Its height above sea level between Spofford and Del Rio is from 1,000 to 1,100 feet; in the vicinity of Austin its level is between 650 and 750 feet; on the Rio Grande, near Santo Tomas, in Webb County, it occupies an elevation of between 600 and 650 feet. It does not cease at the Rio Grande, but extends far into Mexico.

The distinction to be drawn between the Uvalde formation of Rio Grande deposition and that of the rivers of the region farther east lies in the difference in the character of the material composing the gravels. As has already been noted, the gravel in the latter area is composed almost entirely of flints derived from the disintegration of the limestone of the Edwards Plateau. In the former region the material has been derived to a large extent from sedimentary rocks older than the Cretaceous and from eruptive masses occurring along the course of the Rio Grande above the mouth of the Pecos River. However, mixed with the gravels of igneous material and pre-Cretaceous sediments is a large percentage of pebbles of Cretaceous rocks, comprising Edwards limestone, other limestone, and flint pebbles, etc.

Another interesting peculiarity in this region is, the pebbles are frequently cemented into a firm conglomerate by a chalky matrix. Often the pebbles are very rare and the formation is represented by the limestone alone. The two hills in Webb County north of Santo Tomas, known as the Dos Hermanos (Two Brothers), are buttes capped by a white chalky indurated limestone containing a very small number of pebbles. The cap rock is almost a pure limestone of "tierra blanca" (white earth), as described on page 256.

Besides the distribution of the formation in the Rio Grande Plain, it extends up the canyons of the larger streams flowing from the Edwards Plateau, making the highest terrace, standing about 50 feet above the stream bed, and is composed mostly of flint gravel. As the streams descend from the plateau and approach the plain country, their canyoned valleys become wider and the Uvalde formation becomes more and more extensive, spreading out and covering all of the divides after the belt of the Balcones fault disturbance has been traversed.

It is evident that there was a long period of canyon cutting preceding the deposition of the Uvalde, and this period was probably during Miocene and early Pliocene time.

The coastward extent and relations of the Uvalde formation have not been investigated. It extends southward beyond Santo Tomas, in Webb County, and eastward of the International and Great Northern Railroad from Seguin to Laredo.

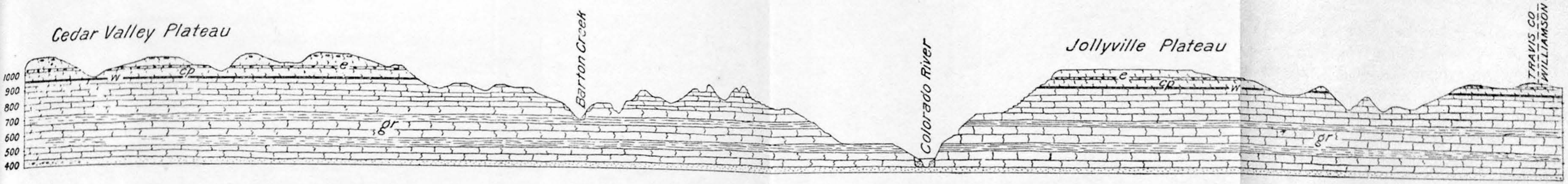
There can be but little doubt that the Uvalde formation is of the same age as the plateau gravel of Arkansas and northeast Texas and the Lafayette formation of the Gulf and Atlantic region east of the Mississippi. These have been correlated by Hill, by Penrose, and by McGee.[1]

Dumble has considered the Reynosa limestone of Penrose[2] as only a phase of the extensive upland gravel deposit described above,[3] and extends the term Reynosa so as to include the whole of the gravels. According to Penrose, this limestone is hard and whitish, and occurs about 50 feet above the Rio Grande in the town of Reynosa, State of Tamaulipas, Mexico. As we have not been able to make studies in the vicinity of Reynosa, we have not sufficient data to pass judgment on Dumble's treatment of Penrose's term. Dumble states that the seaward extension of the Uvalde (Reynosa) passes beneath the Gulf coast clays (Port Hudson clays of Hilgard).[3]

As nearly as can be determined, the deposition of the Uvalde formation took place in late Pliocene time.

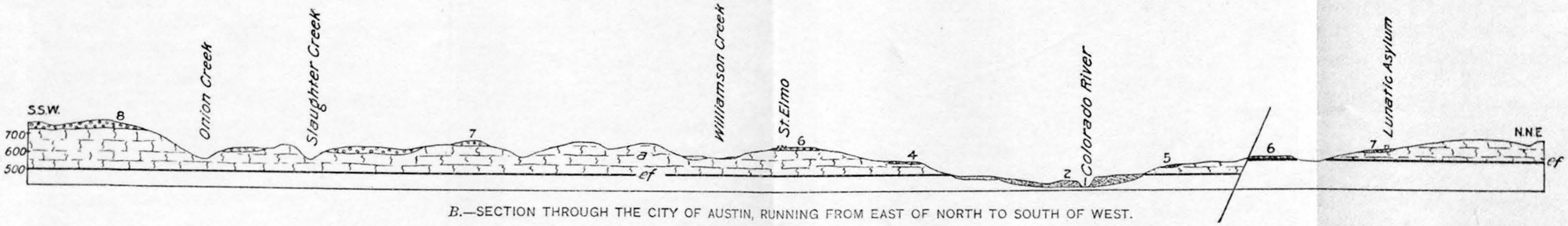
It does not appear to the writers that it is necessary to postulate a marine submergence or an absolutely horizontal deposition level to explain the Uvalde formation within the area considered in this paper. At most a large portion of the deposit was laid down by the process later described under the heading "The wash" (p. 254), when the country stood lower than now, but was undergoing gradual uplift, the streams

[1] The Lafayette formation: Twelfth Ann. Rept. U. S. Geol. Survey, Part I, 1891, pp. 347-521.
[2] First Ann. Rept. Geol. Survey Texas, 1890, p. 63.
[3] Bull. Geol. Soc. America, vol. 3, 1892, p. 230.



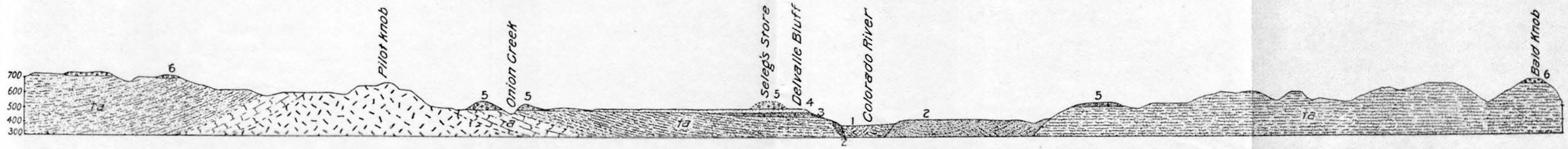

*A*.—SECTION FROM NORTHEAST TO SOUTHWEST ACROSS THE CANYON OF THE COLORADO RIVER, CUTTING IT AT A POINT 9¾ MILES NORTHWEST OF THE CAPITOL BUILDING AT AUSTIN.

Horizontal scale, 1 : 100,000, or, approximately, 1 inch = 1.6 miles. Vertical scale, 1 inch = 862 feet.




*B*.—SECTION THROUGH THE CITY OF AUSTIN, RUNNING FROM EAST OF NORTH TO SOUTH OF WEST.

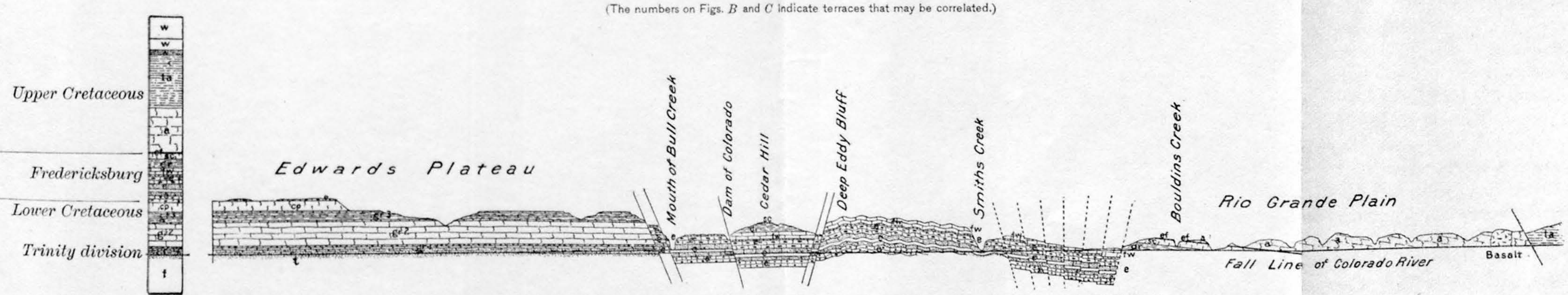
Same scale as *A*.




*C*.—SECTION FROM NORTHEAST TO SOUTHWEST, CUTTING THE COLORADO RIVER AT A POINT 9 MILES SOUTHEAST OF THE CAPITOL BUILDING AT AUSTIN.

Same scale as *A*.

(The numbers on Figs. *B* and *C* indicate terraces that may be correlated.)




*D*.—SECTION EAST AND WEST ALONG THE COLORADO RIVER AT AUSTIN, SHOWING THE BALCONES ZONE OF FAULTING.

Horizontal scale, 1 inch = 0.8 mile; vertical scale, 1 inch = 1,190 feet.

## CROSS SECTIONS IN THE AUSTIN QUADRANGLE.

*ta*, Taylor; *a*, Austin; *ef*, Eagle Ford; *sc*, Shoal Creek; *dr*, Del Rio; *fw*, Fort Worth; *e*, Edwards; *cp*, Comanche Peak and Walnut; *gr*, Glen Rose; *t*, Travis Peak. Heavy black line near top of Section *A* (*w*), Walnut formation; line at bottom of Section *D*, low-water line of Colorada River.

debouching upon the plain seaward of the Balcones scarp, and probably inundating its expanse at various flood periods. The shifting of the channels of the streams was also probably instrumental in distributing the débris.

### PLEISTOCENE TERRACES.

After the deposition of the Uvalde gravel sheet or sheets the rate of elevation of the land increased somewhat, and drainage valleys were successively cut through these older gravel formations. In the valley, not only in the general region of the Coastal Plain, but over the entire Central Denuded region, extending up to the breaks of the Plains and into the canyons, are many alluvial talus fans and terrace deposits, varying in number and character with the age, position, and character of the streams. Streams of larger magnitude, like the Rio Grande, Pecos, Colorado, Nolan, Brazos, and Red rivers, show more of these terraces than those of more recent origin.[1] Theoretically the different deposits must merge in their coastward extension into delta or estuarine beds, and are no doubt represented in the coast sediments.

### TERRACES OF THE COLORADO RIVER.

No better expositions of terrace phenomena can be found than those seen adjacent to the Colorado River as it crosses Travis County, and especially in that portion of its course lying to the east of the main fault line. These phenomena will be illustrated and described in detail in the forthcoming Austin folio of the Geologic Atlas of the United States. Their general relations are illustrated in the sections of Pl. XXXV, *A–C*. In this vicinity several categories of terraces can be distinguished, as shown in the accompanying table (p. 248), including various stages in the history of the Colorado River, from the oldest (Uvalde) plains on the high divides to the present river bottoms. The light-colored, sometimes chalky, matrix of the fundamental formations out of which the Colorado has eroded its valley affords a favorable foundation for the clear demarcation of the various terraces. Their elucidation is further facilitated by the material of the terrace deposits themselves, consisting as they do of three entirely distinct and contrasting materials, viz, (1) white flints and limestones derived from the Cretaceous formations of the Edwards Plateau; (2) the more consolidated and darker-colored materials of the Paleozoic formations of the central region, such as dark limestones, ferruginous sandstones, and a vast quantity of red feldspar and milky quartz; (3) vermilion loams derived from the Red Beds in which the present headwaters of the Colorado have their sources. Each terrace seems to be composed largely of one of these three classes of material, although sometimes an admixture is found.

[1] The terrace phenomena associated with the major streams, such as the Red, Brazos, Colorado, and Rio Grande, as well as those of the Trinity, Guadalupe, and Nueces, have been the subject of detailed investigation by the authors.

The Uvalde beds previously described may be considered the most ancient or earliest of this series of formations, while the present high-water flood deposits of the river are the most recent. Between these there are many steps or stages.

The following table shows the locality, height, and composition of some of the terraces recognized on both sides of the river. The altitudes given represent as nearly as possible the original deposition surface of the formation.

*Table showing relation of terrace phenomena on the sides of the Colorado Valley near Austin, Texas.*

| South side of river. | | | North side of river. | |
|---|---|---|---|---|
| Location. | Composition. | Altitude. | Composition. | Location. |
| | | *Feet.* | | |
| Hills 3 miles southeast of Manchaca | 1 | 775 | | |
| Hills southeast of Creedmoor | 1 | 725 | | |
| Hill at fork of Oak Hill and Manchaca roads | 2 | 675 | | |
| Terrell Hill | 4 | 675 | 1 | Plateau southeast of Sprinkle; Bald Knob, near Manor. |
| St. Elmo divide | 1 | 650-675 | | |
| Lytton Springs | 1, 2 | 635 | 2 | Lunatic asylum, Austin. |
| | | 625 | | |
| | | 615 | 2 | University of Texas. |
| Onion Creek marl | 1 | 550 | | |
| East of Seelig's store | 1 | 525 | 1, 2 | State capitol, Tillottson College, Walnut Creek. |
| | | 500 | 2, 3 | Fourth street terrace. |
| | | 480 | 2, 3 | Railroad station and city bridge bluff, Austin. |
| Delvalle Bluff, lower part | 2 | 475 | | |
| Bottom, Austin | 4 | 430 | 2, 3 | Minor bottom terraces of Colorado below Austin. |
| Stream way, Austin | 4 | 422 | 2, 3 | Stream way, Austin. |

1 = Rolled flints, white limestone, and other débris of Cretaceous formations of the Edwards Plateau.
2 = Largely quartz and feldspar from Burnet region granite, with mixture of Paleozoic limestone.
3 = Red Beds débris.
4 = Mixture of above.

On the north side of the river these terraces can be divided into two categories: (1) A series of higher levels above the present stream valleys, the integrity of which has been destroyed by erosion; and

(2) those nearer the present stream way, which maintain their original integrity. There are many of each class, and the distinction between the two is purely arbitrary. Of the first class are all those above the Capitol (Tillotson) terrace, on which the State capitol building is located, including those upon which the State insane asylum and the university are situated. The topographically lower, wide flats of the Colorado River, composed of fine sand and silt, form a large number of minor terraces and constitute the second class. The latest of this class is the alluvium of the present flood plain of the river.

*The Asylum terrace.*—This terrace lies some 215 feet above the Colorado River, and extends nearly 4 miles back from the present river bed, forming the summit of a high bench that usually marks the exterior margin of the so-called Colorado bottoms. In the city of Austin it commences north of the university and constitutes the sandy post-oak flat in the northern part of that city upon which the lunatic asylum stands. This terrace extends down the Colorado River, preserving its position at the summit of the bluff along the river flat, into the Bastrop quadrangle, where it may be related to the gravel on the divides on the south side of the river in the vicinity of Lytton Springs. West of Shoal Creek, as the line of the Balcones fault is approached, occasional remnantal patches have survived the vigorous erosion to which the formation has been subjected, but the terrace once extended to the foot of the escarpment.

We are not prepared to determine the stratigraphic position of the Asylum terrace south of the Colorado in the vicinity of Austin, although patches of it may still be preserved south of Barton Creek, on the road to Oatmanville, and in occasional places along the river bluff north of Delvalle. South of Onion Creek and extending far down the river this terrace has a great development, and probably spreads out over a large area. Its material consists largely of granite débris, granite pebbles, quartz, and red-colored plastic material derived from the decomposition of feldspar, brought by the river from the Burnet granite region to the northwest. There are also pebbles of schist and Paleozoic limestone from the Burnet-Llano region, and a little flint from the Edwards limestone.

The character of the material of the Uvalde formation shows that during the deposition of the oldest of these beds the Colorado River had probably not cut through the Cretaceous covering that overlay the Paleozoic area of central Texas. The Asylum terrace material, however, tells definitely that at the time of its deposition the river had cut into the older formations of the Burnet-Llano region and was vigorously eroding them.

There are several minor terraces below the Asylum terrace, the débris of which occurs in the university grounds. Some of these are as follows:

*The Capitol terrace.*—The next lower conspicuous terrace is that which forms the surface of the capitol grounds and the residence portion of Austin. The best-preserved example of this is in the city block northwest of the old Catholic Church. It can be traced eastward down the river, Tillotson University being located upon it. East of that institution it gradually merges into the wide flats of the Colorado. South of the river this terrace forms the top of the high bluff or valley wall of the Colorado bottom at the State Asylum for the Deaf and Dumb. Fragments of it can be traced far down the river. The material of this terrace is the same kind as that of the Asylum terrace; i. e., it consists mostly of granitic débris brought from the Burnett-Llano region. It is presumed that the finer silts which originally constituted a part of the Asylum and Capitol terraces have been removed by erosion. This terrace is also well shown west of Austin, between Shoal Creek and the river.

Some 40 to 60 feet above the river at low water and above the present limits of the highest water in times of overflow is a vast flat, or "second bottom," often several miles in width. The bottom in places consists of five or more terraces, forming steplike descents to the river. These are specially well shown near Montopolis Bridge. Their elevation one above another is only 5 to 10 feet. It is very evident that this whole area was once the flood plain of the Colorado, and, although not now subject to overflow, it is locally termed the Colorado bottom. The highest of the second group of terraces is that upon which the main business portion of the city of Austin is built. It is well shown along Fourth street, particularly just back of the Board of Trade building. About 10 feet below this is the top of the terrace exposed in the bluffs of the Colorado at the city bridge. The summit of this is about 60 feet above the river, and is composed largely of red loam. Four distinct later terraces, in all, can be seen above the present stream way in the cross section of the Colorado at Austin from Fourth street to the bluffs just north of the Deaf and Dumb Asylum.

The material of the lower group of terraces is usually fine sandy loam, with occasional beds of gravel or clay. At Austin excellent bricks are manufactured from the clay beds. During the time of the formation of these terraces the headwater drainage of the river had worked its way back and down into the Red Beds country, and was bringing down material similar to that which now gives the Colorado water its characteristic color, and from which the river derives its name.

*The present flood plain.*—The bed of the river is usually shallow and sandy, the stream occasionally cutting into the underlying bed rock. The higher flats are covered by water in time of overflow. The alluvium is reddish, similar to that of the flats above described, having been deposited largely by the "red rises," which are originated by the heavy rains that fall in the Red Beds region near the principal headwaters of the river. There are also occasional deposits of débris of the Paleozoic and Cretaceous rocks, mostly derived from the sediment of the "white rises," which usually originate in the drainage basin of the Llano River.

*Terraces on the south side of the Colorado River.*—These differ from those on the north side in two particulars: First, older and higher deposits are preserved on the south side of the river, such as those seen capping the hills 3 miles southeast of Manchaca and Creedmoor, at elevations as high as 750 feet. No stream deposits of equivalent height are found on the north side of the river. Second, below this level, occupying the same level as the Bald Knob Plateau near Manor, are several gravel-capped outliers. Some of these, such as the one at the fork of the Oak Hill and Manchaca roads, are composed of the old granitic débris, while the St. Elmo divide, occupying the same level, is made up entirely of flint material. This would indicate that at the time of their deposition a stream of the former material was coming down the Colorado, while the flint was being deposited by its laterals. This is the highest elevation at which the granitic material has been found. There are other instances of terraces on the south side of the river occupying the same level as those on the north side but composed of radically different material. For instance, the Onion Creek marls, as seen at Bluff Springs bluff, occupy an elevation almost identical with that of the Capitol terrace, while those of Delvalle are analogous to those of the bluff at the city bridge in Austin. The Onion Creek and Delvalle terraces are composed of calcareous marl derived from the Cretaceous formations, and the others of granite débris and of the fine brick-clay silt derived from Permian and granitic materials. These differences in the material at the same level on opposite sides of the river are readily explained: Barton Creek and Onion Creek, two of the principal laterals of the Colorado, drain only Cretaceous formations, and have brought down only material derived from them, while the granitic, Paleozoic limestone, and Permian materials were brought down by the Colorado itself.

## TERRACES OF THE RIO GRANDE.

The Rio Grande, although a stream of major magnitude, presents, between Del Rio and Laredo, terraces considerably different in detail from those of the Colorado River near Austin. The Uvalde formation extends to the edge of a high erosion bluff that forms the outer border of the stream valley of the Rio Grande from Del Rio an indefinite distance southward. The bluff is steep, and its summit is 150 or 200 feet above the river. The foot of the bluff marks the outer margin of a terrace usually several miles wide and about 120 feet lower than the top. The descent from this wide terrace to the present flood plain of the river is by a succession of minor terraces, usually three or four in number. The usual relation and succession is indicated by fig. 63. Plate XXXVI, reproduced from a photograph, shows the topographic expression.

At Palafox, Webb County, there are remnants of a greater number of terraces than were seen elsewhere. Fig. 64 illustrates their occurrence.

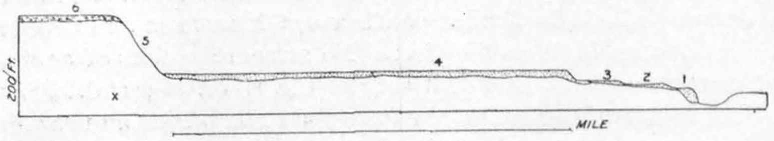



FIG. 63.—Diagrammatic cross section of the Rio Grande terraces. 1, 2, 3, lower silt terraces; 4, wide silt terrace; 5, bluff of Upper Cretaceous or Eocene; 6, Uvalde formation; x, underlying bed rock.

rence. It seems quite probable that once there may have been more terraces than now exist, some having been almost or entirely destroyed by the later erosion of the river.

ONION CREEK MARL AND ALLIED DEPOSITS.

Occupying an intermediate altitude below the level of the Uvalde formation and above the present flood plains of the numerous secondary streams of the Edwards Plateau and Rio Grande Plain, there is a formation which consists of a faint yellow or salmon-yellow calcareous marl, sometimes accompanied by fine pebble conglomerate, all of which is derived from the Cretaceous limestone material. It is usually less than 50 feet in thickness.

A type locality for this kind of material is the valley of Onion Creek from near its mouth to the eastern end of Pilot Knob Canyon, and from the western edge of Pilot Knob Canyon up that stream west beyond

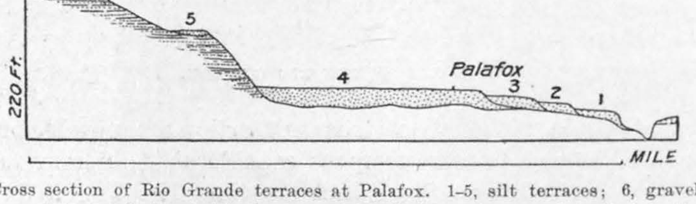



FIG. 64.—Cross section of Rio Grande terraces at Palafox. 1-5, silt terraces; 6, gravel terrace; 7, Eocene sandstone and clays, with the topographically lower terraces deposited on their eroded surface.

Buda. Here, in the portion first described, these marls and pebbles underlie quite an extent of black-land country, and outcrop along the bluffs of the Colorado north of Delvalle about 100 feet above the present river level. Just west of Pilot Knob, on Rinard Creek, the fossil bones of a bison were found in these marls, which, according to Mr. F. A. Lucas, of the Smithsonian Institution, is *Bos scaphoceras* of

ESCARPMENT AND TERRACE OF THE RIO GRANDE AT EAGLE PASS, TEXAS

The upper plain is the Uvalde formation level.

Cope,[1] a species hitherto found only in Nicaragua, where it is associated with an abundant fauna regarded as the equivalent of the Equus fauna of early Pleistocene age.

Marl of this character has been observed south of the Colorado in nearly all the streams of the Black and Grand prairies and the Edwards Plateau, notably along the Conchos, Llano, Guadalupe above Kerrville, the Pedernales above Fredericksburg, at many places along the East and West forks of the Nueces, and on the Frio, Medina, Devils, and Pecos rivers.

North of the Colorado similar deposits occur in all the larger secondary streams of the Central Denuded and the Black Prairie and Grand Prairie regions as far as the Ouachita Mountains, especially in the valleys of the Nolan, San Gabriel, Cow House, Leon, Bosque, Paluxy, and Trinity. Like the Leona formation, next to be described, and some of the later terraces of the major stream ways, the Onion Creek formation is usually deposited in furrows which have been eroded in the older Uvalde formation.

### TERRACES OF THE NUECES AND LEONA RIVERS, AND THE LEONA FORMATION.

The highest terrace of the Nueces River, where it flows through the canyon of the plateau, is the equivalent of the Uvalde formation, as has been previously stated. In the same canyon, below the Uvalde formation terrace, is a second, narrower terrace. Where the stream passes southward out of its canyon the Uvalde formation, as previously stated, spreads out over the divides as soon as the faulted zone is crossed. There is a bluff or low escarpment along the western side of the Nueces from the boca of its canyon entirely across Uvalde County. The eastern side is bounded by an escarpment almost as far south as the crossing of the Southern Pacific Railroad. Four and a half miles south of the railroad the escarpment or bluff, on the eastern side, begins again and continues southward entirely across the county. There is no escarpment on the east side of the river in the space between the Southern Pacific Railroad and a point 4½ miles below it, so that a wide flat extends from the Nueces eastward toward the Leona, limited on the south by a sinuous escarpment extending between the two rivers from a point 3 miles south of Uvalde, near the Leona, westward to a point 4½ miles below the crossing of the Southern Pacific Railway over the Nueces. Where the river is inclosed by escarpments there is a terrace flanking it on each side, occupying a position in general 25 to 50 feet above the stream bed and 75 to 100 feet below the Uvalde formation level. The material of the terrace is fine calcareous silt at the surface, and grades downward into coarse gravel.

Below a point 3 miles south of Uvalde the terraces of the Leona are exactly similar to those of the Nueces. The old (Pleistocene) flood-

[1] Jour. Acad. Nat. Sci. Phila., Vol. IX, 1895, pp. 457–458.

plain deposits of the two streams fused through the low saddle or plain above described. The town of Uvalde is situated on the eastern side of the saddle.

The name Leona formation is proposed for the deposit making the first wide terrace of the Nueces and Leona rivers, below the level of the Uvalde formation, and for the flood-plain deposit extending westward from Uvalde on the Leona to the Nueces River. The Leona formation is a Pleistocene flood-plain deposit bearing certain definite relations to the older Uvalde formation and the streams that have laid down its component materials. Where the streams are flanked by bluffs or escarpments it occurs as a terrace; in one instance it extends across a saddle and constitutes the junction of the flood-plain deposits of two neighboring streams. The width of the formation in any particular portion of the course of a stream depends upon the topography of the country at the time of its deposition. Where the stream flowed through a narrow canyon the Leona formation is a narrow terrace; if the surrounding country was flat the flood that brought the material down may have covered many square miles, spreading the silt and gravel over vast expanses of country. Such is the case along the west side of the Frio.

We designate as type localities for this formation the courses of the Nueces and Leona rivers in the Uvalde district, as has been indicated in the preceding discussion (see map, fig. 70, p. 275). The formation presents the same general characters along the Frio River, and the wide silt terrace that occurs about 150 feet below the Uvalde formation level along the Rio Grande is undoubtedly of the same age. The Leona may ultimately be correlated with the Onion Creek formation.

Along both the Leona and the Nueces rivers there is usually a smaller terrace, some 10 to 20 feet below the Leona terrace level. It is rather insignificant.

The present stream deposit consists mostly of limestone and flint pebbles derived from the Edwards Plateau.

### THE WASH.

The edges of the outcropping ledges of hard limestone forming the scarps and crests of the hills in the Plateau country are shattered into fragments by alternate expansion and contraction, due to the diurnal variation of temperature. The loosened pieces may remain temporarily in situ or may roll down the steep slopes. When a sudden rainfall occurs they are washed down the slopes by the torrents, and scattered in great sheets over local lower levels.

A peculiarity of the cloud-burst type of rainfall of the semiarid region is that, although the water forms great torrents and spreads out into sheets, it usually disappears entirely upon reaching lower, more level areas, where it is either imbibed by the underlying rocks or evaporated without being conducted into the larger streams. Sedimentary material deposited by these rain storms constitutes formations of vast areal extent in the arid and semiarid region, and may be appropriately designated "the wash."

The character and appearance of the wash vary with the geology and topography of the country in which it occurs. In southern Texas, along the Balcones escarpment and up the valleys and canyons of the Edwards Plateau, where the country rock is mostly the Edwards limestone, it consists largely of flint nodules, which are brought down from the decaying plateau and scattered over the Rio Grande Plain at its base. In the trans-Pecos mountain region the wash is of the type of talus fans, spreading out at the mouths of the mountain canyons over the margins of the bolson[1] plains. The surface of the Llano Estacado is all wash.

This material is found extensively in both the plain and the plateau, especially in the wide hemispherical headwater valleys of the minor stream ways of southwestern Edwards and northern Kinney counties, such as that of Griffin, Dry Sycamore, and Hackberry creeks.

The wash is of especial interest because it enables us to a certain extent to interpret the origin of some of the older deposits of the Uvalde and Leona type. Each of these formations begins as a ribbon of aggradational material in the narrow canyons of the plateau, and flares out into a broad sheet in the coastward regions. The Uvalde gravel, for instance, widens out immediately below the Balcones scarp line; the Leona widens out a little farther on. These sheets may ultimately expand into the broad alluvial tracts of the Coastal Plain.

### FRY-PAN DEPOSITS.

The peculiar basin valleys to which the cowboys have given the appropriate name "fry-pan" are pouch-like indentations into monoclinal escarpments opposing the course of a stream. The wider or flaring portion of the valley, constituting the body or pan portion of the fry-pan, lies upstream, and the valley constricts downstream, forming the handle of the pan, until the stream cuts its way out through a narrow gorge. In the arid and semiarid regions at times of flood the water is checked in these gorges, causing a precipitation of the material held in suspension. Typical deposits of this character are found in western Uvalde County, above the point where Turkey Creek crosses the monoclinal scarp line of the Anacacho Mountains, and above that where Elm Creek cuts across the monoclinal scarps of the Shoal Creek limestone. The southern end of the great Pecos Valley between the Rocky Mountains and the thirty-first parallel, where the plaza country ends against the northern edges of the Edwards and Stockton plateaus, is a gigantic fry-pan of this character. The fry-pan alluvium by marginal gradation merges into the alluvial deposits described above as "the wash."

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[1] Bolsons are basin valleys which have not, or had not originally, any outflowing drainage, and are lined with sedimentary débris derived from the surrounding country. Hill, Descriptive topographic terms of Spanish America: Nat. Geog. Mag., Sept., 1896, Vol. VII, No. 9, p. 295.

### TEPETATE AND "TIERRA BLANCA," CHEMICAL LIME DEPOSITS.

Throughout the limestone regions of the hot climates of America a superficial crust of white-lime material is found, called tepetate. Sometimes it is comparatively free from foreign material, or occurs as the matrix or cement of conglomerates. This is a concentrate of the lime which has been dissolved from the surface, transported in solution by the torrential streams, and redeposited through evaporation. In some cases it infiltrates downward into the embedded matrices of permeable formations and is there consolidated, in which case it is called "tierra blanca." If the country rock is a chalky limestone, the slightest rainfall is sufficient to take some of the lime in solution and redistribute it over the surface as "tepetate." For instance, the outcrop of the marly Eagle Ford beds in Kinney and adjacent counties of Texas is everywhere coated by this secondary deposit. The tepetate is forming great incrustations around the margins of the bolson plains of northern Mexico. The material often constitutes the matrix of the Uvalde formation gravel along the interior margin of the Rio Grande Plain and the Black Prairie region south of Austin.

## IGNEOUS ROCKS.

### KINDS OF IGNEOUS ROCKS.

All of the igneous rocks of the Rio Grande Plain are very basic, belonging to various types of basalt, viz, nepheline-basalt, nepheline-melilite-basalt, orthoclase-basalt, plagioclase-basalt, and limburgite.[1] Phonolite occurs at a considerable number of localities between the Frio and Nueces rivers, on both sides of the Southern Pacific Railroad.

The rock from Pilot Knob, Travis County, was the first studied.[2] It is nepheline-basalt. Mr. Whitman Cross has examined for us thin sections and specimens from a great number of localities. In the vicinity of Uvalde phonolite and all the types of basalt above mentioned are found. In Kinney County plagioclase-basalt is found at several localities, one of which is Pinto Mountain. At some places this contains a small amount of nepheline; an instance is the capping of Los Moras Mountain.

### MODE OF OCCURRENCE OF THE IGNEOUS ROCKS.

The igneous rocks all occur along the interior margin of the Rio Grande Plain, usually on the downthrown side of the Balcones fault. The geologic relations of the Pilot Knob area have been described in some detail in an article entitled "Pilot Knob, a marine Cretaceous volcano."[3] The central mass is a volcanic neck or stock of nepheline-basalt that has been pushed up through the Austin chalk, which is

[1] These rocks were identified by Mr. Whitman Cross.
[2] Hill and Kemp, Am. Geologist, Nov., 1890, pp. 292–294.
[3] Hill, Am. Geologist, Vol. VI, 1890, pp. 286–292.

often marmorized along the contact and lies against the flanks of the igneous mass, dipping away from it on all sides.

The basalt has an imperfect columnar structure, nearly vertical at the south extremity of the hill and nearly horizontal at the north side. The flat region between the basaltic hills and the chalky perimeter of the igneous area is filled with a soft, yellow, amygdaloidal, exfoliating material, some of which is undoubtedly the product of basaltic decomposition, while in other places it resembles volcanic ash. . . . In any direction from the basaltic hills which form the center of the whole outcrop, the average distance to its edge is about one-half mile, where excellent contacts with the chalk are found. The chalky stratum forming the margin of this area throughout its whole extent is crenulated into gently waving undulations and presents different degrees of hardness. In places of direct contact with basaltic material the chalk is converted into hard marble; when the ash-like material intervenes between the basalt and chalk the latter retains its soft, unaltered, pulverulent nature.[1]

The occurrence of volcanic material interbedded with chalk is adduced in the same paper as evidence to prove that a part of the volcanic disturbance took place as early as the latter portion of the time in which the Austin chalk was deposited.

Intrusive sheets and dikes of soft, much decomposed material of a basaltic nature occur at numerous other places within a radius of 10 miles.

The only other areas that have been studied in detail are the Uvalde and Brackett quadrangles. The basalts and phonolite probably are all intrusive. No undoubted lava flows and no pyroclastic material of any kind have been found. The rocks present four modes of occurrence: (1) Large laccolithic masses, one of which lies south of the Southern Pacific Railroad west of the Nueces River. (2) Stocks, bosses, or necks, some of which are merely protuberances above an underlying laccolith, as Sulphur Peak (see Pl. XXII). (3) Horizontally intruded sheets (sills) that have been forced between overlying strata from one of the stocks or necks. The basalt cappings of numerous hills, such as those of Pinto (Pl. XXXVII), Los Moras, and other hills, were probably originally sills. (4) Dikes.[2] We have seen no rocks younger than the Anacacho beds that have been cut by the intrusive rocks of this region. In Uvalde County gold has been reported to occur in rocks in contact with this igneous material.

## ARRANGEMENT OF THE STRATA.

### THE MAIN SYSTEM OF DIPS.

The apparently horizontal Cretaceous rocks are laid upon each other in sheets of great persistence and regularity, as is shown in many of the illustrations in this paper. Each of these individual sheets ultimately thins out and disappears beyond the area considered in this

[1] Hill, loc. cit.

[2] The mode of occurrence and petrographic characters of the rocks in the region of Uvalde will be discussed in detail in a paper soon to be published by Mr. Vaughan and Mr. Cross.

paper, but the change is so slow and gradual that the beds are practically uniform throughout the region under discussion. Their arrangement usually appears horizontal to the eye, but in fact the strata are all slightly tilted or inclined toward the coast, so that a rock sheet outcropping at the surface in one locality may be carried down by this dip from 10 to 100 feet beneath the surface a mile or two seaward.

North of the Colorado River the beds, as a whole, have the uniform gently tilted monoclinal arrangement characteristic of the general structure of the Coastal Plain, whereby the various rock sheets dip beneath the surface and each other, toward the coast, at an angle greater than the surface slope in the same direction. The dip of the Cretaceous rocks on a line drawn eastward from Dublin through Waco is estimated to be about five times as great as the average continental slope of the surface in the same region.

South of the Colorado the direction of the dip of the rocks is likewise coastward, but the angles of dip are more varied. The rocks of

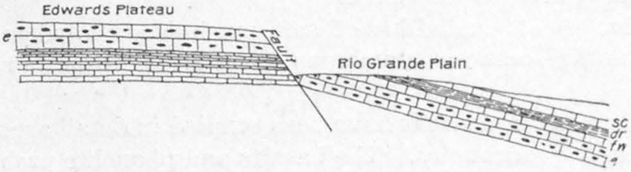


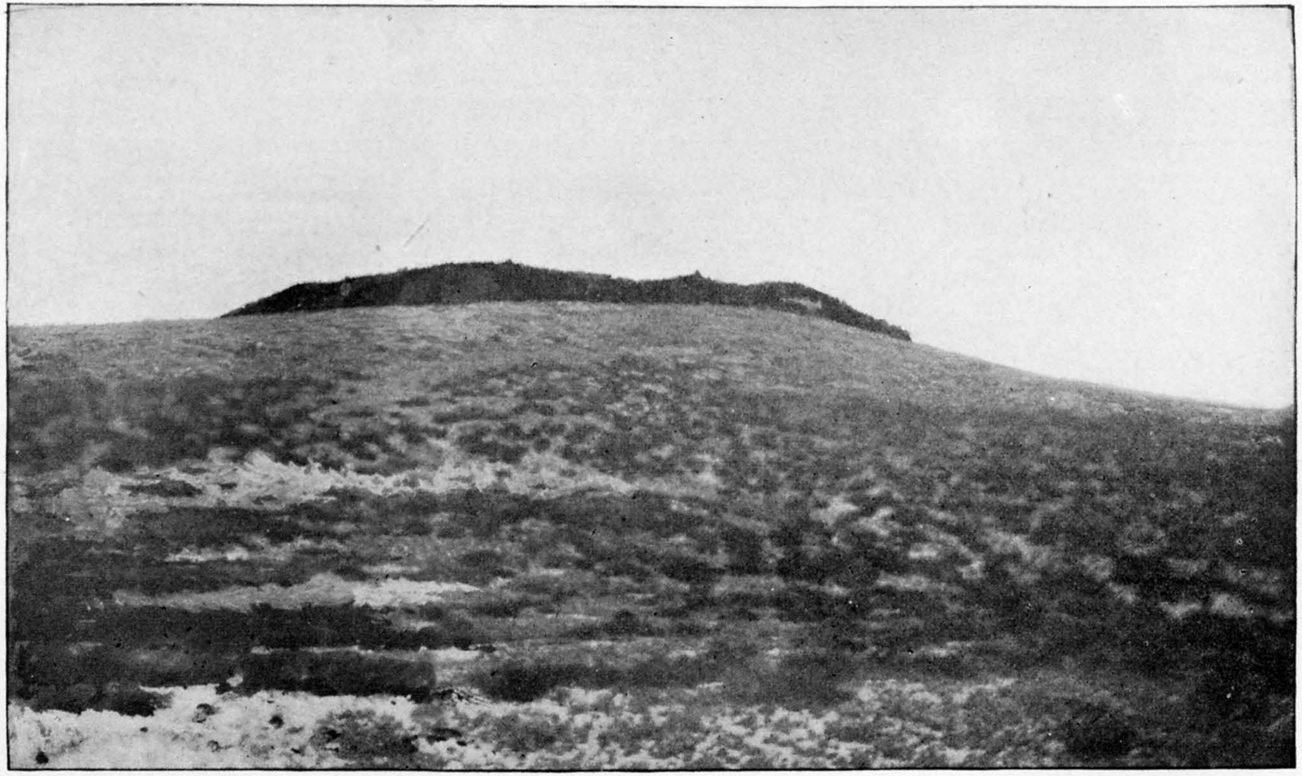

FIG. 65.—Diagrammatic representation of increase in steepness of dip south of Balcones fault line. *sc*, Shoal Creek; *dr*, Del Rio; *fw*, Fort Worth; *e*, Edwards.

the Edwards Plateau are more nearly horizontal than those of the Rio Grande Plain; their average dip is less than 10 feet to the mile, while in the plain the dip ranges between 50 and 100 feet to the mile. In the plateau the average dip conforms almost precisely to the average surface slope, so that a single formation—the Edwards limestone—constitutes nearly the whole surface. In the plain the coastward dip is much more rapid than the slope of the surface, so that each formation has a narrow outcrop and quickly disappears beneath its overlying coastward neighbor.

This change in the dip coincides approximately with the Balcones scarp line, of which an account has already been given (p. 203), and more precisely with the Balcones fault line.

BALCONES FAULT ZONE.

The abrupt southern termination of the Edwards Plateau between the Colorado River and the Rio Grande and the sudden fall in altitude from the summit of the plateau to the lower level of the Rio Grande Plain has been produced by faulting. The line of faulting follows closely the foot of the Balcones escarpment, and for that reason has been named the Balcones fault zone. This fault zone is one of the most important features in the geologic structure of Texas, because it is the only structural break in the continuity of the post-Paleozoic strata in



CAP OF PINTO MOUNTAINS.

The basaltic cap was probably a sill originally.

the vast stretch of country between the Gulf of Mexico and the Rocky Mountains. It makes the southern boundary of the Great Plains region, because to it the Balcones escarpment, which is the southern boundary of the Plateau of the Plains, owes its existence.

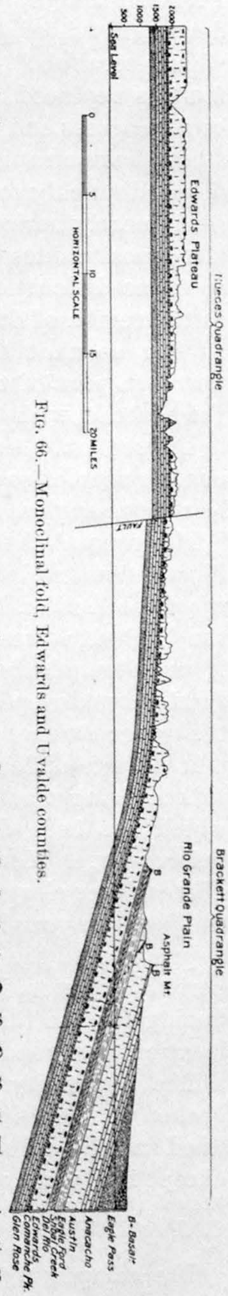



FIG. 66.—Monoclinal fold, Edwards and Uvalde counties.

The strata on the seaward side of the faults have been dropped down, so that any particular stratum—the top of the Edwards limestone, for instance—lies 500 to 1,000 feet lower on the coastward or downthrown side of the fracture than on the interior or upthrown side. Another statement of the same idea is: A stratum on the upthrown side of the fault zone is opposed on the downthrown side by a stratum that belongs 500 to 1,000 feet higher in the geologic section. This explains why the Cretaceous rocks at the surface along the northwestern margin of the Rio Grande Plain belong, geologically speaking, hundreds of feet above the summit rock of the plateau, although the latter are 500 feet higher, topographically, than the former, and it enables us to understand certain laws, not hitherto recognized, that govern the occurrence of much of the underground water in the region. These laws will be elaborated under the heading "Source of the underground waters of Rio Grande Plain" (pp. 313–316).

The fault zone really consists of many faults having subparallel directions, all concentrated within a narrow belt of country, as is shown in the Austin section (Pl. XXXV, *D;* figs. 65, 66). By an examination of this cross section of the faulted belt it will be seen that interiorward there are major faults of large downthrow, attended to the seaward by numerous smaller ones. The effect of the numerous faults is to break the regularity of the dip of the strata and to chop them into numerous blocks tilted at various angles within the faulted zone (fig. 67). To trace and map in detail the many small faults and folds would be an almost impossible task, the sum of them all amounting to a general downthrow of several hundred feet.

The series of subparallel faults has produced a narrow belt of country, occurring in Travis and through other counties southwest, the surface of which is composed of the outcrop of the Edwards limestone, which has fallen below the level of the Edwards Plateau. As this zone is of small area and of exceptional character, we do not devote much space to it. It usually lies between parallel faults. Its type is splendidly developed between Manchaca and Oatmanville, in Travis County. From the rough, broken character of its rocky surface, clothed with a dense growth of junipers and small scrubby oaks, it is locally called "hard scrabble." In Uvalde County, west of the Nueces River, the fault is replaced by a simple monoclinal flexure. Instead of being broken across and dropped down bodily along the plane of fracture, the strata have a rapid increase of dip at the edge of the plateau, so as to descend quickly to a lower level, and then, by another curve, become again nearly horizontal. The two bodies of nearly horizontal strata on the opposite sides of the flexure thus stand at very different levels, and so far as dislocation is concerned the result is practically the same as in the region of faulting. The effect on the topography is also quite similar, so that the line of the Balcones is not interrupted. The northern part of this flexure, connecting the gently dipping rocks of the plateau with the steeply dipping rocks of the scarp line, is shown in the section. Accompanying the fold are numerous small faults that usually are not parallel to the axis of the folding, but cut it at more or less acute angles.

The general character of the fault line is given in the accompanying illustrations and maps. Its details will be shown more thoroughly upon the Austin, Brackett, and Uvalde folios of the Geologic Atlas of the United States, now in preparation. Near Austin, just north of Mount Bonnel, the lowest strata of the Upper Cretaceous, the Eagle Ford, have dropped down so that they now occur against the edges of the upper 100 feet of the Glen Rose beds, as shown in fig. 76 (p. 315).

It is quite difficult to determine with accuracy the time during which this faulting was developed. We know that the Uvalde formation, which has been deposited since this faulting began, is probably of Pliocene age. Its deposition did not take place until after the Edwards Plateau was elevated into its present position relative to the plain and the deep canyons had been cut into it. That period of vigorous erosion approximated the close of Miocene time. As the faulting preceded the period of erosion, it is very probable that it may have been in operation during Eocene time. There is but little doubt, however, that the dislocation continued through later epochs.

The continuity of the stratified rock is likewise, though rarely, broken in places by dikes and sills of igneous rock. These are confined mostly to the Rio Grande Plain, and are quite unimportant except in Uvalde and Kinney counties.

## CAPACITY OF THE VARIOUS ROCK SHEETS FOR WATER.

The composition, texture, sequence, and arrangement of the rock sheets constituting the systems of underground waterworks in the region discussed having been described, the part that each particular rstatum plays in the distribution of water will now be considered.

Of the numerous beds mentioned, all those composed of impervious material, such as clays and close-textured limestones, may be considered as non-water-bearing, and their function in the transmission of water is solely that of control, not of supply, since they retain the water in the water-bearing beds. Nearly all of the Cretaceous rock sheets south of the Colorado lying above the Fredericksburg division belong in this category of impervious beds overlying the water-bearing strata. There may occasionally be a few arenaceous layers, sufficient to furnish a scant supply for a few dug wells, but as a rule, as all those who have endeavored to obtain wells in them can testify, the underground water they contain is neither abundant in quantity nor good

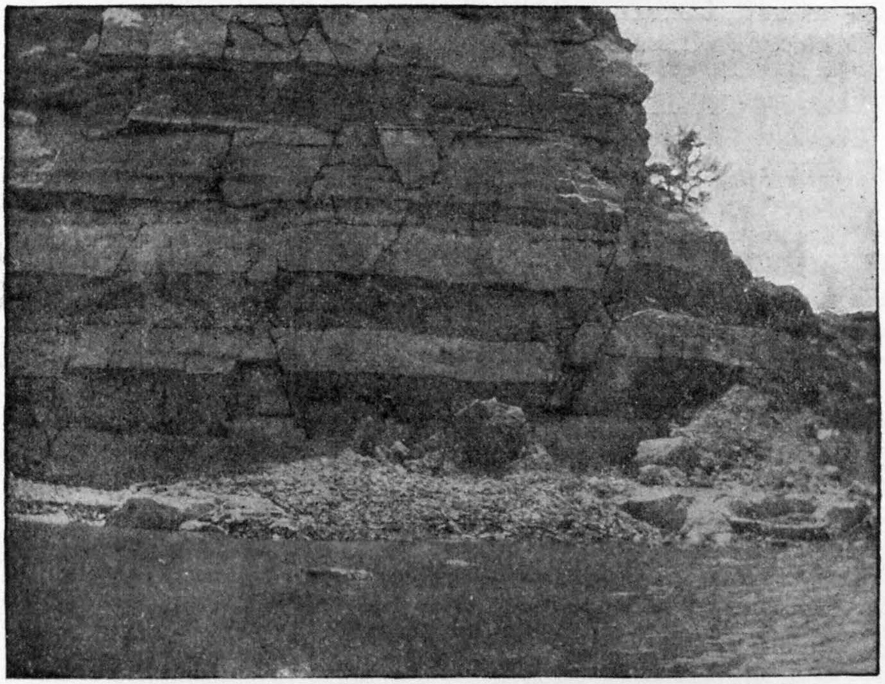


FIG. 67.—Block faults in Edwards limestone at mouth of Bee Creek, Travis County.

in quality. The inhabitants of the regions underlain by the Upper Cretaceous rocks (mostly the so-called "joint clays") have usually had poor success in obtaining surface wells, and have been obliged to rely largely upon cisterns for their domestic supply. We have personally seen but few springs from Cretaceous strata above the Fort Worth limestone. Notwithstanding their great thickness and areal extent their outcrops are usually dry.

Rocks of open texture, such as sands, conglomerates, porous, chalky limestones, and massive rocks broken by joints, fissures, honeycombs, or other openings, are usually water-bearing. These are mostly found below the Del Rio clay.

It is from these lower beds that we are to expect artesian supply in the region south of the Colorado River—south of the Brazos, in fact. That certain of these beds are completely charged with water is demonstrated by observations upon the source of the springs of the plateau region and by the experiments of well-drillers.

*Summary of the water capacities of the formations of the Cretaceous system.*

| Formations. | Permeability. | Conductivity. |
|---|---|---|
| Eagle Pass | Poor (*a*) | Poor. |
| Taylor | Poor | Poor. |
| Austin | Fair | Poor. |
| Eagle Ford | Poor | Poor. |
| Shoal Creek | Poor (*a*) | Poor. |
| Del Rio | Poor (*a*) | Poor. |
| Fort Worth | Poor (*a*) | Poor. |
| Edwards | Poor and good | Poor and good. |
| Comanche Peak | Poor and good | Poor and good. |
| Walnut | Poor | |
| Glen Rose | | |
| Upper | Fair (*a*) | Fair. |
| Medial | Fair (*a*) | Poor. |
| Lower | Fair (*a*) | Fair. |
| Travis Peak | Good | Good. |

*a* Exceptional beds having the opposite conditions may occur in these groups of strata.

The chief water-bearing strata of the Edwards, Glen Rose, and Travis Peak or Gillespie formations will now be described more fully.

The barren limestones of the Edwards formation have little in their appearance to suggest the occurrence in them of underground water; yet in the series there are several horizontal layers of water-bearing rocks which, when cut by streams or penetrated by the well-digger, afford abundant and pure supplies.

The proof that the Edwards limestones are water-bearing is threefold, as follows:

1. Great springs can be seen bursting out of them at the headwaters of the Llano, Guadalupe, Frio, and Nueces rivers.

2. The records of the Manor, San Marcos, and San Antonio artesian wells all point to these rocks as the source of their waters.

3. Deep nonartesian wells are produced from them on the summit of the plateau.

Some of the water-bearing strata are pervious arenaceous beds intercalated between the limestones. The character of these beds is often difficult to discover, the fact that they are very calcareous causing them to resemble the limestone ledges, and their surface being often coated by calcareous incrustations which completely obscure their true nature. Where they are exposed, so that they can be studied, they are seen to consist of from 20 to 50 feet of pervious beds, composed of very fine particles of sand embedded in a white or yellowish limy matrix.

The horizontal distribution of water in the Edwards formation is also facilitated by the occurrence in the series of certain limestone strata that have been honeycombed or made cavernous by underground solution. The limestones, as exposed in cliffs, may appear hard, durable, and of homogeneous nature, but the interior when examined may prove to be very heterogeneous. There will be hard and soft spots, the latter being much more soluble than the former. The soft, soluble portions of the limestone may represent the original loci of fossils, small particles of iron pyrites, or tubular molds of fucoidal character. The irregular cavernous decomposition of the Edwards limestone is well shown in the bluffs along the Colorado River west of Austin. There are fucoidal layers, which weather into honeycombed rock, and peculiar red cavernous blotches develop in certain fossiliferous horizons of the massive limestone. (See fig. 75, p. 315.)

Whatever may be the origin of the honeycombed beds, it is a fact that they transmit immense quantities of water, and it is through them and the arenaceous layers of limestone that the headwater springs of the rivers of the Edwards Plateau above mentioned and the artesian wells of the San Antonio system deliver their water at the surface.

The impervious roofs and floors of the water-bearing limestones can not be so sharply delineated as those of the water-bearing strata of other formations, owing to the fact that the various limestone strata resemble one another so much as to be indistinguishable for mapping purposes. Furthermore, the paleontologic and lithologic research thus far conducted has not been sufficient to secure data to differentiate and define them accurately.

While water is distributed horizontally by thin pervious beds, separation planes, and cavernous strata or honeycombed rock, it is also distributed vertically in the Edwards limestone by joints, fissures, and crevices. Much more concerning the water-bearing beds of the Edwards limestone may be ascertained when that formation is made the subject of still more detailed investigation.

In southwestern Edwards County there are several water-bearing layers that occur at various levels in the Edwards limestone from the base upward. These layers, in ascending series, supply the Kickapoo springs, the Black water hole, and the Justice spring. The supply from the uppermost layer is trivial and unreliable in character; the other two are of great economic importance. The ranchmen of the summit region of the plateau bore deep wells down to these strata.

One of the water-bearing strata of the Edwards beds lies approximately 300 feet below the summit of the formation in sections measured at the headwaters of the Frio River in Edwards County. Another measurement makes the distance 380 feet below the Del Rio clays. The uppermost of the embedded water-bearing strata of the Edwards formation at Austin, San Marcos, and Manor, so far as can be estimated from the meager artesian-well records at hand, lie about 50 feet below the summit of the formation. These beds are further discussed under the head of "Availability and limitations of the underground waters" (p. 316).

The Glen Rose formation contains considerable quantities of water, as is well shown by springs, surface wells, and artesian wells. Some of the springs can be seen in the stream cuttings near Anderson's mill, in Travis County. There are springs along the courses of many of the streams that have cut down into it.

The Travis Peak and Gillespie formations contain a greater quantity of water than any other beds of the Comanche series. The water contained in the Travis Peak and Gillespie formations has previously been spoken of as Trinity water, because the name Trinity has been applied to the sands at the base of the Comanche series farther north. Since the name Trinity is applied to the division including the basal sands, which in the Austin region have been designated the Travis Peak sands and the Glen Rose formation (called also farther north the Paluxy sands), it seems best to dispense with the name Trinity as applied to the water derived from the lowest sands, and to use it only for the waters of the Trinity division as a whole. Therefore, we shall speak of the Travis Peak or Gillespie water and the Glen Rose water, as well as of the Paluxy water. Numerous springs, such as those in the vicinity of Travis Peak and Fredericksburg, derive their water supply from them. There are also numerous surface wells where they outcrop, and many artesian wells are supplied by them where they are embedded. The water that they contain is also the purest found in the strata of the Comanche series. The quantity and purity of the water increase as one goes downward in the series. The water of certain of the higher beds of the Glen Rose and Edwards formations is strongly impregnated with mineral matter and must be cased off in wells, for it is not potable.

## UNDERGROUND WATER OF THE REGION.

### WATER SUPPLY OF EDWARDS PLATEAU.

The summit region and the canyons of the Edwards Plateau present different conditions for supplying underground water. The summits are so high above the surrounding regions that both the surface and the underground water naturally drains from them toward the lower canyons. Therefore the conditions on the plateau and the canyons will be discussed separately.

The whole plateau is underlain by the water-bearing strata of the Edwards, Glen Rose, and Travis Peak formations. The upper strata of the Edwards compose the summits, while the Glen Rose beds are in part exposed in the deeper canyon cuttings, as can be seen in the Guadalupe, Nueces, Frio, and other streams. There are strong probabilities that west of the Frio below the lowest exposures of these beds there are still lower strata containing a great amount of water.

No artesian water has ever been obtained from wells sunk on the summit of the plateau, although several drillings have been made, but in most places nonartesian water is found.

## NONFLOWING WELLS.

The horizontal arrangement of the strata of the Edwards Plateau is not favorable for the obtainment of artesian water upon the summit areas. As the uplands are higher than the outcrop of any known water-bearing beds occurring beneath them, the hydrostatic pressure necessary for flowing wells can not exist.

The ranchmen of the summit region rely solely upon deep pumping wells for stock water. These wells are from 300 to 500 feet deep and obtain their water from the water-bearing strata of the Edwards beds, described in the previous section.

We have seen many records of the deep wells sunk on the summit of the Edwards Plateau in Kerr, Kimble, Menard, Schleicher, Sutton, Edwards, Valverde, and Crockett counties, and in nearly all instances they are drilled to the depths above mentioned through the solid Edwards limestone. Their depths check closely with the measurements of the rock sections in the adjacent canyons.

We present the following records of the deep nonartesian Edwards limestone wells of the plateau region, including in the list some on the monoclinal slope constituting its southern edge in Kinney County.

### DEEP WELLS OF THE PLATEAU SUMMIT.

Grass Valley pasture, Fort McKavett. Total depth, 255 feet. Limestone all the way except a few feet of sandstone at the bottom. Good water. (M. C. Ott.)[1] Water probably struck in the Edwards formation.

Rock Springs, Edwards County. In the Edwards formation. Two deep wells, 400 feet; pumped for use of town. Another well, 129 feet. (Ott.)

Hillcoat well, No. 2, Edwards County, 3 miles west of McKenzie trail, Nueces quadrangle, 300 feet deep. Through white limestone (Edwards) all the way. Soft drilling. (Ott.)

Griffin Creek, Nueces quadrangle, Edwards County. Surface gravel a few feet; soft white limestone (Edwards), 275 feet; sandstone (Edwards), with water. (Ott.)

Valverde County. B. L. Croucher's well, 30 miles north of Del Rio and 10 miles from Devils River. Depth, 475 feet. An unlimited supply of permanent non-artesian water 300 feet from the surface. Raised by an 18-foot windmill.

Valverde County. B. N. White's well, 20 miles north from Del Rio and 12 miles from Devils River. Depth, 300 feet; water struck 250 feet from surface.

R. W. Prosser's well, 20 miles from Comstock, Valverde County. Total depth, 569 feet; depth to water, 340 feet. Entire distance through solid limestone, probably all Edwards.

[1] We are indebted to Mr. M. C. Ott, of Brackett, for the records to which his name is appended.

VALLEY OF GRIFFINS CREEK.

Hillcoat's well, northern Kinney County. Total depth, 124 feet. Bowlders of hard limestone flints (Uvalde gravel), 88 feet; hard (Edwards) limestone, 36 feet. Fine water. (Ott.)

DEEP WELLS OF THE PLATEAU MONOCLINE.

North end of Grass Valley pasture, Kinney County. Total depth, 205 feet. Good hard limestone (Edwards) all the way. Few feet of sandstone at the bottom. Pumps 180 feet. (Ott.)

Hitchcock well, Grass Valley, Kinney County. Total depth, 204 feet. Gravel, 9 feet; limestone (Shoal Creek), 40 feet; slate (Del Rio beds); limestone (Edwards); hard sandstone (Edwards). Water rises to within 60 feet of surface. (Ott.)

Grass Valley pasture, south end. Two holes; no water; gas wells.

Grass Valley, north side of the creek. Total depth, 336 feet. Slate, 186 feet (Del Rio?); drilled 150 feet below slate into limestone; limestone all the way down to sandstone (Edwards); 10 feet of bottom water. Inexhaustible water. (Ott.)

Grass Valley, 1 mile west of Hitchcock ranch. Running stream in cave in mountain rock. (Ott.)

Weymüller well, Kinney County. Surface gravel (thickness not given); mountain limestone (Edwards), 180 feet; whitish sand (Edwards), 20 feet; water in last; non-artesian. Windmill pump. (Ott.)

SURFACE WELLS OF THE CANYON VALLEYS.

These can usually be obtained in those portions of the canyon valleys of the region where the stream way has cut below the level of the Edwards beds. Water is found in the outcrops of the Glen Rose and lower beds and in the alluvial deposits of the Uvalde, Onion Creek, and fry-pan type.

The basement sandy beds of the Cretaceous yield a constant supply of wholesome well water. The Glen Rose beds and alluvial deposits are somewhat variable.

WELLS NEAR BEE CAVES, TRAVIS COUNTY, SUNK IN THE GLEN ROSE FORMATION.

Information concerning the following wells was furnished by T. C. Bohls, Bee Caves, Texas:

At Bee Caves. Depth of well, 160 feet. Water supply constant. Constant depth of water in well, 11 feet. Water found in black sand.

Three-fourths of a mile west of Bee Caves. Depth, 165 feet. Constant depth of water in well, 25 feet.

One and one-half miles west of Bee Caves. Depth, 163 feet. Constant depth of water in well, 40 feet.

On T. C. Bohls's place, 2½ miles east of Bee Caves, there are two wells, as follows: One 221 feet deep; 140 feet of water sometimes; supply not constant; well goes dry. Another 264 feet deep; 80 feet of water; can dip dry.

There is a well at the junction of the two forks of Barton Creek, bored 100 feet above the creek bed; well 54 feet deep. There is 10 feet of water in well, and water can not be lowered.

WELLS OF SOUTHERN PACIFIC COMPANY IN PECOS COUNTY

The following wells drilled by the Southern Pacific Company in Pecos County are west of the Pecos, and therefore beyond the field of this paper, but are allied to the deep wells of the Edwards Plateau system:

Longfellow well; 683 feet deep; unlimited supply of water; capacity, 1,250 gallons per hour.

Sanderson well; 987 feet deep; large supply of good water; capacity, 1,700 gallons per hour.

Dryden well; 1,797 feet deep; capacity, 1,000 gallons per hour; unlimited supply of good water. Well begins at Shoal Creek limestone.

Lozier well; 770 feet deep; limited supply of water; capacity, 1,000 gallons per hour, which exhausts the well in two hours. It then takes two hours to fill up. Well begins in Del Rio beds.

Thurston well; said to have been drilled 1,800 feet through solid limestone.

None of these wells are flowing, nor have their geologic relations been studied.

## GRAVITY SPRINGS OF THE CANYONS OF THE PLATEAU.

In the investigation of the springs of the region to which this paper pertains we have found it important to distinguish two classes. In one case water-bearing strata lying nearly level outcrop on slopes, gentle or steep, and the water is drained away at the horizontal outcrop; the underground water is merely a continuation of the descending surface water, which may have been diffused through sandstone or concentrated in tunnel-like channels dissolved from limestone. In the second case the water-bearing strata do not locally outcrop at the surface, but are deeply buried, and communication with the surface exists through natural fissures. The water in the strata, being under hydrostatic pressure, is forced up through the fissure conduit so as to flow out at the surface of the ground. The two classes are here distinguished as *gravity springs* and *fissure springs*.

In the description of the rivers of the plateau we have mentioned the springs which were met in ascending the canyons of the streams. These springs break out near the water line in the rivers, the water draining by gravity from the horizontal strata of portions of the Edwards, Glen Rose, and Travis Peak or Gillespie formations.

Such springs usually occur in all the streams of the plateau from the Llano to the Pecos. Wherever the gradient of the stream, in descending the summit of the plateau through the canyons to the lower Rio Grande Plain, cuts into a water-bearing bed the water drains out into the stream way. In many cases this forms large, deep pools of clear running water. Of this character are the so-called headwater holes of the various forks of the Llano, Pedernales, Guadalupe, Comal, Medina, Frio, Nueces, and Devils rivers. The constant waters of all these streams are derived from springs of this character. As the rivers have never been measured no idea of the exact quantity of water thus escaping has been obtained, and we can present only general descriptions of it.

About 17 miles north of Leakey is the great headwater hole of the

Frio (Pl. XXXVIII), which occurs at the very head of the flat-bottomed canyon, just below where the steep ascent to the summit region begins. These springs break out from cavernous, slightly arenaceous layers in the Edwards limestone, about 300 feet below its summit. Between this point and Leakey there are many large, deep, and long pools of water in the stream way.

From Leakey the east prong of the Frio was ascended in a northeast direction. Fine springs burst out from the Glen Rose and Edwards beds, and supply the river with much water. About 12 miles northeast of Leakey there are some springs the tufaceous deposits of which make a number of hemispherical concentric pools, arranged in descending series, resembling very much the illustrations of Gardner's springs in the Yellowstone Park. Below Leakey and thence to the mouth of the canyon the Glen Rose beds afford a large number of gravity springs. There are four large ones at Van Pelt's ranch, and at Rio Frio postoffice many acres of land are irrigated with spring water.

The Guadalupe and its several tributaries derive the abundant water which they contain above Kerrville from various strata in the Glen Rose and Edwards beds. At Kerrville the springs can be seen draining out of the Glen Rose beds. The highest springs are about 7 miles above Vix and at Spencers Hole. At the latter place the water breaks out of a horizon high up in the Edwards limestone.

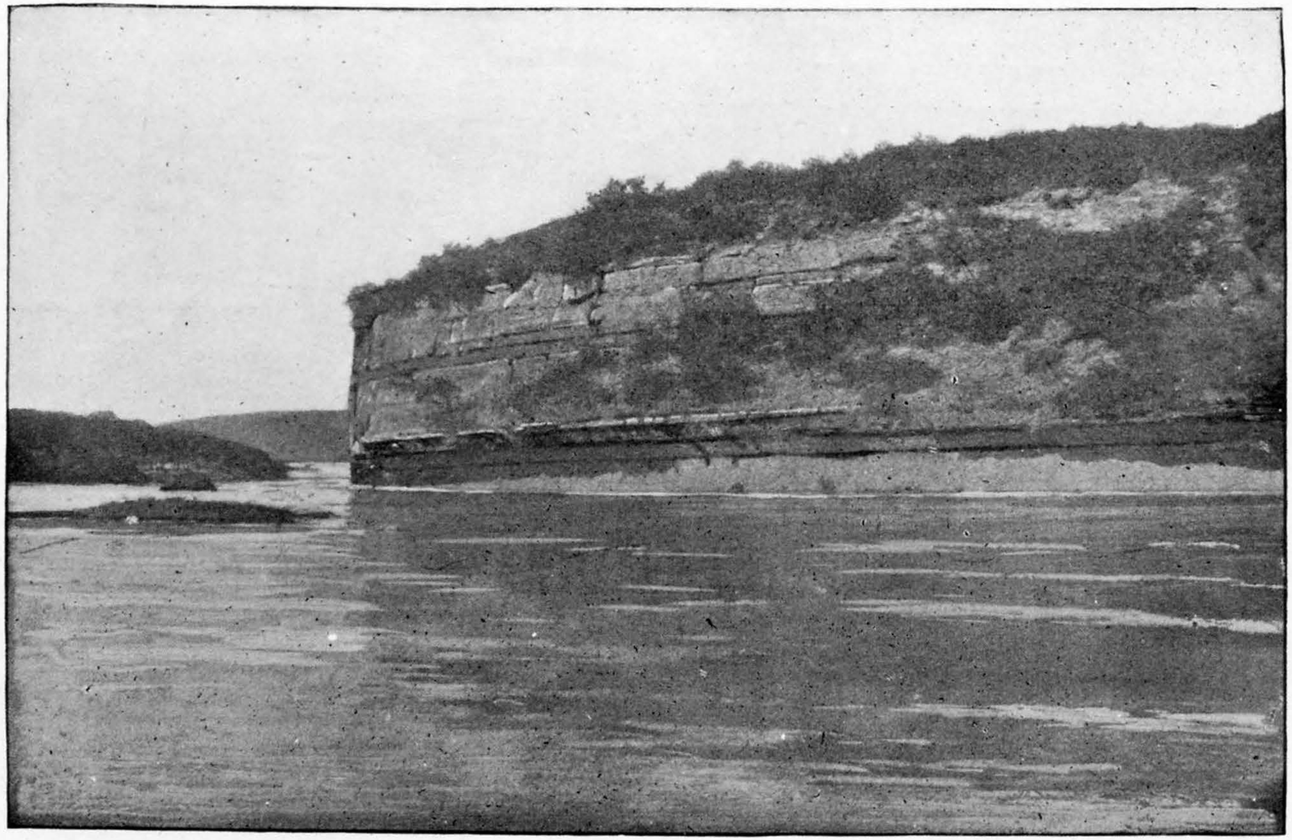
The various forks of the Llano, the East Fork of the Nueces, and its principal tributary, Hackberry Creek, reveal the same wealth of gravity spring water in the upper portion of their flat-bottomed canyons, the water breaking out from various horizons of the Edwards and Glen Rose beds.

The waters of the Pedernales are largely derived from the basement beds of the Cretaceous. Many fine springs break out in the course of the stream and its tributaries, as seen around and below Fredericksburg.

In addition to the gravity springs here enumerated there are, no doubt, many others, both in the plateau and in the plain, which the writers have not had opportunity to see. Among these may be the springs of Howards Creek, the Pecos, and Devils River. Concerning the latter, we have taken the following notes from a letter written to this office by Dr. John T. Nagle:

I desire to state that there are two other large springs in Valverde County, Texas, besides the San Felipe at Del Rio. The water is clear, has a very pleasant taste, appears to be pure, and runs over a rocky bed into the Devils River, which empties into the Rio Grande. These springs might be of interest to your Bureau, and I will give their names and location as near as I can.

The first spring is at R. W. Prosser's ranch, which is about 6 miles north and slightly to the west of Camp Hudson, on the Devils River, between Camp Hudson and Pecan spring. The spring is not named, but is known as Prosser's, on account of its being on his land. The second spring is called Pecan spring; it is another sparkling spring of pure limpid water, and its taste is delicious. There has been



RUN-OFF OF FRIO (GRAVITY) SPRINGS, NEAR THE HEAD OF WEST FORK OF FRIO RIVER.

no chemical test of these waters, but I am inclined to think that there may be a very slight amount of lime in the water. Pecan spring is about 2 miles north and one-half mile west of Prosser's ranch house, close to the Devils River, as is also Prosser's spring.

The following details of gravity springs along the forks of the Nueces will illustrate the general character of the plateau springs.

### KICKAPOO WATER HOLE.

At Kickapoo springs the West Nueces and Kickapoo Creek are cut down to the level of the Comanche Peak limestone. Here enormous springs break forth, creating a wide, running stream of clear water that continues 4 miles. In places it is 100 yards in width, is bordered by exquisite forests, and teems with aqueous vegetation and game fish. Toward its lower end the streamway narrows; the bold flow which up to this point has rushed through the rocky banks suddenly ceases and is succeeded by a whitened pebble bed. We were informed that the gravel drank in this water; but, suspecting that its disappearance was too rapid for such imbibition, cleared away a thin layer of gravel and discovered that it escaped down a large fissure into the underlying limestone rocks (Pl. XLVIII, p. 314). These springs drain out of the strata near the contact of the Edwards limestone and the Comanche Peak bed. Observations tend to show that this geologic horizon is elsewhere completely saturated with water. Springs from this horizon are found along the East Fork of the Nueces and Hackberry Creek. Nearly all the abundant living water in the East and West forks of the Nueces, except the Black water hole and the springs at the immediate head of Hackberry Creek, is derived from this water-bearing horizon.

Where the rocks of the plateau are horizontal, north of where they bend down to the southward, this water-bearing stratum occurs at a level of about 1,750 feet above the sea and about 600 feet below the summit of the plateau, and (except in the lower valley of the East Nueces, which has cut below its level) is available for wells. Along the southern monocline it is reached in the well at Hillcoat's ranch, 124 feet below the surface, at an altitude of 1,447 feet above the sea, and at two wells in the arena of Griffin Creek, west of Hillcoat's ranch, at altitudes of 1,400 and 1,450 feet, respectively.

### BLACK WATER HOLE.

This occurs about 150 feet above the base of the Edwards limestone, and occupies a position at Black water hole 1,900 feet above sea level. The water here is not so abundant as at the Kickapoo springs.

Justice spring, Cedar spring, and Cherry spring, on the western border of the Nueces quadrangle, probably derive their waters from a third and still higher water-bearing horizon of the Edwards beds. The waters in the vicinity of Seep Springs Mountains may also be derived from this source. These springs are feeble, and the horizon has not

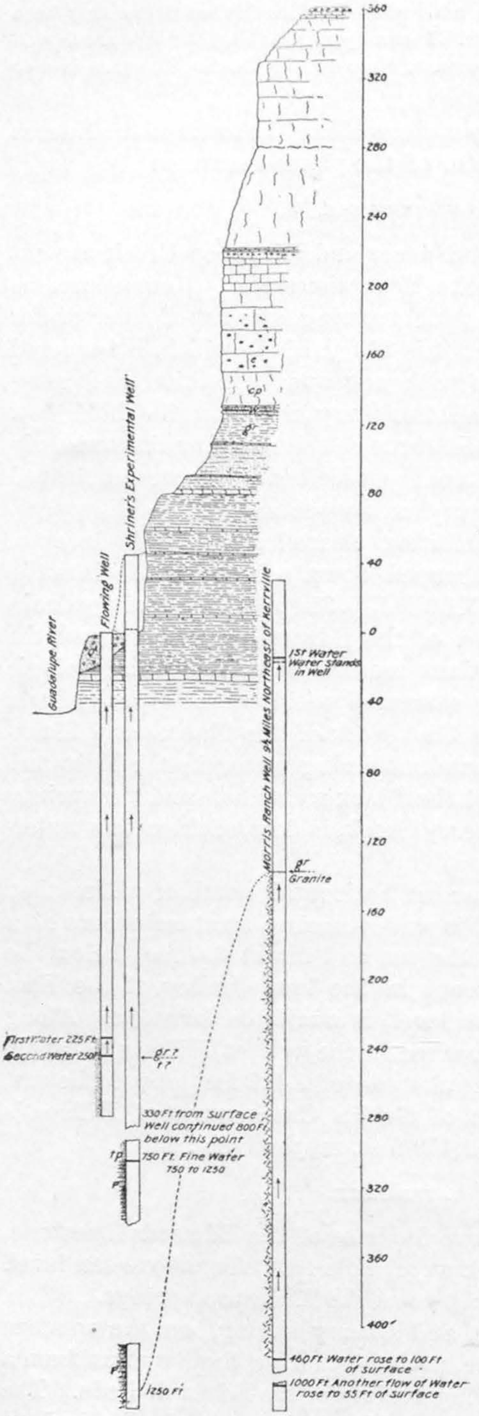



FIG. 68.—Artesian wells at Kerrville and Morris ranch. *e*, Edwards; *cp*, Comanche Peak; *w*, Walnut; *gr*, Glen Rose; *tp*, Travis Peak; *p*, Paleozoic; *Q*, Pleistocene.

been sufficiently studied to justify at present any conclusions tending to show persistent horizontal distribution or availability. There are strong reasons for believing, however, that in the region lying to the east of the Nueces quadrangle this water-bearing stratum becomes more productive, especially in the heads of the Guadalupe, Llano, Medina, and Frio. In the plateau region this water lies at an altitude of from 2,000 to 2,100 feet above sea level, or from 300 to 350 feet above the base of the Edwards limestone. In common with the other water-bearing strata, its plane descends with the dip along the monocline at the south.

Occasional seepage springs also break out in these canyons at the contact of the gravel deposits with the underlying Cretaceous beds, but these are feeble and intermittent in yield, being sympathetic in their flow with the rainfall.

## ARTESIAN WELLS OF THE CANYONS OF THE PLATEAU.

Since the drainage ways of the plateau south of the Pedernales do not cut entirely through the Glen Rose formation, it is evident that a large portion of the water-bearing rocks of the Cretaceous system, including those which elsewhere yield the largest supplies, must be embedded below the lowest cutting of the stream valleys. At various points in the canyons artesian flows have been obtained from the Glen Rose and Travis Peak sands, as at Kerrville and Center Point, in the Guadalupe Canyon, and 7 miles southeast of Utopia, in the canyon of the Seco. A description of the wells in the valley of the Guadalupe at and below Kerrville will serve to illustrate the conditions controlling their occurrence.

Mr. Charles Shriner has made several interesting drillings at Kerrville. He finally obtained flowing water in the river valley at a place 50 feet above the water in the Guadalupe. The well is a 12-inch hole. The first flow of water was struck at 225 feet and the main flow at 250 feet. The water rises about 8 feet above the surface. This well was commenced in the Glen Rose formation, about 150 feet below its contact with the Edwards limestone, which can be seen in ascending the adjacent divides; hence the water-bearing sands struck at 225 and at 250 feet occur in strata about 400 feet below the Edwards, which position corresponds very nearly with the base of the Glen Rose formation (See fig. 68).

Mr. Shriner's first experiment was a well drilled on a hill in the northern suburb of the town, about 50 feet higher than the surface of the present well. At about 750 feet fine water was struck, which rose to within 75 feet of the surface, or not quite to the level of the river. This water must have been very near the bottom of the Travis Peak formation, for below it were salt, clay, and black sand, probably from the underlying Paleozoic rocks. At 1,250 feet it is stated that granite was struck and penetrated to a depth of 75 feet.

A flowing well has been reported at Center Point, Kerr County, in the valley of a small stream flowing into the Guadalupe, but we have had no opportunity to verify this information.

A few other wells of a similar character have been bored in the Guadalupe Valley below Kerrville. One of these is reported 1½ miles below the town; another is lower down. About fifteen have been reported from within a radius of 10 miles of Welfare post-office, Kendall County. These have a depth of from 170 to 175 feet and flow about 5 gallons each per minute.

A single canyon well of this type has been reported in the valley of the Blanco, at Wimberley, Hays County. This is said to be 250 feet deep. No information has been obtained as to the amount of flow.

The only other flowing wells reported to have actually been obtained in such valleys are two shallow wells in the valley of the Seco, 7 miles southeast of Utopia.

Similar wells should probably be found in the valleys of the shallower streams, such as the Pedernales, Cibolo, etc., but we have no data for making definite predictions concerning them.

The well drilled several years ago at Bulverde, on the Cibolo, north of San Antonio, furnished the only record we have been able to procure of this type of canyon wells. This, together with some experiments near Aue, indicates that the water will rise but not flow. The record of the Bulverde well as reported by Roessler (except the names of the formations, which we have supplied) is as follows:

*Record of R. Mecke's well at Bulverde, Bexar County, on the Cibolo, due north of San Antonio.*

[Altitude, 1,600 feet; depth, 361 feet.]

| | Thickness in feet. | Depth in feet to bottom of strata. | |
|---|---|---|---|
| Glen Rose formation (?): | | | |
| Soft yellow stone | 42 | 42 | |
| Blue stone | 20 | 62 | |
| Yellow stone | 18 | 80 | |
| Blue stone | 2 | 82 | |
| Yellow stone | 6 | 88 | |
| Blue limestone | 19 | 107 | |
| White limestone (water bearing) | 30 | 137 | 137 |
| Travis Peak formation (?): | | | |
| Bluish sandrock | 10 | 147 | |
| Blue slate | 40 | 187 | |
| Red clay | 13 | 200 | |
| Limestone | 60 | 260 | |
| Sandstone | 31 | 291 | |
| Quartz sand | 1 | 292 | |
| Yellow clay | 23 | 315 | |
| Limestone | 35 | 350 | |
| Sandstone | 11 | 361 | 224 |
| | 361 | | 361 |

Strong supply of water rises 250 feet, or to within 135 feet of surface.

*Record of G. A. Dueller's well, on the Leon, near Aue.*

[Drilled in the Glen Rose and Travis Peak formations.]

| | Thickness in feet. | Depth in feet to bottom of strata. |
|---|---|---|
| Very hard rock, sandy limestone | 2–3 | 2 |
| Alternations of rock, brown sandstone, inky blue sandstone, etc. | 620 | 622 |
| Water | .... | 622 |
| Same rock as foregoing | 51 | 673 |

Water rises to within 20 feet of surface, and now stands within 4 feet of surface. Well stopped in yellow sandstone.

Except at the localities given, we know of no artesian borings of sufficient depth to reach the Travis Peak sands having been made in the plateau canyons west of the Guadalupe, but there is reason to believe that in these sands, lying below all the strata exposed in the deepest cuttings of the canyons, there exists an abundant supply of water which can, in places at least, be made available. On the Nueces there are large springs at Camp Wood, rising through joints and fissures from rocks lying lower than the Glen Rose beds exposed at the surface, and there is every reason to believe that the basal sands are the source of this supply. This water, if it exists, will be found at least 500 feet below the base of the Edwards limestone.

MORRIS RANCH WELL.

An instructive boring is reported to have been made at Morris ranch, Gillespie County. This ranch is situated on one of the headwater tributaries of the Pedernales, about 11 miles southwest of Fredericksburg and 9½ miles northeast of Kerrville. We are indebted to Mr. Charles F. Morris, of the ranch, for the following information concerning this well: Total depth drilled, 1,100 feet; first water struck at 50 feet; granite struck at 180 feet. At 460 feet more water was reached, and this rose to within 100 feet of the surface. The well was continued and reached another flow of water at 1,000 feet. The water then rose to within 55 feet of the surface. The drilling was continued in solid granite until the depth of 1,100 feet was reached. According to Mr. Morris, "the well was in granite from 180 feet to its bottom. It was all very hard, but would change color at times from red to gray and then to red again. We could only go from 3 to 4 feet in 24 hours."

If the log of this well as given is correct, it shows a remarkable and unusual occurrence of underground water. The specimens sent to this office by Mr. Morris are undoubtedly granitic, but he says that he is uncertain as to the exact depth from which they came. Moreover, the record of this well, if correct, is further confirmation of the belief that the Cretaceous formations are laid down upon an unequal floor of Paleozoic rocks, for the top of the granite in the Morris ranch well, struck at an altitude of about 1,500 feet above sea level, is 1,000 feet higher than the same as met with in Mr. Shriner's first experimental well at Kerrville. This close proximity of the granite to the surface in Gillespie County no doubt also accounts for the intense redness and arenaceous character of the Gillespie beds in the valley of the Pedernales adjacent to Fredericksburg.

The following statement lately appeared in the New York Sun:

> Nordenskjold, the Swedish scientist, has shown that water can be found by boring into granite and other crystalline rocks to a depth of from 100 to 170 feet. Briefly, he proceeded on the theory that the variations in temperature ought to cause shearing strains between the upper and lower layers of the rock, in such way causing horizontal crevices into which water from the surface would percolate, and the water would also be fresh. A well was said to have been sunk in the islet of Arko, off the Swedish coast, in 1894, and at the depth of 110 feet fresh water was found, supplying 4,400 gallons a day, and since then six other wells have been bored and water found at about the same descent, the object of the research being to provide lighthouses and pilot stations with a permanent and plentiful water source.

The occurrence of water in granite depends upon the existence of fissures in the granite. The presence of water in mines sunk along veins where fissures exist is an illustration of this principle. But granite can not be regarded as offering conditions favorable for the procurement of underground water, even though in some instances wells sunk in it may have been successful.

## WATER OF RIO GRANDE PLAIN.

### NONFLOWING WELLS AND GRAVITY SPRINGS OF RIO GRANDE PLAIN.

The nonflowing wells and gravity springs of the Rio Grande Plain may be classified in two groups: (1) Those derived from alluvial (gravel) deposits; (2) those derived from marine Cretaceous deposits.

In the area between the Colorado River and Onion Creek, Travis County, there are many shallow wells which derive their supply of water from the Uvalde upland gravel formation. Similar wells are obtained in the upland gravel beds wherever the latter cover considerable areas in Travis, Hays, Comal, Bexar, Medina, Uvalde, and Kinney counties. In San Antonio they are 40 to 50 feet deep; at Spofford Junction, about 40 feet. The water usually accumulates at the base of the gravel beds, at their contact with an underlying impervious bed rock.

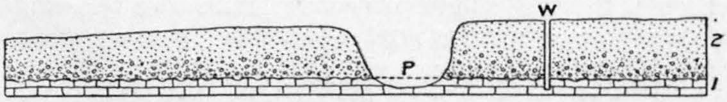



FIG. 69.—Wells in gravel beds. 1, underlying rock floor; 2, Leona formation terrace, silt at surface, grading downward into gravel; P, pool containing water where the stream has cut down into the rock floor; W, well sunk through the silt and gravel into the rock floor.

Water is also found in the gravel of the present stream ways of the interrupted and intermittent drainage courses. In Uvalde and Kinney counties, for instance, the beds of such streams, although dry at the surface, usually contain much water below, which can be procured by digging a few feet into the gravel bed. This condition is especially likely to prevail where the channel is filled with gravel. The character of the stream bed is due largely to the fact that the water flows down into the gravel and disappears from view. In those portions of the water way where the gravel has been removed, exposing the formations underlying it, the water accumulates in large ponds or pools that usually last during the dry season. Such pools may be seen at many places along almost any stream. There is one on the north side of the crossing of the Uvalde-San Antonio road over the Frio River. At this place an exposure of basalt, the gravel having been washed off, forms the bottom of a small reservoir.

Just south of the crossing of the Southern Pacific Railroad over the Leona River the gravel is washed off the Shoal Creek limestone and the Del Rio clay. The latter forms the bottom of a large, deep pond, into which the water from upstream accumulates. Below such pools, when the stream way is filled with gravel, the water again disappears. Besides the pools, along the streams there are frequently springs that flow from underneath the gravel when it is underlain by an impervious stratum of any kind. A good example of this kind of spring is the Soldiers' Camp spring on the Nueces River, about 1 mile below the crossing of the Uvalde-Nunn's ranch road. The gravel at this place is underlain by the Austin chalk. The water rushes out in a stream of considerable size, and is cool, clear, and pure. Such springs are surrounded by splendid growths of pecan trees. Below the Soldiers' Camp spring the bed of the Nueces River is usually formed by the Austin, Anacacho, or argillaceous Eagle Pass beds, none of which absorb water very rapidly. The result is a flowing stream.

### WATERS OF THE LEONA AND KINDRED FORMATIONS.

The terraces above the present flood plains are structurally comparable to the present stream beds, consisting of gravel accumulations deposited in stream valleys upon a bed-rock floor.

Water is usually stored in the bottom of these sheets, above their contact with the underlying formations, and wells sunk in them usually

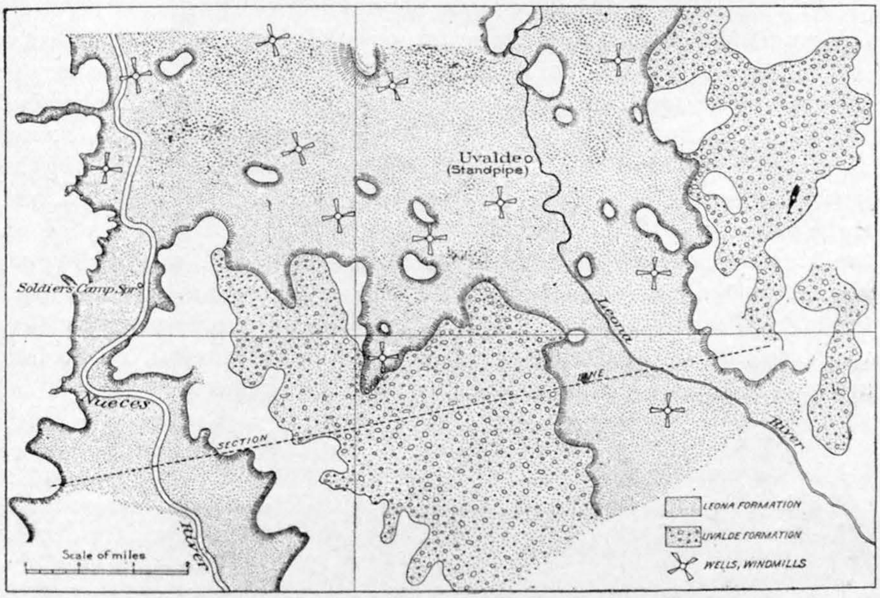



FIG. 70.—Alluvial deposits around Uvalde, showing wells.

penetrate to the underflow level of the present stream beds before obtaining water. (See fig. 69.)

At Uvalde (and perhaps other places) these gravel sheets are many square miles in extent, and the contained water is of great economic

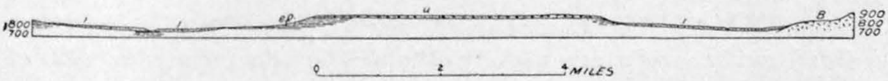



FIG. 71.—Section across valleys of Nueces and Leona rivers and intervening highland, on line shown in fig. 70. *l*, Leona; *ep*, Eagle Pass; *u*, Uvalde; *b*, basalt.

importance (see fig. 70). One well at the waterworks in the town of Uvalde is 50 feet deep, and pumps 40,000 gallons per day.

The Leona River below Uvalde cuts down to the contact of the Leona gravel sheet and the underlying floor of Cretaceous rocks, where a large volume of gravity spring water (the springs of the Leona) breaks forth with a discharge of 11 second-feet, or 7,000,000 gallons per day. The run-off forms large pools below Uvalde, and the water resembles very much that of the typical fault springs, with which it has been confused. This water is drained from the bottom of the Leona gravel sheet and is apparently the same as that obtained in the surface wells sunk in the formation.

Six miles below Uvalde, near the Leona River, there is a well at a Mexican jacal 30 feet deep. Basalt bowlders and rotten basalt were found in the well. The water seems to come from the Leona terrace, and is very good.

The waters of the alluvial deposits are usually abundant and pure.

### WATER OF THE RIO GRANDE TERRACES BELOW DEL RIO.

Very little is known concerning the possibilities of wells sunk in the Rio Grande terraces. Below Del Rio to Eagle Pass the country is almost uninhabited, excepting along such creeks as Pinto, Tulio (Las Moras), and a few others, in which there is running water. In a reconnaissance made down the Rio Grande during the summer of 1895 by the junior author and Mr. T. W. Stanton, only one dug well was seen, and there were no settlements except those above alluded to and a few on the Mexican side of the river. The well mentioned was at Upson, about 16 miles above Eagle Pass. It is located near a stagnant pool, from which the water seeps into it. It seems reasonable to expect that wells sunk in the Rio Grande terraces would be successful, but no data are available upon which to base any definite conclusion.

### WELLS IN THE FRY-PAN DEPOSITS.

Wells are easily obtained in the gravel deposits of the fry pan gorge of Turkey Creek, in western Uvalde County. These have been sunk at Cline Station and at several points on Moore and Allen's ranch to depths of from 20 to 40 feet, and water obtained just above the contact of the gravel beds with the Austin chalk. The Southern Pacific well at Cline Station is 12 by 12 feet and 40 feet deep, and affords an inexhaustible supply of water, being one of the best wells on that line of railroad.

Where the stream bed of Turkey Creek cuts through the gravel down to the Austin chalk, near Cline post-office, many springs break forth. These are large in times of rainfall, but fail in dry weather.

### WATER OF THE TERRACE FORMATIONS OF THE COLORADO RIVER.

Good surface wells are usually obtained in the terrace deposits of the Colorado River, where they are of extensive area and not too much cut by later erosion. Such wells, however, are apt to vary appreciably with the rainfall, more in the coarser deposits than in those of finer texture.

Small seepage springs can be seen breaking out in many places at the contact of the gravels with the lower formations. This is noticeable after a rainfall in the case of the high Asylum terrace in north

Austin. These springs are also quite noticeable along the banks of the Colorado wherever the contact of the terrace and Cretaceous formations is visible.

WATER OF THE ONION CREEK MARL AND SIMILAR DEPOSITS.

Water is obtained from wells in the Onion Creek marl in the vicinity of Delvalle and between Pilot Knob and Buda at depths of from 25 to 50 feet. Shallow wells are also successful in the similar deposits along the Guadalupe River and other streams of that class.

The positions of the wells that have been sunk in the Leona and similar deposits in the vicinity of Uvalde are indicated on the map (fig. 70, p. 275). The depths of the wells are usually from 40 to 60 feet.

Water is obtained along some of the streams by sinking wells near the stream bed through the alluvial deposits. The water from the stream filters through the loose alluvium into the wells. The wells at Eagle Pass, according to reports, belong to this type. The old city waterworks at Austin derived their supply from wells of this class.

*Records of wells in the alluvial deposits in Kinney, Edwards, and Uvalde counties.*[1]

Mouth of Griffin Gap, head of Pinto Creek. Total depth, 256 feet. This well was through cobblestones and red clay from top to bottom; struck a running stream.

Nine miles south of east of Brackett, on Turkey Creek. Gravel of white limestone pebbles, 40 feet; basalt, 60 feet; water.

Moore and Allen's well, No. 3, at Cline. Total depth, 47 feet. All gravel.

Cline. Several wells at Cline Station, Cline post-office, and in the vicinity of these places; all dug in the gravel of Turkey Creek Flat. Depth, 20 to 40 feet.

Spofford. Several wells in the Uvalde gravel. These wells average about 40 feet in depth. The water is alkaline.

Hall's pasture. Well 66 feet deep. Limestone bowlders. Water in gravel.

Pinto Flat, north of above, 9 or 10 miles north of Brackett. Yellow clay, red clay, and gravel. Plenty of water.

SURFACE WELLS AND SPRINGS OF THE CRETACEOUS FORMATIONS.

The outcrops of the various Cretaceous formations of the Rio Grande Plain vary slightly in their water productivity, but in general the yield of all of them is small. This is especially true of those marly beds composed of the so called "joint clays," such as the Webberville, Taylor, Eagle Ford, and Del Rio, in which wells are seldom successful.

The limestone formations are more variable relative to one another, but are in general slightly more productive. Even in these success is not the rule. We have seen a few successful wells in the Austin chalk and the Shoal Creek, Washita, and Edwards limestone, but failure is more frequent.

We have not many records of surface wells in the Rio Grande Plain, and the opinions above expressed are largely the result of casual observations. The following records, however, furnished us by Mr. M. C.

[1] Furnished by Mr. M. C. Ott, of Brackett.

Ott, of Brackett, give valuable data concerning the western portion of the region:

IN THE EDWARDS LIMESTONE.

No. 2, east of Las Moras Mountain, head of Guaxillo Creek. Total depth, 255 feet. Starts in Shoal Creek. Limestone to 15 feet of bottom; then 2 feet of slate; then 12 or 13 feet of sandstone. Plenty of water, probably from Edwards limestone.

IN STRATA BETWEEN THE EDWARDS LIMESTONE AND AUSTIN CHALK.

Brackett. Holmes well in Las Moras Flat, Spring street. Depth, 104 feet. Earth and gravel, 9 feet; white limestone, 30 to 40 feet (Eagle Ford); hard black rock, probably basalt, 76 feet. Could not get to bottom of this. Deep water at 18 feet; very bad.

Brackett. C. A. Windus's well. Total depth, 150 feet. Magnesian limestone, 60 feet (Eagle Ford). Bad water (gypsiferous). Water found at 37 feet (in Eagle Ford?). Hard limestone (Shoal Creek) at 60 to 100 feet. Black, magnetic limestone (possibly basalt). Water.

Moore and Allen's well, No. 4. Dug for 50 feet; drilled 100 feet. Water in sandstone at 150 feet. This well is probably on the Nueces, north of Southern Pacific Railroad, east of Cline.

Moore and Allen's ranch. Well No. 1, near head of Turkey Creek. Soil, 4 to 6 feet; "slate," 143 feet (probably Del Rio). Sulphur water at 149 feet. This well is probably in Del Rio.

Moore and Allen's ranch. Windmill well, 2 miles north of Cline. Soil with gravel, 20 feet; "magnesian" limestone, 20 feet (Eagle Ford); at 150 feet, Shoal Creek limestone, base of Eagle Ford; slate (Del Rio) at 190 feet. Water in black sandstone at 255 feet (possibly near top of Edwards and base of Washita division). Sulphur water below rock.

New York and Texas Land Company's well, 7 miles from Brackett, on Turkey Creek. Total depth, 404 feet. Earth and yellow clay, 17 feet; blue clay, 3 feet (Eagle Ford); lignite or asphalt, 11 feet (Eagle Ford); gray limestone, 12 feet (Eagle Ford); slate, 367 feet (Eagle Ford, Shoal Creek, and Del Rio). First water at 37 feet. Water, 404 feet in sandstone (probably same beds as foregoing).

Near Mariposa ranch, James Dignowity, northwest of Brackett. Depth, 155 feet. Little clay, thin limestone. This well is in a canyon 100 feet deep, and the narrator was uncertain as to horizons.

New York and Texas Land Company's well No. 1, at abandoned windmill 4 miles north of Brackett. Total depth, 269 feet. Hard gray crystalline limestone, 85 feet (Shoal Creek); slate, 184 feet (Del Rio); sandstone, 10 feet. Water rises 131 feet. This water is found in Del Rio.

Four miles east of Brackett, Turkey Creek road. Depth, 105 feet. Gravel on top. Dark blue sandstone struck within 10 feet of water. White sulphur water.

IN THE AUSTIN CHALK.

Fry-pan well. Fry-pan Valley, Brackett quadrangle. Soil, 10 feet; lime rock, 10 feet (Anacacho); blue limestone, 380 feet (Austin chalk). Most of this well is in Austin chalk.

Palmer's ranch, 8 miles west of Cline, 2 miles southwest of Waldo, north edge of Anacacho Mountains. Soil, 10 feet; white limestone, 10 feet (Austin chalk); blue rock, 280 feet (Austin chalk). No water was obtained.

Moore and Allen's ranch, north edge of Uvalde County, 8 miles west of Cline, 2 miles southwest of Waldo Station. Soil, 10 feet; white limestone, 10 feet (Austin chalk); blue rock, 280 feet (Austin chalk). Well quit in last. No water, but a seep.

On the southwest side of Pilot Knob, Travis County, along the branch that flows north into Onion Creek, there is a well sunk into the Austin chalk, 28 feet deep, that yields a constant supply of clear, pure water.

ARTESIAN WELL, CITY WATERWORKS COMPANY, SAN ANTONIO, TEXAS.

IN ANACACHO AND EAGLE PASS FORMATIONS.

James's ranch well. Depth, 170 feet, through top of Anacacho limestone. Water salty and sulphurous.

Furness ranch, southwest corner of Anacacho Mountains. Sandy, muddy rock to sulphur water, 220 feet (Anacacho or Eagle Pass). Last water in white sandstone rises to 11 feet of the surface.

Moore and Allen's well No. 5, 12 miles east of Cline, in Uvalde County, near old California road, south of Nueces River, west of Big Mountain. Total depth, 250 feet. Green marl all the way (Eagle Pass beds). Water at 70 feet.

Another well 4 miles north of above. No coal or hard limestone. Black, clayey material.

Creek 3 miles east of Nueces River, on the road from Uvalde to Carrizo. Black sulphur water. Soft rock, species of slate or shale (Eagle Pass). Struck water first at 170 feet. Struck coal in three places; one vein 5 feet at 150 feet depth.

There is a well on the Frio River, below the Southern Pacific Railroad, near Connor's ranch, sunk into decomposed basalt or phonolite. The water is sulphurous and smells very bad.

From the foregoing data the following generalizations can be formulated concerning the nonartesian wells of the Cretaceous formations of the western portion of the Rio Grande Plain: Water is obtained in the Edwards limestone on the downthrown side of the fault. Water can be obtained in the Del Rio clays, but we do not know its quality. That in the Eagle Ford shales is very bad. The wells in the Austin chalk are often entire failures. The water obtained in the Anacacho formation may be salty or contaminated by sulphur. Good water may be obtained from the Eagle Pass formation, but is sometimes salty. The water obtained in wells that are sunk into the decomposed igneous rocks is likely to be very bad, but that obtained nearby where such rock material is not penetrated does not seem to be especially contaminated.

ARTESIAN WELLS.

The Rio Grande Plain, so far as the inclination of the strata is concerned, with certain important local exceptions to be mentioned, presents conditions favorable for obtaining artesian water. The strata dip with the surface slope from the interior margin of the plain toward the coast at a very gentle angle. The plain as a whole is underlain by several artesian systems of different geologic ages. Two of these systems underlie the portion of the plain discussed in this paper. These are the water-bearing sheets of the Edwards beds and the Travis Peak sands. Water also occurs in some of the Glen Rose beds.

The nature and availability of these reservoirs can best be understood by an examination of the data at hand concerning the various artesian wells, especially those at Austin, Manor, San Marcos, and San Antonio.

AUSTIN WELLS.

At least seven artesian wells have been sunk in the vicinity of Austin—one at the State Institution for Colored Dependents, altitude 650 feet; the Groom well, altitude about 615 feet, and one at the Lunatic

Asylum, altitude 635 feet, both in the northern suburbs of the city, beyond the university; two on the capitol grounds, altitude 520 feet; one on Fifth street east, altitude 500; and one at St. Edward's College, altitude 660 feet, 2 miles south of the river. They all pass through similar formations, as far as they go, but the Asylum and St. Edward's wells are the only ones which have been drilled to the basal Trinity beds.

The Groom and Capitol wells at the present time have only a small discharge, 5,000 or 6,000 gallons a day. The Asylum well had a discharge of 150,000 gallons a day and threw the water to a height of 40 feet. In St. Edward's well, which is 25 feet higher in elevation than the latter and obtains its supply from the same source, the water comes within about 5 feet of the surface and has to be pumped.

The Asylum well is 1,975 feet deep, and the formations, in order downward from the surface, are as follows:

*Log of well at Lunatic Asylum (altitude 635 feet), Austin, as given by Mr. McGillvray, the driller.*[1]

| | Thickness in feet. | Depth in feet to bottom of strata. |
|---|---|---|
| (*a*) Dark shale | 80 | 80 |
| (*b*) Very hard limestone | 25 | 105 |
| (*c*) Blue marl | 90 | 195 |
| (*d*) Limestone and alternations of limestone, marl, and sand | 1, 105 | 1, 300 |
| (*e*) Water-bearing sand[2] | 15 | 1, 315 |
| (*f*) Limestone | 60 | 1, 375 |
| (*g*) Rotten shale | 50 | 1, 425 |
| (*h*) Limestone | 50–60 | 1, 485 |
| (*i*) Sand, water-bearing | 315 | 1, 800 |
| (*j*) Blue shale or marl; no limestone | | 1, 975 |

Mr. McGillvray states that stratum "*b*" is very hard limestone and is easily recognized whenever encountered; stratum "*d*," of 1,105 feet of limestone, contains occasional streaks ot shaly water-bearing sand. In the Capitol well, at 400 feet, a small flow having a disagreeable odor was encountered. The beds marked "*i*," or the 315 feet of Trinity (Travis Peak) sands, consist of alternate bands of from 25 to 35 feet of thin sand and from 5 to 6 feet of brown and reddish shale.

The strata enumerated may be converted into geologic terminology as follows:[3]

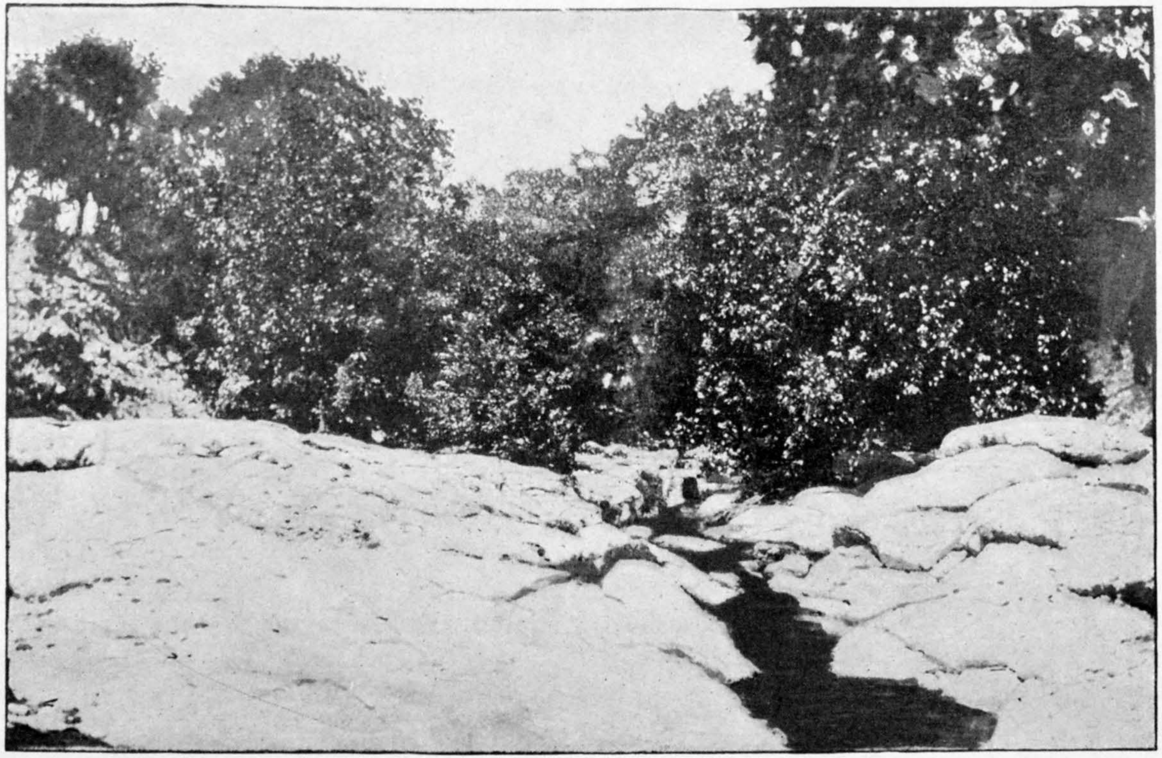
(*a*) The lower part of the 80 feet of dark shale called "*a*" is the Eagle Ford shale, locally known as the "Fish beds." The surface portion is composed of Pleistocene terrace material.

(*b*) The 25 feet of very hard limestone represents the upper portion of the Shoal Creek beds.

[1] Statistics collected by Mr. Cyrus C. Babb, of the U. S. Geological Survey.

[2] *e* to *i* inclusive = 490 feet.

[3] It should be borne in mind that the thickness of strata as reported from churn-well drillings is never accurate within 10 feet or more.



HEAD OF KICKAPOO SPRINGS, IN BED OF WEST FORK OF NUECES RIVER, EDWARDS COUNTY, TEXAS.

(*c*) The 90 feet of blue marls are the Del Rio beds.

(*d*) The 1,105 feet of limestone marked "*d*" includes the Fort Worth limestone, 70 feet; the Edwards limestone, about 250 feet; the Comanche Peak and Walnut beds, about 60 feet; 450 to 500 feet of the Glen Rose formation, and 275 to 225 feet of the Travis Peak.

(*e–i*) The beds from "*e*" to "*i*," inclusive, 490 feet, are probably the basement beds of the Travis Peak formation.

(*j*) May or may not belong to formations older than the Cretaceous.

The Groom well, altitude 615 (?) feet, is 1,300 feet deep, and is said to receive its supply of water from stratum "*e*." The total depth of the new Capitol well is about 1,450 feet. Its water is supposed to have its source in a stratum similar to that of the Groom well, but 100 feet deeper from the surface. The surface of the Capitol well is 65 feet lower in altitude than the Groom well. This would give a difference of 165 feet in the depth of stratum "*e*" between one well and the other. This difference may be real, and may result from the fact that the Groom well is geologically lower at the surface, from downthrow by faulting between the two localities.

The St. Edward's well, sunk in the winter of 1892–93, is 2,053 feet deep and obtains its supply from "*i*." At this point the surface is near the top of the Austin chalk, and is 500 feet above the bottom of the blue marl, or "*c*."

Two artesian wells have been bored by the State authorities on the capitol grounds. The first of these was in the year 1858 and was carried to a depth of only 471 feet. The log of this well, as published by Dr. B. F. Shumard, the first State geologist of Texas, is much more detailed than any other in the neighborhood, and is here given.[1] The character and thickness of the different strata passed through, as shown from an examination of the borings preserved by Mr. Peterson, are given in the section, fig. 2 of Pl. XLI (p. 286).

*Log of old State Capitol well, Austin.*

| | Thickness in feet. | Depth in feet to bottom of strata. |
|---|---|---|
| 13. Soil and subsoil | 5 | 5 |
| 12. Soft, white chalky limestone, disintegrating more or less rapidly on exposure to the air | 18 | 23 |
| 11. Moderately hard, bluish-gray and cream-colored argillaceous limestone, containing teeth and scales of fishes, *Inoceramus* (*Trichites*) *lerouxii*, Ammonites, and other fossil remains | 94 | 117 |
| 10. Dark, bluish-gray, indurated marl | 14 | 131 |
| 9. Compact, bluish-gray limestone | 31 | 162 |
| 8. Blue, marly clay with fossil shells coated with iron pyrites, chiefly *Exogyra arietina*, Janira, and Dentalina | 70 | 232 |
| 7. Hard, dark bluish-gray, earthy pyritiferous limestone and shale, containing *Exogyra arietina*, *Gryphæa pitcheri*, Janira, and Toxaster. Many of the fossils of these strata are wholly or in part composed of iron pyrites | 47 | 279 |

[1] Texas Almanac, Richardson and Co., Vol. III, pp. 161–162, Galveston, 1859.

*Log of old State Capitol well, Austin*—Continued.

| | Thickness in feet. | Depth in feet to bottom of strata. |
|---|---|---|
| 6. Soft, sandy, argillaceous limestone, with fossils like those of 7. | 25 | 304 |
| 5. Soft, earthy, sandy, fine-grained limestone of a dull-gray hue... | 6 | 310 |
| 4. Indurated, bluish-gray silico-magnesian limestone, containing a good deal of sulphuret of iron........................................ | 6 | 316 |
| 3. Grayish-white, earthy, fine-textured sandy limestone (magnesian?), with Toxaster and Exogyra. Water at 323 feet; rose 283 feet to within 40 feet of surface.......................... | 13 | 329 |
| 2. Bluish-gray, sandy, magnesian limestone with thin marly partings and abounding in organic remains—*Exogyra arietina*, *Gryphæa pitcheri*, Janira, Dentalina, and fish teeth. Many of these fossils are coated with sulphuret of iron, which gives to them an elegantly bronzed appearance........................ | 48 | 377 |
| 1. Gray, earthy limestone, of a fine sandy texture, with gypsum, nodules of flint, and masses of iron pyrites, and also a few organic remains, chiefly Exogyra and Toxaster................ | 94 | 471 |

The foregoing section of Dr. Shumard's may be interpreted as follows:

No. 13 is the ancient terrace gravel.

Nos. 12, 11, and 10 collectively represent the Austin chalk and the fish beds ("*a*" strata of McGillvray's logs).

No. 9 is the Shoal Creek limestone ("*b*" strata of McGillvray's logs).

No. 8 is the Del Rio clay ("*c*" strata).

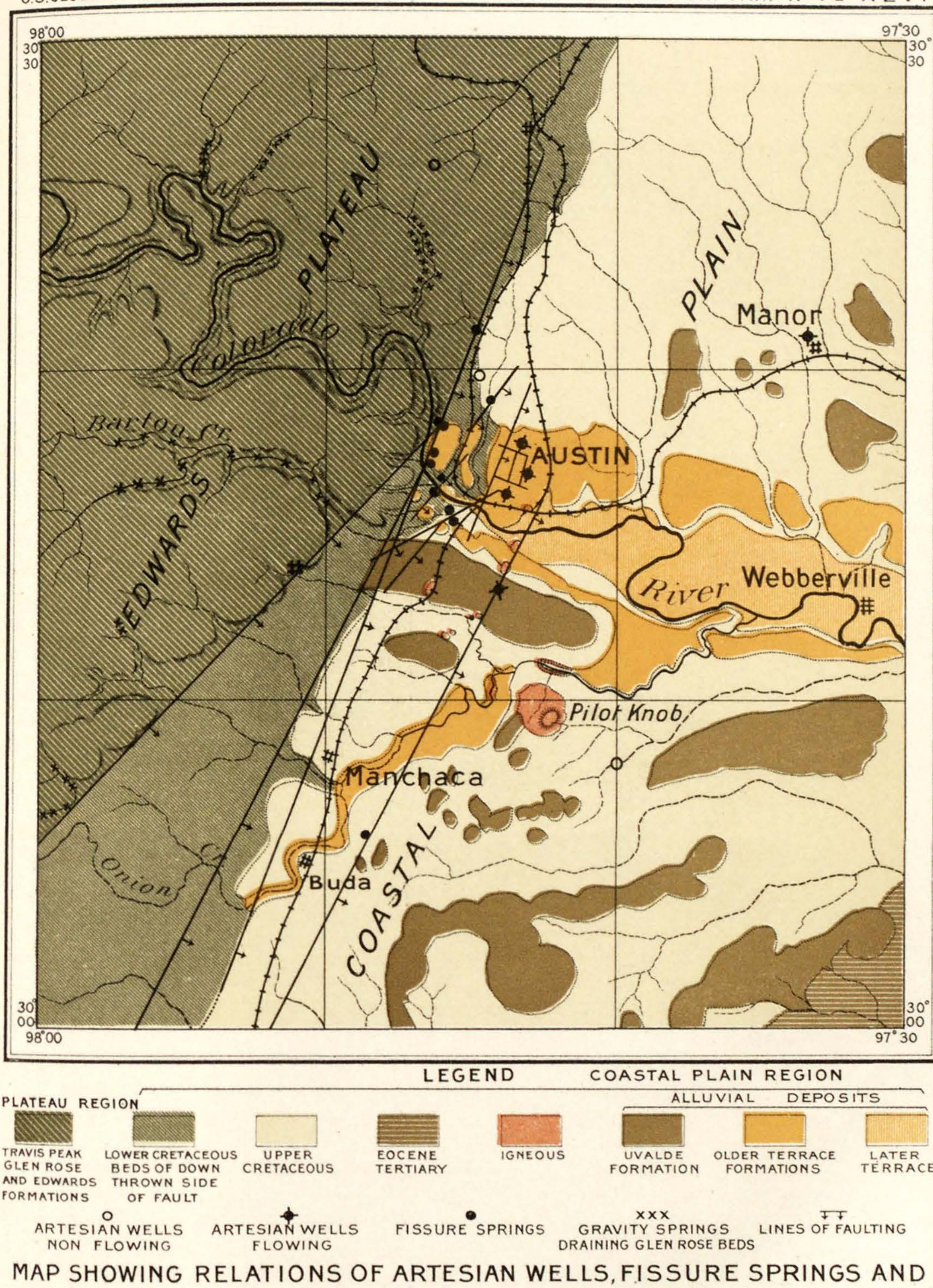
No. 7, at 232 feet, seems to represent the upper part of stratum "*d*" of McGillvray's logs and the top of the combined Fort Worth and Edwards limestones. If this is so and the assumed thickness of the Fort Worth limestones, 70 feet, is correct, the top of the Edwards limestone lies 302 feet beneath the surface of the capitol grounds where this well was dug, and the flow of water struck at 323 feet in No. 11 comes from a horizon not over 21 feet beneath its summit.

No. 6 may be the bottom of the Fort Worth beds, in which case the sulphur water struck would be only 27 feet below the summit of the Edwards limestone, or in the chalky beds immediately below those known as the "lithographic flags."

The beds from 5 to 1, including 192 feet of strata, all undoubtedly belong to the Edwards limestone, extending downward into the magnesian beds exposed about 35 feet above the water in the Bee Caves Bluffs. The interesting fact in relation to this Shumard well section is that every foot of the strata penetrated by the well can be seen outcropping within 2 miles of its location, in the western suburbs of Austin, to the mouth of Bee Creek.

We have long doubted the accuracy of the paleontologic determinations of bed No. 2. Among the fossils enumerated are *Exogyra arietina* and fish teeth, and we know that these do not occur at this horizon. The specimens may have been mixed by the well drill. They certainly do not belong here.

Dr. Francis Moore, in The Texas Almanac for 1860, page 96, has also




LEGEND

PLATEAU REGION

TRAVIS PEAK GLEN ROSE AND EDWARDS FORMATIONS

LOWER CRETACEOUS BEDS OF DOWN THROWN SIDE OF FAULT

COASTAL PLAIN REGION

UPPER CRETACEOUS

EOCENE TERTIARY

IGNEOUS

ALLUVIAL DEPOSITS

UVALDE FORMATION

OLDER TERRACE FORMATIONS

LATER TERRACE

ARTESIAN WELLS NON FLOWING

ARTESIAN WELLS FLOWING

FISSURE SPRINGS

GRAVITY SPRINGS DRAINING GLEN ROSE BEDS

LINES OF FAULTING

# MAP SHOWING RELATIONS OF ARTESIAN WELLS, FISSURE SPRINGS AND IGNEOUS ROCKS OF THE RIO GRANDE PLAIN TO THE BALCONES FAULTING

by

R. T. Hill

SCALE

0 6 12 18 MILES

1897

given a section of this well, which disagrees with the one given by Dr. Shumard both in minor details and in total depth. Dr. Moore's section is as follows:

*Section of old State capitol well, as given by Dr. Francis Moore.*

| | Feet. |
|---|---|
| 1. Limestone containing Ammonites, Inoceramus, etc | 18 |
| 2. Blue argillaceous limestone | 58 |
| 3. Dark carbonaceous marl | 24 |
| 4. Slaty marl | 9 |
| 5. Hard, compact, yellowish limestone with Hippurites | 35 |
| 6. Blue marl with pyrites | 62 |
| 7. Limestone with masses of chert and flint | 61 |
| 8. Siliceous limestone | 2 |
| 9. Blue limestone with pyrites, minute Exogyra, etc | 150 |
| Total | 419 |

Our interpretation of these beds is as follows: Nos. 1 and 2, Austin chalk; 3 and 4, fish beds; 5, Shoal Creek limestone (Del Rio clays); 7, 8, and 9, Edwards limestone.

We have not been able to obtain a complete log of the second well bored by the State, six or seven years ago. It is reported that it was drilled to about 1,450 feet, and stopped just as it reached, and before penetrating, the greater basement group of water-bearing strata. Hence its flow is feeble.

The well at the colored asylum, 2 or 3 miles northwest of the State capitol, commences in the Shoal Creek limestone and passes through the Del Rio and Fort Worth beds and penetrates the Edwards limestone. The water does not rise to the surface.

The following record of a well on East Fifth street, in process of boring on May 1, 1897, which was obtained for us by Mr. E. W. Parker, of this Survey, will also afford an idea of the nature of the rocks encountered.

*Record of well-boring on East Fifth street, Austin.*

| | Thickness in feet. | Depth in feet to bottom of strata. |
|---|---|---|
| Dark soil, old river alluvium | 30 | 30 |
| White limestone (Austin chalk) | 40 | 70 |
| Dark shale (Eagle Ford) | 60 | 130 |
| White limestone (Shoal Creek) | 40 | 170 |
| Blue marl (Del Rio) | 60 | 230 |
| Main limestone (Washita and Edwards) | 120 | 350 |
| Sulphur water, no flow | | 350 |
| Alternate layers of sand and limestone, varying in thickness | 150 | 500 |
| Flow begins (sulphur water) (Glen Rose?) | | 400 |
| Solid limestone | 200 | 700 |

Flowing 20,000 gallons of sulphur water per day. Main Trinity flow expected at 1,400 feet. First sulphur water, 50 feet below summit of Edwards beds.

Further particulars of this well, since completed, have been furnished as follows by Mr. H. McGillvray:

*Log of the Austin Natatorium artesian well, drilled by Hugh McGillvray, at the corner of San Jacinto and Fifth streets, Austin, Texas.*

| Bed. | Thickness. | Depth to bottom of strata. | Remarks. |
|---|---|---|---|
| | *Feet.* | *Feet.* | |
| 19. Dirt | 20 | 20 | |
| 18. Gravel bed | 5 | 25 | |
| 17. Limestone | 100 | 125 | |
| 16. Shale | 70 | 195 | |
| 15. Limestone | 25 | 220 | |
| 14. Blue marl | 40 | 260 | |
| 13. Limestone | 100 | 360 | |
| 12. Sandrock | 10 | 370 | |
| 11. Limestone | 70 | 440 | |
| 10. Sandrock and limestone in strata of about 10 feet. | 150 | 590 | Mineral water of medicinal value.[1] |
| 9. Limestone | 600 | 1, 190 | |
| 8. Sandrock | 25 | 1, 215 | Small flow. |
| 7. Limestone | 300 | 1, 515 | |
| 6. Blue shale | 60 | 1, 575 | |
| 5. Limestone | 100 | 1, 675 | |
| 4. Sandrock | 200 | 1, 875 | Main flow; freestone water. |
| 3. Blue shale | 40 | 1, 915 | |
| 2. Sandrock | 50 | 1, 965 | |
| 1. Blue shale | 60 | 2, 025 | |
| Total depth of well | 2, 025 | | |

[1] See analysis by Prof. H. W. Harper, given on p. 304.

The flow is about 250,000 gallons a day, and is increasing; temperature, 100°. Pressure has not been tested, but is probably between 70 and 80 pounds. Casing, 300 feet 10-inch pipe, 600 feet 8-inch pipe, 300 feet 7-inch pipe, 200 feet 6-inch pipe. The medicinal vein has been separated from the freestone vein by letting the flow come up between the 8- and 10-inch pipes. This water will be used for a natatorium and sanitarium.

There is a discrepancy between the two records given above which we can not explain. So much as is given in Mr. Parker's log is more in harmony with the stratigraphic measurements made in the region. The strata from 1,915 to 2,025 feet may be Paleozoic.

### THE MANOR WELL.

A valuable contribution to the extent of the artesian field in Travis County was made by the drilling of the well at Manor, in 1895. This well is situated about 20 miles east of Austin, near the extreme eastern margin of the Black Prairie and on the outcrop of the base of the lower portion of the equivalent of the Eagle Pass beds. Through the kindness of Mr. E. W. Parker we have been enabled to obtain the following accurate log of this well, which enables us to locate the water vein with definiteness and also to ascertain the thickness of the Taylor and Austin beds of the Upper Cretaceous, which had not until then been accurately measured:

*Log of well at Manor.*[1] (See fig. 7 of Pl. XLI.)

| | Thickness in feet. | Depth in feet to bottom of strata. |
|---|---|---|
| 16. Black soil | 6 | 6 |
| 15. Yellow clay | 11 | 17 |
| 14. Flint rock and gravel (surface water) | 3 | 20 |
| 13. Yellow and joint clay | 30 | 50 |
| 12. Blue clay | 540 | 590 |
| 11. Soft white rock | 74 | 664 |
| 10. Indigo-blue clay | 1 | 665 |
| 9. Rock, hard and soft alternately | 335 | 1,000 |
| *a* 8. Shale (caves badly) | 25 | 1,025 |
| *b* 7. Hard rock | 50 | 1,075 |
| *c* 6. Blue clay (caves) | 60 | 1,135 |
| *d* { 5. Lime rock | 115 | 1,250 |
| *d* { 4. Water-bearing rock (water flows at 1,250 feet[2]) | 150 | 1,400 |
| *d* { 3. Hard rock | 20 | 1,420 |
| [From this point downward we have no detailed record.] | | |

Quantity of water discharged per hour, 4,166⅔ gallons. Size of discharge pipe, 6 inches. Temperature of water, 93°.

There are several records of this well, which show slight discrepancies. The one presented gives the most detail and appears to be most nearly accurate.

Having very recently studied the section of the rocks through which the Manor well passes, we can easily identify the different strata as follows:

16, 15, and 14 are the post-Cretaceous Uvalde formation, here having a total thickness of 20 feet.

13 and 12 are the Webberville and Taylor marls or "joint clays."

11, 10, and 9 are the Austin chalk, having an aggregate thickness of 410 feet.

8 is the Eagle Ford shale, and is the stratum "*a*" of the Asylum well section at Austin (p. 280).

7 is the Shoal Creek limestone ("*b*" of Asylum well).

6 is the Del Rio clay ("*c*" of Asylum well).

5, 4, and 3 are the Edwards limestone, the upper part of the limestone group, marked "*d*" on the Austin section.

2. Salt clay occurs at 1,920 feet.

1. Bottom of well at last information, April 19, 1897, was 2,220 feet.

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[1] Furnished by Calcasieu Lumber Company. Exact location of well: Middle of lot No. 13, in block No. 25, according to map and plot of Manor.

[2] Three separate accounts give this strike at 1,250, 1,265, and 1,280 feet, respectively.

Water will rise in pipe about 30 feet above the surface. The first effort to dig this well failed, as it caved in at the depth of about 1,100 feet. The present well, bored by Mr. G. J. Eppright, was finished about February 13, 1896, and cost $4,060.

All the information we have been able to obtain concerning the wells of Travis County is plotted on Pl. XLVI (p. 282). In this we have endeavored to show the geologic position of the surface at the location of the well, and have referred all the wells to a common geologic datum—the top of the Shoal Creek limestone, which seems to be recognizable in most of the well records.

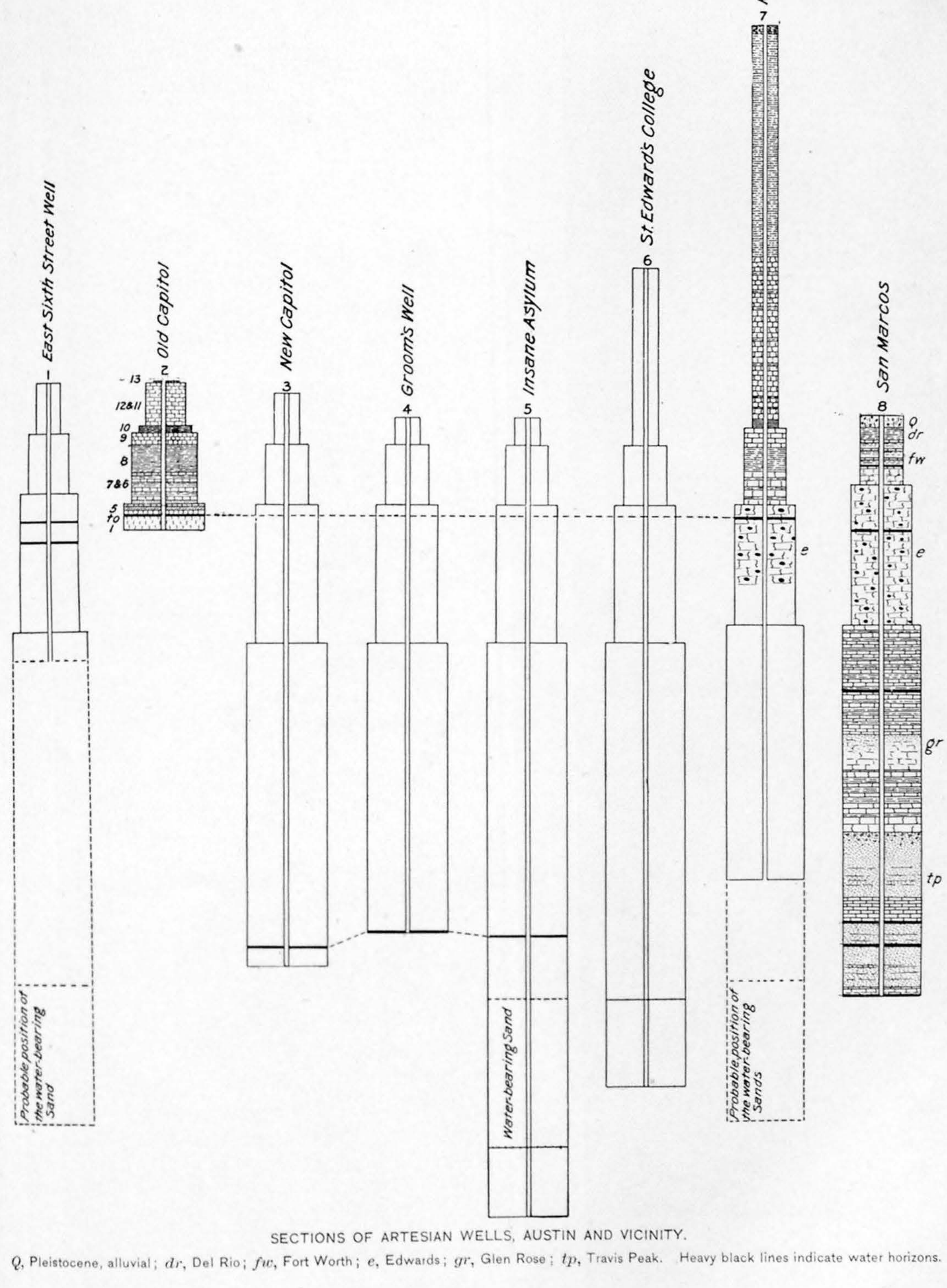
From the records thus plotted valuable generalizations can be made, as follows: There are at least three important water-bearing horizons in the Lower Cretaceous series beneath Austin. The first of these, which is conspicuously illustrated by the Manor flow of 1896, is recognizable in the old and new Capitol wells and in the Sixth street well now under construction. This water was also probably struck in the other three wells, but we have no detailed record thereof. It is highly charged with sulphur and injurious mineral ingredients, as will be described later, and in all instances where encountered should be cased off. This water is undoubtedly obtained in the Edwards beds, from strata which, according to the best computations we can make at present, lie from 21 to 50 feet below their summit, or 90 to 140 feet below the summit of the limestone group, commencing below the Del Rio clays.[1]

The next flow occurs about 1,000 feet lower in the series, in strata which may be considered as practically the transition ground between the Glen Rose and Travis Peak beds. This water is the uppermost of the water-bearing strata characterizing the base of the Cretaceous.

In the Asylum well the above-mentioned flow is separated by about 160 feet of limestone and shale from the third and lowest of the water-bearing beds, the basement, Travis Peak, or "Trinity" sands. These produce the purest water, and ordinarily the most, of all the water-bearing strata of the series. According to Mr. McGillvray's record, these water-bearing beds are 315 feet in thickness, and are marked by thin bands of reddish clay.

It will be seen that only one of the wells has positively penetrated the entire series of beds composing the Cretaceous system, thereby exploiting its fullest capacity and reaching into the underlying impervious Paleozoic formations. This is the well at the Insane Asylum. It is very probable that the well at St. Edward's College has also penetrated the entire thickness of water-bearing strata, but we have been unable to obtain an exact log of it. All the other wells are incomplete, inasmuch as they have not reached the best water of the district—that which lies close to the base of the Cretaceous system,

[1] According to the best computations we can at present make, sulphur water was struck in the Edwards formation at a depth of 120 feet below the bottom of the Del Rio clays in the Sixth street well, 78 feet in the old Capitol well, 50 feet in the new Capitol well, and 35 feet in the Manor well.




SECTIONS OF ARTESIAN WELLS, AUSTIN AND VICINITY.

*Q*, Pleistocene, alluvial; *dr*, Del Rio; *fw*, Fort Worth; *e*, Edwards; *gr*, Glen Rose; *tp*, Travis Peak. Heavy black lines indicate water horizons.

at a depth of from 1,400 to 1,750 feet below the top of the Shoal Creek limestone.

The fact that the water does not flow out at the surface at St. Edward's College (altitude 660 feet) indicates that the 650-foot contour on the Austin geologic sheet marks the possible limit of flowing wells in the vicinity.

THE SAN MARCOS WELL.

Two records of an artesian well drilled for the United States Fish Commission at San Marcos in the year 1895 afford the most complete well section we have been able to obtain in the Rio Grande Plain. The records consist of (1) a tabulated record kept by Mr. Judson, the contractor, and (2) a series of numbered samples furnished us by the United States Fish Commission for determination. These, as will be shown, are not harmonious as to detail, and well illustrate the difficulty of obtaining exact logs.

The well was drilled to a depth of 1,490 feet, and when stopped was still in the Cretaceous formations, about 175 feet above the estimated base of the Asylum well, at Austin.

*Log of the San Marcos well; furnished by Mr. Judson, the contractor.*

[See fig. 8 of Pl. XLI.]

| | Thickness in feet. | Depth in feet to bottom of strata. |
|---|---|---|
| Soil, wash, and gravel | 30 | 30 |
| Blue clay and limestone (Del Rio beds) | 20 | 50 |
| Hard gray limestone (Fort Worth) | 78 | 128 |
| Cave with abundant water | 2 | 130 |
| Brittle white limestone (Edwards) | 72 | 202 |
| Black flints (Edwards) | | 202 |
| Blue shale (Edwards) | 12 | 214 |
| Black flints (Edwards) | 6 | 220 |
| Soft white limestone (Edwards) | 106 | 326 |
| Hard white limestone (Edwards) | 57 | 383 |
| Hard gray limestone | 37 | 410 |
| Rotten limestone | 41 | 451 |
| Hard gray limestone | 35 | 486 |
| Trace of blue clay | | 486 |
| White limestone | 47 | 533 |
| Hard white limestone | 92 | 625 |
| Hard blue clay | 27 | 652 |
| Hard white limestone | 78 | 730 |
| White clay | 5 | 735 |
| White limestone | 25 | 760 |
| White clay | 5 | 765 |
| Limestone | 34 | 799 |
| Hard blue limestone | 11 | 810 |
| Brittle white limestone | 225 | 1,035 |
| Gray limestone | 77 | 1,112 |
| Blue clay | 7 | 1,119 |
| Hard gray limestone | 59 | 1,178 |
| White limestone | 71 | 1,249 |
| Hard yellow limestone | 42 | 1,291 |

*Log of the San Marcos well; furnished by Mr. Judson, the contractor*—Continued.

| | Thickness in feet. | Depth in feet to bottom of strata. |
|---|---|---|
| Sandstone | 54 | 1,345 |
| Hard gray limestone | 27 | 1,374 |
| Soft limestone | 31 | 1,405 |
| Black sandstone | 10 | 1,415 |
| Soft gray sandy limestone | 60 | 1,475 |
| Blue clay to bottom of well | 15 | 1,490 |

*Determination of samples (U. S. Fish Commission) from San Marcos artesian well.*

| Number. | Depth in feet. | Remarks. |
|---|---|---|
| 2 | 40 | Sticky blue clay, probably Del Rio. |
| 3 | 65 | Mixture of above clay and limestone. |
| 4 | 70 | Grayish-white limestone having lithologic characteristics of Fort Worth limestone. |
| 5 | 81 | More chalky limestone mixed with above; yellow ferruginous spots, giving it a yellowish tinge. |
| 6 | 85 | Slightly lighter color, faintly blue, more marly material; glauconitic. |
| 7 | 108 | Same as 6; still slightly lighter in color; very argillaceous. |
| 8 | 140 | Pure white marl; chalk in lumps. |
| 9 | 160 | Gritty, granular, with small fragments of firmer limestone; wet. |
| 10 | 167 | Firm, white chalky limestone, resembling Edwards; specimen dry. |
| 11 | 187 | Wet angular fragments of limestone; some particles seem rounded, one quartz grain. |
| 12 | 200 | Wet brown sand with intensely ferruginous specks; apparently decomposing pyrites. |
| 13 | 215 | Dry, white, limy material with mixture of pyritiferous specks like foregoing; shows decided change. |
| 14 | 295 | Dry, white chalky marl with blue clay specks and pyritiferous spots; minutely granular, mealy gray limestone. |
| 15 | 585 | Wet, thoroughly sticky calcareous clay. |
| 16 | 635 | Wet, blue pasty clay like above. |
| 17 | 703 | Slightly more calcareous, but still as above. |
| 18 | 809 | Gray argillaceous limestone. |
| 19 | 810 | An intensely ferruginous calcareous pack sand, rapidly oxidizing; resembles first sand of Trinity at Travis Peak; fetid. |
| 20 | 895 | Wet limestone, smelling strongly of sulphur; odor of petroleum; *Patellina texana* (Roemer), a typical lower Glen Rose fossil. |
| 21 | 1,025 | Dry chalky marl. |
| 22 | 1,050 | Dry grit composed of small rounded grains, lime, quartz, and calcite. |
| 23 | 1,150 | Wet, fine-grained calcareous sand; apparently downward continuation of above. |
| 24 | 1,179 | Do. |

*Determination of samples (U. S. Fish Commission) from San Marcos artesian well*—Cont'd.

| Number. | Depth in feet. | Remarks. |
|---|---|---|
| 25 | 1,190 | Very fine sand, very calcareous, almost impalpable. |
| 26 | 1,200 | Still finer grit, almost like flour, calcareous, white, free from ferrugination; dry. |
| 27 | 1,229 | Wet, granular blue limestone. |
| 28 | 1,291 | Very fine pack sand, homogeneously ferruginous, with little ferruginous grains. |
| 29 | 1,315 | Very fine sand, chocolate colored; very ferruginous matter and possibly mixture of red clay. |
| 30 | 1,322 | Do. |
| 31 | 1,338 | Wet, very fine sand, free from coloration of above. |
| 32 | 1,345 | Dry, very fine sand, light colored. |
| 33 | 1,398 | Still finer sand, same as above. |
| 34 | 1,405 | Do. |
| 35 | 1,430 | Do. |
| 36 | 1,445 | Pyritiferous ferrugination again. |
| 37 | 1,465 | Dry, granular limy material with black flakes. |
| 38 | 1,480 | Dry, granular limy material with black flakes; finer than 37, very argillaceous; some fine sand. |

The San Marcos well commences in strata about 150 feet lower geologically than the Austin wells, the former commencing near the base of the Del Rio and the latter near the base of the Austin chalk. If the figures for the Insane Asylum well are correct, the San Marcos well has not yet reached the bottom of the Cretaceous strata by some 100 feet. In drilling the San Marcos well seven horizons of underground water were encountered between 128 and 1,450 feet, as follows:

*Horizons of underground water in San Marcos well.*

| Number. | Feet. | Remarks. |
|---|---|---|
| 1 | 128 | Abundant water; sulphur. |
| 2 | 191 | Inexhaustible water; good quality. |
| 3 | 652 | Sulphur impregnation noticed. |
| 4 | 1,178 | Good water; shell formation. |
| 5 | 1,291 | Good water; rose 400 feet. |
| 6 | 1,345 | Flow increased to 4 gallons per minute. |
| 7 | 1,475 | Flow increased to 6 gallons per minute. |

The first of these horizons (No. 1), struck at 78 feet below the summit of the Fort Worth limestone, or about 8 feet below the estimated summit of the Edwards beds, is the same general geologic formation as that giving the first flow of the Austin, Manor, and San Antonio wells.

No. 2 is probably a lower horizon in the Edwards beds, and its freedom from sulphur is a valuable quality.

The sulphur impregnation in No. 3 is probably a local layer in the Glen Rose beds.

No. 4 seems to correspond almost exactly with flow No. 2 of the Austin wells, that from which the new Capitol and Groom wells are supplied.

Nos. 5, 6, and 7 are, beyond any reasonable doubt, from the basement sands.

There are one or two elements of perplexity concerning the San Marcos well which also present themselves at San Antonio. It is located almost upon the line of a great fault, which may be seen on the west side of the spring pool a short distance from the well. It is possible that the drill hole may cross this fault line not far below the surface, and that the so-called caverns are its waterworn fissures. The fact that this water was full of peculiar cave-inhabiting animals[1] indicates that there are cavities beneath the ground, the extent of which, however, can not be stated. These may be pockets, such as are seen in the outcrops of the Edwards limestone, or they may be extensive caves, like the Hillcoat caverns of Edwards County. The fact that the drill passed through only 2 feet of cavity rather opposes the latter hypothesis.

### SAN ANTONIO WELLS.

A large number of wells have been drilled in and around the city of San Antonio. They occur in nearly all parts of the city and its adjacent suburbs, as shown in fig. 73 (p. 310).

These wells, like the well at San Marcos, are all drilled close to the lines of faulting, and hence present many anomalies. It is very possible that the drill often crosses the lines of faulting, producing inconsistencies in the record.

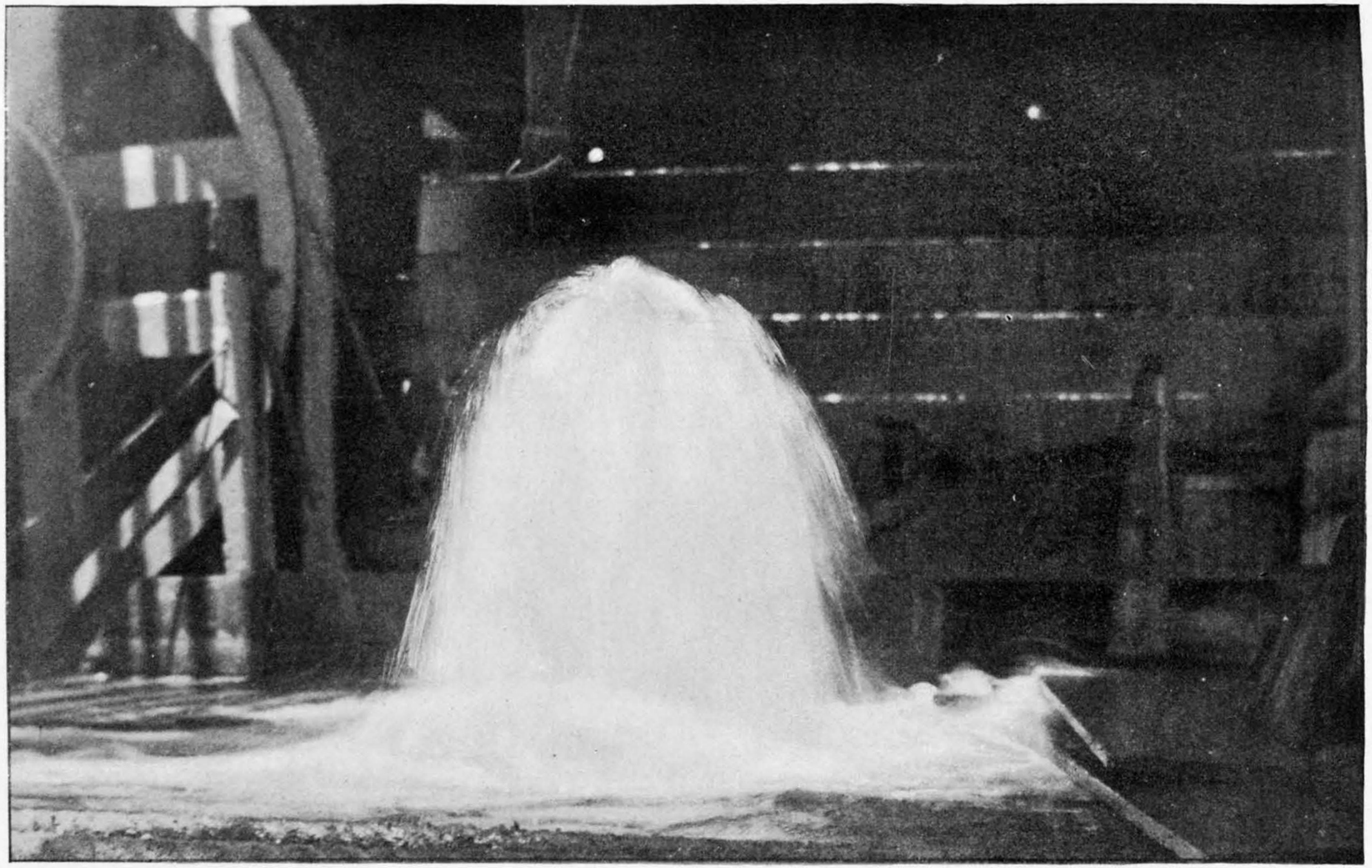
The following list is incomplete, but it will afford an idea of the approximate number and location of the wells.

*List of artesian wells reported at San Antonio.*

| No. | Owner. | Location. | Depth of well. | Depth to water.*a* | Reported flow in 24 hours.*b* | Remarks. |
|---|---|---|---|---|---|---|
| | | | *Feet.* | *Feet.* | *Gallons.* | |
| 1 | West End Town Co. | West End Lake | 260 | | 216,000 | |
| 2 | Russ | do | 250 | | 85,000 | |
| 3 | Brendle | 3 miles north of court-house. | 583 | 493 | None. | Water rises to 50 feet of surface. |
| 4 | Yantis | 2½ miles northwest | 540 | | | Penetrated 90 feet of cavern. |
| 5 | Waterworks Co. No. 1. | ½ mile below head of river. | 630 | | | City water supply. |
| 6 | Waterworks Co. No. 2. | Lower pump-house, 1½ mile below head of river. | 750 | | 1,500,000 | Do. |

*a* Depth to water when not otherwise stated is presumed to be at bottom of well.
*b* These statistics are not in every case trustworthy, being usually exaggerated.

[1] Described by Messrs. Stejneger and Benedict, in Proc. U. S. Nat. Mus., Vol. XVIII, April, 1896, Nos. 1087 and 1088.



ARTESIAN WELL, CITY WATERWORKS COMPANY, SAN ANTONIO, TEXAS.

*List of artesian wells reported at San Antonio*—Continued.

| No. | Owner. | Location. | Depth of well. | Depth to water. | Reported flow in 24 hours. | Remarks. |
|---|---|---|---|---|---|---|
| | | | *Feet.* | *Feet.* | *Gallons.* | |
| 7 | Waterworks Co. No. 3. | Lower pump-house, 1½ mile below head of river. | 780 | ........ | 1,500,000 | City water supply. |
| 8 | Waterworks Co. No. 4. | .....do .............. | 780 | ........ | 1,500,000 | Do. |
| 9 | Collins Manufacturing Co. No. 1. | 900 yards north by northeast from court-house. | 650 | ........ | 1,000,000 | |
| 10 | Collins Manufacturing Co. No. 2. | .....do .............. | 650 | ........ | 1,000,000 | |
| 11 | Collins Manufacturing Co. No. 3. | .....do .............. | 715 | ........ | 750,000 | |
| 12 | Collins Manufacturing Co. No. 4. | .....do .............. | 825 | ........ | 250,000 | |
| 13 | Collins Manufacturing Co. No. 5. | .....do .............. | 1,100 | 880 | 250,000 | Soft, sweet water. |
| 14 | Waterworks Co. No. 5. | Market street....... | 880 | ........ | 5,500,000 | City water supply. |
| 15 | Waterworks Co. No. 6. | .....do .............. | 880 | ........ | 3,000,000 | Do. |
| 16 | Waterworks Co. No. 7. | .....do .............. | 880 | ........ | 3,000,000 | Do. |
| 17 | Waterworks Co. No. 8. | .....do .............. | 880 | ........ | 3,000,000 | Do. |
| 18 | Waterworks Co. No. 9. | .....do .............. | 880 | ........ | 3,000,000 | Do. |
| 19 | Schulz Sanitarium. | Alamo plaza........ | 782 | ........ | 800,000 | |
| 20 | Crystal Ice Co. No. 1. | Eighth street....... | 657 | 600 | 36,000 | Soft, sweet water. |
| 21 | Crystal Ice Co. No. 2. | .....do .............. | ........ | ........ | ............ | |
| 22 | Crystal Ice Co. No. 3. | .....do .............. | ........ | ........ | ............ | |
| 23 | Crystal Ice Co. No. 4. | .....do .............. | ........ | ........ | ............ | |
| 24 | San Antonio Street Rwy. Co. | Tenth street........ | 980 | ........ | 225,000 | |
| 25 | County of Bexar.. | Court-house, center of city. | 870 | ........ | 1,240,000 | Gas and water at 460 and 620 ft. |
| 26 | State of Texas.... | 3 miles southeast of city, at insane asylum. | 1,100 | ........ | ............ | Hot sulphur water. |
| 27 | Colonel Terrell, U. S. A. | 3 miles south of city. | 1,900 | ........ | 800,000 | Do. |
| 28 | Mrs. C. Kampman. | 3 miles east......... | ........ | ........ | ............ | |
| 29 | George Dullnig... | 6 miles southeast... | 2,215 | 1,800 | ............ | |
| 30 | San Antonio Street Railway. | Tenth street........ | 1,140 | ........ | 25,000 | |
| 31 | Santa Rosa Hospital. | ...................... | 1,000 | ........ | 150,000 | Sulphur. |
| 32 | .....do ............ | ...................... | 1,250 | ........ | ............ | |
| 33 | Schulz well....... | ...................... | 822 | ........ | 300,000 | |
| 34 | Union Meat Co ... | Southwest of city... | 1,202 | ........ | 144,000 | |
| 35 | Menger Hotel...... | ...................... | 1,160 | ........ | 250,000 | |
| 36 | Kiebling well..... | Near S. A. & A. P. depot. | 1,100 | ........ | 200,000 | |
| 37 | Dignourty well... | West of town....... | 465 | ........ | 80,000 | |
| 38 | Kampman well ... | Velita street........ | 850 | ........ | 300,000 | |
| 39 | Epps well......... | Southwest of court-house. | 884 | ........ | 100,000 | |
| 40 | Van Dale ......... | 1 mile west ......... | 835 | ........ | 80,000 | |

The underlying geologic structure of San Antonio and immediate vicinity is difficult to discover, owing to the fact that the region is mostly covered by a sheet of superficial gravel and marl (the Uvalde formation). The depth of this alluvial sheet can not be stated or even approximated with accuracy. In some wells it is 40 feet in depth, in others more than 100. The variation in thickness is caused by its having been deposited upon the unevenly eroded surface of the Austin chalk. The Austin chalk outcrops near the San Antonio and the San Pedro springs, striking in a northeast-southwest direction at those points. It is probable that southeast of its outcrop the Taylor formation underlies the Uvalde, and that all the Cretaceous beds descend gently toward the southeast. The concealed strata are much jointed and possibly faulted in this area. The limestones are also very cavernous.

We have been unable to obtain detailed records of many of the San Antonio wells, but the following data will give a general idea of their character.

Mr. C. S. Austin informs us that the wells in the valley of Alazon Creek, just west of the center of the city, range from 225 to 450 feet in depth. Their flow is about as strong as that of the wells in the valley of the San Antonio.

*Log of well of West End Town Company, San Antonio.*

| | Thickness in feet. | Depth in feet to bottom of strata. |
|---|---|---|
| Black alluvial soil | 5 | 5 |
| Yellow clay | 10 | 15 |
| Coarse gravel | 10 | 25 |
| Soft blue rock impregnated with petroleum | 235 | 260 |

Depth, 260 feet; flow, 150 gallons per minute.

Gustave Jermy[1] makes the following notes concerning artesian wells in the vicinity of San Antonio:

One, of a depth of 225 feet, in the western addition to the city of San Antonio, with a fine quality of drinking water which forces its way to the surface. Another, 4 miles east of San Antonio, near the Salado, which was sunk to a depth of 450 feet and also brings a constant stream of water, containing hydrosulphuric acid, to a considerable height. It is clear and is being utilized for the ordinary wants of man and beast, but could be made more useful on account of its medicinal qualities.

The following record has been published by Roessler:[2]

*Log of Banes well, San Antonio.*

| | Thickness in feet. | Depth in feet to bottom of strata. |
|---|---|---|
| Gravel | 8 | 8 |
| Yellow clay | 50 | 58 |
| Blue clay | 175 | 233 |
| Churned mud | 262 | 495 |
| Slate | 5 | 500 |

Water at 225 feet; oil at 250 feet; water at 350 feet.

[1]Texas Geological and Mineralogical Survey, First Rept. of Progress, for 1888, 1889, p. 64.

[2]Report of F. E. Roessler, division field agent for Texas, Fifty-first Congress, first session, Ex. Doc. No. 222, Washington, 1890, pp. 243-319.

Other wells have been reported as follows:

*Log of well of Mr. H. Brendle, located about 1 mile north of west of San Pedro Springs.*

| | Thickness in feet. | Depth in feet to bottom of strata. |
|---|---|---|
| Black soil and gravel | 4 | 4 |
| Cemented gravel | 2 | 6 |
| Yellow clay (Taylor) | 72 | 78 |
| Bluish limestone (Austin) | 180 | 258 |
| Very hard limestone, containing pyrites in 10 or 15 horizons, occurring in hard lumps, probably marcasite | 235 | 493 |
| Water (rises to within 50 feet of surface) | ... | 493 |
| Blue marl and clay, described as sea mud | 90 | 583 |

Eyeless salamanders were found in the water. It is located in a faulted district. A well one-third of a mile southeast of this struck water at 540 feet and went 90 feet through a cave.

*Log of well of Crystal Ice Company, San Antonio (No. 1).*

| | Thickness in feet. | Depth in feet to bottom of strata. |
|---|---|---|
| Black residual soil | 4 | 4 |
| Impervious yellow clay | 12 | 16 |
| Gravel | 20 | 36 |
| Blue clay | 300 | 336 |
| Soapstone (clay) | 250 | 586 |
| Black mud | 60 | 646 |
| Sandstone, very hard | 5 | 651 |

Sulphur water and gas, 375 feet; flow, about 25 gallons per minute; very strong flow at 600 feet; sweet water.

Judson Brothers, who bored the wells for the Waterworks Company, furnish us the following record:

*Log of Waterworks Company wells, San Antonio.*

| | Feet. |
|---|---|
| Alluvial soil | 16 |
| Blue clay | 400 |
| Limestone | 304 |
| Blue clay | 40 |
| Hard limestone | 120 |

Total depth of well, 880 feet. Has a strong flow of water, struck at a depth of 880 feet from the surface. This flow fluctuates slightly, but has a maximum head of 51 feet above the river. The flow is estimated at 5,400,000 gallons per diem. Water slightly impregnated with lime.

The water came from a crystalline limestone out of a fissure 3 feet in depth. We have sunk nine flowing wells for Mr. Breckenridge, varying in depth from 630 to 880 feet. All passed through about the same formation.

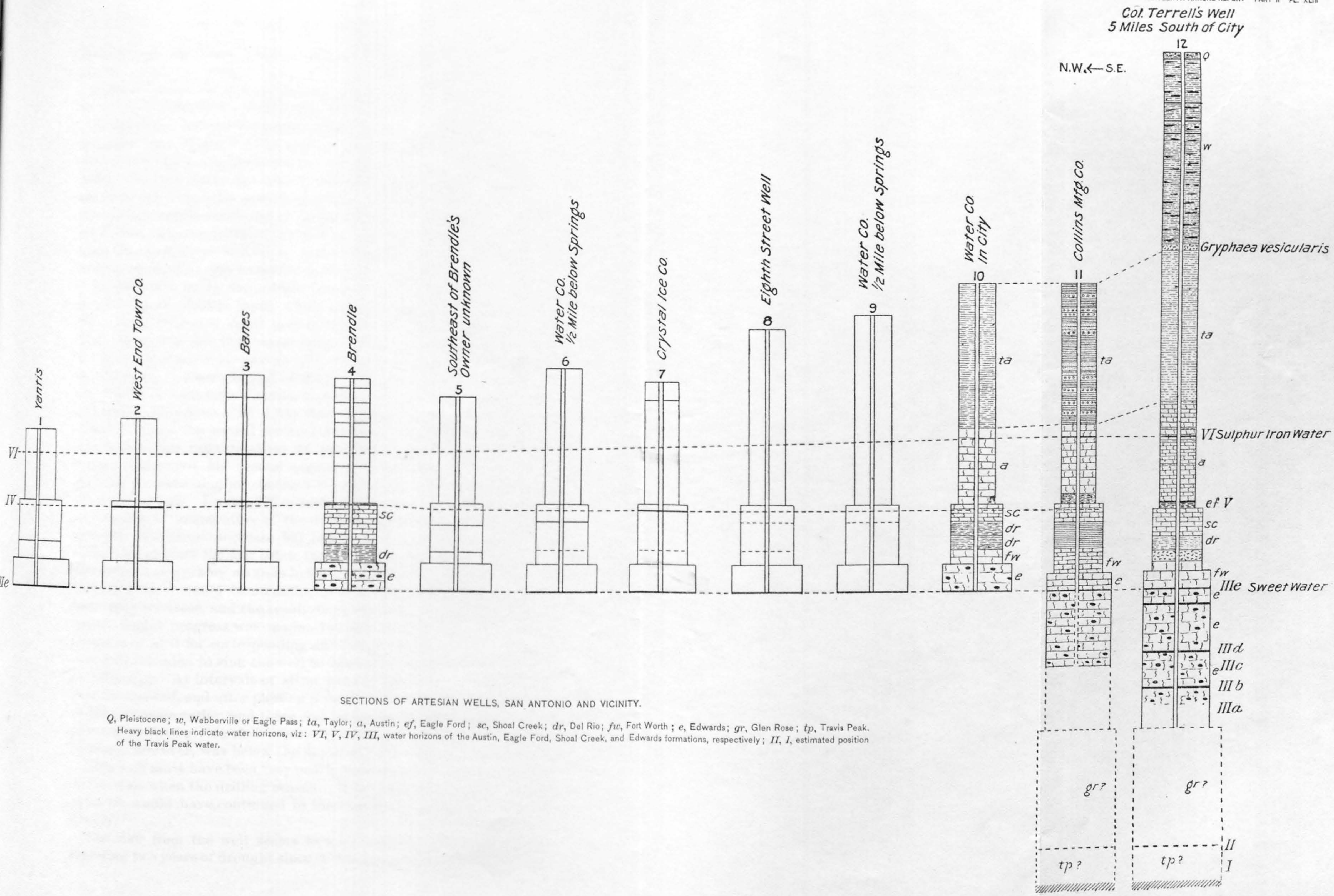
The following information concerning the well of F. F. Collins Manufacturing Company was furnished through the owners by Mr. James Brown, who drilled the well. (See Pl. XLIII.) The well is located 900 yards northeast by north from the new court-house of Bexar County, on the banks of the San Antonio River. It is 20 feet from the water line of the river; the level of the surface is 14 feet above the river. The well was first sunk to a depth of 40 feet, 10 inches in diameter, and cased, then continued to a depth of 440 feet, or until the sulphur and iron water was reached. It was then cased with 6-inch casing from the top, carried down that size through 280 feet of a light-colored limestone rock, and through 75 or 80 feet of a blue clay, and sufficiently far in the next formation to be secure. From this point the boring was continued through this rock, which was firm enough to stand up without any casing, until a final depth of about 1,100 feet was reached. The "sweet" water was found in this rock at a depth of 880 feet from the surface, and gave a flow of 200,000 to 250,000 gallons in twenty-four hours. This water rises in a standpipe to about 40 feet above the level of the San Antonio River. The sulphur-iron water struck 440 feet below the surface gave a flow of 5 to 10 gallons per minute; associated with it was gas, given off at the rate of 30 to 40 cubic feet per minute. The gas burns well in an Argand burner with Welsbach attachment.

The flow of "sweet" water comes from the upper part of the Edwards limestone, 50 to 70 feet below its top.

The most complete data we have been able to receive concerning the strata penetrated at San Antonio and vicinity are those furnished by Col. C. M. Terrell, United States Army, concerning the well drilled by him.

This well is 2 miles due south of the city limits, on the west side of the main road to Pleasanton and on the divide between San Antonio and the Medina River. It is on a summit, and so situated that the water is now carried for purposes of irrigation to all parts of the land through 6½ miles of surface acequias, and the surplus water can be turned into the San Antonio River, 4 miles east, or into the Medina River, 5 miles south.

The well is about 1,900 feet deep; the yield is estimated at 800,000 gallons per day. Temperature, 106° F. The water overflowed with force from the top of a pipe 68 feet above the surface. When a 22-foot joint of pipe was added the water rose in the pipe to about 84 feet above the surface. This well at the surface of the ground is 51 feet 6 inches below the level of the main springs of the San Antonio River, and at its greatest depth is 1,270 feet below the level of the Gulf of Mexico. The elevation above the surface to which the water will rise is 84 feet, or 32 feet 6 inches above the level of San Antonio springs. Colonel Terrell says that this fact convinces him "that the confined basin from which it comes has no direct connection with the source of supply of the other flowing wells in this vicinity," but in this opinion we are inclined to differ from him, for there is every evidence that the sources of the waters are identical. While lower in altitude than the San Antonio springs, the outlet of this well is some 650 feet higher in the geologic series. These differences, and similar conditions in all the wells east and south of the city, such as Kampmann's, Dullnig's,




SECTIONS OF ARTESIAN WELLS, SAN ANTONIO AND VICINITY.

*Q*, Pleistocene; *w*, Webberville or Eagle Pass; *ta*, Taylor; *a*, Austin; *ef*, Eagle Ford; *sc*, Shoal Creek; *dr*, Del Rio; *fw*, Fort Worth; *e*, Edwards; *gr*, Glen Rose; *tp*, Travis Peak. Heavy black lines indicate water horizons, viz: *VI*, *V*, *IV*, *III*, water horizons of the Austin, Eagle Ford, Shoal Creek, and Edwards formations, respectively; *II*, *I*, estimated position of the Travis Peak water.

and that at the State Lunatic Asylum, are due to the dips of the strata.

Colonel Terrell gives the following log of the well, a graphic record of which is given in Pl. XLIII, fig. 12.

Formations: Ordinary surface soil, 6 feet; coarse gravel, 5 feet; yellow joint clay, 50 feet. From this the drill passed into the blue marl. At 140 feet below surface it passed through about 10 inches of brown coal in the blue marl; the same at 240 feet; the same at 280 feet. Six hundred feet below the surface it passed through 20 feet of very hard shell conglomerate composed of *Gryphæa vesicularis* (identified by Robert T. Hill), the largest shell of which is 4 by 3½ inches, thickness of walls one-half to three-fourths inch. The blue-marl cement of this shell conglomerate was hardened into rock.

At 1,000 feet below the surface the drill struck limestone rock, supposed to be the Austin chalk, which crops out northwest of San Antonio. At 1,310 feet a small seep of sulphur water was struck in this rock. This was the first water found in the well. The drill passed through the upper limestone at 1,380 feet and into a bed of yellow joint clay, resting on about 10 feet of shells, together 45 feet. Immediately below this, at 1,425 feet, it struck limestone, supposed to be the Caprina (Edwards) limestone. At 1,535 feet a small overflow of water was struck. When the second contract was completed, at 1,758 feet, there was an overflow equal to about 18 gallons per minute. At this depth, having completed his second contract on the well, the contractor declined to make another contract for depth, and was employed thereafter by the day. From 1,625 to 1,758 feet, Colonel Terrell had noticed an increase of temperature of the water, equal to 1° for each 25 feet of descent. This rapid increase led him to believe that the strata were porous for at least 100 feet below the then bottom of the well, and that the water was working up from below. If his reasoning was correct in this, then there would be no difficulty in drilling deeper with satisfactory daily progress, and the result demonstrated the correctness of his view. Rapid progress was made—142 feet in nine days—with a less increase of heat for corresponding additional depth.

It was intended to sink the well to 2,000 feet, but the contractor quit at 1,900 feet. At intervals of about 70 to 75 feet below the first water that overflowed, and after passing a very hard stratum in the rock, the drilling became easier, and there was an increase of the overflow, each of such sections giving an increase over the preceding. The greatest increase, however, was below the depth of 1,885 feet.

This well must have been very nearly through the Edwards limestone at the time when the drilling ceased. It is probable that the flow and pressure would have continued to increase if the well had been sunk deeper.

The flow from the well seems to have slightly increased, notwithstanding two years of drought since it was completed, and the temperature has gone up from 103.8° to 106° F. Some claim to have found it 108° and 109°. This increase in temperature may be accounted for in part by heat being abstracted, for some time after the well was completed, to warm up the 1,900-foot channel.

The estimate of flow, 800,000 gallons in twenty-four hours, was made by the contractor. Colonel Terrell's own estimate is slightly less. Other wells near this city having less minimum diameter of bore and less heat are estimated at 1,200,000 gallons in twenty-four hours, from which fact Colonel Terrell concludes that there is a tendency to overestimate the flow from artesian wells. Colonel Terrell thinks that the flow of water is slightly influenced by atmospheric pressure, a low barometer causing an increased flow.

An analysis of the water was made some time after the well was completed, and is given on page 301. It will be seen from the analysis that the water contains no chloride of sodium and no organic matter. The carbon dioxide is equal to 34 cubic inches of carbonic-acid gas to the gallon, and the hydrogen sulphide to 8 inches of sulphureted-hydrogen gas to 1 gallon of the water.

When the surface flow in the acequia is obstructed, a floating scum accumulates. This scum has not been analyzed. The acequia terminates in a lake 2,200 feet long on the tract of land. The lake is stocked with rock and black bass, striped perch, cat, minnows, turtles, etc. Minnows have been seen in the acequia within 100 feet of the well, where the water has nearly its maximum heat.

The water has been used for two years for irrigation. A small orchard, 1,700 fruit trees, depends almost entirely on this water, and the trees are doing well. Garden vegetables, corn, cotton, oats, and grass seem to do well when irrigated with the water. Stock drink the water freely—for months at a time having no other—and it is said to free them from ticks. In one case, when given for the purpose, it entirely freed a horse from bots and other intestinal worms.

Many sick people have visited the well to use the water, both for drinking and for bathing. Afflicted persons who have tried it claim that by drinking and bathing in the water they have been completely cured of many diseases.

No commercial use has been made of the water, but it is daily used at the well and carried away for use, by and for sick people, without charge.

We have plotted upon one plate all the well data which we have been able to procure relating to the vicinity of San Antonio. (See Pl. XLIII.) Many of these records are meager, giving only the depth at which water was found, unaccompanied by any description of the rock material passed through. The Terrell and Collins wells, however, may be considered the best available standards of reference, to which the others may be compared. We have referred all these sections to a geologic datum plane, which approximately corresponds with the top of the Shoal Creek limestone. Upon examination of these sections it will be seen that there is a notable absence of sandstone strata, and that the water occurs in limestones and marl. Furthermore, there are two persistent occurrences of water in the general geologic section as far as explored, accompanied probably by one or two minor horizons. The uppermost of these consists of a stratum of sulphur water, which is struck in many wells in a geologic horizon near the base of the Austin chalk, possibly in the Eagle Ford shales. This horizon is accompanied by gas and oil. These impurities are such as one familiar with these rocks would naturally expect to find in them, owing to the bituminous character of the shales and the excessive accompaniment of iron pyrites, which no doubt causes the element of sulphur to enter into the water. It is very probable that there are more than one of these horizons in the lower part of the Gulf series, as will be seen in some of the sections given.

The "sweet water," as it is locally called, however, lies about 200 feet lower than that at the base of the Austin chalk, and undoubtedly occurs in the upper part of the Edwards limestone, or about 100 feet below the bottom of the blue-black clay of the Del Rio beds.[1]

In Terrell's well the pressure and volume of this water increased as the drill was pushed for a distance of over 350 feet below it, being strongest at a depth of 1,850 feet. It is very probable that the drilling was stopped just as the Glen Rose beds of the Trinity division were being encountered, and we possess no record of wells drilled into the basement sands of the Trinity division, which supply some of the Austin and Waco wells. We have heard rumors that certain wells have been drilled to a depth of 3,000 feet in the city of San Antonio and failed to obtain water from the base sands, but we have no positive facts bearing upon this subject.

From our knowledge and observation of the thickness of the strata in adjacent regions, we estimate that the basement sands of the Comanche lie from 1,000 to 1,300 feet below the sweet-water flow at San Antonio. We can not predict whether these will prove water-bearing or not.

These records, together with the depths given in other cases, all show that the lower or "sweet" water is obtained in the foregoing wells in the city at depths varying with the dip, from 500 feet in the extreme northwest to 880 feet at Alamo Plaza.

In general, the wells deepen toward the south and east, as is illustrated by the depths of the wells belonging to the San Antonio waterworks and those at the State Lunatic Asylum and at Terrell's.

The records in the northwestern part of the city show that the water

[1] The exact thickness of the Fort Worth limestone at San Antonio is not known. Estimating it at 50 feet, the water in the Edwards limestone (the San Antonio flow) is about 50 feet below the top of the formation, corresponding with the Austin, Manor, and San Marcos records.

was struck at about 350 feet; the wells to the east, about one-half mile below the head of the springs, are 630 feet deep; 1 mile down the river they are about 780 feet deep, and the five wells in the city, about 1½ miles still lower down, are about 880 feet deep. These records show a progressive deepening of the wells of 250 feet in 3 miles, or about 83 feet per mile. Colonel Terrell's well, the exact distance of which from the city wells can not be ascertained, but is about 4 or 5 miles, did not secure flowing water until over 1,500 feet had been drilled.

This deepening of the water-bearing rocks is in accord with the geologic conditions previously explained. The limitations of the available belt at San Antonio have not been fully ascertained. We know that Cable's ranch well, 8 miles north of west of San Antonio, is probably west of the line of success, an unsuccessful well having been drilled there to a depth of 2,000 feet. This locality is probably west of the fault line, which runs from Helotes northeast, via Leon Springs, to New Braunfels. West of this line no flowing wells are obtainable, except in the canyon valleys of the Edwards Plateau, as previously explained.

Colonel Terrell's well shows that the available water is deepening rapidly to the east and south. Mr. John Wickland, 12 miles east of San Antonio, drilled 700 or 800 feet through the blue clays of the lowermost Tertiary and uppermost Cretaceous. He would probably have to go at least 1,000 feet lower than at San Antonio to obtain the sweet water.

There are many resemblances between these wells and those at Austin and San Marcos. By comparison with the log of the old Capitol well, given on page 282, it will be seen that the San Antonio "sweet water" corresponds approximately in its depth and geologic position to the sulphur water obtained in the old Capitol well at 323 feet. The geologic horizons of the surface are slightly different, but the details of the section below the Austin chalk are similar. The water in both is obtained in the upper part of the Edwards limestone. The geologic position of the "sweet water" is very similar to the second water struck in the San Marcos well.

At the well of the Crystal Ice Company it is said that the "water is apparently an underground stream, the head of which is covered with a red or dark-yellow clay, mixed with decomposed clay." This description suggests the red clay and fossils so frequently found as residuum in the caverns of the Edwards limestone,[1] as can be seen west of Austin.

The occurrence of underground caverns and eyeless cave animals, and alleged underground streams, are also phenomena peculiar to the Edwards limestone of the Austin and San Marcos regions.

[1] The only limestones of the Cretaceous series which leave a bright-red clay residuum are found in the Edwards beds.

THE DEL RIO WELL.

From San Antonio west to the Rio Grande, we possess but few records of well experiments, and know of but one flowing well in that direction. A mile or two south of Del Rio, Valverde County, a well was bored to a depth of 460 feet which struck a flow of unpalatable sulphur water. This well was commenced at a horizon near the top of the Del Rio clays, and the water was obtained from one of the upper beds, probably the Edwards limestone.

Mr. John Wood, of Del Rio, put down a well in the town 760 feet in depth in which water was struck at about 565 feet. This contained alum, sulphur, and iron, and there is little doubt that it came from a mineral horizon of the Edwards beds. The water rose to within 60 feet of the surface. It is to be regretted that this well was not continued lower, for it is reasonably certain that the more potable and abundant waters which supply the adjacent San Felipe springs would have been encountered within the 1,000 feet or more of the Cretaceous strata underlying the present bottom of the well.

An unsuccessful experimental well was also drilled at Spofford Junction some years ago by the Southern Pacific Railway Company. This is further discussed on a succeeding page (p. 318).

CHEMICAL QUALITIES OF THE WATERS.

Each of the various water-bearing beds enumerated in the foregoing pages possesses peculiar chemical qualities. It is exceedingly difficult, for various reasons, to determine the properties of each particular bed. In wells penetrating one or more of these the water is invariably mixed, so that when it reaches the surface it is not representative of any single bed. In the second place, such analyses as have been made were mostly for individual parties and are inaccessible. No systematic comparison of the waters as a whole has been made. It would be an interesting experiment to collect, analyze, and compare these various waters with one another and with those of the spring rivers, and the writers hope that it will yet be done.

From our knowledge of the rocks through which these wells are drilled, we know that the different strata vary greatly in chemical character, some being comparatively free from all mineral ingredients except lime carbonate, which is nearly always present, while others carry various accessory impurities. The Austin chalk, miscalled magnesian limestone, contains no magnesia,[1] and hence the magnesian constituent of the water must originate below that formation. It does contain considerable pyrites, however, sufficient to supply the iron and sulphureted hydrogen derived from this horizon at San Antonio.

The Eagle Ford shales are somewhat similar to the Austin chalk, so

[1] See analyses in Third Ann. Rept. Geol. Survey Texas, pp. 351–352, 354.

far as accessory minerals are concerned, containing, however, more pyrites and bituminous and lignitic matter.

We have never been able to obtain a complete analysis of the Shoal Creek limestone, but there is reason to believe that water penetrating it would take up sulphur, iron, and saline impurities, such as can be seen incrusting it at Austin, where we have seen efflorescences of salt and magnesium sulphate. We have little doubt that any water transmitted through these beds will be strongly "mineralized."

The Del Rio clays are very impervious and not apt to affect underground waters seriously, unless they percolate through the pyritiferous fossiliferous beds, in which case much sulphureted hydrogen will be present.

The waters from the Fort Worth limestone and the upper part of the Edwards are highly impregnated with mineral matters, somewhat analogous to those of the higher Shoal Creek limestone and lower Glen Rose beds. The minerals form an efflorescence on the surfaces of the protected rock ledges, or in the bluff caverns, as can be seen in many places west of Austin.

The waters from the middle and lower parts of the Edwards beds are singularly free from any mineral accessories except magnesium and a trace of sodium, and this condition harmonizes with the composition of the San Antonio "sweet water."

In the upper Glen Rose beds, as can be seen in the bluffs of Mount Bonnel, west of Austin (Pl. XXVIII, p. 222), there are certain strata that contain strontium, magnesium, and sodium, which would materially affect the water. Fortunately we know of no wells in this province affected by them.

Another magnesium horizon occurs near the top of the Travis Peak beds, but below that these beds are very free from any unpleasant ingredients, as is attested by the analyses of the waters and certain of the beds.

The following table shows all the analyses we possess of these rocks. Many of the materials deleterious in water have not been determined, but the variation of some of them is clearly shown.

*Table showing chemical composition of some of the rocks of the Cretaceous formations.*

| | Taylor.[1] | Austin (1).[1] | Austin (2).[1] | Austin (3).[1] | Eagle Ford.[2] | Shoal Creek.[2] | Fort Worth (1).[2] | Fort Worth (2).[3] | Edwards (soil).[4] | Paluxy.[5] |
|---|---|---|---|---|---|---|---|---|---|---|
| Silica ($SiO_2$) | 45.02 | 5.94 | 10.32 | 11.31 | 23.22 | 2.75 | 6.11 | 4.60 | ...... | 98.08 |
| Alumina ($Al_2O_3$) | 16.17 | 1.41 | 5.41 | 5.78 | 1.62 | 1.19 | 1.16 | ........ | 9.21 | .23 |
| Ferric oxide ($Fe_2O_3$) | 4.78 | 1.31 | 1.15 | 1.72 | 1.82 | 1.12 | .44 | 2.60 | 5.76 | .54 |
| Lime (CaO) | 14.26 | 48.73 | 45.31 | 42.61 | 33.36 | 49.32 | 37.53 | 52.13 | 2.61 | .10 |
| Magnesia (MgO) | ........ | ........ | trace | ........ | ........ | ........ | trace | ........ | ...... | trace |
| Potash ($K_2O$) | .975 | .20 | .17 | .33 | ...... | ...... | ...... | trace | .11 | ...... |
| Soda ($Na_2O$) | 3.22 | 2.60 | 2.07 | 2.36 | ...... | ...... | ...... | trace | .13 | ...... |
| Carbon dioxide ($CO_2$) | 10.36 | 37.84 | 34.44 | 33.86 | 25.40 | 30.25 | 25.42 | 40.96 | .28 | ...... |
| Phosphoric oxide ($P_2O_5$) | .113 | .142 | .218 | .131 | ...... | ...... | ...... | trace | .19 | .16 |
| Sulphuric oxide ($SO_3$) | .97 | .42 | 1.04 | 1.13 | ...... | ...... | ...... | trace | 3.85 | .31 |
| Water ($H_2O$) | 4.36 | .82 | .51 | 1.27 | ...... | ...... | ...... | ...... | ...... | ...... |
| | 100.228 | 99.412 | 100.638 | 100.501 | ...... | ...... | ...... | 100.29 | ...... | 99.42 |

[1] Analyses made by G. H. Wooten, Third Ann. Rept. Geol. Survey Texas.

[2] Analyses, incomplete, made by Prof. P. S. Tilson.

[3] Analysis by L. Mangenat.

[4] Analysis of soil by P. S. Tilson, Trans. Texas Acad. Sci., Vol. I, No. 4, p. 45, 1895. The soil contains 62.97 per cent insoluble matter, 39 per cent soluble $SiO_2$, and 15.47 per cent organic and volatile matter, not given in the above table.

[5] Contains both soda and potash. This formation is not encountered south of the Leon River.

We present also, in the table on the next page, for what they are worth, a few analyses of the water from the various beds, such as we have been able to gather from miscellaneous sources. These indicate that there are four or five broad classes of water, varying in chemical impurities as much as in geologic occurrence.

The analyses are expressed in parts per million. Analysis 1 was made by Prof. W. A. Noyes and is stated in grains per liter. The other analyses are stated in grains per gallon. No. 2 is taken from a circular furnished by Colonel Terrell and bearing the signature of Professor Noyes. No. 4 was made by Prof. H. W. Harper, of the University of Texas. Nos. 5 and 6 are taken from Roessler's report, page 271, in "A report on the preliminary investigation to determine the proper location of artesian wells, etc." (Fifty-first Congress, first session, Ex. Doc. No. 222, 1890.) The name of the analyst is not given. All of the analyses have been recalculated and the mode of statement has been changed. The hydrogen is derived from the bicarbonates; $SO_4$ is the acid radical of the sulphates; $CO_3$, of the carbonates; $PO_4$, of the phosphates. In two of the analyses sodium and potassium sulphate were not stated separately, and in one sodium carbonate and bicarbonate were not differentiated.

*Analyses of waters from the various beds.*

| | Terrell's well, San Antonio. | State Insane Asylum, San Antonio. | Kampmann's well, San Antonio. | Spring, power house, Austin dam. | Waco well. | Bell Company well, Waco. |
|---|---|---|---|---|---|---|
| Sodium | 461.5 | 437.6 | 1.9 | 42.9 | 40.5 | 288.8 |
| Potassium | 33.2 | ........ | ........ | ........ | ........ | ........ |
| Magnesium | 115.1 | 194.7 | 123.3 | 16.25 | 4.1 | ........ |
| Calcium | 738.0 | 635.6 | 20.4 | 82.6 | 7.9 | 10.0 |
| Iron | .4 | 7.1 | ........ | ........ | ........ | ........ |
| Strontium | (*a*) | 5.5 | ........ | ........ | ........ | ........ |
| Lithium | (*a*) | trace | ........ | ........ | ........ | ........ |
| Hydrogen | .15 | .25 | ........ | ........ | ........ | ........ |
| $SO_4$ | 1,938.6 | 1,920.2 | 346.8 | 63.0 | ........ | 277.2 |
| $CO_3$ | 8.7 | 201.4 | 61.4 | 183.9 | 22.2 | 328.2 |
| Carbon dioxide ($CO_2$) | 276.0 | 54.3 | 453.6 | 48.7 | ........ | 210.3 |
| $PO_4$ | (*a*) | .28 | ........ | ........ | ........ | ........ |
| Chlorine | 941.5 | 850.3 | 106.8 | 60.1 | 62.5 | 55.6 |
| Hydrogen sulphide ($H_2S$) | 18.5 | 26.2 | ........ | ........ | ........ | ........ |
| Bromine | trace | 4.8 | 6.7 | ........ | ........ | ........ |
| Iodine | trace | 2.3 | ........ | ........ | ........ | ........ |
| Boron | trace | ........ | ........ | ........ | ........ | ........ |
| Alumina | 1.5 | .8 | ........ | ........ | trace | ........ |
| Silica | 22.9 | 8.0 | 45.2 | 34.9 | 17.7 | 12.8 |
| Iron sesquioxide ($Fe_2O_3$) | ........ | ........ | ........ | ........ | 2.4 | ........ |
| Albuminoid ammonia | ........ | ........ | ........ | .14 | ........ | ........ |
| Organic | ........ | 792.8 | ........ | ........ | ........ | ........ |
| Sodium sulphate ($Na_2SO_4$). Potassium sulphate ($K_2SO_4$) | ........ | 43.3 | ........ | ........ | 409.5 | ........ |
| Sodium carbonate ($Na_2CO_3$). Sodium bicarbonate ($NaHCO_3$). | ........ | ........ | ........ | ........ | 353.2 | ........ |
| Alumina and iron oxide | ........ | ........ | ........ | 7.0 | ........ | 2.5 |
| Total | 4,586.05 | 5,183.13 | 1,166.1 | 539.49 | 920.0 | 1,185.5 |

*a* Small amount.

We present here, also, the following sanitary analyses of some of the spring waters from near Austin and of the Colorado River water at the power house at the dam.

*Analyses of waters from Austin and vicinity.*

[Dr. Henry Winston Harper, University of Texas, analyst. Samples collected March 14, 1896. Parts in 1,000,000.]

| | Spring in power house. | Spring on beach half mile below power house. | Large spring south side of river (Barton's spring). | River water from pump in power house. |
|---|---|---|---|---|
| Total solids | 298.0[1] | 280.0 | 325.00 | 296.0 |
| Temporary hardness | 72.0 | 83.06 | 50.00 | 66.13 |
| Permanent hardness | 34.4 | 37.90 | 44.08 | 50.53 |
| Total hardness | 106.4 | 120.96 | 94.08 | 116.66 |
| Chlorine | 28.28 | 12.62 | 22.22 | 29.29 |
| Nitrogen as nitrates | 0.125 | 1.000 | 0.667 | 0.080 |
| $H_2S$ | none | none | none | none |
| Sulphides | none | none | none | none |
| Sulphites | none | none | none | none |

[1] Dr. Harper states that the total solids fluctuate from 260 to 300.

The similarity of results of analyses of water from the spring in the power house and of the river water from the pump in the power house is suggestive of a relationship between the sources of the two waters. This feature presents peculiar interest because the results differ very widely from the analyses made of these two waters during the spring of 1895, and the limit of error does not account for the variance. This feature is worthy of further study.

*Chemical analyses of waters from Austin.*

[Harry W. Clark, chemist in charge of the Lawrence Experiment Station of the Massachusetts State board of health, analyst. Samples collected March 21, 1896; examined March 30, 1896. Parts in 1,000,000.]

| | River water from pump in power house. | Spring in power house. | Spring on south side of river (Stern spring). |
|---|---|---|---|
| Turbidity | decided | very slight | none |
| Sediment | decided | very slight | very slight |
| Color | 3.1 | 1.0 | 0.7 |
| Total solids | 289.5 | 265.5 | 330.0 |
| Free ammonia | 0.108 | 0.002 | 0.0 |
| Albumenoid ammonia | 0.112 | 0.048 | 0.018 |
| Chlorine | 28.300 | 29.800 | 33.800 |
| Nitrogen as nitrates | 0.320 | 0.280 | 1.750 |
| Nitrogen as nitrites | 0.008 | 0.004 | 0.002 |
| Oxygen consumed | 1.400 | 0.600 | 0.300 |
| Hardness | 140 | 130 | 180 |

*Analysis showing saline constituents of the water from the artesian well located near San Jacinto and Fifth streets, Austin, Texas.*

[Dr. H. W. Harper, associate professor of chemistry, University of Texas, analyst.]

| | Parts per 1,000. | Grains per United States gallon. |
|---|---|---|
| Sodium chloride | 6.0343 | 351.9084 |
| Sodium bromide | traces | traces |
| Sodium sulphate | .6561 | 38.2652 |
| Sodium sulphide | .0211 | 1.2332 |
| Sodium hyposulphite | .1140 | 6.6486 |
| Sodium silicate | .1585 | 9.2469 |
| Sodium bicarbonate | 1.3089 | 66.9088 |
| Potassium sulphate | .7096 | 41.3824 |
| Calcium sulphate | .9540 | 55.6405 |
| Magnesium chloride | .4614 | 26.9087 |
| Ferrous carbonate | .0104 | .5086 |
| Total solid contents | 10.4283 | 598.6513 |

Sulphureted hydrogen gas, 6.597 cubic inches per United States gallon. Carbonic acid gas, 43.971 cubic inches per United States gallon. The gases were measured at 16 C., and 750 mm. barometric pressure. Temperature of water as it issues from the well, 21.2 C. Depth of well, 182.88 meters (600 feet).

From the analysis last given, it will be seen that this well yields a sulphur water resembling the waters of the Bluelick Springs, Nicholas, County, Kentucky, of Hanna Springs, Lampasas, Texas, and of the springs at Harrogate, Yorkshire, England. It can be profitably used medicinally in such cases as would be benefited by the waters from the sources named.

With the above statements and analyses in mind, we can now briefly consider the character of the water of the various horizons reported.

We know from the well records that at San Antonio certain ferruginated and sulphureted waters are met with in the Austin chalk and Eagle Ford clays. This water is also accompanied by or is closely adjacent to oil and gas, and is usually piped off so as not to contaminate the purer water below.

We possess no data proving that water has been obtained from the mineral-charged Del Rio and Fort Worth beds, although we have suspected that in Kinney County several sulphur waters were derived from these horizons.

Concerning the Edwards limestone, however, we feel justified in speaking with more positiveness. The upper strata are strongly impregnated with mineral ingredients which the waters take up in passing through them, as shown by wells at Austin, Manor, San Marcos, San Antonio, and possibly Del Rio.

The vein of mineral water struck at the old Capitol well in the Edwards beds at a depth of 325 feet from the surface, according to Dr. B. F. Shumard, had a saline taste, and was strongly impregnated with sulphureted hydrogen gas. A qualitative chemical analysis of this water by Prof. W. P. Riddell, chemist of the Texas geological survey, showed it to contain the following constituents, named in the order of their relative abundance:

Sulphureted hydrogen.
Sodium chloride.
Calcium bicarbonate.
Calcium sulphate.
Sodium sulphate.
Aluminium and potassium sulphate.
Magnesium sulphate.
Iron sulphate (a trace).

A quantity of the Edwards limestone water from the Manor well was sent to the chemical department of the State university at Austin for analysis, but a detailed report had not been received at last advices. Dr. Stromberg made the following qualitative analysis of the water:

Sulphureted hydrogen, in excess.
Calcium sulphate.
Magnesium sulphate.
Sodium sulphate.
Potassium carbonate.
Sodium chloride.
Magnesium chloride.
Calcium chloride.
Iron carbonate.

Dr. Stromberg says that the above ingredients correctly proportioned would represent approximately an analysis of the water.

The following partial analysis showing the chief ingredients of the Manor well has been kindly furnished us by Prof. J. C. Nagle, of the State agricultural and mechanical college:

Sulphureted hydrogen.
Aluminium sulphate.
Calcium sulphate.
Strontium chloride.
Magnesium chloride.
Lithium chloride.
Sodium chloride.

This water is strong in salt, magnesia, and gypsum, while the strontium and lithium are quite high.

A small flow of sulphur water corresponding to the foregoing was also struck in the San Marcos well at a depth of 162 feet, or 32 feet below the estimated top of the Edwards beds.

The mineral water of Terrell's well at San Antonio, according to the record, also appears to be from this horizon.

The highly impregnated mineral waters of the Del Rio well may also come from the same upper horizon of the Edwards. We have records of several strong sulphur wells (nonartesian) in Kinney County apparently from the same source.

A variation is noticeable in all these analyses, which, notwithstanding their imperfection, are sufficient to show that the waters of approximately the same horizon are chemically dissimilar in different localities. For instance, the strontium of the Manor well has not elsewhere been reported from the Edwards beds. Furthermore, in the vicinity of San Antonio the waters vary greatly in different places.

The purer or "sweet water" of the San Antonio wells probably comes from a horizon only a short distance below the mineral horizon of Terrell's well. We have been unable to procure an analysis of this water, although we suspect that the analysis of the Kampmann well given in the table represents these potable Edwards limestone waters. This has magnesium and lime as its chief ingredients—the normal material of the Edwards limestone. The record of the San Marcos well also indicates that two classes of water may be obtained from the Edwards limestone, one of which is soft and potable while the other is highly charged with sulphur and various chemical ingredients.

The minerally impregnated beds of the Glen Rose, containing epsomite and other deleterious substances, have not been encountered in the San Antonio wells, although we strongly suspect that they have contaminated certain of the Austin wells. In the still lower Travis Peak formation the waters are softer, more potable, and freer from sulphureted hydrogen, containing mostly bicarbonates in solution. The waters of the latter formation have been utilized, so far as we know, only in the State Insane Asylum and St. Edward's College wells at Austin. These are so much purer and better than the overlying "mineral" waters that one wonders why all the other wells of that city have not been driven down to them.

From the various analyses given it will be seen that the chief mineral impurities of the mineral waters derived from the Edwards beds are chlorides and sulphates, calcium, magnesium, and sodium. In this they differ greatly from the potable water of the lower Trinity (Waco) beds, in which the chief ingredients are bicarbonates and sulphates of sodium, while the total solids in the former are nearly five times as great as those in the latter.

The fact that the waters of the fissure springs appear to have the properties of the lower Edwards and Travis Peak artesian waters leads us to accept the hypothesis that the latter waters lie below much of the region in which the fissure springs occur, west of San Antonio, and have not yet been penetrated by the artesian drillings.

From these studies of the chemical relations of the rocks and waters we think some useful deductions can be made. Bad waters can and should be cased off wherever encountered, and the well continued until a purer flow is struck or the Cretaceous system entirely passed. This is done in many instances, such as the principal wells of San Antonio. In our opinion, no matter how firm the drill hole, it would pay in all instances to drill wells to the lower waters and case up the well above them in order to insure against the seepage of these mineral waters. In the light of the facts set forth in this report it will hardly be excusable hereafter to allow such waters to flow as those issuing from the State Capitol well at Austin, the State Insane Asylum well at San Antonio, and many other similar wells.

### FISSURE SPRINGS.

At intervals along the interior boundary of the Rio Grande Plain from Austin to Devils River, through a distance of 300 miles, there is a series of remarkable springs which rise out of the ground. They do not break out from bluffs or fall in cascades, but appear as extensive pools, often in the level prairie. These pools or small lakes of limpid blue water find their outlet in swift and silently flowing streams.

The pools are carpeted with rare water plants, among which many fishes may be seen swimming. So transparent are these waters that objects 15 to 20 feet below the surface appear to be only a few feet away. They have been filtered by passing through the pores of the rocks for many miles.

The most conspicuous of these springs are near Del Rio, Brackett, San Antonio, New Braunfels, San Marcos, Manchaca, and Austin. In addition to these there are large springs north of the Colorado at Round Rock, Georgetown, Salado, Belton, and other places, as well as numerous small springs which need not be mentioned. The Cedar springs, north of Dallas, are probably the most northern of the line.

The following paragraphs describe the most important of those which fall within the province of this paper:

Several groups of springs break out in the vicinity of Austin along the line of the secondary faults accompanying the great fault zone which extends approximately north and south through the east foot of Mount Bonnel. The principal are Mount Bonnel and Taylor springs, east of the foot of Mount Bonnel; Sand springs, between the dam and the city; Bee springs, Barton springs, and several unnamed springs breaking out at river level beneath Deep Eddy Bluff, west of the river.

Sieder's spring, on Shoal Creek, in the northwest part of the city, also belongs to this category. Although small in volume, it is educationally instructive, because its relation to the faulting is clearly visible. The other springs are of larger volume.

The Mount Bonnel and Taylor springs are now covered by the back water of the lake. Barton springs occur in and on each side of Barton Creek, about one-quarter mile above its confluence with the river, and give forth large volumes of water. The head lake of these springs is partially shown in Pl. XLIV. They have a discharge of 25 second-feet, or about 16,000,000 gallons per day. The chief springs in the bed of the creek can be seen welling up out of the clean-cut fissure of a fault line just above the old dam site. A mill was until recently run by the water, but the power is now unutilized. These springs are beautifully situated and are the favorite resort of the people of Austin; they are surrounded by fine groves of pecan and picturesque rocks. Their aggregate volume must reach thousands of gallons per minute. The related faulting of the strata is shown particularly well in the vicinity of these springs.

Almost due north of Barton springs, beneath the highest bluff of the river at Deep Eddy, along other fracture lines, there is another group of fissure springs, but, owing to the fact that they are at the base of a high bluff and accessible only by boat and at the low-water level of the Colorado, few people have seen them. They discharge a large volume, but as they break out in the river's edge it is impossible to gage them. The aggregate flow of the springs near Austin is so great that the volume of the river is materially increased.

Manchaca springs, about 13 miles south of Austin, on the old San Antonio road, burst out of a fissure in the Austin chalk, and the run-off

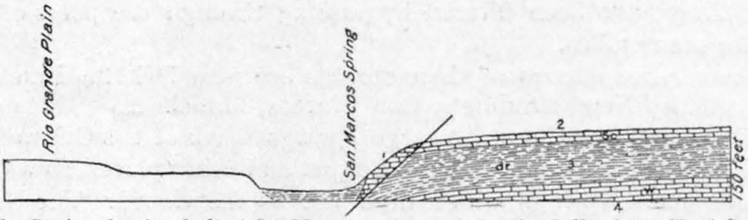
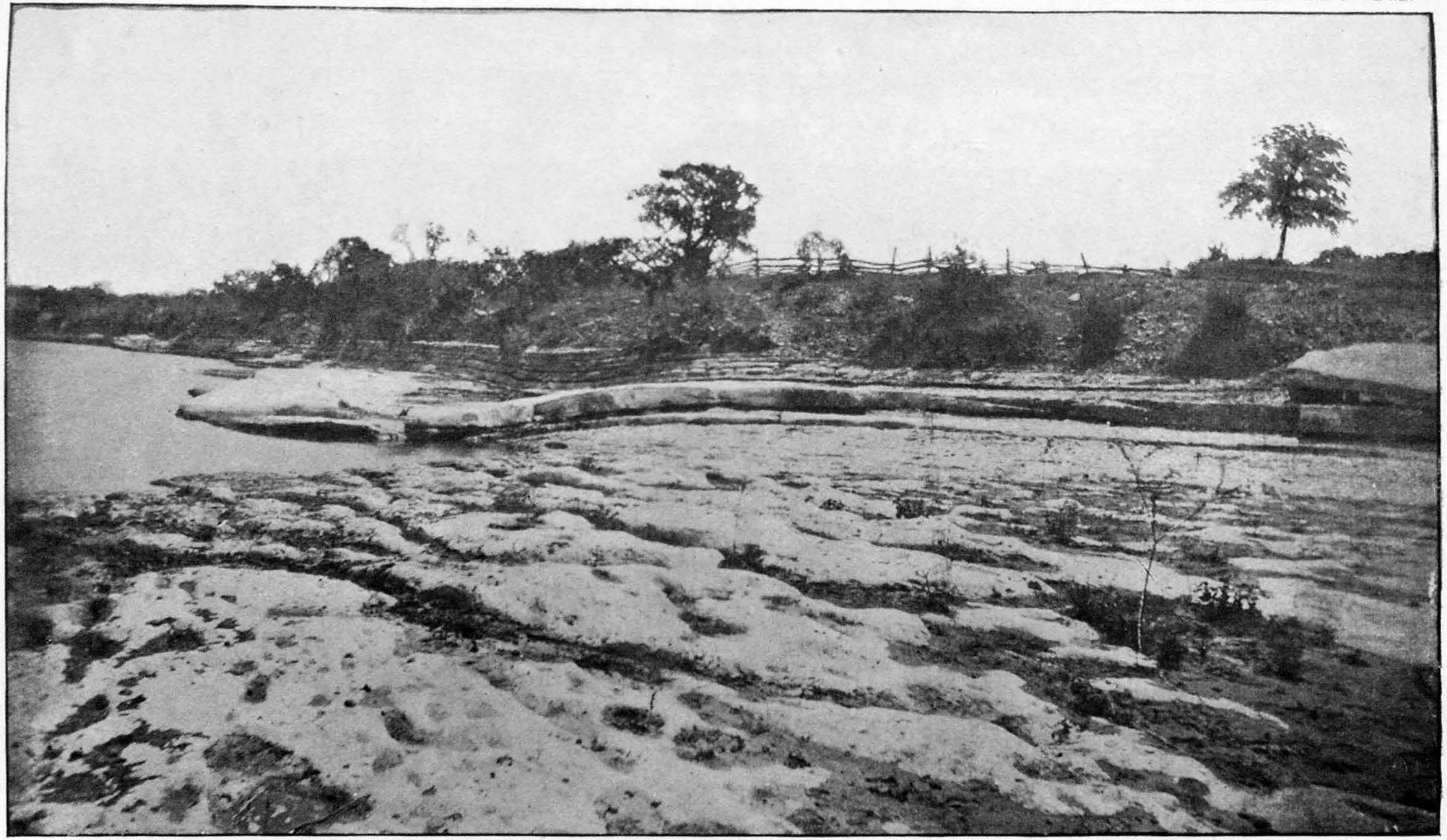



FIG. 72.—Section showing fault at San Marcos springs. 1, Austin chalk; 2 (*sc*) Shoal Creek limestone; 3 (*dr*) Del Rio clays; 4 (*w*) Fort Worth limestone. The stream bed is filled with alluvium.

finds its way into Onion Creek. The flow is large, but less than that of Barton springs.

Southward the next conspicuous springs of this character are in the Blanco River above Kyle. At the village of San Marcos a great group of springs breaks out at the foot of a north-south line of bluffs making the Balcones scarp line in this region, and form the source of a beautiful river flowing 57,000,000 gallons per day. This has long been a famous resort in Texas on account of the exquisite aqueous flora and the beauty of the water. The springs form a lake nearly half a mile long, and its run-off forms the San Marcos River. At the lower end of the lake a mill and an ice factory are run by its water, and the United States Fish Commission has established a culture station here. While none of the water is utilized in irrigation, there is no reason why such a volume of water should not be used to irrigate considerable areas of the fertile Black Prairie lands.

The springs of the Guadalupe and Comal rivers near New Braunfels also belong to this class. Those of the latter stream have a flow of 328 second-feet, or about 200,000,000 gallons per day, and are the



BARTON CREEK, ABOVE DAM, AUSTIN, TEXAS

Shows head of Barton Spring, alternation of dry stream bed and water holes, a low monoclinal fold, and the Edwards limestone and terrace deposits.

largest of the whole group. The water is utilized to run a mill, and could irrigate many thousands of acres. Dr. Evermann has published the following notes on these springs:[1]

Like most of the streams of this part of Texas, the Guadalupe is fed chiefly by numerous springs issuing from the Cretaceous limestone along its course. As a consequence, the water is usually exceedingly clear. During heavy rains, when there is considerable direct surface drainage, these streams, of course, become muddy for a short time. The temperature of the water at 4 p. m., December 3, was 68°, the air being 58°.

*The Comal springs.*—There are a great many springs in the vicinity of New Braunfels, the principal group being known as the Comal springs. There are several springs in this group situated upon the land of Mr. Joseph Landa, a little over a mile northwest of New Braunfels. The largest of these flows, perhaps, as much as 50,000 gallons per minute, and is certainly a magnificent spring. The other springs of the same group flow at least as much more.

The main spring comes out near the foot of a limestone hill, and after running rapidly for a short distance over a pebbly bottom and in a narrow channel, it widens out into quite a pond with mud bottom and filled with vegetation. This pond also receives the water from numerous other springs, and has its outlet in Comal Creek (or the Rio Comal), which, after a course of 2 or 3 miles, joins the Guadalupe River. The water of these springs is, of course, very clear. The temperature is 75°.

About 2 miles north of the town is another group of springs, smaller than that just described. The amount of water is abundant, however. . . .

Many of the large springs, and some of the most noted of their class, occur in the vicinity of San Antonio, the largest being at the head of the San Antonio River, a few miles north of the city. Until recently these flowed out of the ground in great volume—27,000,000 gallons per day—forming an exquisite lake, the run-off of which is the San Antonio River, which flows through the heart of the city of San Antonio and supplies it with water.

Below this group of springs and upon the banks of its outflow was situated one of the most ancient Indian settlements, or pueblos, of Texas. The early Spanish priests, appreciating the beauties and natural advantages of the place, located several missions there within a short distance of one another. The natives were employed in the cultivation of farms and gardens irrigated by the spring waters. The ancient acequias or ditches, followed by the older streets, shape the present outline of the city.

The spring-fed river furnished, until recently, water for the city of 48,000 inhabitants without very appreciably diminishing its volume. Many acres of gardens and farms were irrigated, and there was sufficient water to irrigate many more. As elsewhere shown, the flow of the river has been recently seriously diminished by the drilling of numerous wells around San Antonio.

The San Pedro springs are about 2 miles southwest of those above mentioned, at the head of the river. Besides supplying an irrigation ditch they constitute the nucleus of handsome pleasure grounds. The

[1] Bull. U. S. Fish Commission for 1891, pp. 72-73.

springs here break out of fissures in the *Gryphæa aucella* beds of the Austin chalk. Their flow is estimated at 9 second-feet, or 6,000,000 gallons per day.

The Los Moras springs, at Fort Clark, 125 miles west of San Antonio,

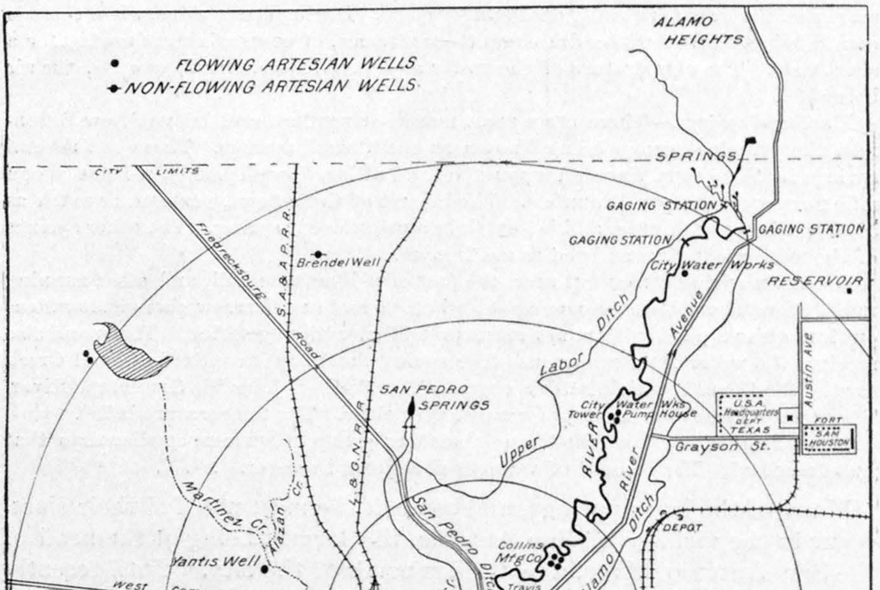
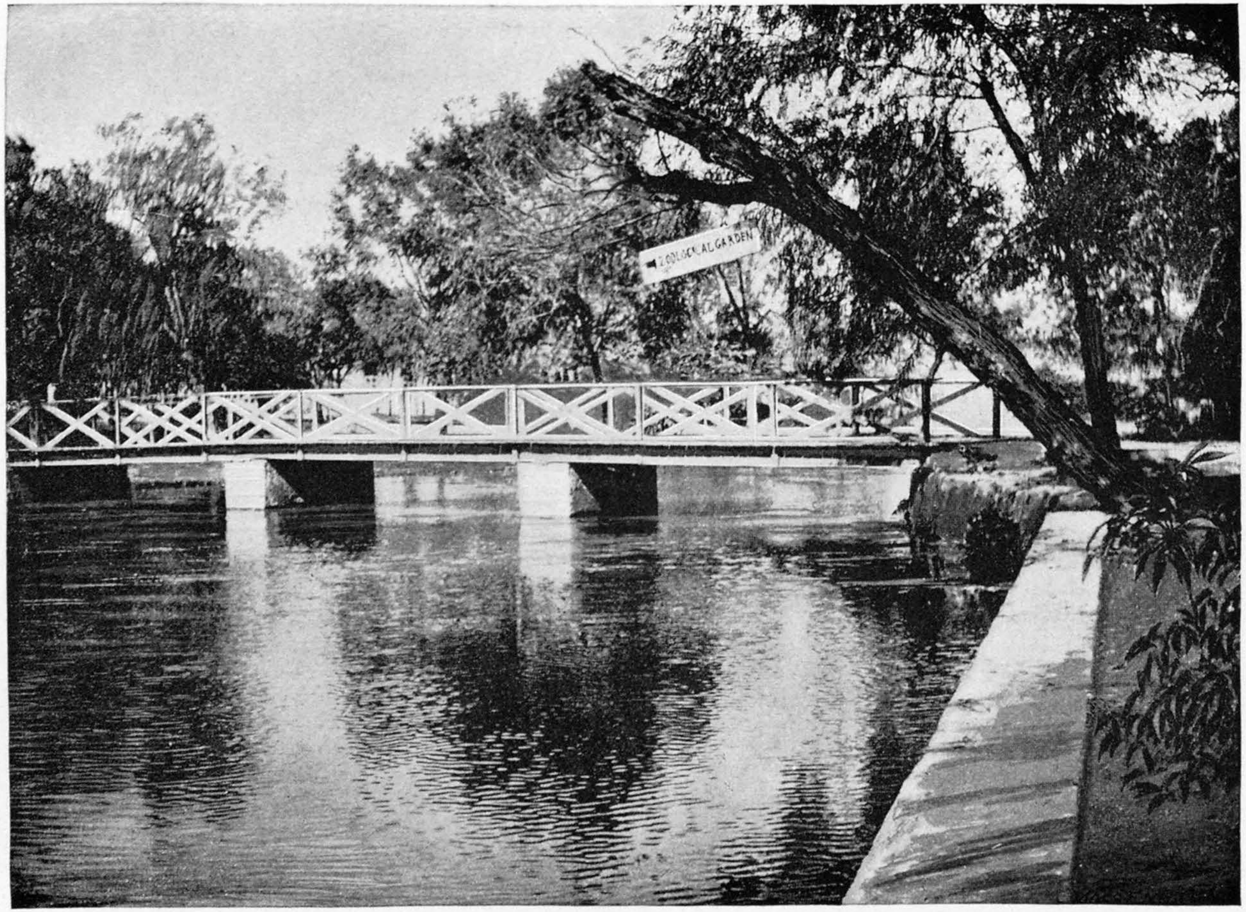



FIG. 73.—Map showing the localities of artesian wells and springs in the vicinity of San Antonio.

are another instance of this type of waters. They break out from the Eagle Ford shales and form extensive headwater pools, around which are built the military post of Fort Clark and the county town of Brackett, both of which are supplied with water from it. The run-off forms a




RUN-OFF OF SAN PEDRO SPRINGS, SAN ANTONIO.

beautiful, limpid stream, which is largely used for irrigation a few miles below Fort Clark. Mr. Babb has measured the flow of these springs and estimates it to be 21 second-feet, or over 13,000,000 gallons per day. About 10 miles north of Los Moras springs are the springs of the Pinto, which probably belong to the same class. They are much smaller than most of those herein enumerated.

The westernmost of the line of fault springs are the San Felipe springs near Del Rio. They break out at the edge of the Edwards Plateau, 2 miles northeast of Del Rio and about 5 miles from the Rio Grande. The pool is almost as large as that at the head of the San Antonio River. From the deep-seated rock at its bottom the water can be seen welling up in a great column, and has the same peculiar greenish-blue color as that of the other streams of this class. No trees surround it; it is alone—a fountain in the desert. The rocks from which it bursts—the Fort Worth limestones—have the same kind of joints and faults as are found at San Antonio and Austin. The outflow from the pool forms a bold, rushing stream that runs off to the Rio Grande, some 5 miles distant. This spring stream, in addition to running a mill and supplying the village with water, is partially utilized to supply 15 miles of irrigation ditch and to irrigate 5,000 acres, and can furnish water for the irrigation of several thousand acres more. Mr. Babb's measurements make a total discharge of 19 second-feet, or about 12,000,000 gallons per day.

The flow of these various springs has never been measured through a period of time sufficiently extended to give their variation. At the writers' request, the Division of Hydrography of this Survey, in 1895, made measurements of the principal spring rivers, and the results as tabulated were as follows:[1]

*Discharge of the various spring rivers.*

| Date. | Stream. | Discharge in second-feet. | Discharge in gallons in 24 hours. |
|---|---|---|---|
| 1895. | *Fissure springs.* | | |
| Dec. 18 | Barton springs | 25.0 | 16,157,921 |
| Dec. 18 | Dam spring, Austin | 4.3 | 2,800,000 |
| Dec. 19 | San Marcos | 89.0 | 57,522,200 |
| Dec. 20 | Comal | 328.0 | 211,981,932 |
| Dec. 21 | San Antonio | 42.0 | 27,145,308 |
| Dec. 21 | San Pedro (at San Antonio) | 9.0 | 5,816,852 |
| Dec. 24 | Los Moras spring | 21.0 | 13,572,653 |
| Dec. 24 | Del Rio Ditch | 19.0 | 12,280,021 |
| Dec. 24 | San Felipe spring | 80.0 | 51,705,350 |
| | Guadalupe river | 48.0 | 31,023,210 |

[1] Bull. U. S. Geol. Survey No. 140, 1895, p. 86. Measurements made by Mr. C. C. Babb.

*Discharge of the various spring rivers*—Continued.

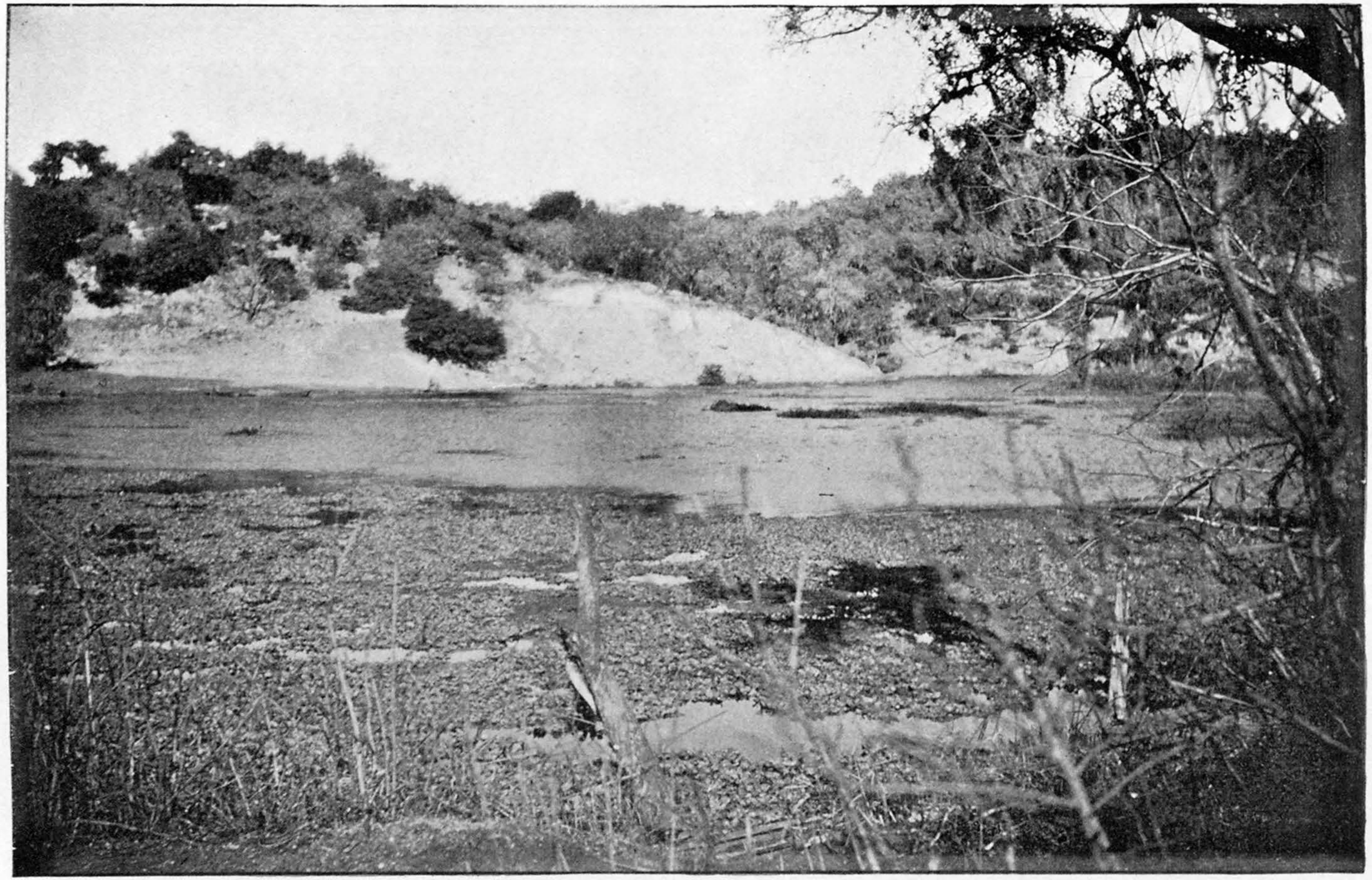
| Date. | Stream. | Discharge in second-feet. | Discharge in gallons in 24 hours. |
|---|---|---|---|
| 1895. | *Gravity springs.* | | |
| Dec. 23 | Sabinal | 0.6 | 387,790 |
| Dec. 23 | Leona | 11.0 | 7,109,486 |
| | *Fissure springs not measured.* | | |
| | Deep Eddy, Austin | | |
| | Sand spring, Austin | | |
| | Manchaca | | |

It is a significant fact that the trend or line of these great springs along the northern margin of the plain coincides almost exactly with that of the Balcones fault line. A study of the rocks in the vicinity of the springs has shown that the springs are associated with the system of joints and fractures accompanying the fault line, and that their waters ascend to the surface along these fissures. In other words, these waters come from the deep-seated rocks, and are forced to the surface by hydrostatic pressure. Hence they are artesian in nature and constitute natural artesian wells.

PROBABLE IDENTITY OF SOURCE OF ARTESIAN AND FISSURE-SPRING WATERS.

From the similarity of color, taste, temperature, etc., of this chain of springs extending in a continuous line 250 miles between Austin and Del Rio, and from their association with the line of Balcones scarp and faults, there can be no doubt that they are all of similar nature and origin. The temperature of the water brought by the springs from their subterranean source, about 75°, does not differ greatly from the mean annual temperature of the air in this part of Texas—68° to 69°; and as the normal downward temperature change requires only 50 or 60 feet of depth for one degree of temperature, the formation from which the water comes can not be many hundred feet below the surface. The great volume of the springs shows their chief source to be a formation transmitting water freely. Their freedom from sulphureted hydrogen and other ingredients that would be detected by taste or smell excludes from consideration the higher water horizons of the Edwards limestone as sampled by artesian wells. These various facts leave no reasonable doubt that their water is derived from either the "sweet water" horizon of the Edwards formation or the Travis Peak sands; that is, they have the same source as the purer waters of the artesian wells.

The fact that the flow from the springs is slow in showing sympathetic variation with drought or rainfall is evidence that the reservori supplying them is of vast extent.



SAN MARCOS SPRING.

The identity of the source of the fault springs and the artesian wells is further confirmed by their relationship at San Antonio. Upon information furnished by Col. G. W. Brackenridge, who owns the city waterworks, which are supplied by both the spring at the head of the San Antonio River and a large number of auxiliary artesian wells, we learn that when the wells are allowed to flow the springs diminish in volume and the river is greatly lowered, being at times almost completely emptied. On the other hand, when the wells are stopped by valves the springs furnish their usual flow volume.

It is also an interesting fact that although the waters of the fissure springs of the Rio Grande Plain and the gravity springs of the Edwards Plateau, respectively, differ in mode of outflow, they are both derived from the same geologic horizons—the rock sheets of the Edwards beds and Trinity division. The difference in mode of outburst is due to the difference in structural arrangement which these horizons present in the regions of their occurrence.

## SOURCE OF THE UNDERGROUND WATERS OF RIO GRANDE PLAIN.

The question naturally arises in the minds of all who reflect upon the phenomena we have described, What is the origin and source of the underground waters of the Edwards Plateau and Rio Grande Plain? The customary explanation is that the waters supplying the artesian wells and spring rivers come from the distant Rocky Mountains. This is impossible, because the continuity of the strata between these rivers and the mountains is completely severed by the drainage valley of the Pecos, the strata being eroded away over thousands of square miles. The real source of the water is the rainfall of the Plateau of the Plains and its adjacent borders, as will now be explained.

Let us first consider the mode of catchment. Much of the rain water is caught directly upon the edges of the Glen Rose and lower beds which outcrop along the western and northern summits, breaks, and margins of the Plateau of the Plains at an elevation higher than that of their embedded continuation along its eastern and southern margin. These outcrops on the higher surface and slopes of the western part of the plateau, between the New Mexican line and the Pecos at the thirty-first parallel, are between 3,000 and 5,000 feet high, while a part of the same strata along the eastern and southern margin of the plateau are less than 1,000 feet above sea level. Gradual as is the dip from Castle Mountain, on the Pecos, to San Antonio, it would not be sufficient to embed the waters at the latter locality below the altitude of their outcrop were it not for the sudden faulting. No doubt some water also enters the basement beds along that portion of the eastern margin of the Plateau of the Plains which constitutes the northern border of the Edwards Plateau. The outcrop of the water-bearing basement beds in this escarpment along the south breaks of the Concho have an altitude of between 1,500 and 2,000 feet, exceeding in height the embedded strata at the southern and eastern margin of the plain at Austin by about that much.

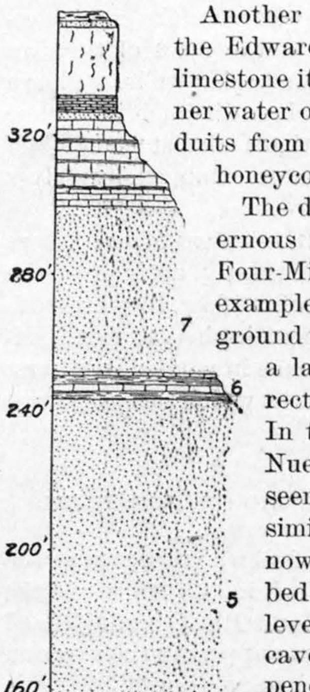



FIG. 74.—Composite section of the Gillespie formation near Fredericksburg, showing the water-receiving beds. 1, 2, Paleozoic; 3-7, Gillespie sands; above 7, same as 2-7 of fig. 55 (p. 221).

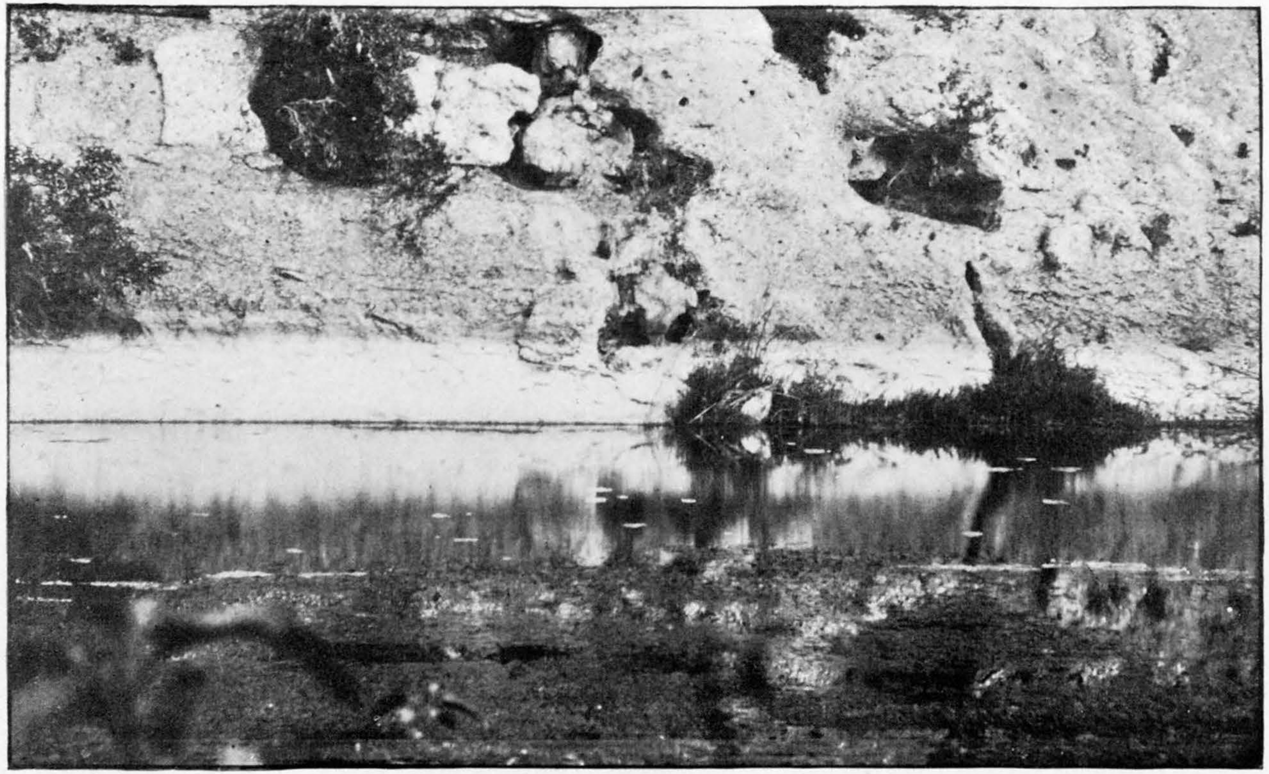
Another large part of the rainfall on the surface of the Edwards Plateau percolates downward through the limestone itself to the water-bearing beds. In this manner water often reaches the embed through intricate conduits from the surface, such as fissures and caves and honeycombed spaces in the limestone strata. (Fig. 75.)

The disappearance of the Nueces River into a cavernous fissure at the lower end of the Kickapoo or Four-Mile water hole, as described on page 269, is an example of the direct transmission of water underground through the aid of caverns and fissures. Here a large volume of water can be seen passing directly into the "bowels of the earth," so to speak. In the adjacent massive limestone bluffs of the Nueces, vertical sections of old caverns can be seen, as shown in Pl. XLVIII. These are no doubt similar in character to those into which the stream now disappears, and at a time before the stream bed had been lowered by erosion to its present level the water disappeared down them. These caves are probably of the same nature as those penetrated in the San Marcos and San Antonio drill holes. The presence of peculiar cave animal life in the wells of these places also demonstrates that some of the underground water comes through cavernous passages.

We do not possess sufficient data to estimate how much water the embed of the Edwards limestone receives throughout the vast extent of the thousands of square miles constituting the Edwards Plateau.

Such are the conditions explaining the catchment and transmission of water in the embedded rocks of the plateau. These conditions are entirely different in the Rio Grande Plain, where the water-bearing beds have no outcrop open to the rain, and where the jointed limestones of the Edwards formation lie beneath impervious clays and shales.

The embedded Edwards limestones of the Rio Grande Plain are charged by the peculiar mechanical arrangement produced by the Balcones system of faulting. The dislocation or throw of the fault or faults breaks or disconnects the continuity of the water-bearing strata. This



NATURAL SECTION OF OLD WATER CHANNELS IN LIMESTONE BLUFF BELOW KICKAPOO SPRINGS

View near lower end of Kickapoo Water Hole, Edwards County, showing cavernous passages in massive limestone analogous to those down which the stream now disappears 100 feet below the point where this view was taken

does not relieve the hydrostatic pressure, and the water is powerfully urged either to push its way to the surface through the fracture or to enter the stratum which is brought into juxtaposition with it and con-

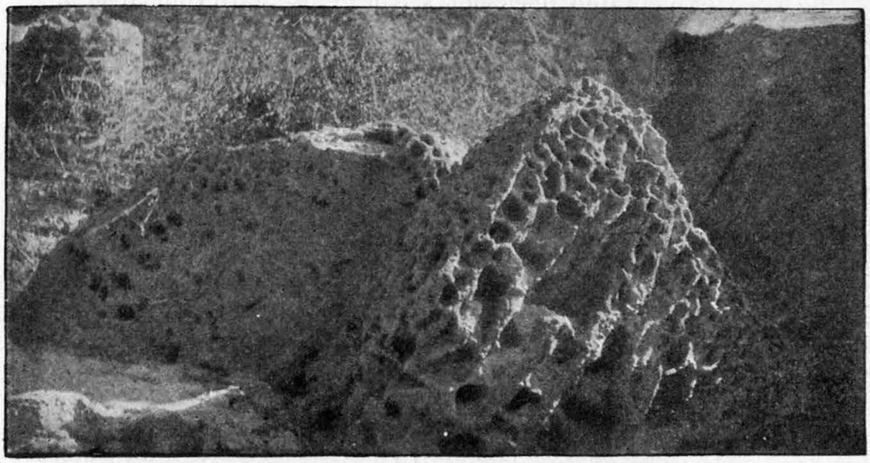


FIG. 75.—Honeycombed strata, Edwards limestone.

tinue beneath the ground. The amount of throw of the fault doubtless varies greatly, but there must be many places where the severed edges of the porous limestones of the Edwards formation on the sea-

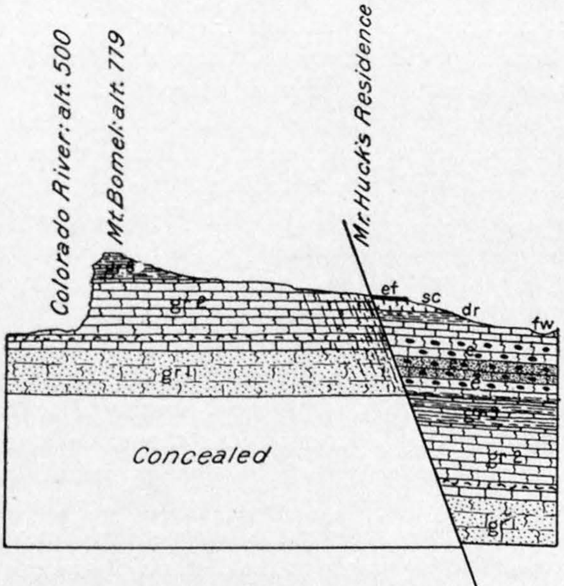



FIG. 76.—Displacement at Mount Bonnel, Travis County. *ef*, Eagle Ford; *sc*, Shoal Creek; *dr*, Del Rio; *fw*, Fort Worth; *e*, Edwards; *gr*, Glen Rose.

ward side are in direct contact with the water-charged arenaceous beds of the lower Glen Rose and Travis Peak formation underlying the Edwards Plateau (as shown in fig. 76). In such places the limestone strata under the plain receive their water supply, and it is forced into them with the same pressure that carries another part of it to the surface in the fissure springs.

The same explanation does not seem to apply to that part of the Travis Peak formation which is embedded under the Rio Grande Plain. The fault wholly separates it from the part underlying the plateau, and that which is opposed to its cut edges is probably some portion of the deep-lying Paleozoic series; and we have no reason to suppose that those rocks could furnish it either a large quantity or a good quality of water. It seems possible that a part of the water from the Travis Peak sands under the plateau may pass downward along the fault plane to the Travis Peak beds under the plain, just as some passes upward along the fault plane to the surface of the ground; but it is perhaps more probable that the passage from west to east takes place farther north, where there is no fault and the basement sands are continuous.

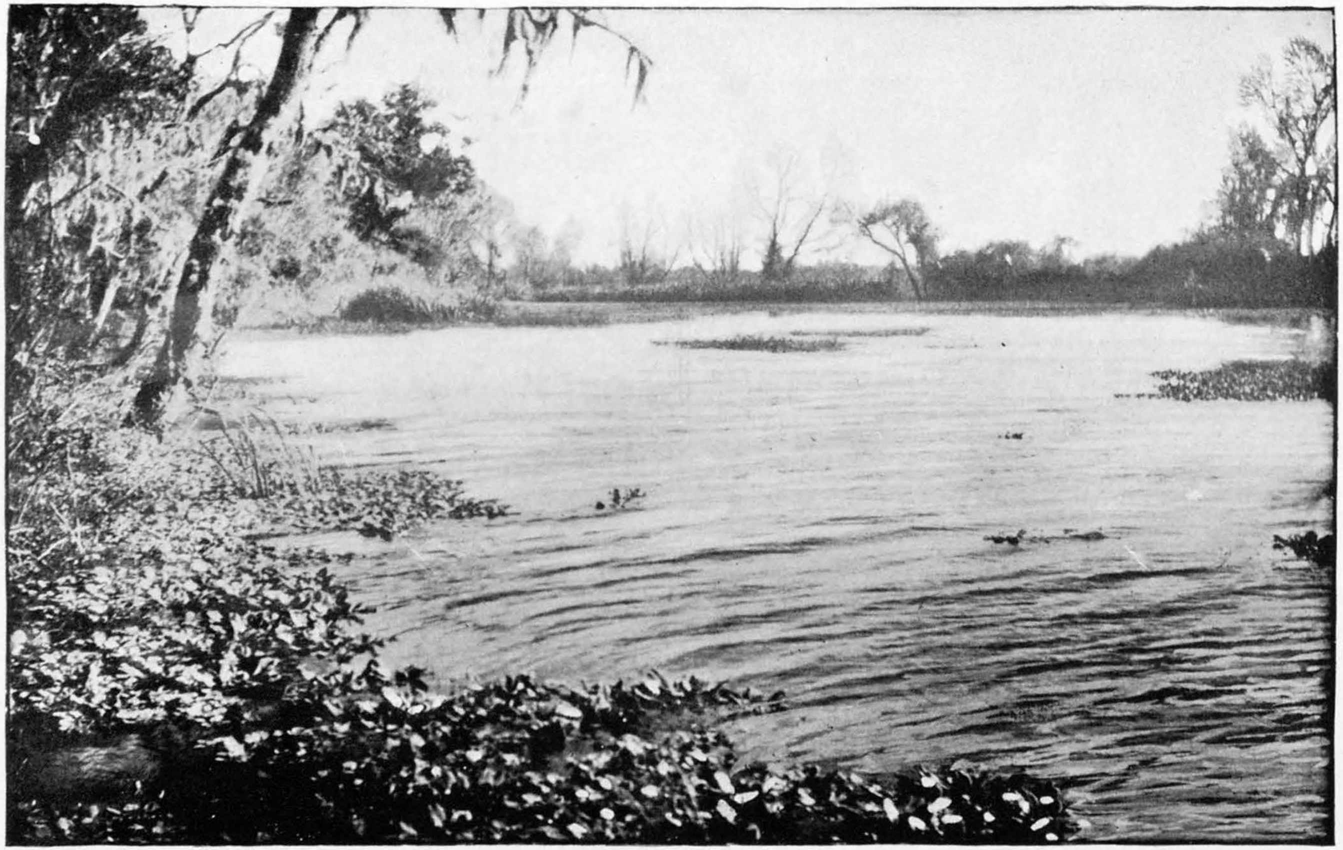
### AVAILABILITY AND LIMITATIONS OF THE UNDERGROUND WATERS.

Having now explained the general conditions controlling the occurrence and distribution of artesian water along the interior margin of the Rio Grande Plain, a more definite word or two concerning the limitations of the possible success of wells may be said. It should be remembered that owing to the lack of good topographic and geologic maps it is impossible to discuss the details of the respective districts with that degree of accuracy which is desirable, and owing to these reasons all opinions herein expressed are tentative and based upon observations largely of the nature of reconnaissance.

For the purpose above mentioned the Rio Grande Plain may be subdivided into four districts, as follows: (1) That portion lying between and including Austin and San Antonio, (2) the portion lying between San Antonio and the Sabinal, (3) the Uvalde laccolithic district between the Sabinal and the east end of the Anacacho Mountains, and (4) the portion lying west of the above, embracing the region westward nearly to the mouth of Devils River.

In the first of these districts the northernmost locality of the occurrence of flowing artesian water derived from the Edwards limestone is at Manor, in Travis County. The most southern locality is Colonel Terrell's well, south of San Antonio. As the same structural conditions which control the success of these wells prevail from north of the Colorado to San Antonio, it is a logical deduction that similar water can be obtained throughout the intervening distance of 90 miles. The only recorded well throughout the distance is the one at San Marcos, and this supports our hypothesis.

From the experiments at Manor and Marlin it is evident that artesian wells can be obtained at least as far east as the eastern margin of the Cretaceous Black Prairie. In view of these facts, it is reasonable to believe that water can be obtained at depths of from 200 to 3,000 feet (the depth increasing to the eastward) in all the region between San



RUN-OFF OF SAN MARCOS SPRING.

Waves produced by spring welling up from below.

Antonio and the San Gabriel, in a belt of country averaging 20 miles in width, lying between the post-oak timbered region on the east and the International and Great Northern Railway on the west.

In the second district mentioned, from San Antonio westward to the Sabinal, the same general conditions prevail, with some slight but important modifications, in the structure. It is our opinion that that portion of the belt lying north of the boundary between the Eagle Pass and Taylor formations, and south of the Balcones fault, except where disturbed by igneous intrusion, presents favorable conditions for procuring artesian wells.

This belt may be limited to the coastward by a line arbitrarily drawn parallel with the Southern Pacific Railroad and 5 miles south of it. We have in our possession no record of any well experiments having been made in this particular region, nor have the structural conditions been minutely studied, but so far as known they are favorable for the securing of artesian wells of the Austin, San Antonio, and San Marcos type.

It would be well for those who contemplate drilling wells to ascertain positively whether any such experiments have been made. The water-bearing strata deepen more rapidly away from the Balcones fault in this region than to the northward, and hence the width of available area is less. It is doubtful if wells of less than from 1,500 to 2,500 feet in depth can be obtained along the line of the Southern Pacific.

The third district is one concerning which we can not express an opinion favorable to the procurement of artesian waters. In prospecting for water in this region it is necessary to consider the effect of the numerous igneous intrusions. Owing to their presence it is not probable that flowing wells can be obtained in this portion of the Rio Grande Plain. These igneous rocks break the continuity of the strata, and how much they may affect the occurrence of underground water is an important question.

The presence of masses of igneous rocks also renders the success of artesian experimentation doubtful in a small section of country belonging to the first district, i. e., the immediate vicinity of Pilot Knob, in Travis County. The recent experiments at St. Edward's College, Manor, and Austin prove that the presence of intrusions of the Pilot Knob group does not affect the artesian conditions for any great distance to the west of them. No experiments have been made to the east.[1]

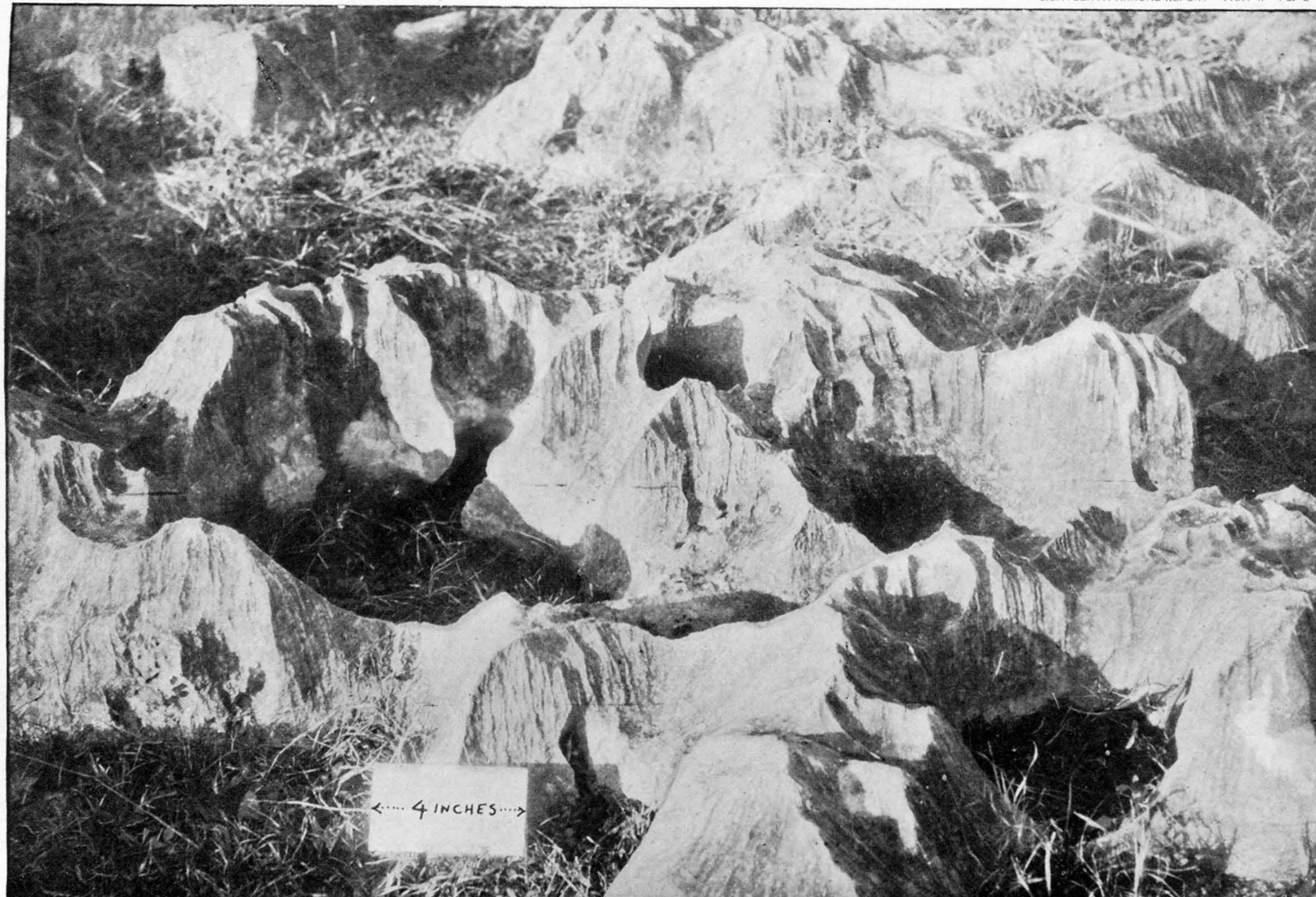
In how wide an area outside of the immediate occurrence of the Uvalde group of laccoliths the artesian prospects are influenced is also

[1] Since this report was prepared the writers have learned that Mr. William Blocker, living about 2 miles southeast of Pilot Knob, has commenced an artesian well. This well begins in the Taylor formation, and necessarily would have to go nearly as far as the depth of the San Marcos well to reach the Edwards limestone, or 1,000 feet lower to the Travis Peak flow. If this well should prove or has proved successful, it will demonstrate the possibility of success for a large area of country between the Colorado and Mayhard Creek as far east as Bastrop. No more important experiment than this well has as yet been undertaken, and the writers regret very much that they have not means at their disposal to study its progress.

problematic. From experiences around Austin, and the occurrence of artesian springs at Fort Clark, just beyond the volcanic outcrops, it seems probable that those intrusions do not seriously affect the artesian conditions for any great distance outside of their own immediate area, i. e., the Uvalde laccolithic region of the Rio Grande Plain between the Sabinal and the east end of the Anacacho Hills.

In the fourth district named, we again meet some evidences of the possibility of obtaining artesian water. Two artesian-well experiments have been made west of the Uvalde laccolithic area, at Spofford Junction and Del Rio. The experiment at Spofford Junction was made many years ago by the Galveston, Harrisburg and San Antonio Railway Company. We have not been able to obtain a trustworthy record of the boring. The depth to which this well was dug has been variously reported at from 1,160 to 1,800 feet. From our geologic studies of this region we know that Spofford is underlain by the elsewhere water-bearing beds of both the Edwards and Travis Peak formations. Whether the drill ever penetrated one or both of the water-bearing beds is uncertain. A report in our possession makes it seem probable that water was struck at a depth of over 1,800 feet, and rose within the well some 600 feet, or to within 1,200 feet of the surface. We have been informed that the fossil *Exogyra arietina*, which is peculiar to the Del Rio clays, was found at a depth of 1,800 feet. If this is true the water obtained was probably the top water of the Edwards beds, and the well should have been continued at least 500 feet before being abandoned. The fact that the water rose 600 feet in this well demonstrates that beneath the place there is water under hydrostatic pressure. The position of Spofford relative to the beds of the Cretaceous section is problematic, owing to the fact that the surface at that place is obscured by the post-Cretaceous Uvalde conglomerate, and the rocks have been faulted in the neighborhood. The geologic position of the place is certainly higher than the Austin chalk, and we are inclined to believe that it is well up in the Anacacho formation, or toward the base of the Eagle Pass. If these hypotheses be correct, then the water alleged to have been struck at 1,800 feet must have been in the Edwards beds, and it is assumed that the basal sands have not been penetrated here as yet. Furthermore, artesian water does rise at the Las Moras springs, 100 feet above the altitude of Spofford; and hence we believe that if a well should be continued to the base of the Cretaceous series at Spofford, it is probable that an artesian flow would be struck at a depth of less than 2,500 feet.

Both Spofford and Del Rio are typical localities of that portion of the Rio Grande Plain lying west of the Anacacho Mountains, and are situated on the downthrown side of the Balcones monocline, where the stratigraphic conditions necessitate the embedding of the water-bearing strata. Furthermore, both localities are immediately south of great artesian springs, those of the Las Moras and San Felipe, respectively, which show the presence beneath the region of water-bearing strata of




KARRENFELDER, EDWARDS LIMESTONE.

great capacity, identical in character with those of San Antonio, San Marcos, and Austin, and the waters are undoubtedly of the same origin, coming up through natural fissures from an embed which is equally accessible to the artesian drill.

Among the data in our possession there is no record of any drill hole in the Rio Grande Plain country on the downthrown side of the Balcones fault between San Marcos and the Devlis River which has as yet penetrated to the water-bearing basement beds of the Travis Peak formation. From the records of the wells and study of the geologic sections of the canyons of the plateau region, we know that these beds are well developed up to the very margin of the Rio Grande Plain, and must exist beneath it. They have been penetrated and are water producing in the Austin, Kerrville, and San Marcos wells. In view of these facts, we are inclined to the belief that the artesian experimentation in this subregion has not been carried to a sufficient depth to reach the Travis Peak sands. This opinion is not to be taken as a positive prediction of success, and might be changed if we possessed more accurate logs of the wells already bored.

## PRACTICAL SUGGESTIONS.

While this paper does not propose to deal with other than the geologic side of the question of underground water, it may not be inappropriate to offer a few direct suggestions bearing upon the location and drilling of wells.

The question of interest to the well-seeker is, How can I estimate the depth of the water-bearing sheet below the spot where I propose to sink my well? In a general way there are two methods. If there is another well within a few miles of the locality, whose log shows the local depth of the desired water, and if the rocks are not faulted, then by determining the dip of the rocks he can compute from dip and distance the difference in absolute altitude of the water horizon at the two places; he can also, by actual measurement with the surveyor's level, determine the difference in height of the surface of the ground at the two places. These two differences can then be applied as corrections to the depth of the water horizon in the existing well, and the result will give an estimate of its depth at the place where the well is desired.

The second method demands a knowledge of the geologic column of the region—that is, of the various formations, their order of superposition, and the thickness of each. Which formation occupies the surface of the ground at the locality where boring is projected must also be known. Then, by adding the thickness of all the formations between the one at the surface and the one carrying the water below, an estimate of the depth to which it will be necessary to drill may be obtained. Accurate geologic surveys, such as have been made of two typical portions[1] of

[1] The Austin and Brackett quadrangles have been surveyed in detail and the data are now in process of publication.

the region, would give this information. Unfortunately, however, only reconnaissances have been made over the rest of the region, and in the absence of maps these determinations for each particular locality should be referred to a professional geologist.

Paleontology is the most reliable guide in determining the position of any bed in the geologic series, in order to ascertain the depth from any particular portion of the surface of the underground waters in the Cretaceous regions of Texas. If a dozen fossils, such as can be found in nearly any locality, be sent to a geologist familiar with the sequence of the beds, he can predict within a few feet the depth below the surface of any particular stratum in the series.

The specimens of the rock sheets as brought up from the drill hole differ quite materially in color, and sometimes in hardness, from the surface outcrops of the same rock, in which the color is changed by oxidation and the texture by induration or decomposition. Thus it is that limestones, marls, and clays brought from considerable depths in these wells are usually bluish or blue-black in tint, while the surface outcrop of the same beds, owing to drying, will be lighter, or if they contain iron, yellow, or occasionally red.

In general, if one boring an artesian well should commence at the uppermost of the Cretaceous formations and penetrate the entire system, he would be able to recognize the different beds as they were encountered by the following characters:

North of San Antonio the Webberville, Taylor, Eagle Ford, and the Del Rio formations are composed of marls, locally called "joint clays." These can be distinguished from one another as follows: The Webberville joint clays always have minute specks of the greenish-black mineral called glauconite. The Taylor marls are very blue and possess no accessory mineral except lime; they are especially free from grit. They weather into lumps with conchoidal fracture. The clays of the Eagle Ford beds are blue-black in color, impure in composition, and have a very finely laminated structure, whereby on the least weathering they separate into leaf-like layers. They also have occasional layers of impure laminated limestone of blue or brown colors. The Del Rio beds are usually blue clays containing much iron pyrites and the characteristic oyster, *Exogyra arietina*, which almost invariably appears in the well débris when the drill penetrates this horizon.

Still other clays are encountered after passing through great thicknesses of limestone beneath the Del Rio clays. The first of these will be the Walnut clays, which contain a different species of oyster, *Exogyra texana*, and have a somewhat granular character. The clays in the upper Glen Rose beds are also very granular, and often contain much minute shell débris. The lowest clays of the Cretaceous formations found intercalated in the sands of the basement beds have faded red colors, even when brought up from these great depths.

The first limestone of much thickness encountered by the driller will be the Austin chalk, which, under the churn drill, appears as an unctuous blue mud. This lies between the Taylor and Eagle Ford marls and is from 400 to 500 feet in thickness.

The few limestone layers of the Eagle Ford beds can be recognized by their laminated character, which distinguishes them from all other limestones of the region.

The Shoal Creek limestone is intensely hard, and the well-driller will have no difficulty in recognizing it when he encounters it.

The Fort Worth, Edwards, and Comanche Peak limestones will constitute practically one formation to the well-driller, having an aggregate thickness of about 450 feet in the Colorado Valley and about 640 along the Nueces. This horizon can be recognized principally by its thickness and hardness. The Edwards limestone is easily distinguished by the flints that it contains.

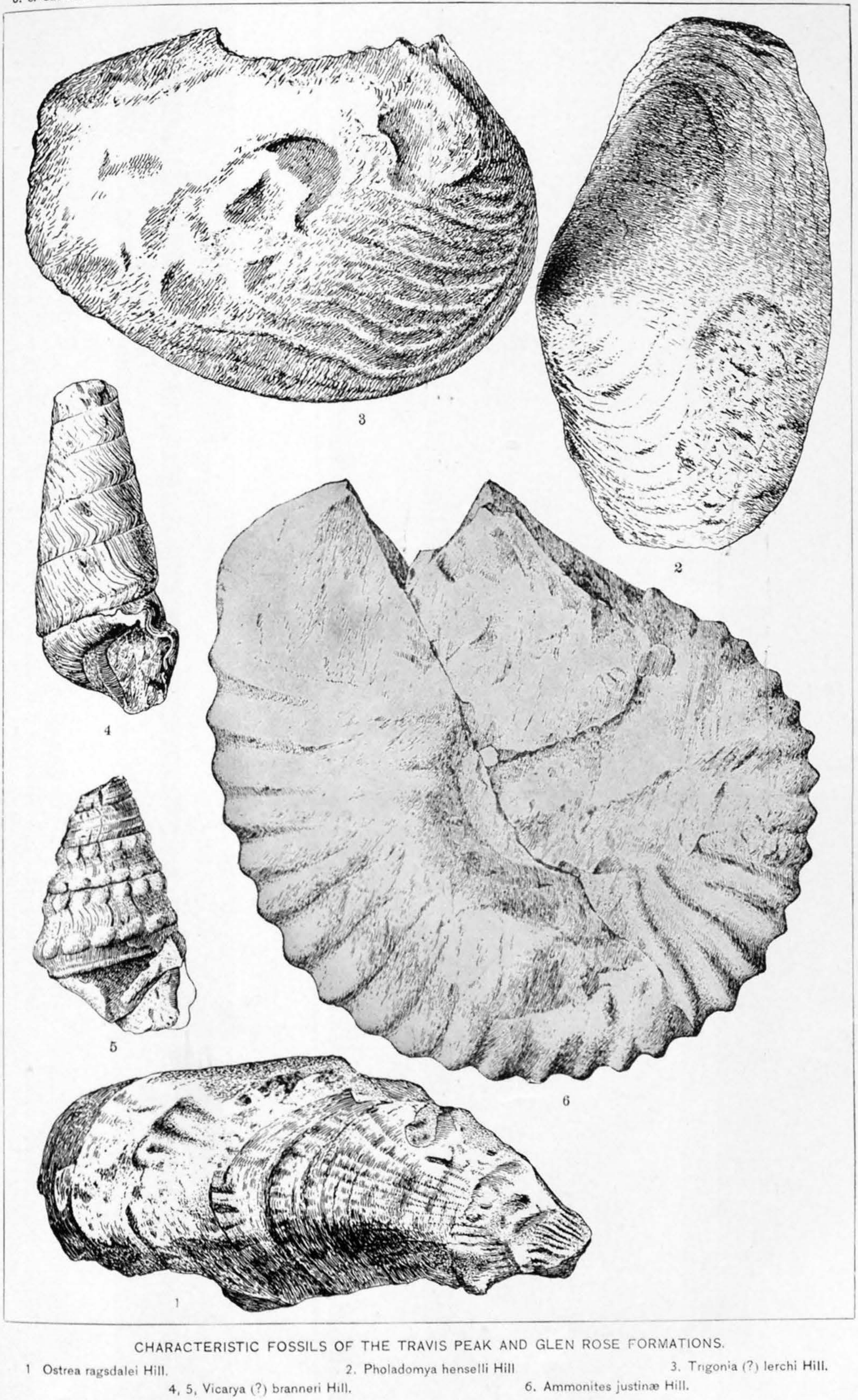
The Glen Rose beds will be recognized by their consisting of alternations of soft flaggy limestones and calcareous clays.

Probably, in boring, the most easily recognized geologic horizon is that of the Del Rio clays. The Travis Peak water in the vicinity of Austin is 1,300 to 1,600 feet below them. The water-bearing sands are slightly more than 300 feet thick. The "sweet water" at San Antonio occurs only 80 to 100 feet below the Del Rio.

The following table gives the variation in thickness, the approximate position of the water horizons, and the most noteworthy characters of each formation.

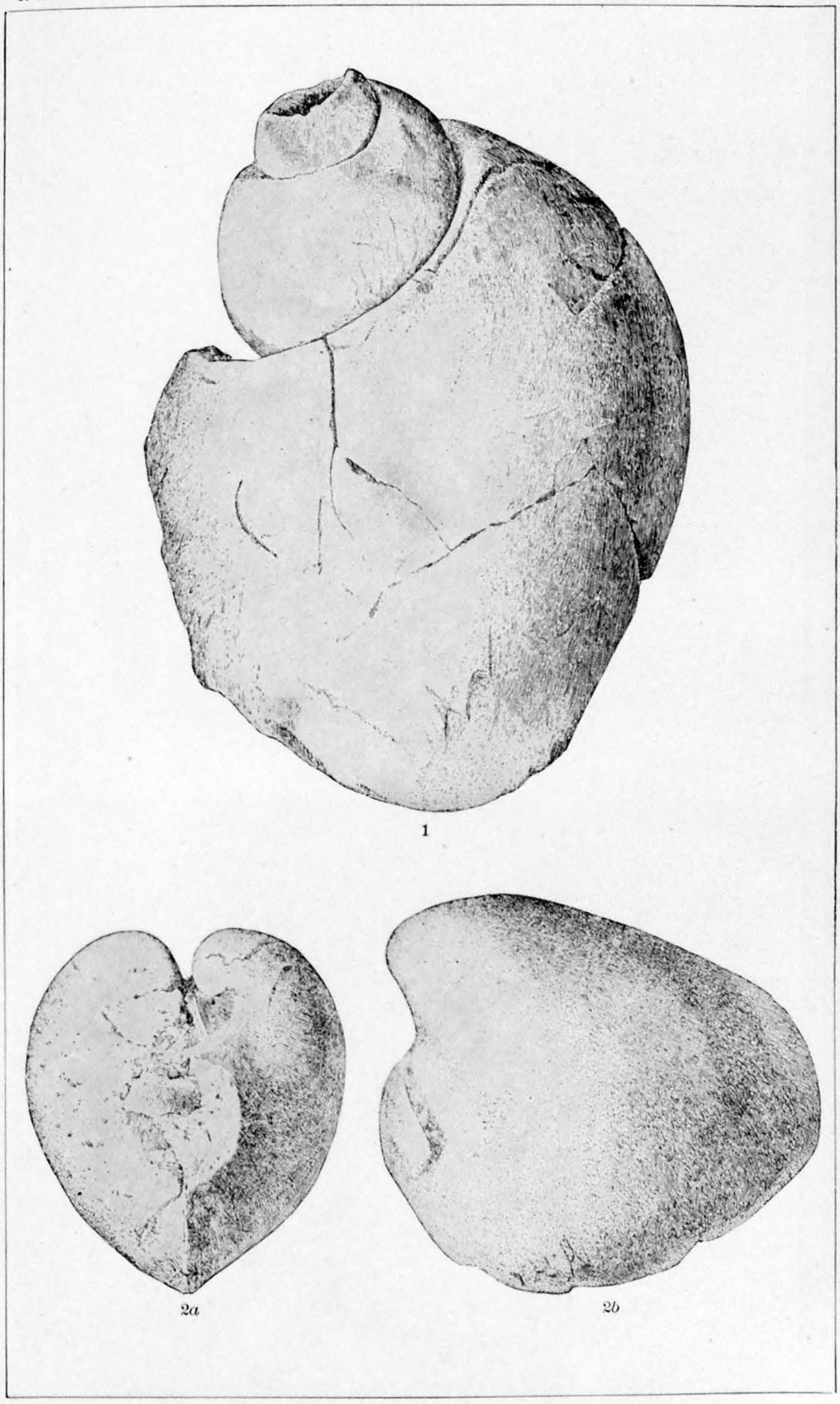
*Thicknesses, water horizons, and characters of the formations.*

| Formation name. | Thickness in feet. | Water horizon. | Character. |
|---|---|---|---|
| Webberville | | None | Glauconitic clays and sands. |
| Taylor clays | 500–550 | ....do | Blue "joint" clay. |
| Austin chalk | 400–500 | Some bad water | Usually soft white or blue limestone. |
| Eagle Ford shales | Usually 25–60; 250 in Kinney County. | Bad water; gas | Dark-blue laminated shale, some limestone bands. |
| Shoal Creek limestone. | 40–100 | Bad water at San Antonio. | Hard limestone. |
| Del Rio clays | 75–100 | None or bad | Blue clay, containing the ram's horn oyster, *Exogyra arietina*. |
| Fort Worth limestone. | 50–75 | Probably none | Soft limestone. |
| Edwards limestone | 250 Colorado River; 640 Nueces River. | Several. The water near the top often bad; in vicinity of San Antonio good or bad; usually good. | Limestone with flints. |
| Comanche Peak limestone. | About 40. | None so far as known. | Soft limestone. |
| Walnut clays | 10–25 | None | Clays, with many *Exogyra texana*. |
| Glen Rose formation. | 400–700 | Several; water usually not good. | Alternations of limestone and clays. |
| Travis Peak formation. | 200–400 | Water excellent | Conglomerate, sands, and clay. The latter may be red even when brought up in the well drill. |



CHARACTERISTIC FOSSILS OF THE TRAVIS PEAK AND GLEN ROSE FORMATIONS.

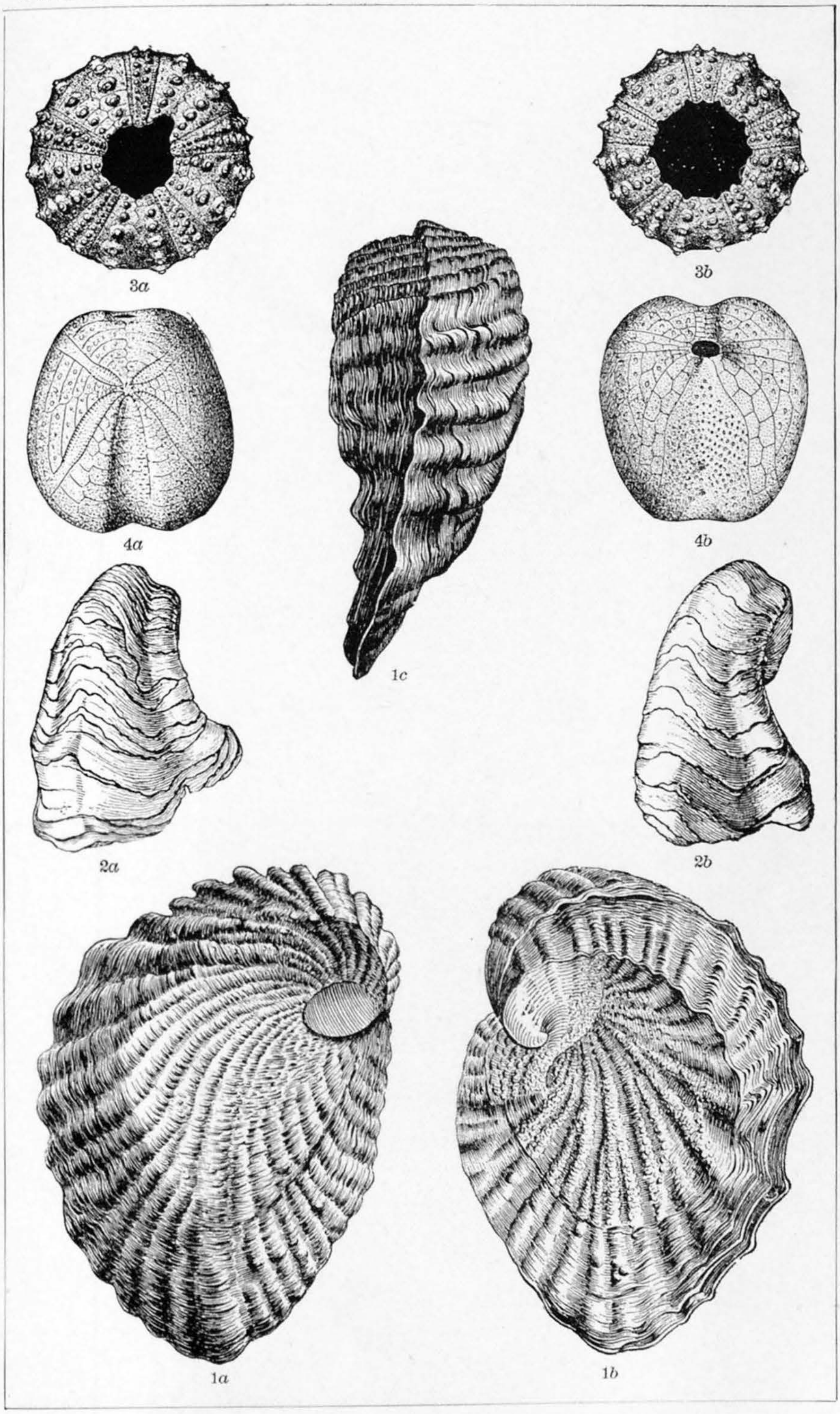
1. Ostrea ragsdalei Hill. 2. Pholadomya henselli Hill 3. Trigonia (?) lerchi Hill. 4, 5, Vicarya (?) branneri Hill. 6. Ammonites justinæ Hill.




CHARACTERISTIC FOSSILS OF THE GLEN ROSE FORMATION.

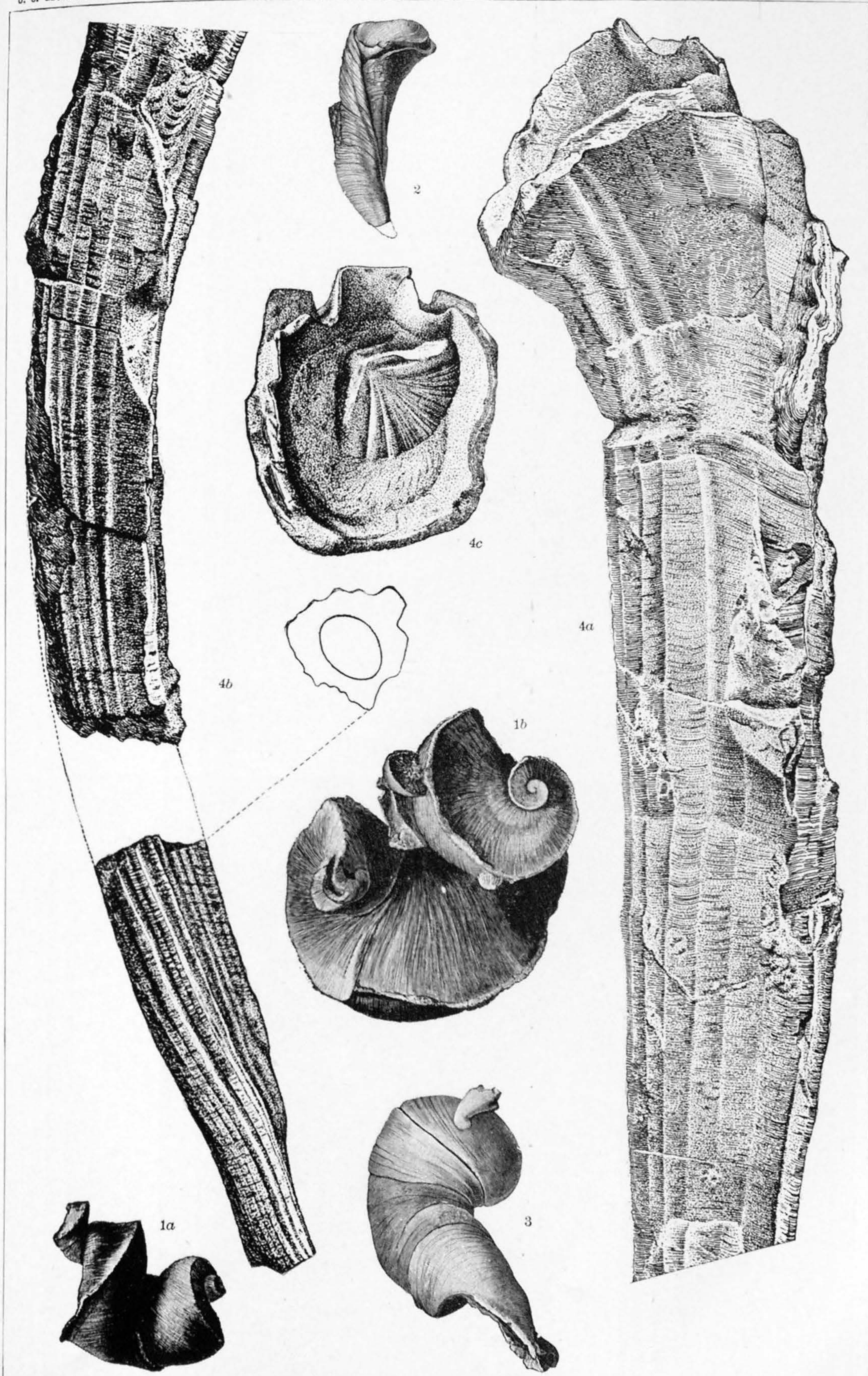
1. Tylostoma pedernalis (Roemer).

2*a*, 2*b*. Cardium mediale Conrad.



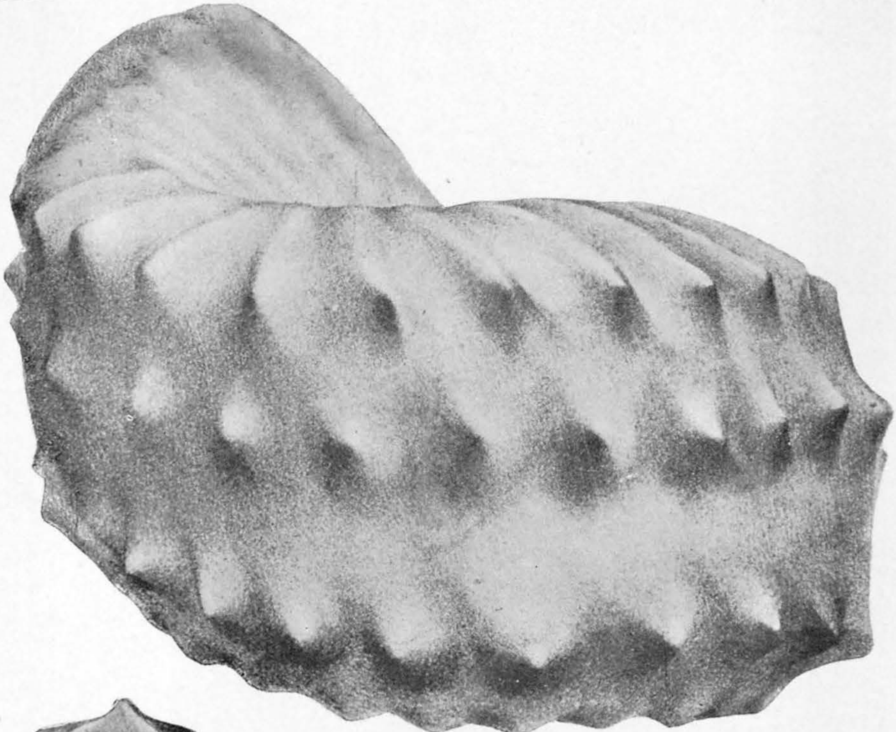

CHARACTERISTIC FOSSILS OF THE WALNUT CLAYS AND COMANCHE PEAK LIMESTONE.

1*a*, 1*b*, 1*c*. Exogyra texana Roemer.
2*a*, 2*b*. Gryphæa marcoui Hill and Vaughan, sp. nov.
3*a*, 3*b*. Pseudodiadema texanum Roemer.
4*a*, 4*b*. Enallaster texanus Roemer.




CHARACTERISTIC FOSSILS OF THE EDWARDS LIMESTONE.

1*a*, 1*b*. Requienia patigiata White.
2. Monopleura marcida White.
3. Monopleura pinguiscula White.
4*a*, 4*b*, 4*c*. Radiolites davidsoni Hill.



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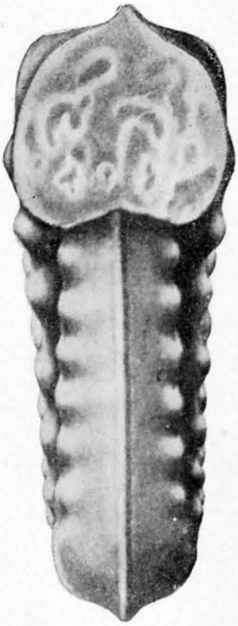
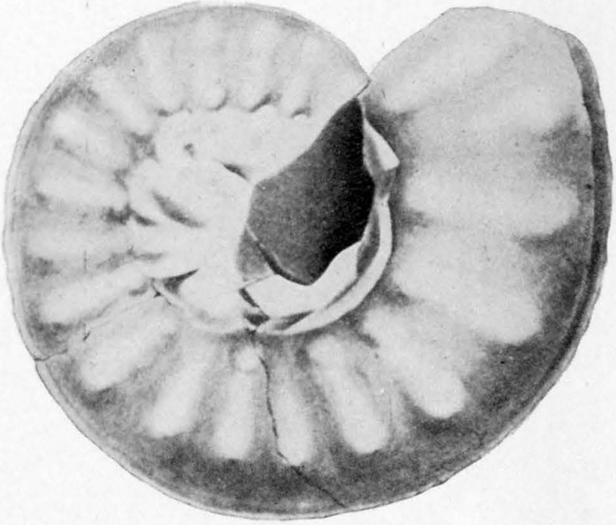


2a



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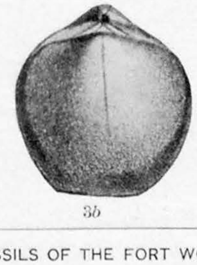
3a



3b

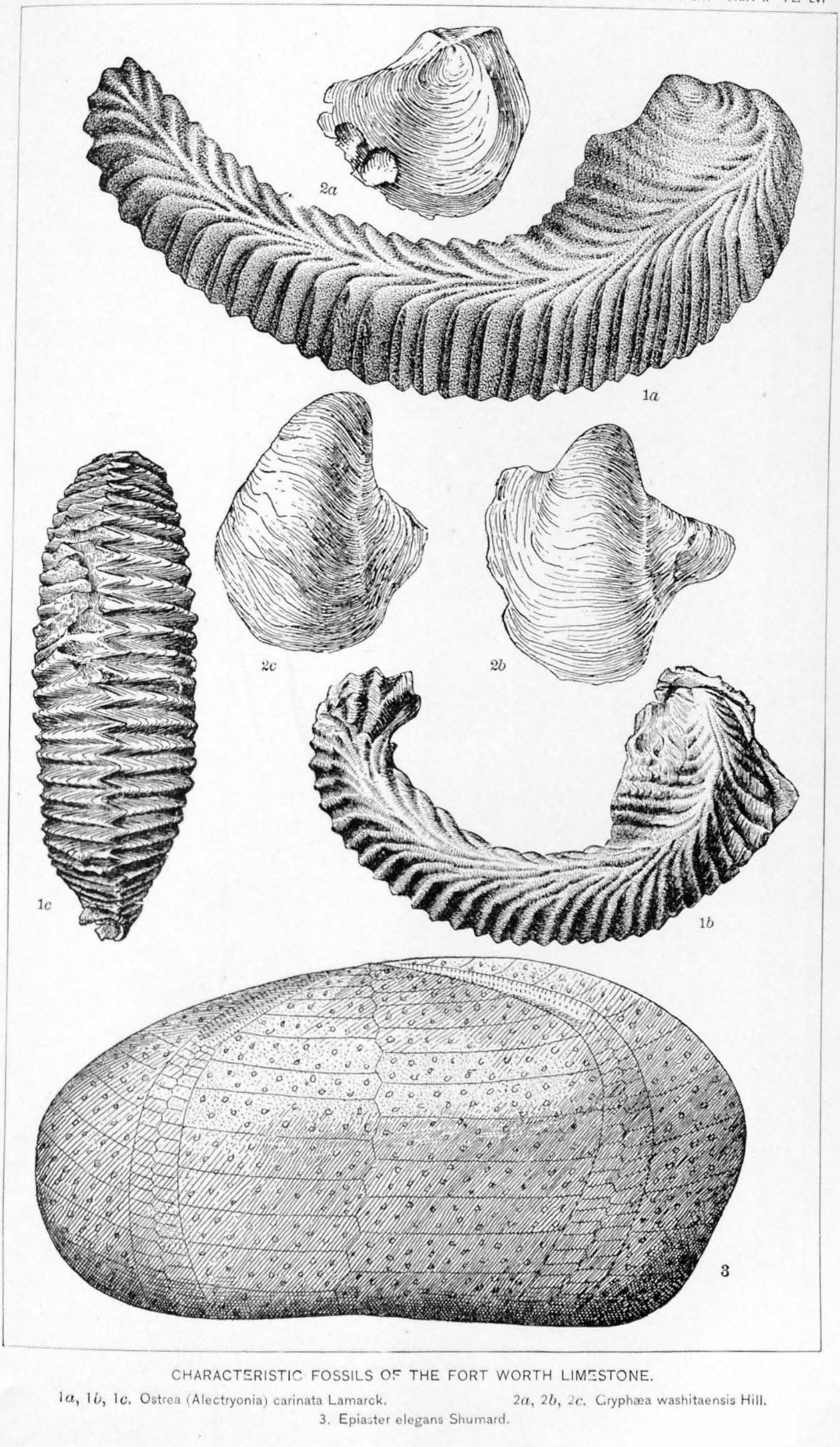
3c

CHARACTERISTIC FOSSILS OF THE FORT WORTH LIMESTONE.

1. Turrilites brazoensis Roemer.

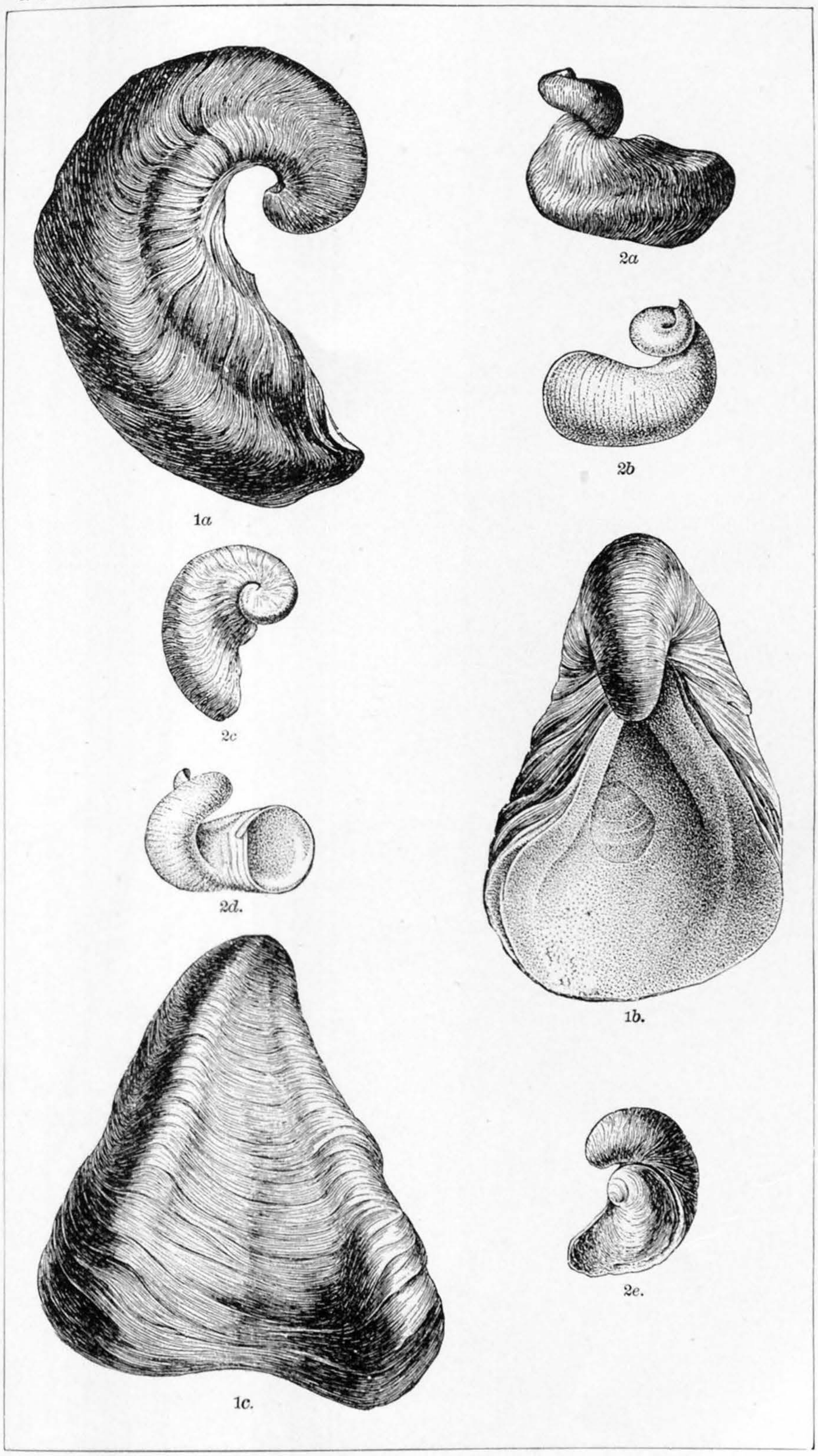
2a, 2b, Schlœnbachia leonensis (Conrad).

3a, 3b, 3c. Kingena wacoensis (Roemer).




CHARACTERISTIC FOSSILS OF THE FORT WORTH LIMESTONE.

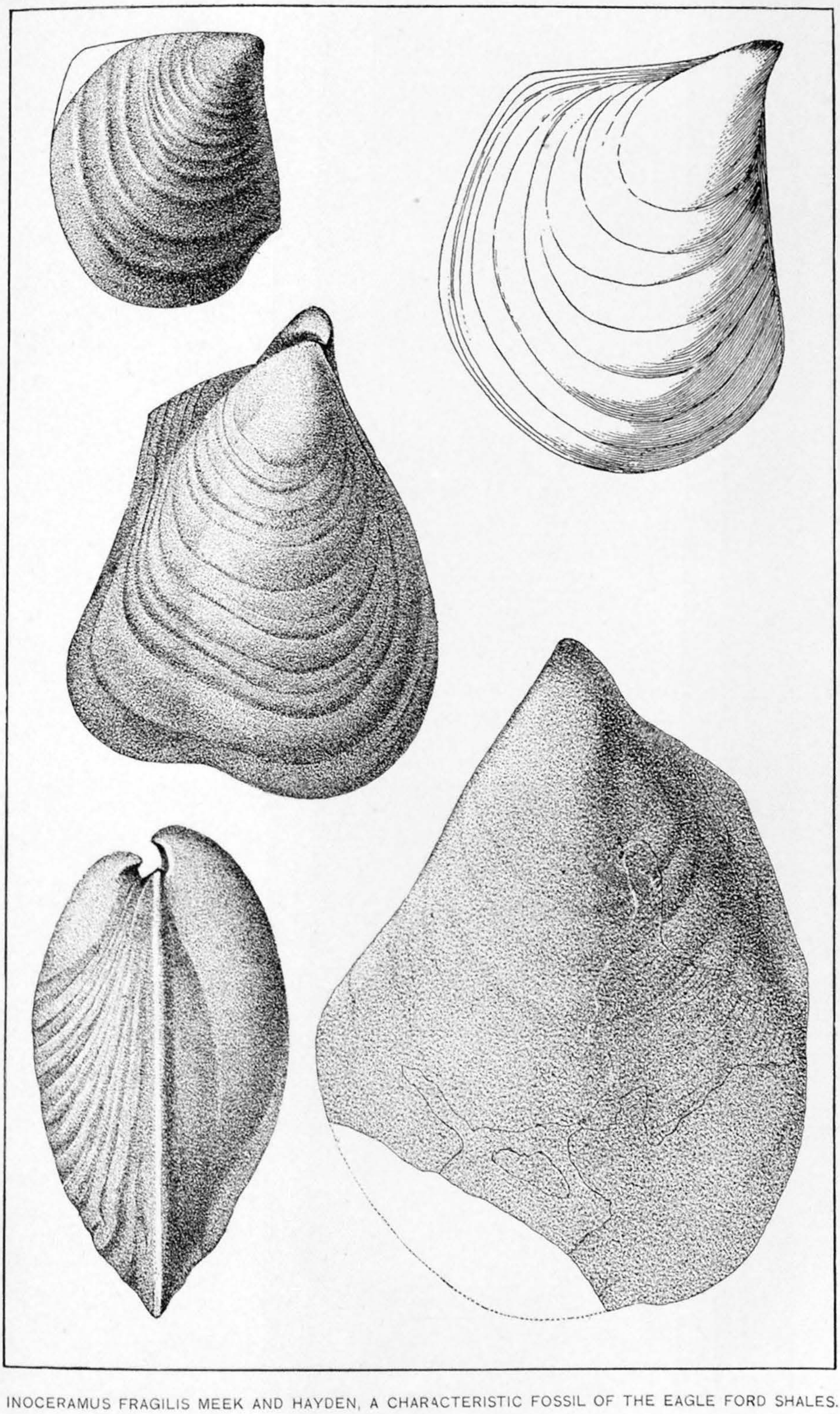
1*a*, 1*b*, 1*c*. Ostrea (Alectryonia) carinata Lamarck. 2*a*, 2*b*, 2*c*. Gryphæa washitaensis Hill.
3. Epiaster elegans Shumard.



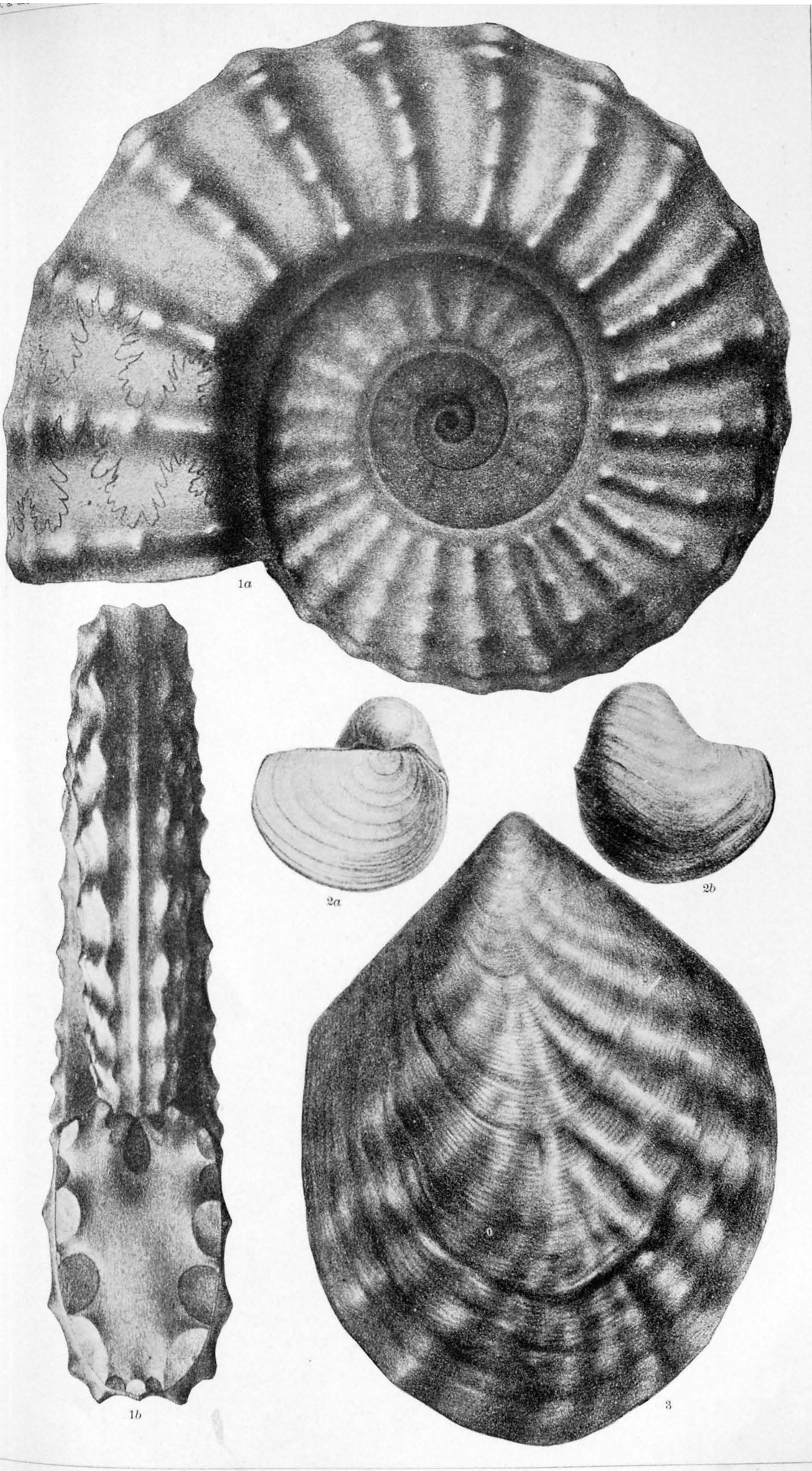

CHARACTERISTIC FOSSILS OF THE DEL RIO CLAYS.

1*a*, 1*b*, 1*c*. Gryphæa mucronata Gabb.
2*a*–2*e*. Exogyra arietina Roemer.

VOLA ROEMERI HILL, A CHARACTERISTIC FOSSIL OF THE SHOAL CREEK LIMESTONE.



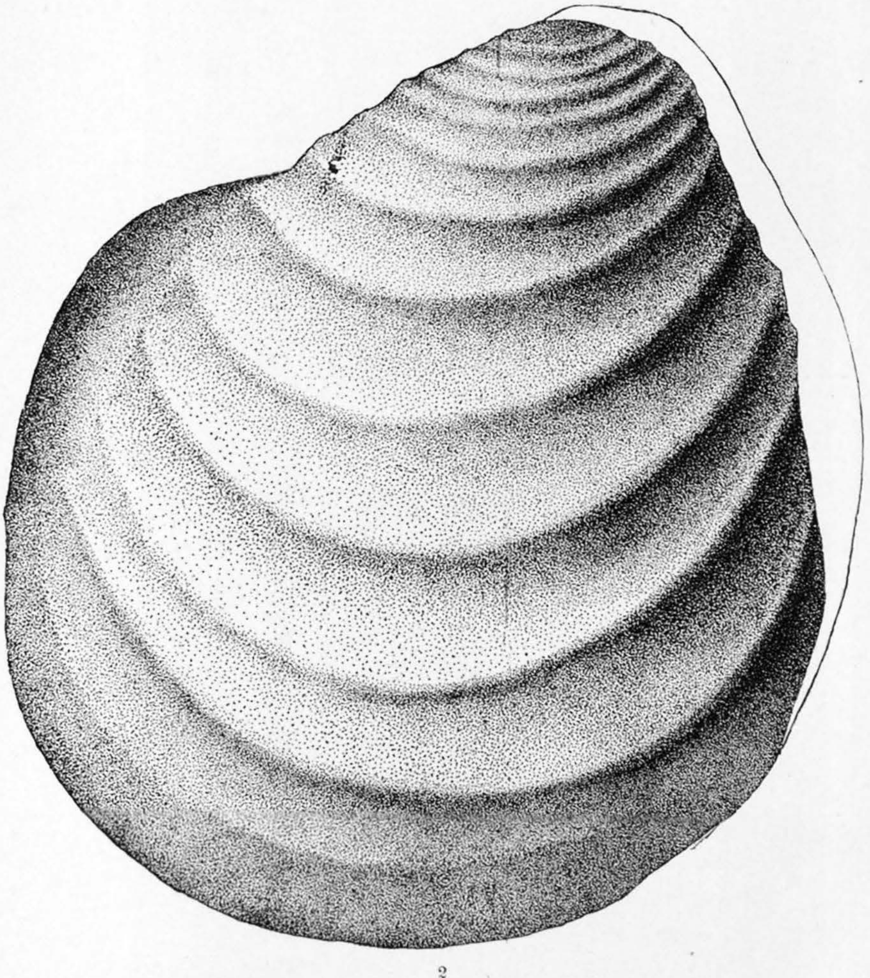
INOCERAMUS FRAGILIS MEEK AND HAYDEN, A CHARACTERISTIC FOSSIL OF THE EAGLE FORD SHALES.



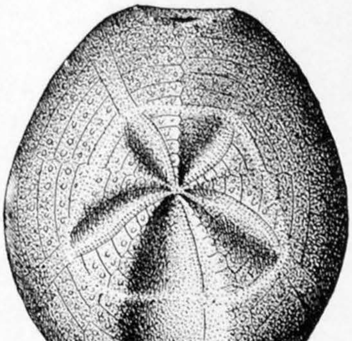

CHARACTERISTIC FOSSILS OF THE AUSTIN CHALK.

1*a*, 1*b*. Mortoniceras texanus (Roemer). 2*a*, 2*b*. Gryphæa aucella Roemer.

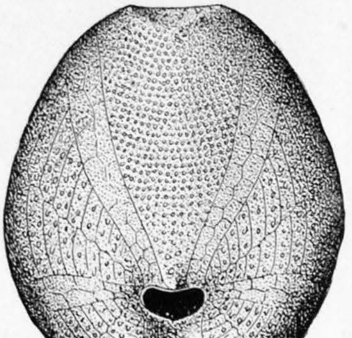
3. Inoceramus digitatus Sowerby — Inoceramus undulato-plicatus Roemer.



2



1*a*

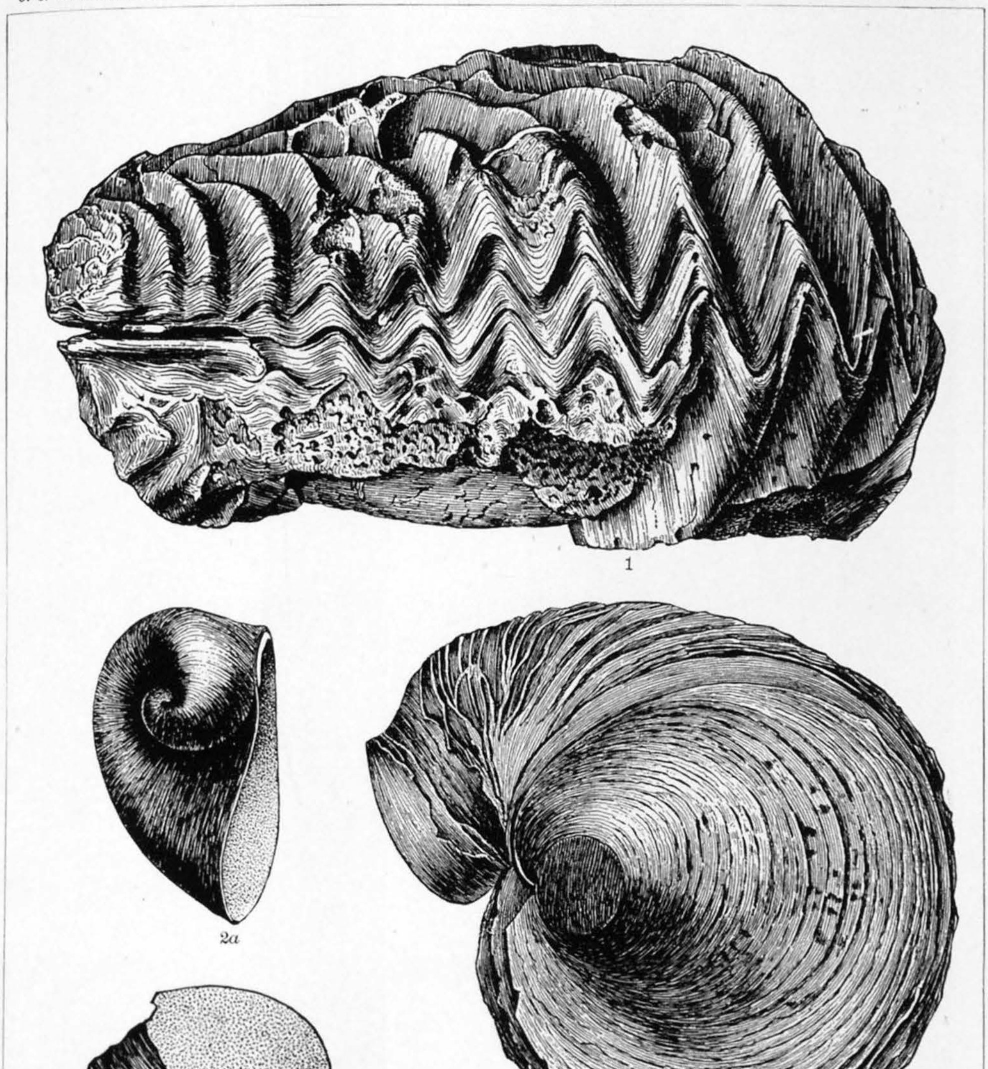


1*b*

CHARACTERISTIC FOSSILS OF THE AUSTIN CHALK.

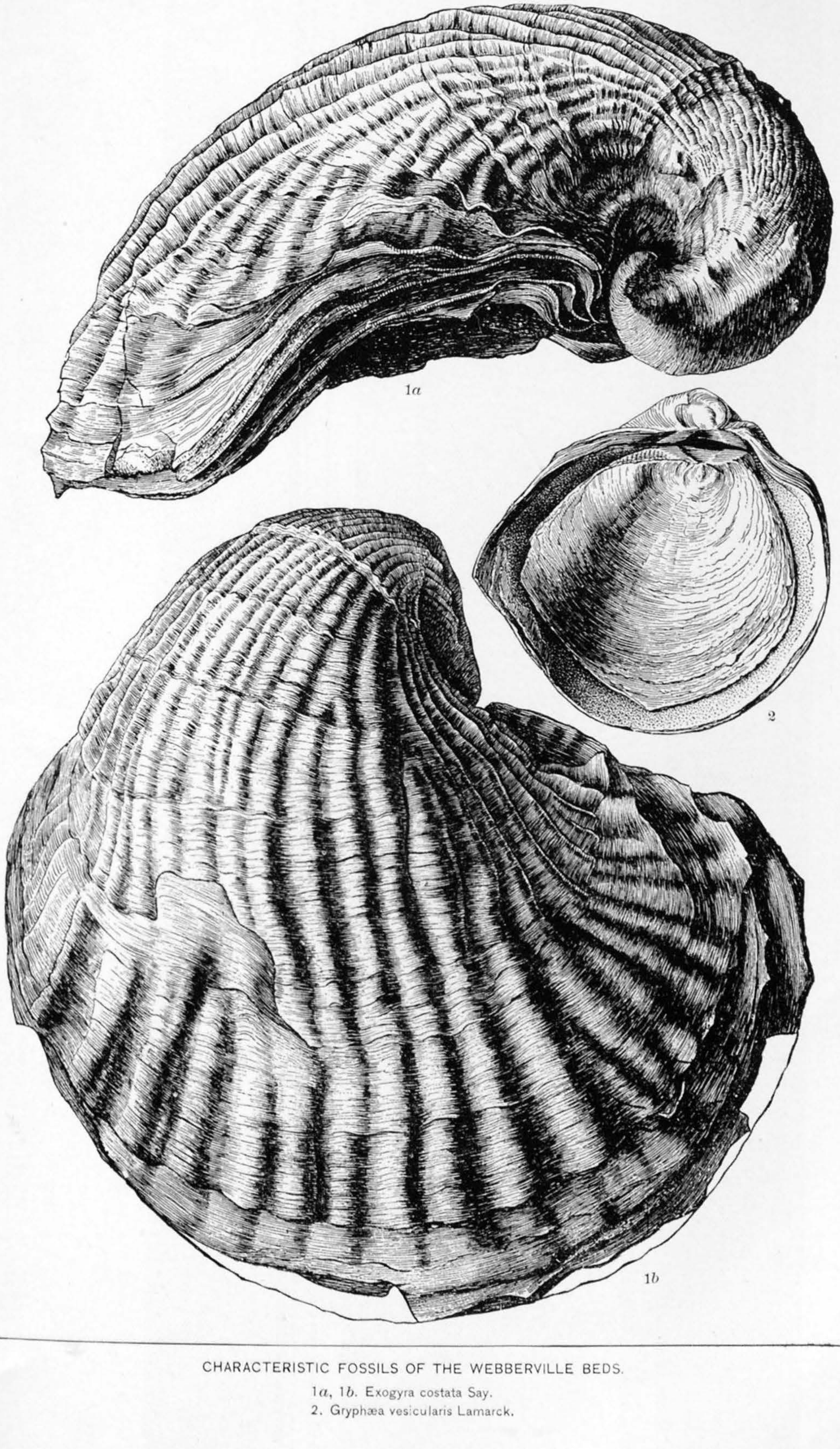
1*a*, 1*b*. Hemiaster texanus (Roemer).

2 Inoceramus deformis Meek.



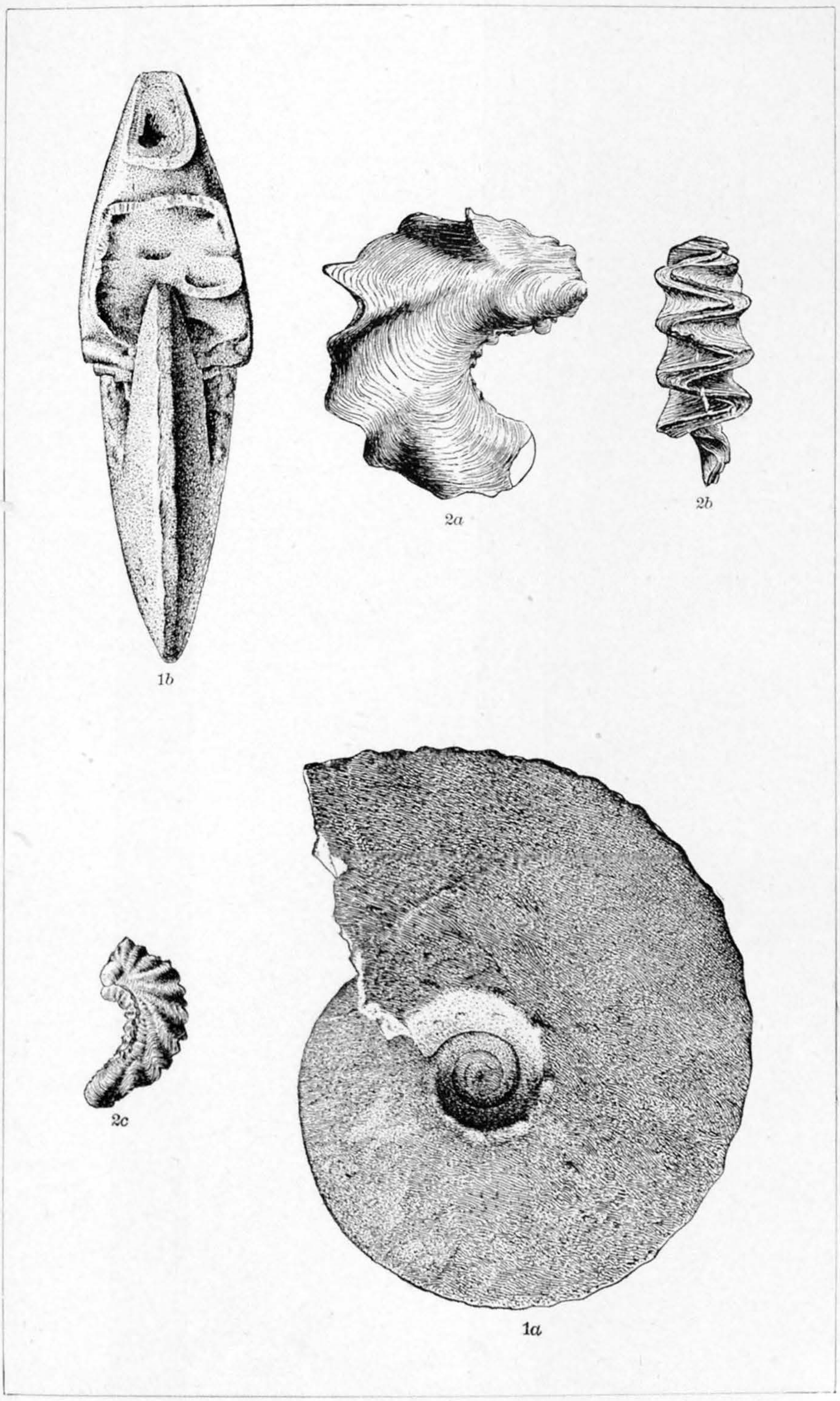

CHARACTERISTIC FOSSILS OF THE UPPER PART OF THE AUSTIN CHALK AND TAYLOR MARLS.

1 Ostrea (Alectryonia) diluviana Linné Austin chalk.
2*a*, 2*b*. Exogyra læviuscula Roemer. Austin chalk.
3*a*, 3*b*. Exogyra ponderosa Roemer. Austin chalk and Taylor marls.




CHARACTERISTIC FOSSILS OF THE WEBBERVILLE BEDS.

1*a*, 1*b*. Exogyra costata Say.
2. Gryphæa vesicularis Lamarck.




CHARACTERISTIC FOSSILS OF THE WEBBERVILLE BEDS.

1*a*, 1*b*. Placenticeras placenta Morton.
2*a*, 2*b*, 2*c*. Ostrea larva Lamarck.

# A TABLE OF THE NORTH AMERICAN TERTIARY HORIZONS, CORRELATED WITH ONE ANOTHER AND WITH THOSE OF WESTERN EUROPE, WITH ANNOTATIONS.

BY

WILLIAM H. DALL.

# CONTENTS.

# A TABLE OF THE NORTH AMERICAN TERTIARY HORIZONS, CORRELATED WITH ONE ANOTHER AND WITH THOSE OF WESTERN EUROPE, WITH ANNOTATIONS.

By W. H. Dall.

## INTRODUCTION.

The present article has grown out of a request from the Director, in 1895, that I should prepare a table of the Tertiary formations, which at first was not intended to carry many details. So many points of unexpected interest developed as the work went on that it gradually assumed its present shape.

The standard column of the Tertiary horizons in the United States is that afforded by the series north of the Gulf of Mexico and east of the Mississippi River, near the Gulf border. Even this series is not to be regarded as completely differentiated, since, with the exception of the Midway limestones and the Claiborne sands, hardly a single one of the successive faunas has been thoroughly worked out. No doubt minor changes will be necessary when this has been accomplished; new zones will be discriminated and possibly some consolidations will be advisable. But, on the whole, the series is the best we have, and the only one which at present can be adopted as a provisional standard for the Tertiary. With this are compared the beds of the Atlantic States and the Antilles, those of the western interior lake basins, and those of the Pacific coast. To the left a column shows the general divisions adopted for the United States, and to the right another gives the equivalents in use in western Europe or elsewhere.

For the data the annotations will show the authority, but I should mention that I am especially indebted to Prof. W. B. Scott for the use of his correlations of the vertebrate faunas of the West with those of Europe.

Criticisms will doubtless suggest themselves to geologists, especially in the treatment of the Pacific coast, where our knowledge of the faunas is least and the data for correlation are insufficient for anything like finality. But the very imperfections of such a table are a stimulus toward its betterment, and the arrangement is submitted with the understanding that it is provisional.

The table is not intended to comprise all the known beds or the minor subdivisions, but rather the principal horizons of which the correlation is reasonably certain. The use of a name is not to be taken as an indication that its synonymy has been thoroughly worked out and that the particular name used has been adopted as a finality. On the contrary, unless otherwise indicated in the notes, the names used are taken because they are well known and distinctly limited, and hence less subject to be misconstrued.

The columns have been lettered and the names in each numbered for convenience of reference. Between the principal divisions a line has been left for transition beds, when such have been identified, but the absence of a name on this line in any column is not to be taken as necessarily indicating a gap in the series. While the names have been arranged as nearly as possible in their supposed stratigraphic order, there are some cases when more than one name may occur in the same column which further researches will show to belong to horizons of the same age but geographically separated.

## THE MAIN DIVISIONS.

A few preliminary remarks on the main divisions may be made here.

### PLEISTOCENE.

For the purposes of this table the Pleistocene epoch is that which began with the Glacial period and ended with the Champlain period, as adopted by Dana.[1] The present or Recent epoch is excluded, and the Pleistocene horizons mentioned are merely illustrative, as this division is extralimital to this paper. It may be added that nothing so far observed in the geology of the Gulf and northern Antillean region lends any support to the hypotheses advanced by Dr. J. W. Spencer in regard to enormous Pleistocene elevations or depressions in this region.[2] On the contrary, the observed conditions appear to be quite incompatible with such extreme changes of level during that epoch.

### PLIOCENE.

The Pliocene epoch in America has been defined by me[3] as beginning with the culmination of the Miocene uplift which connected the American continents and joined the island of Florida to the mainland of Georgia. It terminated, according to this definition, at the beginning of the Glacial period. With the Pliocene there began a slight depression of the continental border, which was followed by an elevation to which are due the perezonal products of erosion known as the Lafayette formation, and in a broad way perhaps also the inception of the Glacial period itself. Everything prior to the period of glaciation, after the culmination of the Miocene uplift, is regarded as Pliocene. The Pliocene for the Atlantic and Gulf coasts was a period of milder climate than the preceding Miocene, a condition perhaps due to changes in the direction of the ocean currents permitted by the initial subsidence.

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[1] Manual of Geology, 4th edition, 1895, p. 940.

[2] Reconstruction of the Antillean continent: Bull. Geol. Soc. Am., Jan., 1895, Vol. VI, pp. 103–140.

[3] Trans. Wagner Inst. Sci., Phila., Vol. III, 1891, p. 208, and Bull. U. S. Geol. Survey No. 84, 1892, p. 18.

## MIOCENE.

The Miocene was a period of elevation, and, as here defined, began when the elevation reached a point sufficient to deflect the warm waters of the Gulf Stream far enough offshore to admit of the invasion, between them and the coast, of colder northern waters, bringing a cool-temperate fauna south to the Suwanee Strait, between the island of Florida and the continent, and westward to the Mississippi embayment. At this time the Americas were not connected, though the subsequent movement in elevation brought about a connection during this epoch, and perhaps long before its close. The Miocene as here understood ended with the culmination of the upward movement, which, according to Gabb, produced an elevation of several thousand feet in some parts of the Central American region, closed the Suwanee Strait, and finally gave the continent, in the main, its present outlines.

In American and Antillean geology until recently the term Miocene has been used, as it formerly was in Europe, to cover a series of beds which are now recognized as having closer faunal and physical relations with the Eocene than with the Miocene, as at present restricted by European geologists. These beds have recently been relegated to their proper place, as indicated by their relations to analogous European horizons,[1] and shown in the accompanying table. In using the term "Miocene" for the older beds of the Chipola epoch, which have a subtropical fauna, as well as for the cool-temperate Chesapeake, earlier work has been in harmony with the existing literature of Antillean geology. I had for some time felt that the arrangement was objectionable, since the faunal break between the two is the sharpest in the whole Tertiary series of our Gulf border, and the change is followed by the most important post-Cretaceous uplift which has affected the Gulf, Antillean, and Central American region.

As soon as an attempt was made to correlate this supposed older Miocene with the European horizons the testimony of the vertebrate faunas was added to that of the invertebrates, and the conclusion was inevitable that these beds should be referred to the Middle and Upper Oligocene, and that only those which had before been referred to the Yorktown epoch, under the name of Chesapeake, represent the typical Miocene of modern European geology.

So far as yet known the deposition of the Miocene upon the Oligocene in our Southern States is not marked by any stratigraphic break of importance, though a nearly total change in the fauna and striking lithological differences reflect important physical changes which must have taken place elsewhere.

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[1] See prefatory remarks to Descriptions of Tertiary fossils from the Antillean region, by R. J. L. Guppy and W. H. Dall: Proc. U. S. Nat. Mus., Vol. XIX, No. 1110, 1896, pp. 303–305.

## OLIGOCENE.

The Oligocene in America dates from the disappearance, at the end of the Jacksonian, of such forms as Zeuglodon and *Venericardia planicosta*, includes the whole of the Vicksburgian and nummulitic beds of Heilprin and the succeeding subtropical faunas, and ends with the climatic change which permitted the invasion of the Gulf region by the more northern fauna of the Chesapeake Miocene. It is a period of intense organic life, especially of the lower calcareous foraminifera and other invertebrates, whose remains in the Lower Oligocene alone were deposited in rather deep water to a thickness of thousands of feet.

Between the typical (Chesapeake) Miocene and the Upper Oligocene there occur in the West beds containing an intermediate vertebrate fauna, according to Cope and W. B. Scott. A strictly analogous fauna is found at Sausans, Department of Gers, in southwest France, according to Scott. A tendency toward transition is recognizable in the Oligocene fauna of the marine Alum Bluff beds and the Oak Grove sands in western Florida and the Ashley River marl of South Carolina.

The Oligocene proper forms a continuation of the Eocene of the Gulf border, without essential change in the manner or character of deposits, except that which is brought about by marked changes in the fauna. They are essentially composed of clay, lime, and sand, variously consolidated, mixed, or modified. None of our Eocene limestones is of purely organic origin like those of the Lower Oligocene, but the Upper Oligocene beds resume an Eocene facies, lithologically. The whole may be conveniently divided into two parts:

1. Upper Oligocene, Chipola epoch { Chipola formation. / Chattahoochee formation.
2. Lower Oligocene, Vicksburg epoch { Ocala formation. / Vicksburg formation.

Between the latter and the typical Eocene are sundry transitional beds.

### UPPER OLIGOCENE.

The Upper Marine Oligocene is differentiated from the Lower by the change in the character of the sediments, which are those of more shallow water, and are consolidated into more compact limestones and marls, with a much more general silicification. The fauna also changes, especially by the disappearance of the more conspicuous foraminifera and a majority of the species of invertebrates of higher type, though most of the molluscan genera persist. There is no marked break in the stratigraphy between the Upper and Lower beds so far as yet observed.

### LOWER OLIGOCENE.

The Lower Oligocene of the coast begins in the United States with the disappearance of such types as *Venericardia planicosta* and the Zeuglodon and the invasion of the fauna by multitudes of foraminifera, especially Nummulites, Orbitoides, and numerous peculiar vertebrates and mollusks. It was a period of profuse invertebrate life and steady sedimentation, especially of oceanic deposits in water of not always great depth. Some 2,000 feet of strata, formed almost wholly of organic débris, were deposited in the peninsular region of Florida. The Lower Oligocene terminated with the elevation of the island of Florida and probably part of the Antillean land above the sea level. The change is marked by the disappearance of most of the larger foraminifera from the fauna, together with many of the larger mollusks. In the southeastern United States there is no marked stratigraphic break between the Eocene and Oligocene. Many of the fossils persist into the upper beds, but the character of the fauna as a whole undergoes a well-marked alteration, showing that physical changes of some sort, such as would profoundly affect the marine fauna, must have taken place. The change by which the Oligocene was brought to a close and the typical Miocene inaugurated, caused, as already described, the most remarkable faunal break in the geological history of the United States after the Cretaceous. The proofs are conclusive of a decided change for the colder in the temperature of the sea, driving away the warm-water Oligocene fauna and admitting a cold temperate fauna in its place. The reaction which took place at the end of the Miocene was only partial, and many of the Miocene types still persist on the Gulf coast.

The intermediate character of the Vicksburg horizon as between the Eocene and the Chesapeake Miocene was early recognized by Conrad,[1] whose keen eye also noted the analogy between the Santo Domingo and Vicksburg faunas. His comparison was defective, as claimed by Gabb,[2] but this was because he was comparing the Lower Oligocene of Vicksburg with the Upper Oligocene of Haiti, and not because his acumen was at fault. The Vicksburg deposits were first referred to the Oligocene (after that term had been invented by Beyrich in 1853) by Conrad,[3] who indicated their equivalence to the "Older Miocene" of McCoy, and continued to regard the Santo Domingo beds as of the same age. The discovery of the Chattahoochee group by Langdon[4] and of other related beds by Burns and Johnson, and the discrimination of their fauna from that of the typical Miocene by the writer, for the first time provided the data[5] by which the Haitian deposits could be correlated with those of the mainland of North America.

The results of a more general correlation, as now summarized in the accompanying tables, show, among other things, the existence in the Gulf States of a well-defined series of Oligocene beds, including not

[1] Proc. Acad. Nat. Sci. Phila. for 1852, pp. 198-199.

[2] Geology of Santo Domingo, 1873, p. 97.

[3] Check list of the invertebrate fossils of North America, Eocene and Oligocene: Smithsonian Miscell. Coll. No. 200, 1866, pp. IV, 26, 37.

[4] Am. Jour. Sci., 3d series, Vol. XXXVIII, 1889, p. 324.

[5] See Trans. Wagner Inst. Phila., Vol. III, Parts I-II, 1890-94, and Bull. U. S. Geol. Survey No. 84, 1892, pp. 101-127.

only the Vicksburgian of Conrad and the nummulitic of Heilprin,[1] already recognized as probably Oligocene, but also the whole series of warm-water "Old Miocene" included in the Chattahoochee group of Langdon and the Tampa group of Dall.[2] Gabb, in his Geology of Santo Domingo, questions Conrad's conclusion, and estimates that one-third of the Santo Domingo species are still living; but a review of his collection at Philadelphia has not sustained this conclusion. His study of the fossils seems to have been hurried, and many of his identifications are unsound when they relate to recent forms. As might be expected, many of the recent species are represented by ancestral forms differing by a moderate differentiation from their descendants; but this differentiation is constant, and they must be regarded as distinct species. Many of the recent representatives are now found only on the Pacific coast, and their extinction in the Caribbean area is doubtless coincident with the disturbances which by elevating the Oligocene deposits united the two continents during the Miocene epoch. The faunal correlations are reached in a somewhat roundabout way. The Trinidad section is stratigraphically quite complete, but the fossils are in poor condition. They are, however, sufficiently recognizable to enable the identification of a number of species in the San Fernando or Naparima beds to be made with those of the Gatun beds of the Isthmus of Darien, while the latter also contain enough Tejon or Claiborne species to settle their place in the Alabama column.

The Oligocene Caroni beds of Trinidad are similarly connected with the Jamaica and Haitian Bowden beds, to the west; and these with the Monkey Hill and Mindi beds along the line of the Panama Canal, to the south; and, to the north, by a fair number of identical species with the Tampa group of Florida. The typical Vicksburgian is recognizable in the Guallava sandstone of Costa Rica. Lastly, the fossils of the deep artesian well of Galveston, strongly tinctured with a faunal element which in the recent fauna is Pacific in distribution, are without doubt Upper Miocene, and show that the exclusion of the Pacific fauna and the extermination of its Caribbean representatives was not complete until about the end of the true Miocene.

## EOCENE.

The Eocene, as everywhere recognized in American geology, follows the great disturbance which terminated the Mesozoic epoch, and, in its typical form, ends with the faunal changes already alluded to as ushering in the Lower Oligocene. In a wide sense it includes both Eocene and Oligocene of the present table, the two not being separated by essential stratigraphic breaks in the Gulf column or by changes in the climatic relations of the fauna.

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[1] Trans. Wagner Inst. Phila., Vol. I, 1887, p. 4.
[2] Bull. U. S. Geol. Survey No. 84, 1892, pp. 112, 335.

## SUBJACENT MESOZOIC BEDS.

The end of Mesozoic time, according to Dana,[1] is marked by the climax of an epoch of mountain building, which began toward the close of the era; and by the character of the fauna, which to this point comprises among mammals only oviparous or nonplacental types; among reptiles, dinosaurs and mososaurs, or closely related forms; and among invertebrates, such types as Ammonites, Baculites, Inoceramus, and Rudistes. This definition accords with the views of paleontologists generally, in all countries. It has been questioned by Cross,[2] in discussing the age of the Denver and Arapahoe formations of Colorado, on the ground that, notwithstanding the persistence of horned dinosaurs (Ceratopsidæ) in these beds, the evidence of stratigraphy shows that, at their base, a nonconformity exists between them and the Laramie, representing the erosion of a considerable mass of strata; that, between the deposition of the Arapahoe and the Denver, volcanic outbursts on a large scale had occurred, such as generally mark the inauguration of the Tertiary epoch; that the evidence of the plant remains, and some of the fresh-water mollusks indicates a transition between those of the Laramie and the Eocene Fort Union beds; that the existence of a similar succession of post-Laramie deposits in other localities[3] shows that the conditions were widespread, and not merely local; and that at present there is no distinct evidence of an especially important earth movement succeeding the Denver formation, while that preceding the Arapahoe at present appears to have been of the character and magnitude usually assumed for the disturbance closing the Mesozoic. He therefore queries whether the dinosaurian fauna may not have been a survival into the Eocene of a special group of Mesozoic types which for some reason did not find the orographic disturbances alluded to inimical to their continued development.

It is clear that the question is one of definition and limitation—of interpretation rather than of fact.

The movement in elevation, the climax of which is taken as indicating the end of the Mesozoic epoch, must have begun and continued for a long period prior to that culmination, and was marked by a gradual recession of the sea, which at the close was definitely excluded from the greater portion of the area covered by Mesozoic deposits. We can not suppose that after the culmination of the movement the sea covered any portion of the Cretaceous area away from its borders; yet 200 feet above the great unconformity in the Livingston formation, which Cross has shown to be the direct analogue of the Denver, Weed found beds of oyster shells, which are a positive proof not

[1] Manual of Geology, 4th edition, 1895, pp. 815–16.

[2] Whitman Cross, On the age of the Arapahoe and Denver formations: Mon. U. S. Geol. Survey, Vol. XXVII, 1896, p. 206.

[3] The Animas River beds of Colorado and New Mexico, the Ruby beds of Grand River, Colorado, and the Livingston formation of Montana are cases in point.

only that the sea had been near by during the development of the nonconformity, but that the conditions had not been such as to exterminate its fauna within that distance, and hence that the culmination of the uplift could not then have been reached. The persistence of the dinosaurs in the Denver is also conclusive evidence that the changes, however great, had not at that time been of a character to interrupt the development of a type which, from its size and peculiarities, must have been very susceptible to differences in the environment, and which a little later did disappear from the fauna before the introduction of unmistakable Eocene types.

The physical evidence, therefore, important as it is, does not seem conclusive that the climax of change had been reached at the end of the Laramie, though it may have been well on toward culmination, and the Denver formation in this table will therefore be left where Cross has left it, to form the uppermost or transitional portion of the Mesozoic column.

In the American Journal of Science for October, 1891, Vol. XLII, Pl. XII, Prof. O. C. Marsh printed a table (also issued separately) in which he indicated the geological horizons of vertebrate fossils. This list, with some amplifications, appeared again in Johnson's Cyclopædia, second edition, Vol. VIII, 1895, p. 491. The following is that part of the table which refers to the succession in the Tertiary:

| | |
|---|---|
| Pliocene | Equus beds.<br>Pliohippus beds. |
| Miocene | Miohippus beds.<br>Oreodon beds.<br>Brontotherium beds. |
| Eocene | Diplacodon beds.<br>Dinoceras beds.<br>Heliobatis beds.<br>Coryphodon beds. |
| Cretaceous (uppermost) | Laramie series or Ceratops beds. |

As these terms are intended to indicate the order of succession, and are not accompanied by stratigraphic data, it has not been practicable to utilize the names in the present attempt at correlation.

It seems proper to note that the writer could not attempt to judge between the divergent views of European authorities as regards the limits and nomenclature of the various European horizons. The names most generally accepted, as formulated by Mayer and adopted by G. F. Harris in 1891, have been used with a few modifications, and effort has been concentrated upon the task of a correct correlation of the American horizons with them as thus understood. The delay in printing this paper has led to various emendations and additions to the original manuscript. Quite recently the writer has had an opportunity to study a paper by Dr. F. Sacco[1] on the classification of the European Tertiary,

[1] Congrès géologique international: Compte rendu de la sixième session, 1894, pp. 309-320, and table 1897.

*Correlation table of Tertiary formations; data to 1895; by W. H. Dall.*

| A. | Epochs and stages. | B. | Pacific coast. | C. | Lake beds. | D. | Gulf States. | E. | Atlantic States. | F. | Foreign analogues. |
|---|---|---|---|---|---|---|---|---|---|---|---|
| | | 1 | Kowak clays | | | 1 | Columbia | 1 | Columbia | | |
| 1 | PLEISTOCENE | 2 | Ground ice | 1 | Equus beds | 2 | Cornfield Harbor clays | 2 | Simmons' Bluff beds | 1 | Tehuelche. |
| | | 3 | San Pedro beds | | | | | | | | |
| | | 4 | Coos conglomerate (?) | | | 3 | Reynosa limestone | 3 | Lafayette | | |
| | | 5 | Mytilus beds | 2 | Blanco | 4 | De Soto beds | 4 | Gay Head sands | 2 | Astian. |
| 2 | PLIOCENE | 6 | Merced group | | | 5 | Limon clays | 5 | Croatan beds | 3 | Crag of Britain. |
| | | 7 | San Diego beds | | | 6 | Caloosahatchie beds | 6 | Waccamaw beds | | |
| 3 | *Transitional* | 8 | Auriferous gravel | 3 | Palo Duro | | | 7 | Duplin beds | 4 | Congeria beds. |
| | | 9 | Ione formation | | | | | | | | |
| | | 10 | Sooke beds | | | 7 | Oakville beds | 8 | Gay Head gravel | | |
| 4 | MIOCENE | 11 | Empire beds | 4 | Loup Fork | 8 | Pascagoula | 9 | Chesapeake — St. Marys | | |
| | | 12 | Astoria sandstone | | | 9 | Chesapeake | 10 | Chesapeake — Jones Wharf | 5 | Helvetian. |
| | | 13 | Monterey beds | | | | | 11 | Chesapeake — Plum Point | | |
| 5 | *Transitional* | | | 5 | Deep River | 10 | Alum Bluff beds | | | | |
| | | | | | | 11 | Oak Grove sand | 12 | Ashley River marl | 6 | Sausans beds. |
| | | | | | | 12 | Tampa beds | 13 | Shiloh marls | | |
| | OLIGOCENE: | | | | | 13 | Chipola beds | | | | |
| 6 | *a.* Upper or Chipolan | 14 | Tunnel Point beds (?) | | | 14 | Upper Chattahoochee | | Bowden beds | | |
| | | 15 | Astoria shales | 6 | John Day | 15 | Altamaha grits | 14 | Monkey Hill beds | 7 | Aquitanian. |
| | | | | | | 16 | Lower Chattahoochee | | Caroni beds | | |
| | | | | | | 17 | Typical Grand Gulf | | | | |
| | | | | | | 18 | Hawthorne beds | | | | |
| 7 | *Transitional* | | | 7 | Protoceras bed | 19 | Shell Bluff group (?) | 15 | Naparima beds (?) | 8 | Tongrian. |
| | | | | 8 | White River — Upper | 20 | Ocala group | | | | |
| 8 | *b.* Lower or Vicksburgian | 16 | Aturia bed | 9 | White River — Middle | 21 | Coral limestone | 16 | Shell Bluff group (?) | | |
| | | | | 10 | White River — Lower | 22 | Vicksburg | 17 | Cooper River marl | 9 | Ligurian. |
| | EOCENE: | | | | | 23 | Red Bluff | 18 | Guallava sandstone | | |
| | | | | | | 24 | Zeuglodon beds | 19 | Santee beds | | |
| 9 | *a.* Jacksonian | 17 | Foraminiferal shales (?) | 11 | Uinta | 25 | Moodys branch beds | | | 10 | Bartonian. |
| | | | | | | 26 | Marks Mill beds | 20 | Manzanilla beds | | |
| | | | | | | 27 | White Bluff marl | 21 | Gatun beds | | |
| 10 | *b.* Claibornian | 18 | Arago beds | 12 | Bridger — Washakie | 28 | Claiborne sand | | | 11 | Parisian. |
| | | | | 13 | Bridger — Bridger | 29 | Ostrea sellæformis beds | 22 | Shark River, N. J | | |
| | | | | 14 | Bridger — Green River | 30 | Lisbon beds | 23 | Wilmington, N. C | | |
| | | | | | | 31 | Tallahatta | 24 | Orangeburg, S. C | | |
| | | | | | | 32 | Hachetigbee beds | | | | |
| | | 19 | Kenai group | | | 33 | Bashi series | 25 | Atane beds (?) | | |
| 11 | *c.* Chickasawan | | | 15 | Wasatch | 34 | Tuscahoma | 26 | *a.* Woodstock beds | | |
| | | | | | | 35 | Gregg's Landing | 26 | Pamunkey, Va., Md | 12 | Suessonian. |
| | | | | | | 36 | Nanafalia | 26 | *b.* Aquia Creek beds | | |
| | | | | 16 | | | | | | | |
| | | 20 | Tejon | 17 | Puerco — Torregon | 37 | Naheola | | | | |
| 12 | *d.* Midwayan | 21 | Martinez | 18 | Puerco — Puerco | 38 | Sucarnochee | | | 13 | Cernaysian. |
| | | 22 | Puget group | 19 | Fort Union | 39 | Midway limestones | | | | |
| 13 | MESOZOIC | 23 | Chico-Shasta | 20 | Denver-Laramie | 40 | Ripley group | 27 | Upper Green marl | 14 | Danian. |

in which some revolutionary changes have been made in the nomenclature of Mayer. It was still thought best to adhere to the latter in this paper, as more widely known, and especially because it seemed to apply to American horizons better than the modified scheme of Sacco. That author includes in his table, though taking no responsibility therefor, the classification of Ameghino for the South American Tertiaries. From the invertebrate fossils I have examined from the Patagonian region I infer that Ameghino has assigned too great an age to his whole series, putting the (Pleistocene) Belgranien in the Pliocene, the (Miocene or Oligocene) Santa Cruzien in the Middle Eocene, etc., a view which can not be sustained from these fossils.[1]

## NOTES ON THE TABLE

### PLEISTOCENE.

B 1. *The Kowak clays.*—The Kowak River of northwestern Alaska falls into Kotzebue Sound. For definition of the Kowak clays see Dall, Bull. U. S. Geol. Survey No. 84, 1892, pp. 261-268. It is in this formation that the Pleistocene mammalian remains of Kotzebue Sound and other Alaskan localities are found preserved. These clays are above the Ground ice.

B 2. *The Ground ice.*—A description of the Ground ice formation will be found under the references given for B 1.

B 3. *San Pedro beds.*—Extensive beds of unconsolidated Pleistocene sand replete with molluscan shells in very perfect condition, best exhibited at Harbor Hill, at the head of San Pedro Harbor, California. These shells are nearly all of recent species, and the same horizon is recognizable above the Pliocene of Pacific Beach, San Diego, California; at various points on the Coronado Beach peninsula opposite San Diego, especially at a cove called Spanish Bight, and also at Santa Barbara, California, and elsewhere. It is very generally unconformably underlain by Pliocene beds, which, when unconsolidated, may be discriminated by their more yellow color, due to greater oxidation. The unconformity is often of the nature of an erosion-interval, not necessarily involving any marked difference of dip, and in some cases, as in the upper part of the Merced series, it is doubtful if any real unconformity exists. See Lawson, Bull. Dept. Geol. Univ. California, Vol. I, pp. 143 et seq.; Ashley, Jour. Geol., June, 1895, Vol. III, pp. 434-454, and Proc. California Acad. Nat. Sci., 2d series, Vol. V, Aug., 1895, pp. 275-367.

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[1] Since the above was written, several important papers have been published, by Messrs. J. B. Hatcher, A. E. Ortmann, and H. A. Pilsbry, in which the stratigraphy of southern Patagonia has been revised and corrected, and the inferences above mentioned are wholly confirmed. The following table will indicate the succession of the Tertiary as established by Hatcher's observations in this region.

| | |
|---|---|
| Pleistocene | Tehuelche formation.<br>Piso Belgranense (Ameghino). |
| Pliocene | Cape Fairweather beds. |
| Miocene | Santa Cruz beds (fresh water).<br>Supra-Patagonian (marine). |
| Eocene ? | Patagonian beds (marine). |

For details the reader is referred to the following publications:

Hatcher, J. B., Am. Jour. Sci., Sept., 1897, 4th series, Vol. IV, pp. 246-248, and the same for Nov., 1897, pp. 327-354.

Ortmann, A. E., Am. Jour. Sci., Nov., 1897, 4th series, Vol. IV, pp. 355-356.

Pilsbry, H. A., Proc. Acad. Nat. Sci. Phila. for 1897, pp. 329-330.

B 4. *Coos conglomerate.*—A singular formation at Fossil Point, Coos Bay, Oregon, probably of Pleistocene age. It consists chiefly of Miocene fossils, small waterworn chert pebbles, sand, and a few fossil forms still found living in the vicinity, cemented into a hard conglomerate, the whole lying upon an eroded surface of the Empire (Astoria) sandstone, with which it agrees in dip. Only fragments of the original deposit remain, the rest having been eroded, though originally some 10 feet thick. The Miocene fossils were derived from the Empire beds and still contain traces of their original matrix. The source of the chert is unknown. The only portion of the conglomerate which may be regarded as strictly contemporaneous with its deposition comprises the rather small proportion of Pleistocene fossils and sand which is intermixed with the rehandled older material.

C 1. *Equus beds.*—From the presence of remains of the horse as a characteristic fossil. See Gilbert, Mon. U. S. Geol. Survey, Vol. I, 1890, p. 393.

D 1. *Columbia sands and loams.*—For a definition and limitation of the Columbia formation see McGee, Am. Jour. Sci., 3d series, Vol. XXXV, 1888, p. 448.

D 2. *Cornfield Harbor clays.*—These beds on Cornfield Harbor, near Federalsburg, Maryland, contain an interesting marine fauna, including Rangia and Cyrena, both unknown in the recent fauna north of the Gulf of Mexico. They have generally been referred to the Pleistocene, but it is possible that further research will show them to be upper Pliocene, like the Gay Head sands. They appear to be represented near Lake Ponchartrain, but might with equal propriety have been put in the next column.

E 2. *Simmons Bluff beds.*—The Pleistocene of Simmons Bluff, South Carolina, carries a rich and finely preserved marine fauna.

F 1. *Tehuelche formation.*—The "Pampean" formation of the southern part of South America has been loosely treated in the literature and as the term was originally used comprised without discrimination horizons of very different geological age. M. Ameghino, according to Sacco (Congrès géol. int; Compte rendu de la 6me sess., 1894, p. 320) restricts the term to a group of four subdivisions which he places in the Pliocene, those portions which he regards as Miocene being included under the name Araucanien. The three upper subdivisions of this Pliocene Pampean, from their marine fossils, appear to be incontestibly Pleistocene. The so-called "Pampean" mammals began their migration to North America in Miocene time when the elevation of Central America had reached a point where it afforded them a passage. Bones of some of them occur under a marine Pliocene limestone in Florida, and in the beds of the rivers they are mixed with Pleistocene forms, so that the lists of species from this region include forms as old as the Loup Fork, and as late as the Equus beds. The immigration was doubtless more or less continuous from the later Miocene until it was checked by Pleistocene climatic changes. The marine mollusks of the "piso belgranense," forming part of the Pampean, have been catalogued by Dr. H. von Ihering,[1] and form an unmistakably Pleistocene assembly. See also the note on this subject, page 335.

## PLIOCENE.

B 5. *Mytilus beds.*—From the presence of Mytilus as a characteristic fossil, in western Oregon. See Condon, Am. Naturalist, Vol. XIV, 1880, p. 457; and Dall, Bull. U. S. Geol. Survey No. 84, 1892, p. 228. Beds of Shoalwater Bay, Oregon.

B 6. *Merced group.*—From the Merced Cliffs of the Seven-Mile Beach of the San Francisco peninsula, extending northward from near Mussel Rock on the coast, and southward in the interior. Named by Lawson in 1893. See Bull. Dept. Geol. Univ. California, Vol. I, 1893, No. 4, pp. 115 et seq. A correlative

[1] Science, April 19, 1895, new series, Vol. I, pp. 421–422.

series in the Wildcat district, Humboldt County, northern California, was subsequently named the Wildcat series by Lawson. See the same Bulletin, Vol. I, 1894, No. 8, p. 255; and Ashley, Proc. California Acad. Nat. Sci., 2d series, Vol. V, 1895, pp. 312–331.

There are several rather characteristic Miocene species enumerated in the lists of the fauna of the Wildcat series, which, if correctly identified, are perhaps derived from the eroded Miocene beds on which the Wildcat series rests. The upper part of the beds included in the Merced series, at the typical locality, has a fauna in which few if any extinct species were detected by the writer during an examination of the section in September, 1897. The impression derived from this examination was that this portion of the Merced series is probably Pleistocene, though strictly conformable with the lower beds.

B 7. *San Diego beds.*—Pliocene sandy marls exhibited in the well explored by Hemphill in the city park; below the Pleistocene at Pacific Beach, San Diego, and on the shores of False Bay near by; and on the Coronado peninsula opposite San Diego. The same horizon crops out in various places northward, especially at Dead Mans Island and Harbor Hill, San Pedro, and on the coast of Todos Santos Bay, Lower California. See references under B 6, and also Dall, Proc. California Acad. Nat. Sci., Vol. VI, 1874, pp. 227 et seq.; and Proc. U. S. Nat. Mus., Vol. I, 1878, pp. 10–30.

C 2. *Blanco formation.*—Rio Blanco of Texas. See Dumble, Jour. Geol., Vol. II, No. 6, 1894, p. 559.

D 3. *Reynosa limestone.*—Reynosa, Texas. See Dumble, Jour. Geol., Vol. II, No. 6, 1894, p. 560.

D 4. *De Soto beds.*—Pliocene lake De Soto of Central Florida. See Dall, Bull. U. S. Geol. Survey No. 84, 1894, pp. 133, 324.

D 5. *Limon clays.*—Port Limon, Costa Rica. See Gabb, Jour. Acad. Nat. Sci. Phila., 2d series, Vol. VIII, 1881, p. 349. A deposit of Pliocene shell-bearing clays and sands with beautifully preserved fossils.

D 6. *Caloosahatchie beds.*—Pliocene marls of the Caloosahatchie River, Florida. See Dall, Am. Jour. Sci., 3d series, Vol. XXXIV, 1887, pp. 167, 169, and Trans. Wagner Free Inst. Science, Vol. III, parts 1 and 2.

E 3. *Lafayette formation.*—From Lafayette County, Mississippi. For an account of the character and synonymy of this formation, see Am. Geologist for August, 1891, pp. 129–131, or Bull. U. S. Geol. Survey No. 84, p. 328. A Pliocene fluvial formation, perhaps partly overlapping with the Glacial period and resulting from a comparatively rapid Pliocene uplift in the Appalachian region. A considerable part of what has been included under this name may prove of a different origin and age, and the whole formation requires detailed examination to define its exact extent, but the wide distribution and distinctive character of the typical Lafayette are beyond dispute.

In New Jersey what was first regarded as a northeastern extension of the Lafayette has been described under the name of the Pensauken formation. See R. D. Salisbury in Ann. Rept. State Geol. N. J. for 1894 (Part I, Surface Geology), 1895, pp. 105–118. More recent study leads him to doubt this correlation.

E 4. *Gay Head sands.*—Pliocene sands with marine fossils of Gay Head, a bluff at the southwest extreme of the island of Marthas Vineyard, Massachusetts. See Dall, Am. Jour. Sci., Oct., 1894, 3d series, Vol. XLVIII, p. 299.

E 5. *Croatan beds.*—Croatan, North Carolina, Pliocene marls. See Dall, Trans. Wagner Free Inst. Science, Jan., 1892, Vol, III, part 2, pp. 201–217.

E 6. *Waccamaw beds.*—Waccamaw River, South Carolina, Pliocene marls. See Dall, Trans. Wagner Free Inst. Science, 1892, Vol. III, part 2, pp. 201–217.

F 2. *Astian* (typical).—Yellow sands of Asti, Italy. The culminating member of the Pliocene, usually correlated with the Red Crag of Britain. See G. F. Harris, in Newton, Systematic list of the Edwards collection in the British Museum, 1891, p. 327, and table. This work will hereafter be cited in these notes as "Harris, Table of 1891."

F 3. *Crag of Britain.*—The correlation of the special horizons of the British Crag with the various Pliocene beds of the New World is not yet practicable. In general the whole crag series is probably comparable with the whole of the American Pliocene below the Lafayette.

## TRANSITIONAL BEDS.

B 8. *Auriferous gravels.*—The auriferous gravels of the Sierra Nevada region of California as a whole may cover a considerable part of the Tertiary column. While part of them are Pliocene a large part are probably Miocene, and they are said by Diller to be laterally continuous with the Ione formation of Lindgren, which is chiefly Upper Miocene. See Whitney, Mem. Mus. Comp. Zool., Vol. VI, part 1, 1877, p. 283; Dall and Harris, Bull. U. S. Geol. Survey, No. 84, 1892, p. 321; the text of Geologic Atlas U. S., folio 5, Sacramento, California, 1894; Turner in Fourteenth Ann. Rept. U. S. Geol. Survey, Part II, 1894, p. 462; and Lindgren, Age of the Auriferous Gravels: Jour. Geol., November-December, 1896, Vol. IV, pp. 881-906.

C 3. *Palo Duro beds.*—These beds, identified in western Texas by Scott as transitional, have also had the absurd name of "Goodnight beds" applied to them. See Dumble, Jour. Geol., Vol. II, No. 6, 1894, p. 559.

E 7. *Duplin beds.*—Late Miocene marls of Duplin County, North Carolina. See Conrad, Am. Jour. Sci., 1st series, Vol. XII, 1841, pp. 335-343.

F 4. *Congeria beds.*—Some at least of the various "Congerien schichten" of eastern Europe seem to be of a transitional character. See Harris, Table of 1891.

## MIOCENE.

B 9. *Ione formation.*—From the town of Ione, Amador County, California; named by Lindgren in the text to Geologic Atlas U. S., folio 5, Sacramento, California, 1894. See note under B 8; also Diller, Fourteenth Ann. Rept. U. S. Geol. Survey, 1894, pp. 403-433.

B 10. *Sooke beds.*—From Sooke Inlet, on the southern coast of Vancouver Island. Miocene beds containing a peculiar fauna, probably later than the Empire or Astoria Miocene. See J. C. Merriam, Bull. Dept. Geol. Univ. California, Vol. II, No. 3, 1896, pp. 101-108.

B 11 and 12. *Empire beds.*—From typical exposures near Empire City, Coos Bay, Oregon. See Diller, Seventeenth Ann. Rept. U. S. Geol. Survey, Part I, 1896, pp. 475-476. These beds are well exposed between Pigeon Point and Fossil Point, 3 miles southwest of Empire City, and abut unconformably upon the Oligocene beds at Coos Head and the Eocene at Marshfield in the same region. At Fossil Point they are overlain by the richly fossiliferous Pleistocene beds of the Coos conglomerate. So far as can be judged from the fauna collected, the Empire beds are the exact equivalent of the upper part of the Miocene beds at Astoria, called by Dall, in 1892, the Astoria sandstones, to distinguish them from the (Oligocene?) Astoria shales (formerly called Miocene), which conformably underlie them. The exact place of the Astoria shales in the column must await a better knowledge of the fauna. If the double use of the name Astoria in this manner is regarded as objectionable, the name Empire beds might be taken for the sandstones.

B 13. *Monterey beds.*—The Monterey beds are mostly composed of soft shales, often of a whitish color, with many infusoria and few but usually well preserved molluscan fossils, and a certain proportion of sandstones. The typical outcrops are those of the southeastern part of Monterey Bay, but they are widely extended elsewhere; in the Santa Cruz mountains, at Carmel Bay, etc.

Portions of this group have been named the Carmelo series, Pescadero series, and it seems to include parts of the San Francisco sandstone; but the sequence and relations of the different Miocene beds south of the Golden Gate are still in need of much elucidation, and can not be clearly made out until the fauna is more thoroughly studied. In confused regions, such as this, a lithologic study alone is entirely insufficient, and dependence upon it can only lead to greater confusion. It would be well if the application of names to minor local series of rocks could be postponed until the geology is worked out for the general succession of the beds of this part of the coast. See Lawson, Bull. Dept. Geol. Univ. California, No. 1, 1893; No. 4, 1894; Ashley, Proc. California Acad. Nat. Sci., 2d Series, Vol. V, 1895, pp. 273-367; Lawson, Am. Geologist, June, 1895, Vol. XV, pp. 342-356; and Whitney, Geol. Surv. California, Geology, Vol. I, 1865, pp. 61-166.

C 4. *Loup Fork group.*—From the Loup Fork of Platte River, central Nebraska. See Bull. U. S. Geol. Survey No. 84, 1892, pp. 293-298, 329; and Dana, Manual of Geology, 4th edition, 1895, p. 889.

D 7. *Oakville beds.*—From Oakville, Live Oak County, Texas. These form part of the assemblage in Texas formerly included under the name of the Fayette beds, but which further study has shown to be heterogeneous. See Bull. U. S. Geol. Survey No. 84, 1892, pp. 174-5, and Dumble, Jour. Geol. Vol. II, No. 6, 1894, pp. 552-3, 556.

D 8. *Pascagoula clays.*—From the Pascagoula River, Jackson County, Mississippi. See Dall and Stanley-Brown, Bull. Geol. Soc. America, Vol. V, 1894, p. 168; and E. A. Smith, Report on the Geology of the Coastal Plain of Alabama, 1894, pp. 91-97.

D 9. *Chesapeake formation.*—Chesapeake Bay, Maryland, around which the formation is exhibited. See Dall and Harris, Bull. U. S. Geol. Survey No. 84, 1892, p. 123; and Geol. Mag. for June, 1891, p. 287.

Prof. R. D. Salisbury, in Annual Report State Geologist New Jersey for 1893 (Part I, Surface Geology), 1894, pp. 39-52, and same for 1894, pp. 99-104, has described certain gravels, under the name of the Beacon Hill formation, which he correlates with the Miocene and describes as physically analogous to the southern Chesapeake, though the analogy between a nearly pure siliceous gravel and the Chesapeake marls is not obvious. Until the unpublished evidence in Professor Salisbury's hands is available, a satisfactory correlation of these gravels is impracticable.

E 8. *Gay Head gravels.*—Gay Head, Marthas Vineyard, Massachusetts. See Dall, Am. Jour. Sci., 3d Series, Vol. XLVIII, 1894, p. 299.

E 9-10-11. *Beds of St. Marys, Plum Point, and Jones Wharf.*—Localities on the Maryland Miocene section. See Harris Am. Jour. Sci., 3d Series, Vol. XLV, 1893, pp. 21-31. Deposits of the Yorktown epoch of Dana (1863), Miocene of Maryland and Virginia.

F 5. *Helvetian.*—Mayer-Eymar, Class. des terraines tertiares, 1884. See Harris, Table of 1891, pp. 327, 329 It is probable that the Langhian and Tortonian are also covered by the American Miocene series.

## TRANSITIONAL BEDS.

C 5. *Deep River beds.*—Ticholeptus beds of Cope, Am. Naturalist, Vol. XX, 1886, p. 367. See Bull. U. S. Geol. Survey No. 84, 1892, pp. 297, 336. The name Ticholeptus, as a generic designation, is said to be preoccupied, and the geographical term Deep River beds has been proposed by Scott, but it is now stated that the correct name of the stream is Smith River, and this, if admitted, will probably take precedence. The horizon is the equivalent of the beds of Sausans (Gers) in France, Steinheim in Wurtemburg, and the Dinotherium sands of Bavaria, according to Scott.

D 10–11. *Alum Bluff beds and Oak Grove sands.*—Fossiliferous beds of Alum Bluff, Apalachicola River, and Oak Grove on the Yellow River, in western Florida. See Dall and Stanley-Brown, Bull. Geol. Soc. America, Vol. V, 1894, pp. 165–167.

E 12. *Ashley River marl.*—Phosphatized marls of the Ashley River, South Carolina, the upper member of the Ashley and Cooper marls of Tuomey. See Dall, Am. Jour. Sci., 3d series, Vol. XLVIII, 1894, pp. 300–301.

F 6. *Sausans beds.*—See note under C 5.

## EOCENE AND OLIGOCENE.

### *a.* UPPER OLIGOCENE OR CHIPOLAN STAGE.

B 14. *Tunnel Point beds.*—Shales and sandstones at and southwest of Tunnel Point, Coos Bay, Oregon, containing a sparse fauna, distinct alike from the Eocene shales below and the Empire beds above; conformable with the former but not with the latter. The faunal change and the aspect of the species suggest that this group of beds, which is about 1,200 feet in thickness, may represent the Oligocene in this section. Observations by Dall in 1897 are in preparation for publication.

B 15. *Astoria shales.*—Typical locality at Astoria, Oreg. These shales are in part sandy, and when the whole fauna is known may be differently limited. See Condon, Am. Naturalist, Vol. XIV, 1880, p. 457.

C 6. *John Day beds.*—John Day basin, Oregon. See Scott, Science, new series, No. 42, 1895, p. 499. Exact European equivalent is the horizon of St. Géraud le Puy, according to Scott.

D 12. *Tampa beds.*—Tampa City, Fla. See Dall, Bull. U. S. Geol. Survey No. 84, 1892, p. 112.

D 13. *Chipola beds.*—Chipola River, west Florida marls. See Dall, Bull. U. S. Geol. Survey No. 84, 1892, p. 112.

D 14. *Upper Chattahoochee beds.*—See note under D 16.

D 15. *Altamaha grits.*—Altamaha River, Georgia. See Dall, Bull. U. S. Geol. Survey No. 84, 1892, p. 81.

D 16, 17. *Chattahoochee beds.*—Chattahoochee River, Florida. See Dall, Bull. U. S. Geol. Survey No. 84, 1892, pp. 106–107, 112, and Dall and Stanley-Brown, Bull. Geol. Soc. America, Vol. V, 1894, pp. 164, 170.

D 18. *Hawthorne beds.*—Hawthorne, Alachua County, Florida. See Dall, Bull. U. S. Geol. Survey No. 84, 1892, p. 107.

E 13. *Shiloh marls.*—Marls of Shiloh and Jericho, New Jersey. See Dall, Bull. U. S. Geol. Survey No. 84, 1892, pp. 40–41.

This marl was deposited in troubled waters under the influence of strong currents, and contains both Oligocene and Miocene species, but which, it is probable, were not contemporaneous. The evidence is conclusive of the conditions under which the marls took their present shape, and the discrimination of the mixed species can only be assumed from the unmixed faunas of other regions.

E 14. *Bowden beds.*—The typical locality is at Bowden, Jamaica, where the marl affords a rich fauna which has until recently been always regarded as Miocene. For a description and discussion of the fauna see Guppy, Quart. Jour. Geol. Soc. London, Vol. XXII, 1866, pp. 281–294; for notes on the Haitian equivalent of the Bowden beds see Carrick Moore and Sowerby, Quart. Jour. Geol. Soc. London, Vol. VI, 1849, pp. 39–53, and also Gabb, Trans. Am. Philos. Soc., Vol. XV, 1873, pp. 83–102. The equivalency of beds at Curaçao is established by collections made by U. S. S. *Albatross*, on which no publication has been made; that of the Monkey Hill beds on the Isthmus of Darien, from collections by Rowell, Blake, and, more lately, by R. T. Hill. These isthmian beds will probably offer three horizons, which are, in ascending order: 1, the foraminiferal beds with *Orbitoides forbesii;* 2, the Gatun beds; 3, the Monkey Hill beds. The latter are those which are the equivalents of the Haitian marls, the Bowden beds of Jamaica, the Caroni beds of Trinidad, and the Bonilla beds of Costa Rica (Hill). The relative position of Nos. 1 and 2 is hardly certain, but, from the fauna, No. 2 is the equivalent of the Claiborne sands of Alabama and the Arago or Upper Tejon of California. For a discussion of the Caroni and associated beds of Trinidad see Guppy, Quart. Jour. Geol. Soc. London, Vol. XXII, 1866, pp. 281–285, 571–579, and Vol. XLVIII, Nov., 1892, pp. 519–538; also Gregory, in the same publication, Vol. LI, 1895, pp. 255–310, and Dall, Proc. U. S. Nat. Mus., Vol. XIX, No. 1110, 1896, pp. 303–305.

F 7. *Aquitanian.*—See Harris, Table of 1891, pp. 327–329. This is not the Aquitanien of Dr. Sacco, which he places in the Miocene.

TRANSITIONAL BEDS.

C 7. *Protoceras bed.*—Upper portion of the White River beds of South Dakota. See Leidy, Geol. Survey, Minnesota (Owen), 1852, pp. 539–572, and Cope, Am. Naturalist, Vol. XVIII, 1884, p. 686; also W. B. Scott, Princeton Coll. Bull., Vol. II, No. 4, 1890, p. 175.

D 19. *Shell Bluff group.*—From Shell Bluff, Savannah River, Georgia. See Conrad, Am. Jour. Sci., 2d series, Vol. XLI, 1866, p. 96, and E. W. Hilgard, in same, Vol. XLII, 1866, pp. 68–70.

E 15. *Naparima beds.*—Naparima or San Fernando series of Trinidad, West Indies. See Guppy, Microzoic of Trinidad, Quart. Jour. Geol. Soc. London, Vol. XLVIII, 1892, pp. 519–538, and Gregory, in same, Vol. LI, 1895, pp. 255–310. Probably the equivalent of the Gatun beds of Darien.

F 8. *Tongrian.*—See Harris, Table of 1891, pp. 327–329. This is the Tongrian of Mayer-Eymar and the majority of European geologists, but not of Dr. Sacco, who uses the name for a lower Oligocene group of beds, here referred, in conformity with the mass of authority, to the Ligurian.

*b.* LOWER OLIGOCENE OR VICKSBURGIAN STAGE.

A 8. *Vicksburgian.*—Vicksburg group of Warren County, Mississippi. See Conrad, Am. Jour. Sci., 2d series, Vol. II, 1846, p. 124. Owing to the number of common species of mollusks, it has for some time been questioned whether the Vicksburg and Jackson beds could be separated as distinct groups, but later researches ally the Jacksonian more intimately to the fauna of the Claiborne sands, and indicate that, when fully explored, the Jacksonian fauna will have individuality enough to retain its place as a distinct member of the Eocene.

B 16. *Aturia bed.*—See Dall, Bull. U. S. Geol. Survey No. 84, 1892, pp. 224, 321. A horizon below the Astoria shales at Astoria, Oregon, characterized by the presence of Aturia and other Eocene forms.

C 8–9–10. *White River beds.*—White River, South Dakota. See Meek and Hayden, Proc. Acad. Nat. Sci. Phila., Vol. XVIII, 1861, pp. 443–444, and Dana, Manual of Geology, 4th edition, 1894, pp. 886, 918. Also W. B. Scott, Science, 2d series, No. 42, 1895, p. 499.

D 20. *Ocala group.*—Ocala, central south Florida. See Dall, Bull. U. S. Geol. Survey No. 84, 1892, p. 331.

D 21. "*Coral limestone.*"—Salt Hill, Clarke County, Alabama. See Smith and Johnson, Bull. U. S. Geol. Survey No. 43, 1887, p. 18. When better known this will preferably take a geographic name.

D 22. *Vicksburg.*—See note under A 8.

D 23. *Red Bluff.*—Red Bluff, Wayne County, Mississippi. See E. W. Hilgard, Agric. and Geol. Mississippi, 1860, pp. 135–136.

E 16. *Shell Bluff group.*—Shell Bluff, Savannah River, Georgia. See Conrad, Am. Jour. Sci., 2d series, Vol. XLI, 1866, p. 96, and E. W. Hilgard, in same, Vol. XLII, 1866, pp. 68–70. Its exact position is still not definitely settled.

E 17. *Cooper River marls.*—Cooper River, South Carolina; wrongly united by Tuomey with the Ashley River marl under the name of Ashley and Cooper beds. See Tuomey, Geol. South Carolina, 1848, pp. 162, 211, and Dall, Am. Jour. Sci., 3d series, Vol. XLVIII, 1894, pp. 300–301.

E 18. *Guallava sandstone.*—Guallava, Costa Rica, sandstone, carrying typical Vicksburg fossils, such as *Orbitoides mantelli;* see R. T. Hill, Bull. Mus. Comp. Zool. Harvard Coll. (In preparation.)

F 9. *Ligurian.*—See Harris, Table of 1891, pp. 327–329, and note under F 8. This is the Tongrian of Sacco, but not of Mayer.

*Santa Cruz formation.*—Having received from Dr. H. von Ihering a number of fossil mollusks from the Santa Cruz formation of Patagonia,[1] it seems desirable to note here that they have a decidedly modern aspect, and if not Miocene, can hardly be referred to a horizon older than the Oligocene, though an Eocene (Parisian) age has been confidently claimed for this formation. See remarks by Ameghino and Smith-Woodward, Geol. Mag., new series, Decade IV, Vol. IV, No. 1, Jan., 1897, pp. 4–23; also this paper, p. 335.

## TYPICAL EOCENE.

### *a.* JACKSONIAN STAGE.

A 9. Jackson City, Hinds County, Mississippi, where, on Moodys Branch, a fine fossil invertebrate fauna has been collected. See Conrad, Proc. Acad. Nat. Sci. Phila., Vol. VII, 1855, p. 257, and Heilprin, in the same for 1882, p. 184.

B 17. *Foraminiferal shales(?)*—A series of beds over 1,000 feet thick, of as yet undetermined age, conformably underlying the Tunnel Point beds at Coos Bay, Oregon. See note under B 14.

C 11. *Uinta group.*—From the Uinta Mountains, Utah. See Clarence King, U. S. Geol. Expl. Fortieth Par. Atlas, 1876, and Dana, Manual of Geology, 4th edition, 1895, pp. 886, 918. The Ludian of D'Orbigny is equivalent, according to W. B. Scott.

D 24. *Zeuglodon beds.*—Cocoa post-office, Choctaw County, Alabama, recently examined by Burns and Schuchert, and presenting an individualized fauna, especially characterized by the remains of Zeuglodon, *Scala ranellina, Ostrea falco,* and other mollusks, and rather numerous brachiopods.

D 25. *Moodys Branch beds.*—Beds at Moodys Branch, Jackson, Mississippi. See A 9.

D 26. *Marks Mills beds.*—Cleveland County, Arkansas. See Harris, Geol. Survey Arkansas, Ann. Rept. for 1892 (1894), pp. 98–100, and Am. Jour. Sci., 3d series, Vol. XLVII, 1894, p. 304.

E 19. *Santee beds.*—Upper Eocene of Santee River, South Carolina. See Tuomey, Geol. South Carolina, 1848, p. 156. Tuomey included Claibornian as well as Jacksonian marls in his series. The term as adopted here refers to the upper green marls from which Zeuglodon has been obtained. The series when undisturbed would probably be divisible into several horizons corresponding to the different members of the Upper Eocene.

F 10. *Bartonian.*—Sables moyens of the Paris Basin, Barton beds of the English series, etc. See Harris, Table of 1891, pp. 327–329.

### *b.* CLAIBORNIAN STAGE.

A 10. This term as employed here comprises the series from the Orangeburg or Tallahatta to the White Bluff marls, inclusive. It is divided by Harris into upper and lower groups, the former including the Claiborne sands and the

[1] This is the Supra-Patagonian of Hatcher, with marine fossils.

White Bluff, Arkansas, marls. See Conrad, Proc. Acad. Nat. Sci. Phila., Vol. VII, 1855, p. 257; Heilprin, in the same for 1882, p. 184, and Harris, Am. Jour. Sci., 3d series, Vol. XLVII, 1894, p. 304.

B 18. *Arago beds.*—From Cape Arago, Oregon, near which they are well exposed. These beds are composed of sandstones and shales, and extend northward from Cape Arago to Cape Gregory and thence eastward to Miners Flat, on the south shore of the entrance to Coos Bay, Oregon. They have an average dip of 70° NE., and a thickness of over 3,000 feet. They contain *Cardita planicosta*, Ampullina sp., and other middle Eocene forms which suggest their correlation with the Claibornian of the Gulf column. They are apparently newer than the blackish rocks of the region about the junction of Little River and the North Umpqua, which also contain *Cardita planicosta*, but which have also a large proportion of distinct species. The name Arago beds is suggested by Mr. Diller. They appear to be older than the brackish-water deposits which inclose the Coos Bay lignites, though the latter have occasional intercalary marine layers which contain species of fossil shells apparently identical with members of the Arago fauna. The brackish-water species comprise forms belonging to Corbicula, Cyrena, and Melania or Cerithiopsis, forcibly recalling those which occur in the lignite beds of the Puget group, some of which prove to be identical.

The Umpqua beds above alluded to contain a notable number of "*Loxonema*" *turrita* Gabb, which is stated by him to be common in the Tejon, and have been referred by Diller to the Tejon group. (Bull. Geol. Soc. America, Vol. IV, 1893, p. 219.) While this disposition of them may be confirmed by a study of their fauna, and is probable, the Arago beds, which were not at first discriminated from those of the Umpqua, now appear, from the differences in their fauna, to require separation as a distinct series of beds.

See Harris on the correlation of the Tejon, in Science, Aug. 18, 1893, Vol. XXII, p. 97. There seems to be a strong probability that part of the beds heretofore referred to the Tejon represent the basal Eocene, but at present we are without sufficient evidence to speak positively.

C 12. *Washakie beds.*—Washakie Station, Sweetwater County, Wyoming. See Bull. U. S. Geol. Survey No. 84, 1892, p. 337. This and the other subdivisions of the Bridger are adopted on the authority of Prof. W. B. Scott.

C 13. *Bridger group.*—Fort Bridger, Uinta County, Wyoming. See Hayden, Prel. Rept. Geol. Surv. Colorado and New Mexico, 1869, p. 91, and note under C. 12.

For a discussion of the Huerfano Lake beds, in which they are divided into two groups, the upper beds regarded as equivalent to the Bridger and the lower beds to the Green (Wind) River or Upper Wasatch, see Osborn, Bull. Am. Mus. Nat. Hist., New York, Oct., 1897, Vol. IX, pp. 247–258. The Poison Canyon series, originally referred by Prof. R. C. Hills to the lower Eocene, are determined by Wortman and Osborn to be Cretaceous from the presence of Baculites.

C 14. *Green River beds.*—Sometimes called Wind River beds, from Wind and Green rivers, Wyoming. See Meek and Hayden, Proc. Acad. Nat. Sci., Phila., Vol. XIII, 1861, p. 447; and Hayden, Prel. Rept. Geol. Surv. Colorado and New Mexico, 1869, p. 90; also note under C 12 and 13.

D 27. *White Bluff marl.*—White Bluff, Jefferson County, Arkansas. See Harris, Ann. Rept. Geol. Surv. Arkansas for 1892, Vol. II, 1894, p. 87.

D 28. *Claiborne sands.*—Claiborne, Monroe County, Alabama. See Conrad, Proc. Acad. Nat. Sci. Phila., Vol. VII, 1855, p. 257, and E. A. Smith, Report on the Geology of the Coastal Plain of Alabama, 1894, pp. 17, 122.

D 29. *Ostrea sellæformis beds.*—Claiborne Bluff, Monroe County, Alabama. See E. A. Smith, Report on the Geology of the Coastal Plain of Alabama, 1894, p. 123.

These, with D 30, 31, are sometimes associated as the Lower Claiborne in contradistinction to D 27, 28, which form the Upper Claiborne.

D 30. *Lisbon beds*.—Lisbon Landing, Monroe County, Alabama. See E. A. Smith, Report on the Geology of the Coastal Plain of Alabama, 1894, p. 130.

D 31. *Tallahatta or Orangeburg formation*.—Hitherto known as the Buhrstone, a local term used for the rough siliceous rocks belonging to the Lower Eocene of the Carolinas, Georgia, Alabama, etc., adopted by Lyell in 1845 (Quart. Jour. Geol. Soc. London, Vol. I, p. 435) for a formation which he regarded as newer than the white limestone. The name Buhrstone had been used by Finch and others as early as 1823 (Am. Jour. Sci., 1st series, Vol. VII, p. 38), but the term is of a colloquial rather than a distinctive character. Owing to the very great objections to these qualitative mineralogical terms as names for formations, and the loose way in which the present one has been used in the literature, it has been thought best to propose a geographical name for the original group. The name selected is that of the Orangeburg district of South Carolina, the typical locality of the Buhrstone, as defined by Tuomey (Geol. South Carolina, 1848, p. 149), in correcting Lyell's stratigraphy. It was in the Orangeburg district that Tuomey found the fossils which fixed the horizon, and also the thickest and finest exposures. For the exposures of nearly the same age, but of different fauna and lithologic character in Alabama, Prof. E. A. Smith suggests the name Tallahatta, from the local name of the hills containing them.

E 20. *Manzanilla beds*.—Eocene beds of Manzanilla Point, east coast of Trinidad, which seem, from their fossils, to be probably coeval with the Gatun beds of the Isthmus of Darien and the Arago beds of the upper part of the Tejon of California. See Guppy, Quart. Jour. Geol. Soc. London, Vol. XXII, 1866, p. 572.

E 21. *Gatun beds*.—Eocene beds of Gatun, near Vamos-Vamos Station, on the line of the Panama Canal. Mr. R. T. Hill has recently brought back a collection of fossils from these beds, containing a number of Claiborne sand and Tejon (Arago) species, and fixing approximately their geological horizon. A similar fauna was found in a different matrix at Mindi Hill, on the line of the canal. See also Dall, Trans. Wagner Free Inst. Science, Phila., 1892, Vol. III, p. 233.

E 22. *Shark River beds*.—Shark River, New Jersey. See Conrad, Am. Jour. Sci., 2d series, Vol. XLVII, 1869, pp. 358–364. This, like the Eocene of North Carolina, is regarded by Harris as newer than that of Maryland and Virginia, which has been included under the name of Pamunkey by Darton. See also W. B. Clark, Prel. Rept. Cret. and Tert. Form. of New Jersey, 1893, pp. 208–210.

E 23. *Wilmington beds*.—See Clark, Bull. U. S. Geol. Survey No. 83, 1891, pp. 48–50; also note under E 22.

E 24. *Orangeburg formation*.—See note under D 31. The original Buhrstone of Tuomey.

F 11. *Parisian*.—See Harris, Table of 1891, pp. 327–330.

### c. CHICKASAWAN STAGE.

A 11. *Chickasawan formation*.—This group represents the upper part of what has been loosely termed the Lignitic or the lignite-bearing beds of the Eocene, which lie beneath the Orangeburg, Tallahatta, or Buhrstone formation. It is not the Lignitic of several authors who have applied the term to the whole or part of the strata formerly included in the so-called Laramie formation. It does not include the earliest lignite beds of the Gulf coast region, nor even the youngest of such beds. It was called the Northern Lignitic by Hilgard (Agric. and Geol. Mississippi, 1860, p. 110); Heilprin has termed it the

Eolignitic (Proc. Acad. Nat. Sci. Phila. for 1881, p. 159, footnote), a name less accurate and otherwise equally objectionable. A portion of it is believed to be included in the Camden series of R. T. Hill, of Camden, Ouachita County, Arkansas (Ann. Rept. Geol. Survey Arkansas, Vol. II, 1888, p. 49), but these beds are not characteristic and not thoroughly known. Believing that, to conform to the international rules for geologic nomenclature, it is desirable that a definite geographic term should be substituted for the present petrologic name, the subject was laid before Messrs. E. W. Hilgard and E. A. Smith, with a request that they should suggest a suitable term to be adopted.

Professor Hilgard writes, under date of December 20, 1895: "The entire Northern Lignitic is within the 'Chickasaw Purchase,' and its most characteristic and conspicuous outcrops are on the four Chickasaw bluffs, of which the Memphis bluff is the last. It would thus seem eminently proper to let the name be Chickasaw, which I think is not preoccupied."

Since communicating with Professor Hilgard I have been informed by Professor Shaler that the name Chickasaw was, he thinks, proposed for this group in some manuscript prepared for the Kentucky State Survey at the time it was under his direction, and while he is not sure that it ever got into print, he remembers distinctly that it was colloquially in use among the members of the Survey for the formation exposed, as stated by Professor Hilgard, at the bluffs of the same name.

Professor Smith entirely accords with Professor Hilgard's substitution; so that those chiefly interested having accepted the change, and a careful search not revealing any conflicting use of the term, nothing seems to stand in the way of its adoption.

The relations of the Lignitic or Chickasaw stage with the Cretaceous and Lower Claibornian in northwestern Louisiana are discussed by Vaughan, Bull. U. S. Geol. Survey No. 142, 1896, pp. 14–27, and Harris, The Lignitic Stage; Part I, in Bull. Am. Paleontology No. 9, June 15, 1897, Vol. II, pp. 195–294.

B 19. *Kenai group.*—Lignitiferous beds of the northwest coast of America situated between the Mesozoic limestones and the marine Miocene, from the latter of which the Kenai group are not separated by any stratigraphic break of importance, the two series being conformable. The Kenai group is most fully displayed on the northwest slope of the peninsula of Kenai, Cook Inlet, Alaska, but extends along the coast and in the interior, in many localities from Norton Sound on the north to British Columbia and perhaps to Oregon. These beds were referred to the Miocene by Heer, Newberry, and other paleobotanists, but by J. Starkie Gardner, Dawson, and Knowlton are believed to be Eocene. They are with little doubt coeval with the Atane beds of Greenland and other arctic leaf-bearing strata. Their exact horizon is doubtful, but some of the plants appear to be common to the lignitic beds of the Mexican Gulf coast, and they are provisionally placed here awaiting more definite information. For a discussion of the subject see Dall, Bull. U. S. Geol. Survey No. 84, 1892, pp. 249–252; Knowlton, Bull. Geol. Soc. America, Vol. V, 1893, pp. 587–590, and Proc. U. S. Nat. Mus., Vol. XVII, 1894, pp. 236–240; also Dall, Seventeenth Ann. Rept. U. S. Geol. Survey, Part I, 1896, pp. 837–842.

C 15. *Wasatch.*—Wasatch Mountains, Utah. See F. V. Hayden, Prel. Rept. Geol. Surv., Colorado and New Mexico, 1869, p. 91; Cope, Proc. Am. Philos. Soc., February, 1872; and Dall, Bull. U. S. Geol. Survey No. 84, 1892, p. 337. See also W. B. Scott, Science, new series, No. 42, 1895, p. 499.

D 32. *Hatchetigbee beds.*—Hatchetigbee Bluff, on the Tombigbee River, Washington County, Alabama. See E. A. Smith, Bull. U. S. Geol. Survey, No. 43, 1887, p. 39.

D 33. *Bashi series.*—Bashi Creek, Clarke County, Alabama, also known as Woods

Bluff series. See Smith and Johnson, Bull. U. S. Geol. Survey No. 43, 1887, p. 43; and E. A. Smith, Report on the Geology of the Coastal Plain of Alabama, 1894, p. 27.

D 34. *Tuscahoma series.*—Tuscahoma Landing, Choctaw County, Alabama. See Langdon, Bull. Geol. Soc. America, Vol. II, 1891, p. 596, and E. A. Smith, Report on the Geology of the Coastal Plain of Alabama, 1894, p. 162. This series is better known as the Bells Landing series, but Professor Smith, who has worked more than anyone else on this series, gives precedence to the name Tuscahoma, whose advantages over the other name are sufficiently obvious.

D 35. *Greggs Landing series.*—Greggs Landing, Alabama River. See Harris, Am. Jour. Sci., 3d series, Vol. XLVII, 1894, p. 304. This series of beds, with those of Bells Landing or Tuscahoma and those of Nanafalia, form a group of closely related faunal horizons, which are by Harris maintained to be the equivalent of the Pamunkey Eocene of Maryland and Virginia.

D 36. *Nanafalia series.*—Nanafalia Bluff, Marengo County, Alabama, on the Tombigbee River. See Smith and Johnson, Bull. U. S. Geol. Survey No. 43, 1887, p. 51.

E 25. *Atane or Atanekerdluk leaf beds.*—Disco, Greenland. See Heer, Flora Fossilis Arctica, Vol. I, 1868, p. 72. These beds are provisionally placed here to be parallel with the Kenai group. It is possible that future researches will result in moving both higher up in the Eocene column.

E 26a–b. *Pamunkey formation.*—Pamunkey River, Virginia. See Darton, Bull. Geol. Soc. America, Vol. II, 1891, p. 439. This formation is intended to comprise all the Eocene of Maryland and Virginia as understood by Rogers and Conrad. Harris, Am. Jour. Sci., 3d series, Vol. XLVII, 1894, pp. 301–304, has shown that the fauna of these beds is paleontologically similar to that of the Bells Landing group of Alabama, a conclusion indicated by both the mollusks and the corals. W. B. Clark is of the opinion that the Pamunkey is the equivalent of a more extended portion of the Eocene column as recognized in Alabama. See Clark, Johns Hopkins Univ. Circ., Vol. XVI, No. 121, October, 1895, p. 3; and Bull. U. S. Geol. Survey No. 141, 1896, p. 39. Professor Clark divides the Maryland and Virginia Eocene (Pamunkey) into two series of beds, the Woodstock, above, which he is inclined to regard as the equivalent of the Claibornian; and the Aquia Creek, below, which is more distinctly identical with the Tuscahoma or Bells Landing series of Alabama. A more thorough knowledge of the fauna in the Pamunkey is essential to a final decision on this question.

F 12. *Suessonian.*—See Harris, Table of 1891, p. 330; and Dana, Manual of Geology. 4th edition, 1894, p. 925.

### *d.* MIDWAYAN STAGE.

A 12. *Midway stage.*—From Midway, a plantation and landing on the Alabama River, Wilcox County, Alabama. See Smith and Johnson, Bull. U. S. Geol. Survey No. 43, 1887, p. 62. For a phase of this series Langdon proposed the name Clayton group, from Clayton, Barbour County, Alabama (Bull. Geol. Soc. America, Vol. II, 1891, p. 594), and E. A. Smith in his latest publication seems disposed to substitute the name Clayton for that of Midway, as a whole (Report on the Geology of the Coastal Plain of Alabama, 1894, p. 192). Some further examination into their equivalency would seem advisable before this change is definitely adopted, though Midway as a proper name is open to obvious objections. The subject has been discussed by Harris in Bull. Am. Pal. No. 4, 1896, pp. 7–13, who adopts the name Midway and fully illustrates the fauna.

B 20. *Tejon group.*—From Fort Tejon, California, near which the formation was first recognized by Conrad in 1855. The fauna at this locality, where the

Tejon beds lie unconformably upon metamorphic rocks, must be regarded as typical for this formation (see note under B. 18). It is asserted that in certain situations (as at New Idria, California) the Tejon is not only conformable to the Chico Cretaceous beds, but represents sedimentation continuous with theirs. If this prove true it is obvious that the lower part of the Tejon must represent the basal Eocene. Harris (Science, August 18, 1892, Vol. XXII, p. 97) has shown that the fauna presents great similarity to that of the Lower Claiborne of Alabama. Dr. J. C. Merriam, of the University of California, after a careful study of the Martinez group of Gabb, finds the typical fauna occurring near the town of Martinez very distinct from that of the typical Tejon which occurs above it. See Jour. Geol., Nov.-Dec., 1897, Vol. V, pp. 767-775. It is therefore probable that the conformity referred to is accidental and local, and that no true continuity exists, in the sense intended, in the Cretaceous-Tertiary sedimentation. The Tejon is provisionally left at the foot of the column, but should probably be raised to the Chickasawan division, or even higher, leaving the Martinez beds to represent the Midway in the Pacific column. The views of Stanton, formulated before Dr. Merriam had completed his researches, and in which the former treats the Martinez group as Lower Tejon, and gives references to the rather copious literature of this vexed question, may be found in the Seventeenth Ann. Rept. U. S. Geol. Survey, Part I, 1896, pp. 1005-1036.

B 21. *Martinez group.*—From the town of Martinez, California, near which the typical outcrops occur. See note under B 20.

B 22. *Puget group.*—Brackish-water lignitiferous beds of Puget Sound, Washington. See White, Am. Jour. Sci., 3d series, Vol. XXXVII, 1888, p. 443. In his discussion Dr. White quotes Mr. Willis as to the enormous thickness of the beds included in the Puget group, and describes them as estuarine. Certain marine beds of Dwamish, Puget Sound, contain Tejon fossils, and are regarded by Mr. Willis as not older than the upper part of the Puget group, which would thus appear to be essentially of Tejon age, and probably represents estuarine local conditions coeval with part of the Chico series, being thus basal Eocene or uppermost Cretaceous. The conditions of estuarine deposition with development of lignite beds probably continued through the Eocene and possibly into Miocene time. If so, the Puget group may eventually be differentiated into several members. See note under B 18. See also Burrard Inlet beds of Sir J. W. Dawson, Trans. Royal Soc. Canada, 2d series, Vol. I, 1895, pp. 137-151.

C 16, 17, 18. *Puerco beds.*—Cope, Ann. Rept. U. S. Geog. Surv. W. One Hundredth Mer. for 1875-1877, and Am. Naturalist, Vol. XVI, 1882, pp. 177-195. The vertebrate fossils of the Puerco beds are discussed by Osborn and Earle, who show that the series contains two principal horizons bearing fossils; the lower, lying near the base of the formation, was found exposed at Pina Verta Canyon in the San Juan basin of northwestern New Mexico and elsewhere. It is characterized especially by the genera Polymastodon and Chriacus. The upper, separated from the lower by an interval of 30 feet of unfossiliferous rock, was exposed, among other places, at the head of Canyon Gallego, and is characterized by the genera Chirox, Indrodon, and Pantolambda. Beside the characteristic genera mentioned, there are numerous others common to the two beds. I have therefore separated the two horizons in the table. See Osborn and Earle, Bull. Am. Mus. Nat. Hist., Vol. VII, 1895. pp. 1-70. More than two years after the above was written, Dr. Matthew reported to the New York Academy of Sciences that the upper and lower Puerco beds contain in common no species and only three or four genera. Retaining for the lower series the name of Puerco, Dr. Wortman proposes for the upper beds the name Torregon formation. See Science, Dec. 3, 1897, new series, Vol. VI, No. 153, p. 852.

C 19. *Fort Union group.*—Fort Union, in western North Dakota; also known as the Great Lignitic group. See Meek and Hayden, Proc. Acad. Nat. Sci. Phila. for 1861, p. 433, and Knowlton, Proc. U. S. Nat. Mus., Vol. XVII, 1894, pp. 207–240.

D 37. *Naheola series.*—Naheola, Marengo County, Alabama; also, and perhaps better known, the Matthews Landing series, from the point of that name in Wilcox County, Alabama. See Smith and Johnson, Bull. U. S. Geol. Survey No. 43, 1887, p. 57, and Smith, Report on the Geology of the Coastal Plain of Alabama, 1894, pp. 181–186, 230, 481.

D 38. *Sucarnochee series.*—From Sucarnochee Creek, Wilcox County, Alabama; also known as the Black Bluff series. See Smith and Johnson, Bull. U. S. Geol. Survey No. 43, 1887, p. 61, and Smith, Report on the Geology of the Coastal Plain of Alabama, 1894, pp. 186, 230; also Dall and Harris, Bull. U. S. Geol. Survey No. 84, 1892, p. 322.

D 39. *Midway limestones.*—From Midway, Wilcox County, Alabama. See note under A 12.

F 13. *Cernaysian.*—After Cernays, a village near Rheims, France. See Dana, Manual of Geology, 4th edition, 1894, pp. 884, 923, 925.

## SUBJACENT MESOZOIC.

B 23. *Chico Shasta group.*—Also widely known as the Chico-Tejon. From the town of Chico, Butte County, California. See note under B 20, and C. A. White, Bull. U. S. Geol. Survey No. 15, 1885, p. 11, and preliminary remarks to this series of notes. Compare the Nanaimo group of Sir W. Dawson, Trans. Royal Soc. Canada, 2d series, 1895, pp. 137–151.

C 20. *Denver group.*—Denver, Colorado. See Marsh, Am. Jour. Sci., 3d series, Vol. XXXIV, 1887, p. 324; Stanton and Knowlton, Stratigraphy and Paleontology of the Laramie and related formations in Wyoming, Bull. Geol. Soc. America, Vol. VIII, 1897, pp. 127–156; and preliminary remarks to this paper, p. 333.

D 40. *Ripley group.*—From Ripley, Tippah County, Mississippi. See Hilgard, Rept. Agric. and Geol. Mississippi, 1860, p. 87.

E 27. *Upper Green Marl.*—According to White and Whitfield, while the upper Green Marl bed of New Jersey has in its lower portion fossils of the Ripley fauna and in its upper part species belonging to the Middle Eocene, there is really a faunal hiatus between the two. The New Jersey marls are deposits formed in troubled water and are stratigraphically unreliable from the tendency to exhibit mixed and heterogeneous fossils. See White, Bull. U. S. Geol. Survey No. 82, 1891, pp. 100, 106; and Whitfield, Mon. U. S. Geol. Survey, Vol. XVIII, 1892, p. 20.

F 14. *Danian.*—See Dana, Manual of Geology, 4th edition, 1894, pp. 858, 864, 866; also Harris, Table of 1891, pp. 330–331. The Upper Danian or Garumnian of Meulan (Pisolitic limestone), though referred to the Eocene by Mayer-Eymar as above, is by later authorities pretty generally assigned to the uppermost Cretaceous, deduction being made of the Calcaire de Mons, or Montian of Cornet and Briart.

# GLACIERS OF MOUNT RAINIER,

BY

ISRAEL COOK RUSSELL;

WITH A PAPER ON

# THE ROCKS OF MOUNT RAINIER,

BY

GEORGE OTIS SMITH.

# CONTENTS.

# ILLUSTRATIONS.

# GLACIERS OF MOUNT RAINIER.

By Israel C. Russell.

## THE PRINCIPAL PHYSICAL FEATURES OF THE STATE OF WASHINGTON.

The State of Washington presents marked geographic and climatic diversities. The Cascade Mountains cross the State in a general north-south direction and divide it into two portions which are as different from each other in nearly all their features as two adjacent regions can well be.

The Cascade Mountains begin at the south in northern California and extend northward across Oregon and Washington and into British Columbia. In northern Washington the range is composed of granitic rocks, and its scenic features are here much more diverse than in its southern extension, where only igneous rocks, mostly andesite and basalt, have as yet been reported to occur. Many of the summit peaks of the Cascade Range in Washington are from 6,000 to 8,000 feet or more in altitude. At least five passes are known which are considered practicable for railroads. These range in elevation from 3,100 to 5,500 feet above the sea. The Columbia River, flowing from east to west, passes through the mountains, forming in part the boundary between Washington and Oregon, and dissects the range nearly to sea level. The elevation at Cascade Locks is but 106 feet above the ocean.

East of the foothills of the Cascades lies the Great Plain of the Columbia, as it is termed, which merges on the south with a vast plateau, forming southeastern Washington and extending far into Oregon. The elevation in this region ranges from about 300 feet along the Columbia to 2,500 feet near the Washington-Idaho boundary. Central and southeastern Washington, embracing a region about 20,000 square miles in area, was once a level plain of basalt formed by many successive flows of molten rock. In the neighborhood of the Cascades the once level basaltic sheets have been broken by numerous lines of fracture and the intervening blocks have been variously tilted, but in the southeastern portion of the State the lava sheets are still horizontal and form a nearly level plateau, in which there are many deeply eroded stream channels, some of which, termed coulées, are no longer lines of drainage.

The vast lava flows which covered central and southeastern Washington, and a still greater region to the southward, met the mountains of metamorphic rock of Idaho and Washington in an irregular line coinciding in general with the Washington-Idaho boundary from the Oregon line northward to Spokane River; thence westward the border of the lava is followed by the Columbia in making its westerly detour, known as the Big Bend. To the north of the Big Bend of the Columbia are rugged mountains of granite rocks which merge with the northern Cascades. When the geology and geography of Washington are more thoroughly studied it will probably be found that the granitic portion of the northern Cascades belongs in reality to the great series of highlands and mountains north of the Big Bend, which is designated in general as the Okanogan Mountains.

The Cascade Mountains are precipitous on the west and descend abruptly to a region of mild relief in which the extremely irregular basin occupied by Puget Sound is sunk. This low country, deeply covered with glacial drift, extends westward to the Olympic Mountains and to the Pacific.

As shown in this brief outline, the main geographic features of Washington are a great mountain range trending approximately N. 15° E., a roughened lava plateau on the east, and a dissected plain of glacial drift on the west.

Associated with the Cascade range in Washington, but of later date, and distinct from it both geographically and geologically, are four especially prominent volcanic mountains. About 15 miles east of the crest line of the Cascades in south-central Washington stands Mount Adams, 9,570 feet high. This is a volcanic cone much defaced by disintegration and erosion, but steam escaping from its summit reveals the fact that the rocks within the snow- and ice-covered cone are still hot. Fifty miles west of the Cascades, and near the Canadian boundary, rises Mount Baker, 10,877 feet high, also a volcanic mountain, with a well-defined crater near its summit. Similar to Mount Baker in geographical position, but in the southern part of the Puget Sound country, stands Mount St. Helens, 9,750 feet high. Mount Baker and Mount St. Helens have been in eruption in modern times, and it is believed, although perhaps not proved, that eruptions from them have occurred within the last fifty years. The fourth great volcano referred to, and the grandest of all, is Mount Rainier, to be described later, on which the glaciers forming the subject of this paper are situated.

## CLIMATE.

Of special interest in the study of the glaciers of Mount Rainier is the character of the climate of Washington. The plateau country to the east of the Cascades is arid and partakes of the desert-like nature of the Great Basin region, of which in many respects it is a northern extension. The rainfall is small. Over an area of 20,000 square miles the mean annual precipitation varies from 5 to 20 inches. The summers are long and hot; the winters mild and rainy. Snow lies on the higher portions of the plateau for several weeks each year, but seldom remains more than a few days in the valleys. The plateau is treeless. In the lower valleys sagebrush and other desert shrubs impart their neutral tints to the landscape. The higher portions of the plateau in a state of nature were covered with luxuriant bunch grasses. Much of the former prairie is now under cultivation. The deep, rich soil formed by the decay of basalt absorbs the rain and retains it in such a manner as to render wheat raising possible with an annual precipitation of less than 20 inches.

West of the Cascades the entire country, except certain limited areas where soil conditions are unfavorable, was originally clothed with magnificent forests. The forests of fir, cedar, spruce, etc., from which the Evergreen State derives its popular name, extend from the shore of the Pacific over the Cascade Mountains and down their eastern slopes to an elevation of about 2,000 feet, leaving the higher summits bare. The annual rainfall to the west of the mountains ranges from 50 to over 100 inches. On the higher mountains it is probable that the mean annual precipitation, mostly in the form of snow, is in excess of the heaviest precipitation observed in the valleys. The temperature throughout the year is surprisingly mild and equable. The winters are humid, but with little snow in the valleys.

The prevailing winds in Washington are from the west, and come from the Pacific charged with moisture. In passing Mount Rainier and other lofty mountains they are forced upward, rarefied, and chilled, and part with much of their humidity before reaching the plateau east of the Cascades.

Glaciers are no less dependent on climatic conditions than forests. We find them especially where high mountains rise in the paths of warm, humid winds. On the mountains of Washington, and particularly on Mount Rainier, these conditions are abundantly fulfilled.

## POSITION AND ELEVATION OF MOUNT RAINIER.

Mount Rainier is situated approximately 43 miles southeast of the city of Tacoma and 11 miles west of the crest of the Cascade Mountains. As determined by the United States Geological Survey its highest summit, Crater Peak, is in latitude 46° 51′ 04.84″ and longitude 121° 45′ 28.50″, and is 14,526 feet in height.

The country intervening between Puget Sound and the foothills of Mount Rainier is low and for the most part heavily forest covered. From Tacoma, which is at tide level, the whole height of the majestic mountain is in view; that is, the visual height of the peak embraces the entire elevation of its summit above the sea. The appearance of the mountain from near Tacoma is shown in the photograph forming Pl. LXV. Although much of the detail in the sculpturing of the mountain's slopes is lost in this picture, yet it serves to show the isolation of the great peak and the manner in which it towers above all its neighbors. As seen from many points on Puget Sound, even as far north as the Straits of Fuca, the mountain is wonderfully impressive. It rises boldly into the sky as a magnificent snow-clad dome, and appears sufficiently rugged and precipitous to bid defiance to the boldest mountaineer. In fact, its northern side is so precipitous that no ascents have been attempted from that direction.

To the east of Mount Rainier, and intervening between its base and the Cascade range, there is a rugged and deeply eroded country, not yet thoroughly explored, in which the highest peaks are, by estimate, between 7,000 and 8,000 feet high. This region, in common with the lower slopes of Mount Rainier and nearly all of the adjacent country, as previously stated, is densely forest covered.

The important place that Mount Rainier occupies in the minds of all who are familiar with the charming scenery of western Washington has created a desire for more detailed information concerning its higher and more inaccessible portions, and particularly in reference to the glaciers visible on its sides from great distances.

## DISCOVERY AND EARLY EXPLORATION OF MOUNT RAINIER.

Spanish explorers entered what is now known as Puget Sound in 1790, and must have been familiar with Mount Rainier as seen from a distance, but, so far as the records seem to show, they did not give a name to the mountain.

Vancouver explored and mapped Puget Sound in 1792, and named not only the waterways and their immediate shores, but several of the mountains seen in the distance as well. Mounts Baker, Rainier, and Hood were named at this time in honor of the lords of the British Admiralty whose names they still bear. In the narrative of his voyage,[1] in describing the region about Port Townsend, Captain Vancouver stated that a "very remarkable high round mountain, covered with snow, apparently at the southern extremity of the distant range of snowy mountains before noticed [the Cascades] bore S. 45° E." When farther south in Puget Sound, he recorded (p.79): "The weather was serene and pleasant, and the country continued to exhibit, between us and the eastern snowy range, the same luxuriant appearance. At its northern extremity Mount Baker bore by compass N. 22° E.; the round snowy mountain now forming its southern extremity, and which, after my friend Rear-Admiral Rainier, I distinguished by the name of Mount Rainier, bore N. 42° E."

Before the coming of Vancouver the Indians had a name for the mountain in which we are interested. To them it was known as Tacoma,

[1] Captain George Vancouver, A Voyage of Discovery to the North Pacific Ocean and Round the World, London, 1801, 8°, vol. 2, pp, 73, 79.

MOUNT RAINIER FROM STEILACOOM PLAINS, NEAR TACOMA.

as nearly as their pronunciation can be rendered in English. Personally, I am strongly in favor of retaining the aboriginal name, but as the United States Board on Geographic Names has decided that the name which Vancouver gave shall be used on Government maps and in official publications, I have no choice in this paper but to accept their decision.

The first ascent of Mount Rainier was made by Gen. Hazard Stevens and P. B. Van Trump in August, 1870. An exceedingly graphic and entertaining account of this pioneer climb was published by Stevens in the Atlantic Monthly, Vol. XXXVIII, 1876. This ascent was made on the south side of the mountain, by the way of what are now known as Paradise Park and Gibraltar, which is practically the route followed in recent years by many tourists. After spending a night in the crater at the summit, they made the descent by the same route.

In October, 1870, Messrs. S. F. Emmons and A. D. Wilson, of the United States Geological Exploration of the Fortieth Parallel, ascended Mount Rainier, taking essentially the same route as that followed about two months before by Stevens and Van Trump. During this excursion much valuable information concerning the geology of the mountain, and especially in relation to the numerous glaciers on its sides, was obtained, and this will be presented later.

During the past ten years the south side of Mount Rainier, between the Nisqually and Cowlitz glaciers, has become a favorite and much frequented resort for camping parties. The park-like region near timber line presents unusual attractions on account of the cool summer temperature, bracing air, and magnificent scenery, and also furnishes a convenient starting point for persons desiring to climb the great peak. An especially beautiful portion of the southern slope, known as Paradise Park, is one of a number of open, meadow-like tracts, strewn in profusion with charming flowers during the short summer and diversified by thrifty groves of firs, to which much of its beauty is due. From Paradise Park the way to the summit of Mount Rainier is easily found, and the climb, considering the elevation attainable, is by no means difficult.

## CHARACTERISTICS OF MOUNT RAINIER.

Mount Rainier is an extinct volcano. The residual heat of its once molten rocks still gives origin to steam jets, slightly impregnated with sulphurous gases in a few instances, which escape from crevices in the now partially snow-filled craters at the summit.

As has been determined by Bailey Willis, the mountain stands on a slanting peneplain, which consists of granites, schists, and coal-bearing Tertiary rocks; that is, the region where Mount Rainier is situated was eroded during late Tertiary times until it was reduced practically to a plain at sea level. Such a plain is known among geographers as a peneplain. This peneplain was then upraised and tilted so as to slope gently westward. Since the plain was elevated it has been deeply dissected by erosion, and the land masses between the sunken stream channels have been worn into mountain forms. The general level of the summits which mark approximately the position of the tilted peneplain, in the region adjacent to Mount Rainier on the north, is about 6,500 feet. The streams that have roughened the plain by excavating deep channels in the rocks composing it were guided westward by the slope produced by the uplifting; that is, their courses were determined by the slope of the land, due to tilting, and for this reason they are classed as consequent streams. Several rivers have their sources in the glaciers of Mount Rainier, and flow away from the mountain in all directions, and have deeply eroded its sides. These, too, are consequent streams; that is, their courses were determined by the original slopes of the mountain.

The latest marked eruption of Mount Rainier ceased before the work of erosion, now evidenced by such conspicuous results on its lower slopes and in the platform on which it stands, was far advanced. The rocks all about the base of the volcano are mainly Tertiary sediments and the products of ancient volcanic eruptions. Of older date than the Tertiary, and apparently rising through the rocks of that age at a few localities, are isolated areas of light-colored granite. Outcrops of granite were noted on the south side of Mount Rainier, near the extremity of the Nisqually Glacier, by Kautz in 1857, and also by Emmons in 1870. Similar granite, forming bold, rounded knobs and mountain-like masses on the north side of the mountain, between Carbon and Winthrop glaciers, was discovered in 1896 during the reconnaissance which formed the basis of this paper. A line connecting the granite outcrops on the two sides of the mountain bears slightly east of north, and possibly indicates the direction of the line of fracture that admitted of the extrusion of the lavas of which Mount Rainier is built. An hypothesis suggested by the observations made by myself last summer is that the granite was raised by faulting after the adjacent Tertiary rocks were deposited but previous to their being worn down to a peneplain. No opportunity was found, however, for testing this suggestion.

The main mass of Mount Rainier is composed of andesite and basalt, which were ejected to a considerable extent in a fragmental condition as scoria, pumice, lapilli, bombs, etc. Lava flows were not abundant during the later stages of eruption. The mountain ranks as a composite cone, but so far as its structure is revealed in the canyons and amphitheaters sculptured in its sides, and as is indicated also by the profiles of the great cone, it was built largely of material thrown out by explosions from a summit crater. The profiles of the mountain and the character of its summit show that at the time of its greatest perfection and beauty it rose as a tapering cone, with gently concave sides, to a height about 2,000 feet greater than its present elevation. At a later date it was truncated, probably by an explosion, which removed the upper 2,000 feet and left a summit crater from 2 to 3 miles in diameter. Remnants of the rim of this immense crater now form Peak Success and Liberty Cap. Subsequently explosive eruptions partially filled the great crater and formed two smaller craters within it. The rims of the smaller craters are still clearly traceable, although at present the depressions they encircle are nearly filled with snow. A moderately prominent point between the two younger craters, known as Crater Peak, is now the actual summit of the mountain.

As Mount Rainier stands to-day it has lost much of its youthful grace and symmetry. Its rocks have yielded to frost and storms and have been deeply sculptured by glaciers. The characteristic features produced by the decay, dissection, and erosion of the rocks forming the mountain, and the future topographic changes that may be expected to take place if the present destructive agencies continue their work, will be considered after the nature of the glaciers, to which the changes in the physiography of the mountain are mainly due, are briefly discussed.

## NARRATIVE.

The reconnaissance during which the notes for this essay were obtained began at Carbonado, a small coal-mining town about 20 miles southeast of Tacoma, with which it is connected by a branch of the Union Pacific Railroad. Carbonado is situated on the border of the unbroken forest. Eastward to beyond the crest of the Cascade Mountains is a primeval forest, the density and magnificence of which it is impossible adequately to describe to one who is not somewhat familiar with the Puget Sound region. From Carbonado a trail, cut through the forest under the direction of Willis in 1881, leads to Carbon River, a stream flowing from Mount Rainier, which it formerly crossed by a bridge that is now destroyed, and thence continues to the west of the mountain to Busywild. A branch of this trail leads eastward to the north side of the mountain, making accessible a beautiful region near the timber line, known as Spray Park. A portion of the main trail and the branch leading to Spray Park is shown on the accompanying map, Pl. LXVI.

Our party consisted of Bailey Willis, geologist in charge, George Otis Smith and myself, assistants, and F. H. Ainsworth, Fred Koch, William B. Williams, and Michael Autier, camp hands.

### IN THE FOREST.

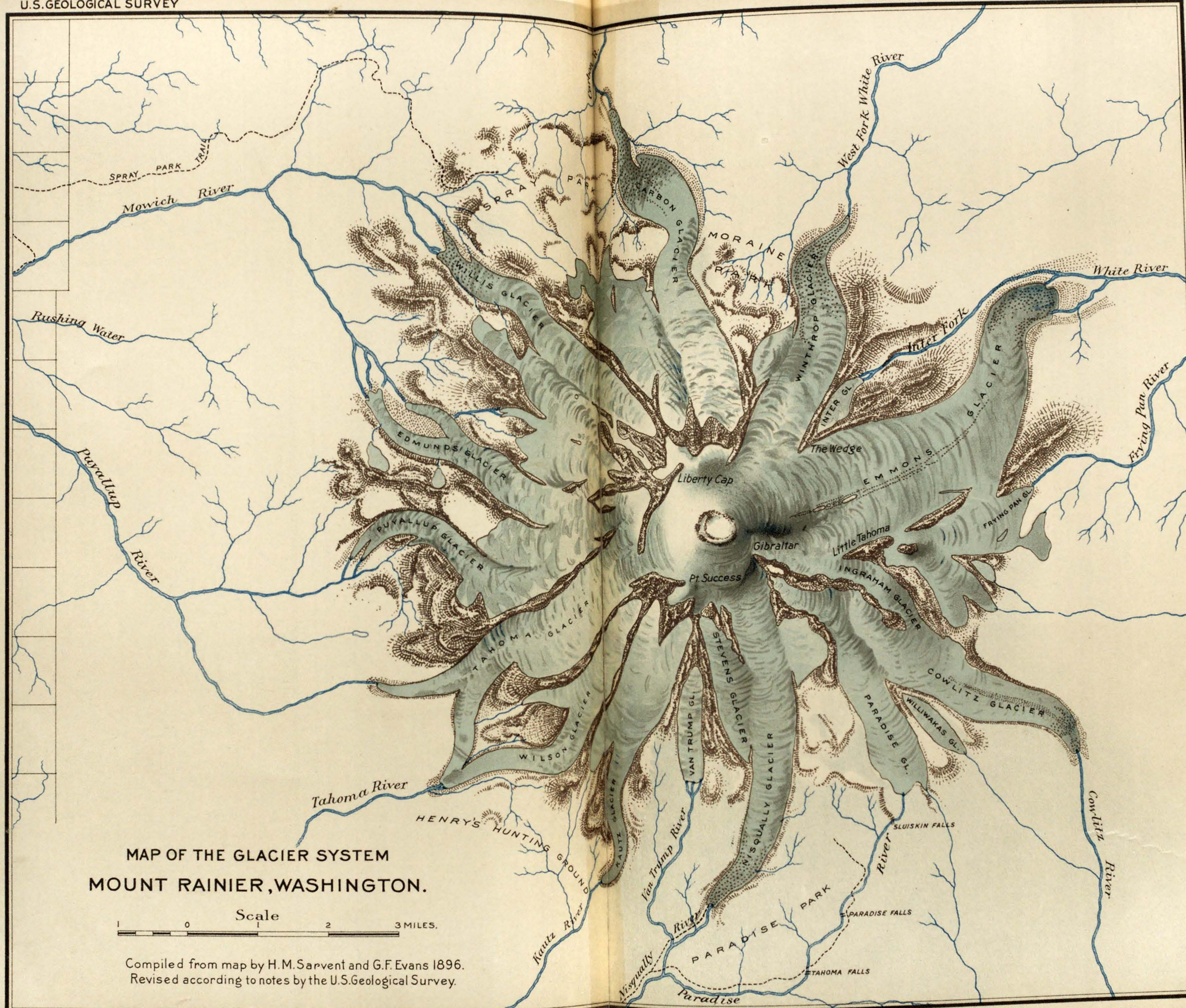
From Carbonado we proceeded with pack animals along the Willis trail, already mentioned, to the crossing of Carbon River. We then left the main trail and went up the right bank of the river by a trail recently cut as far as the mouth of Chenuis Creek. At that locality our party was divided. Willis and myself, taking blankets, rations, etc., and crossing the river, proceeded up its bowlder-strewn left bank to the foot of Carbon Glacier. The remainder of the party cut a trail along the right bank, and in the course of a few days succeeded in making a depot of supplies near where the river emerges from beneath the extremity of the glacier. The pack train was then taken back to near Carbonado for pasture.

The tramp from Carbonado to the foot of the Carbon Glacier was full of interest, as it revealed the characteristics of a great region, covered with a dense forest, which is a part of the deeply dissected Tertiary peneplain surrounding Mount Rainier. The rocks from Carbonado to Carbon River crossing are coal bearing. Extensive mines are worked at Carbonado, and test shafts have been opened at a few localities near the trail which we followed. At Carbonado the river flows through a steep-sided canyon about 300 feet deep. Near where the Willis trail crosses the stream the canyon broadens, is deeply filled with bowlders, and is bordered by forest-covered mountains fully 3,000 feet in elevation. On account of the dense forests, the scenery throughout the region traversed is wild and picturesque. At a few localities glimpses were obtained of the great snow-clad dome of Mount Rainier, rising far over the intervening tree covered foothills.

The forests of the Puget Sound region are the most magnificent on the continent. The moist atmosphere and genial climate have led to a wonderfully luxuriant growth, especially of evergreens. Huge fir trees and cedars stand in close-set ranks and shoot upward straight and massive to heights which frequently exceed 250 feet, and sometimes are even in excess of 300 feet. The trees are frequently 10 to 12 feet or more in diameter at the height of one's head and rise in massive columns without a blemish to the first branches, which are in many instances 150 feet from the ground. The soil beneath the mighty trees is deeply covered with mosses of many harmonious tints, and decked with rank ferns, whose gracefully bending fronds attain a length of 6 to 8 feet. Lithe, slender maples, termed vine-maples from their habit of growth, are plentiful, especially along the small water courses. In many places the broad leaves of the devil's club (*Fatsia horrida*) give an almost tropical luxuriance to the shadowy realm beneath the lofty canopies formed by the firs and cedars.

In writing of life in this forest while prospecting for coal in 1881, Willis[1] states that one of the fir trees which was measured rose like a huge obelisk to a height of 180 feet without a limb and tapered to a point 40 feet above. "The more slender trees, curiously enough, are the taller; straight, clear shafts rise 100 to 150 feet, topped with foliage whose highest needles would look down on Trinity spire. Cedars, hemlocks, spruce, and white firs mingle with these giants, and not competing with them in height, they fill the spaces in the vast colon-

[1] Bailey Willis, Cañons and glaciers, a journey to the ice fields of Mount Tacoma: The Northwest, April, 1883, Vol. I, p. 2.


MAP OF THE GLACIER SYSTEM
MOUNT RAINIER, WASHINGTON.
Scale
1 0 1 2 3 MILES.
Compiled from map by H. M. Sarvent and G. F. Evans 1896.
Revised according to notes by the U.S. Geological Survey.
SPRAY PARK TRAIL
Mowich River
SPRAY PARK
CARBON GLACIER
MORAINE PARK
West Fork White River
White River
Inter Fork
WINTHROP GLACIER
INTER GL.
The Wedge
EMMONS GLACIER
Frying Pan River
FRYING PAN GL.
WILLIS GLACIER
Rushing Water
EDMUNDS GLACIER
Liberty Cap
Puyallup River
PUYALLUP GLACIER
Gibraltar
Pt. Success
Little Tahoma
INGRAHAM GLACIER
TAHOMA GLACIER
STEVENS GLACIER
COWLITZ GLACIER
WILLIWAKAS GL.
PARADISE GL.
VAN TRUMP GL.
WILSON GLACIER
KAUTZ GLACIER
NISQUALLY GLACIER
Tahoma River
HENRY'S HUNTING GROUND
Van Trump River
SLUISKIN FALLS
River
Cowlitz River
PARADISE PARK
PARADISE FALLS
TAHOMA FALLS
Kautz River
Nisqually River
Paradise

nade. * * * The silence of these forests is awesome, the solitude oppressive. The deer, the bear, the panther, are seldom met; they see and hear first and silently slip away, leaving only their tracks to prove their numbers. There are very few birds. * * * The wind plays in the tree tops far overhead, but seldom stirs the branches of the smaller growth. The great tree trunks stand immovable. The more awful is it when a gale roars through the timber, when the huge columns sway in unison and groan with voices strangely human."

The mighty forest through which we traveled from Carbonado to the crossing of Carbon River extends over the country all about Mount Rainier and clothes the sides of the mountain to a height of about 6,000 feet. From distant points of view it appears as an unbroken emerald setting for the gleaming, jewel-like summit of the snow-covered peak.

In spite of the many attractions of the forest, it was with a sense of relief that we entered the canyon of Carbon River and had space to see about us. The river presents features of geographical interest, especially in the fact that it is filling in its valley. The load of stone contributed by the glaciers, from which the stream comes as a roaring turbid flood, is greater than it can sweep along, and much of its freight is dropped by the way. The bottom of the canyon is a desolate, flood-swept area of rounded bowlders, from 100 to 200 yards broad. The stream channel is continually shifting, and is frequently divided by islands of bowlders, heaped high during some period of flood. Many of the stream channels leading away from Mount Rainier are known to have the characteristics of the one we ascended, and show that the canyons were carved under different conditions from those now prevailing. The principal amount of canyon cutting must have been done before the streams were overloaded with débris contributed by glaciers—that is, the deep dissection of the lower slope of Mount Rainier and of the platform on which it stands must have preceded the Glacial epoch.

### ON THE GLACIERS.

After a night's rest in the shelter of the forest, lulled to sleep by the roar of Carbon River in its tumultuous course after its escape from the ice caverns, we climbed the heavily moraine-covered extremity of Carbon Glacier. At night, weary with carrying heavy packs over the chaos of stones that cover the glaciers, we slept on a couch of moss beautified with lovely blossoms, almost within the spray of Philo Falls, a cataract of clear icy water that pours into the canyon of Carbon Glacier from snow fields high up on the western wall of the canyon.

I will ask the reader to defer the study of the glaciers until we have made a reconnaissance of the mountain and climbed to its summit, as he will then be better prepared to understand the relation of the glaciers, névés, and other features with which it will be necessary to deal. In this portion of our fireside explorations let us enjoy a summer outing, deferring until later the more serious task of questioning the glaciers.

From Philo Falls we ascended still higher, by following partially snow-filled lanes between the long lateral moraines that have been left by the shrinking of Carbon Glacier, and found three parallel, sharp-crested ridges about a mile long and from 100 to 150 feet high, made of bowlders and stones of all shapes, which record the former positions of the glacier. Along the western border of the oldest and most westerly of these ridges there is a valley, perhaps 100 yards wide, intervening between the abandoned lateral moraine and the western side of the valley, which rises in precipices to forest-covered heights at least 1,000 feet above. Between the morainal ridges there are similar narrow valleys, each of which at the time of our visit, July 15, was deeply snow-covered. The ridges are clothed with spruce and cedar trees, together with a variety of shrubs and flowering annuals. The knolls rising through the snow are gorgeous with flowers. A wealth of purple Bryanthus, resembling purple heather, and of its constant companion, if not near relative, the Cassiope, with white, waxy bells, closely simulating the white heather, make glorious the mossy banks from which the lingering snow has but just departed. Acres of meadow land, still soft with snow water and musical with rills and brooks flowing in uncertain courses over the deep, rich turf, are beautiful with lilies, which seemed woven in a cloth of gold about the borders of the lingering snow banks. We are near the upper limit of timber growth, where parklike openings, with thickets of evergreens, give a special charm to the mountain side. The morainal ridge nearest the glacier is forest-covered on its outer slope, while the descent to the glacier is a rough, desolate bank of stones and dirt. The glacier has evidently but recently shrunk away from this ridge, which was formed along its border by stones brought from a bold cliff that rises sheer from the ice a mile upstream. Standing on the morainal ridge overlooking the glacier, one has to the eastward an unobstructed view of the desolate and mostly stone and dirt covered ice. Across the glacier another embankment can be seen, similar to the one on the west, and, like it, recording a recent lowering of the surface of the glacier of about 150 feet. Beyond the glacier are extremely bold and rugged mountains, scantily clothed with forests nearly to their summits. The position of the timber line shows that the bare peaks above are between 8,000 and 9,000 feet high. Looking southward, up the glacier, we have a glimpse into the wild amphitheater in which it has its source. The walls of the great hollow in the mountain side rise in seemingly vertical precipices about 4,000 feet high. Far above is a shining, snow-covered peak, which Willis named the Liberty Cap. It is one of the culminating points of Mount Rainier, but not the actual summit. Its elevation is about 14,300 feet above the sea. Toward the west the view is limited by the forest-covered morainal ridges near at hand and by the precipitous slopes beyond, which lead to a northward-projecting spur of Mount Rainier, known as the Mother Mountains. This, our first view of Mount Rainier near at hand, has shown that the valley down which Carbon Glacier flows, as well as the vast amphitheater in which it has its source, is sunk in the flanks of the mountain. To restore the northern slope of the ancient volcano as it existed when the mountain was young we should have to fill the depression in which the glacier lies at least to the height of its bordering ridges. On looking down the glacier we see it descending into a vast gulf bordered by steep mountains, which rise at least 3,000 feet above its bottom. This is the canyon through which the water formed by the melting of the glacier escapes. To restore the mountain this great gulf would also have to be filled. Clearly the traveler in this region is surrounded by the records of mighty changes. Not only does he inquire how the volcanic mountain was formed, but how it is being destroyed. The study of the glaciers will do much toward making clear the manner in which the once smooth slopes have been trenched by radiating valleys, leaving mountain-like ridges between.

Another line of inquiry which we shall find of interest as we advance is suggested by the recent shrinkage of Carbon Glacier. Are all of the glaciers that flow from the mountain wasting away? If we find this to be the case, what climatic changes does it indicate?

## TO CRATER LAKE.

From our camp among the morainal ridges by the side of Carbon Glacier we made several side trips, each of which was crowded with observations of interest. One of these excursions, made by Mr. Smith and myself, was up the snow fields near camp; past the prominent outstanding pinnacles known as the Guardian Rocks, one red and the other black; and through Spray Park, with its thousands of groves of spire-like evergreens, with flower-enameled glades between. On the bare, rocky shoulder of the mountain, where the trees now grow, we found the unmistakable grooves and striations left by former glaciers. The lines engraved in the rock lead away from the mountain, showing that even the boldest ridges were formerly ice covered. Our route took us around the head of the deep canyon through which flows Cataract Creek. In making this circuit we followed a rugged saw-tooth crest, and had some interesting rock climbing. Finally, the sharp divide between Cataract Creek and a small stream flowing westward to Crater Lake was reached, and a slide on a steep snow slope took us quickly down to where the flowers made a border of purple and gold about the margins of the snow. Soon we were in the forest, and gaining a rocky ledge among the trees, could look down on Crater Lake, deeply sunk in shaggy mountains which still preserve all of their primitive freshness and beauty. Snow lay in deep drifts beneath the shelter of the forest, and the lake was ice covered except for a few feet near the margin. This was on July 20. I have been informed that the lake is usually free of ice before this date, but the winter preceding our visit was of more than usual severity, the snowfall being heavy, and the coming of summer was therefore much delayed.

The name Crater Lake implies that its waters occupy a volcanic crater. Willis states that Nature has here placed an emerald seal on one of Pluto's sally ports; but that the great depression now water-filled is a volcanic crater is not so apparent as we might expect. The basin is in volcanic rock, but none of the characteristics of a crater due to volcanic explosions can be recognized. The rocks, so far as I saw them, are massive lavas, and not fragmental scoriæ or other products of explosive eruptions. On the bold, rounded rock ledges down which we climbed in order to reach the shore, there were deep glacial scorings, showing that the basin was once deeply filled with moving ice. My observations were not sufficiently extended to enable me to form an opinion as to the origin of the remarkable depression, but whatever may have been its earlier history, it has certainly been profoundly modified by ice erosion.

Following the lake shore southward, groping our way beneath the thick, drooping branches which dip in the lake, we reached the notch in the rim of the basin through which the waters escape and start on their journey to Mowich River and thence to the sea. We there found the branch of the Willis trail leading to Spray Park, and turned toward camp. Again we enjoyed the luxury of following a winding pathway through silent colonnades formed by the moss grown trunks of noble trees. On either side of the trail worn in the brown soil the ferns and flowering shrubs were bent over in graceful curves, and at times filled the little-used lane, first traversed fifteen years before.

The trail led us to Eagle Cliff, a bold, rocky promontory rising as does El Capitan from the Yosemite, 1,800 feet from the forest-lined canyon of Mowich River. From Eagle Cliff one beholds the most magnificent view that is to be had in all the wonderful region about Mount Rainier. The scene beheld on looking eastward toward the mighty mountain is remarkable alike for its magnificence and for the artistic grouping of the various features of the sublime picture. In the vast depths at one's feet the tree tops, through which the mists from neighboring cataracts are drifting, impart a somber tone and make the valley's bottom seem far more remote than it is. The sides of the canyon are formed by prominent serrate ridges, leading upward to the shining snow fields of the mighty dome that heads the valley. Nine thousand feet above our station rose the pure white Liberty Cap, the crowning glory of the mountain as seen from the northward. The snow descending the northwest side of the great central dome is gathered between the ridges forming the sides of the valley and forms a white névé from which flows Willis Glacier. Some idea of the grandeur of this scene may be gained from the illustration, Pl. LXVIII. In

SCENE IN THE FOREST SOUTH OF MOUNT RAINIER.

looking up the valley from Eagle Cliff the entire extent of the snow fields and of the river like stream of ice flowing from them is in full view. The ice ends in a dirt-covered and rock-strewn terminus, just above a huge rounded dome that rises in its path. In 1881 the ice reached nearly to the top of the dome and broke off in an ice cliff, the detached blocks falling into the gulf below. The appearance of this cliff in 1885 is shown by Pl. LXXXI. The glacier has now withdrawn its terminus well above the precipice where it formerly fell as an ice cascade, and its surface has shrunk away from well-defined moraines in much the same manner as has already been noted in the case of Carbon Glacier. A more detailed account of the retreat of the extremity of Willis Glacier will be given later.

From Eagle Cliff we continued our tramp eastward along the trail leading to Spray Park, climbed the zigzag pathway up the face of a cliff in front of Spray Falls, and gained the picturesque and beautiful park-like region above. An hour's tramp brought us again near the Guardian Rocks. A swift descent down the even snow fields enabled us to reach camp just as the shadows of evening were gathering in the deeper canyons, leaving the silent snow fields above all aglow with reflected sunset tints.

## ACROSS CARBON GLACIER.

Taking heavy packs on our backs on the morning of July 21, we descended the steep broken surface of the most recent moraine bordering Carbon Glacier in its middle course, some idea of which is conveyed by Pl. LXXIV, and reached the solid blue ice below. Our course led us directly across the glacier, along the lower border of the rapidly melting covering of winter snow. The glacier is there about a mile across. Its central part is higher than its border, and for the most part the ice is concealed by dirt and stones. Just below the névé, however, we found a space about half a mile long in which melting had not led to the concentration of sufficient débris to make traveling difficult. Farther down the glacier, where surface melting was more advanced, the entire glacier, with the exception of a few lanes of clear ice between the ill-defined medial moraines, was completely concealed beneath a desolate sheet of angular stones. On reaching the east side of the glacier we were confronted with a wall of clay and stones, the inner slope of a moraine similar in all respects to the one we had descended to reach the west border of the glacier. A little search revealed a locality where a tongue of ice in a slight embayment projected some distance up the wall of morainal material, and a steep climb of 50 or 60 feet brought us to the summit. The glacier has recently shrunk—that is, its surface has been lowered from 80 to 100 feet by melting.

On the east side of the glacier we found several steep, sharp-crested ridges, clothed with forest trees, with narrow, grassy, and flower-strewn dells between, in which banks of snow still lingered. The ridges are composed of bowlders and angular stones of a great variety of sizes and shapes, and are plainly lateral moraines abandoned by the shrinking of the glacier. Choosing a way up one of the narrow lanes, bordered on each side by steep slopes densely covered with trees and shrubs, we found secure footing in the hard granular snow, and soon reached a more open, parklike area, covered with mossy bosses of turf, on which grew a great profusion of brilliant flowers. Before us rose the great cliffs which partially inclose the amphitheater in which Carbon Glacier has its source. These precipices, as already stated, have a height of about 4,000 feet, and are so steep that the snow does not cling to them, but descends in avalanches. Above the cliffs, where the inclination is less precipitous, the snow lies in thick layers, the edges of which are exposed in a vertical precipice rising above the avalanche-swept rock-slope below. Far above, and always the central object in the wild scenery surrounding us, rose the brilliant white Liberty Cap, one of the pinnacles on the rim of the great summit crater. Our way then turned eastward, following the side of the mountain, and led us through a region just above the timber line, which commands far reaching views to the wild and rugged mountains to the northeast. This open tract, leading down to groves of spruce trees and diversified by charming lakelets, bears abundant evidence of having formerly been ice-covered, and is known as Moraine Park.

In order to retain our elevation we crossed diagonally the steep snow slopes in the upper portion of the Moraine Park. Midway over the snow we rested at a sharp crest of rock, and found that it is composed of light-colored granite. Later we found that much of the area between the Carbon and Winthrop glaciers is composed of this same kind of rock. Granite forms a portion of the border of the valley through which flow the glaciers just named, and furnished them with much granitic débris, which is carried away as moraines and later worked over into well-rounded bowlders by the streams flowing from the ice. The presence of granite pebbles in the courses of Carbon and White rivers, far below the glaciers, is thus accounted for.

A weary tramp of about 4 miles from the camp we had left brought us to the border of Winthrop Glacier. In the highest grove of trees, which are bent down and frequently lie prone on the ground, although still living, we selected a well-sheltered camping place. Balsam boughs furnished luxuriant beds, and the trees killed by winter storms enabled us to have a roaring camp fire. Fresh trail of mountain goats and their but recently abandoned bed showed that this is a favorite resort for those hardy animals. Marmots were also abundant, and frequently awakened the echoes with their shrill, whistling cries. The elevation of our camp was about 8,000 feet.

From our camp on the cliffs above the west border of Winthrop Glacier we made excursions across that glacier and to its heavily moraine-covered extremity. The snow mantle that is spread over the region about Mount Rainier each winter melts first on the rugged plateau surrounding the base of the mountain, and, as the summer's heat increases, gradually withdraws up the mountain sides, but never so as to uncover the more elevated region. The snow line—that is, the position to which the lower border of the mantle of perennial snow withdraws late in summer—has an elevation of about 9,000 feet. The lower margin of the wintry covering is always irregular, however, extending farthest down on the glaciers and retreating highest on the rocks. At the time of our visit the snow had melted off of nearly all the region below our camp, leaving only dirt-stained snow banks in the more completely sheltered recesses and in deeply shaded dells in the adjacent forests. On the glaciers all the region at a greater elevation than our camp was white and free from dirt and stones, while the hard glacial ice was abundantly exposed at lower altitudes and ended in a completely moraine-covered terminus. Above us all was barren, white, and wintry; below lay the flowery vales and grass parks, warm and inviting, leading to the welcome shade of noble forests. Our course led upward into the frozen region.

### TO THE WEDGE.

On leaving the camp on the border of Winthrop Glacier we began our alpine work. There were five in the party selected for the difficult task of scaling Mount Rainier, namely: Willis, Smith, Ainsworth, Williams, and myself. Taking our blankets, a small supply of rations, an alcohol lamp, alpenstocks, a rope 100 feet long to serve as a life line, and a few other articles necessary for traveling above timber line, we began the ascent of Winthrop Glacier early on the morning of July 23. Our route was comparatively easy at the start, but became steeper and steeper as we advanced. The snow was firm and, except for the numerous crevasses, presented no great difficulties to be overcome. In several places the névé rises in domes as if forced up from beneath, but caused in reality by bosses of rock over which the glacier flows. These domes are broken by radiating crevasses which intersect in their central portions, leaving pillars and castle-like masses of snow with vertical sides. At one locality, in attempting to pass between two of these shattered domes, we found our way blocked by an impassable crevasse. Considerable time was lost in searching for a practicable upward route, but at length, by making a detour to the right, we found a way which, although steep, allowed us to pass the much crevassed area and gain the sharp ridge of rock which divides the névé snow flowing from the central dome of the mountain, and marks the separation between Winthrop and Emmons glaciers. This prow-like promontory, rising some 500 feet above the glaciers on either hand, we named The Wedge. This is the upward pointing, acute angle of a great V-shaped portion of the lower slope of the mountain, left in bold relief by the erosion of the valleys on either side. As will be described later, there are several of these remnants about the sides of the mountain at the same general horizon, which record a somewhat definite stage in the destruction of the mountain by ice erosion.

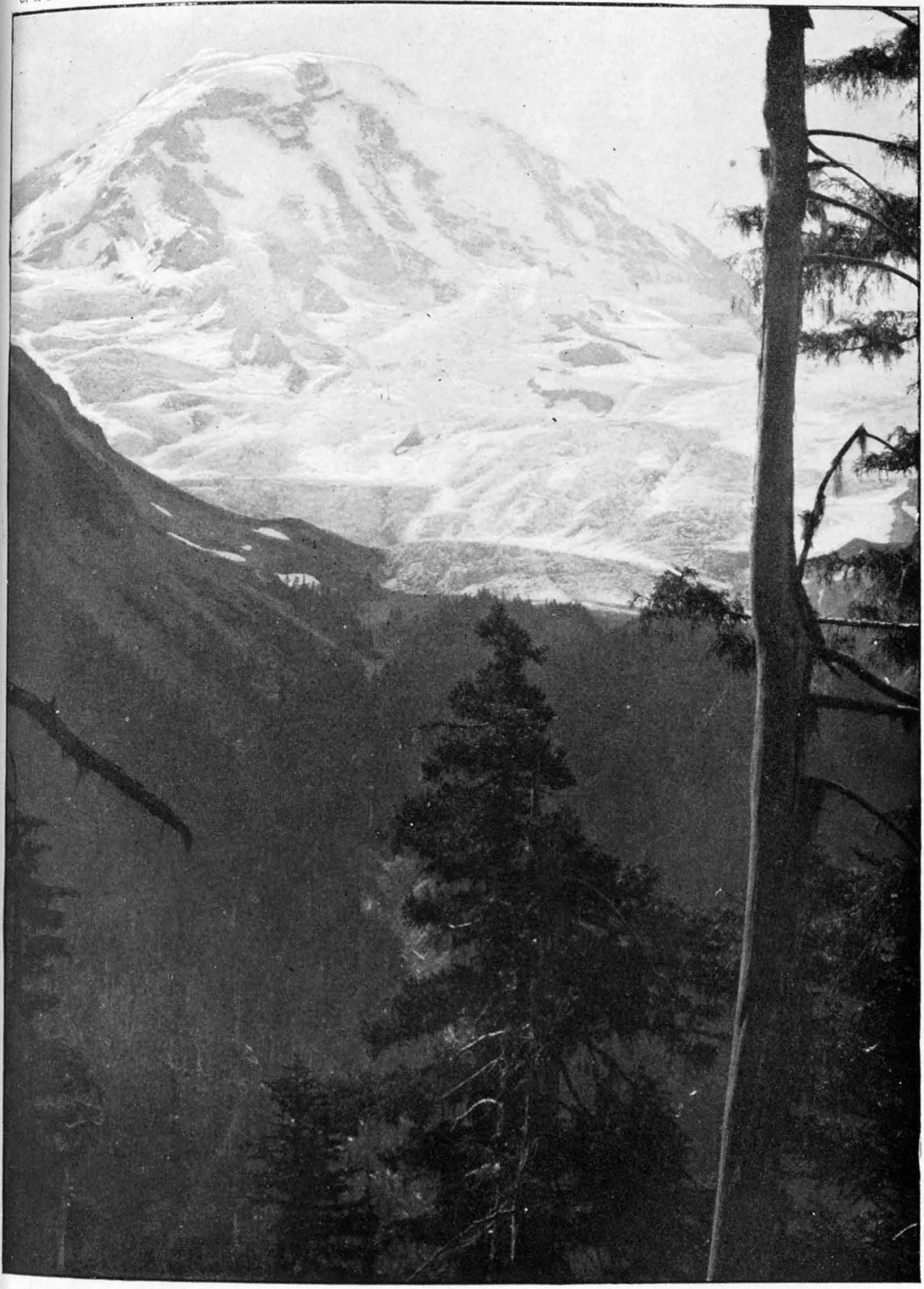
On reaching The Wedge we found it an utterly desolate rocky cape in a sea of snow. We were at an altitude of about 10,000 feet, and far above timber. Water was obtained by spreading snow on smooth rocks or on rubber sheets, and allowing it to melt by the heat of the afternoon sun. Coffee was prepared over the alcohol lamp, sheltered from the wind by a bed sheet supported by alpenstocks. After a frugal lunch, we made shelf like ledges in a steep slope of earth and stones and laid down our blankets for the night. From sheltered nooks amid the rocks, exposed to the full warmth of the declining sun, we had the icy slopes of the main central dome of the mountain in full view and chose what seemed the most favorable route for the morrow's climb.

Surrounded as we were by the desolation and solitude of barren rocks, on which not even a lichen had taken root, and pure white snow fields, we were much surprised to receive passing visits from several humming birds, which shot past us like winged jewels. They came up the valley occupied by the Emmons Glacier, turned sharply at The Wedge, and went down the way of the Winthrop Glacier. What tempts these children of the sunlight and the flowers into the frozen regions seems a mystery. That the humming birds are bold explorers was not new to me, for the reason that on several occasions in previous years, while on the snow-covered slopes of Mount St. Elias, far above all vestiges of vegetation, my heart had been gladdened by glimpses of their brilliant plumage.

When the sun declined beyond the great snow-covered dome that towered above us, and the blue shadows crept down the previously dazzling cliffs, the air became cold and a strong wind made our perch on the rocks uncomfortable. Wrapping ourselves in our blankets we slept until the eastern sky began to glow with sunrise tints.

## TO THE SUMMIT.

Early on the morning of July 24 we began the climb of the steep snow slopes leading to the summit of the mountain. Roped together as we had been on the previous day, we slowly worked our way upward, in a tortuous course, in order to avoid the many yawning crevasses. The way was steep and difficult. Some members of the party felt the effects of the rarefied air, and as we lacked experience in true alpine work our progress was slow and laborious. Many of the crevasses that our course crossed were of the nature of faults. Their upper rims stood several feet above their lower margins, and thus added to the difficulty of passing them. Our aim at first was to traverse the névé of Emmons Glacier and gain the less rugged slope bordering it on the south, but the intervening region was greatly broken and, as we found



MOUNT RAINIER; THE LIBERTY CAP AND WILLIS GLACIER, FROM THE TRAIL NEAR EAGLE CLIFF; LOOKING SOUTHEAST.

after several approaches to it, utterly impassable. The climb presented no special difficulties other than the extreme fatigue incident to climbing steep snow slopes, especially while attached to a life line, and the delays necessitated by frequently turning and retracing our steps in order to get around wide crevasses.

Once while crossing a steep snow slope diagonally, and having a wide crevasse below us, Ainsworth, who was next to the rear of the line, lost his footing and slid down the slope on his back. Unfortunately, at that instant, Williams, who was at the rear of the line, removed his alpenstock from the snow, was overturned by the pull on the line, and shot head first down the slope and disappeared over the brink of the crevasse. A strong pull came on the members of the party who were in advance, but our alpenstocks held fast, and before assistance could be extended to the man dangling in midair, he climbed the taut rope and stood unhurt among us once more. The only serious result of the accident was the loss of an alpenstock.

Pressing on toward the dark rim of rock that we could now and then catch glimpses of at the head of the snow slopes and which we knew to be the outer portion of the summit crater, we crossed many frail snow bridges and climbed precipitous slopes, in some of which steps had to be cut. As we neared the summit we met a strong westerly gale that chilled us and benumbed our fingers. At length, weary and faint on account of the rarity of the air, we gained the lower portion of the rim of stones marking the position of the crater. While my companions rested for a few moments in the shelter of the rocks, I pressed on up the rugged slope and gained the top of the rim.

The stones exposed at the summit are bare of snow, possibly on account of the heat from below, and are rounded and their exposed surfaces polished. The smooth, black bowlders shine in the sunlight much the same as the sand-burnished stones in desert regions. Here on the mountain's brow, exposed to an almost continuous gale, the rocks have been polished by drifting snow crystals. The prevailing rounded form that the stones present may be the result of weathering, or possibly is due to the manner in which the fragments were ejected from the volcano. My hasty examinations suggested the former explanation.

Descending into the crater, I discovered crevices from which steam was escaping, and on placing my hands on the rocks was rejoiced to find them hot. My companions soon joined me, and we began the exploration of the crater, our aim being to find the least uncomfortable place in which to take refuge from the freezing blast rather than to make scientific discoveries.

The crater that we had entered is one of the smaller and more recent ones in the truncated summit of the peak, and is deeply filled with snow, but the rim is bare and well defined. The steam and heat from the rocks have melted out many caverns beneath the snow. In one of these we found shelter.

### A NIGHT IN THE CRATER.

The cavern we chose in which to pass the night, although irregular, was about 60 feet long by 40 wide, and had an arched ceiling some 20 feet high. The snow had been melted out from beneath, leaving a roof so thin that a diffused blue light penetrated the chamber. The floor sloped steeply, and on the side toward the center of the crater there was a narrow space between the rocks and the descending roof which led to unexplored depths. As a slide into this forbidding gulf would have been exceedingly uncomfortable, if not serious, our life line was stretched from crag to crag so as to furnish a support and allow us to walk back and forth during the night without danger of slipping. Three arched openings or doorways communicated with other chambers, and through these drafts of cold air were continually blowing. The icy air chilled the vapor rising from the warm rocks and filled the chamber with steam which took on grotesque forms in the uncertain, fading light. In the central part of the icy chamber was a pinnacle of rock, from the crevices of which steam was issuing with a low hissing sound. Some of the steam jets were too hot to be comfortable to the ungloved hand. In this uninviting chamber we passed the night. The muffled roar of the gale as it swept over the mountain could be heard in our retreat and made us thankful for the shelter the cavern afforded.

The floor of our cell was too uneven and too steeply inclined to admit of lying down. Throughout the night we leaned against the hot rocks or tramped wearily up and down holding the life line. Cold blasts from the branching ice chambers swept over us. Our clothes were saturated with condensed steam. While one side of the body resting against the rocks would be hot, the strong drafts of air with a freezing temperature chilled the other side. After long hours of intense darkness the dome of snow above us became faintly illuminated, telling that the sun was again shining. After a light breakfast and a cup of tea, prepared over our alcohol lamp, we resumed our exploration, none the worse for the exposures of the night.

### ON CRATER PEAK.

Following the inner rim of the crater so as to be sheltered from the gale still blowing steadily from the west, we gained its northern border and climbed to the topmost pinnacle, known as Columbia's Crest. This pinnacle rises about 50 feet above the general level of the irregular rim of the crater, and is the highest point on the mountain. Its elevation, as previously stated, is 14,526 feet.

The magnificent view described by former visitors to this commanding station, which we had hoped would reward our efforts, was concealed beneath a canopy of smoke that covered all of the region about the mountain to a depth of about 10,000 feet. The surface of the layer of smoke was sharply defined, and appeared like an undulating sea surrounding the island on which we stood. Far to the northward rose the regular conical summit of Mount Baker, like an isolated seagirt island. A few of the rugged and more elevated summits, marking the course of the Cascade Mountains, could be discerned to the eastward. The summits of Mount Adams and Mount St. Helens were in plain view and seemingly near at hand. All of the forest-covered region between these elevated summits was blotted out by the dense, heavy layer of smoke, which rose until it met the westerly gale of the upper regions.

During the ascent of Mount Rainier by Emmons and Wilson, previously referred to, more favorable atmospheric conditions prevailed than at the time of my visit, and the region about the base of the mountain was clearly revealed. In describing the view from the summit Emmons says:

From the northeastern rim of the crater we could look down an unbroken slope of nearly 10,000 feet to the head of the White River, the upper half or two-thirds of which was so steep that one had the feeling of looking over a perpendicular wall. [It was up this slope that the climb briefly described above was made.] The systems of glaciers and the streams which flowed from them lay spread out as on a map at our feet; radiating out in every direction from the central mass, they all with one accord curve to the westward to send their waters down toward Puget Sound or the Lower Columbia. [Attention has already been directed to the westward curvature of the streams from Mount Rainier on reaching the tilted peneplain on which the mountain stands, and the explanation has been suggested that they are consequent streams the direction of which was determined by the original slope of the now deeply dissected plateau.]

Looking to the more distant country, the whole stretch of Puget Sound, seeming like a pretty little lake embowered in green, could be seen in the northwest, beyond which the Olympic Mountains extend out into the Pacific Ocean. The Cascade Mountains, lying dwarfed at our feet, could be traced northward into British Columbia, and southward into Oregon, while above them, at comparatively regular intervals, rose the ghost-like forms of our companion volcanoes. To the eastward the eye ranged for hundreds of miles over chain on chain of mountain ridges, which gradually disappeared in the dim, blue distance.

### THE MOUNTAIN'S SUMMIT.

In the truncated summit of Mount Rainier there are three craters. The largest one, partially filled by the building of the two others, is the oldest, and has suffered so greatly from subsequent volcanic explosions and erosion that no more than its general outline can be traced. Peak Success and Liberty Cap are prominent points on the rim of what remains of this huge crater. Its diameter, as nearly as can be judged, is about 2½ miles. Within the great crater, in the formation of which the mountain was truncated and, as previously stated, lost fully 2,000 feet of its summit, there are two much smaller and much more recent craters. The larger of these, the one in which we took refuge, is about 300 yards in diameter, and the second, which is an incomplete circle, its rim having been broken by the formation of its more recent companion, is perhaps 200 yards across. The rim of each now partially snow-filled bowl is well defined, and rises steeply from within to a sharp crest. The character of the inner slopes shows that much rocky material has been detached and has fallen into the cavities from which it was ejected. The rock in the crater walls is in fragments and masses, some of them well rounded and probably of the nature of volcanic bombs. In each of the smaller craters there are numerous steam jets. These show that the rock below is still hot, and that water percolating downward is changed to steam. These steam jets evidently indicate the presence of residual heat and not an actual connection with a volcanic center deep below the surface. All the evidence available tends to show that Rainier is an extinct volcano. It belongs, however, to the explosive type of volcanoes, of which Vesuvius is the best-known example, and there is no assurance that its energies may not be reawakened.

## THE DESCENT.

In descending we chose the south side of the mountain, knowing from the reports of many excursionists who had ascended the peak from that direction that a practicable route could probably be found. Threading our way between numerous crevasses we soon came in sight of a bold, outstanding rock mass, which we judged to be Gibraltar, and succeeded in reaching it with but little difficulty. On gaining the junction of the rock with the snow fields rising above it, we found evidences of a trail, which was soon lost, however, and only served to show that our general course was the right one. A deep, narrow space between the border of Nisqually Glacier and the precipitous side of Gibraltar, from which the snow and ice had been melted by the heat reflected from the cliffs on our left, led us down to a shelf on the lower side of the promontory, which proved a safe and easy way to the crest of a rocky rib on the mountain side which extended far down toward the dark forests in view below.

Gibraltar is a portion of the cone of Rainier built before the explosion which truncated the mountain. It is an outstanding and very prominent rock mass, as may be seen in Pls. LXXIV and LXXVII, left in bold relief by the ice excavation which has carved deep valleys on each side. The rock divides the descending névé in the same manner as does The Wedge, and causes a part of the snow drainage to flow to the Cowlitz and the other part to be tributary to the Nisqually Glacier. The rocks forming Gibraltar consist largely of fragments ejected from the crater above, but present a rude stratification due to the presence of lava flows. When seen from the side and at a convenient distance, it is evident that the planes of bedding, if continued upward at the same angle, would reach above the present summit of the mountain. Gibraltar, like The Wedge and several other secondary peaks on the sides of Mount Rainier, are, as previously explained, the sharp, upward-pointing angles of large V-shaped masses of the original volcanic cone, left in bold relief by the excavation of deep valleys radiating from the

NÉVÉ ON THE EAST SIDE OF MOUNT RAINIER.

central peak. On the backs, so to speak, of these great V-shaped portions of the mountain which now seem to rest against the central dome, secondary glaciers, or interglaciers as they may be termed, have excavated valleys and amphitheaters. In the V-shaped mass of which Gibraltar is the apex, a broad amphitheater-like depression has been cut out, leaving a bold cliff above it. The excavation of the amphitheater did not progress far enough up the mountain to cut away the apex of the V-shaped mass, but left it with a precipice on its lower side. This remnant is Gibraltar. An attempt will be made later to describe more fully the process of glacial erosion of a conical mountain, and to show that the secondary topographic features of Mount Rainier are not without system, as they appear at first view, but really result from a process which may be said to have a definite end in view.

Below Gibraltar the descent was easy. Our life line was no longer needed. Tramping in single file over the hard surfaces of the snow field, remnants of the previous winter's snow, we made rapid progress, and about noon gained the scattered groves of spruce trees which form such an attractive feature of Paradise Park.

Fortunately, we found Prof. E. S. Ingraham, of Seattle, and a party of friends, including several ladies, encamped in Paradise Park, and the hospitality of the camp was extended to us. During the afternoon we basked in the warm sunshine, and in the evening gathered about a roaring campfire and enjoyed the society of our companions, who were enthusiastic in their praise of the wonderful scenes about their camp.

## PARADISE PARK.

The southern side of Mount Rainier is much less precipitous than its northern face, and the open park-like region near timber line is broader, more diversified, and much more easy of access. The general elevation of the park is between 5,000 and 7,000 feet, and it is several thousand acres in extent. Its boundaries are indefinite. It merges into the heavily forested region to the south, and into more alpine regions on the side toward the mountain, which towers above it on the north. To the east it is bordered by Cowlitz Glacier, and on the west by Nisqually Glacier. Each of these fine ice rivers descends far below timber line. The small interglacier, known as the Paradise Glacier, may be considered as lying within the limits of the park.

Paradise Park presents many and varied charms. It is a somewhat rugged land, with a deep picturesque valley winding through it. The trees grow in isolated groves. Each bunch of dark-green firs and balsams is a cluster of gracefully tapering spires. The undulating meadows between the shady groves are brilliant in summer with a veritable carpet of gorgeous blossoms. In contrast to the exquisite charms of the groves and flower-decked rolling meadows are desolate ice fields and rugged glaciers which vary, through many tints and shades, from silvery whiteness to intense blue. Added to these minor charms, and

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Something of the magnificence of Mount Rainier as seen by visitors to Paradise Park is revealed in the accompanying photograph (Pl. LXXI), here reproduced through the kindness of Mr. E. Curtis, of Seattle. The view was taken near sunset, when the chill of evening was just beginning to dissipate the vapors that enshrouded the summit of the great peak.

While the central attraction to the lover of mountain scenery in Paradise Park is the vast snow-covered dome of Mount Rainier, there are other mountains in view that merit attention. To the east rises the serrate and rugged Tattoosh range, shown on Pl. LXXII, which is remarkable for the boldness with which its bordering slopes rise from the forested region about it and the angularity of its many serrate summits. This range has never been explored except by miners and hunters, who have made no record of their discoveries. It is virgin ground to the geologist and geographer. Distant views suggest that the Tattoosh Mountains have been sculptured from a plateau, probably an upraised peneplain in which there existed a great mass of igneous rock surrounded by less resistant Tertiary sediments. The softer rocks have been removed, leaving the harder and more resistant ones in bold relief, to become sculptured by rain and frost into a multitude of angular peaks. This attractive, and as yet unstudied, group of peaks is in plain view from Paradise Park, and may be easily reached from them by a single day's tramp. Many other delightful excursions are open to one who pitches his tent in the alpine meadows on the south side of Mount Rainier.

### TO LITTLE TAHOMA.

Bidding our friends in Paradise Park good-by, we resumed our journey early on the morning of July 26. Ascending toward Gibraltar until an elevation of about 10,000 feet was reached, we turned eastward for the purpose of traversing the eastern slope of the mountain and regaining our camp at Winthrop Glacier. After crossing the upper portion of Paradise Glacier, we traversed broad and but little broken snow fields to the brink of the valley down which Cowlitz Glacier flows. Beyond Cowlitz Glacier, at about the same level that we had reached, we could see the bold, cathedral-like crags of Little Tahoma, the upward-pointing angle of a secondary mountain mass which divides Cowlitz and Emmons glaciers. Not wishing to descend into the deep valley before us and climb out again on the farther side, we chose to cross the névé fields to our left and endeavor to pass over a rugged and much broken region where the main current of Cowlitz Glacier descends a rocky slope about a thousand feet high. In following the route chosen we became involved in a succession of crevasses and ice precipices, which caused much delay. Slowly working our way upward, we reached the base of the highest ice wall, but a vertical cliff of ice about 50 feet high barred all further progress in that direction. Reluctantly we turned back and, losing all the advantage we had gained by three or four hours of excessively hard climbing, went down the central portion of the Cowlitz Glacier until we reached the level of the highest grove of trees on its left bank, and crossing to the land chose a delightful and well-sheltered spot beneath low pine trees at which to rest for the night.

Our camp was perhaps half a mile below where the ice stream flowing southward from Little Tahoma, and named Ingraham Glacier on the map forming Pl. LXVI, joins the main Cowlitz Glacier. Our bivouac was in a delightful locality, and would have furnished a pleasant camping place if we had been provided with the necessary blankets and rations with which to make life comfortable. As it was, we had to sleep on the moss without covering, our feet to a blazing fire. For food each man had one hard-tack and a cup of tea, without sugar or milk, for each meal. Near at hand was a cascade of clear water, furnished by the melting snow fields on the back of the spur of which Little Tahoma is the culminating summit. A larger cascade a mile below tumbled over the rugged cliffs and awakened the echoes on the precipices across the glacier. The music of the falling waters filled the intervals of our broken sleep. The full moon shed a soft light over the white glacier and rugged rocks, and added a charm to a wild scene that under more comfortable conditions would have been considered one of the most fascinating we met during our excursion.

All of the Cowlitz Glacier in view was heavily snow-covered, and broken in a systematic manner by marginal crevasses which left the shore at angles of about 45° and trended upstream. The appearance of this glacier at a later stage in the melting of its covering of snow is shown on Pl. LXXVI. The glacier is deeply sunk in a well-defined valley, and is, perhaps, the most characteristic example of a glacier of the alpine type to be found on the slopes of Mount Rainier.

Rising with the sun and partaking of our allotted single hard-tack and a cup of tea, we climbed the rugged cliff at the base of which we had passed the night, taking a route explored the previous afternoon by Smith and Ainsworth, and gained the summit of the cliff overlooking Ingraham Glacier.

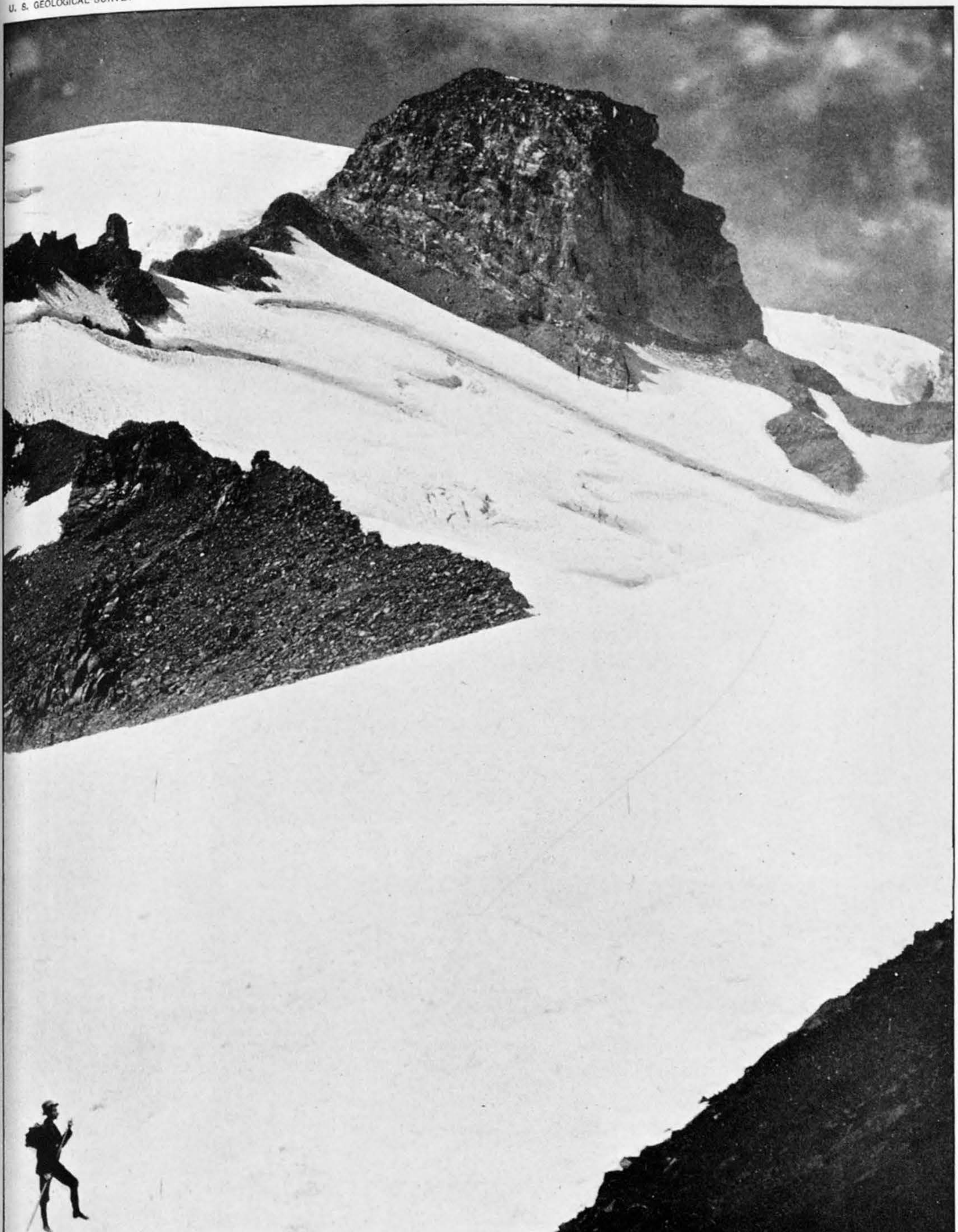
A deep descent of 500 feet over ledges of crumbling lava brought us to the hard ice below. An hour's tramp up the glacier and along the base of magnificent cliffs of volcanic agglomerate and lava in rude, irregular layers brought us to the sharp angle of rock at the base of Little Tahoma, against which the névé snows from the central dome divide. The descending snow meets the wedge of rock and is parted by it as the current is divided by the prow of a ship at anchor. The ice and snow, much shattered and standing in pinnacles, rises a hundred feet more against the obstruction in its downward course. The cliffs are or being cut away by the strong ice current, and rise above it in vertical precipices that culminate in sharp spires a thousand feet above the glacier. This proved a splendid locality for studying the manner in which the névé divides on reaching the medial portion of the main mountain side, and for seeing the modifications in the topography due to the erosion of glaciers flowing from the central dome.

To the north of Little Tahoma and flowing eastward is Emmons Glacier, the largest ice stream of Mount Rainier. This glacier is greatly broken, especially in its upper course, and has a peculiar feature not seen in connection with any other glacier. In its central portion there are two nearly parallel ridges of rock, which begin far up toward the summit of the main dome and descend for a mile or more below the level of the prow-like point of Little Tahoma. These parallel ridges appear from a distance to be medial moraines, and in part this is their true character. All along the dark belts in the center of the glacier, however, there are narrow and angular crests of rock in place, which project above the surface of the snow and ice. These crests of rock cause great crevasses to form in the ice flowing on either side of them and between them. For this reason Emmons Glacier is more difficult to cross than any other met with during our reconnaissance. The origin of the parallel rocky crests, the nature of which is shown to some extent on Pl. LXXIII, and their place in the topographic development of the mountain was not fully made out.

### ACROSS EMMONS GLACIER.

The névé of Emmons Glacier abreast of the prow of Little Tahoma was so greatly shattered that, looking down on it from the cliffs against which the névé divides, no practicable way for crossing could be chosen. Going down onto the glacier and following its side under the shadow of Little Tahoma for about a mile, we found the slope more gentle, with fewer crevasses, and then turned our faces toward The Wedge, where we had left our blankets four days previously. A way across the glacier was finally discovered, however, by patiently making trial after trial and going about the ends of the widest breaks. Many narrow crevasses were jumped, others were crossed by frail snow bridges. By persistent effort we at length gained the narrow crest of rock in the center of the glacier, and although our goal was in plain view, many difficulties had still to be overcome before we reached our former camping ground.

A hasty lunch and a cup of coffee renewed our strength. Taking



GIBRALTAR, A REMNANT OF THE CONE OF MOUNT RAINIER.

our blankets on our backs once more, we started on the homeward tramp down the Winthrop Glacier. The footprints made five days previously were clearly visible at first, but each impression stood in high relief, owing to the more rapid melting of the uncompacted snow about it. The descent was easy and rapid. By 5 in the afternoon we again joined the members of our party who had remained in the timber-line camp on the west border of Winthrop Glacier.

### THE RETURN.

Our plan was to carry our reconnaissance about the west slope of Mount Rainier, so as to gain at least a general idea of all of the glaciers flowing from the mountain, and of other features in the geography and geology of the region, but learning that the forest along the trail leading to Carbonado was on fire, a change became necessary.

Returning to our former camp on the west side of Carbon Glacier, plans were quickly made for dividing the party. Willis, with the camp hands, returned to the lower extremity of Carbon Glacier and thence to Carbonado, while Smith and myself, taking a small supply of rations, but without tents or blankets, turned our faces westward and visited Willis Glacier, Eagle Cliff, Crater Lake, and thence by way of the Spray Park trail and the main Willis trail also reached Carbonado.

Our reconnaissance extended from July 15 to 31, inclusive. Our route may be traced on the accompanying sketch map, Pl. LXVI, on which the glaciers and main topographic features of the mountain are indicated. The glaciers on the southwest side of the mountain between Willis and Nisqually glaciers were not visited, but some of their principal features were seen from distant points of observation.

## ICE EROSION OF AN ISOLATED, CONICAL MOUNTAIN.

The study of the manner in which the topography of Mount Rainier has been modified by ice erosion has suggested certain general principles which control the sculpturing of an isolated mountain peak sufficiently lofty to be crowned with perennial snow and to give origin to glaciers. Although these conclusions were reached after detailed studies of individual glaciers and of the topographic changes resulting from long-continued ice erosion, it is convenient in describing the glaciers of Mount Rainier to have in mind at least the general laws governing their distribution and behavior.

The geology of Mount Rainier and its present general form show that when in its greatest perfection it was a conical mountain, with gracefully concave sides. The upper portion of the mountain is formed to a large extent of fragments thrown out during explosive eruptions. Lava flows are also abundant, but did not greatly modify the character of the slope as determined by the falling of projectiles shot out of the summit crater. The primitive form was that of a typical scoria and lapilli cone of the character of Fusiyama, Japan, and other conical volcanic piles which still retain their youthful perfection of form.

Whether or not Rainier was truncated before the glaciers had greatly modified its lower slopes is unknown, but the truncation was not sufficient to have much influence on the character of the glacial mantle formed about it. In either case the height of the mountain—between 15,000 and 16,000 feet before the explosion that truncated its summit, and about 14,500 feet after that event—insured the gathering of perennial snows and the formation of névé fields and glaciers on its more elevated portions without an intervening period of weathering and stream erosion. The main topographic changes that have resulted must therefore be due to glacial action and to the eroding power of streams fed by the melting of the ice.

The weathering of rock masses unprotected by snow and ice has also assisted in the work of deforming the once symmetrical peak. Climatic conditions similar to those to which Mount Rainier is now exposed would lead to the covering of the mountain above an elevation of about 10,000 feet with a sheet of perennial snow. The snow would change to the condition of a névé, which would give origin to glaciers. Assuming that the peak was originally a perfectly symmetrical cone with smooth, even sides, and that the névé formed a uniform covering over the upper third of its surface, the downward flow of the névé would be equal in all directions. The spreading of the ice as it flowed down the cone, a progressively greater area being covered by it as it descended, would insure equal melting, except so far as that might be influenced by the unequal amount of heat reaching the southern and northern sides of the peak, and would either thin away uniformly in all directions or be gathered into local streams.

Many disturbing conditions come in, however, in the case of a peak like Mount Rainier, composed of loose agglomerate and lava sheets. Irregularities in the surface of the cone, erosion by streams flowing from the ice, unequal drifting of the snow as well as unequal melting owing to variations in exposure on the northern and southern sides, etc., would lead to the gathering of the descending ice into more or less well-defined streams. Individualized ice streams once established would hold their position and by their erosion would sink deeper and deeper into the rocks. From the extremity of each glacier a stream fed by the melting ice would carve a gorge or canyon leading to rivers on the plain below. As the ice gathered in well-defined streams melting would be retarded and the glaciers consequently extended farther and farther down the water-cut gorges. In this manner what may be termed *primary* glaciers would originate from the dividing of the descending névé.

Below the horizon where the primary glaciers divide the ice erosion would be confined to comparatively narrow channels, and would cut radiating trenches in the sides of the mountain. Excavation would be continued below the extremities of the glacier by the streams flowing from them, and the valleys would there be narrower than in their ice-filled depressions higher up. As the glaciers deepen their beds they sink into the mountain and are more completely sheltered from the sun, thus tending to perpetuate their own existence. Between the primary glaciers there would be portions of the lower slopes of the mountain left in relief by the excavation of the valleys between them. These V-shaped masses pointing up the mountain would form wedges against which the descending névé would divide to form primary glaciers. The Wedge and Little Tahoma are typical examples of such wedges.

The surfaces left as V-shaped masses between the primary glaciers when the division occurs above the snow line, or when a climatic change causes the perennial snow to descend lower on the mountain's sides, would become covered with snow fields, which would give origin to secondary glaciers. Interglacier, below The Wedge, and the small glaciers on the back of the V-shaped mass of which Little Tahoma is the culminating point, are examples of these secondary glaciers, or *interglaciers*, as they may be termed, after the typical example just mentioned.

The interglaciers excavate valleys and cut back amphitheaters, so that the surface of the original V-shaped mass between any two primary glaciers becomes hollowed out, leaving rocky crests along their sides. The ridges bordering the primary glaciers converge upward, and, uniting, form wedges, which, crumbling under the attacks of the destructive agencies of the air, become broken into pinnacle or other forms, the details in their sculpturing depending on the nature of the rock. The original V-shaped masses left by the intrenching of primary glaciers thus become skeleton forms, their borders and high, wedge-like, upward-pointing extremities alone projecting above the snow, except when the summer melting is far advanced.

As is well known, the erosive action of a glacier, other conditions being the same, depends on the gradient of its bottom. The abrasion of the bed of a normal alpine glacier increases from its terminus with increase of gradient up to a certain point, and then, if the gradient still increases and approaches the vertical, becomes less and less. The gradient that insures greatest erosion is not definitely known. It varies with the amount of débris with which the ice is charged, as well as with other conditons, but is apparently in the neighborhood of 30°. This law has an important bearing on the topographic changes that an isolated glacier-covered mountain passes through. Judging from the present condition of Mount Rainier and other similar isolated peaks on the Pacific coast, it appears that the most intense erosion occurs in a zone about half a mile broad, where the primary glaciers become distinct ice streams. In this zone the glaciers excavate canyons and thus increases the slope of the central mass of the mountain above the extremities of the V-shaped residual masses on its lower slopes. The heads of these valleys tend to become amphitheaters.

As has been shown, especially by Willard D. Johnson, of the United States Geological Survey—the results of whose studies, however, are not yet fully published—the cliffs encircling an amphitheater in which a glacier has its source gradually recede, owing to the disintegration of the rocks in the great crevasse, termed a *bergschrund*, which is formed near where the upward-sloping névé meets the rock walls inclosing it. The rocks in the bergschrund are shattered by changes in temperature and by the freezing of water in their crevices and interstices, and the loosened fragments are plucked out by the outward flow of the névé snow. The bergschrunds are filled with snow each winter and reopened the following spring. This process leads to an energetic sapping of the bases of the cliffs and a consequent recession of their walls. The faces of the cliffs encircling an amphitheater or cirque are commonly too steep to admit of the accumulation of snow upon them, but allow it to fall in avalanches and to be blown away. The cliffs, being bare of snow, are exposed to changes of temperature and are wind-swept. These various destructive agencies assist in the extension of the amphitheater toward the summit of the mountain, and also in its general enlargement.

The primary glaciers excavate deep, steep-sided valleys, but accomplish little toward altering the profile of an isolated peak as it appears when beheld from a distance. The even slopes of the mountain become broken by the intense erosion and the deepening of amphitheaters above the extremities of the buttressing wedges. When seen in profile the mountain slopes become broken at the horizon referred to, as may be seen in Pls. LXX and LXXI. There is then a long, sweeping ascent from the base of the mountain to the horizon where the wedges terminate, and then a downward slope toward the vertical axis of the pile, leading to a belt of moderate inclination which terminates above in the steeper slopes forming the sides of the central dome.

At the summit of a lofty, snow-covered mountain there is but little erosion. The snows are mostly blown away or heaped in pyramids on the crowning summits. On mountains of sufficient height—such height varying with climatic conditions—the snow does not melt so as to be changed to a névé, but is always dry and powdery. Under such conditions relief from indefinite accumulation is secured by the blowing action of the wind, by evaporation of the snow, and by its descent in avalanches. When bare rocks are exposed on a lofty mountain, as in the case of Mount Rainier, they may become polished by drifting snow, but this action is too slight to lead to noticeable topographic changes.

By the process outlined above a regular, smooth volcanic cone which reaches well above the snow line becomes sculptured by glaciers, so as to present a precipitous central mass, with a conical or dome-shaped summit, according to the shape of the original peak, bounded by the steep walls of amphitheaters and surrounded by secondary peaks and crags forming the apices of V-shaped residual masses between primary

MOUNT RAINIER, FROM PARADISE PARK; LOOKING NORTH.

Elevation of the foreground, about 6,500 feet; elevation of the summits, 14,000 to 14,526 feet.

glaciers. From these *tahomas*—using the name of the best example on the sides of Mount Rainier as a generic term—descending lines of crags forming rocky ribs lead down to the platform on which the mountain stands.

The extension of the amphitheaters at the heads of the primary glaciers renders the sides of the central dome more and more precipitous as glacial erosion progresses. Certain of the primary glaciers may advance with their task of excavating the slopes which they descend more rapidly than others and cut back their amphitheaters faster than do their neighbors, thus making the central dome unsymmetrical. This is the case in the present stage of the erosion of Mount Rainier. Carbon Glacier, flowing northward and having its amphitheater sheltered from the noonday sun, has excavated a great recess or cirque in the side of the mountain, while the glaciers on the south side of the peak have scarcely more than begun to form similar recesses.

The process outlined above, by which an isolated snow-covered cone is sculptured by glaciers and by the streams flowing from them, is not an ideal picture of what may occur, but is rather a statement of what has taken place on Mount Rainier. The regularity in the development of topographic forms that would result from the sculpturing of a conical mountain of homogeneous rock has been modified in the example before us by irregularities in the primitive form, due in part to the truncation of the summit, but to a much greater extent by variation in rock texture. Mount Rainier, as already stated, is a composite cone built of projectiles and lava streams. The portions formed of agglomerate are open in texture and yield readily to mechanical erosion. The lava flows are frequently dense and compact and stand in relief when the adjacent agglomerate is worn away. Inequalities in snowfall also occur, owing to the prevalence of westerly winds, and, as previously mentioned, snow melting is greatest on the southern, or possibly the southwestern, side of the mountain. At an early stage in the history of the mountain after it had ceased to be an active volcano the heat of the rocks must have influenced the rate at which the snow falling on it was melted. This influence was greatest and longest continued at the summit, as some effects of the residual heat of the rocks are still to be seen there. From this cause the summit may have been bare for a long time after glaciers originated on the sides of the mountain and had begun their task of eroding the rocks.

The varied conditions just enumerated, and possibly still others, would lead to modifications in the orderly sequence of topographic changes that a lofty isolated peak of homogeneous rock would pass through.

Changes in the topography of Mount Rainier have also been influenced by great variations in climatic conditions. The mountain is supposed to be of Tertiary age, probably Pliocene (as to the definite time at which it was built I can offer no direct proof), and was exposed to the climatic changes which were such a marked feature of the Glacial period. The extent of the glaciers that flowed away from the peak during the Glacial period and the variations they experienced have not been made out. Such facts as are in hand bearing on this question indicate that the primary glaciers were well established before the coming of the Glacial period and that the ice drainage during its maximum followed valleys previously outlined, which were greatly deepened.

In spite of the modifying conditions enumerated above, and possibly still others that may have influenced the manner in which Mount Rainier has been sculptured, the changes in topography that the mountain has undergone make a near approach to what should be expected from the erosion by ice on the upper portion and by water on the lower slopes of a symmetrical peak of homogeneous rock of the height of the example before us and under the climatic conditions to which it has been exposed.

## COMING TOPOGRAPHIC CHANGES.

With the postulate that existing climatic conditions will remain practically unchanged for a great length of time—the tendency of the glaciers to recede, discussed later, being checked—it is not difficult to sketch in outline the main topographical changes that Mount Rainier will pass through during the ages to come.

The primary glaciers are cutting back the cliffs encircling the amphitheaters from which they flow, and in their middle courses are slowly sinking into the sides of the mountain. This entrenching of the glaciers, by affording greater shelter for the ice, tends toward their preservation. The greatest changes in topography now in progress about Mount Rainier are at the heads of the glaciers that have made a marked advance in excavating their amphitheaters.

As an amphitheater recedes farther and farther into the side of an isolated peak the region favorable for the accumulation of snow and the growth of a névé becomes less and less extensive. For this reason the enlargement of an amphitheater leads to a decrease in the size of the glacier flowing from it. This is illustrated at the present time on Mount Rainier by Carbon Glacier, which has formed a large amphitheater on the north side of the peak. The snow on the less steep slope above the cliff leading to Liberty Cap creeps down to the verge of the precipice and there breaks off and forms avalanches, which descend to the glacier below. Carbon Glacier in reality has no true névé at present, except the snow-covered area just mentioned, above the summit of the cliffs encircling its amphitheater. The gathering ground of the glacier has been decreased by the extension of the amphitheater until it is but a fifth or a tenth of its original extent. Some compensation for the decrease in the size of the névé, as a glacier enlarges its amphitheater, is found in the fact that as the enlargement takes place the divides between the amphitheater and adjacent névé fields crumble away, and the glacier whose amphitheater recedes most rapidly diverts some of the snow drainage that was previously contributed to the neighboring glacier. This process is illustrated by Carbon Glacier, which has enlarged the western side of its amphitheater so as to divert some of the névé which formerly supplied Willis Glacier.

Carbon Glacier is still enlarging its amphitheater, and, if the process does not check itself by decreasing the area on which snow for the supply of the glacier accumulates, will cause such a recession of the cliffs at its head that the central dome of the mountain will become broken. The present symmetrical form of the mountain's summit will then be modified and replaced by a crest having a steep descent to the north and a much gentler southern slope.

If other glaciers about Mount Rainier also cut back their amphitheater so as to break the summit curve of the mountain, pinnacles and crests will replace the present rounded dome. In this stage there would be a central core, or a *matterhorn*, as it may be termed, with precipitous ribs and angular summit, surrounded by deep amphitheaters. When this stage is reached the gathering ground for névé fields will probably be so reduced that the glacier will shrink and have but little influence on the succeeding topographic changes which the mountain will pass through.

After the glaciers have ceased their work, weathering and stream erosion will continue. The mountain will become more and more angular, owing to the sculpturing of deep radial valleys down its sides, and the peaks and crests left between them. The history in store for the mountain when it is so altered in shape and reduced in elevation—with the exception of the central core—that glaciers can no longer be supplied by its snow, is illustrated by many ancient volcanic piles and need not be discussed at this time. Briefly stated, the topographic diversity of the mountain will increase for a time, and gradually the central core of hardened lava, filling the pipe through which eruptions formerly reached the surface, will be brought out in relief by the removal of the less compact agglomerates and lava sheets about it and stand as a tower above the ruins surrounding it. This "volcanic neck" will slowly crumble and sink to the condition of a rounded hill with rock fragments on its summit. The ultimate fate of the once glorious mountain is reduction to baselevel. When this topographic phase is reached, a plain, but little elevated above the sea, will occupy the place where now stands Mount Rainier.

## PRESENT CONDITION OF THE GLACIERS.

As stated on a previous page, the ice bodies on Mount Rainier may be classified, in reference to their position on the mountain, as primary and secondary glaciers. The secondary glaciers have also been termed interglaciers, the type being Interglacier, situated on the broad V-shaped remnant of the middle mountain slope between Winthrop and Emmons glaciers.

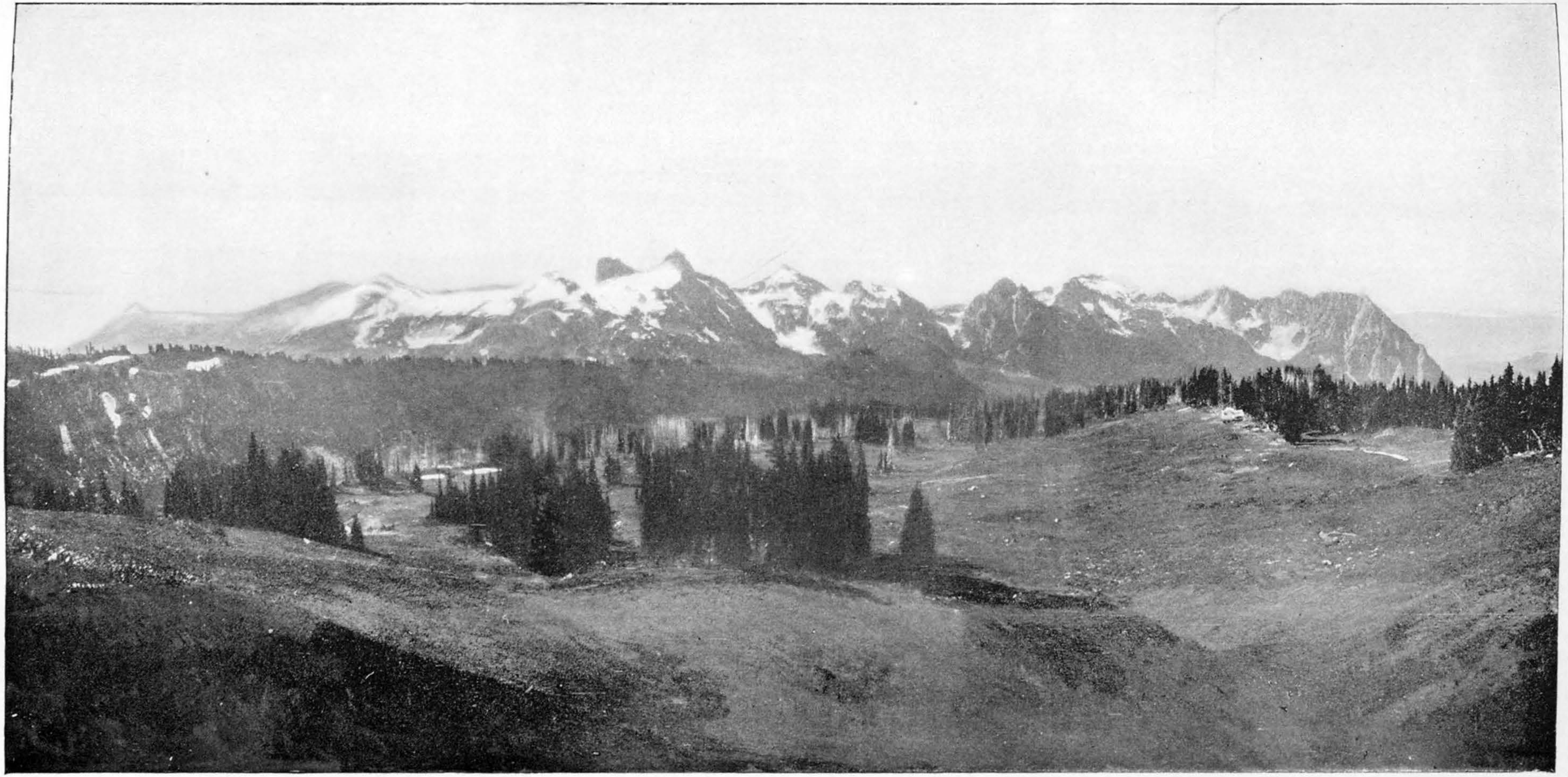
The primary ice streams in the order of their occurrence, beginning on the north side of the mountain and going about it toward the east, south, etc., are, as indicated on the accompanying map, the Carbon, Winthrop, Emmons, Ingraham, Cowlitz, Nisqually, Kautz, Wilson, Tahoma, Puyallup, Edmunds, and Willis glaciers.

The secondary streams, or interglaciers, as it is convenient to term them, in the order just stated, are Interglacier, Frying-pan, Little Tahoma, Williwakas, Paradise, Van Trump, and others not named. All the glaciers of this type are not represented on the accompanying map (Pl. LXVI), partly for the reason that they merge with indefinite snow fields, and some of them have not been recognized by those to whom we are indebted for the map. It is only in late summer or early autumn that the existence of glacial ice beneath the general snowy covering of the higher portions of the mountain can be distinguished. The interglaciers, as a rule, do not form well-defined ice streams, but are rather broad névés, from which a protrusion of glacial ice can be seen when the summer melting is far advanced. In some instances, as about the Guardian rocks, the glaciers of this type scarcely merit the name here applied to them, as they are little more than névés. These small ice bodies grade into snow accumulations which endure perhaps for several years, but are occasionally completely melted. A former extension of the interglaciers, and the previous existence of true glaciers where only deep snow accumulations now occur, is shown by the polish and grooves on the rocks below the positions they occupy. In common with the primary glaciers, those of the secondary type were formerly much more extensive than at present, but their condition during the Glacial period has not been fully determined.

## CARBON GLACIER.

The amphitheater in which Carbon Glacier has its source, as already stated, is the largest excavation that has been made in the sides of Mount Rainier. The wall of rock rising above the head of the glacier is about 4,000 feet high. On this vast precipice little snow accumulates, but on its summit there is a vertical cliff of stratified névé snow about 200 feet high. This wall of snow, conspicuous on account of its contrast in color with the dark rocks below, exposes a section of the névé which rests on the slope leading from the crest of the precipice to Liberty Cap. The slow downward creep of the névé causes portions of it to break off from time to time and descend in avalanches to the bottom of the amphitheater.

In the amphitheater is a névé, it is true, but not of the ordinary type. A deep accumulation of snow is formed there each winter by the avalanches and by snow blown from neighboring heights, in addition to that which falls directly, but the surface of the deposit thus formed is uneven, and, as may be seen in the numerous crevasses, lacks the well-marked stratification which is such a characteristic feature of normal névés.



TATTOOSH RANGE, FROM PARADISE PARK; LOOKING SOUTH.

The avalanches from the cliffs encircling the head of Carbon Glacier bring down considerable rock débris, and in summer rock-falls are common, owing to the action of the wind, the loosening of blocks by changes of temperature, etc. The snow in the amphitheater thus becomes heavily charged with débris, which is carried down by the outward flow of the névé, and becomes concentrated on its surface as melting progresses. On account of the strong winds that prevail much dust is also scattered over the snow flooring the amphitheater, which serves to darken its surface.

With the recession of the walls at the head of the amphitheater there has been an increase in its breadth. Some of the snow drainage previously contributed to Willis Glacier on the west, and to a former interglacier on the east, has been diverted and now feeds Carbon Glacier. As previously stated, the recession of the cliffs at the head of the amphitheater has decreased the area of snow accumulation leading to it, and thus has led to a diminution in the size of the glacier originating in it. The loss due to the recession of the cliffs, and the consequent decrease in the extent of the elevated region of snow accumulation, has apparently been much greater than the gain due to a broadening of the amphitheater and the consequent robbing of adjacent glaciers of their snow supply. Carbon Glacier, by enlarging its amphitheater, is slowly destroying the conditions on which its existence depends. On the central portion of the great wall at the head of the amphitheater there is a buttress-like ridge, which indicates the position of a former dividing wall. At an early stage in the sculpturing of the mountain there were evidently two amphitheaters which contributed their snow to Carbon Glacier. With the enlargement of these recesses in the mountain side the ridge that divided them has nearly disappeared.

The sides of the amphitheater are rocky crests, as may be seen especially on the east, where a dark ridge of rock, marked by pinnacles and crags, separates the snow drainage of Carbon Glacier from that of Winthrop Glacier. This line of crags and pinnacles is the crest of the wall of bare rock sinking westward to Carbon Glacier, but its eastern slope is much less precipitous and is heavily covered with névé snow. A small enlargement of the amphitheater will cause a break in the divide on the summit of the cliffs, and some of the snow now flowing to Winthrop Glacier will be diverted to feed its more energetic neighbor on the west.

The outstretching ridges forming the side walls of the amphitheater extend northward and merge at their lower or northern extremities into broad, interglacier spaces at an elevation of about 9,000 feet. The amphitheater is about 8,000 feet across from east to west, and in the neighborhood of 5,000 feet in length from the base of the great cliffs at its head to the break through which the glacier outflows.

At the outlet of the amphitheater the snow, still having the characteristics of a névé, is much crevassed, especially where it passes over bosses of rock on the floor beneath. Dome-like elevations are thus caused in the glacier, which are broken by radiating fissures into castle-like blocks with precipitous walls. Domes of this nature are a characteristic feature of many of the glaciers on Mount Rainier, and the fact that they are due to the ice passing over prominent rock masses is clearly shown near the extremities of several of the primary glaciers where the rocks have been exposed by the melting and recession of the ice. The nature of these domes will be noted more in detail in connection with the descriptions which follow of Winthrop and Willis glaciers.

Just below the outlet of its amphitheater Carbon Glacier passes down a somewhat steep descent, and is much broken. Opposite Andesite Cliff the surface gradient becomes much more gentle, and for about a mile and a half downstream the glacier descends a very gentle grade. To one walking up the glacier the rise in this portion is scarcely noticeable. At the lower end of the nearly level reach just referred to the ice again descends a steep slope—which, however, scarcely merits the name of an ice fall—leading to the terminus. This last steep descent is about 1,000 feet in a mile.

In brief, the glacier descends a moderately steep slope on leaving the amphitheater, flows for a mile and a half with a very gentle grade, and then goes over the edge of a precipice and descends a steep slope to its end. The alternate breaks and level reaches of the glacier, resembling a great stairway, are not a novel feature, as is well known, but a characteristic of many alpine glaciers. The level reaches separating breaks in a glacier indicate similar topographic features of the rock surface beneath. In studying existing glaciers it is instructive to examine valleys formerly glaciated. The best reported illustrations of the alternate step and terrace features in high-grade glaciated valleys occur in the Sierra Nevada, and have been described by me in a previous article.[1] In ascending Tuolumne Valley, near its source on Mount Lyell, for example, one encounters a series of steep escarpments separated one from another by almost level reaches, in some of which there are shallow rock-basin lakes. When the great valley referred to was occupied by a glacier each of the steep descents caused an ice fall. These conditions are reproduced on a small scale by Carbon Glacier.

As is explained in part in the report just referred to, and as has been still more clearly stated by Willard D. Johnson, a glacier cuts back its beds from one ice fall to another in much the same way that a cascade in a stream recedes. The formation of scarps and shelves, as the level reaches may be termed, is due to a process similar to that by which amphitheaters are enlarged.

Opposite Andesite Cliff Carbon Glacier is about half a mile broad, but it soon increases to nearly a mile in width, and maintains this increase all the way to the brink of the steep descent a mile and a half

[1] I. C. Russell, Quaternary history of Mono Valley, California: Eighth Ann. Rept. U. S. Geol. Survey, 1889, pp. 348, 354–355.

THE SOURCE OF EMMONS GLACIER, SHOWING THE SUMMIT AND EASTERN SLOPE OF MOUNT RAINIER.

below. Indeed, but little diminution in breadth occurs until the final descent toward the terminus begins. It then contracts in width somewhat abruptly to about 1,000 feet, and ends in a precipitous slope.

Opposite Andesite Cliff the hard blue ice of the glacier is exposed. In this region there is a reach of the glacier about half a mile long, intervening between the lower edge of the névé and the heavily moraine-covered ice farther downstream, which is comparatively free from dirt and stones. The extent and character of this surface vary from day to day during the summer, the névé receding and the dirt-stained and stone-covered area at the same time increasing upstream, owing to the concentration of débris at the surface as melting progresses.

Downstream from the belt of clear ice just referred to the glacier is progressively more and more deeply covered with stones and dirt. In this region many of the minor features characteristic of moraine-covered ice, such as glacier tables, sand cones, surface streams, moulins, etc., may be recognized. About 1,000 feet downstream from Andesite Cliff four somewhat prominent medial moraines make their appearance, and may be traced to the brink of the steep descent a mile below. Two of these moraines are near the west side of the glacier, the nearest being about 700 feet from its western border, and separated from its companion by a lane of less completely débris-covered ice about 150 feet broad. The other pair of medial moraines occupy a similar position adjacent to the east side of the glacier. These medial moraines are not conspicuous features, but are marked by irregular débris pyramids, rising from 10 to 30 feet above the adjacent surface.

The débris along the west side of the glacier, derived largely from Andesite Cliff, is gray, corresponding with the color of the cliffs from which it comes; but on all other portions of the surface the prevailing color of the moraines is dark brown. Practically all of the morainal material is angular. Rounded or smooth and striated stones are seldom seen.

Below the beginning of the steep descent leading to the terminus of the glacier no ice can be seen in a general view. The entire surface is buried beneath a sheet of brown, angular débris. The larger stones range in size from a few inches to several feet, and mingled with them are large quantities of fine, earth-like material. This portion of the glacier is rugged, on account of numerous crevasses and unequal melting, due to variations in the thickness of the débris. Something of the manner in which surface morainal material is concentrated in depressions and then raised in relief by the melting of the adjacent surface so as to form débris pyramids may there be seen. This process, however, goes on most actively when a glacier has but little motion or is stagnant, and about Mount Rainier may be best studied near the terminus of Winthrop Glacier. At the end of Carbon Glacier the ice descends precipitously 300 or 400 feet, and presents a dirt-stained face too steep for one to climb without cutting steps. At the foot of this steep descent Carbon River emerges from a cavern in the ice, as a brown, roaring torrent, heavily encumbered with bowlders. The river is overloaded with coarse débris, and is filling, or aggrading, its valley all the way to the narrow canyon 3 miles above Carbonado. The condition of the valley bottom below that locality is not known to me.

Where the river leaves the icy cavern from which it emerges many of the stones in its bed are angular; these have come directly from the steep ice slope above. Other stones, many of them 2 feet or more in diameter and composed of hard rock, are well rounded; these have been brought out of the subglacial tunnel, and show that much erosion is performed by the stream before it comes to the light. Less than half a mile below the terminus of the glacier nearly all the stones that form its banks are well rounded.

The end of Carbon Glacier was seen by Willis in 1881. At the time of our visit the glacier had retreated about 100 yards, as nearly as could be estimated, above the position it occupied fifteen years previous, and the precipice at its terminus had become much less steep.

Accompanying the recession of the terminus during recent years, there has been a general lowering of the surface of the glacier all the way up to the névé, but no measure of the amount that the ice had shrunk could be made. The lowering of the surface is of the nature of a general shrinking, which is greatest near the crest of the lower ice fall, and progressively decreases upstream. One conspicuous result of this surface melting is a marked increase in the amount of superglacial débris, as noted by Willis.[1]

A recent lowering of the surface of the glacier is recorded by abandoned lateral moraines. These are conspicuous along each side of the glacier, from the brink of the lower icefall up to Andesite Cliff on the west bank and to the entrance to the amphitheater on the east side. These moraines, composed largely of sandy clay heavily charged with stones and angular rock masses, rise from the ice in precipitous escarpments from 100 to 140 feet high, and are too steep in many places to be climbed. The angle of slope toward the glacier is from 40° to 50°. These slopes are broken faces, from which stones frequently fall, and are entirely bare of vegetation. The slope referred to rises to a sharp crest, from which the slope away from the ice is more gentle, averaging between 30° and 40°. On these outer slopes spruce trees, some of them 20 to 30 feet high, are growing.

On leaving the glacier on either side and climbing the fresh slopes of morainal material bordering it, one finds other similar, parallel ridges, each of which is clothed with forest trees. These older moraines are in several instances higher than the most modern one, and show in general a progressive lowering of the surface of the ice as the width decreased.

[1] Seventeenth Ann. Rept. U. S. Geol. Survey, Part I, 1896, p. 53.

The narrow valleys between the abandoned moraines are without forest trees, but are carpeted with moss, grasses, and a profusion of brilliant flowers. The snow lingers until late in summer in these shaded dells, and the black, humus soil, after the snowdrifts disappear, is water-soaked and in many places swampy. Thus in several ways the ridges favor the growth of forest trees, while the marshy dells between furnish conditions suitable for the growth of more lowly plants.

The abandoned lateral moraines below Andesite Cliff are in an embayment on the side of the valley. Their formation illustrates the manner in which a glacier builds moraines where the valley widens, and thus tends to make its channel even sided. The most recent moraine—the one overlooking the ice—starts from Andesite Cliff, but not at its most exposed portion. The next lateral moraine to be formed if the glacier continues to shrink will, it is to be expected, start from the extreme end of the cliffs where they now project into the glacier.

The older lateral moraines, of which there are sometimes three, and again four, abreast, terminate at their upper or proximal ends abruptly, without uniting with the cliffs from which they derived much of the material composing them. The reason for their abrupt ending is that a lateral stream following the side of the glacier cut a channel across them at their junction with the side of the valley. This process is plainly illustrated on the west border of Winthrop Glacier near its terminus.

The moraines on the east side of Carbon Glacier are in general like those just mentioned, and again all but the last formed are without connection at their proximal ends with the more elevated region above.

All of the moraines just described pertain to the present topography and were formed when the glacier had its present characteristics, except that below Andesite Cliff, when the earliest pair was formed, it was about a mile broader and its surface about 250 feet higher than now. Whether the valley was ever more deeply filled with ice than is recorded by these old moraines remains to be determined.

## WINTHROP GLACIER.

The névé of Winthrop Glacier extends to the summit of Mount Rainier. A part of the snow that accumulates in the great summit crater between Crater Peak and Liberty Cap flows eastward down the precipitous slope of the central dome and contributes to the growth of the extensive névés covering all that side of the mountain. The eastern slope of the mountain is more heavily snow covered than any other portion, mainly for the reason that the prevailing westerly winds cause the snow to be deposited there in greatest abundance. The great peak rising in the path of the moist winds from the Pacific produces something like an eddy in the air currents on its eastern side, and thus favors deep snow accumulation.

All of the eastern side of the mountain above an elevation of 8,000 to 10,000 feet, except the precipitous ridges and jutting crags, is névé covered. The appearance of this side of the mountain in summer is shown on Pl. LXIX. As may be seen in the illustration to some extent, the snow is much crevassed, and is broken by faults where the cliffs are steepest. This is due to the ruggedness of the rocky slopes beneath, which yield unequally to the erosion of the descending ice and snow, but no amphitheaters or cirques have been excavated.

Near its lower limit the névé is divided by two rocky promontories, as has been previously described, known as The Wedge and Little Tahoma. These prowlike rock masses divide the névé into three primary glaciers—the Winthrop, Emmons, and Cowlitz—as shown on the accompanying map.

The névé of Winthrop Glacier descends below The Wedge, and terminates above timber line at an elevation of approximately 8,000 feet. Below the lower margin of the névé the solid blue ice of the glacier proper, in places heavily covered with débris, extends far down the valley between rugged mountains, and ends at an elevation of between 4,000 and 5,000 feet.

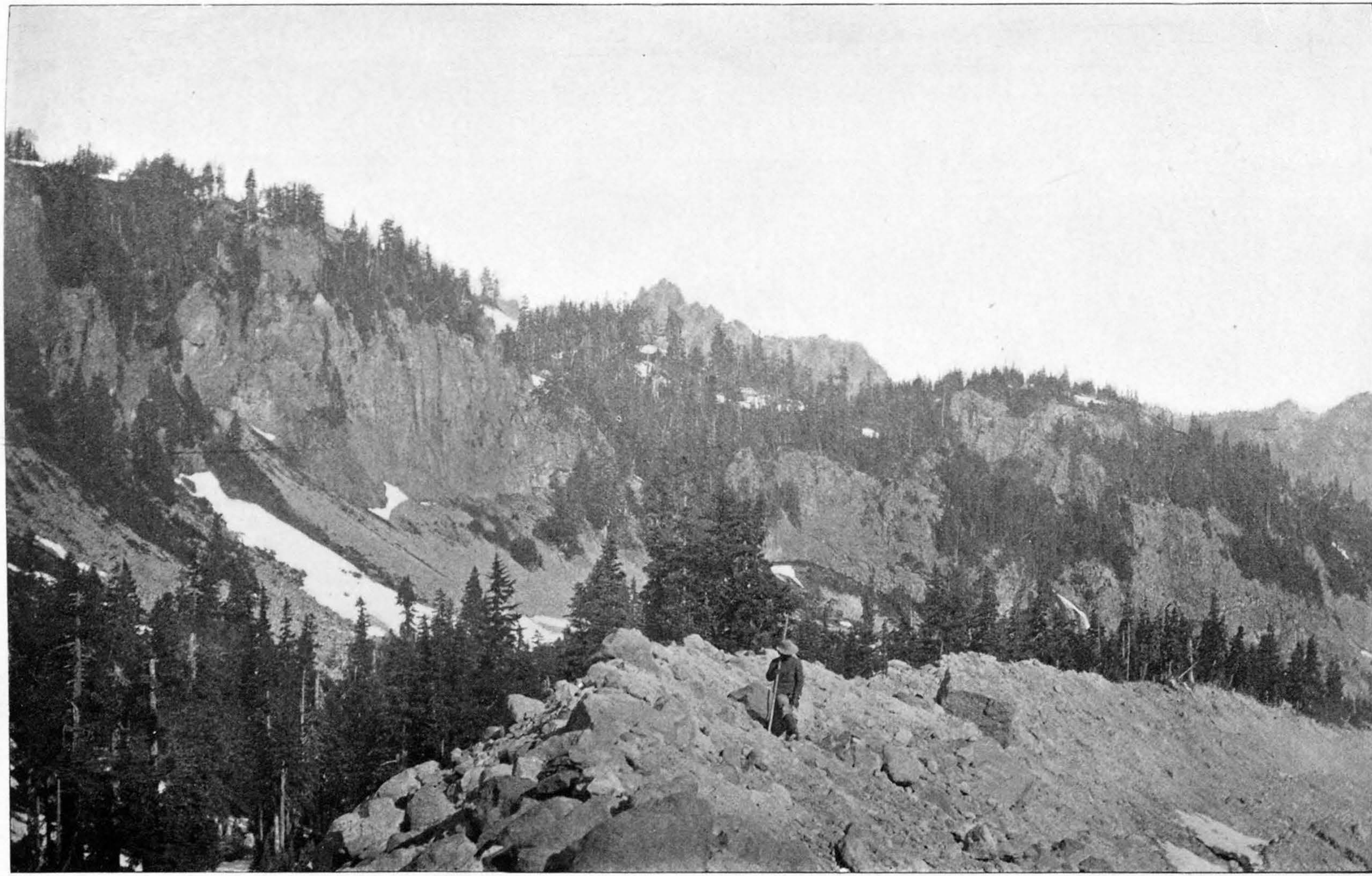
From the end of the glacier one branch of White River flows out as a swift turbid stream, heavily loaded with coarse débris.

One of the characteristic features of the glaciers about Mount Rainier, as already mentioned, is the occurrence of well-marked domes, the summits of which are commonly fractured so as to produce radiating crevasses. Several of these domes occur in Winthrop Glacier, both in the névé portion and in the glacier proper, and furnish abundant material for study.

The domes in the glaciers have the appearance that might be expected to result if a sheet of ice or of névé snow 100 to 200 feet thick could be lowered down vertically onto a surface on which there were haystack-like domes of rock 100 to 300 feet high. In such a case the ice over the domes would become fractured, while on the generally even surface between it would settle down with less breaking and conform to the general contour of the rocks beneath.

We know, however, that the snow and ice supplying the glaciers flow from higher regions, and that it must advance over the domes. This means that the ice has in fact an upward motion, for the glaciers rise in passing over the elevations in their beds. From the summits of the domes the snow, or ice, as the case may be, descends in all directions, but usually the slope is steepest on the downstream side.

A series of ice domes on the glaciers about Mount Rainier might easily be selected ranging from regular domes with practically equal slopes on all sides, through other similar forms with precipitous lower face, to precipices down which the ice descends, forming what are termed ice cascades. From analogy with the ice cascades I shall designate the elevations on the surfaces of the glaciers here described *ice domes*. Like the cascades, the domes vary in their characteristics according as they occur in a névé region or in a glacier proper.



RECENTLY ABANDONED MORAINE ON WEST SIDE OF CARBON GLACIER, JUST BELOW ANDESITE CLIFF.

The domes in the névé region, as on the upper portion of Winthrop Glacier below The Wedge, differ in appearance from those in the glacier proper, owing to differences in the physical properties of névé snow and ice. The domes in the névé have more or less of a network of wide fissures about their summits, with radiating fissures extending down the sides, which gradually contract and die out in the course of 100 or 200 yards. The radiating fissures on the upstream sides of the domes are commonly less extensive and much less conspicuous than those extending in other directions. The fissures in the summits of the domes in the névé at the time of our visit were widely expanded, and when the temperature was below freezing one could safely walk over the rough cavernous snow partially filling them and penetrate to the very center of the system of breaks. In some instances a nearly rectangular column of snow 30 feet on a side and rising 50 or 60 feet above the partially filled crevasses bordering it rose in the center like a huge obelisk.

Farther down Winthrop Glacier, in the glacier proper, the domes are equally conspicuous, but the crevasses in them are less regular and not nearly so wide as in the instances just cited. Just below the skirt of the névé as it existed at the time of our visit there are four domes inclosing a basin in which there was a shallow pond about an acre in area, which was conspicuous on account of its blue color. These domes rose, by estimate, about 50 or 60 feet above the ice on the upstream side, but the ones farthest downstream, when seen from below, had nearly twice this height. Their slopes on the upstream side descended at an angle of about 4° to 6°, but on the downstream side they were considerably steeper, probably 10° to 12°, and in one instance at least 20° or 25°. Just above each dome there is an absence of crevasses, and the ice seems to be compressed, but small breaks appear soon after the ascent begins. These crevasses are curved and tend to surround the dome as contours, but near their ends trend away from it in curves concave toward the center and die away. The crevasses become broader on the summit of the dome, especially on the brink of the steep downstream side, and as the ice descends are closed. The lower side of one of the larger domes was visited, and a hollow in the surface of the glacier was found below it. Standing in this hollow and facing the dome, one sees a steep descent, much like an icefall, with pinnacles of blue ice along the crest. On each side of the dome there is a curved ridge of ice, making the east and west walls of the depression in which the observer stands. These curved ridges become lower and lower as they leave the dome, and, uniting, completely inclose the depression, the lowest point on the rim being opposite the dome. A stream fed by surface-melting coursed along in the bottom of the depression, flowing in the direction of glacial motion for 5 or 6 rods, and then plunged into a crevasse. In the depression below other similar domes there were shallow ponds.

The barriers below the domes are evidently due to a strong flow of ice about their sides, which formed the lateral curved ridges that unite a short distance below.

That the domes are due to bosses of rock rising in the bed of the glacier is shown, as previously stated, by the appearances of such domes at the extremities of some of the glaciers which have been uncovered by the melting of the ice. One such dome stands near the extremity of Winthrop Glacier, and now forms a portion of its west wall, but was formerly completely surrounded and overridden by the glacier. The exposed rock domes are rounded, strongly glaciated, and more or less covered with morainal material. Perched bowlders, which a strong push would dislodge, are frequently seen on them. The rock domes are, in all observed instances, composed of dense, hard rock. One in Carbon Glacier, on the brink of the lower fall, is of granite. The others observed are of dense igneous rock, probably andesite. The rock about the bases of these domes is perhaps softer and more easily eroded, but that this is really the case has not been proved by observation, owing to the débris that occupies the depression and conceals the rock beneath. The impression that one gains on examining the rock domes exposed by the retreat of the glaciers is that they result from ice erosion and have been left in bold relief by the wearing away of softer rocks about them.

The manner in which a glacier rises over a dome is of interest in connection with the much-discussed problem of the cause of glacial motion. A detailed study of ice domes would apparently furnish evidence for deciding whether a glacier behaves as a rigid body that is thrust forward by a force acting from above, or as a plastic body moving under the influence of gravity. The manner in which the ice, in the case of a dome in the glacier proper, flows about the obstruction so as to form lateral ridges which gradually approach and inclose a depression below certainly favors the idea of plasticity. The ice appears to be forced over the summit of a rock dome and to rise higher than the adjacent surface upstream, both by the pressure of ice above and by the drag of the deeper ice current on either side.

It is difficult to conceive of the manner in which a rigid body would behave under the conditions here referred to, but there appears to be no reason for assuming that it would close in below the domes; we should expect, rather, to find an open channel below each obstruction, with precipitous and much-broken walls.

On the east side of Winthrop Glacier, below The Wedge, the rocks rise precipitously and form cliffs, from which much débris falls. The glacier is evidently sapping the cliffs that border it from The Wedge to near its termination. On the west side the limit of the névé fields is indefinite, there being many crags that rise through the snow above the timber line.

Below that horizon, however, the glacier's margin is sharply defined by bordering precipices. The ice has there melted back from rugged

UPPER ICE CASCADES OF THE COWLITZ GLACIER.

cliffs so as to leave a marginal valley some 200 feet deep, in which a stream flows. The margin of the glacier is heavily moraine-covered and much broken by crevasses. In places it is impassable. The extremity of the glacier, as already stated, flows past a bold rock dome, which was formerly covered by the ice, and at a later stage, as the glacier receded, divided into two branches, the eastern one being the broader. As the glacier continued to retreat, the tongue of ice to the west of the rock dome melted back, and a heavy lateral moraine was deposited as a free ridge, having a curved course, which extends out from the shore and joins the rock dome. The end of the moraine near the shore has been cut away by the stream following the side of the glacier, and when traced toward the border of the valley is found to end in a precipitous slope. Standing on the crest of the moraine, on which a few small spruce trees are growing, one can look down into a deep valley to the west of the rock dome, across which the moraine has been built. The slope of the moraine on that side is steep and the descent is about 400 feet. On the side overlooking the glacier the descent is still more precipitous. The ice has shrunk away from the moraine and is now fully 100 feet below its crest.

Above the rock-dome and bordered by the sharp-crested moraine just mentioned there is an embayment in the side of the glacier, occupied by heavily débris-covered ice that is fast melting away. This stagnant ice has a markedly different aspect from that on other portions of the glacier where motion is still in progress.

The topography of stagnant moraine-covered ice is difficult to describe, but easily recognized. A prominent feature is the great number of short, steep ridges or crests, rising to a broad, gable-like angle in the center and bounded on one side by a steep crescent-shaped slope of dirty ice and on the other by a much more gentle slope, which is heavily moraine-covered. The steep slopes usually face southward, but this is not an invariable rule. Below the steep slopes, and filling the spaces between adjacent crests, there is usually a deep accumulation of stones and dirt. As melting progresses these concentrated masses of morainal material are left as débris pyramids. The moraines on stagnant ice are usually darker than on other portions of a glacier, probably on account of the greater abundance of fine, earth-like débris, which retains moisture and is always wet. The steep slopes of the ice crests in the case of the Mount Rainier Glacier are dark-brown and frequently appear nearly black, as is also true of the stagnant border of Malaspina Glacier, Alaska. These dark, wet slopes are not common on moving ice.

The eastern branch of Winthrop Glacier ends in a low slope to the east of the rock-dome referred to, and has but little morainal material in it. Above the low terminus the ice rises steeply, and has been washed clean by surface streams. Above this steep rise the ice is profoundly crevassed and can be crossed only by patiently choosing a way from one narrow ice blade to another between deep gashes in the solid blue ice.

There are many other features of Winthrop Glacier that demand attention, but in the absence of a detailed map and other illustrations an attempt to describe them is probably not advisable.

One interesting feature near the end of Winthrop Glacier, reported by E. S. Ingraham, but not seen by me, is a deep, narrow cleft in the rock, on the border of the valley and parallel to its longer axis, about a mile below where the glacier now terminates. This gorge is probably the result of stream erosion along the side of the glacier when it was much more extended than now. That this is the true explanation of its origin, however, remains for future travelers to determine.

### EMMONS GLACIER.

The best example of the manner in which the general névé field descending the slopes of the central dome of Mount Rainier is divided by wedges of rock into primary glaciers is furnished by Emmons Glacier and its neighboring ice streams. The origin of the wedges now dividing the névé, by the erosion of the outward-flowing ice, has already been explained.

Below The Wedge and Little Tahoma, Emmons Glacier is a well-defined ice stream, about 5 miles in length, with bold, rocky cliffs on each side. The glacier becomes heavily charged with débris along its borders from the adjacent cliffs, and in the lower portion of its course is completely covered with stones and dirt on each side. These lateral moraines become broader and broader toward the terminus of the glacier, leaving a tapering, lane-like tongue of clear ice between, but before the actual terminus is reached the ice over the entire surface is concealed by a continuous sheet of brown and barren débris. On the right-hand side of the glacier, for one or two miles above the terminus, abandoned lateral moraines occur in parallel ridges, marking a gradual shrinking of the ice. A similar record occurs also on the left side, but the moraines are there broader, and show by their color and by the relief of the surface that they rest on stagnant ice. On the side of the valley, above the stagnant moraine-covered ice just referred to, there are abandoned moraines which are banked against the steep cliffs forming the valley side.

The tongue of clear ice near the extremity of the glacier is some 2 or 3 miles long and, although gradually tapering downstream, is much of the way about one-third of the width of the valley. The grade is there low and the ice not much broken by crevasses. Down the central portion of this tongue there are two light medial morainal bands, derived from rocky crests in the central part of the névé mostly above the horizon of Little Tahoma. These narrow rocky ledges may be seen in Pl. LXXIII.

SURFACE OF COWLITZ GLACIER.

The manner in which débris causes a decrease in the flow of the ice containing it, and final stagnation if the rock material reaches a large percentage,[1] leads to interesting suggestions in the case of the glacier here described.

The excessive load of stones and earth on the sides of the glacier, the central portion being notably free from such accumulations, causes the marginal portions to become heavily charged with englacial débris as melting progresses. The flow of the ice is thus checked, and the marginal portions of the glacier become stagnant. This is equivalent to a narrowing of the valley through which the ice flows, and the clear portion is enabled to progress farther before melting than it would if the border had not become stagnant. Emmons Glacier, like all the other primary glaciers on Mount Rainier, is evidently wasting away and its terminus receding. The process just referred to, by which the channel available for the flow of the clear ice becomes contracted, might lead to an advance of the terminus in spite of the fact that a general wasting away is in progress. The stagnant ice along the sides of the valley not only causes a decrease in the width of the channel, but shields the clear ice from melting more effectually than cliffs of rock would do. The clear ice under the present conditions is also able to advance farther than it would in a broad valley where its sides would be exposed to melting; or than it would between cliffs and rocks which, by reflecting heat, would cause even greater marginal melting than would occur in a broad valley.

Emmons Glacier is deeply intrenched. Near its terminus it is bordered on each side by bold, rugged mountain ranges, left in relief by the excavation of the valley in the which glacier lies. The valley becomes narrower and its sides still more rugged below the terminus of the glacier. The stream flowing down it, a branch of White River, like other similar rivers already mentioned, is overloaded with coarse débris and is aggrading its channel.

## INGRAHAM GLACIER.

The portion of the névé descending the east side of the central dome of Mount Rainier, to the right or south of Little Tahoma, and divided by that promontory from the portion of the névé tributary to Emmons Glacier, forms a primary glacier of an abnormal type. This well-defined ice stream does not descend the mountain slope in direct course, but is deflected southward or becomes tributary to Cowlitz Glacier. Its course is oblique to what may be considered the normal flow of a primary glacier originating on an isolated peak.

Where the descending névé is split by the sharp edge of the great wedge-shaped remnant of the side of the mountain, of which Little Tahoma is the culminating peak, the ice rises against the rocks and is

[1] I. C. Russell, The influence of débris on the flow of glaciers: Jour. Geol., Vol. III, 1895, pp. 823–832.

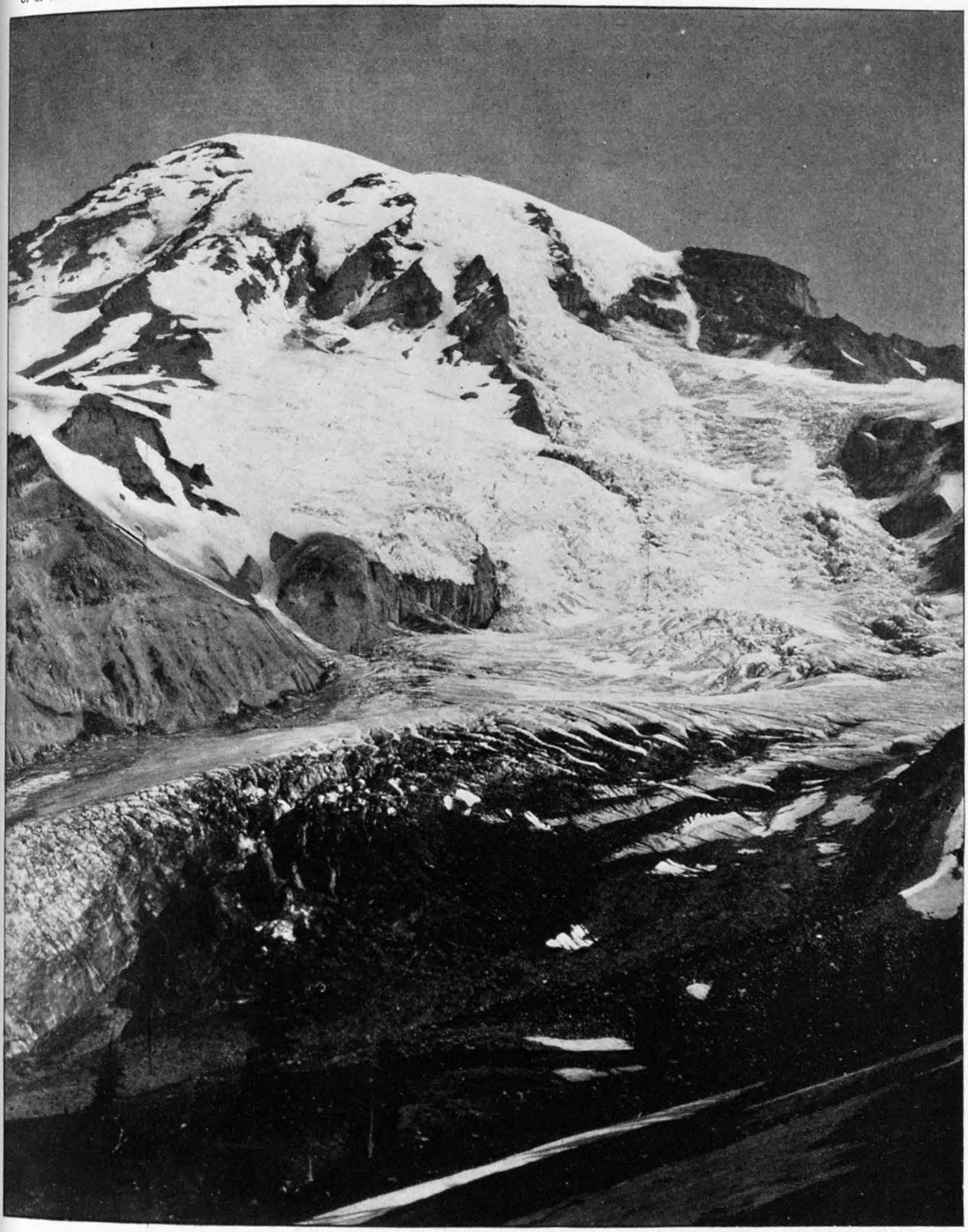
greatly shattered. The portion of the névé turned southward by the obstruction now descends a steep slope and is much crevassed; below the slope the descent is more gentle, and the hard, blue ice, but little encumbered with débris, flows on as a well-defined glacier. All along the left side of the glacier the cliffs of Little Tahoma rise in rugged precipices, from which débris is continually falling. On approaching its junction with Cowlitz Glacier, Ingraham Glacier descends a precipice about 800 feet high and forms a fine ice cascade. Something of the appearance of this steep ice-covered descent is shown in the accompanying illustration (Pl. LXXV), in which the head of Cowlitz Glacier and the bold cliffs of bedded lava and agglomerate between the two glaciers are also shown. In this illustration, also, the bedded character of the rocks forming Little Tahoma, and the manner in which the irregular strata end in the air on the left—that is, toward the summit of Mount Rainier—may be recognized. Before erosion altered the topography of the mountain, and previous to the blowing away of its summit, the strata in Little Tahoma were continued upward to the summit of the perfect volcanic cone.

The illustration to which attention has been directed is unsatisfactory on account of its "flatness." The distance from the rocks in the foreground, across Cowlitz Glacier, to the cliff of inclined beds between the two ice falls is fully a mile. The peak of Little Tahoma is about 3 miles distant. With these measures in mind the photograph reproduced in Pl. LXXV becomes more intelligible.

## COWLITZ GLACIER.

The Cowlitz Glacier, above where Ingraham Glacier joins it, expands somewhat and occupies an irregular depression, having some of the features of an amphitheater. The slopes at the head of the depression are so sharp that the snow descends in avalanches. The main portion of the névé is comparatively low on the mountain side, but some of the snow drainage comes from near the summit. The valley occupied by the névé contracts just above where Ingraham Glacier comes in, and it is there that the lower limit of the névé probably occurs, but at the time of our visit the glacier was snow-covered and white as far down its course as we were able to see. The snowfall during the winter preceding our visit was unusually heavy, especially on the south side of the mountain, and the melting of the snow in spring was long delayed. Photographs taken in previous years (Pl. LXXVI) show that ordinarily in late summer the solid glacial ice is exposed and forms an extremely rugged, dirt-covered surface, where we walked with ease over hard snow.

The portion of Cowlitz Glacier below Ingraham Glacier is inclosed by bold cliffs, and is well defined. There is less evidence of shrinkage along its sides than in the case of the other glaciers examined. This is possibly due in part to the fact that Ingraham Glacier has a high grade



NISQUALLY GLACIER.

and probably conducts away much of the snow that might be considered as belonging to Emmons Glacier. A sharp-crested lateral moraine across a side expansion of the valley, a mile below Ingraham Glacier, is evidence, however, of a recent lowering of the surface of at least 75 or 100 feet.

The lower portion of Cowlitz Glacier was not seen by me, but, judging from what I observed, I should say that it furnishes the most typical example of an Alpine glacier that exists about Mount Rainier. It is readily accessible from Paradise Park, and could conveniently be made the subject of special study.

### NISQUALLY GLACIER.

The Nisqually Glacier flows past Paradise Park on the west, and as this beautiful region is visited each summer by hundreds of tourists, it is, in a general way, the best known of the glaciers on Mount Rainier. My visit to the southern side of the mountain was too brief, however, to admit of my becoming acquainted with the glacier that is so familiar to many others, and instead of attempting a description of it I wish rather to suggest observations that visitors to Paradise Park may easily make.

Nisqually Glacier heads in two névé fields, which occupy what may be termed incipient amphitheaters, situated below the level of Gibraltar. Each of these depressions receives snow from avalanches that descend the steep slopes above them. The easterly névé, the one nearest Gibraltar, however, is fed by two snow streams, which endure through the summer, and from ice cascades, on which small avalanches frequently occur. These features are shown on Pl. LXXVII, which also illustrates the manner in which the glacier is crevassed and the extent of the sheet of débris covering its borders just below where the two névés unite. In this same region there are several small ice domes, one of which may be recognized in Pl. LXXIX. Visitors to Paradise Park could render assistance to students of glaciers by obtaining good photographs of this ice dome from time to time, and thus furnishing a record of the changes it passes through. By boring 2-inch auger holes to a depth of 6 or 8 feet in the ice and inserting vertical stakes, the daily rate of surface melting could be measured. A row of such stakes placed in line across the glacier would furnish a means of measuring the flow of the ice.

The Nisqually Glacier narrows to a well-defined stream to the west of Paradise Park, and at its terminus there is an archway from which Nisqually River rushes out. Photographs of the terminus taken at several dates during the summer from the same point of view—which should be so marked that it could be rediscovered year after year, for the purpose of obtaining additional pictures—would furnish data for ascertaining the advances and recessions of the glacier. To aid in these observations records should be made on the cliffs, either by chiseling into the rock or by painting an appropriate mark upon it, which would mark the position of the ice front at a definite time and admit of measurement of the fluctuations of the end of the glacier in subsequent years.[1]

Nisqually Glacier affords abundant opportunity for observing and studying the various minor features that glaciers present, such as the nature of crevasses, and their different types; the nature and origin of moraines; the manner in which lateral moraines are left stranded on the sides of a valley by the lowering of the ice; glacial tables, and the changes they pass through during the season of melting; sand cones; débris pyramids, etc. The delights of camp life in Paradise Park can be greatly enhanced by anyone who chooses to avail himself of the advantages for glacier study there afforded.

The glaciers on the southwest side of Mount Rainier have not been seen by me, and no data are available concerning them except such as are given on the accompanying map.

## WILLIS GLACIER.

On the northwest side of Mount Rainier and at the head of the deep, narrow valley through which the north branch of Mowich River flows, there is a glacier, known as the Willis Glacier, which is perhaps the most interesting of any here considered. It has many of the features of the primary glaciers already described, but is of small size, and one may see all its characteristic features in a single day's excursion.

From the summit of Eagle Cliff—where may be seen the most magnificent of the views about Mount Rainier, and in fact one of the most sublime pictures of noble scenery to be had anywhere in America—the whole of Willis Glacier, from the snow fields that give brilliancy to Liberty Cap down to the dirt-stained and crevassed extremity of the ice stream, is embraced in a single view. Below the end of the glacier the river, rising in two branches from its divided extremity, rages over its bowlder-filled channels and between craggy mountains which rise in precipices on either hand. To an observer on Eagle Cliff the distant roar of the troubled waters in the wild valley at his feet is mingled with the softer music of creeks and rills that plunge down the bordering cliffs to join the torrent below.

From Eagle Cliff the manner in which Willis Glacier is divided at its extremity into two moraine-covered tongues of ice is a noticeable feature. The bold, rocky eminence that causes the division rises steeply in the center of the valley to a height of fully 1,000 feet, and is clothed on its downstream side with forest trees. For the sake of a name, I

[1] A committee of the International Congress of Geologists was appointed in 1894 for the purpose of recording observations of the nature above suggested. Prof. H. Fielding Reid, of Johns Hopkins University, Baltimore, Maryland, is a member of this committee, and will furnish instructions for making the desired observations. Photographs, or other data concerning the fluctuations of glaciers, forwarded to Professor Reid will be carefully preserved and utilized to the best advantage, full credit being given those furnishing the information. See the Journal of Geology, Vol. III, 1895, p. 284.

MARGINAL CREVASSES, NISQUALLY GLACIER.

propose to call this bold, isolated eminence Division Rock. When Eagle Cliff was first reached by Willis in 1881 the glacier since named in his honor was larger than now, and extended to near the summit of Division Rock. On gaining the brink of the cliff immediately to the north of the summit of the rock, which is conspicuous from many points of view lower down the valley, the ice broke off so as to form a precipice at the summit of the rock cliff. As will be shown below, this observation enables us to make a rough measure of the amount that the end of the glacier has receded during the past fifteen years. On reaching Division Rock the glacier divides into two tapering tongues, more fully described below, and as seen from Eagle Cliff is marked by three rather faint medial moraines, separated by lanes of clear ice. The central moraine ends at the division produced by the rock, and the ones on each side follow the central portion of the tongue of ice in the valley at its sides.

In visiting Willis Glacier I approached it from the east by way of the Guardian Rocks, and came to the brink of the canyon through which it flows midway between its terminus and the steeper portions of its névé. The crest of the cliffs of jointed andesite bordering the glacier on the northeast rise more than 1,000 feet above its surface and furnish exceedingly favorable localities for observing the glacier at their base. From such a station the entire glacier, from the clear white snow fields culminating at Liberty Cap down the steep side of the central dome, where the névé is much broken, to the blue ice below, and on to the moraine-covered terminus, is in full view. The entire distance from Liberty Cap, where the snow accumulates, to the extremity of the glaciers, where it melts away, is approximately 5 miles. The distance from the head of the slightly defined amphitheater in which the glacier heads to its terminus is about 3 miles. The breadth of the glacier where its borders are best defined, about a mile above its terminus, is approximately 3,000 feet. These measures, or rather estimates, show that the glacier is small, but the many interesting features it exhibits make it one of the most instructive of the glaciers in the system to which it belongs.

The north wall of the canyon through which the glacier flows is exposed to the sun, and the snow is melted from it early in the summer, while the south wall nearly to the end of the glacier remains heavily snow-covered throughout the year. From these snow fields broad, irregular névés descend to feed the glacier. Owing to this difference in exposure and to the manner in which the opposite sides of the canyon are affected by névé erosion and weathering, the southward facing wall is steep and rugged, with some vegetation in favored places, while the northward-facing slopes are gentle.

At the head of the canyon there is a steep ascent to the summit of the mountain, resembling the higher and more precipitous cliffs at the head of Carbon Glacier. On these cliffs irregular patches of névé snow are to be seen. The snow from these small accumulations, as well as from the less steep slopes about Liberty Cap, descends in avalanches to the head of the canyon. There is a noticeable enlargement of the canyon near its head, but it is not extensive enough to be classed as an amphitheater. The snow falling in avalanches, and also that which accumulates at the head of the canyon during the winter seasons, forms a névé, which is rugged and much broken by ice domes and crevasses. The lower limit of the névé is about 3,000 yards above Division Rock and just above a conspicuous ice fall which crosses the entire width of the glacier. In the névé region there are eight ice domes, which give to its surface an unusual topography. In these elevations the snow rises so as to form well-defined domes, which are, by estimate, from 20 to 50 feet high and from 150 to 200 feet in diameter. As the snow rises on the upstream side of a dome it becomes crevassed. The breaks at first are narrow and curve about the domes, but are highest and broadest in the central part. As the snow advances over the dome the crevasses widen and expose sections of dirt-stained snow beneath a clear, white surface layer 4 to 6 feet deep. On passing the crest the ridges between the crevasses crumble and fill the breaks with a confused mass of shattered blocks. This breaking is a characteristic feature of the downstream sides of the domes in the Willis Glacier, although not more noticeable than in many other examples seen, and gives to the lower sides of the elevation the characteristics of ice falls. In none of the crevasses about the domes could the underlying rock be distinguished.

From the behavior of the névé in passing over the domes just described we should expect the upstream sides of the rocks beneath to be rounded and otherwise glaciated, while their downstream faces would be broken and angular. This, as is well known, corresponds with what is frequently seen in glaciated regions. Division Rock, at the present terminus of Willis Glacier, illustrates the character of the bosses which cause the domes now forming such a characteristic feature of the névé region a mile or two above. This rock is worn and rounded on its upstream side, but presents a broken and angular precipice when seen from below.

The grade of the névé portion of the glacier is gentle; in fact, to one standing by its side, in its lower portion, the surface appears almost level, except that it is higher in the middle than at the margins. Below the ice fall just below the lower limit of the névé the descent to the terminus of the glacier above Division Rock is 840 feet, or about 1 foot in 10.

Above the ice fall mentioned above the névé is not only nearly flat for some 2,000 feet, but practically without crevasses. The absence of crevasses in this region is mainly due, no doubt, to the absence of inequalities on the rocky floor beneath, but is in part explained by the fact that the previous winter's snow, forming the surface, conceals the breaks that may exist in the ice beneath.

Below the névé, at the time of my visit, July 30, there was clear ice forming a belt some 500 feet broad, which extended across the glacier and was limited below by the ice fall. In this portion crevasses begin, which at first are narrow, and grow broader and broader as the ice nears the brink of the precipice over which it falls. The first crevasses to be seen are marginal, and trend upstream at angles of about 40° with the shore, but toward their distal extremities they curve downstream. Soon the crevasses from either side meet in the center of the glacier. These long breaks at first have a gentle downstream curvature in the central portion. Nearer and nearer the brink of the fall the crevasses become wider and wider, and at the same time more pronouncedly curved, until the brink of the precipice is reached. The walls of ice there become broken and fall in blocks and fragments of many shapes and sizes into the intervening crevasses. On the brink of the fall there are numerous blades and tower-like pinnacles, from 10 to 15 feet thick and 20 to 30 feet high, which are inclined forward as they pass over the precipice and fall from time to time with a crash.

The most interesting feature of the ice fall is that the surface of the glacier, as the ice approaches the steep descent, rises sharply, as it does in passing over a dome. The backward or upstream slope of the surface of the glacier in the central part, as measured with a clinometer, while on a level with it on the adjacent shore, is between 2° and 2° 30′. The total rise is about 20 feet. This rise of the surface on nearing the brink of the precipice seems to show that a ridge of hard rock there crosses the bed of the glacier. The nature of the rock in the precipice down which the ice descends is revealed at the end of the fall, but whether there is softer rock above it or not was not observed. When the ice passed over Division Rock a rise of the glacier, like that just described, must have occurred, since the surface of the rock slopes downward toward the east, forming an inclined plane, up which the glacier traveled. The inclination of the surface of the rock was not measured, but by estimate is certainly 3° to 5°. The length of the inclined plane now exposed is about 500 feet.

At the ice fall described above the glacier descends precipitously 400 feet. On the face of the fall the crevasses so conspicuous above are closed and the surface slopes sharply downstream, probably at an angle of 6° or 10°, for about 500 feet, and then becomes more gentle. The ice at the base of the fall is evidently strongly compressed. Blue bands at right angles to the flow of the glacier were observed, but a careful study of them was not made. Surface streams are common on this portion of the glacier during midday in summer, but they soon reach crevasses and disappear.

Descending the glacier, one finds that the ice near the terminus is much wasted by melting, surface débris becomes abundant, and in the central portion above Division Rock deep crevasses appear, which are parallel in a general way with the direction of flow. The crevasses indicate that the ice is split on coming in contact with the rock on which it divides.

On reaching Division Rock we found that in the middle of the V formed by the division of the glacier we could step from the ice onto the rock, although at the sides of the rock the ice had melted away so as to form deep gulfs.

Just where the ice divides, a monument of rough stones about 4 feet high was built. This monument is on the top of a small rounded rock dome somewhat detached from the main mass of Division Rock, and records the position of the terminus of the glacier at the center of the V referred to on July 30, 1896. Division Rock, as stated above, rises steeply from the monument to its summit, and is strewn with stones. On its summit there are about ten small evergreen trees, which show that the actual summit was not occupied by the glacier when it was seen by Willis in 1881.

The distance from the monument described above to the brink of the cliff at the summit of which the ice broke off, as observed by Willis, at the date just stated, is about 500 feet. The glacier has retreated this distance during the past fifteen years, but this is not an accurate measure of total recession, since at the time the position of the terminus was first recorded it formed a vertical precipice of ice on the summit of the cliff referred to. In the summer of 1881 the glacier formed an ice cascade at the lower side of Division Rock, but the fallen ice did not reunite below the cliff.

The appearance of Division Rock in the summer of 1885, when the glacier still ended in a precipice at its lower margin, is shown in Pl. LXXXI.

The glacier, after being parted by Division Rock, sends a sloping and rapidly tapering débris-covered tongue of ice down the deep valleys on each side of it. The southern tongue is the larger of the two, and now ends about abreast of the highest portion of the rock.

Upstream from Division Rock the ice rises somewhat steeply, is heavily covered with débris, and so broken by crevasses that it is with difficulty one can climb it. Just above the first sharp rise, and on the right or northern bank of the glacier, there is another bold rock, similar to Division Rock, but close to the border of the canyon. This eminence was formerly covered by the glacier, but has been abandoned sufficiently long to allow forest trees to grow on its summit and northwestern side.

The retreat of the glacier within recent years has been accompanied by a lowering of its surface, as is plainly recorded by fresh-looking ridges of débris along its border. On the northern side of the glacier, for a mile above the fall of 400 feet, there are three well-defined abandoned lateral moraines. The youngest of these, the one nearest the glacier, rises about 50 feet above the ice, and at the fall is about on a level with the pinnacles on its crest. The second moraine is about 40 feet higher

ICE DOMES, NISQUALLY GLACIER.

than the first, and, unlike the more recent one, is partially clothed with small trees; above this again is a thick moraine, the crest of which is lower than that of the second.

Below the 400-foot fall there is an abandoned lateral moraine, probably of the same age as the youngest moraine described above, which is 100 to 120 feet above the glacier. The vertical interval between this moraine and the present surface of the ice increases gradually downstream. This moraine shows a recent recession of the ice, and probably marks a stage in the lowering of its surface, which was determined in the reach between the 400-foot fall and the present terminus by the obstruction made by Division Rock.

There are many other instructive features to be seen on Willis Glacier which would well repay detailed study, but the time available did not enable me to make accurate notes concerning them. It is to be hoped that future travelers will report especially the position of the ice with reference to the monument built on the east end of Division Rock, and, if possible, secure photographs of the glacier from the point of view from which the photograph reproduced in Pl. LXXXI was obtained.

### INTERGLACIERS.

I can offer but scanty information concerning the interglaciers about Mount Rainier, as the time available for making the reconnaissance on which this report is based was too short to admit of more than a casual examination of a few of them. That the small glaciers originating on the remnants of the mountain side, between the deep canyons carved by the primary glaciers, however, are of importance in the process of topographic development through which Mount Rainier is passing, is evident from even the hasty observations available.

The type of the interglaciers is furnished by the example from which the generic name is derived. Interglacier, named by E. S. Ingraham, occupies a broad, deep depression on the V-shaped remnant of the eastern slope of the mountain, between Winthrop and Emmons glaciers. The apex of this V-shaped mass is The Wedge, described on a previous page. The glacier receives no snow drainage from the main central dome of Mount Rainier, but is supplied by the snow falling on the V-shaped area referred to. The glacier has deepened its basin by excavation until something of an amphitheater has been formed, the rim of which is the divide between Interglacier and Winthrop Glacier on one side and Emmons Glacier on the other. The rocky ridge separating Interglacier from Winthrop Glacier has been broken at one locality, principally by the lateral erosion of the latter ice stream, so as to form what has been named St. Elmo Pass.

Interglacier flows eastward, but melts away before reaching the valley which Emmons Glacier occupies. The stream formed by its melting is tributary to White River.

The interglacier below Little Tahoma has a broad snow field, below which there is a protrusion of true glacial ice in summer. Instead of descending the slope of the mountain so as to excavate a symmetrical amphitheater, however, this glacier divides and is tributary to Emmons Glacier on the north and to Cowlitz Glacier on the south. The névé to the east of Little Tahoma divides into two branches, in much the same manner as the main névé field about the central dome of the mountain gives origin to primary glaciers. The two small glaciers fed by this snow field have excavated depressions for themselves, leaving a sharp, wedge-like mass of rock at the place of division. On the eastern slope of this secondary V-shaped remnant of the mountain side, left in relief by the cutting away of its borders, there is a small snow field, shown on the accompanying map, which probably has true glacial ice beneath it. Under the tentative system of classification used above for the glaciers about Mount Rainier the little glacier here referred to would belong to a tertiary series. This small nameless glacier demands attention from future travelers, as it supplements in an instructive manner the evidence furnished by the larger glaciers of the topographic development of an isolated peak under the influence of glacial erosion.

As may be seen on consulting the map just referred to, each of the branches of the interglacier to the east of Little Tahoma again subdivides, leaving in each instance at the place of bifurcation a wedge-like rock mass facing toward the névé from which the glaciers have their source. A similar division of glaciers into branches is illustrated on a much larger scale by several of the ice streams on the southwest side of Mount Rainier. In these instances, however, it is probable that rock domes similar to Division Rock at the extremity of Willis Glacier cause the ice streams to divide, and not the increase in the area of the mountain side with distance from its apex.

The information referred to above concerning the topography of the V-shaped area east of Little Tahoma and the existence of a small glacier in that region representing a tertiary series is based entirely on the excellent sketch map forming Pl. LXVI, which is here reproduced, with some slight changes, from a manuscript map prepared by Messrs. H. M. Sarvent and G. F. Evans, of Tacoma, Washington, from surveys made by themselves.

Between Cowlitz and Nisqually glaciers there is another secondary glacier or interglacier known as Paradise Glacier. Other small ice bodies on the sides of Mount Rainier, beginning at about timber line and occupying areas more or less completely inclosed by primary glaciers, may be easily recognized, but to attempt a detailed description of them at this time would be premature.

The interglaciers were formerly much more extensive than now, and much of the beauty of the park-like regions in the neighborhood of the upper limit of timber growth is due to the changes they made in the relief of the mountain side, both by rounding and smoothing the rocks over which they flowed and by heaping moraines upon them. Many of

END OF NISQUALLY GLACIER.

the crags and pinnacles which give diversity to the scenery on the steep mountain slopes, like the Guardian Rocks near Spray Park, Gibraltar and the numerous crests near it, and other similar crags in Henry's Hunting Ground, etc., are remnants spared by the glaciers which once enveloped nearly the entire surface of the mountain, but still in their deeper portions flowed in most instances in well-defined channels.

## RECESSION AND SHRINKAGE OF THE GLACIERS.

Every glacier about Mount Rainier that was examined by the writer furnished evidence of a recent recession of its terminus and of a lowering of its surface. In two instances—the Carbon and Willis glaciers—rough measurements of the amount of these changes during the past fifteen years were obtained.

A recent recession of the extremities of the glaciers is shown by barren areas below them on which vegetation is advancing. All of the primary glaciers extend below timber line, and in a climate like that of the region about Mount Rainier any area where there is sufficient soil soon becomes covered with young trees. The barren area about the ends of the glacier may in part be accounted for by the fact that the streams from the glaciers shift their positions from time to time and sweep away the earth in their courses or make deposits of bowlders and stones; but about the extremity of each glacier there are areas not thus affected which are similar in soil conditions to adjacent areas on which large trees are growing. The forests are advancing on the barren areas and gradually taking possession of them. This evidence, even if actual observations of the recession of the extremities of the glaciers were not available, is sufficient to show that the ice streams have for a number of years been growing shorter and shorter.

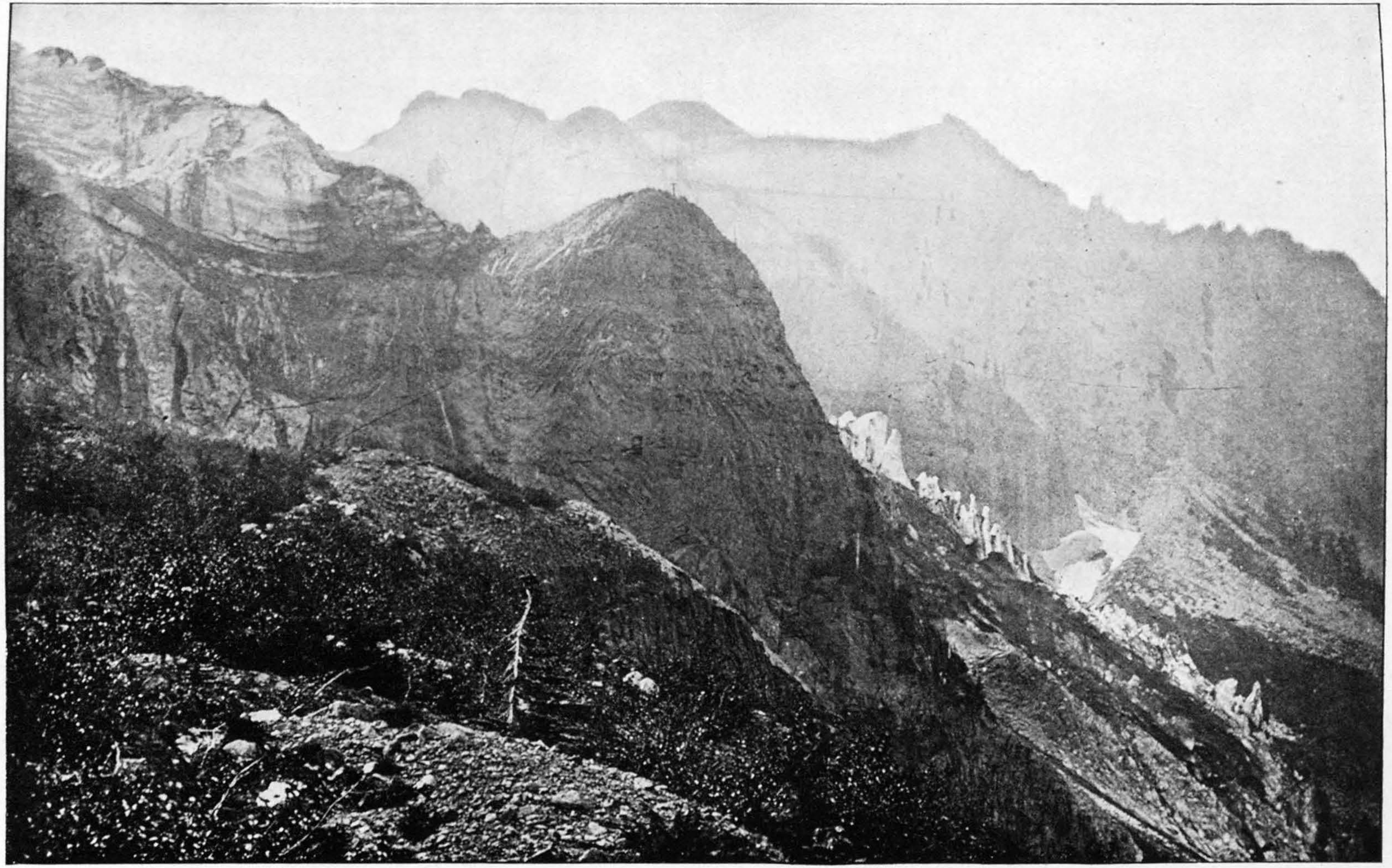
The fact that abandoned lateral moraines bordering the glaciers below their névé fields are in many instances not yet clothed with vegetation, while older moraines of a similar character are forest covered, is evidence that their surfaces have been recently lowered. The elevation of the crests of these recently abandoned lateral moraines above the surface of the glacier that they border is least, or perhaps not at all, noticeable just below the lower borders of the névés, and becomes progressively greater and greater downstream. The lowering of the surfaces of the glaciers is due largely to surface melting. This leads to a concentration of englacial débris at the surface. For this reason the lower portions of the glaciers are almost always heavily covered with stones and earth. As the superficial débris increases in thickness it affords greater and greater protection to the ice beneath, and surface melting is correspondingly retarded. The influence of climate on the melting of the glaciers is thus counteracted to a marked degree. While in the case of glaciers that are free of morainal material or but lightly loaded surface melting would be in excess of subsurface melting, it is apparent that in the case of heavily moraine-covered ice bodies the rate of melting is greatest below the surface portion. The subsurface melting is apparently due to the descent of surface streams into crevasses and moulins and to the heat furnished by streams from the adjacent land which flow through englacial or subglacial tunnels.

The marked recession and shrinkage in progress in the case of the glaciers on Mount Rainier is evidence of a climatic change which is accompanied either by a decrease in the snowfall or by an increase in the mean annual melting, or, what is more probable, of both. Whether the recession and surface lowering of the glaciers is a continuous process or is varied by minor periods of advance and rise of the surface is not known. Judging by what has been observed in the case of the glaciers of the Alps, however, it is to be expected that the glaciers about Mount Rainier will be found to pulsate, as it were—that is, their extremities, although in process of retreat when a period of ten or twenty years is considered, will be found to alternately advance and retreat when smaller periods of time are considered, the algebraic sum of the fluctuations being in the direction of recession. It is for the purpose of obtaining more detailed information in reference to the fluctuations of the glaciers that records of the positions occupied by their extremities, suggested in a preceding page, are desired.

The rapid recession and marked lowering of the surfaces of the glaciers about Mount Rainier is not peculiar to that system of ice streams, but is shared by all of the glaciers of the west coast of North America, with but one known exception. In no instance have the minor fluctuations of the glaciers referred to been observed, but the fact of a general retreat of the extremities of the glaciers, from those in the High Sierra of California to western Alaska, is well known.[1] Not only are the glaciers of the Cordilleran region of North America undergoing a gradual decrease, but all others in the northern hemisphere, with the possible exception of those of Greenland and adjacent regions, are likewise shrinking back into the higher portions of the mountains from which they flow, although in some instances exhibiting local advances.

The recession of the glaciers referred to is evidence that a climatic change has been in progress for at least a score of years, and probably for over a century, which is unfavorable to the existence of perennial ice. Evidently the mean annual snowfall is becoming less, the mean summer temperature greater, the prevalence of clouds and fogs in the summer season decreasing, or a combination of these changes in conditions throughout the entire northern hemisphere is leading, on an average, to an excess of melting over snow accumulation. Observations on the fluctuations of the extremities of the glaciers about Mount Rainier thus have increased interest because these fluctuations form a part of a series of changes which is affecting probably the entire earth. The study of the climatic changes indicated by the fluc-

[1] I. C. Russell, Glaciers of North America, Ginn & Co., Boston, 1897, pp. 146–159.



DIVISION ROCK, WILLIS GLACIER.

tuations of existing glaciers has a still wider bearing, as it may be expected to throw light on the much-discussed problem of the cause of the cold of the Glacial period.

## RÉSUMÉ.

Mount Rainier is a typical example of a lofty volcanic cone built largely of projectiles, but containing also many lava streams. It belongs with the class of volcanic mountains known as composite cones.

At one time the mountain was more lofty than it now is, its reduction in height being due to an explosive eruption which blew away the upper 2,000 feet of the original cone, leaving a great crater in the truncated remnant. After the loss of its summit the mountain was not symmetrical; the rim of its great summit crater was highest on the west, and lowest and probably breached on the eastward side.

At a more recent date two smaller craters were formed by mild explosive eruptions within the great crater and nearly filled it. The building of these secondary craters partially restored the symmetrical outline of the top of the mountain, but gave to it a dome-shaped instead of a conical summit. Whether glaciers were formed on the mountain previous to its truncation, or appeared only after that event, is unknown.

There is no evidence to show that the higher portion of the mountain was exposed to stream erosion previous to the gathering of perennial snows and the formation of glaciers. Broad névé fields were formed about the sides of the mountain, and extended from the summit far down its sides. The lower limit of the snow fields fluctuated with climatic changes, which also caused many variations in the size and extent of the glaciers flowing from them. The descending névé, on spreading as the area to be covered increased with decrease of elevation, became divided and gave origin to primary glaciers which carved deep canyons in the middle slope of the mountain. The backward cutting of the heads of the canyons led to the excavation of more or less well-defined amphitheaters.

Below the extremities of the primary glaciers the rivers formed by the melting ice carved deep canyons, not only in the lower slopes of the mountain, but in the platform or plateau on which it is situated.

Secondary glaciers, originating on the interspaces between the primary glaciers, excavated depressions, the rims of which now stand as ridges and peaks about the middle slopes of the main mountain mass.

The modifications in the general history due to a great extension of the glaciers during the Glacial epoch remain to be studied. All of the glaciers about Mount Rainier are receding and shrinking within the valley walls that confine them. Evidence of a marked climatic change unfavorable to the existence of perennial ice is thus indicated, which is in harmony with similar evidence furnished by nearly all the known glaciers of the northern hemisphere.

## WASHINGTON NATIONAL PARK.

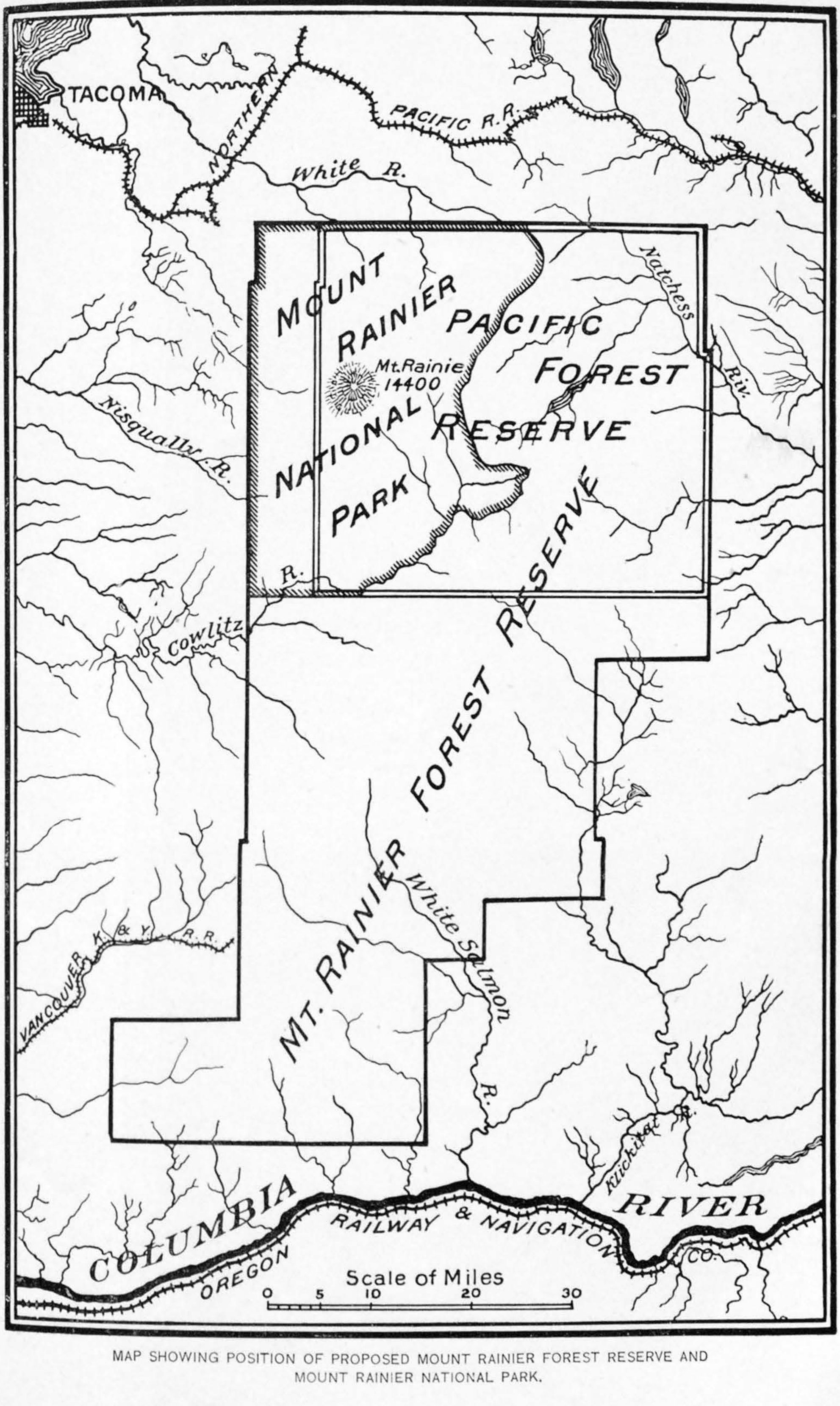
The magnificence of Mount Rainier and the wonderful beauty of the region immediately surrounding it have led to an earnest desire on the part of nearly all who are familiar with the Puget Sound country and are considerate of the future welfare of the State of Washington to have the mountain and its environs reserved in a state of nature as a national park.

Not only is the setting aside of the proposed park thought desirable on what may be considered æsthetic grounds—as the preservation of a tract of primitive forest from the inroads of lumbermen and from fires kindled through carelessness or with a ruthless desire to destroy may be termed—but students of forestry see clearly that to preserve the natural vegetation about Mount Rainier means a check on the floods in the river flowing from the mountain and a lessening of the danger from inundations in the thickly settled valleys leading to Puget Sound and Columbia River.

In 1893, by proclamation of President Cleveland, a rectangular tract of land, indicated on the map forming Pl. LXXXII, about 35 miles square, and including Mount Rainier in its western portion, was made a forest reserve. As may be seen from the map, more than half of this reserve lies east of the crest of the Cascade Mountains. Owing to a mistake due to the imperfection of the maps available, the western boundary of the reserve was made a north-south line passing only about 3½ miles west of the summit of Mount Rainier, and thus leaving much of the western slope of the mountain, including some of its snow fields and glaciers, outside of the protected area. To correct this mistake, and also in order to secure more complete protection for the region about Mount Rainier, a bill was introduced in Congress on June 11, 1896 (H. R. No. 4058), for the purpose of creating a national park, to be known as the "Washington National Park," which should include an area about 25 miles square, of which the summit of Mount Rainier would be about the central point. The boundaries of this proposed park are also shown on Pl. LXXXII.

Owing to the precedence of other business, the bill referred to did not come up for action in 1896, but before the close of the last session of the Fifty-fourth Congress it was passed—too late, however, to receive the President's signature.

The movement toward making the region about Mount Rainier a national park has attracted much attention and received the hearty support of many persons, among whom are those best qualified to judge of the desirability of preventing the region referred to from passing into the ownership of individuals or corporations. In order to indicate to the reader the widespread interest that has been awakened in this project, and to present additional information concerning the region to be included in the proposed park, I can not do better




MAP SHOWING POSITION OF PROPOSED MOUNT RAINIER FOREST RESERVE AND MOUNT RAINIER NATIONAL PARK.

than append here a memorial from committees appointed by several of the scientific societies of the United States, which was presented to the United States Senate by Senator Squire on July 16, 1894.[1] This memorial reads as follows:

*To the Senate and House of Representatives of the United States of America in Congress assembled:*

At a meeting of the Geological Society of America, in Madison, Wis., August 15, 1893, a committee was appointed for the purpose of memorializing the Congress in relation to the establishment of a national park in the State of Washington to include Mount Rainier, often called Mount Tacoma. The committee consists of Dr. David T. Day, Mr. S. F. Emmons, and Mr. Bailey Willis.

At a meeting of the American Association for the Advancement of Science, in Madison, Wis., August 21, 1893, a committee was appointed by that body for the same purpose as above mentioned, consisting of Maj. J. W. Powell, Prof. Joseph Le Conte, Prof. I. C. Russell, Mr. B. E. Fernow, and Dr. C. H. Merriam.

At a meeting of the National Geographic Society, held in Washington, D. C., on October 13, 1893, there was appointed a committee for the purpose above mentioned, consisting of Hon. Gardiner G. Hubbard, Hon. Watson C. Squire, Mr. John W. Thompson, Miss Mary F. Waite, and Miss Eliza R. Scidmore.

At a meeting of the Sierra Club, held in San Francisco December 30, 1893, a committee for the same purpose was appointed, composed of Mr. John Muir, President D. S. Jordon, Mr. R. M. Johnson, Mr. George B. Bayley, Mr. P. B. Van Trump.

At a meeting of the Appalachian Mountain Club, held in Boston April 11, 1894, a similar committee was appointed, consisting of Mr. John Ritchie, jr., Rev. E. C. Smith, Dr. Charles E. Fay.

The committees thus appointed were instructed by the several bodies to which they belong to cooperate in the preparation of a memorial to Congress, setting forth the substantial reasons for the establishment of such park.

Pursuant to their instructions, the committees present the following memorial to the Congress, and pray that such action may be taken by the honorable Senators and Representatives as will secure to the people of the United States the benefits of a national park which shall include the area mentioned above. In support of their prayer they beg to submit the following statement:

By proclamation of the President, in compliance with the statutes provided therefor, a Pacific Forest Reserve has been established in the State of Washington, the western portion of which is nearly coincident with the tract of land to be included in the national park for which your memorialists pray.

The western part of this reserve includes many features of unique interest and wonderful grandeur, which fit it peculiarly to be a national park, forever set aside for the pleasure and instruction of the people. The region is one of such exceptional rainfall and snowfall that the preservation of its forests is of unusual importance as a protection against floods in the lower valleys; but the scenic features, which mark it out for a national park, attract tourists, who set fire to the timber. This destruction goes on notwithstanding it is a forest reserve, and will continue until protection is afforded by adequate supervision of the area, whether as a reserve or park.

### GENERAL DESCRIPTION.

The reserve is traversed through the middle from north to south by the crest of the Cascade Range, which has an elevation varying from 5,300 to 6,800 feet. This is the divide between tributaries of Puget Sound, flowing west, and those of Yakima River, flowing east. Mount Rainier, the isolated volcanic peak, 14,400 feet high, stands 12 miles west of the divide, from which it is separated by a deep valley.

[1] Senate Mis. Doc. No. 247, Fifty-third Congress, second session.

than append here a memorial from committees appointed by several of the scientific societies of the United States, which was presented to the United States Senate by Senator Squire on July 16, 1894.[1] This memorial reads as follows:

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At a meeting of the Appalachian Mountain Club, held in Boston April 11, 1894, a similar committee was appointed, consisting of Mr. John Ritchie, jr., Rev. E. C. Smith, Dr. Charles E. Fay.

The committees thus appointed were instructed by the several bodies to which they belong to cooperate in the preparation of a memorial to Congress, setting forth the substantial reasons for the establishment of such park.

Pursuant to their instructions, the committees present the following memorial to the Congress, and pray that such action may be taken by the honorable Senators and Representatives as will secure to the people of the United States the benefits of a national park which shall include the area mentioned above. In support of their prayer they beg to submit the following statement:

By proclamation of the President, in compliance with the statutes provided therefor, a Pacific Forest Reserve has been established in the State of Washington, the western portion of which is nearly coincident with the tract of land to be included in the national park for which your memorialists pray.

The western part of this reserve includes many features of unique interest and wonderful grandeur, which fit it peculiarly to be a national park, forever set aside for the pleasure and instruction of the people. The region is one of such exceptional rainfall and snowfall that the preservation of its forests is of unusual importance as a protection against floods in the lower valleys; but the scenic features, which mark it out for a national park, attract tourists, who set fire to the timber. This destruction goes on notwithstanding it is a forest reserve, and will continue until protection is afforded by adequate supervision of the area, whether as a reserve or park.

### GENERAL DESCRIPTION.

The reserve is traversed through the middle from north to south by the crest cf the Cascade Range, which has an elevation varying from 5,300 to 6,800 feet. This is the divide between tributaries of Puget Sound, flowing west, and those of Yakima River, flowing east. Mount Rainier, the isolated volcanic peak, 14,400 feet high, stands 12 miles west of the divide, from which it is separated by a deep valley.

[1] Senate Mis. Doc. No. 247, Fifty-third Congress, second session.

The eastern half of the reserve differs from the western in climate, in flora, and in fauna, in geographic and geologic features, and in aspects of scenery. The eastern slope of the Cascade Range within the reserve is a mountainous region, with summits rising to a general elevation of 6,500 to 7,600 feet above the sea. It is forest covered and presents many attractions to the tourist and hunter; but it is not peculiar among the mountain regions of America either for grandeur or interest, and it is not an essential part of the area to be set apart as a national park.

The western slope of the Cascades within the reserve is short and steep as compared with the eastern. Much of it is precipitous, particularly opposite Mount Rainier, where its bare walls would appear most grand were they not in the shadow of that overpowering peak. North and south of Rainier this slope is more gradual and densely wooded.

The western half of the Pacific reserve, that portion which it is proposed shall be made a national park, is characterized by Mount Rainier, whose summit is but 4 miles from the western boundary of the reserve and whose glaciers extend beyond its limits.

Mount Tacoma is not simply a volcanic cone, peculiar for its hugeness. It was formerly a vast volcanic dome, 30 miles in radius to the north, west, and south; but rivers have cut deep canyons, glaciers have carved ample amphitheaters back into the mass, and now many serrate ridges rising from a few hundred to 10,000 feet above the sea converge at that altitude to support the central pyramid, which towers more than 4,000 feet above its base.

This grand mountain is not, like Mount Blanc, merely the dominant peak of a chain of snow mountains; it is the only snow peak in view, Mount St. Helens and Mount Adams being, like it, isolated and many miles distant. Rainier is majestic in its isolation, reaching 6,000 to 8,000 feet above its neighbors. It is superb in its boldness, rising from one canyon 11,000 feet in 7 miles. Not only is it the grandest mountain in this country, it is one of the grand mountains of the world, to be named with St. Elias, Fusiyama, and Ararat, and the most superb summits of the Alps. Eminent scientists of England and Germany, who, as members of the Alpine Club of Switzerland and travelers of wide experience, would naturally be conservative in their judgment, have borne witness to the majesty of the scenery about Rainier.

In 1883 Professor Zittel, a well-known German geologist, and Prof. James Bryce, member of Parliament and author of the American Commonwealth, made a report on the scenery about Mount Rainier. Among other things, they said:

"The scenery of Mount Rainier is of rare and varied beauty. The peak itself is as noble a mountain as we have ever seen in its lines and structure. The glaciers which descend from its snow fields present all the characteristic features of those in the Alps, and though less extensive than the ice streams of the Mount Blanc or Monta Rosa groups are in their crevasses and serracs equally striking and equally worthy of close study. We have seen nothing more beautiful in Switzerland or Tyrol, in Norway or in the Pyrenees, than the Carbon River glaciers and the great Puyallup glaciers; indeed, the ice in the latter is unusually pure, and the crevasses unusually fine. The combination of ice scenery with woodland scenery of the grandest type is to be found nowhere in the Old World, unless it be in the Himalayas, and, so far as we know, nowhere else on the American Continent."

These eminent and experienced observers further say:

"We may perhaps be permitted to express a hope that the suggestion will at no distant date be made to Congress that Mount Rainier should, like the Yosemite Valley and the geyser region of the Upper Yellowstone, be reserved by the Federal Government and treated as a national park."

But Mount Tacoma is single not merely because it is superbly majestic; it is an arctic island in a temperate zone. In a bygone age an arctic climate prevailed over the Northwest, and glaciers covered the Cascade Range. Arctic animals and arctic plants then lived throughout the region. As the climate became milder and glaciers melted, the creatures of the cold climate were limited in their geographic range to the districts of the shrinking glaciers. On the great peak the glaciers linger still. They give to it its greatest beauty. They are themselves magnificent, and with them survives a colony of arctic animals and plants which can not exist in the temperate climate of the less lofty mountains. These arctic forms are as effectually isolated as shipwrecked sailors on an island in mid-ocean. There is no refuge for them beyond their haunts on ice-bound cliffs. But even there the birds and animals are no longer safe from the keen sportsman, and the few survivors must soon be exterminated unless protected by the Government in a national park.

### ECONOMIC RESOURCES.

The area of the Pacific forest reserve includes valuable timber and important water supplies. It is said to contain coal, gold, and silver.

The timber on the western slope differs from that on the eastern in size and density of growth and in kinds of trees. The forests of Puget Sound are world-renowned for the magnitude and beauty of their hemlocks, cedars, and firs. Their timber constitutes one of the most important resources of the State. Nowhere are they more luxuriant than on the foothills west and north of Mount Rainier. But their value as timber is there subordinate to their value as regulators of floods. The Puyallup River, whose lower valley is a rich hop garden, is even now subject to floods during the rapid melting of the snow on Mount Rainier in the limited area above timber line. In the broader area below timber line, but above 3,000 feet in elevation, the depth of snow in the winter of 1893 was 9 to 15 feet. Protected by the dense canopy of the fir and hemlock trees this snow melts slowly and the river is high from March to June. But let the forest be once destroyed by fire or by lumbermen and the snows of each winter, melting in early spring, will annually overwhelm the Puyallup Valley and transform it into a gravelly waste. The same is true of White River and the Nisqually.

The forests of the eastern slope, tributary to the Yakima, are of even greater importance as water preservers. They constitute a great reservoir, holding back the precipitation of the wet season and allowing it to filter down when most needed by crops. In the Yakima Valley water gives to land its value. Storage of flood waters and extensive distribution by canals is necessary. The forests being preserved to control the water, the natural storage basins should be improved and canals built. For these reasons it is most important that no part of the forest reserve should be sacrificed, even though the eastern half is not included in the national park.

The boundaries of the proposed national park have been so drawn as to exclude from its area all lands upon which coal, gold, or other valuable minerals are supposed to occur, and they conform to the purpose that the park shall include all features of peculiar scenic beauty without encroaching on the interests of miners or settlers.

### ACCESSIBILITY.

None save those who can march and camp in the primeval forest can now visit Mount Rainier; but it is the wilderness, not the distance, that makes it difficult of approach. On the west the distance up the Nisqually River from the railroad at Yelm Prairie to the reserve is but 40 miles. Though heavily timbered, the valley of the Nisqually affords an easy route for a railroad. The Cowlitz Valley also offers a line of approach without difficulty by rail, it being about 50 miles from the railroad to the reserve.

On the northwest the railroad at Wilkeson is but 23 miles from the summit of Mount Rainier, and the glaciers can be reached by riding 25 miles through the great forest.

On the north the Cascade branch of the Northern Pacific Railroad crosses the range, only 13 miles in a direct line and 19 miles along the summit from the northern limit of the reserve.

On the east the city of North Yakima is but 62 miles from the summit of Mount Rainier.

The proposed park covers a mountain region which lies across the line of travel from east to west. The railroad winds northward; the travel down the Columbia River turns southward to avoid it. The great current of tourists which flows north and south through Portland, Tacoma, Seattle, Vancouver, and Alaska passes to the west within sight of Mount Rainier, and when the grand old mountain is obscured by clouds the travelers linger to see it, or, passing regretfully on their way, know that they have missed the finest view of their trip.

When a railroad is built up the Nisqually or Cowlitz Valley to the park and connection by stages is assured northward to the Cascade branch of the Northern Pacific Railroad and eastward to Yakima, the flood of travel will be diverted through the park.

## ROUTES WITHIN THE RESERVE.

The point which combines accessibility with surroundings of great beauty, and which is therefore most appropriate as a hotel site, is southeast of Mount Rainier, on one of the spurs of the Tootoosh Mountains, near the Cowlitz Valley. To open this region to travel it would be sufficient to establish the hotel and its connections down the Nisqually or Cowlitz Valley, together with trails to points of interest within the park. From the hotel a principal trail would extend north to the Emmons and White River glaciers, which would thus be easily accessible, and thence the railroad at Wilkeson could readily be reached on horseback over the old Northern Pacific trail. In the future, stage roads, or possibly a railroad, would be extended over the Cowlitz Pass to the eastern slope, North Yakima would be reached via the Tieton or Tannum Valley, and Tannum Lake would become a favorite resort.

But the highway which would challenge the world for its equal in grand scenery would extend from the Cowlitz Pass northward along the crest of the range to the Cascade branch. The distance is 50 miles, 31 in the park and 19 beyond it to the railroad. Within the reserve the summit is open and park-like. On the east is a sea of mountains; on the west is a bold descent of 3,000 feet to the valleys of Cowlitz and White rivers, beyond which Tacoma rises in overpowering grandeur, 8,000 feet above the road and only 12 miles distant.

## CONCLUSION.

A committee of your memorialists has carefully examined the existing maps of the State of Washington with special reference to the position of this reserve, and finds that the boundaries of the reserve are farther east, in relation to Mount Rainier, than was supposed. The western boundary traverses the slope of Mount Rainier at altitudes of 7,000 to 9,000 feet, and the glaciers extend several miles beyond it. In order to include all of the glacial area and the immediately adjacent forest on the west, your memorialists respectfully recommend that the western boundary of the park be drawn one range west of that of the reserve, viz, at the range line between ranges 6 and 7 east of the Willamette meridian. By this change no part of the Wilkeson-Carbonado coal field would be included in the park.

Your memorialists find, as already stated, that it is not necessary to include the eastern slope of the Cascades in the park, and furthermore that it is desirable to leave the Natchez Pass on the north and the Cowlitz Pass on the south open for the construction of railroads. Your memorialists therefore pray that the park be defined by the following boundaries: Beginning at the northwest corner of sec. 19, T. 18 N., R. 7 E. of the Willamette meridian; thence south 24 miles more or less to the south west corner of sec. 18, T. 14 N., R. 7 E.; thence east 27 miles more or less to the summit of the Cascade Range; thence in a northerly direction to a point east of the place of beginning, and thence west 26 miles more or less to the place of beginning.

Your memorialists respectfully represent that—

Railroad lines have been surveyed and after the establishment of a national park would soon be built to its boundaries. The concessions for a hotel, stopping places, and stage routes could be leased and the proceeds devoted to the maintenance of the park. The policing of the park could be performed from the barracks at Vancouver by details of soldiers, who would thus be given useful and healthful employment from May to October.

The establishment of a hotel would afford opportunity for a weather station, which, in view of the controlling influence exerted by Mount Rainier on the moisture-laden winds from the Pacific, would be important in relation to local weather predictions.

Your memorialists further represent that this region of marvelous beauty is even now being seriously marred by careless camping parties. Its valuable forests and rare animals are being injured and will certainly be destroyed unless the forest reserve be policed during the camping seasons. But efficient protection of the undeveloped wilderness is extraordinarily difficult and in this case practically impossible.

Therefore, for the preservation of the property of the United States, for the protection from floods of the people of Washington in the Yakima, Cowlitz, Nisqually, Puyallup, and White River valleys, and for the pleasure and education of the nation, your memorialists pray that the area above described be declared a national park forever.

For the National Geographic Society:

GARDINER G. HUBBARD,
*President.*

For the American Association for the Advancement of Science:

J. W. POWELL.

For the Geological Society of America:

BAILEY WILLIS.

For the Sierra Club:

JOHN MUIR.

For the Appalachian Mountain Club:

JOHN RITCHIE, Jr.

WASHINGTON, D. C., *June 27, 1894.*

# THE ROCKS OF MOUNT RAINIER.

By George Otis Smith.

## INTRODUCTORY.

The earliest geological observations on the structure of Mount Rainier were made in 1870 by S. F. Emmons, of the Geological Exploration of the Fortieth Parallel. The rock specimens collected at this time were studied later by Messrs. Hague and Iddings, of the United States Geological Survey.[1] This petrographical study showed that "Mount Rainier is formed almost wholly of hypersthene-andesite, with different conditions of groundmass and accompanied by hornblende and olivine in places." The only other petrographical study of these volcanics is that of Mr. K. Oebbeke, of Munich,[2] upon a small collection made on Mount Rainier by Professor Zittel in 1883.

On the reconnaissance trips on the northern and eastern slopes of Mount Rainier, during the seasons of 1895 and 1896, the writer had opportunity to make some general observations on the rocks of this mountain, and the petrographical material then collected has since been studied. The observations and collections were of necessity limited, both by the reconnaissance character of the examination and by the mantle of snow and ice which covers so large a part of this volcanic cone.

Two classes of rock are to be discussed as occurring on Mount Rainier: the lavas and pyroclastics which compose the volcanic cone and the granitic rocks forming the platform upon which the volcano was built up.

## VOLCANIC ROCKS.

### GEOLOGIC RELATIONS.

On Crater Peak a dark line of rock appears above the snow, and here the outer slope of the crater rim is found to be covered with blocks of lava. A black, loose-textured andesite is most abundant, and from its occurrence on the edge of this well-defined crater may be regarded as representing the later eruptions of Rainier. Lower down on the slopes of the mountain opportunities for the study of the structure of the volcanic cone are found in the bold rock masses that mark the apexes

[1] Am. Jour. Sci., 3d series, Vol. XXVI, 1883, pp. 222–235.
[2] Neues Jahrbuch für Min., etc., Vol. I, 1885, pp. 222–226.

of the interglacial areas. Examples of these are Little Tahoma, Gibraltar, Cathedral Rock, the Wedge, and the Guardian Rocks. These remnants of the old surface of the cone, together with the cliffs that bound the lower courses of the glaciers, exhibit the structural relations very well.

Even when viewed from a distance these cliffs and peaks are seen to be composed of bedded material. Projecting ledges interrupt the talus slopes and express differences of hardness in the several beds, while variations in color also indicate separate lava flows and agglomeratic deposits. Gibraltar is thus seen to be composed of interbedded lavas and pyroclastics, and on the Wedge a similar alternation is several times repeated, a pink agglomerate being exceptionally striking in appearance.

These lava flows and beds of volcanic ejectamenta thus exposed dip away from the summit at a low angle. The steepest dip observed was in the amphitheater at the head of Carbon Glacier, where in the dividing spur the dip to the northeast is about 30°. Some exceptions in the inclination of the beds were noted on the southeastern slope, where in a few cases the layers are horizontal, or even dip toward the central axis of the cone. In general, however, the volcanics composing Mount Rainier may be said to dip away from the summit at an angle somewhat lower than that of the slopes of the present cone. In the outlying ridges to the north, the Mother Range, Crescent Mountain, and the Sluiskin Mountains, the structure seems to be that of interbedded volcanics approximately horizontal. The extent of the volcanics from the center of eruption has not been determined. Similar lava extends to the south, beyond the Tattoosh Range, and volcanics of similar composition occur to the north, in the Tacoma quadrangle. The latter lavas and tuffs may have originated from smaller and less important cones, now destroyed by erosion.

A radial dike was observed at only one locality, near the base of Little Tahoma. In several cases the lava masses, as seen in cross section, are lens-shaped, and where associated with fragmental beds have unconformable relations. This shows that some of the lava flows took the form of streams, relatively narrow, rather than of broad sheets. Such a feature is in accord with the distribution of rock types. Thus along Ptarmigan Ridge for considerable vertical and horizontal range the rock shows only slight variation. The distribution of rock types will be more fully discussed in a later paragraph.

Of how large a part of the lava flows the crater still remaining was the point of origin is a question to be answered only after more detailed observation has been made. The best section for the study of the succession of flows and ejectamenta is the amphitheater at the head of the Carbon Glacier. The 4,000 feet of rock in this bold wall would afford an excellent opportunity for this were it not that frequent avalanches preclude the possibility of geologic study except at long range.

## MEGASCOPIC CHARACTERS.

The volcanic rocks of Rainier are of varying color and texture. Dense black rocks with abundant phenocrysts of glassy feldspars, rough and coarse lavas of different tints of pink, red, and purple, and compact light-gray rocks are some of the types represented upon the slopes of this volcanic cone. In color, the majority of the rocks may be grouped together as light gray to dark gray. The black and red lavas are less common. In texture, the Rainier lavas are, for the most part, compact. Slaggy and scoriaceous phases are common, but probably represent only a small part of the different flows. Near the Guardian Rocks large masses of ropy lava are found which suggest ejected bombs. Agglomeratic and tuffaceous rocks are of quite common occurrence, although less important than the lavas. Vesicular lavas occur at several localities, and fragments of a light-olive pumice, many as large as a foot in diameter, wholly cover some of the long, gentle slopes southeast of Little Tahoma and in Moraine Park.

Contraction parting or jointing is often observed, being especially characteristic of the basaltic types. The platy parting is the more common, but the columnar or prismatic parting is well exhibited at several localities. The black basaltic lava east of Cowlitz Glacier shows the latter structure in a striking manner. The blocks resemble pigs of iron in size and shape, and where exposed in a vertical cliff these seem to be piled in various positions.

The rocks on the higher slopes of Mount Rainier are in general very fresh in appearance. An exception may be noted in the case of the rocks at the base of Little Tahoma, where some alteration is evident. The bright coloring of the surfaces of the lava blocks and the general appearance of the face of the cliff may indicate fumarole action at this point. There is also some decomposition along the inner edge of the crater rim, near the steam vents. On the lower slopes, some distance below the snow line, the freshness of the rock is not a noticeable feature, and it is seen that here weathering is of the nature of chemical decomposition as well as of mechanical disintegration.

## MICROSCOPIC CHARACTERS.

Microscopically these lavas show more uniformity than is apparent megascopically. Rocks which in color and texture appear quite diverse are found to be mineralogical equivalents. The majority of these rocks are andesites, the hypersthene-andesites predominating, as was shown by Hague and Iddings; but over large areas the andesites are decidedly basaltic, and, indeed, many of the lavas are basalts. The megascopic differences are mostly referable to groundmass characters, the color of the rock being dependent upon the color and proportion of glassy base present. Therefore the degree of crystallization of groundmass constituents is of more importance in determining the megascopic appearance than is the mineralogical composition, and the basaltic lavas are for the most part light gray in color, while the more acid hypersthene-andesites are often black or red.

In petrographic character the lavas range from hypersthene-andesite to basalt. This variation is dependent upon the ferromagnesian silicates, and four rock types are represented—hypersthene-andesite, pyroxene-andesite, augite-andesite, and basalt—any of which may carry small amounts of hornblende. A rigid separation of these rock types, however, is impossible, since insensible gradations connect the most acid with the most basic. In the same flow hypersthene-andesite may occur in one portion, while in close proximity the lava is an augite-andesite.

These lavas have groundmass textures that vary from almost holocrystalline to glassy. The felted or hyalopilitic texture is the most common, and plagioclase is the principal groundmass constituent. The feldspars are lath-shaped, often with castellated terminations. In the more basic phases anhedrons of augite and of olivine appear, and magnetite grains are usually present. Flowage is often beautifully expressed by the arrangement of the slender laths of feldspar.

Among the phenocrysts feldspar is the most prominent. It has the usual twinning characteristic of plagioclase and belongs to the andesine-labradorite series, extinction angles proving basic andesine and acid labradorite to be the most common. Zonal structure is characteristic, being noticeable even without the use of polarized light. Zonal arrangement of glass inclusions testifies to the vicissitudes of crystallization, and often the core of a feldspar phenocryst is seen to have suffered corrosion by the magma and subsequently to have been repaired with a zone of feldspar more acid in composition.

Of the darker phenocrysts, the pyroxenes are more abundant than the olivine or hornblende. Hypersthene and augite occur alone or together, and are readily distinguished by their different crystallographic habits as well as by their optical properties. The hypersthene is usually more perfectly idiomorphic and occurs in long prisms, with the pinacoidal planes best developed, while the augite is in stout prisms, usually twinned. Both are light colored, and the pleochroism of the hypersthene is sometimes quite faint. According to the relative importance of these two pyroxenes, the lavas belong to different types, hypersthene-andesite, pyroxene-andesite, or augite-andesite.

Olivine occurs in certain of the Rainier lavas, in stout prisms somewhat rounded and often with reddened borders. The usual association with apatite and magnetite crystals is noted. The olivine varies much in relative abundance, so as to be considered now an accessory and now an essential constituent, and in the latter case the rock is a basalt.

Hornblende is not abundant in any of the rocks studied, although typical hornblende-andesite has been described among the specimens collected by Professor Zittel. Where it occurs it is in brown crystals, which have usually suffered magmatic alteration. In one case, where this alteration is less marked, the idiomorphic hornblende is found to inclose a crystal of labradorite, and thus must have been one of the latest phenocrysts to crystallize. It also surrounds olivine in this same rock,[1] which is a hypersthene-andesite, the hornblende and olivine being only accessory.

The different textures of these lavas are doubtless expressive primarily of diversity in the physical conditions of consolidation, but also in part of variations in chemical composition. The variations in mineralogical composition are likewise referable to these two factors, but here the latter is the more important. The hypersthene-augite-olivine variation, already referred to, doubtless well expresses the chemical composition of the magma, and deserves to be taken as the chief criterion in the classification of the lavas. As was noted by Hague and Iddings, the hypersthene and olivine play a like rôle, the former occurring when the silica percentage is somewhat higher than in basalt. It is exceptional to find the two in the same specimen, the one being absent whenever the other is present. The following analysis[2] of the typical hypersthene-andesite from Crater Peak shows the lava to be a comparatively acid andesite:

*Analysis of hypersthene-andesite from Crater Peak, Mount Rainier.*

| | Per cent. |
|---|---|
| $SiO_2$ | 61.62 |
| $Al_2O_3$ | 16.86 |
| FeO | 6.61 |
| CaO | 6.57 |
| MgO | 2.17 |
| $Na_2O$ | 3.93 |
| $K_2O$ | 1.66 |
| | 99.42 |

An analysis[3] of one of the light-gray, olivine-bearing rocks on the northern slope of the mountain gives a silica percentage of 54.86, and is doubtless representative of the more basic of the Rainier lavas.

The sporadic occurrence of hornblende in these andesites is principally the result of physical conditions rather than of chemical composition. The magmatic alteration of the phenocrysts of hornblende affords evidence of this variation in consolidation conditions, a diminution of pressure with continuance of slow cooling giving rise to the magmatic

[1] Observed by Iddings: Twelfth Ann. Rept. U. S. Geol. Survey, p. 612.
[2] Hague and Iddings, op. cit,, p. 225.
[3] Oebbeke, op. cit., p. 226.

alteration of the hornblende. That this change took place during the later stages of consolidation is shown by the relative age of the hornblende, noted above, and also by the fact that in one case a phenocryst of augite, where it abuts against the hornblende, has protected the latter from this alteration. The alteration is in part pseudomorphic, the hornblende retaining its characteristic outlines, but often there has been resorption. In one andesite the abundance of these remnants of hornblende and also of augite anhedrons in the groundmass may justify the conclusion that this augite andesite is of derivative origin, of the class described by Washington.[1] It may be noted also that hypersthene shows a tendency to magmatic alteration, although only rarely.

In a basal flow in Moraine Park, the slaggy and compact phases show differences in phenocrysts as well as in groundmass. The glassy rock has hypersthene as the predominant phenocryst, while feldspar is the more important in the compact and more crystalline andesite.

The distribution of the rock types described above is of interest. On the northern slope of the mountain, between Willis and Carbon glaciers, the characteristic lava is a gray andesite, smooth to rough in texture, and showing platy and columnar parting. Hypersthene is not the prevailing pyroxene, and olivine is usually present, often in such abundance as to make the rock a basalt.

In Moraine Park gray andesites also predominate, with both pyroxenes as phenocrysts, but here hypersthene is the more important. On the eastern slope on the Wedge, between Winthrop and Emmons glaciers, the lavas are pyroxene-andesites and vary much in megascopic appearance, although little in microscropic characters. These rocks are quite distinct from any seen to the north. The nunatak in Emmons Glacier is composed of hypersthene-andesite, but on Little Tahoma the lava shows more variety. Both augite-andesite and hypersthene-andesite occur, while at the southern end of this interglacial rock mass, just east of Cowlitz Glacier, the cliffs are composed of the prismatic black basalt. On Crater Peak, and below on Gibraltar, hypersthene-andesite occurs with considerable variation of color and texture. On the spurs west of Nisqually Glacier the andesites contain both pyroxenes, the augite being somewhat the more important.

The distribution of the volcanic rocks, as determined in the study of reconnaissance collections, indicates that the cone has been built up by eruptions of lava and of fragmental material. The successive lava streams were doubtless of considerable thickness, but were limited in lateral extent. The beds of fragmental material are of the nature of flow breccias and of coarse agglomerates on the higher slopes, while tuffs occur at a greater distance from the center of eruption. This composite cone appears to be remarkably free from radial dikes, which may indicate that the volcanic energy was expended chiefly at the crater. The variation in rock types on different sides of the volcanic

[1] Jour. Geol., Vol. IV, 1896, p. 276.

cone may be evidence of changes in position of the center of eruption. The destruction of an earlier crater and the eccentric position of a later would give rise to such a radial distribution of lavas as has been described above.

## GRANITE.

### OCCURRENCE.

The presence of an acid holocrystalline rock on the slopes of Mount Rainier was first reported by Lieutenant Krautz in 1857, from whose accounts Dr. George Gibbs was led to announce the occurrence of granite as a dike in recent lavas.[1] Emmons in 1870 observed a cliff of "beautiful white syenitic granite" rising above the foot of Nisqually Glacier and correctly interpreted the geologic relations. In 1895, on a reconnaissance trip, the writer identified granite among the bowlders composing the lateral moraines of Carbon Glacier, as well as on the surface of the glacier itself, and in the following season bowlders of granite were found to be plentiful in the river bed at the foot of this glacier. This anomaly of granite bowlders coming from a volcanic peak was also noted in the canyon of the Nisqually by Emmons.

In the somewhat more careful study of the Mount Rainier rocks, search was made and the granite was found in place at several points on the northeastern slope. A biotite-hornblende-granite was observed on Carbon River at the mouth of Canada Creek, about 12 miles from the summit of Mount Rainier, and at Chenuis Falls, 2 miles up the river, a finer grained holocrystalline rock occurs, apparently an aplitic phase of the granite. In the lower portion of Carbon Glacier, near its eastern edge, a nunatak of granite can be seen, while the same rock occurs farther to the east, beyond the older of the lateral moraines. Higher on the slopes of Rainier a more marked ridge of granite was traced. A knob rises above the eastern moraine of Carbon Glacier at an altitude of between 7,000 and 8,000 feet, and the more prominent features to the east in Moraine Park also owe their survival to the greater erosion-resisting power of the granite.

### PETROGRAPHIC DESCRIPTION.

These granites have few features worthy of special mention. Hornblende and biotite are the ferromagnesian constituents and vary much in relative importance. The variations from hornblende-granite to biotite-granite occur in the same knob or ridge, and considering all occurrences the two varieties seem to be of equal development. There is also some variation in the amount of quartz present, and in the relative importance of the orthoclase and plagioclase. All of these characters are also found in the granites of the Northern Cascades.

[1] Emmons, Bull. Am. Geog. Soc., 1877, No. 4, p. 45.

## RELATION TO THE VOLCANIC ROCKS.

Along the side of the knob overlooking Carbon Glacier the granite as seen from a distance appears to be intrusive. Blocks of andesite cover the slope, deposited there by the glacier at a time when it possessed greater lateral extent, and the granite talus from above crosses this same slope in a narrow band. The relations prove less deceptive on close examination, and the granite is seen to constitute an older ridge. Farther along this ridge, at the cliffs on the northeastern edge of Moraine Park, the granitic rock is found overlain by the lava. The actual contact of the two rocks is concealed by soil filling the crevice left by disintegration along the contact plane. The granite, however, exhibits no intrusive characters, while the overlying andesite becomes scoriaceous in its lower portion, although compact immediately above. This contact is on the southern side of the granite ridge, the crest of which is approximately east-west. This position of the lava contact considerably below the highest occurrence of the granite indicates that the topographic features of this old granite ridge were even more marked at the time of the eruption of the lavas and the building of the volcanic cone. Above this ridge of granite on the one side tower the cliffs of bedded volcanics which compose the Sluiskin Mountains, and on the other is the andesite ridge bounding the canyon of Winthrop Glacier. Thus Mount Rainier, although a volcanic peak, rests upon an elevated platform of granite which is exposed by erosion at a few points on the slopes of the mountain.

## SUMMARY.

The volcanic rocks of Mount Rainier include both lavas and pyroclastics. The breccias, agglomerates, and tuffs, although of striking appearance, are, perhaps, less important elements in the construction of the composite cone.

The lavas vary much in color and texture, but these megascopic differences are referable rather to the degree of crystallization of the magma than to its chemical character. The variation in the chemical composition of the lavas expresses itself in mineralogical differences, and thus four rock types are distinguished—hypersthene-andesite, pyroxene-andesite, augite-andesite, and basalt. The distribution of these types indicates a radial arrangement of lava streams, and hypersthene-andesite is the more abundant variety of lava.

Granite is exposed on the slopes of Rainier where erosion has cut away the overlying lava, and it is plain that the volcanic cone rests upon an elevated platform of older rock, approximately 8,000 feet above sea level.

# THE AGE OF THE FRANKLIN WHITE LIMESTONE OF SUSSEX COUNTY, NEW JERSEY.

BY

JOHN ELIOT WOLFF AND ALFRED HULSE BROOKS.

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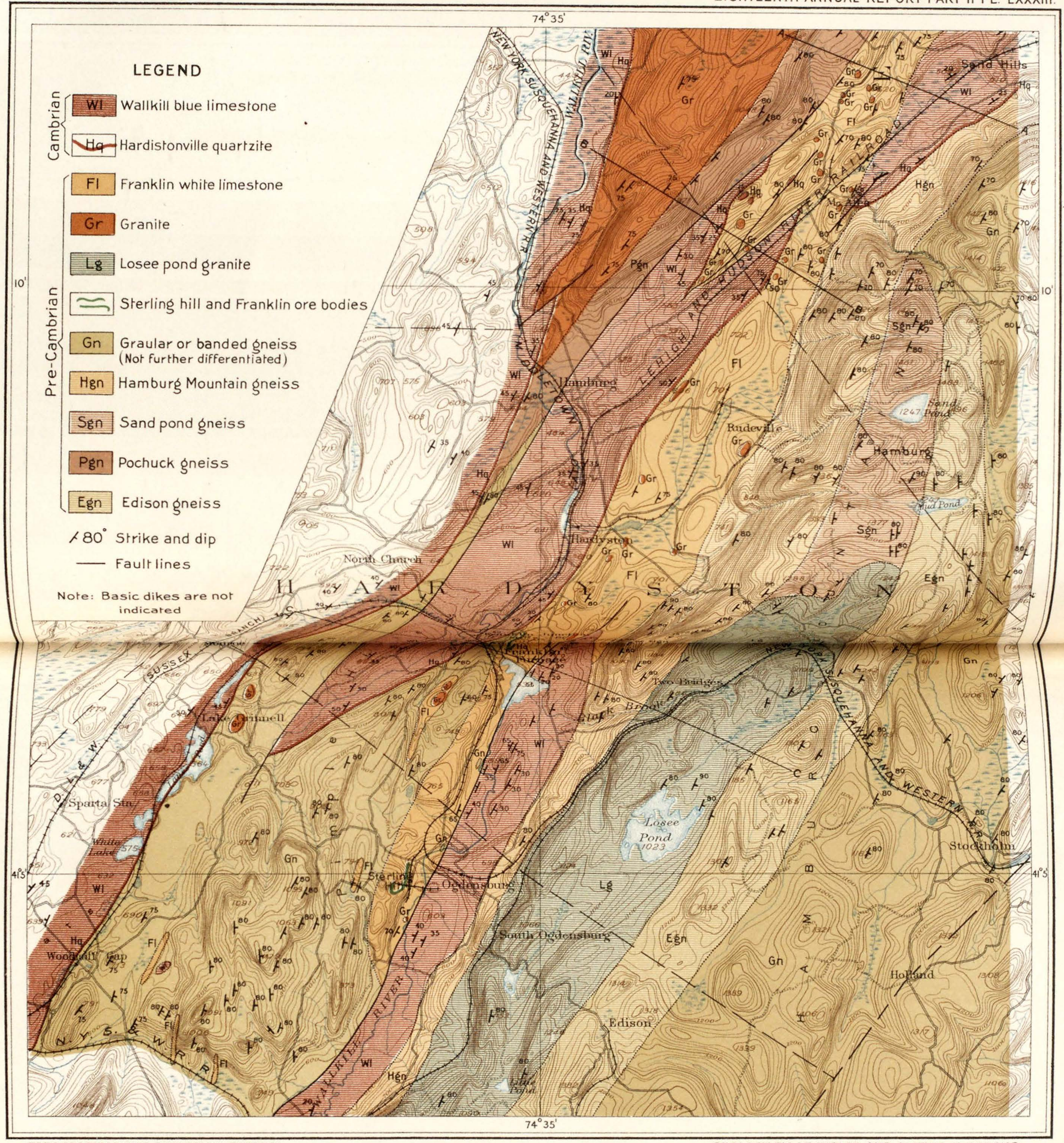
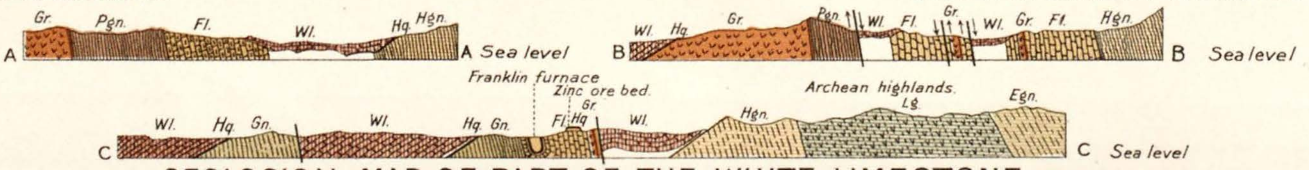



TOPOGRAPHY BY U.S. GEOLOGICAL SURVEY
GEOLOGY BY J.E. WOLFF AND ALFRED H. BROOKS




# GEOLOGICAL MAP OF PART OF THE WHITE LIMESTONE OF SUSSEX COUNTY, NEW JERSEY

Scale $\frac{1}{62,500}$

0 ½ 1 2 3 MILES

Contour Interval 20 feet

*Datum is mean Sea level*

# THE AGE OF THE FRANKLIN WHITE LIMESTONE OF SUSSEX COUNTY, NEW JERSEY.

By J. E. Wolff and A. H. Brooks.

## GENERAL DESCRIPTION OF THE REGION.

The Archean[1] Highlands of New Jersey occupy a belt about 20 miles wide, which runs across the State and continues northeastward into New York and southwestward into Pennsylvania. With the exception of a few longitudinal valleys, in which the younger Paleozoic rocks occur, the whole area is occupied by gneisses, representing in general a few recurring lithological types, with nearly constant northeast strike and southeast dip of the foliation, and with a frequent linear parallel structure, both of the gneisses and of the associated magnetite ore deposits, which is, with rare exceptions, inclined at a moderate angle to the northeast, lying generally in the plane of dip, and is called "pitch." The senior author has given reasons for the belief, in the case of a typical area, that this must be structure of primary crystallization of the rocks as they now exist, and not purely secondary (dynamometamorphic) structure.[2]

At their western margin the Highlands proper are succeeded by the broad Appalachian Valley, in which the Cambrian and Silurian limestones and Hudson River slates extend westward to the Kittatinny Mountain, where the Oneida and Medina quartzites overlie the slates. The topographical transition from the high gneissic plateau to the limestone valleys is generally abrupt, but the boundary line throughout the State is irregular, owing to the outlying gneissic areas, which are cut off in part or completely from the main Highlands by the valleys occupied by the Paleozoic rocks. This feature is well shown on the accompanying sketch map (fig. 77). The Archean area of Scotts and Jenny Jump mountains, as well as that of the Pimple Hills and of Pochuck Mountain, form such outliers.

The white limestone of New Jersey, broadly speaking, occupies two belts, which are shown on the accompanying map. The eastern and

[1] In referring to the gneisses "Archean" is used as a convenient term, but the authors do not wish to commit themselves to anything but the pre-Cambrian age of the series.

[2] J. E. Wolff, Geological structure in the vicinity of Hibernia, New Jersey: Ann. Rept. State Geologist New Jersey for 1893, pp. 359-369.

less important belt lies entirely within the gneissic area. It is marked by isolated areas of white limestone, having a general linear arrangement which is parallel to the general strike of the gneiss. The western belt is much larger and lies near the western margin of the gneisses. A large portion of it is more or less closely associated with the valley rocks. There are three important areas in this belt—the Oxford area, the Jenny Jump Mountain[1] area, which is second in size, and the Franklin area. The last is by far the largest, extending from Sparta,

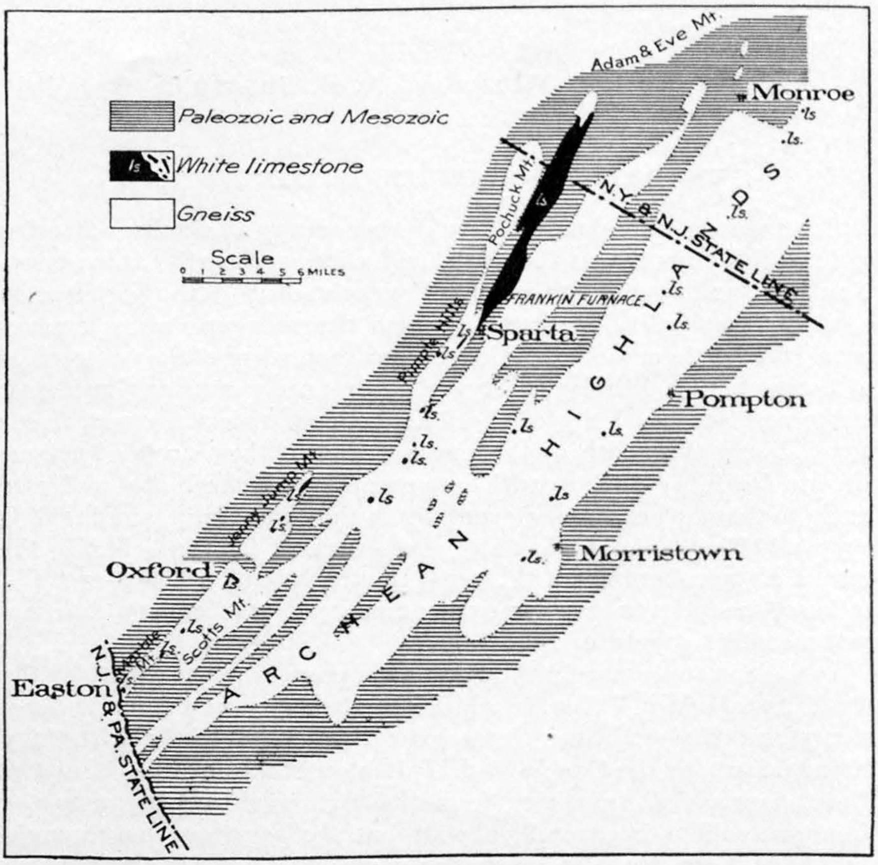



FIG. 77.—Sketch map showing the distribution of the white limestone in New Jersey.

Sussex County, New Jersey, in a northeasterly direction to Mounts Adam and Eve, in Orange County, New York.

The relation of the Franklin white limestone to the Cambrian and gneissic rocks forms the subject of this paper. The detailed field work was done in that portion of the belt lying between Sand Hills on the north and Sparta on the south, which includes the most favorable exposures for the study of the relations. This work was done incidentally to the investigation of the areal geology of the Franklin quadrangle.

[1] Lewis G. Westgate, The geology of the northern part of Jenny Jump Mountain, Warren County: Ann. Rept. State Geologist New Jersey for 1895, pp. 21-61.

## HISTORICAL REVIEW OF OPINIONS.

The literature relating to the geology, minerals, and mines of Franklin Furnace and vicinity, and of the whole belt of white limestone, is voluminous. A partial bibliography has been given by J. F. Kemp,[1] F. L. Nason,[2] and L. G. Westgate.[3]

The ores and other minerals of the white limestone belt attracted the attention of scientific men at the very beginning of the geological sciences in this country, and several of the earliest papers on geological and mineralogical subjects were devoted to the region.

The earliest mention we have been able to find is by William MacClure in 1809,[4] who describes "the large-grained marble on the edge of the Primitive, near Sparta" with the associated ore beds. Although the zinc ores were possibly noticed and prospected as early as 1640, and several tons of the red oxide of zinc were shipped to London about 1770, from which specimens probably found their way to European collections, yet Dr. Archibald Bruce gave the first description and analysis of this mineral in 1810, with an incidental mention of the occurrence.[5]

The first detailed geological description is by Vanuxem and Keating, who visited the spot in 1821.[6] They describe the relative position of the white limestone bed of magnetite and the bed of the zinc ores in relation to the banded gneisses and granite, which they apparently class together as syenite, passing into hornblendic bands or greenstone, and forming the prevalent rock of the "Primitive" in the New Jersey Highlands. They place these rocks in one series, in parallel superposition, and of "almost contemporaneous origin," and describe distinctly the position of the graywacke and overlying blue limestone resting unconformably on the upturned edges of the syenite. They therefore place the white limestone in the Primitive, or Archean.

Pierce,[7] in 1822, alludes to the white limestone of Orange County, New York, and of New Jersey as belonging to the transition, and noticed marine fossils in the blue limestone.

[1] Trans. New York Acad. Sci., Vol. XIII, pp. 76-77; also Kemp, Arthur Hollick, and Heinrich Ries, Annals New York Acad. Sci., Vol. VII, pp. 638-654.

[2] Am. Geologist, Vol. VII, pp. 241-252.

[3] Op. cit., pp. 42-43.

[4] Observations on the geology of the United States, etc.: Trans. Am. Philos. Soc., Vol. VI, 1809, p. 421; also the same, new series, Vol. I, 1818, p. 34.

[5] On red oxide of zinc, etc.: Bruce's American Journal of Mineralogy, Vol. I, 1810, pp. 96-100. This venerable volume, which represents the beginning of American mineralogical and geological research, contains in the first part (1810) analyses and descriptions of two new minerals by Dr. Bruce (brucite and zincite, as they are now known), which are probably the first such descriptions published in this country.

[6] Lardner Vanuxem and William H. Keating, On the geology and mineralogy of Franklin, etc.: Jour. Acad. Nat. Sci. Phila., Vol. II, 1822, pp. 277-288.

[7] James Pierce, Geology, mineralogy, scenery, etc., of the Highlands of New York and New Jersey: Am. Jour. Sci., 1st series, Vol. V, 1822, pp. 26-33.

Thomas Nuttall,[1] in 1822, gives an account of the minerals of Sparta, and says of the white limestone:

The crystalline calcareous rock, which here alternates with granitines of felspar and quartz, or with beds of sienitic granite (near to Dr. Fowler's house, the proprietor of the Franklin works), disappears, and a confluent grauwacke, almost porphyritic, and contemporaneous apparently with the other formations, appears directly overlaid by a bed of leaden, minutely granular, *secondary* limestone, containing organic remains of the usual shells and corallines, and layers of blackish hornstone or petrosilex. This rock, as well as the grauwacke *beneath*, presents disseminated crystals of blue fluate of lime. * * * Thus we have here before us, as at Lake Champlain, the novel and interesting spectacle of an union of every class of rocks, but *passing* decidedly *into each other*, as if almost *contemporaneous!* If they are not contemporaneous, how do they happen to penetrate each other by veins?

C. U. Shepard,[2] in 1832, considered the white limestone Primitive.

Henry D. Rogers,[3] in his first report of 1836, speaking of the white limestone, says:

I regard it as having been originally the blue limestone of the district, invaded at some period subsequent to its formation by these veins of mineral matter [i. e. the zinc ore veins, and veins of syenite and quartz] in a highly heated or molten state, effecting a series of changes similar to those known to have been caused by injection of trap into limestone strata under pretty analogous conditions.

He attributes the peculiar minerals of the white limestone to contact metamorphism.

In his final report,[3] in 1840, he reaffirms the same conclusions at much greater length. He gives a description of the extent and boundaries of the white limestone belt, with a list by Dr. Fowler of the principal minerals associated with it. A detailed description is given of the change from blue limestone to white crystalline limestone, with increasing development of graphite, at the contact with a dike of eruptive granite 4 miles southwest of Sparta, which is selected as representative of many localities showing the same metamorphic change due to the igneous rock. The presence of chondrodite is also ascribed to the same cause.

Charles T. Jackson,[4] in 1852, says that the white limestone is erroneously supposed to be a metamorphic Silurian rock, which it can not be, considering that it is below the Potsdam sandstone, which cuts off the Silurian rocks from any contact with the pyrogenic rocks accompanying the franklinite.

William Kitchell,[5] in 1855, places the white limestone in the Azoic and describes the dikes of granite and syenite cutting it in various places. He also describes two small belts of metamorphic limestone on the western edge of the Azoic, several miles southwest of the Franklin

[1] Observations and geological remarks on the minerals of Paterson and the valley of Sparta in New Jersey: N. Y. Med. and Phys. Jour., 1822; Am. Jour. Sci., 1st series, Vol. V, 1822, pp. 239-248.

[2] A sketch of the mineralogy and geology of the counties of Orange, New York, and Sussex, New Jersey: Am. Jour. Sci., 1st series, Vol. XXI, 1832, pp. 321-334.

[3] Rept. Geol. Survey New Jersey for 1836, p. 132; Geology of New Jersey, Final Report, 1840.

[4] Reports of the New Jersey Zinc Company, 1852. (Geological, by C. T. J.)

[5] Second Ann. Rept. Geol. Survey New Jersey, for the year 1855.

area (one near Pinckneyville, the other between Andover and Tranquility), lying between the blue limestone and the gneiss, which are cut by dikes of granite. As to the age of these no distinct mention is made.[1]

Dr. George H. Cook, in 1861[2] and later, held to the pre-Cambrian (Azoic) age of the white limestones. This view was especially elaborated in the report of 1868,[3] in which full descriptions of the local boundaries of the formations and of their lithological characters, of the zinc deposits, and of the field relations are given. He says (pp. 43–44, 61):

In regard to the crystalline limestones, he [Rogers] was mistaken. They are everywhere conformable to the gneiss and interstratified with it. * * * There is good reason to believe that many beds of coarser granite gneiss have been mistaken for granite dikes and described as such. This has been the case particularly in the crystalline limestone, where the alternating or included beds of gneiss have been thought by many to be dikes. They are, however, conformable to the limestone, and are of the same age, and are metamorphic rocks.

This view of Azoic age, as expressed by Dr. Cook, was held in the publications of the New Jersey Survey,[4] until 1890, when an important paper, by F. L. Nason, appeared in the annual report of that year, readvocating the Cambrian age of the white limestone, and attributing its crystalline condition to the intrusions of igneous rocks. The paper[5] was accompanied by a geological map of the white limestone region and a number of geological cross sections, with descriptions. This paper, by its detail and areal completeness, represented by far the most elaborate and careful work on this question which had been made in the eighty odd years covered in the present review, and must form the basis for future discussion of this problem. The present writers, although obliged to differ from its main conclusion, and to interpret differently some of the most important observations, yet heartily acknowledge their obligation to the work of Nason.

After a preliminary account of the topography and of the areal

[1] In the historical review cited on page 433, Nason incorrectly ascribes the description of the gradual change from blue to white limestone at the latter locality to the Franklin belt. The statement that Rogers in his first report (1836) "says nothing in this report of his idea of the contemporaneous origin of the white and blue limestones" should also be corrected.

[2] Note on the probable age of the white limestone at Sussex and Franklin zinc mines, New Jersey: Am. Jour. Sci., 2d series, Vol. XXXII, 1861, pp. 208–209.

[3] Geology of New Jersey, 1868; George H. Cook, State geologist; John C. Smock, assistant geologist.

[4] N. L. Britton, in 1885 and 1887, placed the white limestone conformable in his middle or gneissic group of the New Jersey Archean. Ann. Rept. Geol. Survey New Jersey for 1885, pp. 36–55. Same for 1886.

[5] The post-Archean age of the white limestones of Sussex County, New Jersey: Ann. Rept. State Geologist New Jersey, for 1890, pp. 25–50. Other subsequent articles by Mr. Nason, elaborating or reasserting the views of the original paper, are: The post-Archean age of the white limestones of Sussex County, New Jersey: Am. Geologist, Vol. VII, pp. 241–252 (historical review already cited). [Same title; a reply to a review]: ibid., Vol. VIII, 1891, pp. 166–171. The franklinite deposits of Mine Hill, Sussex County, New Jersey: Trans. Am. Inst. Min. Eng., 1894, Vol. XXIV, pp. 121–130. The chemical composition of some of the white limestones of Sussex County, New Jersey: Am. Geologist, Vol. XIII, 1894, pp. 154–164. Summary of facts proving the Cambrian age of the white limestones of Sussex County, New Jersey: ibid., Vol. XIV, 1894, pp. 161–169.

distribution of the formations, he describes the rocks, and first gives the distribution of the granite, describing its wide occurrence throughout the area and giving the proofs of its eruptive origin. These proofs are: that it cuts obliquely across the line of strike of the white limestone at Franklin Furnace and has been proved by the diamond drill to be interlaminated irregularly with the limestone, and that it produces distinct contact phenomena, such as the development of fluorite, chondrodite, tourmaline, etc.; also, that it is firmly knit to the Cambrian sandstone at a contact near Hamburg, and at the Pochuck mine road has fragments of sandstone and other rocks embedded in it; that in this sandstone and the blue limestone graphite is developed near the contact with granite, and mica and feldspar in the sandstone; also that the granite changes the blue limestone gradually into a white limestone near the contact with a dike. The eruptive nature of the granite with reference to the white limestone, and we may add to the associated gneisses, is unquestionable, and Nason has established this beyond doubt. The facts adduced to prove a similar relation to the Cambrian quartzite and limestone seem to us to be capable of a different interpretation.

Scapolite rock, or "geflecter gabbro," is also described as a frequent dike rock in the white limestone. The discovery of fragments of trilobites in the sandstone underlying the blue limestone, by Dr. Charles E. Beecher, establishing the Olenellus age of the sandstone, is also announced.

The detailed descriptions of the geological sections, which will be referred to later in connection with our own observations, establish, according to Nason, the transition of the white and blue limestone by gradual change when in proximity to a granite dike, the line of contact being a brecciated zone in many sections, and also the interbedding of the two limestones with the sandstone or quartzite. As summarized in a later paper,[1] his conclusions are as follows:

1. The white limestones are continuous with the blue limestones (now accepted as of Cambrian age), and every degree of transition may be found between them.

2. Both have the same dip and strike.

3. Both are conformable with a quartzite also containing Cambrian fossils.

4. Both are unconformable with the gneiss upon which they rest.

5. Both have in sum total the same chemical composition and are magnesian.

6. The altered crystalline condition of the white limestone is due to the intrusion of igneous masses and to regional metamorphism [the latter agency is not mentioned in the original paper], while the blue limestone never contains such igneous injections.

7. The presence of certain minerals, especially chondrodite, is not indicative of geological age.

The conclusion under (5) is considered in detail in a previous paper,[2]

[1] Summary of facts proving the Cambrian age of the white limestone of Sussex County, New Jersey: Am. Geologist, Vol. XIV, 1894, pp. 161-169.

[2] The chemical composition of some of the white limestones of Sussex County, New Jersey: Am. Geologist, Vol. XIII, 1894, pp. 154-164.

in which by a series of chemical analyses a general magnesian character of the white limestone is indicated.

The northern termination of the white limestone at Mounts Adam and Eve, in Orange County, New York, was described by H. D. Rogers;[1] it was placed in the Primitive by W. Horton[2] in 1839, and considered metamorphic [Cambrian] by Mather in 1843, who adopted Rogers's views.[3]

J. F. Kemp and Arthur Hollick[4] in 1894 described the contact phenomena of the granite of Mounts Adam and Eve with the white limestone, and were inclined to adopt the view of its Cambrian age.

On the other hand, J. D. Dana,[5] in reviewing Nason's original article, in 1891, dissented from his conclusion, holding it not proven. He doubted whether the granite were not foliated gneiss, and in any case doubted the possibility of dikes of granite causing the metamorphism, concluding that to allow the granite to assume this uniform, coarse, foliated structure the limestone must have been hot enough to cause its own metamorphism. He doubts the apparent transitions, and suggests as a possible cause the superposition of the later limestone on the earlier, and subsequent changes; he also asserts that like strike and dip do not necessarily imply contemporaneity.

In a review of the paper by Kemp and Hollick, H. S. Williams[6] doubts the sufficiency of the evidence for the identity of the two limestones in the absence of fossils, and gives examples of occurrences of two apparently conformable and continuous limestones elsewhere which are of distinct geological age.

Westgate, in 1894[7] and 1895,[8] in describing the area of white limestones at the northeast end of Jenny Jump Mountain, in Warren County, believes them to be distinct from and older than the blue Cambrian limestone, for the following reasons:

1. They differ lithologically from the blue limestone in being thoroughly crystalline and in containing large amounts of accessory metamorphic minerals.

2. They are intimately associated with and apparently interbedded with the older gneisses, and gneisses occur also interbedded in the limestones.

3. They show no intimate association in areal distribution with the blue limestone, nor any tendency to grade into it.

4. The metamorphic changes to which the blue limestones have been subjected are general in their nature, and not due to the action of the eruptives by which they are cut, so that no sufficient agent is at hand to account for the supposed change from blue into white limestone.

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[1] Geology of New Jersey, Final Report, 1840, p. 67.

[2] Report on the geology of Orange County: Third Geol. Rept. New York, 1839, pp. 135-275.

[3] Report on the First Geological District, 1843, p. 465 et seq.

[4] The granite at Mounts Adam and Eve, Warwick, Orange County, New York, and its contact phenomena: Annals New York Acad. Sci., Vol. VII, pp. 638-654.

[5] Review of annual report of State geologist for 1890: Am. Jour. Sci., 3d series, Vol. XLII, 1891, pp. 70-72.

[6] The age of the white limestones near Warwick, Orange County, New York: Am. Jour. Sci., 3d series, Vol. XLVII, 1894, pp. 401-402.

[7] The age of the crystalline limestones of Warren County, New Jersey: Am. Geologist, Vol. XIV, 1894, pp. 369-379.

[8] The geology of the northern part of Jenny Jump Mountain, Warren County: Ann. Rept. State Geologist New Jersey for 1895, pp. 21-61.

The white limestones are believed to be older than the blue Cambrian limestone, because (1) they occur in intimate association with the gneisses, which are of admitted pre-Cambrian age, and because (2) they have been subjected to general metamorphic forces, resulting in great changes, of which the neighboring blue limestone shows no trace.

In 1893 A. F. Foerste[1] showed that the thin bed of Olenellus Cambrian quartzite which lies at the base of the blue limestone was much more continuous than indicated on previous maps. He added to the localities at which fossils had been found.

J. F. Kemp,[2] in 1893, in a paper on the ore deposits of the white limestones, attributes the original deposit of the ores to solutions set in circulation by the intrusive rocks. After deposition walls and ore were folded by mountain-making pressure and metamorphosed.

## DETAILED GEOLOGY OF THE AREA.

### TOPOGRAPHY.

The general topography of the region has already been noted, but it seems well to add a few notes in regard to the white limestone belt proper. The belt is flanked on the eastern side by the Archean Highlands and on the western side by the Archean outliers. The two limestones under consideration occupy the valley of Wallkill River and Black Creek. The blue limestone usually forms the floor of the valley and is limited to the lower portion of the hill slopes. The white limestone frequently forms rather sharp ridges and hills with frequent bluffs in the valley lowland, and sometimes occurs high up on the flanks of the gneissic highlands.

The drainage of the belt is northward through the Wallkill River and its tributary, Black Creek, to the Hudson. The original topography of the area is much obscured by the glacial drift, which is in places quite heavy.

### LITHOLOGICAL DESCRIPTIONS AND STRUCTURE.

#### GNEISSES.

*The gneisses of the Highlands, immediately east of the white limestone area.*—The larger part of this area is covered by the usual complex of gneisses of varying mineralogical character and massive or schistose structure, which are common to the Archean Highlands, and in which it is difficult or impossible to form any continuous subdivisions. Certain belts, however, of somewhat uniform character can be identified, and are indicated on the map. The easternmost belt, which lies

[1] New fossil localities in the early Paleozoics of Pennsylvania, New Jersey, and Vermont, with remarks on the close similarity of the lithologic features of these Paleozoics: Am. Jour. Sci., 3d series, Vol. XLVI, 1893, pp. 435-444.

[2] The ore deposits at Franklin Furnace and Ogdensburg, New Jersey: Trans. New York Acad. Sci., Vol. XIII, pp. 76-98.

between Stockholm and the western crest of the Highlands, is a series of ill-defined, often granular gneisses, intimately mixed with long parallel bands of pegmatite. The intrusive character of the latter is shown in favorable localities by tongues from the main band which cut across the foliation of the gneisses as apophyses.

This is bounded on the west by a belt of gneisses characterized by their richness in disseminated magnetite, which, from the extensive mining operations at the old Ogden mines, now called Edison, we have called the Edison gneiss. Toward the northern boundary the scarcity of outcrops has made it impossible to differentiate this band in the general complex.

Lying between the higher crest occupied by the Edison gneiss and the extreme western crest is a very characteristic and well-defined band of a greenish-white, gneissoid, binary granite, which from its typical exposures around Losee Pond is named the Losee Pond granite. In several places in this belt bands of hornblendic gneiss occur, which are cut by the former rock as an intrusive, and it is therefore considered an eruptive mass. At a point about east of Hardistonville this band gradually narrows and is cut off by a belt of coarse, foliated, syenitic gneiss (the Sand Pond gneiss), which dies out to the north or can not be differentiated.

The extreme west crest and slopes of the Highlands are formed by a complex of gneisses (Hamburg Mountain gneiss), of which the most prominent member is a coarse, banded, hornblendic gneiss, resembling phases of the Edison gneiss, with which possibly it might be correlated. In this belt are frequent granitic phases, which are probably intrusive bands.

All these belts have the usual northeast strike and steep southwest dip of the foliation. As will be seen by the map, the easternmost gneisses, including the Edison belt, curve westward just north of the termination of the Losee Pond granite, so as to nearly join the Hamburg Mountain gneiss, which fact, together with their lithological identity, suggests a possible anticlinal arch.

Lying west of the white limestone is the gneissic outlier of the Pimple Hills, connected by a narrow line, nearly buried by the blue limestone and glacial drift, with the other area of Pochuck Mountain. The Pimple Hills outlier is a complex of various granular and foliated gneisses, with masses of eruptive pegmatite and granite, and belts of amphibolitic gneiss, which can not be delineated on a small-scale map, and have not been divided into areal belts. This complex bounds the white limestone on the west for a distance of 4 miles—from its termination south of Ogdensburg to a point north of Mine Hill in Franklin. The tongue of blue limestone west of Franklin separates this from an extreme western belt, the outcrops of which run north to the railroad tracks half a mile south of North Church.

From this point to the southernmost outcrops of the Pochuck area west of Hamburg (a distance of 2 miles) only two small exposures of the gneiss are found.

Three small exposures of gneiss occur along the eastern line of contact of the white and blue limestone between Franklin Furnace and Ogdensburg.

*Pochuck gneiss.*—The gneissic outlier of Pochuck Mountain is represented by two bands of rock. The eastern band, which adjoins the valley and forms the eastern slope of Pochuck Mountain, we have called the Pochuck gneiss. This gneiss is about half a mile in width, and has been traced from near Hamburg to the northern edge of the area mapped. The prevalent rock type is a finely foliated gneiss, rich in biotite and hornblende, and often grading into a mica-schist. With this mica-hornblende gneiss are often associated bands of amphibolite-gneiss. The latter rock consists essentially of plagioclase, green hornblende, and quartz, with some biotite, and is probably a squeezed dioritic rock.[1] Dr. N. L. Britton has described a cyanite-schist from the eastern base of Pochuck Mountain, which we did not find.

West of the Pochuck gneiss occurs a mass of granite, which pinches out near Hamburg and increases in width to the north. It is a greenish-white granite, which is often gneissoid, and is composed essentially of quartz, orthoclase, microcline, and an acid triclinic feldspar, with a little biotite. Its intrusive relation to the gneiss is well shown at numerous contacts along the crest of Pochuck Mountain, where it is often seen cutting the foliation.

The Pochuck Mountain gneissic series has a strike of about N. 20° E., and dips 75° to 80° E. Attention should here be called to a small band of this gneiss which occurs near the road about a mile southwest of McAfee, near the contact of the white and blue limestone. This is the only occurrence we found of this type of rock within the limestone area.

## FRANKLIN WHITE LIMESTONE.

Typically this is a coarsely crystalline rock, of which the crystalline grains of calcite (or dolomite) often measure an inch in diameter. It frequently contains graphite in disseminated scales, which are often sheared into thin films, flakes of a yellowish mica, and lenticular grains of greenish silicates (diopside, etc.). When these accessory minerals are wanting the foliation is often indicated by the lenticular shape of the calcite grains. The foliation is produced by the arrangement of these minerals, and is the only structure by which strike and dip can be measured. We have seen nowhere in the white limestone what could be regarded as planes of sedimentation. Our use of "strike" and "dip," applied to both white limestone and the gneisses, refers to the foliation only. It is not the purpose of this paper to discuss all of the many minerals which are found in the limestone, about which there is a voluminous literature.

[1] Ann. Rept. State Geologist New Jersey for 1886, p. 104.

The crystalline limestone is typically white in color, but often has a slight bluish tinge. This blue phase has, according to our observation, a coarsely crystalline character, which serves to distinguish it from the Cambrian blue limestone, except in a few localities where they have been modified by shearing. We can not agree with Nason's statement[1] that the degree of crystallization depends to any great extent upon the proximity of an igneous rock. Our observations rather point to the conclusion that the white limestone was thoroughly crystalline before the intrusion of the granite.

In the area under consideration the white limestone forms a long, more or less irregular belt, which is less than a mile in width. The belt enters the area of the map from the north about one-half mile west of the Sand Hills, and southward to about McAfee varies little in width, being bounded on the east by the Cambrian series and on the west by the Pochuck gneiss. Near McAfee a diverging arm of the white limestone forms a wedge-shaped area, which is bounded on both sides by the Cambrian rocks. The main belt continues southward from McAfee, bounded on the east by the Hamburg Mountain gneiss and on the west by the overlying quartzite. The quartzite continues to form the western boundary to a point near Mine Hill, where the Pimple Hills complex lies adjacent to the white limestone and continues to form its western boundary until it wedges out 1 mile south of Sterling Hill. From Franklin Furnace southward the belt is quite narrow, because a portion is covered by the Cambrian limestone, which here forms its western boundary. Several small areas of white limestone are represented on the map, which lie entirely within the Pimple Hills gneissic area.

In strike and dip the white limestone shows, as a rule, little variation, and almost always shows a parallelism to the foliation of the gneisses. The strike is N. 20° to 30° E., and the dip is generally easterly, varying from 70° to 80°. The question of the relation of the white limestone to the other rocks will receive further consideration later.

As no planes of sedimentation were found in the white limestone it is evident that its structure and thickness can not be determined. That there are folds in this rock is evident from the well-known synclinal form of the ore bed at Franklin Furnace.

## GRANITE.

The granite which occurs in the white limestone is found as small knobs or as sheets approximately parallel to the foliation, but sometimes cutting across, as can be seen at the Franklin quarry and at other localities. The intrusive nature of the granite is established not only by this, but also by the contact phenomena, similar to those described by Kemp and Hollick at Mounts Adam and Eve in the paper cited.

An examination of the map will show that there is a certain amount of linear arrangement of the granite intrusions, but the individual

[1] Ann. Rept. State Geologist New Jersey for 1890, p. 37.

masses are often widely separated. Intrusive masses of granite are found also in the gneisses, and these are often of considerable size. In line with the southern extension of the granite area of Pochuck Mountain, which has already been described, considerable masses of granite are found in the Pimple Hills complex. The Losee Pond granite has also been referred to. This Losee Pond granite does not differ very much lithologically from the other granites which are represented by the one color on the map, but it is so isolated that it has been thought best to give it a separate name and pattern.

The granite of the white limestone belt and adjacent gneisses is generally coarse, often pegmatitic, with more or less distinct foliation parallel to the bounding planes. In the large areas this structure is parallel to the foliation of the associated gneisses. In the smaller stock-like masses the foliation is often obscure.

The rock is of a whitish or greenish color, the latter due to the feldspar. It is composed principally of quartz and feldspar and small amounts of biotite and hornblende, with various accessory minerals. The feldspar varies in different localities; in some cases microcline with microperthitic albite predominates, in others orthoclase with an acid plagioclase (oligoclase). The structure of these minerals is distinctly gneissic, analogous to that of the minerals of the gneisses, and unlike the normal structure of ordinary intrusive granite. All the granites of the region have a general lithological similarity and probably belong to the same age of intrusion.

Nason and others have described the basic bands in the white limestone and the later dikes. As these rocks are not germane to the subject of the paper, a further reference to them is not deemed necessary.

### HARDISTONVILLE QUARTZITE.

The oldest fossiliferous rock of the region is the Hardistonville quartzite, the Olenellus age of which has been established by Beecher and Nason. This is the basal bed of the blue limestone, with a thickness varying from 30 feet to a foot or less, and grading upward into the overlying Wallkill limestone. Where the contact of the blue limestone and the older rocks is exposed, we have always found this quartzite present, except where there has been faulting.

An examination of the map will show the quartzite to occur in long narrow bands along the normal contacts of the blue limestone. It is found along the western flank of Pochuck Mountain and the Pimple Hills, along the eastern contact of the Hamburg blue limestone belt, and again along the eastern contact of the northwest fork of the Hamburg belt. It occurs along the contact of the Sand Hills limestone area and the gneisses. At the western boundary of the Sand Hills area and the eastern boundary of the Ogdensburg blue limestone belt no outcrops of this rock were found, although its probable existence was indicated by quartzite débris. We have not thought it best to indicate it on the map in these doubtful localities.

The quartzite also occurs in several small isolated patches lying on top of the white limestone and without the usual succeeding blue limestone. Three localities where these relations exist have been observed. The first is one-quarter of a mile west of McAfee, where the relations are somewhat obscured by faulting; another is about a mile south of Hardistonville, and the third is at the new quarry at Franklin Furnace. All these occurrences will be described in greater detail below.

When fresh, the quartzite is usually bluish gray, weathering near the surface to a yellow or brown, often porous, limonitic rock. Frequently it contains considerable pyrite, and varies in coarseness from a fine conglomerate to a quartzite. A shaly phase is often present in the upper part of the bed, where it merges into the limestone. It is composed of large and small grains of clastic quartz, which are usually cemented by calcite, mixed with a fine aggregate of quartz. In many localities it is filled with fragments of clastic feldspar (microcline, etc.), and plates of light-colored mica, which are distinctly clastic, as seen in thin section. Where it occurs near white limestone it frequently contains graphite, as stated by Nason. This graphite occurs in round plates, often bent, and in thin section they show no evidence of having been formed in situ. While the evidence of their clastic origin is not so distinct as in the case of the mica, yet we do not doubt that it is an allothigenous constituent of the rock which was derived from the adjacent white limestone or micaceous gneiss.

At the new limestone quarry in Franklin southeast of the Buckwheat mine, at the brook section in Hardistonville, and on the hill, 720 feet high, 1 mile south of Hardistonville, we found distinct pebbles of typical white limestone embedded in the quartzite. These should not be confounded with the concretionary masses of calcite mixed with quartz, and often fluorite, which are sometimes found in the quartzite as well as in the overlying blue limestone.

The quartzite in many places becomes an arkose, and is then composed of quartz, feldspar, and mica, with fragments of granite. These minerals have a plainly clastic character when examined under the microscope, and are identical with the same minerals found in the adjacent crystalline rocks. For instance, a pale-yellowish mica is a common constituent of the white limestone, and clastic fragments of the same mica are found in the quartzite. The microcline found in the clastic grains in the quartzite is also a common constituent of the granite. There can be no doubt that these minerals in the quartzite are clastic, and we emphasize this fact because Nason[1] implies that the development of the graphite, mica, and feldspar crystals in the sandstone is due to contact metamorphism produced by the granite.

## WALLKILL LIMESTONE.

This is normally a blue limestone, which requires little description. It is a fine-grained homogeneous rock, becoming arenaceous or shaly as

[1] Ann. Rept. State Geologist New Jersey for 1890, p. 31.

it merges into the quartzite, and occasionally having a cleavage which may easily be mistaken for bedding. Chert is sometimes present in the limestone in small nodules. Small, rounded masses of white calcite, and sometimes fluorite, which are found in the blue limestone, are apparently of concretionary origin, but in some cases may be pebbles of white limestone. Like the white limestone, the blue limestone is somewhat dolomitic, but we do not attach any weight to this fact as a proof of their identity.

Along certain lines of contact between the white and blue limestone apparent transitions are found between the two rocks, which we attribute to shearing and brecciation accompanying faulting. The details of these localities will be considered below.

On the accompanying map four areas of blue limestone are represented. The most northeasterly of these areas is called the Sand Hills belt. Another, which lies west of the central portion of the main belt of white limestone, we have termed the Hamburg belt. The third is the Ogdensburg belt, and occupies the Wallkill Valley south of Franklin Furnace; while the fourth is part of the valley series, and flanks Pimple Hills and Pochuck Mountain on the west. The geological sections and map should be consulted in reading these descriptions.

### SAND HILLS BELT.

This area of Cambrian rocks lies in the northeast corner of the Franklin quadrangle. It has a general synclinal structure, with some minor plications in the central portion of the basin (see section A of Pl. LXXXIII). On the west it is bounded by white limestone. Two miles north of Sand Hills, beyond the limit of the quadrangle, the normal succession is observed on the west side of the syncline. Here the white limestone flanks the gneisses, both dipping at a high angle to the east and succeeded by the less steeply dipping quartzite and limestone. Within the quadrangle the quartzite does not outcrop on the west side of the syncline. The relation of the blue and white limestones is well shown in a small side valley about a mile north of McAfee, at which point Nason's section No. XI is drawn. We can not agree with Nason's statement of the facts,[1] for our field observations at this locality are as follows: Our section No. 1 passes through this locality. The white limestone forms the hill to the south of the brook and is well exposed along the extension of the strike in the bed of the stream at the falls. The foliation of the graphite plates gives a strike of about N. 20° E. and an eastward dip of about 80°. Following along the strike up the hill to the north of the brook, we find outcrops of white limestone extending about 60 feet up the slope, all showing a nearly vertical dip. Continuing up the hill, which necessitates crossing the strike diagonally, we traverse 40 feet or more of talus with no exposures, and then find outcrops of typical unaltered blue limestone. From this point the blue limestone outcrops continually to the top of the hill, the upper 20 feet

[1] Ann. Rept. State Geologist New Jersey for 1890, pp. 47-48.

forming a cliff. The blue limestone strikes about N. 50° E. and dips at 20° to 25° SE. From this description it will be seen that there is a marked unconformity in the strike and dip of the two limestones, and that the blue limestone lies topographically above the white. We observed none of the transitional phases described by Nason, and believe that the facts bear the strongest kind of evidence of the unconformity of the two series. It should be observed that while no quartzite outcrops in the section, it probably is present and covered by the talus. South of this brook no outcrops of the Cambrian rocks were found except in the hill with an elevation of 629 feet.

In the southeastern arm of the syncline the quartzite rests directly on gneiss and is overlain by the Wallkill blue limestone, both dipping at a low angle to the westward. The marked discordance of the strike

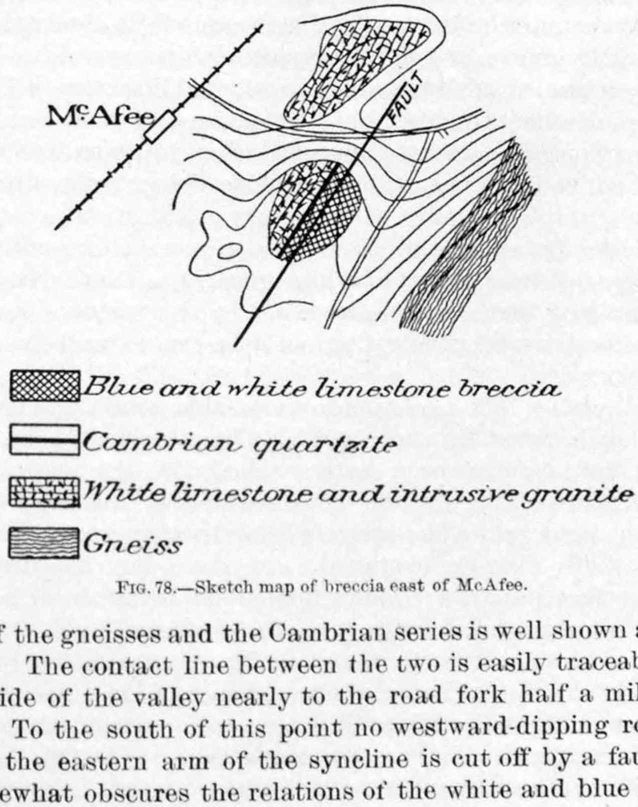


FIG. 78.—Sketch map of breccia east of McAfee.

and dip of the gneisses and the Cambrian series is well shown at several localities. The contact line between the two is easily traceable on the eastern side of the valley nearly to the road fork half a mile east of McAfee. To the south of this point no westward-dipping rocks were found, as the eastern arm of the syncline is cut off by a fault. This fault somewhat obscures the relations of the white and blue limestone at the southern end of the Sand Hills syncline.

The hill 629 feet high, one-fourth mile east of McAfee, affords some very interesting exposures, which are included in Nason's section No. IV. At this locality our observations agree in many respects with Nason's description, but we differ from him in conclusions.[1] The accompanying sketch map (fig. 78) shows the general distribution of the various rocks at this locality. On the west slope of this hill there are

[1] Op. cit., pp. 40-42.

almost continuous exposures of white limestone with considerable intrusive granite. The strike of the foliation is about N. 30° E., and it has an easterly dip of about 80°. A smaller hill, which lies to the north of the road (see map), and which is in the extension of the line of strike of this white limestone, is made up almost entirely of that rock and granite. On the crest of the large hill (629 feet) immediately east of the white limestone is found a band of quartzite and graphitic sandstone. The strike of the sandstone is about N. 30° E., but the dip is variable. At a point on the southern end of the main crest of the hill and near the southern end of the band of quartzite, a westerly dip was found. Here the sandstone dipped toward the white limestone, the latter rock having its strong easterly dip. This band of quartzite, which is but a few feet wide, can be traced continuously from the bluff at the north end of the hill to where the crest begins to fall away at the south end.

East of the quartzite is a mass of blue and white limestone breccia. In this rock fragments of typical graphitic white crystalline limestone are found embedded in a blue, almost unaltered limestone. These fragments are sometimes rounded, but oftener are quite angular, and vary from a fraction of an inch to several inches in diameter. Quite frequently the blue limestone contains considerable graphite disseminated in small plates. Often veins of secondary calcite have penetrated the unaltered blue limestone and give the appearance of fragments of blue limestone buried in a cement of white limestone. Large fragments of mica were also observed in the breccia. In thin section fine veins of quartz were observed penetrating both the cement and the fragments of the breccia.

Nason describes this locality in considerable detail, and asserts that a transition between the white and blue limestone can here be traced, and that the blue has been metamorphosed to the white limestone by the influence of the granite. The succession from west to east is as follows: First, white limestone with intrusive granite, overlain by quartzite, both dipping eastward, and the quartzite in turn succeeded by this peculiar white and blue limestone breccia. As the quartzite is not intruded by the granite or altered by it, it is difficult to conceive how the breccia, which is Nason's transition zone, could be produced by the influence of the intrusive rock. Our explanation of the facts is that the white limestone was faulted over and against the blue limestone. The breccia was produced by the fault movement and was accompanied by a certain amount of recrystallization. No white limestone actually outcrops east of the breccia, but the extension of the strikes of some exposures one-fourth mile to the south would carry it between the breccia and the gneiss. Aside from the presence of the breccia, there is pretty strong evidence of faulting at this locality, for the Cambrian quartzite dips directly toward the gneiss, which outcrops close at hand, while the eastern arm of the syncline, which has been traced nearly to this locality, is entirely wanting.

We confess that the outcrops are not sufficient to enable us to prove our explanation, but at the same time are unwilling to admit that the breccia can have had any other than a mechanical origin.

HAMBURG BLUE LIMESTONE BELT.

This belt of Wallkill limestone averages about a mile in width and extends from near McAfee for about 7 miles southwestward. At its northern end the belt forks and is dovetailed with white limestone, while at its southern end it points out in the gneissic area of Pimple Hills. The general structure of the Hamburg belt is that of a westerly dipping monocline, which is faulted off on its west side. The strikes have a northeasterly trend and the strata dip at a low angle to the west. The belt is bounded on the west by the gneisses along a fault contact, and on the east by the quartzite, which overlies the gneisses and white limestone. (See section B of Pl. LXXXIII.)

At the northern end of the belt the simple monoclinal structure is complicated by an anticline, the eroded crest of which has exposed the white limestone. This is represented on the map by the wedge-shaped mass of white limestone, which separates the two forks of blue limestone. This anticline probably dies out not far south of the southern point of the white limestone.

The relations on the western arm of the anticline are quite simple. At several localities the quartzite is found dipping at low angles to the westward in close proximity to the white limestone dipping steeply eastward. At the hill indicated as 750 feet high, one-half mile west of McAfee, the quartzite is found lying almost horizontal near the top of the hill and is overlain by the blue limestone. On the west flank the quartzite and blue limestone are found dipping at low angles to the west. The lower portion of the eastern slope is occupied by white limestone, dipping steeply to the east. The top of this hill seems to be an anticline, the west limb of which occupies the west slope of the hill. Nason's section No. V crosses this hill. We could find no evidence whatever of the sandstones dipping under the white limestone on the east flank of the hill, as he has represented in the section.

Everywhere along this contact line of the Cambrian beds and the white limestone we found evidence of the unconformity of the two series. This is notably the case at the extreme southern end of the white limestone wedge which separates the two tongues of blue limestone. At some exposures about 1½ miles northeast of Hamburg, which are close to the road leading to the old Pochuck mine, the quartzite is found in contact with the granite. The basal member of the quartzite at this point is a highly feldspathic arkose, which grades into the granite by almost imperceptible changes, but the regenerated character of which can be recognized both in the outcrop and in the thin section. Little stringers of quartzite are observed penetrating the apparently homogeneous granite. In thin sections of this transitional rock the presence of a cement can be recognized. Of this locality Nason says: "At the Pochuck mine road it [granite] has fragments of sandstone and other rocks embedded in it."[1] There are apparent crystalline schist fragments in the granite, but we observed no fragments of sandstone.

The eastern limb of this anticline is faulted off, and in only one place along this axis was any easterly dipping Cambrian found. This locality is at the old Hematite mine on the hill one-fourth of a mile west of McAfee. Here a small mass of quartzite is associated with the white limestones. The general relations are represented on the accompanying diagram (fig. 79). This sandstone has a lenticular form, being about 75 yards long and 75 feet thick. It thins out in both directions along the strike. To the west of the sandstone, and higher on the hill slope, is found white limestone, which dips under the quartzite. They both strike and dip about the same, namely: strike, N. 30° E.; dip, 80° E. The sandstone is tough, pyritiferous, and ferruginous, and contains many nodules of hematite, limonite, and calcite. Some of these may be altered pebbles, but are too much altered to permit of definite determination. The rock also contains quartz pebbles. In the old mine opening the sandstone is well exposed, and is seen to be much sheared, slickensided, and shattered. East of the sandstone, and overlying it, is another mass of white limestone, which also dips steeply to the east. The contact between the sandstone and the upper limestone is well exposed in the mine and is undoubtedly a fault contact. The white limestone and the sandstone are both beautifully slickensided. The fault plane strikes about parallel to the foliation of the limestone and dips at about 80° to the west. Nason's section No. IV[2] crosses at this point, and he argues from this exposure that the quartzite is interbedded with the white limestone. The contact between the upper white limestone and the sandstone being a fault, we can not see how these exposures bear out Nason's views. Our interpretation of the phenomena is that this is a local thickening of the Cambrian sandstone and that the upper white limestone has been thrust over it.

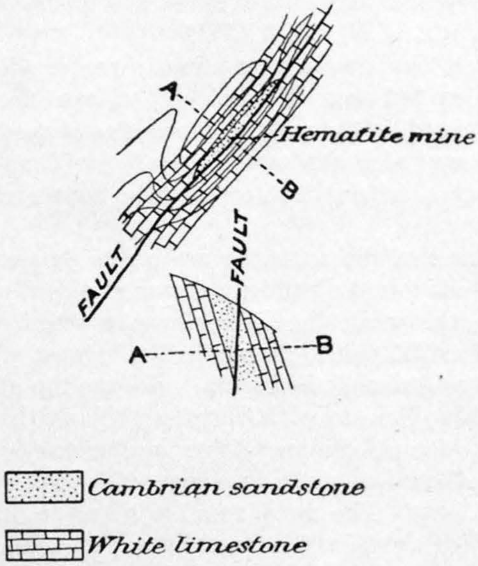

FIG. 79.—Sketch showing relation of Cambrian sandstone and white limestone at Hematite mine, one-fourth mile west of McAfee.

[1] Ann. Rept. State Geologist New Jersey for 1890, p. 31.
[2] Op. cit., p. 40.

The eastern fork of the Hamburg belt has a monoclinal structure, with a fault along its west boundary. The series is well exposed on the northern slope of the hill whose elevation is 722 feet. Near the backbone of the ridge the quartzite is well exposed. It has considerable thickness and is somewhat vitreous. It is overlain by the blue limestone, which is exposed at several places along the west flank of the ridge.

At the northern end of the crest of the hill, at the old Simpson mine, the relation of the quartzite and the white limestone is well shown. In the pits of the mine the quartzite, which is here quite ferruginous, is seen dipping west at an angle of about 75°, while it strikes N. 30° E. Close at hand is the white limestone, with a strike of about N. 20° E., and an easterly dip of 50°. There can be no doubt of the discordance of the two series at this locality.

South of this contact good exposures are afforded again 1 mile east of Hamburg. On the crest of the hill at this point we found outcrops of the blue limestone, partly shaly, underlain by 2 or 3 feet of arkose. Within 2 or 3 feet of the lowest outcrops of the arkose the white limestone, which forms the east slope of this hill, crops out, cut by a long dike of granite which runs the length of the hill. The quartzite dips west, and evidently rests on the edges of the white limestone and granite. Only a very narrow strip separates the granite from the arkose and the overlying blue limestone.

The same contact is again exposed in the brook opposite the hotel at Hardistonville. Nason has described this locality, and his section No. X[1] includes it. The presence of several faults and the occurrence of pockets of Cambrian sandstone in the white limestone make the relations of the various strata at this point rather puzzling. In the creek bed above the outlet of the mill the white limestone is found striking N. 80° E. and dipping 40° to the south. Underlying the white limestone is found the Cambrian quartzite, striking N. 60° E., and dipping 80° to the northwest. The contact between the two rocks is undoubtedly a fault, for the slickensided fault contact can be seen in the bank of the brook. In our opinion the white limestone has been faulted up over the quartzite. South of this point the brook cuts across the strike of the quartzite for about 50 feet, and the northwest dip decreases to about 35°. Close to the millrace the white limestone again outcrops and is overlain by sandstone. Just under the millrace large pebbles of white limestone were found in the sandstone. At this point there is apparently the crest of an anticline, for the northwest dip of the sandstone changes to a southeasterly dip. Below the mill in the stream bed the sandstone and white limestone outcrop irregularly. Pockets of the sandstone are here found entirely surrounded by white limestone. It seems probable that at this point the stream has cut down to the original uneven surface of the white limestone on which

[1] Op. cit., pp. 43-44.

the Cambrian quartzite was deposited. Still farther downstream, near the spring, the quartzite is again found underlying the white limestone. This seems to be another case of faulting, for along the contact there is a zone of very much sheared rock. Above this white limestone, which we believe has been faulted over the quartzite which lies below, the quartzite occurs again in its normal position. It is succeeded by the blue limestone farther down the stream.

From Hardistonville the white limestone lies east of the blue, with the quartzite between them, until near the "Green Spot," just north of Mine Hill, where the pointed gneiss area of the Pimple Hills lies between the two limestones. There are no actual contacts shown in this distance. The first outcrop of gneiss, which lies on the west side of the road near the point of this area, strikes N. 20° E., dips 75° E., pitches 20° NE., and is distinctly cut by bulging masses of granite, while the quartzite rests on the west side of the outcrop with the usual discordant structure. A few feet southwest an outcrop of granite with easterly dipping foliation lies adjacent to the quartzite. Going across the narrow tongue of gneiss to the white limestone at a point just south of the shaft of the Trotter mine one observes that the gneiss and limestone are within a few feet of each other and show a discordance of strike of 10°, the limestone and ore bed striking N. 10° E. and the gneisses N. 20° E., while the angle of dip is about the same. This is mentioned to illustrate an exception to the general rule, that the two rocks have the same strike and the same dip. The strike of the Cambrian west of the gneiss tongue is about coincident with that of the gneiss, and the white limestone resumes its coincidence with the general strike a few hundred yards north.

In the tongue of gneiss which lies between the two limestones in Franklin the linear parallel structure of the minerals, or "pitch," is well marked. At a point just west of the southern end of the zinc vein, in the first outcrops of gneiss west of the white limestone, its inclination is 28° in a northeast direction, which, as is well known, is exactly the inclination at which the zinc ore pitches from the Buckwheat mine northeasterly, and in fact at several places we have observed in the white limestone itself the same lenticular shape of its component minerals pitching northeast. This structure, while undoubtedly originating under the influence of mechanical causes, is yet a primary structure of crystallization in the rocks as they now exist, and distinct from any later shearing action. It therefore establishes another point of identity and of contemporaneous origin between the white limestone, the inclosed ore bodies, and the associated gneisses, and incidentally argues for the contemporaneous formation of the ore bodies without distinct causal connection with the granitic intrusions, however much the latter may have acted locally, since the granite has been intruded into the gneisses as a later eruptive, as well as into the white limestone.

From here to the southern point of the Hamburg belt the Cambrian shows the usual discordance in dip with the foliation of the gneiss, but the strike remains about the same.

The western boundary of the Hamburg belt has few outcrops. The monoclinal western dip continues in the outcrops nearest to the western tongue of the gneiss of the Pimple Hills. At a point along the boundary west of the Pochuck mine road, and at the east base of Pochuck Mountain, we find the limestone and gneiss within 200 feet of each other. While the gneiss has the usual foliation, the limestone strikes N. 25° E., and dips 45° W., so that its layers would abut against the gneiss if continued. No outcrop of the quartzite has been found along this contact, so that we regard it as a fault line.

The main westerly contact of the Cambrian with the gneisses of the Pimple Hills and Pochuck Mountain requires little special mention, except that at several places the quartzite or arkose is found resting on the granite. This is the case a mile south of Monroe Corners, along the west base of Pochuck Mountain, and especially in the railroad cut one-quarter of a mile west of Hamburg, where the actual contact between the quartzite and granite is well exposed. Nason says of this locality (loc. cit., p. 30): "The granite in this cut is in contact with the sandstone and appears to be fused to it. At least the union between the two rocks is so close, that a specimen may be broken off part of granite, part of sandstone." And this is taken as one of the proofs that the granite is eruptive through or against the sandstone. The quartzite here dips 35° W., while the granite has a distinct foliation dipping steeply east, or nearly at right angles to the stratification of the quartzite. The minor detail of the contact line is very irregular. The quartzite is, however, full of detrital feldspar and material derived from the granite, and it is therefore a duplication of the similar arkose contact described above at the Pochuck mine road. The general foliation of the granite shows no parallel relation to the contact line, but is perpendicular to it. Had the granite really flowed against the quartzite, it is inconceivable that this structure, considered as original flow structure, should stand in such relation to the contact; and it would be even more difficult, if the foliation of the granite be considered as due to subsequent metamorphism, to conceive that the change should stop short of the quartzite, which is absolutely without any such secondary foliation, or not affect the overlying limestone. We dwell on this rather self-evident proposition, because the same discordance between stratification of the quartzite and blue limestone and foliation of the adjacent crystalline rocks, whether granite, gneiss, or white limestone, is found universally.

### OGDENSBURG BELT OF BLUE LIMESTONE.

The Ogdensburg belt of Wallkill limestone, occupying the valley of the Wallkill, is poor in outcrops, especially along the east border against the gneisses at the foot of the Highlands; and while, as mentioned previously, the abundant fragments of sandstone east of Ogdensburg make it probable that the sandstone really lies under the limestone against the gneisses, there is no outcrop in place and no outcrop of the limestone itself near the gneisses. In Sparta the nearest outcrops of blue limestone dip gently west. The belt, as a whole, shows several gentle folds, the most prominent of which is developed as an anticlinal axis in the northern half of the belt.

The western contact of this belt requires a detailed description, for it is the principal line along which the two limestones are found in contact without the usual intervening quartzite.

About one-third of a mile north of Ogdensburg a large isolated outcrop of gneissic rock is situated 30 feet from the nearest outcrop of blue limestone, the bedding of which is obscure, but probably dips 40° W. The gneiss is a massive greenish rock, with irregular coarse and fine bands, the latter almost felsitic in appearance, evidently a stretched and brecciated gneiss or granite. In thin sections the autoclastic character of the rock is confirmed. The broken cores of the original rock lie in a confused aggregate of large and small angular fragments of the same quartz and feldspar, with the development of minute greenish micaceous scales in the finer portions. The original rock was a gneiss or granite. The belt of white limestone runs west of this outcrop, but the contact is not shown.

If one follows the border a quarter mile north, the outcrops of the two limestones are found within 50 feet of each other, and the blue dips 40° W. A few hundred yards farther the Wallkill River crosses the contact line at the southwest base of the hill with an elevation of 689 feet (see map). Here, in the space between the railroad and the river bank, are numerous outcrops of white limestone, the foliation of which strikes N. 20° to 30° E. and dips 75° E. On the bank of the river are large ledges of a granular speckled white gneiss, the foliation of which strikes N. 30° E. and dips 75° E. Lying in contact with this is a ledge of granite, which sends apophyses into the gneiss, cutting across the foliation and even bending the edges of the gneiss bands. The nearest outcrops of white limestone are 35 feet distant, west of the granite. We therefore have here gneiss with foliation parallel to that of the white limestone distinctly cut by intrusive granite. The gneiss is a normal, well foliated muscovite-biotite-gneiss. The nearest outcrop of the blue limestone, some distance east across the river, dips 60° W.

Still following the line north and crossing the furnace pond on its north side, the blue and white limestones come close together just north of the road. The blue limestone steepens its westerly dip on approaching the white, the nearest outcrop of which is 40 feet distant. Following the line a little farther, at a spring on the hillside we find on the east the blue limestone dipping 70° W. Six feet below it vertically and 10 feet west horizontally is a banded biotite gneiss or granite, the structure of which dips 60° E. It has fine-grained, greenish, felsitic bands, which are apparently shear planes, and this interpretation is confirmed by the thin section, in which the bending and shearing of the minerals is plain. The thickness of this rock as exposed is about 10 feet. The western part of the outcrop is a coarse, indistinctly foliated rock, probaby granite. One hundred feet across the strike is the nearest outcrop of white limestone, the foliation being well marked by silicate bands and dipping 68° E. Here again, within a foot or two, we have the strata of the blue limestone abutting against the gneissic rocks, succeeded by white limestone, the foliation of both being discordant to the dip of the limestone and the gneiss nearest the contact sheared.

Still farther northeast, where the line crosses the next road, we find, as before, the west dip of the blue limestone steepening as it approaches the white, and between the two are outcrops of rock, which have alternating bands of whitish and bluish limestone standing nearly vertical, so that it is difficult or impossible to draw a line between blue and white.

On crossing the road, passing a little swamp, and approaching the extreme point of the Hamburg belt of blue limestone, granite and the two limestones are found together in a puzzling relation. Nason's section XIII crosses near this point. A large bulging mass of granite (marked on the map) runs parallel to the trend of the white limestone, and at its southern termination comes within 10 feet of the blue. A little farther north the two limestones are connected by a series of apparent transitions, which are so well described by Nason that we can not do better than quote his words:

At this point the white limestone is seen in direct contact with the granite. Within 50 feet to the east of this the blue unchanged limestone is seen outcropping. Between these two points there is 30 feet which is concealed by soil. At various points along the strike of the limestone are to be observed numerous small outcrops of limestone in every degree of transition from the unchanged blue to the white. Absolute continuity was not visible. To settle the question of continuity a trench was dug across the strike exposing a continuous series. As was expected, the rock exposure showed a progressive transition through almost every shade of color, from the blue limestone to the coarsely crystalline snowy white.

The first change was that the blue limestone was broken into closely adhering prismatic fragments, showing light irregular or mottled streaks with cloudy blue aggregates Next, the limestone showed evidences of strong compression, with pressure planes developed. Along these planes, through which the rock split easily, were cloudy carbonaceous bands, with the rock a lighter color. Then the bands became more pronounced, with occasional crystalline scales of graphite. Then, next, the bands become yet more distinct, and the black earthy bands of carbon were changed to bright welts of crystalline graphite. The pressure planes were still strongly developed. Finally, within 6 feet of the granite, the pressure planes become less apparent, and coarse crystals of limestone with scales of graphite, chondrodite, and other minerals were abundantly developed.[1]

It will be seen that there is a shear zone along the contact of the two limestones. We wish to add that the apparent dip in this transition

[1] Am. Geologist, Vol. XIII, p. 160.

zone is a laminated structure caused by bands of limestone of varying coarseness of grain which have a vertical position or very steep southeast dip, and seem to us to represent the position of the shear planes rather than stratification.

We have described this whole western contact line of the Ogdensburg belt of blue limestone in considerable detail because the apparently transitional relations of the two limestones in the part just described have been a strong argument for their identity. It will be seen that along a part of the contact a narrow belt of crystalline rocks (gneiss and granite) lies between the two limestones and has been greatly sheared, and that the blue limestone dips discordantly against the foliation of these rocks, the quartzite being absent. It seems to us plain that it is a fault line accompanied by shearing, and that the transitions of the two limestones at the northern end represent the same

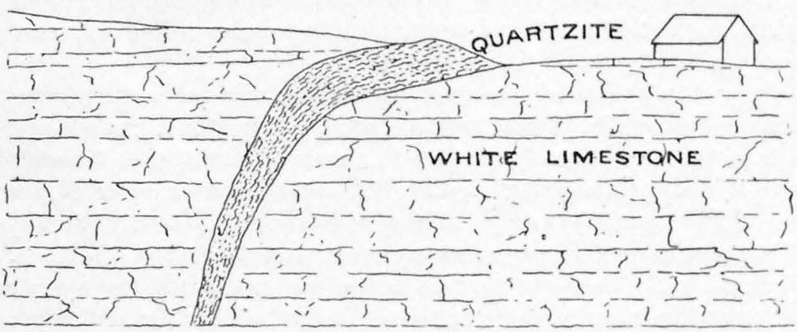


FIG. 80.—Vertical quarry face showing band of quartzite in white limestone, Franklin Furnace, New Jersey.

shear zone with considerable recrystallization of the carbonates. The interposition of a band of gneiss with eruptive granite between the limestones at several places along the contact also makes their identity impossible by any simple and rational theory.

The lack of outcrops makes it impossible to determine the relations of the two limestones along the east side of the point of the Hamburg belt.

AREAS OF HARDISTONVILLE QUARTZITE IN THE WHITE LIMESTONE.

The southernmost of the three patches of quartzite lying isolated in the area of the white limestone is in Franklin, north of the pond. In 1896 a new quarry in the white limestone was in operation, situated just north of the road, and in September of that year a vertical face of the rock about 35 feet high was exposed, facing west. At the top of the quarry and running a few feet back from the face a small patch of quartzite (arkose) lies on top of the white limestone. It is 10 or 15 feet thick, strikes about east and west, and dips 10° N. This was exposed in the quarry face nearly to the bottom and had the shape shown in the sketch (fig. 80). It entered the white limestone, curving gradually downward and contracting until it was about 3 feet thick at the bottom. The top and bottom contacts showed no sign of faulting. It had evidently been formed in place. The rock is a pyritiferous quartzite, composed of clastic quartz, feldspar, light yellow muscovite, and a cement partly fine-grained silica, partly calcareous. It has been quarried out in large blocks, exposing the rock in ideal freshness. The important feature seen in these blocks is that large pebbles or bowlders of typical white limestone are inclosed in the quartzite. The rock composing the pebbles has scales of graphite and yellow mica, and the broad cleavage surfaces of the calcite masses show the dull spots (luster mottling) due to small pœcilitic grains of calcite with different orientation, which is characteristic of the white limestone. Pieces from the main ledge of white limestone have the same minerals and the same structure, so that their nature as pebbles and their source are beyond question. The largest observed had the shape shown in the cut (fig. 81), and was 10 inches long and 3 wide; others are rounded. The rock also contains round masses of purple fluorite associated with quartz and calcite, which are probably concretionary, but have no resemblance to these pebbles. There are occasionally black slaty stratification bands in the rock, which conform to the contacts, so that the stratification curves with the band of rock. A few yards south of the surface outcrop another small patch is exposed by a shallow excavation, sufficient to show its apparent continuity with the limestone. Nason says:[1] "This excavating showed the sandstone going under the limestone with a northwest dip."

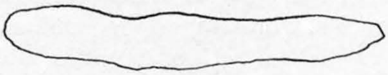

FIG. 81.—Pebble of white limestone 10 inches long and 3 inches wide, occurring in quartzite band.

The foliation of the white limestone here strikes N. 20° E. and dips 70° E., so that this quartzite layer is about transverse in strike. It is plain that this quartzite (arkose) contains the débris of the white limestone and was laid down on its eroded surface, which was necessarily creviced and irregular, and the peculiar form of the deposit we think can be explained only as due to the washing of the sand into a large cavity. It is difficult to understand why the stratification curves downward instead of lying horizontally across the original cavity. We simply note that such is the fact.

The second patch of quartzite is on the west side of the hill, with an elevation of 720 feet between Franklin Furnace and Hardistonville. This is on Nason's section XII, and is described in his text. It is a little difficult to identify the areal distribution of the rocks as described; but the main point is that a bed of sandstone is there described as dipping under the white limestone, which with granite forms the main part of the west slope of the hill. We found the quartzite in place here in a small exposure in contact with the white lime-

[1] Ann. Rept. State Geologist New Jersey for 1890, p. 46.

stone. The line of junction was about vertical, the two rocks being in firm contact. Large pebbles of the white limestone were found inclosed in the quartzite, as before. On the top of the hill another little patch of quartzite lies between outcrops of white limestone. We regard these as similar in occurrence to that at Franklin, just described, and to those in the brook bed at Hardistonville, where little residual patches of the quartzite deposited on the uneven and creviced surface of the white limestone have been preserved from the general erosion which removed the later rocks covering the white limestone.

## SUMMARY AND CONCLUSIONS.

In the area under consideration the white limestone forms a belt less than a mile wide, which runs in a northeast-southwest direction. Its strike and dip, as measured by the foliation, are quite uniform, and the dip is generally steep to the east. These structures are usually parallel to similar structures in the gneiss, and the pitch structures of the gneiss, white limestone, and associated ore deposits have a general parallelism both in direction and in angle. Nowhere was any trace of stratification planes observed in the white limestone. The supposed cases of interbedding of the white limestone and the Cambrian quartzite we have found to be due to faulting or to peculiar conditions of deposition. While it is difficult to prove that the white limestone and gneiss are actually interbedded, narrow bands of true gneisses do occur within the white limestone belt and seem to be an integral part of the series.

The intrusive relation of the granite to the white limestone and gneisses is evident, and it is equally clear that its injection took place previous to the deposition of the Cambrian quartzite and blue limestone. This is proved by the nature of the contacts of the quartzite and granite. We do not regard the main crystalline condition of the white limestone as due to the granite intruded in it, but believe that this crystallization antedated the granite intrusion, and was contemporaneous with the crystallization of the gneisses to their present form. At the same time we do not deny local metamorphism, sometimes of considerable extent, along the contacts of the granite.

The Wallkill or blue limestone of Cambrian age occurs in three separate belts within the mapped area. The Sand Hills belt is a shallow syncline (see section A of Pl. LXXXIII), the eastern arm of which rests on the gneiss and the western arm on the white limestone, and it is faulted near its southern apex. The Hamburg belt is bounded on the east side by a normal basal contact, along which the basal arkose contains pebbles of limestone, detrital mica, and feldspar. The general structure of this belt is that of a low-dipping monocline which is faulted on its west side against the gneisses. At the northern end the monocline is complicated by the development of a central anticline which is faulted on its southeast side. (See section B of Pl. LXXXIII). The

Ogdensburg blue limestone belt is faulted on its west side against the white limestone and gneisses. In this belt the dips are of moderate steepness. (See section C of Pl. LXXXIII.)

Along the normal contacts of the blue and white limestone the quartzite intervenes between the two. We everywhere find the bedding of the blue limestone and underlying quartzite conformable, while the dip of the foliation of the white limestone and the gneisses is discordant with them. The strikes may or may not be parallel.

Isolated patches of the quartzite are found within the white limestone area, which were originally deposited on the eroded and creviced surface of the white limestone, filling in the inequalities and sometimes causing the peculiar intermingling which has been described. The fortunate preservation of one of these crevices filled with quartzite containing undoubted pebbles of white limestone, as well as arkose material, leaves no doubt as to their origin. Where this has been complicated by local faulting we get such obscure relations as those exposed in the brook at Hardistonville. Along the fault lines, where the blue and white limestones come in immediate contact, we find apparent transition zones composed either of breccia or of sheared limestones. These phases we attribute to a mechanical origin, accompanied by some recrystallization.

We conclude, therefore, that the white limestone was deformed and metamorphosed to its present condition and partly eroded before the basal member of the Cambrian series was laid down; that the deformation which the area has suffered since the deposition of the Cambrian, which has manifested itself in folding and faulting, has been but slight compared with the pre-Cambrian deformation; and that the Cambrian rocks overlie the white limestone, as well as the gneisses, unconformably.

As a result of our observations, we are compelled to adopt the pre-Cambrian age of the Franklin white limestone.

# A GEOLOGICAL SKETCH OF SAN CLEMENTE ISLAND.

BY

WILLIAM SIDNEY TANGIER SMITH.

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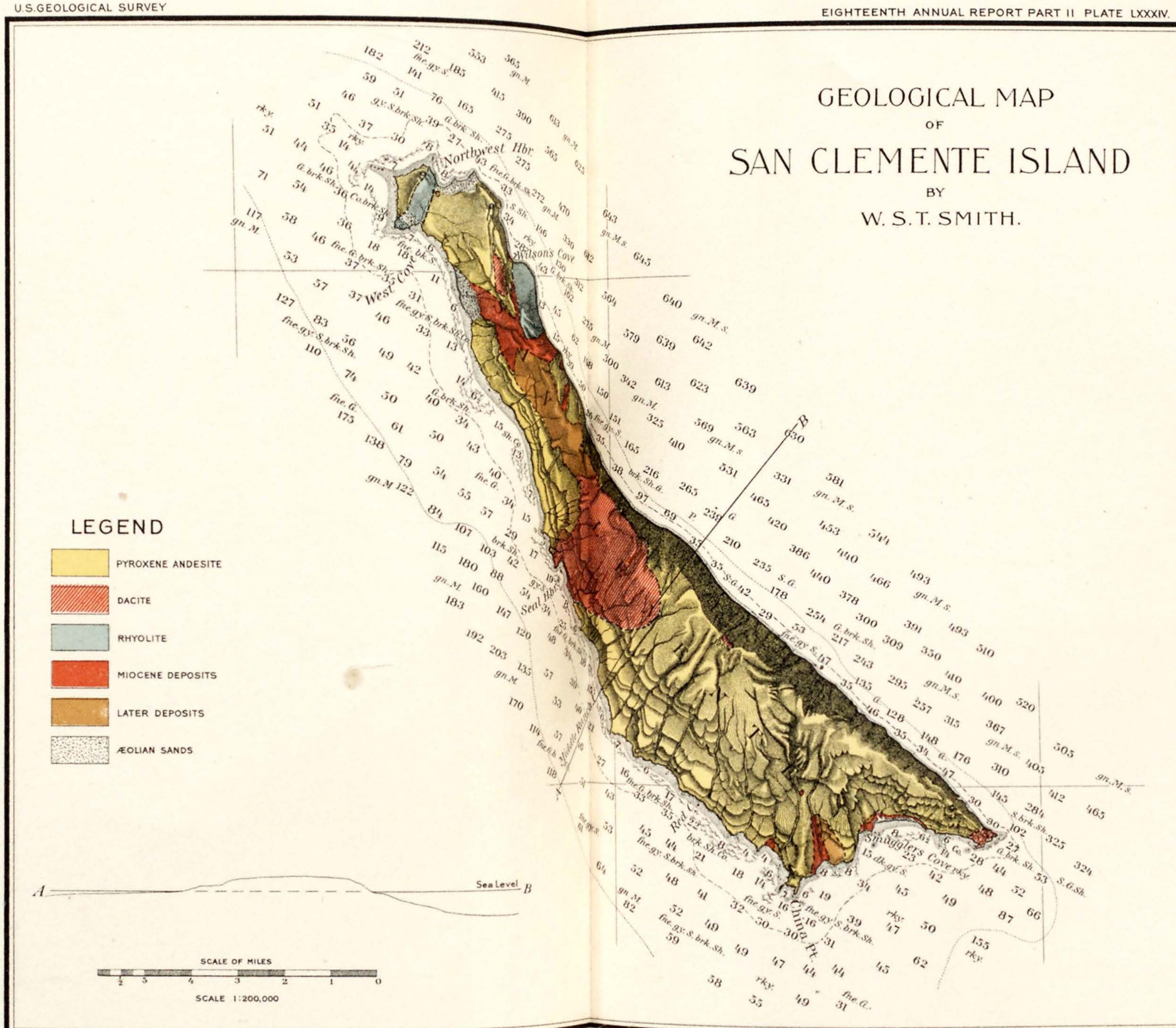
# ILLUSTRATIONS.




*Topography from U. S. C. & G. Survey Chart.*

# A GEOLOGICAL SKETCH OF SAN CLEMENTE ISLAND.

By W. S. T. Smith.

## INTRODUCTION.

The northern part of the California coast is marked by its freedom from islands, the Farallones, a small group opposite the Golden Gate, being the only ones of any importance. Within the last 300 miles of the coast, as we go south, however, we find two well-defined groups, commonly known as the Channel Islands, the individual members of which range in length from less than a mile to nearly 25 miles. The islands of the northern group form a nearly straight line, with only narrow channels between them, and appear to be genetically one. The four islands of the southern group, however, are widely separated, and must be considered as distinct units. The islands of both groups have a general trend parallel to that portion of the coast to which each is nearest. San Clemente is the southernmost island of the southern group.

*Literature.*—The only literature known to the writer bearing on the geology of San Clemente Island is a note by Dr. J. G. Cooper [1] and a later and more extended account of the island by Professor Lawson.[2] Both of these deal especially with the physiography of the island, and in particular with its remarkable terracing. Lawson further characterizes the island as a tilted orographic block.

## GENERAL DESCRIPTION.

San Clemente is habitable at but few points, owing to the scarcity of water. At Wilsons Cove rain water is collected in tanks of considerable size, furnishing the only supply at this point. There are few springs on the island, and the waters of these are decidedly alkaline. Several wells have been dug, of which three were successful, and the supply from these is abundant. The water of one of them, however, is rendered very brackish by the infiltration of sea water, and another is somewhat brackish, probably from the same cause.

Besides these sources of supply, water is found in occasional natural tanks in the larger canyons. These tanks are simply cavities in the rocky stream-beds at the foot of waterfalls, in which a portion of the

[1] Geol. Survey of California, Geology, Vol. I, 1865, pp. 182-186.

[2] The post-Pliocene diastrophism of the coast of Southern California, by Andrew C. Lawson: Bull. Dept. Geol. Univ. Cal., Vol. I, No. 4, December, 1893, pp. 135-139.

winter's rain water collects. Here it is kept, in many instances through the entire summer, furnishing drink for the birds and wild goats, for the sheep, and even for man. The tanks, like other features of San Clemente, remind one of the desert. The island has, in many respects, a desert-like character, though surrounded by water and doubtless receiving as much rain in the winter months as any of its sister islands.

The vegetation is limited almost entirely to low shrubbery and herbage. Occasional small trees (*Lyonothamnus floribundus*) are found in the largest canyons, and shrubs of some size grow on alluvial fans at the mouths of some of them; but aside from these not a tree can be seen in traveling from one end of the island to the other. That trees once grew on other parts of the island is shown by petrifactions found in the sand area opposite Wilsons Cove. A few of these petrifactions, in the form of tree trunks and branches up to a diameter of 6 inches, were seen projecting above the sand, while at numerous other points the sand contains root-like masses, doubtless the remnants of shrubs, the woody fiber in all cases replaced by calcite, or by sand cemented by calcite. In other instances the form is preserved as a hollow tube of sand cemented by calcite.

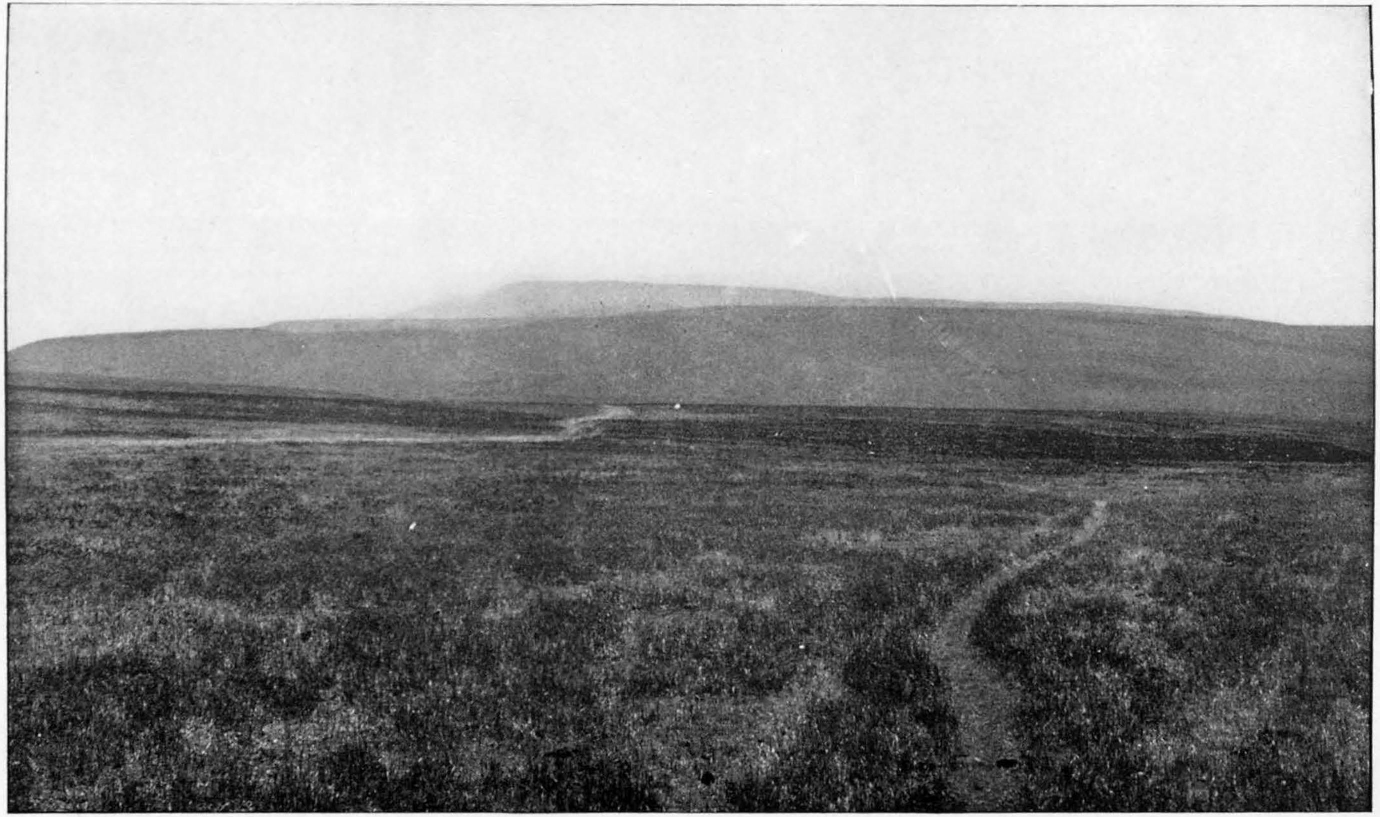
Several varieties of cactus are found, and this plant appears to form the chief growth of the island. There is also a large amount of what is locally called "salt-grass"—two species of Mesembryanthemum—particularly toward the northwestern end of the island. Two species of "foxtail" (Hordeum) grow thickly on the upper, sandier parts. Occasionally a hardier wild flower and the remains of dead plants point to a considerable short-lived growth in the spring. Over a large portion of the island, however, a very thin soil and a superabundance of rocks preclude the growth of almost everything except cactus. It is said that the vegetation of the island had an entirely different aspect twenty years ago, and that the change is due to the ravages of the sheep, which are pastured here in thousands.

Besides the sheep, and the cattle pastured with them, there are many wild goats, originally introduced from Santa Catalina. Occasional foxes are seen, but the squirrel, so well known in other parts of the State, is not found at all here. Lizards are numerous, but snakes and frogs are unknown. Landshells—*Helix* (*Arionta*) *intercisa* W. G. Binney, *H. gabbi* Newcomb, and *H. stearnsiana* Gabb—abound in portions of the island. San Clemente has no permanent human inhabitants except one old man, who has lived there most of the time for the last thirty years.

## TOPOGRAPHY.

As has already been shown by Professor Lawson,[1] San Clemente is a simple, tilted block, little modified by the forces of erosion. The topography in general is very monotonous, the only variation being in the greater or less development of its peculiar characters, so that an

[1] Op. cit.: Bull. Dept. Geol. Univ. Cal., Vol. I, No. 4, December, 1893, pp. 135–139.



SUMMIT, ABOUT 2½ MILES SOUTHEAST OF WILSONS COVE, SAN CLEMENTE ISLAND; LOOKING SOUTHEAST.

inspection of one part of the island gives a fair idea of the whole. The trend of the island is northwest and southeast. It lies some 50 miles to the south of San Pedro Hill, the nearest point on the mainland, and about 25 miles south of Santa Catalina Island. It has a length of nearly 21 miles, with a maximum width of a little over 4 miles, narrowing near the northwestern end to about 1 mile. The greatest altitude is 1,964 feet, at a point a little to the east of the center. From this highest point the descent to either extremity of the island is so gradual that a horse and wagon could easily travel the greater part of the entire length. As seen on the map, the main watershed lies very near the northern coast, and is continuous from one end of the island to the other. From this watershed, for most of its length, the descent on the northern side is very abrupt—in one case amounting to 1,800 feet in half a mile. Along the higher parts of the northern coast there are only two or three places where the shore may be reached from above; and although one may descend some of the larger canyons on this side for several hundred feet, the descent is sooner or later checked by waterfalls of considerable height. On the southern side of the island the average descent is gradual, broken only by the remarkable terraces which are characteristic of the island, and of this side particularly.

Owing to its terraced condition the upper portion of the island presents the appearance in transverse section of an uplifted table, marked by moderate relief in some parts, in others scarcely modified save for a gentle slope toward either side. The relief is not very pronounced at most, and is due to terracing, to subaerial erosion, and to faulting. Where the topography is carved from soft sandstones and shales, we should expect to find the least contrast, and in such parts the surface is, at a few points, much like a gently rolling prairie. In one or two places the surface appears almost level as far as the eye can reach. These gentler slopes are found along the main ridge for a stretch of several miles to the southeast of Wilsons Cove, and also for some distance near the southeastern end of the island. Pl. LXXXV shows this character of the summit, and Pl. LXXXVI shows the summit toward the northwestern end of the island.

The main reasons for believing San Clemente to have been formed as a tilted crust block are its form, minor faulting parallel to the island's longer axis, and the dip of the volcanic rocks. From the description already given it may be seen that the island still presents the appearance of such a block, with one side marked by an abrupt cliff, the other by a gradual slope. Further, the contrast of these slopes is continued below sea level, as shown by the section near the middle of the island, along the line *A—B* on the map, Pl. LXXXIV. The greatest contrast is seen a little to the east of Wilsons Cove, where on the north a depth of 3,600 feet is reached in less than 2 miles from the shore, while at the same distance on the opposite side of the island the depth is only 300 feet.

We can not doubt that this northern scarp is due to faulting. Faulting more or less parallel to this edge is seen at a number of points, particularly toward the northwestern end. One of these minor faults is about $2\frac{1}{2}$ miles southeast of Wilsons Cove, where it presents a scarp with a maximum height of 50 feet. Another, the largest minor fault found, is seen to the northwest of Wilsons Cove, and was followed for a distance of about a mile through a change of altitude of over 200 feet. The fault scarp in this instance reaches a maximum height of 70 feet. This line of faulting makes an angle of about 30 degrees with the trend of the island. Another fault seen at Wilsons Cove makes an angle of about 40 degrees with the general direction of the island. Here the amount of faulting could not be determined, and it may be as much as 100 or 200 feet. In this instance the forces which produced the fault have continued in operation to comparatively recent times, for the fault line seen at the shore shows a displacement of the lowest terrace deposit of between 10 and 20 feet. All of these faults hade in the same general direction as the major fault, which gave rise to the island.

FIG. 82.—Submarine sections between San Clemente Island and San Pedro Hill. Scale 1:200,000.

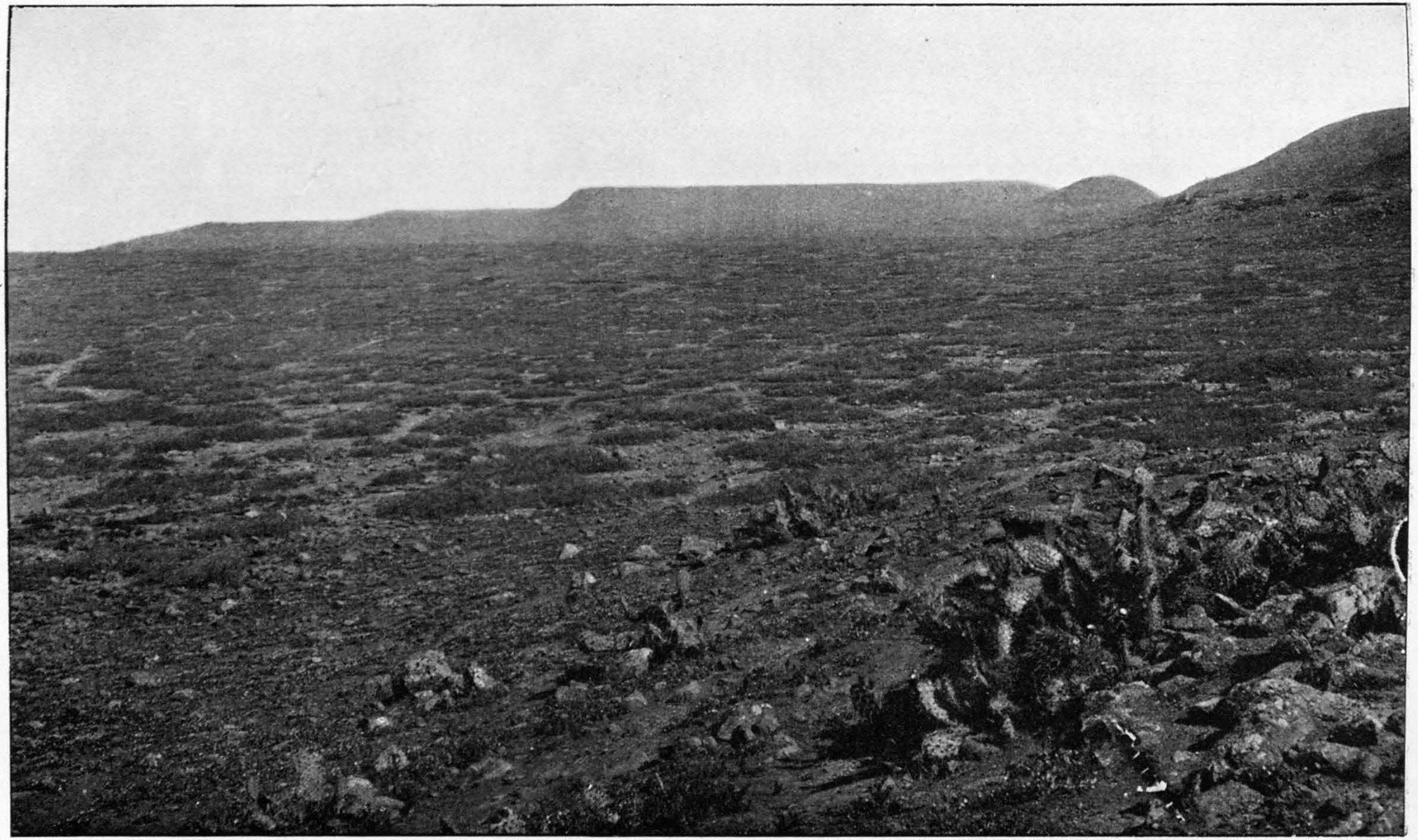
Faults of this sort present cliffs or scarps similar in appearance to those of the terraces, and under certain circumstances it would be very difficult, if not impossible, to distinguish a fault from a terrace. Circumstances favorable to their discrimination are found: when the line of scarp and platform is not a horizontal one (provided, of course, the lines of the true terraces do not show a similar deformation); when the fault cuts across the terraces, causing a displacement of the latter; and, occasionally, when the fault occurs upon a terrace in such a way that the scarp of the fault faces the cliff of the terrace, thus forming a trough. In the last case circumstances may be such as to lead to the formation of a small lake or playa. The desiccated remains of such a small lake were seen at one point.

Besides the faults above described, which are roughly parallel to the main fault line, a number of faults occur transverse to the island's length.

From a study of the submarine contours, and from the gradual decrease in altitude from the summit of the island toward either end, it appears that the major fault dies out gradually in either direction and probably extends no great distance beyond the extremities of the island.

## DRAINAGE.

As San Clemente may be considered a typical fault block, and as it is isolated from other land masses, thus forming a unit in itself, a study of its drainage should give us the characteristics of the drainage of a faulted block—seen to best advantage here, as this feature of the topography is still in its infancy.



SUMMIT, ABOVE WILSONS COVE, SAN CLEMENTE ISLAND; LOOKING NORTH

Most of the streams on the southern side nave their sources in the upper, plateau-like stretches of the island. Here many of the minor streams flow over the upper terraces in shallow channels which form but slight surface indentations. Some of these minor streams disappear entirely after flowing over a few terraces, while others persist nearly, if not quite, to the ocean. Many of them are lost on the lowest terrace, just above the shore. The minor streams here referred to are those which run from the main ridge toward the ocean (on the south) and comprise most of the streams in the western half of the island. The streams from Seal Harbor to Smugglers Cove, on this side of the island, have an entirely different character. Most of them, beginning on the upper terraces, as in the other cases, soon cut deep V-shaped canyons, which grow deeper as they advance, in some places with almost perpendicular walls. The stream flowing in the bottom of the canyon is sometimes as much as 300 feet below its upper edge. These large canyons cut across the terraces with startling abruptness, as they frequently can not be seen until one has almost reached the edge. Professor Lawson has noted this effect, and the term "saw-cuts," applied to them by him, is most apt. The steepness of their walls, together with their great depth, makes a crossing impossible in most places.

Most of these canyons have a simple V shape, but in at least two instances toward the southeastern end of the island the V expands suddenly at the very bottom of the canyon, so that the stream channel is almost shut in overhead by the canyon walls. Some of the larger streams have cut their channels down to sea level at the mouth, while in other cases the canyon ends on one of the lowest terraces, where the streams have formed alluvial fans of considerable size, in which are cut numerous channels. The best example of those streams which have cut down to sea level is in Red Canyon. The canyon walls at the mouth rise almost perpendicularly, and the stream bed here is separated from the ocean only by a bar thrown up by the waves. Making due allowance for exceptions, it may be stated in a general way that as the southeastern end of the island is approached the stream canyons are cut deeper and the condition is more nearly that of graded streams. To the east of China Point, at the mouth of one of the stream courses, there is a small lagoon, containing a diminutive island. Two such lagoons are indicated on the Coast and Geodetic Survey's large manuscript map of San Clemente (scale, 1:20,000), but only this one was noted by the writer.

The stream beds in all the larger canyons of the southern side of the island have a very moderate average grade. The course of the stream is interrupted by occasional falls, some of them of considerable height. In general these falls are due to a variation in the character of the rocks over which the water flows, though in a few instances they are probably due to faulting. One such fault, in Middle Ranch Canyon, has a throw of some 30 feet. The larger canyons differ considerably in width in their different parts, the maximum width reached by any of them being about 100 feet at the bottom. The streams which flow in the largest canyons seldom preserve a straight course for any considerable distance in the lower portions of the channels, but wind about continually. This is due to erosion on the sides of the canyons. There are few important branches of the main streams, and the branches in general are roughly parallel to the direction of the principal drainage line—that is, a direction approximately at right angles to the trend.

On the northern side of the main divide of the island the grade of the streams is necessarily greater; the courses are shorter and falls are more frequent. On account of the steeper grade the small stream will erode more rapidly than a similar stream on the gentle slope of the opposite side, and will, therefore, form a more pronounced canyon. As a consequence, the number of noticeable canyons on the northern slopes of San Clemente is greater than on the southern side. The sides of the canyons being cut more rapidly, there is a tendency to more open stream courses. As the canyons are so numerous, rapid side cutting will sooner or later cause those which are close together to coalesce, thus forming wider canyons. Branch streams on the steep slope have comparatively wider canyons than similar streams on the opposite side of the island, thus tending still more to the broadening of the main canyon, especially in its upper portions. For these reasons there is a marked tendency on the steep face of a tilted block to the formation of cirques, while more trough-like channels characterize the gentle slope. The trough-like stream courses are well marked on San Clemente, and many stages in the formation of cirques may be seen on the large Coast and Geodetic Survey map of the island. The stream development on the northern side, however, is not sufficiently advanced to show the open cirques to the best advantage. They are better seen where the erosion has been carried further, as on the ocean side of the Montara crust-block, to the south of San Francisco.[1]

As the slope toward the line of faulting is by far the steeper, erosion on that side is much more rapid, and consequently the main watershed of the unmodified crust-block migrates slowly away from that side toward the other. In the case of a crust-block, then, the movement of the divide is from the line along which the elevation takes place—that is, from the line of faulting. This differs from an uplift along an axis, without faulting, in a topographically simple region of homogeneous rocks, as in the latter case the axis of uplift itself forms the ultimate divide, while in the former case the resulting divide is situated at some distance to one side of the line of faulting. Thus the effect on a simple drainage system already established will be different for the two kinds of crustal movements, and the law for the migration of divides as given by Campbell[2] must be modified in order to make it applicable to an uplift accompanied by pronounced faulting.

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[1] Fifteenth Ann. Rept. U. S. Geol. Survey, 1895, p. 472.

[2] Jour. Geol., Vol. IV., No. 5, July-Aug., 1896, p. 580.

TYPICAL VIEW OF A LARGE CANYON, SAN CLEMENTE ISLAND.

This may be illustrated as follows, assuming the simplest possible conditions, as Campbell has done, in order to eliminate disturbing elements from the problem.

In figs. 83, 84, and 85, C represents the divide between the symmetrical drainage slopes of the two stream profiles C B A and C D E. If faulting occurs parallel to this divide, with a downthrow toward the left, it may occur on either side of C (leaving out of the question the possible occurrence of a fault at C itself). Suppose, first, that the fault is to the left of C, at B (fig. 83), and that the faulted portions have assumed the positions indicated by B″ C′ D′ E and A B′, A and E representing the limit of movement in either direction.[1] The part B C will have had its angle of slope decreased by elevation, and therefore the rate

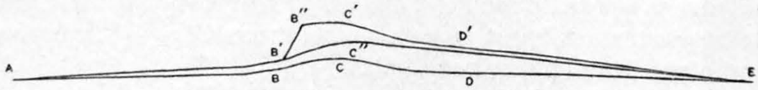



FIG. 83.—Diagrammatic section to illustrate the effect of uplift with faulting parallel to a stream divide, the elevated portion including the divide.

of erosion on this slope will be diminished. The angle of the slope C D E, however, will have been increased, and its rate of erosion will be correspondingly greater. The slope C′ D′ E, then, will be cut away more rapidly than the opposite slope C′ B″, and the divide C′ will migrate toward B″, in accordance with the normal operation of the law. This condition, however, will last but a comparatively short time, for on account of the high angle of B′ B″ erosion on this slope will be very vigorous, and the point B″ will be rapidly carried back toward C′ till the two unite. When this stage is reached the edge of the faulted block will be represented by C″ and the profile by B′ C″ E. Further erosion will tend to carry the crest C″ toward E, owing to the more rapid cutting on the steeper slope B C″. The final result will be that

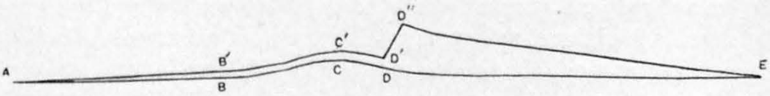



FIG. 84.—Diagrammatic section to illustrate uplift with faulting parallel to a stream divide, the elevated portion excluding the original divide.

the point C″ is carried to some point, such as D′, the exact position of which depends on the relative attitude of the points which correspond to the points B′ and E when erosion has reached this stage.

The second case (fig. 84), when the faulting is between the divide C and the point E, is similar to the last in the movement of the crest of the faulted block toward E, or away from the line of faulting. The initial movement in the former case, however, was toward the fault line, while in this case that element is eliminated, as the faulted portion D″ E excludes the original divide C. In this second case the movement is away from this divide C also, and the final result of the

[1] These sections are diagrammatic and do not attempt to give the exact relative positions of the two parts, as these positions depend on the circumstances obtaining at the time of faulting.

uplift is the establishment of a divide at some point between D′ and E, depending on the relative attitude of these two points.

If the faulting has been sufficient, this resultant divide will form the main watershed for the region, the original divide C being now of insignificant proportions. The faulting may, however, be such that the portion A B C D can not be left out of account. If, as shown in fig. 84, the final position of A B C D is more elevated than its original position, the slope C D will have been decreased, while A B C will have been slightly increased. As a consequence, there will be a migration of C′ toward D′. If the faulting is sufficient in amount, the migrating crest will finally reach D′, and there will be no divide other than that of the faulted block D′ D″ E. As the movement of the crest line in this case is toward E for the parts on both sides of the fault line, the resultant crest must be at some point between C′ and E, whether the amount of faulting be great or small.

If, on the other hand, the movement of A B C D is one of depression (fig. 85), the result is more complicated, and the final position of the crest C will depend on the relative attitudes of the faulted portions A B′ C′ D′ and D′ D″ E. In other words, if the sum total of movement in the two parts produces elevation, the migration of the divide

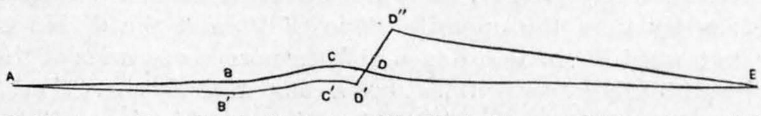



FIG. 85.—Diagrammatic section to illustrate uplift with faulting parallel to a stream divide, the elevated portion excluding the original divide, the portion A B C D having suffered depression.

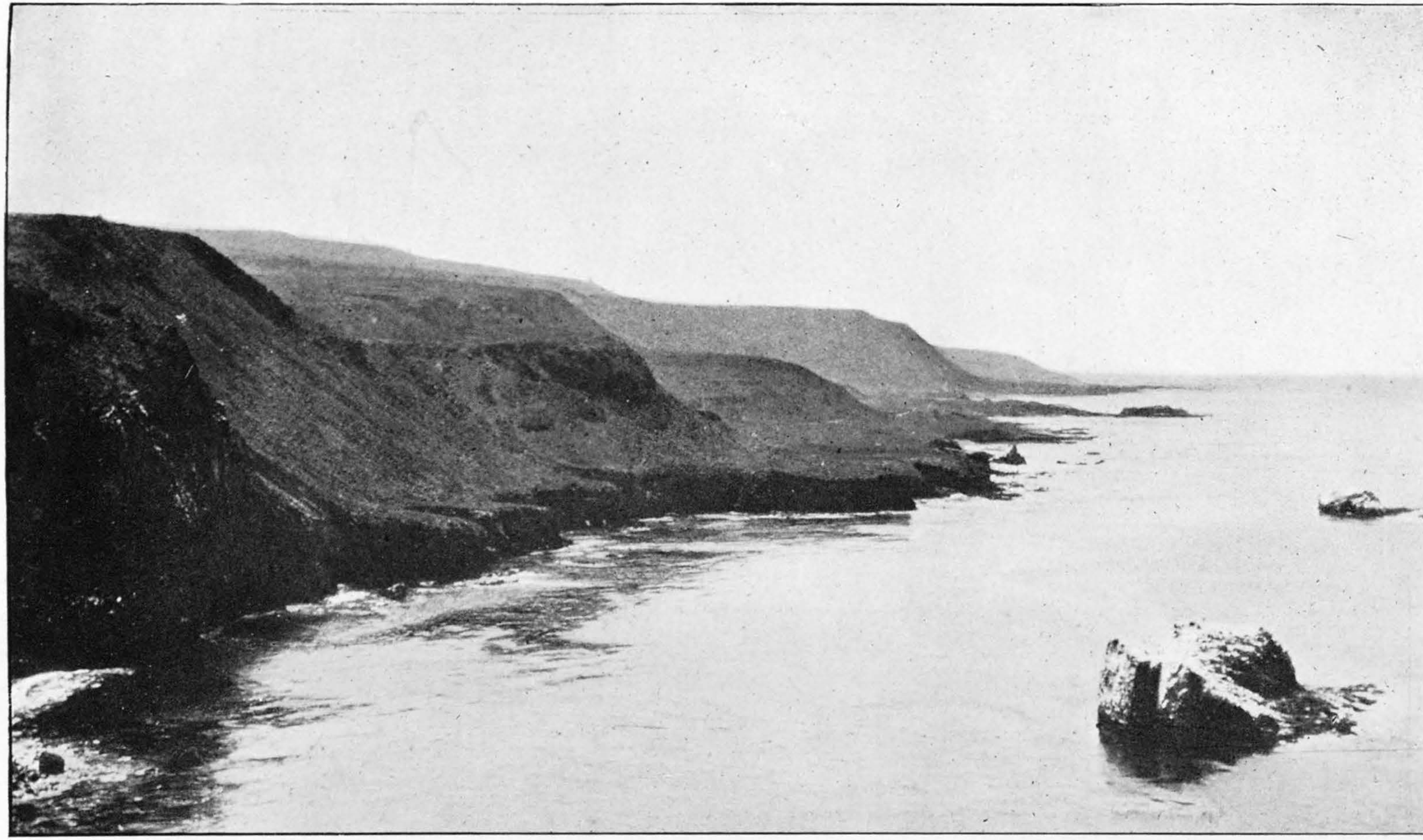
will be in the direction of E; while if the resultant of the movement is a depression, the divide will migrate toward A.

If the faulted block D′ D″ E were elevated from submarine depths to a position similar to that of San Clemente, the final result of erosion would be a divide midway between the limiting waters on the two sides.

The results here arrived at must be true in all cases, whether the movements causing faulting are slow or rapid, continuous or intermittent in their action, and small or great in amount. Variation in these factors, however, will cause a variation in the rate of the migration or in its extent. Other modifying factors are the relative positions of the divide and the line of faulting, and the dip of the fault plane.

To sum up: Where crustal movements occur, causing faulting with resultant elevation, a migration of the stream divide will follow—in the direction of the line of faulting when the fault scarp faces the divide, away from the line of faulting when it does not; or, in other words, the migration is from the axis of faulting when the faulted block includes the divide and toward the axis of faulting when it does not.

There are remnants of what appear to be the stream valleys of an older topographic cycle on parts of San Clemente, best seen near the southeastern end, where a valley is indicated by the outlines of the coast and terraces. The recent streams here are influenced to the extent of flowing approximately at right angles to the coast, rather than at



TERRACES, FROM ABOVE SEAL HARBOR, SAN CLEMENTE ISLAND; LOOKING SOUTHEAST.

right angles to the island's crest. They thus tend somewhat, about Smugglers Cove, to converge toward the cove. This broad, valleylike form is also distinctly seen in the submarine contours at this end of the island, as far out as soundings are given (for a distance of several miles), and there can be no doubt that it extends much farther. There is evidence of considerable folding at the time of the island's formation or later, and it is possible that this valley is structural and not due to the forces of erosion.

### SHORE FEATURES.

The shore features of San Clemente may be considered under the heads of beaches, bays, and cliffs.

There are few sandy beaches. They are found on both sides of the isthmus near the northwestern end of the island, also about China Point and at Smugglers Cove. The sand at the latter place is noticeably free from shell fragments and contains abundant grains of green pyroxene. Beaches of coarse shingle are found in many of the coves along the shore and for short stretches at the base of the cliffs, on the southern coast particularly. At the extreme northwestern point of the island there is no cliff, but a low platform, little above high water, and along the margin of this the waves have built up a barrier of coarse shingle 5 or 10 feet high.

Including all sorts of coastal indentations under the general term "bay," we find this feature of the shore line but little developed on San Clemente. There are many minor irregularities along the shore which to the eye appear as small bays, but which scarcely indent the outline of the large map. Even the most prominent openings will scarcely afford more than an anchorage to vessels, and there are none which will furnish protection in all kinds of weather.

The cliffs are the most prominent of the shore features, and almost completely surround the island. On the northern side they naturally reach their greatest development, and for most of the length of this coast they extend from the water line to the upper plateaulike portions of the island. Toward the northwestern end, however, there is a low terrace for some distance on this side and a correspondingly low cliff. The cliffs on the southern side of the island are comparatively low, averaging perhaps 25 feet in height. At only one or two places do they reach any considerable altitude, the highest point along the shore on this side being at Seal Harbor, where the cliffs rise to a height of 300 or 400 feet.

### TERRACES.

The terraces form the most marked feature of the topography, occurring on both sides of the block, though only occasional and slightly developed on the northern side. On the southern side they are remarkable for their size, continuity, and distinctness. Although terraces are found along the greater part of the California coast, they are nowhere so well developed as on San Clemente, a fact due in large part to the especially favorable conditions for such development found in the low angle of the slope of the crust-block on the side where the waves are strongest. Professor Lawson has discussed these terraces in considerable detail in his Post-Pliocene Diastrophism of the Coast of Southern California,[1] and there is little to be added to his description.

Clear-cut and distinct terraces may be seen up to an altitude of about 1,320 feet; above this, to about 1,500 feet, the island still shows pronounced terracing, but the individual terraces are not so distinct as those below that level. The cliffs between these upper terraces are only a few feet in height, and through the influence of erosion the angles at the base and top of the cliffs are so rounded and grade so gently into the slopes above and below that it is almost impossible to make out the exact point at which cliff and terrace meet, and all the readings are doubtful. The low cliffs of these terraces are due partly to their position on a surface of very gentle grade and partly to the slow rate of the island's elevation at the time of their formation. The terraces as then formed, though of considerable depth from front to rear, had no great vertical interval between them. Their cliffs, therefore, have had ample time since the island's emergence to be almost wholly worn down by erosion. The breadth of these terraces, together with their low cliffs, indicates that the ocean remained for a long time at the levels between 1,320 and 1,500 feet.

Above 1,500 feet no terracing exists, so far as could be determined. There is a faint notching of one of the ridges near the summit of the island which may be due to terracing, but the evidence is extremely doubtful. It is of course possible that the island has been terraced to the very top; but if so, either these older terraces must have been less developed than those below 1,500 feet, so that they have yielded more readily to the action of erosion, or else those above 1,500 feet have been much longer exposed to erosion than those immediately below this level.

Most, if not all, of the terraces below 1,320 feet are of sufficient size and development at some point on the island to be easily distinguished on the large-scale contoured map with contour intervals of 40 feet. These terraces may frequently be traced for miles, giving us the outlines of the older shores, and so the form of the island at different stages of its rise. These lines are in all cases roughly parallel to the present coast. Many of the shore lines, however, remain only in fragments. As the island emerged, step by step, the newly formed coast frequently encroached upon its predecessor. The present coast is doing this now, so that if it is given sufficient time all traces of the preceding terrace will have been removed, and the ocean will be attacking the cliffs below the higher terraces. In the same way some of the earlier terraces have become mere remnants, and others may have been lost entirely. The original number was without doubt greater than that which may be found to-day.

When at any point the cliff cutting has been so extensive as to remove one or more of the preceding terraces, the resulting cliff will be corre-

[1] Bull. Dept. Geol. Univ. Cal., Vol. I, No. 4, December, 1893, pp. 135-139.

VIEW FROM FRONT EDGE OF 480-FOOT TERRACE, ABOUT 2½ MILES SOUTHEAST OF SEAL HARBOR, SAN CLEMENTE ISLAND; LOOKING SOUTHEAST.

spondingly higher. The best example of this is seen at about the middle of the southern side of the island, where at one point, at the back of the 60-foot terrace, the cliff is some 600 feet in height. This cliff extends for some distance at nearly the same altitude, and forms the highest terrace cliff found on the island. It is in this region—between Seal Harbor and China Point—that the terraces reach their greatest development. Their cliffs range from the height just mentioned down to a few feet, while the terrace surface, from front to rear, varies in width from a few feet to more than half a mile. The average vertical interval between the recognized terraces is somewhat less than 100 feet.

The following table shows the altitudes of the principal terraces:

*Altitudes of principal terraces.*

| 1. | 2. | 3. |
|---|---|---|
| *Feet.* | *Feet.* | *Feet.* |
| 1,500 | ........ | ........ |
| 1,375 | 1,365 | ........ |
| ........ | 1,325 | 1,320 |
| ........ | 1,265 | 1,280 |
| 1,250 | 1,200–1,220 | 1,240 |
| ........ | 1,135–1,150 | ........ |
| 1,040 | 1,085–1,100 | 1,040–1,080 |
| 960 | 975 ? | 960 |
| 930 | 900 | ........ |
| ........ | 845–860 | ........ |
| 785 | 775–795 | 760–800 |
| ........ | 710–725 | ........ |
| ........ | 655 | 640–680 |
| 580 | 575 | 560 |
| 550 | 510–530 | ........ |
| 470 | ........ | 480 |
| ........ | 455 | 440 |
| ........ | 410–420 | ........ |
| 380 | 360–375 | 360–400 |
| 325 | 300–305 | 280–320 |
| 280 | 255 ? | ........ |
| 225 | 200–220 | 240 |
| 170 | 155 | 120–160 |
| 120 | ........ | ........ |
| 80 | 60 | 40–80 |
| 40 | ........ | ........ |
| 12 | 12 | ........ |

1. "Altitudes of terraces, as determined by aneroid on trail from Wilsons Cove to the Summit." (Lawson, Post-Pliocene Diastrophism, p. 131.)

2. Altitudes of terraces, as determined by aneroid by the writer.

3. Altitudes of terraces on the southwestern slope of the island, as read from Coast and Geodetic Survey manuscript map, to the nearest 40-foot contour.

There are probably other minor terraces too indistinct to warrant determination. Two or three doubtful cases may possibly have crept into the table. It is also possible that several of the terraces given in the second column are duplicated, owing to variations in the barometer, as the readings were made on different days. Owing to the fact that the best-developed terraces are cut by the largest and most impassable canyons, they can be followed only for short distances. Detritus from the cliffs, and in a few instances large alluvial fans, cause the terraces to grade into the cliffs behind them, so that the exact height of the old shore line is rendered difficult of determination.

The figures in the last column are approximate at best. In several instances it is by no means certain which one of two or three contours is nearest the correct altitude of the terrace. The readings as given appear to represent distinct terraces, though there may be a duplication in the 440- and 480-foot readings. The chief difference between this list and the corresponding one given by Professor Lawson[1] lies in the fact that in several instances where two of his readings appear to refer to the same terrace they have been combined.

That these are of the nature of true wave-cut terraces there can be no doubt. Even if there were no more positive evidence, their parallelism to the present shore line would be suggestive, at the least. The scarcity on this island of the rolled material usually found in connection with such terraces has been noted,[2] but the failure to find it appears to be due only to its concealment by later detritus. Shore conglomerate was found by the writer at a number of points in the angle between the terrace and its cliff, or rarely spread over the narrow terrace itself, and covered in most instances by 5 feet or more of detritus from the cliffs above. At one point numerous bivalves (*Lucina californica* Conrad) and rolled pebbles were found resting on the upturned and beveled edges of Miocene deposits, the whole covered to a depth of about 5 feet with wash from the hill slopes. Shell borings were not seen in connection with the terraces, except at one point where they occur in the terraced surface of the Miocene shale at an elevation of a little over 300 feet.

Further evidence of the character of the terraces is found in their seaward slope and the level line at the rear representing the old shore line. Projecting points of rock on the terraced surface, somewhat removed from the cliff at the rear, were seen at a number of points on the island, both on the higher and on the lower terraces, corresponding in position to the stacks which mark the present coast. At the northwestern extremity of the island circumstances seem to favor the formation of stacks. A series of them is shown on the map off the present coast, and, likewise, they are especially numerous on the low platform which marks this end of the island. An unusually large "stack" is seen near the center of Pl. LXXXVI.

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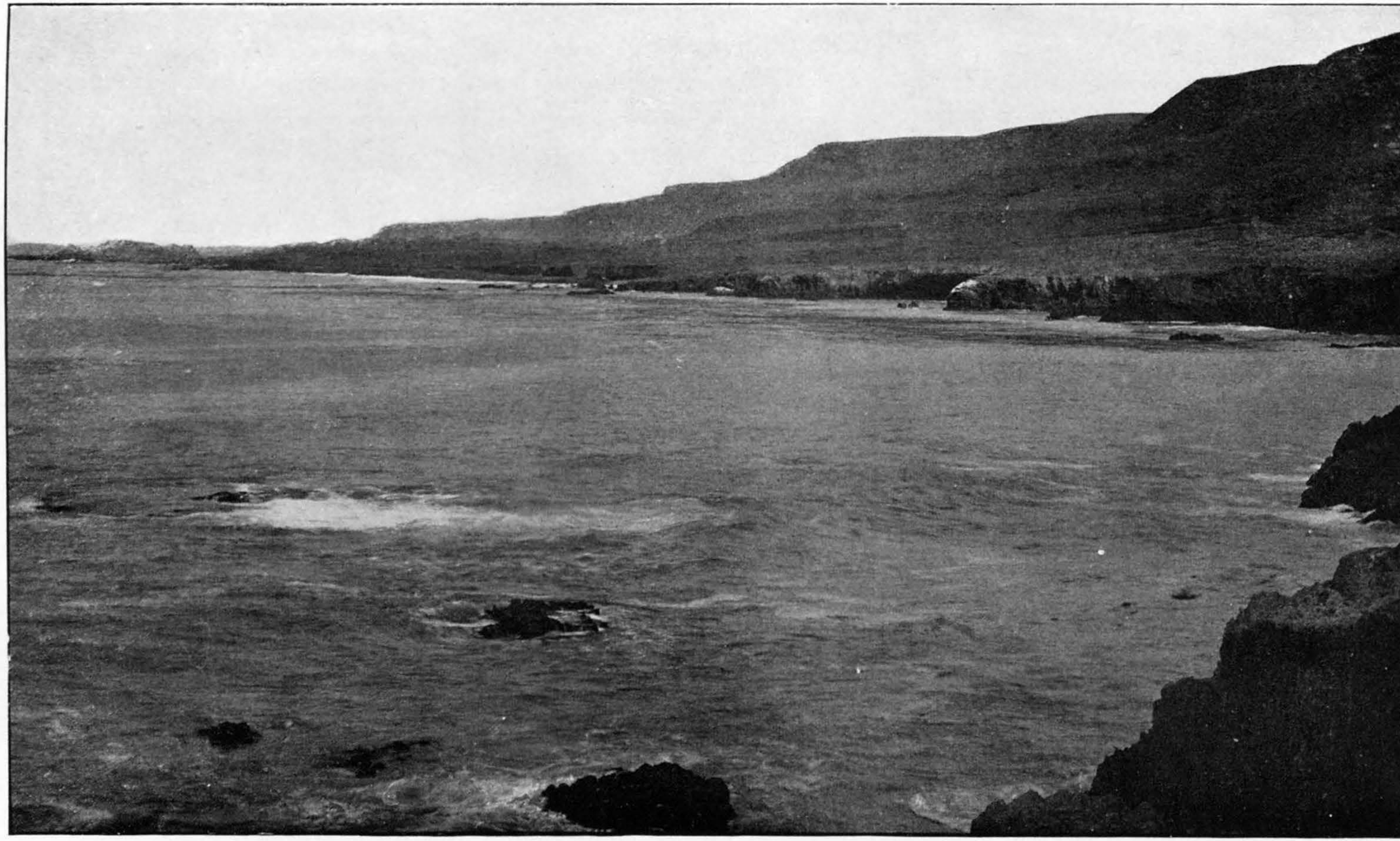
[1] Post-Pliocene diastrophism, p. 131.

[2] Lawson, loc. cit., p. 130; also Cooper, Geol. Surv. of Cal., Vol. I, 1865, pp. 182–186.



TERRACES, SOUTHWESTERN COAST OF SAN CLEMENTE ISLAND, FROM ABOUT 2½ MILES SOUTHEAST OF SEAL HARBOR; LOOKING NORTHWEST.

Caves are occasionally seen in the cliffs back of these terraces, and they were thought by Cooper[1] to have been formed by wave action—doubtless because of their general similarity to caverns formed by the waves along the present shore. They do not appear, however, to have been thus formed, for the cliffs have, as a rule, suffered considerably from erosion since their elevation, so that if such caves had once existed they would now be partly or wholly obliterated by the wearing away of the cliff; further, the caves occur frequently in the larger stream canyons, and therefore out of reach of wave action. On the other hand, the caves do not seem, in general, to be a part of the original structure of the rocks,[2] though a few small ones were seen which doubtless are of this character. The majority of them, however, seem to be formed by a sort of undermining process through the agency of percolating waters. The caves generally occur in the fragmental volcanic rocks, and where these are found in abundance, as toward the southeastern end of the island, caves are numerous. Here they occur both on the face of the terrace cliffs and in the cliffs of the stream canyons, and in many places they are so numerous that the cliff at a distance looks as if honeycombed by the small caverns. The shape of the caves is usually very irregular.

As no direct information could be obtained as to the part played in their formation by running water, it could only be inferred from their position, form, and the character of their surfaces. Possibly many or most of the caves originated through the sliding of water-soaked portions of the rocks forming the cliff; but to whatever forces of erosion the beginnings are due, the process is undoubtedly continued on the sides of the cave and overhead by a gradual loosening and falling away of the fragments of which the pyroclastic rocks are made up. The finer fragments between the coarser blocks fall first, leaving these large angular pieces projecting from the surface, many of them appearing as if on the point of falling. A further proof of this origin for most of the caves is that an exactly similar cave of considerable size—large enough to furnish shelter for perhaps fifty persons—was found on the southern side of the island in a coarse shore conglomerate, and the larger fragments of the conglomerate projected in the same way from the sides and top of the cavern.

## THE ROCKS OF SAN CLEMENTE.

San Clemente is built up almost entirely of lava flows, with intercalated volcanic breccias and ash deposits. A small proportion of sedimentary deposits, Miocene and later, make up the remainder. These deposits are seen about China Point, Smugglers Cove, and toward the northwestern part of the island, occurring only as a thin covering of the rocks beneath.

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[1] Loc. cit.

[2] Lawson, loc. cit., p. 133.

The volcanics have a general dip at a low angle in the direction of the island's tilting, though for most of the northwestern third of the island the structure is in part anticlinal, the dip being to either side from the main ridge, and on the north averaging about 20°. The volcanics and the Miocene sedimentaries are in part gently folded in the direction of the island's length.

The sedimentaries and the fragmental volcanic rocks are in places impregnated with sodium chloride, even at a considerable altitude above the ocean. This is true of these rocks not only at or near the surface but also in the bottoms of the largest canyons, several hundred feet above sea level. So great is the amount of contained salt, that many of the caves at the bottom of the largest canyons are coated with an efflorescence of salt in thread-like fibers.

## ERUPTIVES.

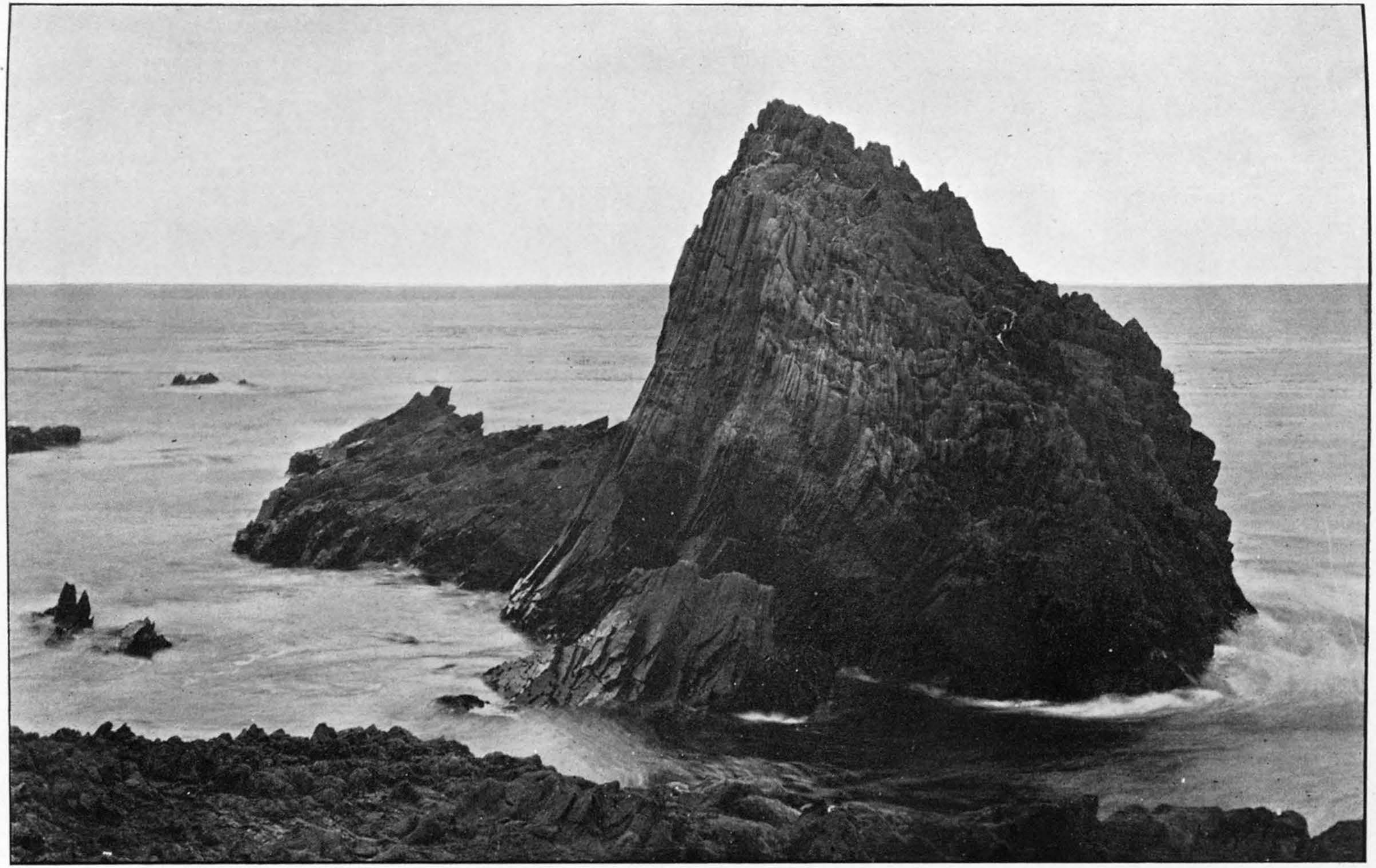
The eruptive rocks of which San Clemente is so largely formed occur mainly as flows. With but two or three exceptions they are normal pyroxene-andesites, ranging in chemical composition from basic to acid. The exceptions to this grouping consist of the youngest volcanic rocks of the island, which have been classed in one case with the dacites and in another with the rhyolites, though closely related to the andesites in the character of their phenocrysts. The silica content in these younger rocks is 67 and 70 per cent respectively.

### ANDESITE.

The most pronounced feature presented by the andesites in the field is the jointed structure which many of the flows have assumed on cooling. This structure is not wholly characteristic of individual flows, but it may be assumed by any of them, and the same flow at one time may be jointed and at another may show no jointing at all, depending on the rapidity of its cooling. In the case of one flow, resting on a breccia, the rock is distinctly jointed to within two or three inches of its lower contact. Between this point and the contact the rock changes in color from a dark gray to black, and in structure it is no longer jointed, but very vesicular. The jointed structure causes the rock to break into plates or lenses of varying thickness, usually between one and three centimeters. A characteristic view of this jointing is shown in Plate XCI. This jointing, so common to the rocks of this island, is not developed to any noticeable extent in the same series of rocks on Santa Catalina.

Many of the flows occur massive, with no definite planes of fracture, and usually more or less vesicular. The rock varies from vesicular to scoriaceous. A local development of a frothy and glassy andesite occurs at China Point, the cavities of the volcanic rock being completely filled with a yellowish-gray limestone.

The flows range in thickness from a few feet up to 50 feet or more.



CHARACTERISTIC JOINTING OF THE ANDESITE, SOUTHWESTERN COAST OF SAN CLEMENTE ISLAND.

In texture they are fine-grained and compact, or coarser-grained, with a rough, uneven fracture, the grain depending on the relative abundance of the phenocrysts and the development of the minerals of the groundmass. The rocks with the coarsest grain are naturally the jointed rocks, while the more vesicular rocks are fine-grained. An occasional alteration of the iron oxide in patches through the rock causes it to present a mottling in black and deep red. The rock ordinarily varies in color from a rather light gray to black, the lighter colors, as a rule, accompanying the coarser crystallizations. Occasionally the rock is red. The andesites are usually fresh, and in general, where any alteration has begun, the hypersthene is the only mineral which has suffered. The andesites are in all cases porphyritic, though in a few instances the more compact rocks show only a small number of scattered phenocrysts.

The essential minerals of these rocks, as shown by the microscope, are a medium basic plagioclase feldspar, augite, or hypersthene, or both, and magnetite. All these minerals occur as phenocrysts, and usually in the groundmass also. The groundmass ranges from nearly holocrystalline to almost wholly glassy. Iddingsite[1] occurs in one specimen, obtained at the base of the cliff about a mile and a half southeast of Wilsons Cove. Some of these andesites are indistinguishable microscopically from those of Santa Catalina.[2] Where there is a difference, it is chiefly due to the fact that the San Clemente andesites are frequently much coarser-grained than those studied from Santa Catalina, and the hypersthene of many of the former rocks presents somewhat different properties.

The feldspars range from andesine to an acid labradorite ($Ab_1\ An_1$). They present excellent crystal boundaries in some instances, though the majority of the sections are more or less resorbed. Zoning is not usually pronounced in these rocks, but it is not infrequent, and is shown not only by the different layers of feldspar substance growing progressively more acid outward, but also by the inclusions. The feldspars contain, as inclusions, rarely magnetite grains, small and rounded grains of pyroxene, more or less glass or inclosed groundmass, and numerous dust-like particles. In several instances the outer borders of some of the feldspars have a micropœcilitic structure, and contain abundant minute pyroxene grains, somewhat smaller than those of the groundmass, thus showing a growth of the feldspar up to a point not long preceding the consolidation of the groundmass. The inclosures of glass or portions of the groundmass are usually very irregular in shape. Occasionally inclusions of glass are zonally arranged, rarely in the form of negative crystals. Rarely glass fills the larger part of the crystal, and in one slide areas of glass are seen, either nearly unbroken or as a coarse network, with a narrow border of feldspar

[1] The geology of Carmelo Bay, by Andrew C. Lawson: Bull. Dept. Geol. Univ. Cal., May, 1893, Vol. I (1893–1896), pp. 31–36.

[2] The geology of Santa Catalina Island, by W. S. T. Smith: Proc. California Acad. Sci., 3d series, Geology, Vol. I, No. 1, February, 1897, pp. 30–41.

substance, and a few small fragments of feldspar through the glass, all the feldspar having a uniform extinction. The dust like particles included in the feldspars occur in patches or zones, and usually such areas show by their extinction a slightly different feldspar substance from the adjoining clear spaces. This dust is largely indeterminable, but some of it, as seen with high powers, consists of liquid inclusions. In general, the feldspar is quite free from alteration products of any sort.

The pyroxenes are usually much less abundant than the feldspar phenocrysts, and considerably smaller. Both augite and hypersthene are present in most of the slides, but the relative amount of each varies considerably. The hypersthene frequently presents excellent crystal forms, the best, as a rule, being in the smaller phenocrysts. Both minerals have a prismatic habit, though in cross section they frequently show an elongation parallel to the $\bar{b}$ axis. The cleavage is usually (though not always) pronounced. Liquid and small glass inclusions are found in both minerals, occasionally as negative crystals. Twinning is common in the augite, in some slides, and is rarely polysynthetic. The two minerals, in a few instances, are intergrown in parallel position, and in one case the intergrowth is quite intricate. Usually, however, the hypersthene appears to have crystallized a little earlier than the augite. Both minerals, in a number of instances, show a period of growth following a period of considerable resorption, for when they occur as inclusions in the feldspar these inclusions are always small, rounded grains, never as large as the phenocrysts, and in only one instance showing any indication of crystal form. In one slide several of the hypersthenes are rounded where in contact with the feldspar in which they are partly included, while the portion in the groundmass presents complete crystal form. In another instance a twinned augite is surrounded by an untwinned border having an extinction between the extinctions of the two halves of the twin. That a part of this marginal growth occurred shortly before the consolidation of the groundmass is seen in two slides. In one a section of augite has near its outer border three small included feldspars closely resembling those of the groundmass; in the other, several sections of augite have an outer border which can be distinguished from the groundmass only in polarized light, having in ordinary light the same deep yellow color as the glassy groundmass.

In color the augite is very pale green. The color of the hypersthene differs in different slides. In some it resembles the augite very closely, and is but slightly pleochroic, like that of the Catalina andesites. In other specimens the hypersthene has a deeper color and shows a pronounced pleochroism. In these $\mathfrak{c}$ = dirty grayish green, $\mathfrak{b}$ = reddish yellow, $\mathfrak{a}$ = pale brownish red. The absorption formula is $\mathfrak{c} > \mathfrak{b} > \mathfrak{a}$. In general the more highly colored hypersthene tends to alter very readily, while that which is most like the augite is usually quite fresh.

The alteration is into fibrous serpentine, generally well charged with a yellow to deep-brown coloring of iron oxide in the portions first altered, and takes place, as a rule, along the borders and cracks. In one slide alteration of the hypersthene has given rise to a yellowish aggregate characteristic of the alteration of this mineral in the rhyolite, and described more fully under that head (p. 487). In a number of slides many of the hypersthenes are bordered on the sides by augite, occurring usually as narrow rods, but in several instances as a series of small grains. The rods extend along those portions of the hypersthene parallel to the $\dot{c}$ axis, and where the hypersthene presents an unbroken line the augite frequently forms a single rod along its border, in two or three instances even projecting beyond the end of the hypersthene into the groundmass. The width of these rods averages 0.01 mm. They are doubtless crystallizations belonging to the groundmass, their vertical axes oriented to accord with the orientation of the hypersthene.

Magnetite usually occurs as small grains and octahedra, and most of the slides are well peppered with it.

Most of the iddingsite mentioned as occurring in one of the slides is of a bright-yellow color and is associated with a varying amount of serpentine. It is quite irregular in form and occurs as small, scattered phenocrysts. The other phenocrysts in this rock are principally labradorite and augite. The rock yields on analysis 55.38 per cent of silica.

The groundmass is usually medium-grained, hypocrystalline, less often hyalopilitic. Flow structure is not common, though pronounced in some cases. In the rocks of coarse-grained groundmass the structure approaches holocrystalline. Feldspar occurs as laths, and is usually oligoclase. The pyroxene is largely augite, occurring in flakes and irregular prismatic forms. The magnetite of the groundmass is usually found as a multitude of microscopic dots in the glassy base. Occasionally these are so abundant as to render the groundmass nearly or quite opaque. The glass in the groundmass varies greatly in amount, in parts forming only a small interstitial residue, but occasionally constituting the bulk of the entire slide. In color it is usually a dirty gray-brown, varying to almost black from contained magnetite. Where the iron occurs in another state of oxidation the groundmass is almost brick-red in color. In one slide the groundmass was a clear but very deep yellow. The vesicular cavities seen in the groundmass of many of the slides show no compression and are only occasionally lined with a thin film of pale-greenish secondary products.

The frothy rock from China Point is seen under the microscope to consist almost wholly of a pale brownish-yellow glass, in which occur very few scattered sections of labradorite, principally as small laths. The cavities of the rock, as already stated, are largely filled with calcite. This deposit, as shown by qualitative tests, contains some magnesia, and in portions of the rock the part soluble in acid gives an abundant precipitate of iron and alumina. This product in its general appearance much resembles the ordinary Miocene limestone.

A little over 3 miles to the northwest of China Point a dike-like mass of andesite about 2 feet wide is seen extending from the shore for a short distance out into the water. The rock is jointed transversely to the direction of the dike, which thus presents the appearance of a pile of wood. The rock is dark-gray in color, and contains minute, scattered vesicles. It shows few phenocrysts under the microscope, and most of these are labradorite. The groundmass is quite coarse-grained, consisting of feldspar laths and scattered grains of pyroxene, with a considerable amount of blackish glass. Flow structure is quite pronounced.

There are associated with the andesites at least two narrow beds of ash, containing more or less coarse, fragmentary material. One of these beds occurs in the upper part of the andesite series and perhaps 100 feet below the rhyolite. This ash is dark-gray in color, somewhat reddened at one point, at its contact with the overlying vesicular lava. The other bed of ash, seen in Red Canyon, has a thickness of from 1 to 3 feet, is of a light-gray color, and rather compact. It occurs finely bedded. Under the microscope these rocks are seen to be made up largely of grains of a pale brownish-yellow to deep-brown glass, with which are associated numerous mineral fragments, almost wholly feldspar, besides more or less blackish or dark-colored, glassy to partly crystalline volcanic fragments.

The volcanic breccias mentioned as occurring with the andesites consist of very coarse material held together by a moderately fine-grained cement. The fragments thus contained are sometimes many feet in diameter. These breccias constitute a considerable proportion of the volcanic rocks of the island, and reach their greatest development toward the southeastern end. They are yellowish, reddish, or nearly black in color, the red predominating. The cementing material of the rock is usually firm and compact. Microscopically it is made up of volcanic fragments in a usually glassy matrix. Many of these included fragments are principally of glass. The matrix contains, besides glass, angular mineral fragments, chiefly feldspathic.

### DACITE.

The dacite, wherever it was found, was the upper flow or flows of the entire volcanic series, with the exception of the rhyolite. This last is without doubt the youngest volcanic rock on the island, although it was not seen with the dacite at any point, so that their relation could be determined only indirectly. The main occurrence of the dacite is near the center of the island, extending in the direction of the island's length from the large sedimentary area to a point not far from the highest part of the island. This area reaches a maximum width at Seal Harbor, where it comes down to the water's edge, its northern limit being beyond the crest on the northern side of the island. Besides this main area there is also a small occurrence of the rock at the extreme southeastern end of the island and another on the crest above Wilsons Cove. There are also very small patches along the summit to the southeast of the center of the island.

The color of this rock varies from a rather light purple through gray to almost black, but whatever the color, it has usually a purplish tinge, particularly if it is at all weathered. It weathers through purple and yellow to a dirty white, which yields on disintegration a very fine and very light-colored soil. The rock occurs in many places in a jointed condition. It is frequently very fine grained and compact. In parts of the main area the rock shows vesicular structure, the vesicles, however, usually completely flattened in the direction of flow. This facies tends somewhat to break into irregular plates along the flattened vesicular surfaces.

In most of the specimens there are few phenocrysts, and these generally small. They are largely plagioclase feldspar, together with more or less augite, hypersthene, and magnetite. The feldspar is andesine, and its characters are in general the same as those given for the andesites. It occasionally tends to more angular forms, and the dusty areas are more common, for the most part zonally arranged. The pyroxenes are usually smaller than the feldspars, and in amount the hypersthene, as a rule, greatly exceeds the augite. The augite usually occurs in very irregular forms, and is always found fresh. The hypersthene is of the variety which is readily altered, and in nearly all cases the sections of this mineral are more or less serpentinized. It is generally considerably resorbed, though in a few instances it shows good boundaries, either complete or partial. It is pleochroic, though occasionally the pleochroism is scarcely noticeable. The mineral is frequently surrounded by a border of varying width, heavily stained with iron. Although the alteration of the hypersthene is usually into serpentine, greenish to pale yellow in color, in a number of slides sections of hypersthene occur, wholly or partly altered into the yellowish aggregate which is particularly characteristic of the rhyolite.

The groundmass of this rock is in most cases holocrystalline, the few exceptions showing more or less interstitial glass. It is generally very fine grained and is composed almost wholly of feldspar and magnetite, the feldspar laths usually showing very pronounced flow. In a few instances pyroxene occurs as an essential mineral of the groundmass in the form of flakes and prisms, but in general it is entirely absent. Quartz occurs in the groundmass in connection with some of the structures to be described.

The magnetite, sometimes altered in part to hematite, occurs as dust-like particles, minute grains, needles or arborescent growths, and frequently shows a more or less parallel arrangement. It is well scattered through the groundmass, and frequently appears to form a complete network.

The feldspar occurs both as laths and rectangles, principally of oligoclase, and as allotriomorphic microgranules. The proportion of laths to the granules is variable. Besides its occurrence in this form, the feldspar also appears in pronounced patches of varying size. These patches, when sufficiently large, have generally a pronounced micropœcilitic structure, which is characteristic of most of the slides of this rock, and which forms the most marked feature of the rock as a whole, distinguishing it from all other rocks of the island. Both the smaller and the larger patches have, in general, extremely ragged borders. Where two patches come in contact they are allotriomorphic, sometimes with their borders intergrown. In a few instances there is an intricate intergrowth of these areas in the groundmass, with development of micrographic structure.

The patches occur as scattered areas, sometimes few in number, but occasionally filling the larger part of the groundmass. Though at times developed as small flakes, these areas are generally comparable in size to the feldspar phenocrysts. These areas and their structure are not distinguishable in ordinary light, except occasionally by their comparative freedom from the network of magnetite microlites. Under crossed nicols, however, the uniform extinction throughout these patches at once calls attention to them, besides the fact that their polarization is usually much sharper and brighter than that of the rest of the groundmass. The microgranular feldspar of the latter, in particular, generally shows a feeble polarization. The appearance of these patches is frequently that of a network of feldspar laths (shown in Pl. XCII, *A*). The chief difference between them and such an assemblage of simple laths is the uniform extinction of the entire area. In many cases, however, the parallel arrangement of the feldspar laths of the groundmass gives these laths a nearly uniform extinction over a considerable area, and in some instances closely simulates the phenomenon before mentioned. The very smallest patches do not form such a lath-like network, but more nearly resemble a very irregularly bounded simple grain of feldspar. Nor do these smallest areas have the pronounced micropœcilitic character seen in most of the larger sections, the reason probably being that their small size prevents their containing an appreciable amount of the other minerals of the groundmass, except, perhaps, magnetite.

On extinction the larger areas show their micropœcilitic character, not only by their containing considerable magnetite, but by inclusions of feldspar laths and granules without uniform extinction. This structure is brought out only on close inspection, on account of the minute size of many of the feldspars of the groundmass and their feeble extinction in many instances.

The areas appear to be largely of orthoclase, with a moderate amount of associated quartz. In one slide, in which the structure described is best developed (see Pl. XCII, *B*), the areas form in part an allotrio-

A

B

A. CHARACTER OF MICROPŒCILITIC PATCHES IN THAT FACIES OF THE DACITE IN WHICH THEY ARE COMPARATIVELY FEW. × 39.

B. FACIES OF THE DACITE IN WHICH THE GROUNDMASS IS MADE UP ALMOST WHOLLY OF MICROPŒCILITIC PATCHES. × 39.

The dark phenocryst near the center of B is hypersthene; the light one, feldspar

morphic aggregate in the groundmass, several of the sections showing simple twinning. In two of the slides there are one or two vein-like lines of allotriomorphic grains of both clear quartz and feldspar, together with several grains showing the micropœcilitic structure. A biaxial figure was obtained from a number of these patches. Although having in many of the slides such a large areal distribution in the groundmass, the actual amount of these micropœcilitic portions is probably small, forming simply a residual crystallization about the other elements of the groundmass.

In the majority of the occurrences of micropœcilitic structure thus far described, the host has been quartz and the included minerals feldspar, and in two cases,[1] at least, these structures had centers of clear quartz. The occurrence of micropœcilitic quartz in these two instances has been attributed to the orienting influence of the previously crystallized quartz centers. In the rocks from San Clemente the matrix consists of both quartz and orthoclase, the inclusions being principally of the previously crystallized feldspars of the groundmass. The cause of the uniform orientation of the parts of the matrix here can not be any previous crystallization, but it rather appears probable that the flow of the rock had some influence, at least, in bringing about a common orientation of the molecules of the final crystallizations over considerable areas.

## RHYOLITE.

There are two main occurrences of the rhyolite, one to the southeast of Wilsons Cove, and the other near the northwestern extremity of the island. The first area begins at Wilsons Cove and extends to the southeast about 2 miles, averaging half a mile in width. The other area is a little narrower, and forms a band across the island. Besides these main areas there are at least two minor occurrences, one of which forms a small patch not far from the northern coast and about midway between the two main areas.

At several points where this rock occurs as loose bowlders, or in the form of stacks on the terraces, they have assumed, by weathering in flakes or scales from the surface, a rounded, boss-like form. A similar weathering was seen in a very few instances in connection with the other volcanic rocks, and it is probably due to alternations of heat and cold. The daily range of temperature on the island during the summer months is sometimes considerable, and would be sufficient to account for such effects.[2] A more characteristic form of weathering in the rhyolite is seen along the water's edge a little to the southeast of Wilsons Cove. All along the northern border of this area of the rock there is a considerable northeasterly dip at an angle averaging about 20°. The rocks here are coarsely jointed, and in weathering large blocks

[1] The volcanics of the Michigamme district of Michigan, by J. Morgan Clements: Jour. Geol., Vol. III, No. 7, Oct.-Nov., 1895, pp. 811-817. The geology of Santa Catalina Island, by W. S. T. Smith: Proc. California Acad. Sci., 3d series, Geology, Vol. I, No. 1, February, 1897, p. 24.

[2] Cf. G. P. Merrill, Principles of rock weathering: Jour. Geol., Vol. IV, No. 6, Sept.-Oct., 1896, p. 713.

break off and fall to the base of the cliffs. For a considerable distance those blocks which are within reach of the waves' action have been eaten into on the sides, as shown in Pl. XCIII. The peculiar and irregular chiselings which have been developed here occur only on the sides of the block-like masses and in the direction of flow. This mode of weathering at this point is doubtless due to structural variations in the rock.

The color of the hand specimens of this rock varies from a yellowish or grayish white to almost black. Generally the rock is quite light in color, with more or less of a reddish tinge. The fracture is usually rough and uneven. Macroscopically the rock is composed of a variable amount of phenocrysts, principally of clear glassy feldspar, embedded in a compact matrix, dark reddish-brown to blackish in color. Occasional banding is seen. In many of the specimens yellowish decomposition products of some of the phenocrysts are scattered in spots over the surface, and, at one point at least, the rock is yellowish in color from disseminated grains of this same product. Considerable pyrite in the form of minute grains was seen in a number of the specimens. At several points the rock is very light in color and appears to be filled with a honeycomb of black compact material, having much the appearance of narrow veins. Under the microscope there is no sharp line of demarcation between the light and the dark portions, and the fact that the light part contains minute yellow secondary grains, from which the other is free, would point to weathering as a cause of the structure.

Microscopically the rhyolite is seen to contain numerous feldspar phenocrysts, with fewer and scattered phenocrysts of hypersthene and magnetite, and occasionally augite and pyrite. The groundmass is generally microgranular in part and in part glassy, showing pronounced flow in nearly all cases.

The feldspar phenocrysts seldom show complete crystal boundaries, and they occur, as a rule, either in angular broken forms or rounded and considerably resorbed. Zoning is very common, but the zones do not wholly agree in outline with the boundary of the section in the angular forms. That many of the sections have been under strain is plainly shown by secondary twinning lamellæ. The feldspars are quite fresh. Their principal inclusions are dust-like areas like those of the andesite and occasional grains of magnetite. A number of feldspar fragments were removed from the crushed rock, and their specific gravity was determined in Klein's solution by the Westphal balance. The range in gravity was from about 2.40 to 2.66, but the majority of the grains were in suspension between 2.65 and 2.66, which corresponds to andesine. This agrees with the determination of the feldspars by means of symmetrical extinction angles to either side of the albite twinning lamellæ. Quartz is occasionally seen in the slides, much cracked and presenting only corroded forms. The scattered magnetite grains have also suffered considerable corrosion. Pyroxene is never very abundant, and one of the slides is entirely without ferromagnesian minerals of

WEATHERING OF RHYOLITE, SHORE SOUTHEAST OF WILSONS COVE, SAN CLEMENTE ISLAND.

any sort. The pyroxene is principally hypersthene; augite occurs as an accessory, and is not found at all in many of the slides. The hypersthene occurs with the habit already described, has a pronounced pleochroism, and is very readily altered, being wholly replaced by secondary products in a majority of the slides. The alteration of the hypersthene is into a very finely granular aggregate, frequently stained with iron. These aggregates are chiefly of two elements—pleochroic epidote, usually deep lemon-yellow in color, and serpentine, with no pronounced color, but a faint yellowish tinge. The serpentine appears to be the chief constituent and to form a sort of matrix in which the epidote, in the form of micro-grains, is embedded. These grains seem readily to undergo solution and redeposition, for the sections show that the amount of this material in the serpentine is variable, and it is common to find it scattered through the groundmass of some of the slides, occasionally as fine veins or lines extending from the phenocrysts out into the groundmass. In two or three of the slides a number of the sections are entirely free from the yellow grains, and nothing has been left but a clear serpentine. The groundmass is usually distinctly banded in narrow lines and lenses, the appearance being due to structural variations, but emphasized by variations in color. It consists essentially of quartz and orthoclase, with occasional laths of plagioclase and more or less glass.

The quartz and orthoclase occur usually as allotriomorphic aggregates of varying grain, in lines and lenses, separated by narrow bands of glass, which have formed "eyes" about the smaller fragmental phenocrysts. The glass in general is deep reddish-brown in color, varying to nearly black. The rest of the groundmass, though colorless, contains a great deal of microscopic dust, pretty evenly distributed. The glassy portions of the rock are filled with this microscopic dust, which from the colors in mass appears to be magnetite.

The crystalline portions of the groundmass vary considerably in grain, from cryptocrystalline in some of the bands to a more coarsely granular condition in which the individual grains attain a maximum length of about 0.15 mm. These coarser portions, though principally granular, are frequently composed in part of laths of plagioclase feldspar. In these areas the quartz and feldspar are seen in all stages of crystallization, and many of them exhibit only a feeble polarization of light. There is a constant tendency to the formation of partial minute spherulites, particularly along the borders of bands, the central portion of the band tending to crystallize as definite grains. The spherulitic rays are occasionally branched. The extinction, which is very feeble, is either radial or the entire spherulite extinguishes simultaneously. The groups are sometimes positive toward the quartz wedge and sometimes negative. The granular portions of the coarser areas are either distinct grains, frequently intergrown along the borders, or they exhibit a micropegmatitic structure.

Closely connected with this compact rhyolite are one or more sheets of pumice, of a very light yellowish to a pinkish or reddish color. The pumice underlies the large rhyolite area near Wilsons Cove. One occurrence, to the southeast of Wilsons Cove, is of variable thickness, reaching a maximum of about 20 feet, and wedging out completely some 200 yards to the northwest of its greatest thickness. In the cut back of Wilsons Cove pumice is also seen, slightly different in color and microscopic character, and possibly forming a different bed. The first-named occurrence is in part free from foreign material, but at one point it contains a great deal of fragmental andesite, the size of the fragments ranging up to a diameter of about 2 feet. It also contains small included fragments of a light-colored pumice, much like that forming the second occurrence.

These rocks under the microscope are seen to be composed very largely of glass, in which are scattered fragmental and corroded feldspars, hypersthene or its alteration products, occasional magnetite, and rarely augite. In most of the rocks the glass is dirty reddish-brown in color, and is filled with microscopic magnetite dust. In one specimen, however, the glass is perfectly clear, and of a pale-green color. This color is not original, but is induced by the heating necessary in making the slide, which process changes the color of the rock from light yellow to a deep green, almost black.

*Analyses of eruptive rocks.*

| | 1. | 2. | 3. | 4. |
|---|---|---|---|---|
| $SiO_2$ | 70.39 | 66.85 | 59.34 | 55.38 |
| $Al_2O_3$ | 14.09 | 14.08 | .......... | } |
| $Fe_2O_3$ | .53 | } 3.06 | .......... | } 27.92 |
| $FeO$ | 2.12 | } | .......... | } |
| $CaO$ | 3.08 | 4.69 | 5.74 | 6.78 |
| $MgO$ | .62 | .91 | 4.69 | 3.91 |
| $Na_2O$ | 3.70 | 3.80 | .......... | .......... |
| $K_2O$ | 3.51 | 2.57 | .......... | .......... |
| Ignition | $H_2O$, S } 2.50 | 2.07 | .......... | 1.62 |
| | 100.54 | 98.03 | .......... | .......... |
| Sp. gr | 2.41 | 2.53 | 2.74 | 2.70 |

1. Rhyolite; Northwest Harbor. W. S. Tangier Smith, analyst.
2. Dacite; Summit east of Seal Harbor. W. O. Smith, analyst.
3. Partial analysis of andesite; Middle Ranch Canyon, altitude 620 feet. W. O. Smith, analyst.
4. Partial analysis of andesite containing iddingsite; shore, 2 miles southeast of Wilsons Cove. W. S. Tangier Smith, analyst.

### GABBRO.

Besides the volcanic rocks already described as occurring on San Clemente, there were found at Wilsons Cove a number of pebbles of gabbro, ranging in diameter up to 2 or 3 inches. These pebbles are found on the present beach and in the terrace conglomerate seen in the neighboring stream-gulch at an altitude of about 50 feet. They are not abundant at either point; on the contrary, they are quite rare, and it is only by careful search that they can be found. Their source is not known, for gabbros were not found in place by the writer, although they may exist at some point not reached by him, as at the base of the cliffs near the middle of the northern coast.

## SEDIMENTARY DEPOSITS.

The sedimentary deposits of San Clemente consist of sandstones, shales, and limestones, with a small amount of terrace conglomerate. The limestone and shale are of Miocene age, as is also a part of the sandstone. The rocks aside from these are in part of uncertain age, but it is probable that they are all post-Pliocene.

### MIOCENE.

Besides the main Miocene area, extending nearly across the island toward the northwestern end, there is a narrow border of rocks of this age along the shore of Wilsons Cove. A good section of a considerable part of the series is obtained in the pass above this cove. There is at least one small isolated patch 2 or 3 miles to the southeast of Wilsons Cove, on the northern side of the island. Toward the southeastern end Miocene deposits are found on either side of China Point as well as about Smugglers Cove. Wherever found, these rocks are thin-bedded and show evidence of disturbance, usually by gentle folding, but also in places by numerous small faults. The areas about China Point and Smugglers Cove have a general dip toward the ocean, ranging from about 10° to about 20°. The greatest observed thickness of the Miocene deposits is about 50 feet. An estimate based on those sections which represent different stages of deposition would give a total thickness for these deposits of about 100 feet.

The sandstones of the Miocene occur in the lower portion of the series, and are found principally in the pass back of Wilsons Cove and in the area just above the æolian sands on the opposite side of the island. They are medium-grained and light in color, containing more or less calcium carbonate, and grade into the shales above them. The sandy grains are in part simple minerals from the andesites, particularly the feldspars, and in part minute volcanic fragments. The carbonate of lime with which these sands are cemented usually forms a considerable, and sometimes the largest, proportion of the rock. These sandstones yielded no fossils.

The shales vary in color from yellowish or grayish to white. The specific gravity also varies considerably, and the less compact the shales the more evenly and thinly are they bedded.

The more earthy shales take up and hold water quite readily, and in a number of places along the shore they are saturated with salt water and are correspondingly soft and clay-like. The shales present the characteristics shown by the Miocene shales in general along the coast.[1] The description by Dr. Hinde[2] of the Miocene rocks of Santa Catalina would apply with little modification to both the limestones and the more earthy shales of San Clemente. Radiolaria are more abundant in general in the Clemente rocks, and foraminifera are more plentiful in local areas. Even in the shales with the lowest specific gravity the minute, angular volcanic fragments appear to be more numerous than in the rocks described by Dr. Hinde. The bulk of these lighter shales consists of diatoms, either entire or in fragments. Minute cavities occur in some of the shales and in most of the limestones. These cavities, where they present definite forms, are seen to be molds of foraminifera, and even where the forms are partly filled with calcite, the majority are undoubtedly of the same nature. All the Miocene rocks except the flaky, thin-bedded shale with the lowest specific gravity, show more or less effervescence with dilute hydrochloric acid.

The limestone occurs as lenses in the shale, with a diameter varying from half a foot to four feet or more; also in beds intercalated with the shales. Along the shore at Wilsons Cove the limestone contains abundant impressions of *Pecten peckhami* Gabb. No shells were seen in the shale, but fish scales are common.

The limestone is magnesian, a fact which would point to a primary origin in part at least. It is probable that it is in part secondary, the calcium carbonate being derived from the foraminifera. The percentage of magnesium carbonate in the rock is far in excess of that of foraminiferal oozes,[3] and it is this which has led the writer to the conclusion that the carbonates are in part original.

The shale at times occurs opalized, though such occurrences are not common. One such specimen, dark grayish-brown in color, has a hardness of about 4.5, is compact and banded, with a pronounced opaline luster. Foraminifera are rather abundant, occurring as minute, translucent spots of a light-gray color. The rock effervesces very slightly

[1] Some of the recent descriptions of the Miocene shale are found in the following:

The geology of Carmelo Bay, by Andrew C. Lawson: Bull. Dept. Geol. Univ. Cal., May, 1893, Vol. I (1893–1896), pp. 22–29. The geology of Point Sal, by Harold W. Fairbanks: Bull. Dept. Geol. Univ. Cal., May, 1896, Vol. II, pp. 9–15. The geology of Santa Catalina Island, by W. S. T. Smith: Proc. California Acad. Sci., 3d series, Geology, Vol. I, No. 1, February, 1897, pp. 43–51.

[2] W. S. T. Smith, loc. cit., pp. 45–48.

[3] In that portion of the limestone from Santa Catalina which was soluble in hydrochloric acid there was 27.3 per cent of magnesium carbonate. The limestones of San Clemente all show qualitatively a considerable percentage of magnesia, too great for a simple foraminiferal ooze. In the Report on the Scientific Results of the Voyage of H. M. S. *Challenger* during the years 1873–1876, Deep Sea Deposits, p. 219, the table of analyses of globigerina ooze shows a range in Mg $CO_3$, in that part which was soluble in hydrochloric acid of from 0.19 to 2.58 per cent, while the range of Ca $CO_3$ was from 37.51 to 93.14 per cent. Ca $CO_3$ in the Santa Catalina rock was 49.9 per cent.

DISTURBED MIOCENE ROCKS AT CONTACT WITH BRECCIA (ON EXTREME RIGHT), WILSONS COVE, SAN CLEMENTE ISLAND.

with dilute acid. Under the microscope it shows a distinct banding. The bulk of the rock is of opaline silica, though occasionally a small amount of chalcedony is seen, and at one point is a small aggregate of minute quartz grains. Scattered through the rock are numerous rhombohedra of calcite and occasional angular mineral fragments. Foraminifera are abundant, and are frequently well defined, with the walls well preserved. The interior of the tests is usually filled with chalcedonic silica. The chamber walls may occasionally be distinguished under crossed nicols by a different grade of crystallization from that found in the interior. In all cases the original carbonate of lime is replaced by silica.

The heavier varieties of shale contain abundant diatomic remains, and the bulk of the rock is isotropic.

From its specific gravity, however, it is highly probable that the percentage of volcanic matter in the form of glassy ash is in excess of the organic débris, and that the rock is to be called an ash rather than a diatomaceous earth. Coarse tuffs like those associated with the shales of Santa Catalina were not found on San Clemente, but a single bed of moderately coarse tuff, one foot thick, was found intercalated with the heavier shales at Wilsons Cove. This bed is light in color and contains scattered angular fragments of andesite.

The Miocene rocks were nowhere seen interbedded with the volcanic flows as they are on Santa Catalina, but they appear in all cases to rest on the volcanics. At Wilsons Cove they rest on volcanic breccia, in a disturbed condition. The contact is imperfectly shown in Pl. XCIV. The rhyolite, apparently the latest flow, is overlain by the shale, which is therefore later than the volcanic flows. The association with andesites, as in the case of Santa Catalina, and the similarity of some of the andesites here to those of Catalina, make it probable that on San Clemente, as on the other island, the andesites and the sedimentary deposits are closely associated in time, and that both are of Miocene age, the volcanic flows preceding the deposition of the shales and sandstones by only a comparatively short interval. As far as could be determined from the few contacts seen, the surface upon which the Miocene deposits were laid down had been somewhat, though not greatly, eroded before their deposition, a conclusion borne out by the presence of volcanic sands in the basal portion of the series of Miocene rocks.

### LATER DEPOSITS.

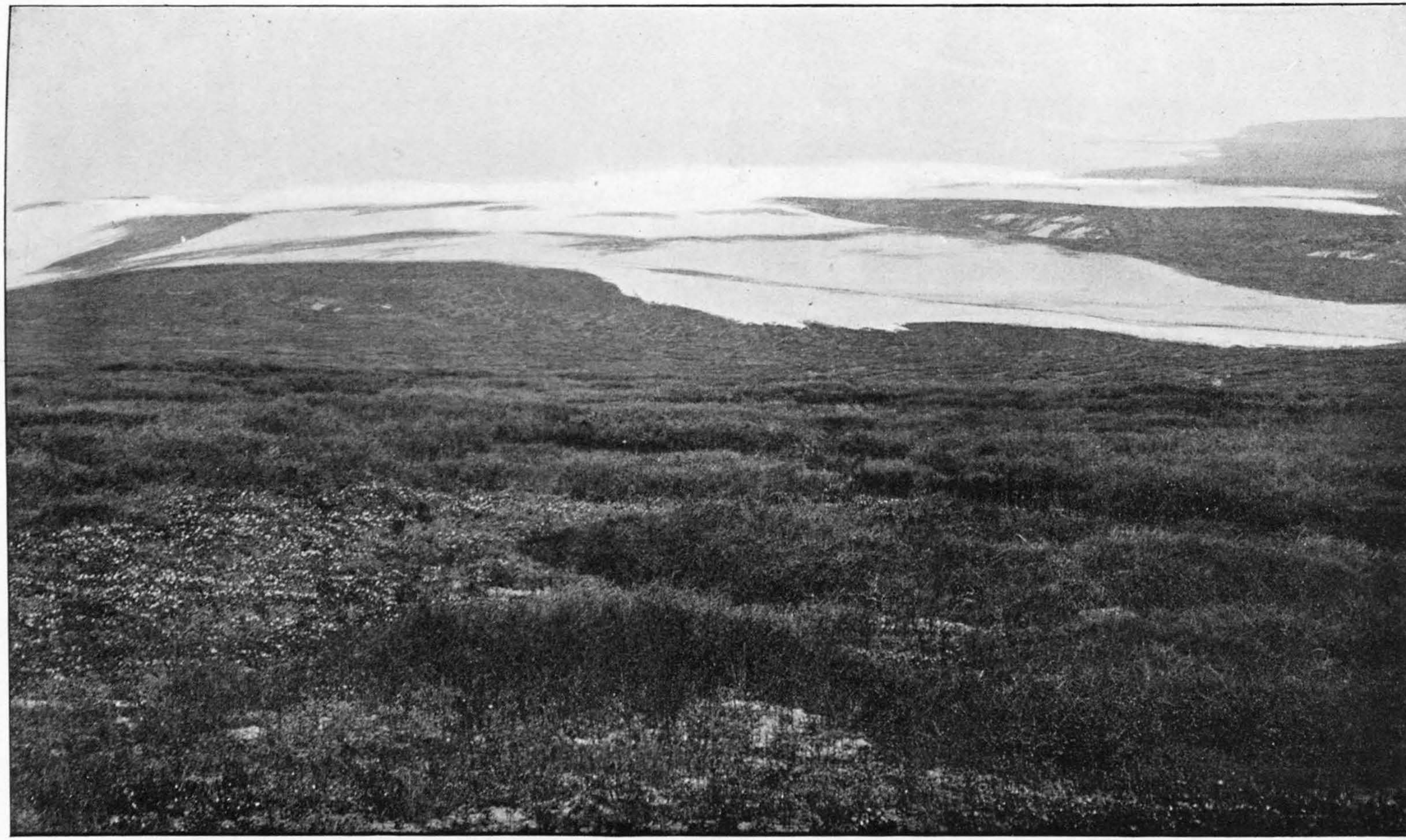
The age of the deposits later than the Miocene has already been referred to. If San Clemente existed as an island during the Pliocene, it could have been only a small one, consisting of the highest parts of the present island, as is shown by the upper terraces. The highest point where the later deposits are now found must have been at the very least 200 feet below the sea level as it then was, and at a considerable distance from the shore line of the island then existing. If they belong to that time, therefore, the deposits under discussion should give evidence of having been laid down in moderately deep water. They are not of such a character, however, but are in large part moderately coarse sandstones, with occasional pebbles, or else they are clayey or calcareous deposits usually containing numerous large pebbles of the volcanic rocks. In no case do they show any bedding. Further, some of the fossils found in these deposits are undoubted littoral forms. It is probable, therefore, that they are all post-Pliocene deposits, and that they were formed along or near the shore lines of the island as it was successively elevated.

As has already been mentioned in another connection, shore conglomerate has been found at a number of points near the angle of terrace and cliff or upon a terraced surface. The rest of the later deposits are found principally in the neighborhood of Miocene rocks, largely because these are softer than the volcanic rocks and therefore more readily furnish detritus. In other cases the reason for their association is that remnants of the Miocene deposits are found in the embayments. It is in such places that the greatest deposition usually takes place along a coast, and thus it is natural that we should find in them some of the post-Pliocene deposits.

Besides the occurrences near the Miocene rocks, there is a small area near Seal Harbor, another smaller one of sandstone, not shown on the map, in the upper part of the island toward its southeastern end, and a third, also of sandstone, near the northwestern end on the neck of land between Northwest Harbor and West Cove, known as the Isthmus. The formations at this point are largely concealed, but it is probable that most of the surface rocks of the Isthmus are of sandstone, both on account of the character of the topography and because areas of drifted sand are seen on each side. The sandstone of the small patch indicated on the map near Northwest Harbor is light-gray and friable, while the usual color of the sandstone is a dirty-brownish. The sandstone near Seal Harbor is scarcely more than a remnant, and resembles the Miocene sandstone in general appearance, but it seems likely that it is of post-Pliocene age. It contains a considerable proportion of fragmental volcanic material.

At no point do these deposits form more than a thin mantle over the volcanics or Miocene sedimentaries, their maximum thickness being perhaps 10 feet. In places they thin out till the merest film of sand or whitened soil masks the rocks beneath. Their boundaries are, therefore, not always definite. In several instances these deposits rest on the beveled edges of the upturned Miocene strata. In the upper part of the main sedimentary area they can not always be definitely separated from the Miocene, owing to the gently rolling character of the region, the comparatively thick soil and abundant growth of grass, and the few definite channels.

Fossils were found in these later deposits at only three points. One



ÆOLIAN SANDS, SAN CLEMENTE ISLAND, OPPOSITE SIDE FROM WILSONS COVE.

of these has already been mentioned, where shells of *Lucina californica* Conrad occur with rolled pebbles on the terraced Miocene deposits, nearly 2 miles southeast of Wilsons Cove. A short distance southeast of this occurrence there is found near the shore a deposit having the appearance of a travertine, and containing, besides numerous andesitic pebbles, a large number of land shells. Only two species were found among these shells—*Helix* (*Arionta*) *ruficincta* Newcomb, and *H.* (*Euparypha*) *tryoni* (?) Newcomb.

The third fossiliferous deposit occurs nearly 4 miles southeast of Wilsons Cove, at an altitude of nearly 800 feet. Here there is a small patch of compact limestone, largely composed of the remains of calcareous algæ, with a few fragments of marine shells. There are also associated with the organic remains small fragments and grains of volcanic rock, besides mineral grains, principally feldspar.

Concerning the occurrence of the calcareous algæ, Dr. J. C. Merriam, who first recognized their generic character, has kindly furnished the following note:

> The rock is composed largely of stem fragments of calcareous algæ, which were examined by Prof. W. A. Setchell and pronounced by him to be closely allied to Lithothamnium. This genus has not heretofore been known fossil from California, but is not rare in the post-Paleozoic rocks of Europe. Mingled with the Lithothamnium stems are a few fragments of gastropod and lamellibranch shells, which resemble species known from the Quaternary deposits of southern California.

There are considerable areas of æolian sands at both the northwestern and southeastern ends of the island. The largest and most interesting area, in the form of dunes, is on the opposite side of the island from Wilsons Cove. This area does not reach the present shore line at any point, this part of the coast being bordered by cliffs of the volcanic rocks. The sand can not have come, then, from the present shore by drifting, but probably belongs to an older shore line. The source of this sand is probably in small part the volcanic rocks, but doubtless the larger part of it comes from the neighboring underlying sandstones. This is the area which contains the roots and tree trunks replaced by sand and calcite (see page 466), but the most interesting feature is the countless number of land shells which fill the sand over the greater part of the area. They are for the most part the same as the principal species of the present day on the island—*Helix* (*Arionta*) *intercisa* W. G. Binney. Their occurrence here in such multitudes, associated with the roots and trunks already mentioned, points to a time when these sandy slopes were covered with a low, dense growth of shrubbery, about the roots of which flourished the animals whose shells we now find. There also occur a great many marine shells, similar to those now found along the shore, but their number is insignificant in comparison with that of the land shells. It is probable that a large part of the marine shells were brought here by the Indians, the remains of whose camps are still seen here and there.

## HISTORY.

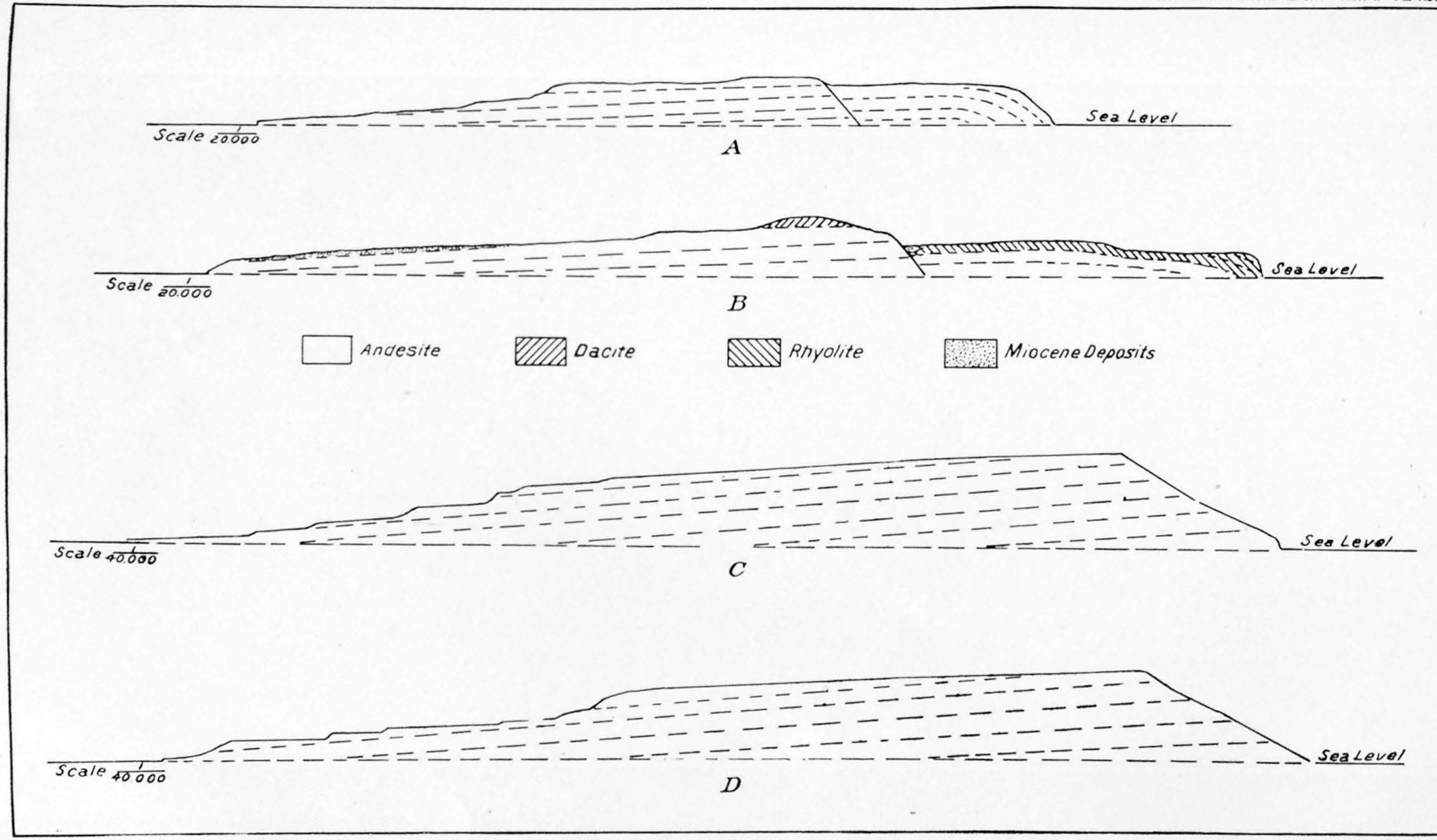
The probable age of the volcanic rocks forming the body of San Clemente Island has already been given as Miocene, little antedating the deposition of the Miocene sedimentaries. It has been shown[1] that the andesites of Santa Catalina were poured forth during a period of slow submergence in Miocene times. Since these rocks rest on the shales of Catalina, and since the volcanics upon which the shales of San Clemente rest were apparently eroded before the deposition of the shales, these volcanics must not only have formed a continuous series through a considerable part of the Miocene but they must have continued to pour out upon the surface of Catalina long after they had ceased to reach the area now occupied by Clemente. San Clemente did not then exist as a differentiated land mass, but formed a part of a larger area of low relief, over which the volcanic flows were spread. The Clemente area appears to have been close to the coastal margin.

A part, at least, of the volcanics of San Clemente must have been laid down in water, since the narrow sheet of ash found in Red Canyon, well down in the series of volcanic rocks, is distinctly bedded. The ash found in the upper part of the series, however, is not bedded. This fact, together with the erosion of the volcanics, would indicate one of two things: either the submergence had not yet set in when these rocks were laid down, or, if it had, it was slower than the upbuilding action of the volcanic flows—all this, of course, on the supposition that the bed of ash first mentioned was deposited in marine waters, as was probably the case. In either of the above conditions of submergence the facts would seem to point to a shore line not far from the place now occupied by Clemente.

The Miocene sandstones containing sands from the andesitic rocks also indicate a period of erosion before the deposition of the shales; and further, the transition from sandstones to shales would show a depression at this time. So here, as at Santa Catalina, it is probable that the coast was sinking, at least during the latter part of the time in which the extravasation of the volcanics occurred. During the time of Miocene submergence the shales were deposited. The heavier shales, with the intercalated bed of tuff, would indicate that the effects of volcanic activity had not at this time ceased to reach this point.

The Miocene depression was followed by a post-Miocene elevation. A further faulting of the Catalina crust-block at this time may have led to the differentiation of San Clemente, by producing strains in the crust to the south, relieved by faulting, which formed the Clemente crust-block. Whatever the cause, San Clemente appears to have been formed by faulting, with slight folding from end to end, at the close of the Miocene or in the post-Miocene interval of erosion. In this interval the coast was more elevated than now, with the ocean strand somewhat to the westward of its present position. The period of erosion

[1] The geology of Santa Catalina Island, by W. S. T. Smith: Proc. California Acad. Sci., 3d series, Geology, Vol. I No. 1, February, 1897.




SECTIONS OF SAN CLEMENTE ISLAND.

*A*, Northeasterly-southwesterly section, a little over one-half mile northwest of Wilsons Cove.
*B*. Easterly-westerly section, about three-fourths of a mile southeast of Wilsons Cove.
*C*. Northeasterly-southwesterly section, about 1 mile southeast of Middle Ranch Canyon.
*D*. Northeasterly-southwesterly section, about 2 miles northwest of Red Canyon.

was long, allowing for the excavation of broad, deep channels along the California coast, afterwards filled in with Pliocene delta deposits.[1]

If the broad valley-like form near the southeastern end of San Clemente (seen also in the submarine contours) is due to erosion, the crust-block at this time must have had no great elevation, and only a part of its faulting could have taken place. Its main surface, as now, sloped to the southwest, but it must have been at a much smaller angle. During the post-Miocene erosion, according to this view, broad and shallow channels were cut in the gently sloping surface, and these extended some distance to the west and south beyond the present shore line, as is seen in the example at Smugglers Cove. If Clemente had had more than a moderate elevation at this time, while subject to the forces of erosion, its topographic forms should be older and more like those of Santa Catalina than we find them. It would follow, then, that the greater part of the faulting which produced this block must have occurred early in the succeeding Pliocene depression of the California coast. If, however, this valley at Smugglers Cove is structural in its character, we have no definite information as to the attitude of San Clemente or its relation to the coast line at this time.

During the Pliocene depression San Clemente became an island for perhaps the first time in its history. It may be that this island was wholly beneath the water in Pliocene times. If so, it must have been for a comparatively short time, for during the greater part of the Pliocene the sea stood in the neighborhood of 1,500 feet above the present sea level, as shown by the broad terraces with low cliffs near that elevation. After the Pliocene depression the island slowly emerged from the waters, resting at intervals while the great terraces on the southern slopes were being cut.

The present shore line does not necessarily mark the limit of this elevation, for it can not be positively stated that the most recent movement has not been one of depression. There is developed on the southern side of the island a broad submarine bench, well shown in the sounding-contours and the section on the map (Pl. LXXXIV, p. 464). The slope of the bench, as is evident in this section, can in no way form a part of the original slope of this side of the crust-block. It is rather of the nature of a terrace, in part, at least, wave-cut. As has been shown for the corresponding submarine terrace about Santa Catalina, so here the most nearly level portion of the terrace is between, approximately, the 200- and 400-foot submarine contours, the grade of the slopes increasing above and below those points. The more rapid descent below 400 feet is shown on the map. Above 200 feet the increase of grade, though a fairly constant feature on the map, is not so evident at the point where this section was taken. The existence of this submarine terrace, with other minor evidence, suggests that the post-Pliocene elevation may have continued to the 400-foot contour, or even farther. It must then have been followed by a submergence which

[1] Post Pliocene diastrophism, p. 158.

has continued to the present time. This hypothesis, however, could be established or disproved only by a study of the other Channel Islands and the coast, in connection with San Clemente.

The greater part of the faulting of the Clemente crust-block must have occurred at some time between the close of the Miocene and early Pliocene times. Even though this is so, it may be that the faulting, once begun, has continued at intervals ever since. It is uncertain whether any movement occurred during post-Pliocene time along the main axis of faulting, though some movement certainly has taken place in other parts of the block, as is evidenced by the minor faults on the island, somewhat oblique to the direction of the major fault.

If there has been any pronounced differential upward movement of the crust-block during the post-Pliocene coastal elevation it could be ascertained by accurate measurements of the altitudes of terraces developed on both sides of the island. One of the principal obstacles to such a determination, however, is that it is impossible to obtain the altitude of any terrace with absolute accuracy, owing to the accumulation of detritus which conceals the actual line between cliff and terrace. Further, it is necessary to be certain that it is the same terrace whose altitudes are compared on the two faces of the block—a difficult task on San Clemente, where the few terraces on the northern coast are fragmentary and little developed at best. If such differences exist between the two sides, they can not be great. For any given differential movement, differences in the altitudes of the terraces on the two sides of the island would be greater near the shore than near the crest, on account of the greater width of the island near sea level. Even if differential movements had been continuous through post-Pliocene times, as the terraces were being formed, it is probable that, owing to this difference in width, the accumulated differences in altitude of terraces in the upper part of the island would not greatly exceed the differences of those at lower levels, which had suffered less of the differential movement.

What the differences of altitude might be for any given differential movement in the Clemente crust-block may be roughly calculated. Taking the altitude of San Clemente as 2,000 feet, with 5,500 feet as the total elevation of the block above the floor which surrounds it; assuming, further, an angle of 5° for the gentler slope of the block, and only 22° for the steeper slope, we should have a minimum distance of 12 miles from the apex of the island to the limit of movement on the side of the gentler slope. These figures would give a width of about 5 miles at the shore line, which is somewhat more than the maximum width of San Clemente. If under these conditions (which are very nearly those of the Clemente block to-day, along a section through the highest parts of the island) there were to be a differential movement causing an elevation of the summit amounting to 100 feet, the difference in the amount of elevation at the shore line on the two sides would be about 45 feet, while at an altitude of 1,600 feet the difference on the two sides would be less than 10 feet.

# GEOLOGY OF THE CAPE COD DISTRICT.

BY

N. S. SHALER.

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# ILLUSTRATIONS.

# GEOLOGY OF THE CAPE COD DISTRICT.

By N. S. Shaler.

## INTRODUCTORY NOTE.

It was at first intended that this paper should include the geology of the peninsula of Cape Cod only, but the progress of the work has made it necessary to extend the consideration so as to take some account of the structure and the succession of deposits exhibited in portions of the mainland, as well as on the islands to the southward. The progress of the inquiry has made it necessary to limit the scope of the work to a somewhat extended discussion of the evidence that goes to show the series of geological events which have occurred in this district since the beginning of the Cretaceous period. Such of the facts as pertain to this discussion are given in this paper; further details will be set forth in the geological folios of the area, which it is expected will shortly be ready for the press.[1]

Inquiries in this field have been very limited in their number and range. Those instituted by the United States Geological Survey have resulted in a Report on the Geology of Marthas Vineyard, in the Seventh Annual Report of the Director (for 1885–86); a report on The Geology of Nantucket, being Bulletin No. 53 (1889) of the Survey series (both by the writer of this paper); a report on The Glacial Brick Clays of Rhode Island and southeastern Massachusetts, by N. S. Shaler, J. B. Woodworth, and C. F. Marbut, in Part I of the Seventeenth Annual Report of the Survey (for 1895–96); and the unpublished folios above referred to.

The reader of the above-named published reports and of this paper will perceive that the Cape Cod district has unexpectedly revealed a considerable range of phenomena, the discussion of which is certain to throw much light on the geological history of the Atlantic coast line. Unfortunately the evidence concerning the succession of these phenomena is of a very obscure nature, and it is therefore not surprising that in the reports above referred to some of it was misapprehended and much was not discerned. Nor must it be supposed that in the following pages anything like a final statement of the facts or of the conclusions to be drawn from them is to be found. Such a statement can not be expected until investigation has gone much further.

I take pleasure in acknowledging indebtedness to Mr. J. B. Woodworth for advice in some parts of the work, and to Messrs. Mark S. W. Jefferson and John Gardner for help in obtaining the photographs from which the illustrations are taken.

[1] Folios of the Geologic Atlas of the United States.

## ORIGIN OF CAPE COD PENINSULA.

The origin and structure of the peninsula of Cape Cod have been a matter of passing interest to all who have considered the geology of the southeastern portion of New England. The peculiar spit-like form of this promontory was at first, and naturally, supposed to be accounted for by the action of the marine currents to which are due the construction of so many of the lesser capes along this portion of the Atlantic shore. When it became evident that a large portion of the materials composing the higher parts of the cape had been brought into position by the action of ice during the last Glacial period, the spit theory was abandoned, and it was at once assumed that the greater part of this area owed its existence as dry land to the morainal and stratified drift deposits which are so evident on the surface, the northeastern extremity being a later addition, made by the action of marine waves and currents.

The last-noted hypothesis as to the origin of Cape Cod, by glacial action, long appeared to have much support from the view, so generally entertained, that the outer morainal deposits formed during the advance of the ice were likely to be massive and of great extent; so that it thus seemed reasonable to suppose that the portion of this cape that was evidently not due to marine agencies was accumulated as a frontal moraine. An inspection of this field alone, without the use of corrections which may be obtained from other parts of the country, almost necessarily leads the observer to adopt the view last mentioned. It was not until I had seen much of the morainal deposits of the region between the Cordilleras and the Atlantic shore, and had made a study of the relations of those accumulations to the Tertiary and Cretaceous rocks of Marthas Vineyard, Nantucket, and other parts of the Atlantic coast line between southern New Jersey and Boston Bay, that I gained what seemed to me to be a truer insight into the nature of the singular peninsula of Cape Cod. On this account it appears desirable to preface the study of this district by an account of the facts revealed in neighboring fields which seem to throw light on its problems.

## GENERAL RELATIONS OF THE DISTRICT.

A glance at a map of the eastern shore of North America will show that the peninsula of Cape Cod is in some respects the most peculiar feature of this coast line; geographically considered, it is in a high measure exceptional. Its crescentic form, as before remarked, is by no means unique, except as to the great size of the hook, many of the sand spits imitating in a small way the general coastal outline of this peninsula; but in the bold manner in which this salient projects from the shore, in its strong topographical relief, and in the character of its coast line, it finds no parallel, so far as I have been able to ascertain, in any country. This exceptional geographical character naturally leads an observer who is aware of the indicative value of such features to seek the origin of this cape in conditions of an unusual sort, such as will become apparent in a discussion of the general relations of the district.

It has long been known that the Cretaceous and Tertiary deposits so extensively developed in the southern portion of the Atlantic States of this country are continued in an interrupted belt lying to the east of the more ancient rocks as far north as southeastern Massachusetts, the Cretaceous extending up to the deposits on Marthas Vineyard and the Miocene Tertiary reaching to Marshfield, a point some distance north of the northern border of Cape Cod. Associated with these Mesozoic and Cenozoic deposits are extensive series of stratified sands and gravels which have hitherto been commonly classified with the glacial drift. South of New York these beds show little signs of disturbance by orogenic action; such distortions as have been noticed in the beds can apparently in most cases be explained by accidents of deposition. North of New York, on Long Island and in the isles to the eastward, these beds have been subjected to dislocation, which in Marthas Vineyard becomes profound, so far as is indicated by the attitude of the beds, exceeding on the average the distortions of the Appalachian Mountain district or of the neighboring field of the Narragansett Basin.

Certain observers have sought to account for the dislocations of these newer rocks on the New England shore district by supposing them to be due to the action of the glaciers of the last ice epoch. As I have elsewhere noted, this view seems quite inadmissible, for the reason that the uplifting and folding of the beds took place long before the advent of the last ice epoch. As this point is of much importance in the discussion of the problem as to the origin of Cape Cod, it will be well to present the facts in some detail, especially as certain excavations recently made on Marthas Vineyard have somewhat extended our knowledge concerning the history of the glacial work in that field.

On Marthas Vineyard the Cretaceous and Tertiary strata, exhibiting a total section of probably 1,000 feet or more, are cast into folds of considerable amplitude, some of them apparently exceeding 1,500 feet in transverse extent. These folds are compressed, overturned, and faulted; in a word, they exhibit all the marks of mountain-building actions working on stratified deposits of weak resistance to compression and not deeply buried. So general and effective has this dislocation been that it has involved all the rocks which are exposed to view, the average dip of the strata perhaps exceeding 40 degrees.

In these exceedingly disturbed strata river valleys were excavated which had their position determined in the usual manner, the greater streams following in general the strike of the beds, the lesser—those occupied by the temporary streams—running at right angles thereto. The larger of these valleys, that of Tisbury River, is about one-third of a mile wide and more than 100 feet deep. Upon this normal and well-developed topography, which indicates a continuance of stream erosion that must have occupied a period to be measured by tens of thousands of years, came the glacier of the last ice epoch. I have elsewhere[1] noted the fact that this ice sheet had little erosional effect upon the topography of this island, and the impression made by my first studies has been confirmed by recent inquiries in the same field. The facts may be briefly stated as follows:

The ice sheet failed to obliterate many details of the topography which were due to differential erosion before the advent of the glacier. At many points the ridges of harder rock, though at most no firmer than compacted sand or soft clay, stand evidently as they were originally formed. So imperfectly did the ice abrade the surface that the white and red colors of the clays is rarely traceable to a height of a foot above the contact of the till with the underlying beds. Although along the crests of the greater ridges there are morainal accumulations which have in places a thickness of from 20 to 50 feet, these are limited to the northern side of the island; the southern part has only slight moraines. Over nearly one-half the area in which the Cretaceous and Tertiary strata rise above the level of the sea the till coating does not average 3 feet in thickness, and many fields of a hundred acres or more in extent are essentially driftless. On the southern shore the evidence at present afforded by the rapidly retreating cliffs is to the effect that a deeply incised topography formed in the Nashaquitsa clays was not effaced, the sharp valleys being merely filled in with the drift deposits. In a word, the conditions of this area indicate that the glacier of the last ice epoch was of such slight dynamic value that it produced little erosion and that all the important dislocatory work was done long before it came upon the district.

It is to be said that there is some evidence of ice action shown by the character of the latest-formed deposits of the disturbed strata, seen in the presence in one of the conglomerates exhibited at Gay Head of pebbles and bowlders apparently derived from the region of Narragansett Bay, including one fragment of the very characteristic ilmenite from Iron Hill, in the town of Cumberland, Rhode Island. But this ice period of the Pliocene or Pleistocene time was, if it existed, so far as we can discern, an even less effective invasion than that of the last Glacial epoch, and, as it came before the dislocation of the beds, can not possibly be made to account for their disturbance. There is thus no reason to doubt that the extensive stress phenomena of this field must be explained by supposing that they are in some way the result of orogenic action. We are, indeed, justified in assuming that along the section of the shore line extending, it may be, from western Long Island to the island of Nantucket, mountain-building movements involving stresses of considerable intensity have been developed.

As to the operation of these mountain-building actions in the district of Cape Cod, the evidence, though not perfectly clear, leads to the

[1] Report on the geology of Marthas Vineyard; Seventh Ann. Rept. U. S. Geol. Survey, 1885-86 (1888), p. 310.

conclusion that they worked on the ill-disclosed foundations of that peninsula in much the same manner that they have done in the well-exhibited beds of Marthas Vineyard. As will be noted in the sequel, the strata which are known on Cape Cod include nothing below the level of the Nashaquitsa series as described in the report on Marthas Vineyard; but the presence of the Tertiary greensands at Marshfield causes the presumption that beds of earlier age lie within the peninsula.

The limited extent of the exposures of the foundation materials of Cape Cod makes it desirable to take into account the structure and history of the adjacent areas both on the south and on the north. It is, indeed, necessary to do this in order to arrive at an understanding as to the history of the particular area. This consideration should include the origin of the sediments, the nature of the transporting agents which brought them to their sites, the orogenic accidents, the development of the drainage, and the oscillations of the sea level which have taken place on this portion of the shore.

The sediments of the Cretaceous and Tertiary rocks in the district between Washington and Boston exhibit certain peculiarities which are not found elsewhere in the eastern United States. The section is in part made up of colored clays and sands, which, except for the admixture of peaty matter in the lignite beds, are evidently derived from the rapid deposition of land waste washed from an area which had been long subjected to interstitial decay, which was followed by rapid erosion. In the lower portion of the beds the conditions are not so abnormal, the clays and sands in general resembling those of the Southern States. They appear to have been deposited from the discharge into the sea of ordinary rivers. The structure of the lignites, which, so far as observed, contain much clay, indicates that they were formed in an estuarine district, subjected to frequent floodings of muddy water and to slight subsidences, which permitted the peaty accumulations to be buried beneath silt.

In passing to the higher marine strata, we find at once that we are in very different clastic conditions. The beds in the Marthas Vineyard district consist of alternating clays and sands, which have evidently been deposited in a rapid manner. The clays show scarcely a trace of lamination, and the sands are exceedingly coarse, often being made up of bits of decayed granite, the crystals running together in one mass. Much of the deposit is composed of detached, not rounded, crystals of feldspar, which are so far softened by decay that they can be crushed in the fingers. It is, in a word, a true arkose, lacking only the usual consolidation of that material, and so destitute of admixture of such substances as are inevitably brought into detrital beds where the transportation which bore the waste to its resting place was by rivers or shore currents, that a careful study of sections many square yards in area has failed to show a trace of any other material than the broken-up crystalline rock from which it was derived.

After passing up through a section having a total thickness of several hundred feet in which the above-noted alternations are exhibited, we come suddenly to a level where beds of conglomerate, composed of ordinary compound hypogene rocks, occur in pebbles of moderate size not differing much in character from those formed during the last Glacial period, except that they are more decayed and somewhat more waterworn. Yet higher in the section we attain to the Nashaquitsa series, which are also somewhat dislocated. These are beds of sands and clays, in general character like those formed during the last Glacial period, though on account of their greater age they have been much more changed in texture than those of that epoch. The interpretation of this section is difficult. The most probable explanation is that which will now be set forth.

In the first place, we may note the fact that the shore line of the old crystalline district of the Appalachians appears always to have lain near this seat of deposition. The arkose in the Tertiary shows this to have been the case in that period. It was so again at the time of the higher conglomerate, and the character of the clays and arkose beds shows that they were not offshore deposits. The structure of these beds suggests that they were laid down in a swiftly accumulated delta at the mouth of a river, which might well have been a continuation of the Connecticut.

The lower Cretaceous deposits, being in nature such as would be discharged from streams draining a land subjected to ordinary conditions of erosion, demand no special explanation. As before noted, it is quite evident that the rocks beneath the land from which they came had been deeply decayed in a long period of stable conditions, such as has prevailed in the southern Appalachians. Suddenly this zone of decay was to a great extent swept away into the neighboring sea, the process continuing until, as the conglomerates which cap the Tertiary section show, the firm-set undecayed rocks were, in certain places at least, exposed to the eroding agents.

The supposition that there was in the Mesozoic period a deep zone of decayed rock in New England which might have afforded, if subjected to rapid erosion, detritus such as is contained in the clays and arkoses of the Tertiary rocks of southeastern Massachusetts, finds some support in the occurrence at many points in that area, particularly in the southern half thereof, of rocks decayed in place under conditions which clearly show that the disintegration has not been brought about since the last Glacial period. Rocks in this state, exhibiting decay to the depth of some score of feet, occur at various points in and about the Boston Basin, and in a number of places in the Berkshire Hills and elsewhere. A notable instance of this decay of the strata in place was found in the excavation of the Hoosac Tunnel, where, for a length of several hundred feet, near the western portal, the mica-schist was found completely softened at a depth of 400 feet below the surface. Owing to the deep covering of glacial drift which hides so much of the surface of New England, as well as to the fact that the ice of the last ice period removed all projecting rocks of this nature, it is only chance excavations, such as are rarely made, that give one an opportunity to see these remnants of a decay which was once widespread. From a careful examination of the evidence, I am of the opinion that at least one-thirtieth part of the crystalline rocks of Massachusetts, Connecticut, and Rhode Island would, if bared, exhibit decay of the type so well known in the plateau district of the southern Appalachians.

The cause of the sudden removal of this material from its old to its new bed place is not easily determined; the following suggestions seem, however, worth consideration. It is doubtful that the result was brought about by the invasion of the sea during a period of subsidence; the singularly unmixed character of the deposits, the entire absence of marine organic waste, which is likely to be found in beds of this nature, and the perfect assortment into thick layers of like sediments are also against this view. Moreover, the cutting rate of coastal erosion agents is normally slow, while these beds indicate very rapid work of this kind. So, too, the hypothesis of exaggerated land erosion due either to a great increase in rainfall or to a steepening of valleys brought about by a change in the attitude of the land, seems inadmissible for the reason that the detritus from any ordinarily conditioned area would have been stained by the organic waste that all such streams normally bear to the sea. It is difficult to conceive a large river carrying and depositing in succession red and white clays and arkoses without a trace of vegetable detritus.

The difficulty which is encountered in the effort to explain the erosion of the detritus of the Gay Head beds by marine action is well illustrated by what is now taking place on the rapidly wasting cliffs of that part of Marthas Vineyard. The materials of the section are to a certain extent rearranged along the shallow-water belt of the shore, but the various forms of detritus are intermingled, and are mixed with organic matter to such an extent as to make the intervention of the sea unmistakably manifest.

It may be suggested that the beds in question have in some way been bleached or colored since they were deposited. This view can not, as is at once seen, be maintained in the case of the clays, for the lignite beds of the Cretaceous which are mingled with them carry the carbonaceous stain with no trace of bleaching. I have been unable to conceive any chemical action occurring in either the clays or the arkoses which might possibly account for the disappearance of original organic waste.

In this state of the problem I have been forced to bring in the hypothesis that the erosion work which removed the materials of these strata from their parent rocks was effected by glaciation, the ice not attaining to the place of deposition, but delivering the detritus to streams, one or more of which debouched near this part of the coast into the sea, or perhaps at times into a lake. Glacial action accounts for all the facts which we have noted concerning the character of these deposits in a way that no other operation could well do; in fact, without using this hypothesis we are left quite without an explanation of a very interesting series of phenomena.

In favor of the hypothesis that glacial erosion delivered the detritus of the Cretaceous section to the currents which bore it to its present resting place, we may note the fact that occasionally, though rarely, in the clays of the Gay Head cliffs we find large, subangular masses of a yellowish-red sandstone embedded in the strata. One of these, visible for some years, has recently fallen and broken to fragments. It was originally not less than 20 cubic feet in volume. In the course of the thirty-six years that this slowly retreating cliff has been under my observation, five or six of these interesting fragments have been noted which were certainly not to be classed with the ordinary glacial bowlders that often work down the slopes so as to appear as if they were embedded in the strata. It will be observed that these apparently ice-rafted rocks are of sandstone, a material which sometimes resists the process of decay where all ordinary hypogene rocks yield to it. It is just such a petrographical species as we should expect to find affording the rare bowlder which would be formed where glaciation took effect on an area indicated by much decayed crystalline rocks. Thus in the Connecticut Valley, whence these floated masses possibly came, borne by ice rafts, the sandstones have suffered but little interstitial decay, while the older rocks of the neighboring Berkshire Hills have, as noted, been in places much disintegrated.

The orogenic history of the Cretaceous and Tertiary strata of the New England islands is even more puzzling than are the conditions of their formation. So far, no evidence has been adduced to show the action of mountain-building forces in any of the beds of this age in the region south of New York. To find on this portion of the coast much evidence of dislocation of a high order is surprising; it justifies, indeed, the effort of those geologists who have endeavored to account for these movements by the thrusting of the ice sheet. We have seen that this explanation is for several reasons inadmissible, and the question arises as to the origin of the compressive strains which have acted in this area. To determine this, so far as it is at present determinable, we should begin by noting the following facts:

The amount of the shortening of the beds as shown on Marthas Vineyard, where a cross section having a length of about 5 miles is exposed, is probably nearly 2 miles. This is shown by the fact that the average dip of the beds, as determined by many observations, is about 45 degrees. There is good reason to believe that the area involved in these disturbances is much greater than the exposures on the island seem to indicate. In my opinion the beds may reasonably be supposed to have at least twice the extension across the strike that is known to exist, and it is not improbable that the area involved may have a width of 30 miles. The dislocations on Nantucket, though not well known, and those noted by Mr. Woodworth on Block Island, and also those on Cape Cod, hereafter to be described, seem in a way to validate this conjecture.

We have next to note that while the strikes of the folds on Marthas Vineyard are somewhat irregular, their commonest direction is from north-northwest to south-southeast, or nearly at right angles to those of the Appalachian folds of the neighboring mainland. This feature at first raised a doubt as to the orogenic nature of these foldings, for the reason that it seemed unlikely that such a departure from the normal strike of the district would take place if the movements were in character like those ordinarily involved in mountain building. But a comparison of the facts with those observed in other areas makes it clear that this discrepancy is not of great significance. In the Cordilleras and elsewhere it is not uncommon to find that the later movements in any mountain system show the effect of stresses acting at high angles, or even normal, to those which were originally effective. It seems, indeed, that the compressive strains of any district tend in the course of time to satisfy themselves through folds running in more than one direction; that when the strains in a certain axis are relieved there is often a tendency to form others in contrasted directions rather than to develop those which were first made. Therefore, the peculiar position of the axes of Tertiary disruption in this area can not be urged as a weighty argument against their true orogenic character.

It is to be observed that the dislocations of the Marthas Vineyard and Cape Cod sections differ in a notable way from those which occur in the older rocks of the Appalachian district. The folds are small, none of them, so far as clearly observed, exceeding a few hundred feet in horizontal amplitude; they are much compressed, and frequently overturned; they are cut by numerous faults, none of which appear to have a throw of more than 100 feet. In some places these accidents of stressing are so numerous and have so intermingled their effects that the result is a confused jumble of entangled beds which can not well be unraveled. At first sight these peculiarities of movements of the Marthas Vineyard section suggest to anyone familiar with ordinary mountain-building work that the strains which have effected them were of a different order from those which uplifted the Alleghenies and other normally folded mountains. It is, however, to be noted that the stresses which acted on these newer rocks took effect under very different conditions from those under which the old strata of the Appalachians were dislocated. There the beds were rigid and deeply buried; here they were soft and had little overburden to oppose their movements when the stress was applied to them.

As yet we have little information concerning the nature of the work done by orogenic action in the superficial portions of a section on which it has taken effect, but all the considerations derived from laboratory experiments, as well as from the principles of dynamic action, lead us to believe that near the surface of a stressed area the folds are more likely to be small and of varied form than in the deeper-lying parts of it, and that in soft strata without beds of such rigidity as to control the movements slight local accidents are likely to determine the formation of many small folds rather than a few of large size.

It is worth while here to note that these Vineyard dislocations, in case they are accepted as of truly orogenic nature, may well be taken as examples of what is likely to be the type of mountain folding as exhibited in weak beds which, at the time of disturbance, lie within a few hundred feet of the surface. So far as I am aware, there is no better place known in which to study this interesting phase of mountain building.

Assuming, then, that the rocks of the sections exhibited on Marthas Vineyard owe their very great dislocation to forces which had their origin in the under earth, I shall consider certain possibilities as to the exact source of these strains. It may be suggested that a slipping movement has occurred in these beds, due to the formation of a great inclined fault extending parallel to the shore and dipping toward the sea at a high angle, the resulting movement being in effect a landslide. This view is inadmissible for the reasons that there is no trace of such a slip fault; that the section moved is of a rank in size of which we have no knowledge elsewhere; that the transverse shortening of the beds is too great to be accounted for in this manner, and that the direction of the axes of the folds is at about right angles to such as would be formed in such a movement.

It may be worth while to set forth another hypothesis which I have been led to apply to these movements in order to arrive at an explanation of them without recourse to true mountain-building action. This is as follows: A large part of the materials in the Marthas Vineyard section is feldspar, which had apparently been imperfectly kaolinized before it was brought to its present site. Is it not possible that the considerable increase of bulk attendant on this conversion of feldspathic matter into kaolin may have led to internal pressures? It appears, however, that of the mass not over 15 per cent can be reckoned as feldspathic, while even if the whole of it were of that nature and all the changes had come about since the beds were laid down the amount of enlargement would be too small to account for the observed disruption of the beds.

It seems evident that we must account for the folding and other movements of the Tertiary and Cretaceous rocks of these New England islands by the ordinary process of mountain building. Questions then arise concerning the nature of the dislocations by which the compressive strains were applied to these superficial beds. The general slope of the hypogene and Carboniferous rocks on the neighboring mainland makes it eminently probable that these old deposits underlie the newer beds at a depth of not more than 1,000 or 2,000 feet below the surface of the sea. How, then, were these lower beds affected in order that they might transmit the strain to the Mesozoic and Cenozoic deposits? and what, if any, were the dislocations on the mainland that were produced at the same time? It seems to me that neither of these questions is, in the present state of the inquiry, answerable with any measure of affirmation, but some suggestions may be made which are perhaps not without value.

As to the movements of the rocks, presumably crystalline, that constitute the foundation of the Marthas Vineyard section, the conditions are substantially the same as those which have existed beneath the Carboniferous deposits of the Narragansett Basin. When they came to be folded, with a measure of compression quite like that which has affected the beds with which we are now concerned, as we may see when the folded Carboniferous beds have been stripped away from the hypogene rocks, the yielding of the crystallines appears to have been made mainly by the interstitial movements of those rocks, and not to any great extent by faulting. All the evidence we have goes to show that while ordinarily massive crystalline rocks may be folded, as is sometimes indicated by the dikes they contain, a frequent method of accommodating themselves to pressure is by squeezing. This action is, indeed, common enough in all rocks which have been subjected to compression strains, as is shown in the distortion of fossils. The absence of any distinct indication of recent faulting on the mainland near Marthas Vineyard affords some support to the supposition that the giving way of these basement rocks was rather more by interstitial movement than by dislocation.

It should also be said that the peculiar position of the masses of decayed crystalline rocks which, as above noted, occur in the three more southern States of New England, long ago led me to the supposition that, after this decay had been effected, a certain amount of faulting occurred, which lowered wedges of the disintegrated rock down to levels to which the surface actions had not penetrated. Moreover, efforts which I have made to account for the details of the topography at several points in southern New England have led me, quite independent of the problems of the Tertiary dislocations, to the idea that at a time not long before the last Glacial epoch there was a certain amount of disruption by faulting in that part of the mainland. This problem of recent faulting on the mainland needs more study than I have been able to give to it; the matter is only suggested here to show that there may be more extensive evidence of orogenic action on this part of the continent than has hitherto been supposed.

A part of the difficulty connected with the question as to the nature of the movements involved in the dislocation of the Marthas Vineyard section arises from the fact that the development of these folds and faults has apparently taken place in a basin without a border of harder rocks on either side through which the compression strains might have been carried. It seems likely that the contraction of the basement beds, however great, would have failed effectively to compress a thick mantle of soft material unless it had been in a basin. The natural result, if the beds lay on a sloping floor without a rim, would appear to be a mere slipping of the bed rocks upon the softer materials, without any such folding as we find. There are obviously no marks of a basin structure made up of the older rocks in the district where these disturbed beds lie. It is possible, however, though there is no evidence whatever to support the suggestion, that there may be a rim of such older rocks lying below the level of the sea.

There is, as is generally admitted, good reason to believe that this portion of the continent stood, in the period extending from the close of the Trias to the beginning of the Cretaceous, at a much higher level than it does at present. During that time a broad river basin may have been excavated in which the Marthas Vineyard Cretaceous and Tertiary were laid down. I have elsewhere called attention to the fact that the Carboniferous beds of the Narragansett district and other local accumulations along the Atlantic coast appear to have been formed in drowned river valleys, and that the beds have since been subjected to mountain-building actions, the trends of the resulting folds having often departed widely from the trend of the neighboring axis of the Appalachian system. I am disposed to think that the Tertiary and Cretaceous beds of Marthas Vineyard have had the same general history, but the fact must be recognized that the evidence to support the conclusion is defective. All that can be said in its favor is that it is consistent with the basin-like origin and structure and the nature of the folding which characterize the localized mountain-built areas of the Atlantic coast.

We now come to the subject of the erosion phenomena of this district, and here we find ourselves in an interesting but difficult field. Beginning with the island of Marthas Vineyard—Capawok, as the Indian name has it—a district which gives us the most information concerning the structure and history of the field, we find there good evidence that the Tertiary and Cretaceous rocks have been subjected to a great amount of erosion since they were dislocated. On the western half of that island these beds rise to a height of about 300 feet above the present level of the sea. As there is an area of about 30 square miles where the crests of the divides have about the same elevation, there is good reason to believe that the existing topography was carved from a surface which, by base-leveling, or more likely by marine erosion, had been brought to an approximate level before the last Glacial period. If I am correct in supposing that there had been a tolerably complete base-leveling or benching with reference to a sea level above the tops of the present divides, then the development of the present topography began long after the close of the mountain-building work.

The pre-Glacial topography of Marthas Vineyard has been but little disturbed either by glacial erosion or by the resulting drift coating. We can see that the course of the principal brooks—rivers, as they are locally termed—has been determined in general by the strikes of the beds coinciding therewith, while the smaller water courses cut across the folds in a normal way. The beds being all of slight hardness, the topography is smooth, but here we find sharp and continuous ridges which owe their relief altogether to differential erosion. These serve to show us how slight has been the wearing effected by the glaciers of the last ice epoch. The absence of such ridges on the remnants of the ancient upland plain also indicates that this plain was due to some action which wore it to a tolerably perfect level.

So well has the pre-Glacial topography of this island been preserved through the accidents of glacial invasion, that we can not only trace two or three cases of ancient stream robbery, but a close inspection makes it evident that all of the brooks of considerable size follow at the present time the channels they had before the ice came. In only two cases have I found that the morainal or other accumulations have changed in an important way the course of the waters. I note these points in order to show that the evidence from these streams as to the general drainage of the district is of value.

Taking the distribution of the brooks of Marthas Vineyard, we note that they are divisible into two groups—those which, turning south, fall into the broad ocean, and those which, descending from the northern side of the island, enter Vineyard Sound. The first-named group of brooks gives us little information except that they enter the sea through what appear to be drowned valleys, and are therefore evidence that the level of this land has been materially lowered since the existing topography was formed. The streams of the northern shore exhibit even more distinctly the same feature of drowning at the mouths, though this is marked, not by their entrance into lagoons, but by the filling of their channels near the sea level by moving sands. On this northern shore also we find in the distribution of the brooks the suggestion that when the land was at the level at which the river topography of the district was developed they entered a large stream occupying the central part of the broad valley now covered by Vineyard Sound.

The general structure of Vineyard Sound is easily misconceived. It has been suggested by Mr. Clarence King that the long range of the Elizabeth Islands, which form the northwestern boundary of this water body, is essentially morainal. There is undoubtedly a covering of morainal drift on the top of these islands, but on examination they prove to be composed mainly of beds similar to if not contemporaneous with the Nashaquitsa section. They probably contain also some part of the Gay Head Tertiary beds. There are no good exposures, but enough is shown to make it clear that the Tertiary portion of that section is above the level of the sea along this line of islands. This condition of a sheet of moraine capping a divide is seen also on Marthas Vineyard, where the broader valleys are in their lower parts almost

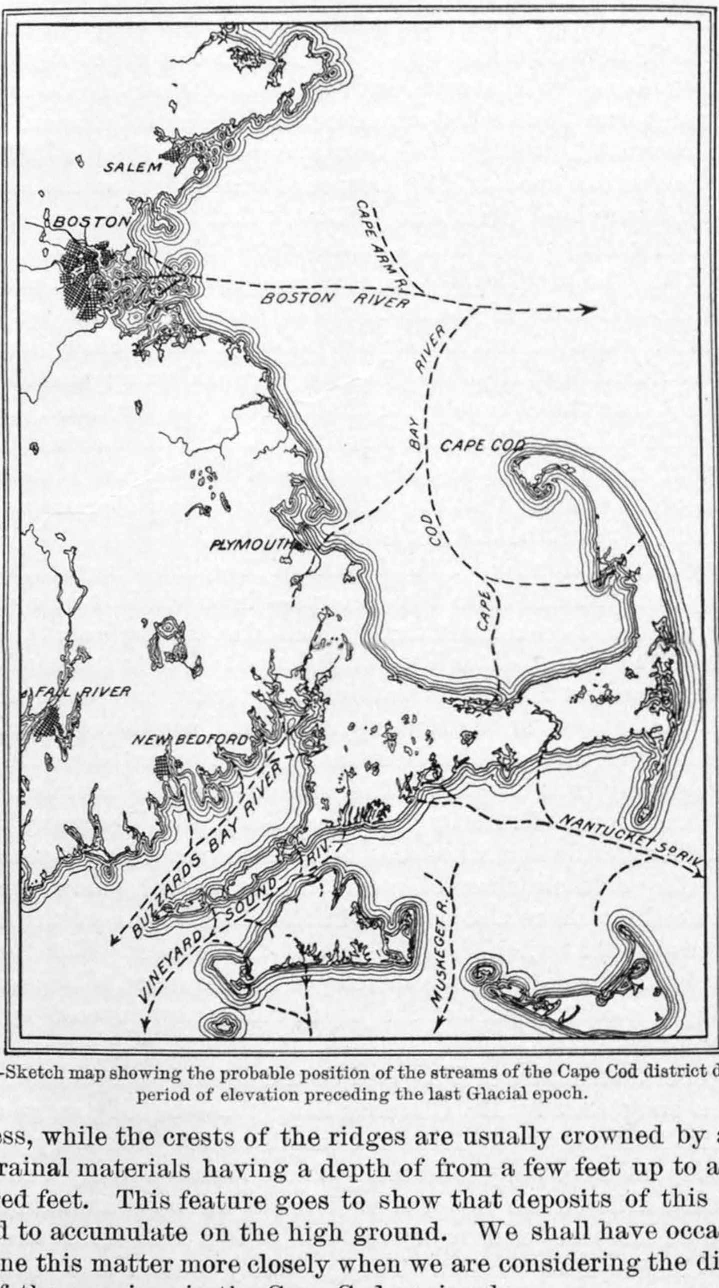



FIG. 86.—Sketch map showing the probable position of the streams of the Cape Cod district during the period of elevation preceding the last Glacial epoch.

driftless, while the crests of the ridges are usually crowned by a layer of morainal materials having a depth of from a few feet up to about a hundred feet. This feature goes to show that deposits of this nature tended to accumulate on the high ground. We shall have occasion to examine this matter more closely when we are considering the distribution of the moraines in the Cape Cod peninsula.

The fact that the Elizabeth Islands are not essentially morainal greatly increases the probability that the valley of Vineyard Sound was excavated by river action. The stream which occupied it during the period of elevation, when the erosive work was done, appears to have had its source on the southern side of Cape Cod. On the west of the Elizabeth divide, in the valley of Buzzards Bay, there was, if this conception of the history of the district be correct, another river which headed in the region about Wareham, taking about half its drainage from the district underlain by the ancient rocks of the mainland. In both these valleys we find the general features of drowned valleys exhibited quite as they are shown in the more southern bays of the Atlantic coast. The basins slope to the seaward, but not in a perfectly regular manner, for the reason that they are much encumbered by drift accumulations and by the waste that has been rearranged by the strong tidal currents which sweep through the bays and sounds of this district. They both widen to the seaward, as we should expect them to do if they had been excavated by the action of land waters.

The position of a third stream is perhaps traceable on the Muskeget Channel, which separates Marthas Vineyard from Nantucket. This headed against Cape Cod and against the upper tributaries of the Vineyard River, as we may call the stream which occupied the sound of that name. It is likely that to the inosculation of these headwaters we owe the formation of the channel which now separates the islands last mentioned from the peninsula of Cape Cod. In the sketch, fig. 86, the conception of the drainage of this district as it was before the last great upward movement of the sea is indicated. In such a figure it is inevitable that many features which are highly conjectural should be shown along with those which are well supported by evidence. In this case the doubt which is the most serious attaches to all that relates to the channels between the eastern end of Marthas Vineyard and the western side of Nantucket and between the last-named island and the mainland. As the tidal currents which flow through this water way are strong, some part of the erosion may have occurred both before and after the last Glacial period, at those stages of elevation and subsidence when the sea was free to pass through these channels. There is also a question as to the nature of the submarine ridges which so abound in these waters.

As regards the shoals of Vineyard and Nantucket sounds, it may be said that some of them, particularly those in the eastern portion of the last-named sound, are of moving sands, and, therefore, may have no relation to the continental topography. Some of these submerged ridges are more reasonably to be considered as preexisting, though their shapes may have been modified by tidal currents. Thus the long shoal, known as the "Middle Ground," which extends along the north shore of Marthas Vineyard from near the west chop of Holmes Hole halfway or more toward Gay Head is, as we may judge from the soundings, a bit of submerged land topography. Although it is now the line of a strong division of tidal currents, and is consequently the seat of a "rip," this perturbation in the movement of the water appears most likely to be the consequence and not the cause of the elevation. So, too, in the case of the shoals to the northward and eastward as far as near Monomoy, there is nothing in the tidal movements which are competent to produce them, though the resistance which they offer to the movement of the currents has doubtless served to effect changes in their forms. If the statements of those fishermen and pilots who know these waters well may be trusted, these submerged ridges often contain on their surface considerable bowlders, which, if true, indicates that they are not in most cases the products of current action, but were formed mainly by subaerial agents of erosion.

It is to be noted that the channels south of Cape Cod to the west of Monomoy Point have in general a definite topography, characterized by steep slopes from the neighboring shores. This form of bottom seems to me inconsistent with the supposition that any great amount of sand is in the possession of the currents along these depressions. It is also noticeable that there is little trace of shifting sands along the shores on either side of Vineyard Sound. Furthermore, it is to be remarked that these valleys, as is shown by the protraction seaward of their very definite land slopes, have not been cut back on the average more than from 500 to 1,000 feet since the shore came to occupy its present level. All these considerations lead me to believe that the floor of these basins is not occupied to any great extent by drifting sands. For the reasons given above, the shoals to the east of Monomoy have been in general regarded as evidence of minor divides formed in the great submerged valleys.

The oscillations of sea level in this region have been more than once referred to in the preceding pages of this report. We have now to review the evidence, with a view to formulating it in a definite manner. It should be noted that there is in this district little, if anything, in the way of ancient beaches to afford data as to the altitude of the land in the periods which are under consideration.

On the mainland to the northwest of this region there are evidences of a base-level of river erosion or of marine planation, which Professor Davis and others regard as of Cretaceous age. The portion of this level at about 400 feet above the present shore line possibly corresponds with the present summit of the Cretaceous deposits of this island nearly enough to warrant the supposition that the sea stood at a height of some 400 to 600 feet above its present position when the lower Cretaceous of Marthas Vineyard was laid down. It should, however, be noted that those beds, owing to their dislocation by mountain-building action, may have been moved either above or below the general plane on which they were deposited.

It is evident that the lignitic portion of the Cretaceous beds was laid down rather above than below the sea level, while the deposits containing marine fossils were formed below the plane of the sea. There is thus evidence of shore swaying in this portion of the section. As yet it has not been clearly determined which of these two elements of the Cretaceous lies the higher. The facts show, however, that in this part of the formation the shore was near its present level, and that it was instable.

Between the lower Cretaceous and the middle Tertiary there is a great blank, which includes the uppermost Cretaceous and the Eocene. As yet, the much-disturbed condition of all the beds showing the contacts of those horizons makes it impossible to say what measure of unconformity existed between them when they were laid down. It seems probable, however, that no mountain-building action had taken place in the district during this interval.

It is not yet perfectly certain that the middle Tertiary strata of this district were deposited in salt water. The marine fossils contained in the beds are found under conditions that admit of the supposition that they were not living when the strata were formed, but were swept in from previously existing deposits. I am forced to regard the determination of the age of this section as in some measure uncertain, but it is clear that it is newer than the Eocene and older than the Pleistocene. The general nature of the beds is most consistent with the supposition that they were formed in an estuary. Assuming that they were made at or about sea level, we should have to conclude that there had been no great change in the position of the shore line between the lower Cretaceous and the Miocene periods, or, what is more likely, that there had been a return of the seashore to about its same altitude in relation to the land after whatever oscillations it had undergone in this long interval.

In the Pliocene, as is shown by the fossils contained in the small locality, now destroyed, at the top of the Gay Head cliff, it is evident that, for a time at least, a shallow sea lay over the surface of the Tertiary beds, which were still in their horizontal position, for these Pliocene beds were evidently involved in the mountain-building movements. It can not be inferred that the altitude of these fossil-bearing Pliocene beds above the sea (about 100 feet) is evidence of a general upward movement of the shore, for the reason that the change of level may have been due to the folding of the strata.

The deposits of the Nashaquitsa series apparently indicate the existence of the shore line at least 100 feet below its present altitude. These beds may be regarded as closing the Pliocene record, and as formed, in part at least, before the orogenic movements took place.

After the series of constructive processes above noted had been accomplished, the beds of this district appear to have been established at a level some 200 or 300 feet lower in relation to the sea level than their present position. During the time in which they occupied this inferior level the upper base-level or bench of Marthas Vineyard probably was formed. It may, in passing, be remarked that the general topography of the bottom of the sea, from the southern end of Nantucket Shoals to Nova Scotia, is in favor of the supposition that we have in this district a surface that preserves in a general way the contours impressed upon it by subaerial erosion.

The down-sinking of this region some time during the Glacial period probably brought about the drowning of the great valleys. It evidently resulted in a lowering of the land at least 100 feet below its present altitude, as is shown by the fact that the morainal aprons or sand plains which are so conspicuous a feature in Marthas Vineyard, Nantucket, and Cape Cod attain about that altitude. These aprons are, as elsewhere noted, composed of sand and gravel, with occasional bowlders of considerable size, which evidently attained their present sites by ice rafting. They have, however, a characteristic submarine topography, such as could not well have been made by any form of subaerial action. The "scour ways" noted in the report on Marthas Vineyard[1] as existing on these aprons are of themselves sufficient to establish the presence of the sea over these plains.

As yet there is no evidence concerning the upper limit of this submergence, which occurred in and possibly before Glacial time. After a careful search throughout southeastern New England, the shore line of the sea in the time when the morainal aprons were formed has not been found. Here and there, at the height of about 200 feet, there are what may be faint traces of a coastal shelf, but they are too indefinite to afford any clear evidence of such a line. The distribution of the drift in the southern part of Marthas Vineyard, in the towns of North Tisbury and Chilmark, is such as to suggest that the glacier did not, save in certain small tongue-like projections, extend south of Tisbury River, and that the drift south of that stream was all rafted to its present site. If this view be confirmed by closer study of the deposits, it will affirm the hypothesis that the whole of the island was under water, for the materials which appear to have been thus transported are found on the very highest land, at an elevation of 300 feet above the sea.

I have elsewhere[2] endeavored to show that on the southern face of the hills of Mount Desert, Maine, we have good evidence of a depression at the close of the Glacial epoch amounting to at least 1,100 feet, and possibly extending up to the highest summits, or 1,527 feet. It seems likely that a depression of even the lower of those levels on the coast of Maine would have involved a submergence in the region of Cape Cod sufficient to have covered the highest lands in that vicinity. It thus appears probable that the Glacial submergence of this district carried the whole of its area below the level of the sea.

The emergence of the Cape Cod district from the Glacial depression must have been very rapid, for the surface appears substantially as it

[1] See Seventh Ann. Rept. U. S. Geol. Survey, 1885–86, p. 316.

[2] See Eighth Ann. Rept. U. S. Geol. Survey, 1886–87, p. 1009 et seq.

was left by the ice. The delicately molded topography of the drift has not been effaced by wave action, nor, as before remarked, are there any traces of ancient beaches. At first sight this absence of any effect of the waves on the surface seemed to me clear evidence that the area could not have been under water since the disappearance of the glacier. In order to determine this point, I visited the region below the elevated beaches in central New York, and found there the same unaffected conditions of surface which exist in southeastern Massachusetts. The conditions of submergence and emergence must have been approximately the same in both areas. It appears, therefore, that we have to accept the conclusion that the uprising of the land after the Glacial period was so sudden that the waves and shore currents did not have time to do effective work. It is possible, however, that the waters were at this time so far obstructed by floating ice that no great amount of wave action took place during a considerable length of time in which the elevation was going on. The last change of level of the Cape Cod district evidently brought the land somewhat above the plane at which it at present stands. This is shown by the occurrence at various points of submerged forests, as in Nantucket,[1] and in the marshes bordering the harbor of Holmes Hole, Marthas Vineyard. The amount of this recent downward motion is not known, but it may have been sufficient to obliterate the land connection which united the islands of Marthas Vineyard and Nantucket with the mainland. The reason for supposing a connection between these islands and the mainland is found in the substantial identity of their faunas and floras with those which exist on the neighboring continent. Considering the width and current-swept nature of the sounds which separate those islands from the mainland, it appears unreasonable to suppose that all these species of animals and plants could have found their way to the outlying stations in the short time which has elapsed since the Glacial period. The conditions of passage are almost as difficult as they are at the straits between the islands of Bali and Lombok, which separate the biological provinces of Australia and southern Asia. As there has been a subsidence since the forests regained possession of these islands, it is reasonable to suppose that the previous elevation was great enough to bring about a connection with the mainland.

It is to be noted that the changes in the relative elevation of sea and land have all been spoken of as if they were due to changes in the altitude of the land. It should be observed that this assumption is incorrect. There are at least two main and efficient causes of alteration in the position of the sea in relation to the land, as well as many others of minor importance which probably have some value. One of these, local in its nature, is the swaying of the land against which the sea lies; the other is change in the form of the sea bottom affecting the height of the open waters along all the shore. It is rarely possible to differentiate these causes. In the case of the Iroquois beaches in central New

[1] See Bull. U. S. Geol. Survey No. 53, p. 28.

York, the fact that they rise to the north supplies a criterion, the like of which may be found in other places, which indicates that the post-Glacial elevation was due to a land movement. Again, as I have elsewhere endeavored to show,[1] the very general drowning of the lower valleys of the rivers of all the continents affords evidence that there has been, in very modern times, a general rise of the sea level to the amount of 100 feet or more. It is to this general inundation that we may perhaps attribute the destruction of the land bridge which for a long time after the close of the Glacial period united the New England islands with the mainland.

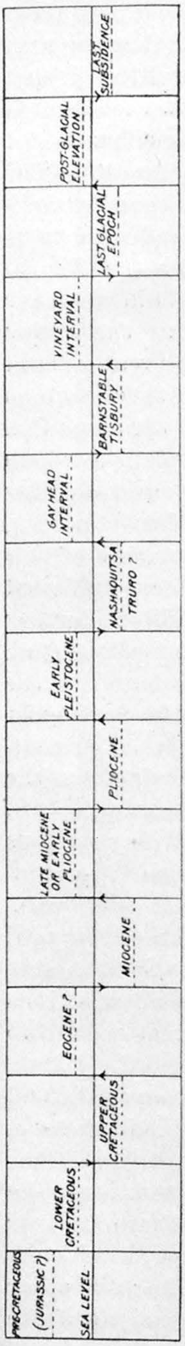



FIG. 87.—Diagram showing the probable movements of the Cape Cod district since the Jurassic period.

There is some evidence as to the duration of this land bridge to be derived from the possible rate of marching of the oaks and other heavy-seeded trees northward from their southern refuge during the last ice time. These trees, in the process of repossessing an abandoned field, do not, according to my observations, advance at an average rate of more than 5 feet a year. To assume a rate of 10 feet a year would be to allow the utmost that could be supposed. This would require about 500 years for a mile of journey, or about 2,000 years for the passage across a land bridge from the mainland to the island of Marthas Vineyard. I have elsewhere urged this slow rate of northward march of the heavy-seeded trees as an argument against the hypothesis that the close of the glacial advance, when the ice lay at the southernmost point it had attained, was not more than from 10,000 to 20,000 years ago. If the argument be valid, the return of the oaks to southern New England, after their expulsion to regions farther south, must have required somewhere near 200,000 years.

It is possible that some of the topographic features of the submerged channels of this area are due to the erosion which took place during the elevation that followed the Glacial period. The cutting away of the divide between what we have supposed to be the drainage basins of the Vineyard and the Muskeget rivers may have been in part effected by the energy of the tidal currents, which, in case the passage were diminished in sectional area, would act with something of the strength they now have in passing from Vineyard Sound to Buzzards Bay through the shallow passage known as Woods Hole.

[1] See Bull. Geol. Soc. Am., 1895, p. 153 et seq.

We have next to consider the erosion performed in this district by the ice of the Glacial epoch. It is now well known that the original conceptions as to the amount of this wearing on the general surface of the country were very much exaggerated. There are few students of the phenomena in the field who would be disposed to believe that the average of this erosion in New England amounted to as much as 100 feet. I doubt whether it was as much as 50 feet. It has commonly been supposed, however, that in incoherent materials the action was more effective than in firm-set, highly changed rocks. So far as we can judge from the conditions seen on Marthas Vineyard, where the contact of the drift with the underlying sands and clays is well revealed by the cliff sections and the clay pits, the glacier did little more than smooth over the soft materials without effacing their original outlines. In no case is any considerable amount of the characteristically colored clays and sands mingled with the till covering. It is usually impossible to find a trace of them at a height of 2 feet above bed rock. When we consider that this occurs at points where the ice has journeyed for a mile or more over the soft beds the absence of all marks of erosion is seen to be very remarkable.

I have already noted the fact that the glacier failed to destroy even the minor features of the topography of this field. It is difficult to exaggerate the extent to which the pre-Glacial topography survives in the Vineyard area. The facts must indeed be seen, and this carefully, to be appreciated. There is one way, however, in which the glacial conditions effected a considerable amount of local erosion in this district. This was by the action of the subglacial streams which flowed upon the surface of the bed rocks. All along the northern face of Marthas Vineyard we may trace these subglacial river ways, which extend from the shore in winding, sometimes beautifully curved, channels cut down into the soft rocks to where they discharged beyond the front of the ice. Sometimes these channels are coincident with earlier-formed valleys. Again they are carved where it is evident that open-air streams have never flowed. The largest, and in some ways the best, example of these interesting and elsewhere unnoted features is seen at Chappaquonset or Tashmu Pond and in the valley which continues the depression to the southward. As elsewhere noted, these grooves, extending from the shores of Vineyard Sound upward, terminate on the south in the morainal apron, or, rather, are continued over that apron in other shallower and wider troughs cut in the sands which extend downward to the ocean shore. In some cases the bottoms of the grooves which were under the ice have been worn down to the depth of 60 feet or more into the soft bed rocks.

It is likely that the low places in the Elizabeth Islands divide, as at Woods Hole, Quicks Hole, and Robinsons Hole, as well as certain others which do not go below the level of the sea, are due to a like action of subglacial streams when they crossed the ridge lying between the sound and the bay. Similar valleys will be noted in the detailed account of the structure and topography of Cape Cod.

We have last to consider the extent of the post-Glacial marine erosion which has occurred in this district. This has been and is still great; at present it is some hundredfold greater than that done by land waters, but the distribution has been in certain regards peculiar. As before noted, the marine benching of this district which occurred after the deposition of the Pliocene strata was apparently pushed to the point where the whole area of the Cretaceous and Tertiary rocks was brought to a level. This may possibly have been due to base-leveling by stream action, but whoever will observe the practical absence of such work on the strong topography of Marthas Vineyard, where the brooks are never colored by the soil waste and where there is not a single stream scar, will hesitate to hypothesize this slow process of lowering to a common level such rocks as here occur. There are on Marthas Vineyard pretty clear evidences of the existence of a second level of marine benching at the height of about 150 feet above the sea. This bench is recognizable around nearly the whole of the western elevated section of the island, and as there is no structural basis for it it must be regarded as of marine origin. Its original width can not now be determined with accuracy, for it is much diminished by marine erosion, but on the average it can not well have been less than one-third of a mile, and may have been more than twice as great as that amount. Thus the pre-Glacial erosion due to the sea apparently includes two large operations, the formation of the original level surface of the island and the cutting of the lower bench, which was a much less extensive work. As to the benching which may have been done in the sections below the water level, the evidence, though in a way interesting, is too perplexing to warrant discussion.

The post-Glacial marine erosion of this district has been extensive and is now in process of very rapid development; no other portion of the coast of North America is undergoing such a complicated and rapid readjustment. When the problems which are there presented concerning the action of the sea on the shore and the arrangement of the detritus derived from the process of erosion are worked out, this area will become the classic ground for students of coastal action. For our purpose it will be well to divide the question into two heads, the first relating to the amount of erosion already done, the second concerning the manner in which the work is effected.

In this, as in most other shore lands, the only possible way of determining the extent to which the sea has gained on the land is to take the slopes which extend downward to the sea and ascertain where, if protracted, they would cut the water level. This method is particularly applicable on shores such as those of southeastern New England, where there is a gentle slope toward the sea. On this basis, using for data observations made at some scores of points, I have come to the conclusion that if the land had been stable since the last ice time the average retreat of the shores might safely be estimated at rather more than about one-half a mile. Unfortunately for the sufficiency of this method, it is certain that there has been a subsidence, which may be, and most likely is, now going on, so that the resulting work of the sea is not the formation of a horizontal shelf but of a sloping scarf, much of which is submerged. Thus we can only say that, the rate of the down-sinking being unknown, we are deprived of any means of accurately fixing the extent of the incutting since Glacial time, and are driven to the ruder method of noting the amount of recession of particular cliffs.

Turning to the evidence afforded by the cliffs, the best obtained is that due to the surveys of the late Prof. H. L. Whiting, long the senior assistant of the United States Coast and Geodetic Survey, who fixed the rate of erosion of the Nashaquitsa cliffs by very careful observation for a period of fifty years at 3 feet per annum.[1] The evidence is clear that this is the rate for nearly if not quite the whole southern shore of Marthas Vineyard. At Gay Head there are only approximate data, which serve to show that most likely the retreat does not amount to as much as a foot per annum. Along the north shore of Marthas Vineyard the process of erosion is very slow, save at a few salient points west of the steamboat landing. At West Chop the rate does not average a foot in five years, but at the east and west chops of Holmes Hole the cliffs have for some years been retreating at an average rate of at least 2 feet per annum. This wearing is probably in some way connected with slight alterations of the shoals which direct the tidal currents against the shore. On the Cottage City or eastern face of the island the rate of wasting is also great. The recession of the shore has amounted to at least 30 feet in fifteen years, and this despite some slight efforts made to resist the action of the waves. The region near Edgartown is amply protected on its east side by the extensive system of hooks about Cape Pogue. As a whole the shores of Marthas Vineyard in process of erosion, excluding the island of Chappaquiddick and the Cape Pogue hooks, are probably entering the land at an average rate of about a foot a year. The mean height of the sea-cliff face may safely be taken at 30 feet, and the total face subjected to erosion at 35 miles. This would make the quantity of material removed amount to a total of about 1,000,000 cubic feet per annum.

On the island of Nantucket, owing to the extent of the sand-barrier beaches, the proportion of the total shore line which is exposed to active erosion is less than that of Marthas Vineyard, but the wearing action of the sea is much more effective, for the reason that clays rarely appear in the escarpments, which are mostly of stratified drift, such as is found in the morainal aprons. The southern shore of the island and a portion of its northeastern face are apparently retreating at the rate of

[1] Geology of Marthas Vineyard: Seventh Ann. Rept. U. S. Geol. Survey, 1888, p. 361.

more than 4 feet per annum. Yet, for the reason above given, it seems likely that the average encroachment of the sea is much less rapid than it is in the island of Marthas Vineyard.

The Elizabeth Islands are wasting for the greater part of their length, but the process is now being arrested by a simple action which has brought protection to much of the shore lands of southeastern Massachusetts. The mass of these islands is, as before noted, of incoherent sand, but the surface is generally occupied by a layer of coarse till or moraine, having a depth of a few feet. As this pebbly and bowldery matter falls to the shore it forms a stony beach. This plating over the soft underlying beds is sufficient to prevent the shore currents from wearing them away. The result is that a platform is made on which the waves break before attaining the shore, and often a barrier beach is formed which to a great extent keeps even the swash from attaining the cliffs. When the adjustment has gone thus far, the shores erode only so fast as is necessary to supply the place of the pebbles which are worn out or the larger waste brought about by the action of the shore ice in rafting away the stones, as it does in a very effective way. An excellent example of the value of these conditions in hindering marine erosion is shown at Gay Head, where, despite the ease with which the strata slip downward into the sea, the vigor of the assault of the waves, and the complete and rapid removal of the sands, the retreat of the escarpment is slower than that of many other less exposed shores on this part of the Atlantic coast line. The bowldery drift, though not large in amount, is enough to have formed a shelf extending irregularly out from the face of the cliffs to the distance of nearly a mile. On this the heavier seas break, so that when running from their prevailing direction only the secondary waves attain the shore, with so little effect that the retreat of the face is at the present time less than a foot per annum and appears to be rapidly diminishing in its rate. On the other hand, the Nashaquitsa cliffs, which in their retreat contribute but little bowldery material to the sea, are, as above noted, retreating at the rate of 3 feet a year, the sea having no difficulty in deepening the bottom as it works in, so that its waves are able to assault the base of the cliffs.

On the peninsula of Cape Cod we find evidence of marine erosion essentially like that on the islands which lie to the southward. The details of this action will be noted in the section of this report which is devoted to the topography of that area, but the general features may well be considered here. The most interesting point is that probably all of the invasion of the sea occurs on the southern and eastern (or outer) part of the peninsula, there being little trace of it on the northern (or inner) shore. This is in part for the reason that the seas strike in times of heavy storm with greater effect in this portion of the coast, but in larger measure it is owing to the fact that strong tidal and shore currents sweep by this part of the coast, which carry away the débris delivered to the sea by many of the cliffs, so that it does not encumber and protect the shores.

On the northern shore of Cape Cod the surface has, as will hereafter be noted, a long riding slope of the glacier, composed usually of clay, which descends gradually to the sea level. This slope extends from the western border of the town of Barnstable to Yarmouth and is partly indicated as far as the town of Brewster. Where this gentle declivity passes beneath the sea, or where it attains the fit depth of water, a beach is formed which incloses the great marine marshes that are so prominent a feature of this part of the coast. This beach is slowly working inland, but the amount of sand which accumulates in the bay it faces is so great that the excavation of the bottom necessary to the inward march of the beach hinders the movement.

On the north shore of Cape Cod the distribution of the products of coastal erosion indicates the weak action of marine currents. On the south side, in the waters between Hedge Fence Shoal and the open sea to the eastward, the distribution of the shoals and spits shows a considerable amount of movable débris in the possession of the sea and its conveyance by strong currents. In the field about Monomoy the struggle between the accumulating sands and the currents is so active that there is evident danger of the passage to the seaward between Nantucket and the cape being closed before many decades have elapsed. If in any time of great storm this channel should become so shallowed that the waves would break across it, the result would be the immediate construction of a barrier beach. If this construction failed to attain to or near the surface, it would doubtless be swept away by the tidal currents. If, however, it made an effective barrier to their flow the island of Nantucket would be again joined to the mainland. An accident of this nature is possible; it is likely, indeed, to be the next great change in the conditions of this part of the Atlantic coast.

As before remarked, the unstable sands of the bays on the eastern side of Cape Cod appear to be mainly limited to the eastern portion of these waters. In Vineyard Sound the evidence from the shores and soundings indicates that the amount of sand at the disposition of the currents and waves is not large. The shores are generally pebbly and the soundings are not to any extent variable. The harbors, such as Tarpaulin Cove, Woods Hole, Holmes Hole, and Edgartown, though so placed that they would naturally be obstructed by moving sands were large quantities of such materials in unstable positions, show little tendency to fill in. The sand beaches, such as those at Menemsha light and between Sengekontacket Pond and the sea, are evidently not gaining in width. These conditions are in very distinct contrast to those which are found in and about Nantucket Sound and Muskeget Channel, where there is a ceaseless oscillation of the shoals and where the harbors which exist are in constant process of closure, against which, as at Nantucket, the precautions of the engineer seem to be of little avail.

The reason for this difference is not perfectly clear; it is probable that it is in part due to the relatively large amount of waste contributed to the sea by the degradation of the shores of Nantucket and Cape Cod, but in a measure also to the fact that the average run of the tides seems to bear the sands to the eastward to a point where the energy of the Atlantic surges, rolling in from the eastward, tends to beat them back into Nantucket Sound.

It seems likely that something like this peculiar condition which we now find in the shoals of the Monomoy group existed a short time ago in the region of Nantucket Shoals, although as yet we do not know enough of these shallows accurately to determine their history. It is probable that it represents the remains of a system of low islands, shoals, and tidal channels which were depressed beneath the sea at the last subsidence. If we should conceive the shore to be 50 feet higher than it is at present, the struggle of the tides and other currents which now exists about Monomoy would be transferred to the region of Nantucket Shoals.

## UNDERSTRUCTURE OF CAPE COD.

On first inspection the body of Cape Cod, i. e., that part of it which lies between Monument River on the west and the sand spit which sets in just east of Highland light, appears to be made of glacial débris. It is true that nearly the whole surface is covered either by the extensive moraines which are to be described in the next section of this report or by the deposits of sand and gravel which are spread on the south and east of the morainal accumulations. These superficial accumulations are so extensive that they very effectively mask the true character of the underlying deposits. None of the streams form sections which reveal the underlying beds, and the only cliff shores which do this are near Highland light, where the evidence, as will hereafter be noted, is not very indicative. I therefore deem it necessary to give in some detail the evidence which goes to prove that there are large areas of relatively old strata lying beneath the glacial beds of this district and above the level of the sea.

Beginning with the southwestermost portion of the cape district, that which is in the town of Falmouth, we observe in the fields about Quamquisset Harbor a quality of surface which clearly indicates the existence of deposits other than those of glacial origin. The topography is evidently older than the last ice time, the valleys being somewhat encumbered with deposits of drift. Sections through the ridges show beneath the thin detrital coating a series of somewhat indurated sands, gravels, and clays, usually thin-bedded, though some of the clay layers are 2 feet or more in thickness. The sands and gravels are rather ferruginous, and sometimes the iron oxide is sufficient in amount to produce a distinct cementation. The clay beds range in color from whitish, through brown, to distinct reds. The materials and their association are essentially like those belonging to the Nashaquitsa series, as shown on Marthas Vineyard. The pebbly matter is rarely of crystalline rocks; it consists almost altogether of quartz and quartzite. In the places where shown in rather small openings it seemed likely that the few pebbles of a granitic nature had been brought to the ground by the glacier and crushed into the mass by ice action.

The most notable feature in the "Quisset" Harbor section is the considerable dislocation to which the beds have been subjected. The layers are thrown into short, abrupt folds, the resulting dips being at several points as much as 30 degrees of declivity. The strikes are irregular, but, as on much of Marthas Vineyard, incline to a general northwest-southeast direction. The condition of the folded beds, especially the fact that a topography somewhat obstructed by glacial deposits but otherwise undisturbed was carved on them in pre-Glacial time, clearly indicates that here, as on the island last named, the disturbances can not be accounted for by the movement of the ice. The important exposures which have yielded this evidence were made in 1896 by a land company in grading the roads of its property. They are, unfortunately, of a nature to be soon effaced.

The topography and the distribution of the "spring levels" (or places where the water contained in the drift is turned to the surface by the clays) of the region about Woods Hole indicate that this Nashaquitsa series—for such we shall term it—rises to the prevailing height of about 60 feet above the sea, being capped by the ridge of the moraine which runs parallel with the shore of Buzzards Bay. Northward along the shore of that bay the conditions of surface, as explained by the above-noted facts, indicates that essentially the same materials continue to Buzzards Bay, the ancient series being interrupted only by the indentations which are formed by several "drowned" valleys and by Monument River. In a railway cutting just west of Buzzards Bay station there was exposed in 1896 a section about 4 feet deep which showed stratified ferruginous sands that were slightly folded. These beds appeared to belong to the same series as those at "Quisset" Harbor.

Just east of Falmouth the stream beds near the shore at several points reveal by a little excavating the presence of indurated ferruginous sands and gravels of the same type as those found north of Woods Hole. Moreover, the streams that drain from the eastern face of the Falmouth moraine show that the percolating drainage which is normal to a sand-plain country is interrupted by some resisting layers, which hold the water near the surface. If there were not water-turning beds under these sand plains they would, like those of the similar plains of Marthas Vineyard and Nantucket, drain by percolation to the sea.

The facts above noted warrant the hypothesis that the western section of Cape Cod, say for a strip some 8 miles in width, has a foundation of ancient sands, gravels, and clays which rises to a considerable height above the water level, and is, in parts at least, much dislocated.

It is barely possible that the unseen water-holding layers on the eastern side of this area are of till or other clay beds of Glacial age, but the improbability of this view will be made apparent by the account of similar deposits in other parts of the cape.

On the northern shore of the peninsula, from near Monument River eastward to Yarmouth, and less distinctly still farther eastward to Orleans, there is an often indistinct but clearly traceable slope leading upward from the sea level to a height of 60 feet or more. This slope is often more or less masked by local accumulations of till, or even by small ridges of a morainal nature; but wherever its structure is revealed it is found to be made up of a deposit of dark-blue and gray stratified clays. Its presence is generally attested by the fact that it is not penetrable by water, and the fields which lie upon it are quite different in character from those found elsewhere in the cape district. The agriculture of the northern portion of Barnstable County has indeed been to a considerable extent founded on the quality of this underlying material, which affords a much more enduring soil than is found elsewhere in the area.

Occasional wells on this northern slope of the cape, and particularly the brick pits in Barnstable, show this clay to have the thickness of at least 20 feet. At no point, so far as I have been able to find, has it been passed through.

The clay which so generally forms the northward slope of the cape, between Orleans and Monument River, apparently underlies all the characteristic deposits of the last Glacial epoch. Upon it rests a number of small areas of an evidently morainal nature, as well as a general though rather thin covering of till, which appears at some points, particularly at the brick pits above referred to, to be somewhat churned up with the lower clay. Nevertheless, the distinction between the two deposits is sufficiently clear to show that they are only accidentally associated. Although this clay is to a great extent masked by the usually thin coating of drift, it appears to be continued as a tolerably connected deposit from the western extremity of the cape to Orleans, and perhaps still farther eastward. The fields of this section owe their relative fertility in part to the fact that the clay keeps the water table nearer the surface of the ground and in part to the commingling of this clay with the glacial waste, which in this district is distinctly more clayey than it is in other parts of the cape.

The northern clay of Cape Cod does not appear to have been dislocated by compression strains, at least none such have been seen in the scanty sections which are exposed to view. As to the origin of the deposit, the evidence is not yet clear. So far as ascertained, the material contains no fossils. In its general aspect it is like the well-known brick clays of southeastern Massachusetts, which were, in some cases at least, clearly formed at the time of, and in front of, the glacial sheets during the last ice advances. Yet, as no pebbles have been found in the deposit, and as its resemblance to some of the beds of the Nashaquitsa series of Marthas Vineyard, which clearly antedates the last ice epoch, is evident, it will not be safe to class it as of Glacial age. It is, perhaps, the equivalent of the Tisbury clays of Marthas Vineyard.

South of the glacial clays, and generally beneath the moraine which extends from Monument River to Orleans, the evidence, though imperfect, goes to show the existence of another series of clays, which exhibit a general likeness to those noted as occurring in Falmouth near Quamquisset Harbor. In my opinion this ridge of older clays forms the greater part of the considerable elevation, which at first sight appears to be entirely of a morainal nature. The evidence in support of this proposition is as follows:

Along the line of the moraine, which attains at several points an altitude of nearly 200 feet, we find that the depressions of the surface, up to 150 feet, often contain water for a large part of the year. Even where they are not temporary pools, these kettles commonly exhibit up to or above the last-named height a degree of wetness which indicates that they rest upon more impervious materials than the very porous moraine affords. An index of the same nature is to be found in the height of the considerable lakes, of which a score or more are shown on the topographical map of this part of the cape. Thus, Peters Pond, in Sandwich, is about 95 feet above the sea, and a number of the other lakes exceed 50 feet in altitude. In general, it may be said that the lakes in the central portion of the cape, particularly those within the limits of the distinctly moraine topography, stand at heights above the sea which clearly indicate that the barriers which retain the water are much less pervious than the sandy, pebbly, and bowldery matter of the moraine itself, which in this regard is but slightly more effective than the washed drift. The same considerations lead us to extend the clay area much to the southward of the southern face of the moraine. The lakes on the sand plains for some distance out from the face of the moraine lie at heights which exclude the supposition that their waters are retained by the interstitial friction or resistance to percolation which is normal to washed drift. On the sand plains of Marthas Vineyard the value of this friction, as is shown by the depth at which water has been struck in the central portion of the island, is not more than 2 to 3 feet to the mile. On Cape Cod many of these lakes of the plain are at heights above the sea which would afford grades of from 12 to 15 feet to the mile from their low-water mark to mean tide. On this account, as well as from the general statement which is had from all those who are familiar with the history of the wells of the district, that they usually strike clay before attaining the level of the sea, I judge that the central clays of the island probably extend some distance south of the moraine.

As to the character and altitudes of the central or submorainal clays of the northern part of Cape Cod, the evidence yet gathered, though not extensive, is, taken with what has been found near Woods Hole, sufficient to show something of these conditions. In the section between Great Pond and West Barnstable the roads show some small sections of grayish-white bedded sands unlike any glacial beds known to me. The beds dip to the southeast at angles of from 8 to 10 degrees. Traces of reddish-brown clays are revealed in the same district.

In the town of Dennis, at the side of the State road, the cuttings at the time they were made revealed dark clays and gray sands, thinly bedded, resembling the Nashaquitsa series. These beds are folded on a north-south axis, the amplitude of the arches, so far as could be ascertained, not exceeding 50 feet. Although these foldings were not very plain, they were recognized by my companion, who had no special knowledge of geology, as arches having the form shown in the diagram, fig. 88.

In the town of Brewster, about one-half mile east of the station of that name, there is a considerable area (100 acres or more) in which the drift covering is so thin that the under clays are revealed. These appear to be somewhat confused next the surface by the rubbing and scouring action of the glacier. It is thus impossible, in the slight exposures, to determine the exact attitude of the beds, which are of grayish and blue sandy clays and red clays, the last of the general aspect of those at Gay Head. Near Griffiths Pond—now a cranberry bog—a considerable exposure of the red clay on the north side of the road to the station shows not very clearly a rather steep dip to the northwest. Although the scanty showing of these clays in the small pits by the roadside does not afford distinct evidence of steep dips, the distribution of the outcrops indicates such attitudes, with a prevailing strike N. 45° E. These clays rise to about 100 feet above the level of the sea.

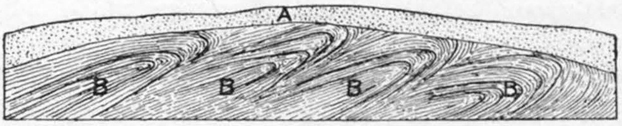


FIG. 88.—Diagrammatic section showing position of folded clays, State road, Dennis. A, glacial drift; B, folds in (Truro?) clays. 1 inch = 100 feet.

From Schoolhouse Pond to East Brewster station the clays, apparently of the same general nature as those last described, lie everywhere near the surface on the south side of the main road. The deposit rises to the height of from 100 to 110 feet above the level of tide. Its upper surface forms a tolerably gentle slope to the northward, on which rests the morainal heaps and into which the kettles seem to be cut, with the result that they are usually very wet at their bottoms. The red or reddish sandy clays appear also, though obscurely, in some of the roads of East Harwich which lie to the south of the section last described.

In Orleans the older series seems to be present thoughout the greater portions of the area of the town. It is scantily revealed in the road cuttings, and is shown in an effective way only on the southwest side of Town Cove, just east of Tonset. Along this shore the deposits are not enough exposed to give a clear idea of their attitude. They are all dark colored and are related to the glacial beds in the manner indicated in the accompanying diagram, fig. 89.

The above-described clays appear to continue for some distance northward into Eastham. The precise line where they cease to rise above the level of the sea has not been determined; it can not be ascertained without pits or borings. It seems likely that this limit is not far from North Eastham station.

On the south side of the moraine there exists a series of clays which are revealed in the ordinary domestic wells that are driven in this section. I have been unable to find any of these wells in process of excavation, and am therefore limited as to information concerning the section to what I have been able to gather from various persons whose

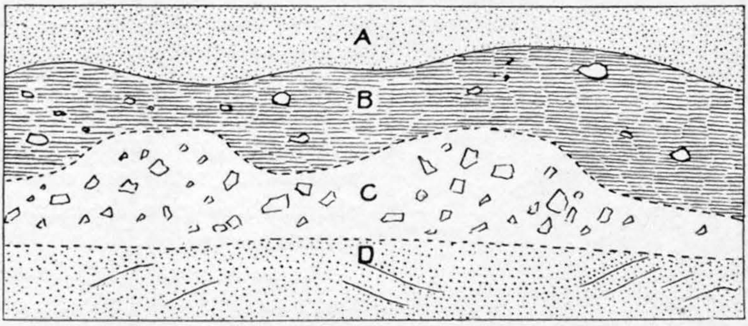



FIG. 89.—Diagrammatic section of post-Glacial clays, west side of Town Cove, one-half mile south of its head. A, glacial sands; B, stratified clays; C, till; D, folded clays and sands. 1 inch = 10 feet.

statements seem worthy of trust. These persons agree that for a considerable distance south of the moraine in Harwich, to within a mile or so of the sea, these wells strike a dark-colored clay ordinarily at a depth of 10 to 20 feet below the surface. This deposit usually has to be passed through in order to obtain water, which is found in thin layers of gravel above a lower-lying clay. The thickness of the upper clay is said to be 10 feet or more; the lower clay deposits do not appear to have been passed through.

The statements concerning the existence of a series of clay layers beneath the sand plain on the southern side of the Cape Cod moraine are in accordance with the evidence afforded by the levels of the lakes in this area. As before remarked, these lakes, especially those which are situated within 2 or 3 miles of the morainal area, have a height above the level of the sea which is inconsistent with the supposition that they are fenced in by no more effective barriers than would be formed of the open-textured sand and gravel of the plain in which they lie. These lakes are, in part at least, to be regarded as occupying old valleys, which slope down to the northward and have been barred across by morainal accumulations of a clayey nature, which have in good part effaced them.

As to the attitude of these clay beds, which lie beneath the whole or a large part of the southern morainal plain of the cape as far out as Orleans, there is no basis for accurate determination. The reports from those who have sunk wells in the area leads, however, to the supposition that the deposits are not dislocated after the manner of the central series, but lie in approximately horizontal attitudes, dipping gently to the southward. It is thus tolerably clear that the fundamental structure of the body of the cape—at least that part of it lying between Monument River and Eastham—consists of a central axis of more ancient clays and sands, having in part at least the general aspect of the Nashaquitsa series of Marthas Vineyard, and being likewise much disturbed by orogenic action. On the flanks of this older axis of elevation lie the clays, which, as before noted, appear to form the north and south slopes of the area, and on which rest the relatively thin layer of glacial waste—the moraines and sand plains which give the surface aspect to the region.

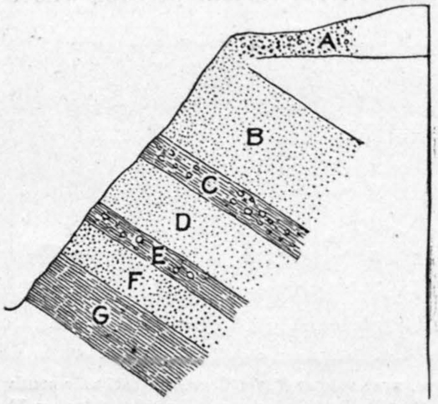
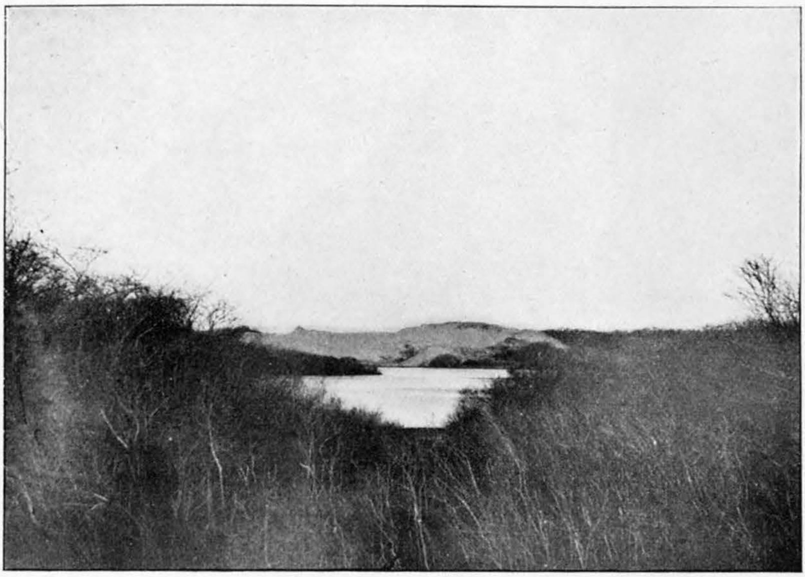
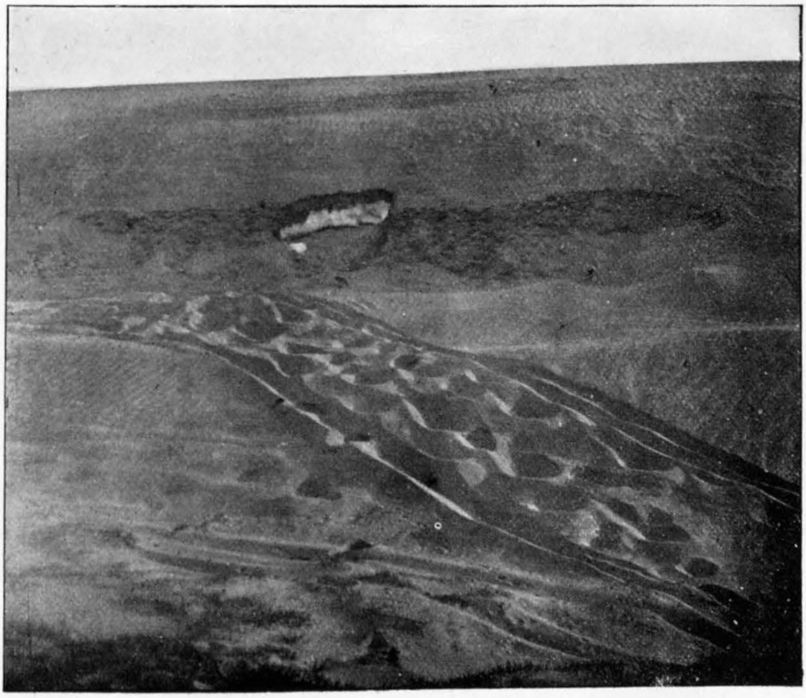


Fig. 90.—Diagrammatic section near Wellfleet bridge, in Truro. A, glacial drift; B, D, F, stratified sands; C, E, pebbly sandy clays; G, fine clay. 1 inch = 5 feet.

The outer portion of the cape—i. e., that between Orleans, or perhaps the northern portion of Eastham, and the extremity of the peninsula—has a structure which indicates a history somewhat different from that just noted. In this outer part the older dislocated beds with the red clay layers do not appear above the water level, or if they rise above that plane they are completely covered by the later-formed deposits of clays and sands. In this area we have a succession of beds, indicated in the accompanying fig. 90, which shows a series of events somewhat different from that exhibited in the district to the westward. In this section, from central Eastham to the end of the highland at Moon Pond and Salt Meadow Pond, the beds consist of certain clays which appear, so far as can be determined by their general nature and relations, to be essentially similar to those beds which were noted as occurring on either side of the Nashaquitsa series of the western district of the peninsula. Upon these clays occur beds of much-decayed sands and gravels, which in character are somewhat like those found in the subglacial portion of the section on the west side of Town Cove in Orleans. Owing to the large showing made by the glacial deposits of



*A*. DUNE POND, PROVINCETOWN.



*B*. NEAR VIEW OF DUNE SURFACE, SHOWING MASS OF BURIED SNOW.

this part of the cape, this series of sands and clays has been assumed to be the product of the ice age. Although the question as to the age of the several divisions of rocks of this field is in the main to be dealt with in a later part of this report, it may here be said that there is good reason to doubt whether the beds shown in Wellfleet and Truro should be reckoned as of glacial origin, at least in the sense that the sand plains in the more eastern section are to be so reckoned. They have exhibited no glaciated pebbles; they lack the surface slope so characteristic of morainal aprons, and they fail to exhibit the occasional large ice-rafted bowlders so common in such deposits.

From the northern end of the highland of the cape to the extremity of the peninsula the land is, so far as its surface is concerned, made up altogether of sands, which have been brought into their position by the recent action of marine currents or of the wind. As has been shown by Professor Davis, the form of the slope which terminates the elevated ground toward Salt Meadow Pond indicates marine erosion before the outermost part of the cape had been built. As will be hereafter noted in more detail, this agglomeration of sand hooks and spits most likely rests upon a portion of the land which had been cut away by a set of currents different from those now prevailing on this shore.


FIG. 91.—Diagrammatic section across Cape Cod from West Barnstable station to Osterville, showing the general structure of the area west of Orleans.

The construction work involved in the formation of dunes is admirably shown in this portion of the cape. By the exercise of a certain amount of care in planting, the local and State authorities have succeeded in arresting the movement over a large part of the area, but the seaword portion of it is still in constant motion. The speed of this movement may be judged by the fact that in April, 1897, a mass of snow 20 feet in length and 2 feet in thickness was revealed where it had been covered with sand during the preceding winter to the depth of 12 feet, the mass having been subsequently cut through by a change in the scouring movement of the wind. (See Pl. XCVII, *B*.)

The irregular deposition of the dunes has led to the formation of a number of small lakes, which, though of no geological significance, are very picturesque. They are generally bordered by a fine growth of scrubby trees, nourished by the moisture they afford, while beyond this fertile margin rise the desolate slopes of sand. (See Pl. XCVII, *A*.)

## HISTORY OF THE CAPE COD SERIES.

As already indicated, the beds exhibited in this area may, so far as they have been interpreted, be provisionally divided into five groups, of very unequal value as regards their extent or the time occupied in their formation. These groups are, in order of age, as follows:

First and lowest, the series of gravels, sands, and sandy clays which, on the basis of general aspect, are here reckoned as the equivalent of the Nashaquitsa series of Marthas Vineyard, and which, as on that island, have been subjected to a considerable amount of stressing.

Second, the dark-colored clays which are revealed at the brick pits in West Barnstable, at the base of the section on the west shore of Town Cove, at the base of the section at Highland light, and at various other points; these are known as the Barnstable series. These pits are occasionally much filled. Their position in relation to the other groups remains somewhat doubtful.

Third, the sands and clays characteristic of the Wellfleet and Truro district, found along the shore northward to Plymouth Harbor, and probably northward to Egypt or Coleman Heights, in Scituate. It is not unlikely that remnants of these beds occur in other portions of southeastern New England.

Fourth, the glacial deposits, including the morainal accumulations, the eskers, and the sand plains which lie south of the moraines.

Fifth, the beds formed since the Glacial period, consisting of dunes, spits and hooks, submarine coast shelves and shallows, and the organic deposits of swamps and marshes.

These five groups of deposits will now be considered from the point of view of their geological history.

## NASHAQUITSA SERIES.

The identification of this series, as exhibited in Cape Cod, with that found at the typical locality on Marthas Vineyard, rests altogether upon the general, though close, resemblance of the physical characteristics of the deposits. In both we have the same gray measures—sands intermingled with sandy clays, which have a red or reddish hue; in both the pebbly element is scanty. In the Cape Cod exposures the red beds are more prominent than on Marthas Vineyard, but in both cases the hue is less pronounced and the clayey element less considerable than in the more ancient deposits of the Gay Head Miocene series. It seems likely that these reddish clays have in each case been derived from the washing over of the older Tertiary deposits. That such is the case on Marthas Vineyard admits of scarcely any doubt, for the reason that there the later beds contain fossils which evidently came from the erosion of the earlier series.

It is hardly to be supposed that these red clays of Cape Cod could have been derived from a field so remote as that of Gay Head. As may be seen at the last-named point, the eroded clay is not carried any distance by the tidal currents. We are, therefore, compelled to suppose that the beds in Cape Cod were derived from some areas of the Gay Head series which have been completely eroded away, or at least lowered beneath the sea level by the wearing to which they have been subjected. The remnant of the Miocene rocks which exists in Marshfield, on the mainland to the north of Cape Cod, lies in a depression of the crystalline rocks, where it has been in a measure protected from erosion. It is eminently probable that these beds once occupied the district between Marshfield and the base of the cape. They may, indeed, have had a much greater areal extent.

The probability that the deposits of the series found at Gay Head and elsewhere on Marthas Vineyard once extended over a wide field in the cape district is made the clearer by the fact that remnants of the greensands, as is well known, occur at Marshfield, Massachusetts, north of Plymouth, and apparently also below the level of the sea in one or more points in the Monomoy group of shoals. The evidence as to the existence of the beds at the last-named point is incomplete, but it deserves a brief statement.

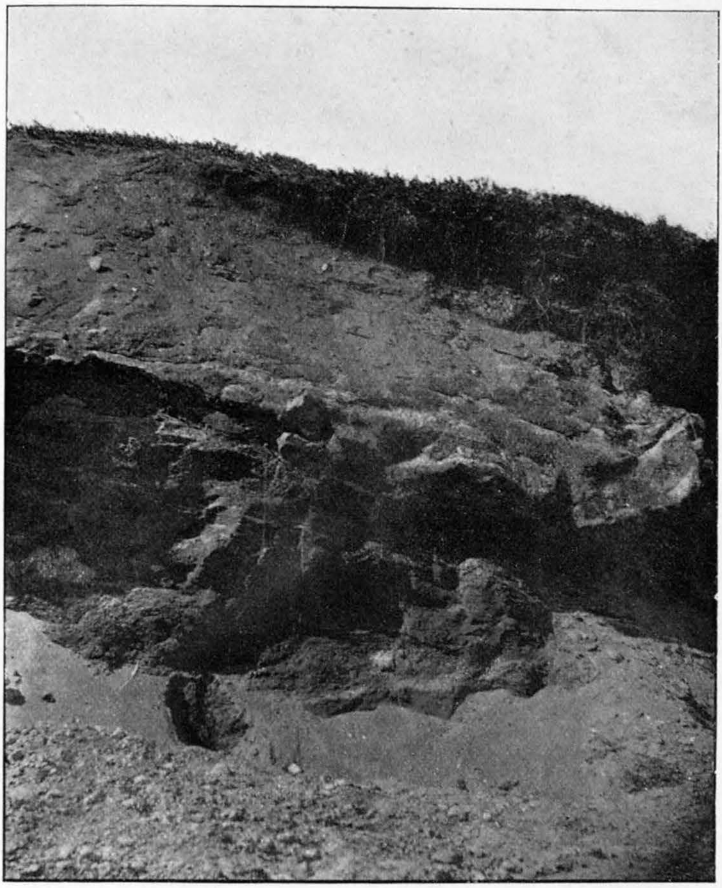
Among the shallows in the Monomoy area is one sometimes known as Stone Horse Shoal. This eminently curious name points to some peculiar feature in the history or structure of the place. I am told by Capt. John L. Veeder, of Woods Hole, that some years ago he was engaged in breaking up the wreck of a ship which had been for some time lying on that shoal and had become partly embedded in the sands. When the hulk rolled over it brought up a quantity of "dark sand" which contained many fragments of bones. In answer to my inquiry Captain Veeder stated that the material was like the greensands of Gay Head. It is well known that sailors are apt to class any bones as those of horses.

As to the nature of the erosion which provided the material for the Nashaquitsa series, there is little distinct evidence, and that is of a negative character. The beds on Marthas Vineyard have afforded fragments of magnetic iron ore, apparently from Cumberland, Rhode Island, and other materials which may be from the same field; but it is to be observed that the pebbles may not have been derived directly from that locality, but may have come, as is the case with much of other materials, intermediately from deposits of Tertiary or Cretaceous age. As these beds were apparently deposited not long before the advent of the last Glacial period, the question arises whether they indicate any form of ice action. To this inquiry a negative answer must in general be given. None of the pebbles are scratched or faceted; there appear to be no ice-rafted blocks; the fragments are all small, the greater part of them of quartzitic or felsitic nature, the ordinary crystalline rocks, such as are so plentifully exhibited by the glacial deposits of the last ice period, being of scant occurrence. In general the pebbles are much waterworn and affected by superficial decay, which shows that they have been long separated from their original bedding places.

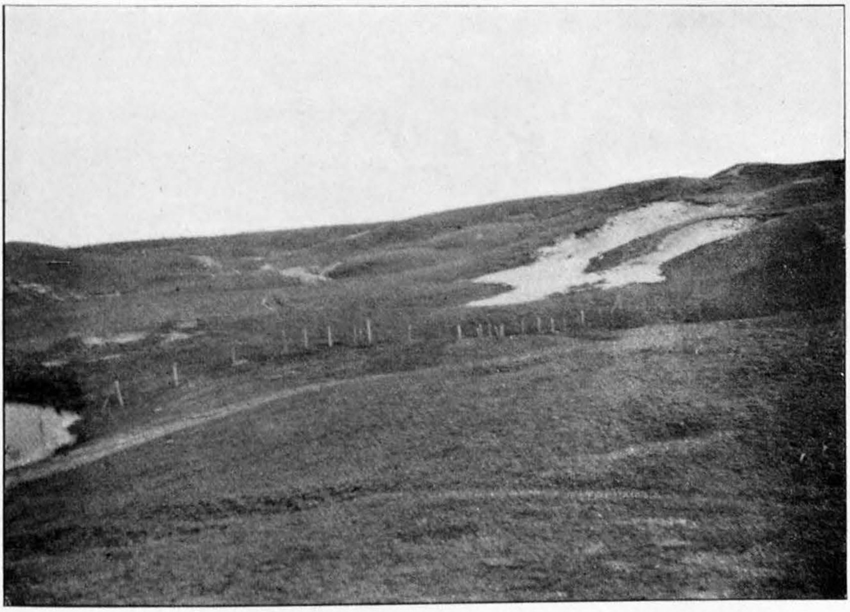
The transportation of the materials of the Nashaquitsa series appears to have been effected, in part at least, by strong and variable currents, as is shown by the stony cross bedding of the sands. At other times, and with sudden alternations, the conditions were such as allowed the deposition of fine-grained clays in these layers. It is a noteworthy feature of the formation that it contains, so far as ascertained, no indigenous fossils of a recognizable nature. This, taken in connection with the fact that on the west end of Marthas Vineyard (where alone the series is well exhibited) there are a great many organic remains of animals derived from the Tertiary rocks, goes to indicate that, though the conditions of deposition and of subsequent time favored the preservation of fossils, none were contributed to the formation by creatures living in the waters. This inorganic aspect of the beds may be due to any one of several conditions existing in this district at the time of their formation. It may have been due to the presence of a glacial sheet; but this hypothesis is less warranted than is the supposition that the deposition took place rapidly in a fresh-water basin much in the manner in which deposits are now accumulating in the basins of certain great lakes, as, for instance, in Lake Ontario near the mouth of the Genesee River. The evidence afforded by the beds is, indeed, most consistent with the view that they were thus formed in a fresh-water or estuarine body into which large and sediment-laden streams were discharged.

At first sight the supposition that this portion of the continent was the seat of considerable lakes during or about the Pliocene epoch may seem to require an excessive difference from the existing geographical conditions. It is, however, evident that the Atlantic shoreland from the Carolinas to Nova Scotia has from the beginning of the Mesozoic to the present geological time tended to develop extensive lacustrine areas. In Triassic time these areas of fresh water were numerous and large, the basins having a character and an extent comparable with those of the eastern flank of the Rocky Mountains during the Cretaceous period. A part at least of the Gay Head series, including the plant beds of the Cretaceous and a portion of the Miocene, appears to have been deposited in areas of fresh water. It is not necessary to suppose that these areas of fresh water were completely separated from the sea; they may have been estuarine in their nature, much as are the sounds of North Carolina and other portions of the southern coast of the United States.

The question arises as to the original extension of the deposits of the Nashaquitsa series. As yet they have been definitely observed only on Marthas Vineyard, in the islands of the Elizabeth Archipelago, and in the area we are now considering. It is likely, however, that beds of equivalent age exist in Block, Fishers, and Long islands. Deposits of perhaps the same age may exist farther to the south, though until fossils are found in the Massachusetts area there will be no sufficient means of fixing the age. The fact that these beds are found scattered over a considerable area in the manner before noted indicates



*A*. SECTION OF PART OF TRURO SERIES ON NORTH SIDE OF PAMET RIVER, NEAR BRIDGE.



*B*. VALLEY EXCAVATED IN TRURO SERIES, CHILTONVILLE, PLYMOUTH.

that they were at one time extensive. The height they now occupy, notwithstanding the considerable erosion to which they have been subjected, shows that they must have been formed when the level of the sea was much higher than at present. Thus on Marthas Vineyard they lie at not less than 200 feet above the tide, and their upper surface has shared in the erosion which has served to develop an old and deeply incised topography on the area. It is, indeed, necessary to assume that the upper surface of the deposit originally lay at a far higher level, perhaps 100 feet or more above its present plane.

As to the dislocation of the beds, this seems to have occurred before the erosion which formed the valleys in which lie the bays and sounds that separate the known location of the deposits. The time of the dislocation can not be more definitely stated than that it was after a part, at least, of the Pliocene had been deposited and before the deposition of the Barnstable clays or the Tisbury beds, which apparently lie above them. The interval between these stages was evidently of considerable duration, even in the geological sense of the word, for it included not only the time occupied in the folding but also the period required for a considerable erosion of the beds.

Concerning the extent of the dislocations which have affected the Nashaquitsa series, it may be said that it was much less intense and general than that which is recorded in the Gay Head section. In the area occupied by the last-named group of strata, about 30 square miles in extent, the average dip of the beds is about 45°, and no part of the layers, so far as seen, remains in a horizontal attitude. In the case of the Nashaquitsa series the greater portion of the Marthas Vineyard area is but little dislocated, and on Cape Cod the average departure from the original horizontal attitude is apparently only a little greater than it is on the Vineyard, probably not averaging more than 10° of declivity.

The foregoing considerations justify the supposition that the Nashaquitsa series originally occupied an area along this portion of the shore of the continent; they warrant also the belief that this area was, after a slight though distinct dislocation, carved into an extended topographical relief and that the surface of its more salient points was considerably lowered in the process. We have to suppose that this carving was, in the main at least, due to river action, though the valleys may have been affected by marine agencies after they were lowered beneath the plane of the sea.

## THE BARNSTABLE SERIES.

After the formation of the topography cut in the Nashaquitsa series had been effected the district was again depressed beneath the sea. The downward movement certainly brought the coast line at least 100 feet above its present level, for the Barnstable clays attain the elevation of 60 feet above tide, and the Tisbury clays, their probable equivalent, rise to about 90 feet. As these clays, particularly those of the Barnstable area, have the character which belongs to deposits formed at some distance from the shore line, it is likely that the down-sinking was to a much greater depth than is here indicated.

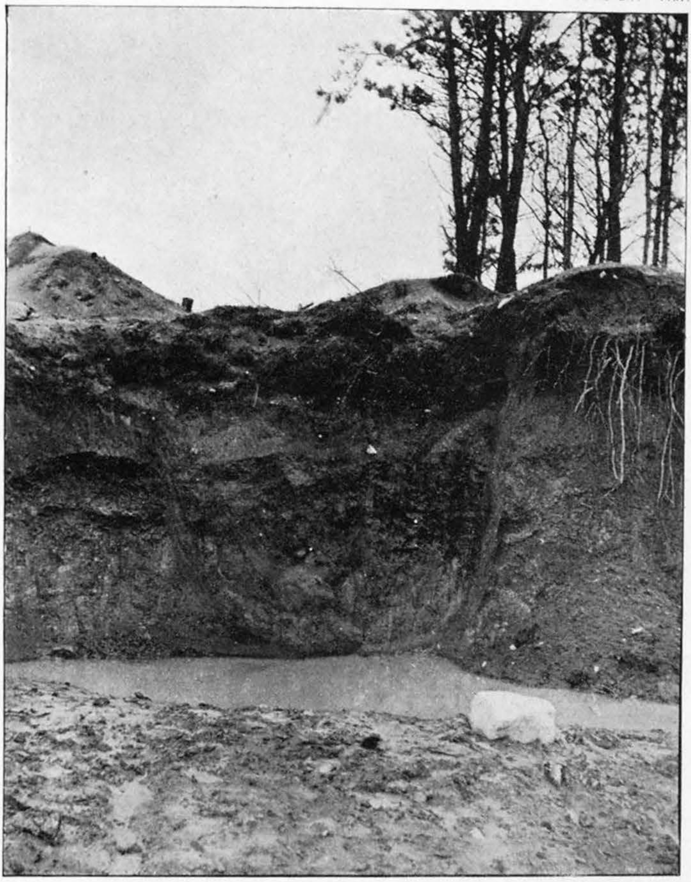
Clays of the same general nature as those of the Barnstable series occur along the shore to the eastward as far as Chatham, though the best exposures known to me are those on the present marine escarpment and in the clay pits at Barnstable; there they seem everywhere to underlie the glacial deposits, being usually separated from them by a variable thickness of apparently pure glacial sands and clays.

It is not unlikely that some of the brick clays lying farther northward and westward on the mainland, as well as other deposits in Harwich, are of the same age as those of the Barnstable series, but their discrimination is difficult and has not yet been effected for the reason that they do not apparently differ in any distinct way from those of later date and of undoubted glacial origin.

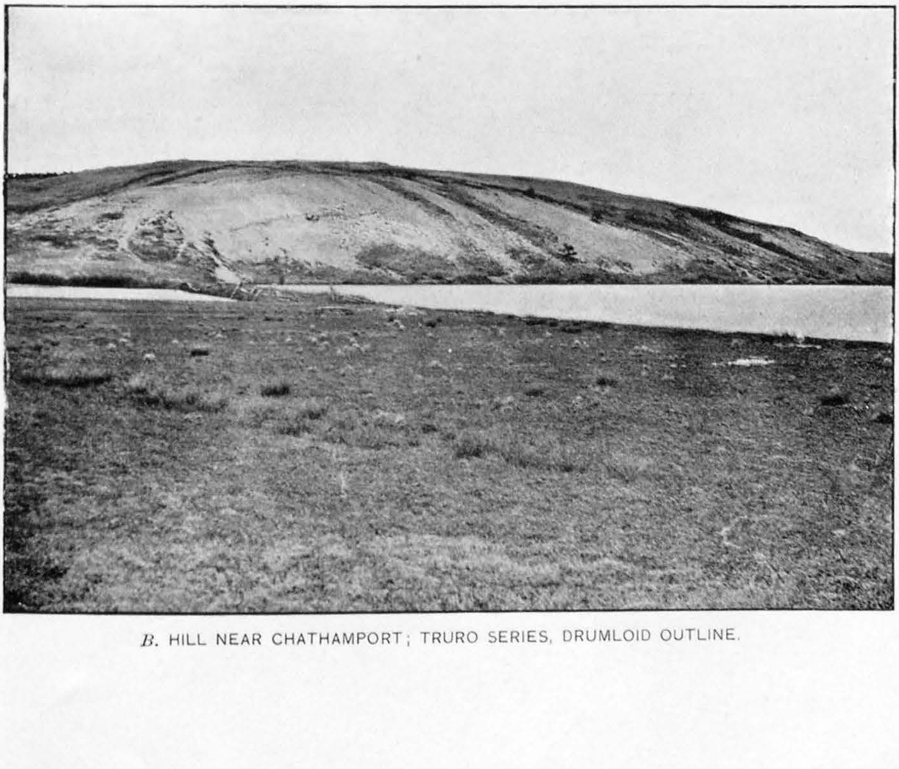
The gravels in the clays of the Barnstable series are known to me only by the reports of those who have penetrated the beds in sinking wells. They are described as composed of small pebbles, mingled or coated with iron oxide.

The Barnstable beds, as has already been suggested, may be the equivalent of the Tisbury beds of Marthas Vineyard. The evidence of the identity of age is, it must be confessed, not very strong. It rests altogether on the fact that in both cases clay beds not greatly disturbed by the mountain-building forces rest upon the disturbed Gay Head series, and that they have both been elevated to a considerable height and carved by erosive agents. To suppose that the two series are of diverse age would require the assumption that there had been one more cycle of erosion, subsidence, and elevation in the Pleistocene period, which is already overcrowded with actions of this nature that I have been compelled to postulate in order to explain the geological structure of the district.

As against the supposition of the identity of age of the two sets of beds, it may be said that the Tisbury series forms a distinct, though much eroded, bench on the north side of the island of Marthas Vineyard. There is no evidence that they ever had a very wide lateral extent. The beds are mottled yellow and bluish clays and sands, with occasional bowlders of small size, which may possibly have been ice-rafted to their present positions. The materials of the strata have apparently been derived from the erosion of the Cretaceous and Tertiary beds of the dislocated area against which they lie. It seems quite possible that with the advance of our knowledge of this district it will be found that the Barnstable beds, which appear to have been formed in deep water in an offshore position, are not to be regarded as in age the equivalents of the Tisbury beds, which were evidently formed nearer the shore and in a shallower depth.



*A.* STREAM CHANNEL CUT IN CONTORTED CLAYS, NORTH WARWICK STATION.



*B.* HILL NEAR CHATHAMPORT; TRURO SERIES, DRUMLOID OUTLINE.

It is to be understood that evidence of a diversity in age of the glacial clays and of the beds here referred to as related to the Barnstable series is not perfectly clear. I see no reason to doubt that the formation of the deposits which lie beneath the cape and the region to the northward as far as Plymouth Harbor clearly antedate the last ice epoch, but some of the clay beds of the cape district may have been deposited during the time when the ice work was in progress. It should also be noted that even if the glacial origin of the Barnstable series should be proved, the evidence is still to the effect that the ice action was not that of the last advance, but an epoch separated from it by events which indicate the lapse of a great interval of time. This will be more evident in the sequel.

## TRURO SERIES.

The characteristic Truro series is even more generally concealed than are the beds which lie beneath along the eastern shore from Wellfleet to Highland light. They are, it is true, revealed in the wasting cliffs, but the amount both of slipping and of loose débris is so great that it is not possible to determine further the character of the section than that it is composed of a hundred feet or more of fine, gray, micaceous sands and sandy clays in frequently alternating beds. These beds apparently contain no fragments of compound rocks; the only pebbles they carry—and these are small and of infrequent occurrence—are composed of white quartz. The beds appear to be somewhat disturbed, but the irregular sliding of the cliff as it is undercut by the sea makes this apparent evidence of orogenic stress untrustworthy.

At only two places has it as yet proved possible clearly to ascertain the true attitude of the beds in the Truro series. One of these is a pit whence clay for hardening roads has been taken. It is on the north side of Pamet River, immediately south of the road, and a few hundred feet from the bridge over that tidal stream. The section is as shown in Pl. XCVIII, *A*, and in fig. 90, p. 534. The materials consist of alternating clays and sands, such as are shown on the cliff at Highland light, even bedded and quite without pebbly matter except bits of rounded quartz. They lie at an angle of about 18°, dipping northwestward. There is a thin layer of pebbly drift on the top of the section. (See Pl. C.)

Another exhibition of these beds is in a clay pit 200 or 300 feet north of South Wellfleet station. Here the beds are at higher angles than in the section near Pamet River bridge; in part the slopes are of 30° or more. The bedding, indeed, seems to be crushed as it is at certain points on Marthas Vineyard. Here, as in the last-named section, there is a thin overlay of pebbly washed drift, with small rounded bowlderets.

It should be noted that the sections above described were obtained at the only points where the attitudes of the Truro series could be clearly discerned. Taken in connection with what has been observed on the cliff shore and in a considerable number of obscure artificial exposures, there is evident reason for believing that the strata of this series are generally dislocated much as are the beds of Tertiary age on Marthas Vineyard.

It appears to me unreasonable to suppose that the steep dips of the beds of the Truro series are due either to cross bedding or to glacial thrusting. The sections examined are sufficiently extensive to reveal the true structure. They show nothing to arouse the suspicion that these slopes are due to deposition on the construction point of the stratum. As for the thrusting, there is, as is elsewhere noted, no good reason to believe that the glacier ever eroded this surface. If it did so, its action was not vigorous enough to have eroded the delicately molded pre-Glacial topography.

The feature which most distinctly separates the surface aspect of the Truro sands from that of the morainal aprons is their slope. This is not, as in the sand plains of Barnstable and elsewhere, toward the open sea on the east, but distinctly toward the bay on the west. The surface is, it is true, to a certain extent encumbered by the waste left upon it in the last advance of the ice; but making allowance for this coating, it is quite evident that the slope, instead of being outward from the ice front, was inward toward the face of the glacier. If, indeed, the deposit is to be regarded as a sand plain, it will have to be assumed that the ice lay outside of the cape, discharging its waste westward toward the bay, a view which is manifestly inconsistent with all we know of the distribution of the glacial envelope on this part of the shore.

Taking no account of the deformations of the surface in Truro and Wellfleet, which have been brought about by the small amount of glacial waste which the area bears, the westerly slope is clearly indicated either to the eye of the observer in the field or by the inspection of the topographic map, where the contours are seen to lower as we pass from the outer or eastern to the inner or western side of the area.

As we pass from the eminently characteristic surface of the sand deposits of Truro and Wellfleet toward the southern and western parts of the cape, the glacial deposits thicken and become more irregular, until in Orleans the Truro sands are to a great extent concealed by this drift. Nevertheless, beds of the same general nature are noticeable at most points where a natural or artificial section is carried to any considerable depth in all the area as far west as Yarmouth. They are particularly well shown in Dennis and Harwich, and are also revealed in the southern parts of Brewster and the northern portion of Harwich. It is here, as in the typical localities of the series, quite evident that the surface was deeply incised by the action of streams before the last invasion of the ice, which served to encumber and at times efface the preexisting valleys, though the erosive action which conveyed this waste was not sufficiently intense to cut away this rather delicately molded topography. Throughout the area in which these

*A*. TALUS SLOPE, HIGHLAND LIGHT.

*B*. CONTORTED TRURO CLAYS AND GRAVELS NEAR CHATHAMPORT.

ancient sands are traceable they generally rise on the crests of the ridges which they occupy to the height of about 100 feet above the level of the sea. It seems likely that while glacial erosion, mainly if not altogether due to streams from beneath the ice, may have cut down these crests to a certain amount, the extent of this wearing has probably been not more than a few feet.

The original extent of these Truro sands is, on account of the erosion to which they have been subjected, not clearly determinable. It seems, however, to have been great, as will be seen from the following notes concerning the distribution of beds apparently of that age which occur in a fragmentary manner in and about Cape Cod. In the western part of the cape there is reason to suspect that deposits of this age underlie the ridge of the Falmouth moraine. Exposures of yellow sands are shown at a number of points on the western face of the moraine in positions which indicate that they are portions of a large area which extends beneath this mass and rises to a considerable height beneath it. On the island of Naushon, the northernmost of the islands of the Elizabeth Archipelago, orange-colored sands with characteristic absence of coarse waste underlie nearly, if not quite, all of the area and rise to the usual height of about 100 feet. Southward throughout this group of islands to Penikese beds of this character and presumably of the same age are seen here and there, evidently lying in the Nashaquitsa series. On Marthas Vineyard the series is less well shown, yet it is tolerably well indicated on the northern side of the island at Copoggan Head (misnamed on the maps Cape Higgon), as well as at other points between Menemsha Creek and West Chop. They are also seen in the retreating escarpments of East Chop and West Chop and the sea face at Cottage City.

Beneath the great plain of Marthas Vineyard, the upper portion of which is clearly of the character proper to morainal aprons, there is revealed by occasional wells extending to the depth of 70 to 90 feet a deposit of yellow sands with no large pebbles, which appears to belong to this group. Here we have to suppose that the beds had been worn down by marine or other action to a level somewhat below that which they occupy elsewhere and then sheeted over with the deposits formed during the last advance of the ice. In the valleys of Tisbury and Tiasquan rivers, in the central part of Marthas Vineyard, and at Gay Head there are traces of the same sands, which are scantily revealed and only discriminable from the deposits of the Nashaquitsa series by the characteristic yellow hue.

West of the base of the cape, along the southern shore of Massachusetts, deposits which may be compared with those of the Truro series are not clearly disclosed and may not exist, though the search for them has not been carried so far as to make their absence certain. In Rhode Island, as has been suggested to me by my colleague, Mr. J. B. Woodworth, beds of this age may underlie the Charleston moraine, where, as remarked by the late J. D. Dana, stratified sands appear to underlie the morainal deposits. The scanty outcrops of these beds in their appearance warrants the supposition that the formation has the same general character that it exhibits in the localities before described.

North of Cape Cod, along the shores of Massachusetts Bay, the Truro sands, overlain by distinct glacial deposits, are abundantly exhibited. On the shore of southern Plymouth, from point to point, they form the marine escarpment. (See Pl. CIV.) In the high ridge of Manomet Hill they probably attain the height of 250 feet or more above the level of the sea. The erratics, which are so abundant on the ridge and which give it the character of a moraine, form only a relatively thin coating on the summit of a pre-Glacially-formed ridge, resulting from extended subaerial erosion of the inferior sands. (See fig. 92, p. 555.)

North of Plymouth the curious table-land known as Egypt Heights, in the southern part of Scituate, appears to be composed of beds in character quite like those at Truro. The general form of this curious deposit of sand can best be explained by the supposition that it is the remnant of a large area, and not a local deposit accumulated during the last Glacial epoch. This view is supported by the general character of the material, which is much the same as that of the Truro section, though it is more deeply covered with recent glacial waste. Scattered patches of the same decayed sands continue to the northward as far as Boston Harbor. In that basin, mostly below the level of the sea, a thick deposit of sands clearly antedating the last ice epoch has been revealed by artificial sections, as in the tunnel for the Moon Island sewer, and particularly in a well boring made on Deer Island. At the last-named locality a thickness of 300 feet was passed through, the beds being in general character like those before described, except that the oxidation was less complete than at the other parts. The evidence goes to show that here, as elsewhere, this section of decayed sands with few pebbles is immediately, though discordantly, overlain by the bowldery drift. North of this point on the shore I am not aware of any sands which may be referred even conjecturally to the Truro series.

In the district of southeastern Massachusetts, remote from the shore, I am aware of but one locality where deposits of much-oxidized sands having the general character of those before described are clearly revealed. This is at Prospect Hill, in the southern part of the town of Raynham and the western part of Taunton. At this place we find an irregular ridge, composed mainly of sands with a few pebbly beds, capped in part by a layer of bowldery nature, which gives the mass something of the aspect of a moraine. Some years ago I came to the conclusion that the greater part of this ridge represented a much-eroded deposit, which was formed before the last ice advance and which had been scantily affected by a morainal accumulation made

TRURO SERIES AND GLACIAL BEDS, HIGHLAND LIGHT.

in that stage of the Glacial period. It now seems most reasonable to regard this as a remnant of the deposits of the Truro series.[1]

It is probable that many other deposits of well-oxidized sands which exist in southern New England will eventually be found to represent the same epochs of Pleistocene time as those above catalogued. At present they are naturally, and it may be inextricably, confused with the accumulations of washed drift which were so plentifully formed in front of the ice during the later advances of the glacier. As will be further noted in the study of the glacial deposits of Cape Cod, the Truro series may possibly be but the outer remnant of a broad sheet of shore sands formed during the earlier epochs of the Glacial period, when the margin of the ice lay at some distance north of the present shore, and that this moraine accumulation passed into other types of glacial waste as it approached the ice front.

The facts before noted make it probable that at some time before the advent of the ice of the last Glacial period in the region about Cape Cod the surface of the land was at a much lower level than it is now—at least 100 feet lower—and that at that time an extensive sheet of water-borne sands was deposited on the sea bottom. It seems necessary to suppose that this sheet was laid down as a tolerably continuous outward-sloping formation, such as is now found in the continental shelf along the Atlantic coast. It certainly could not have accumulated in the patches and ridges in which it now appears. We can not, for instance, suppose that the crest which forms the foundation of the Elizabeth Islands, and which rises about 150 feet above the present level of the sea floor, or that of the Truro Plateau, which attains a like or greater height above the bottom of Cape Cod Bay, was formed as we now find it. We are forced to assume that these evident remnants of erosive work were originally parts of a widespread deposit, by far the greater part of which has been swept away.

The time of the erosion of the Truro sands, which reduced the area of the formation to the few remnants we now find, clearly antedated the last advances of the continental glacier. This is indicated by the facts that the position of the remnants is that which they would occupy if they were left by water erosion, but not such as would exist if the wearing had been effected by ice, and that the preexisting rather delicate topography, such as would have been carved by stream action, was not destroyed by the glacial erosion. In illustration of the first of these points it may be said that the ridge of the Elizabeth Islands is precisely such as would be brought about if it had been a divide between the supposed rivers of Buzzards Bay and Vineyard Sound, but in no way could it well be explained by glacial or marine erosion. As for the second, the many pre-Glacial channels on the Truro-Wellfleet plains show how even delicately sculptured valleys were not completely

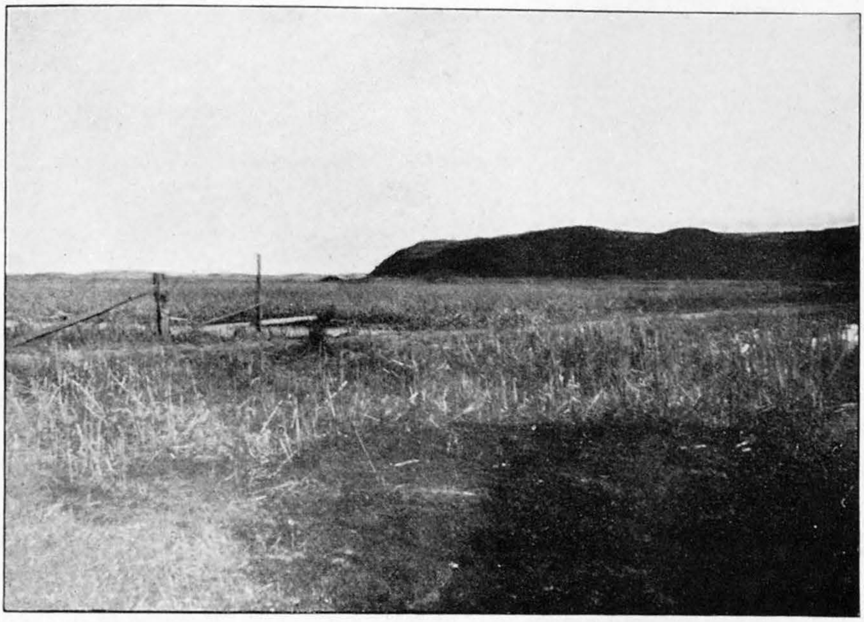
[1] Since this report was written beds apparently the equivalents of the Truro series have been found by the writer at a number of points in southeastern New England.

defaced by the wearing influence of the glacier which came upon them in its marginal and attenuated form, if indeed they were ever actually beneath the ice.

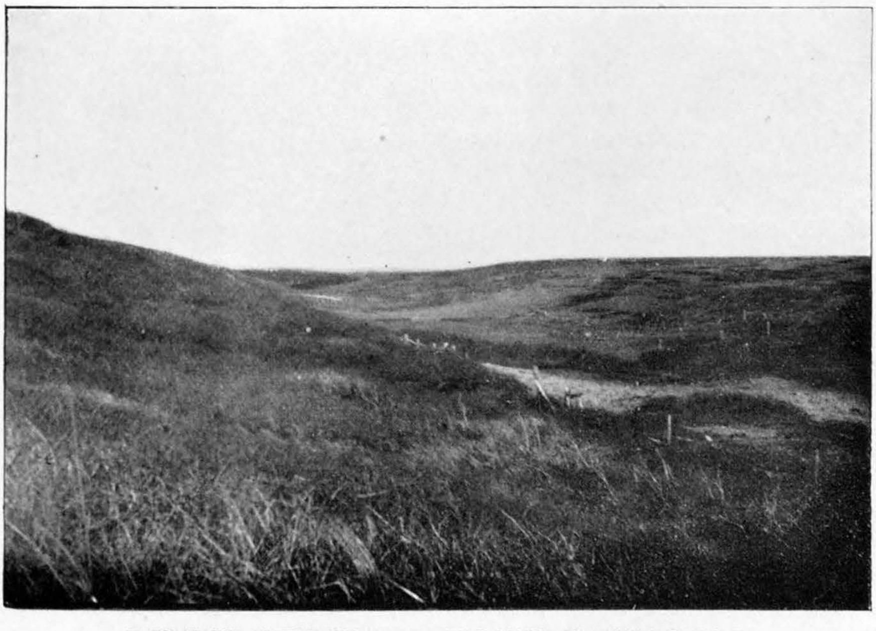
As for the time when the erosion of the Truro sands was effected, we may confidently place it in the later part of the long interval which is partly, at least, recorded in the well-developed subaerial topography which was made on Marthas Vineyard after the cessation of the dislocation of the underlying beds and before the advent of the ice of the last Glacial epoch. As before remarked, this interval was long, for the work done during it was vast. It is clear that the Truro beds were formed after this topography was pretty well completed, for beds referable to the age lie partly in the valleys due to the erosion in question. It is thus evident that the greater part of the erosion of these later sands came after the shape of the Vineyard topography had been in large part determined, but probably before the valleys thereof had attained anything like the present development. (See Pls. CII, CIII.)

The reduction to a plain of the Truro sands was probably in part effected by the action of the sea. As may be noted along the shores where these beds are subjected to the action of the waves and marine currents, the beds wear away with exceeding rapidity. It may, however, have been in considerable part accomplished by ordinary stream action, as is shown by the persistence of many ancient valleys in those parts of the cape district underlain by these deposits. It is, however, difficult to believe that this stream erosion took place under the present conditions of climate and geography, for the reason that the beds of these ancient water ways are no longer occupied by streams, except, perhaps, on the rare occasions when, on a frozen earth, melted snow or rain is deprived of its usual exit by percolating into the porous underlying sands. The absence of water in these channels is probably to be attributed in part to the fact that they have been greatly shortened by the cutting away of their headwaters, so that the water now flowing seaward in their drainage is less in amount than of old, being no longer more than can pass through the interstices of the sands, through which it more readily finds a passage because the way to the sea is not so long as of old. It is probable, however, that the amount of the rainfall has in geologically modern times diminished in this region, as elsewhere on this and other continents, so that the capillary channels are able to afford storage and passage to all the precipitation. It may be observed that in times of any great rainfall sandy plains occasionally for a short time develop superficial streams, the water quickly ceasing to flow when the precipitation stops.

The rate of the erosion of the Truro beds wherever they are assailed by either marine or fluviatile agents is made the greater by the fact that the beds are destitute of coarse débris, which, in the case of the till, brings about the formation of a more or less effective revetment on the erosion face that hinders the action of waves or currents.



*A*. MOON POND ESCARPMENT, FACING PROVINCETOWN SPIT.



*B*. BEHEADED WATERLESS VALLEY JUST SOUTH OF HIGHLAND LIGHT.

Moreover, the very lean nature of the soil causes the growth of vegetation to be slight in amount, so that the protection of this sort which is usually so important is scanty. Thus the wearing rate on this group of deposits is likely to be very much greater than it is on such beds as form the Tertiary strata of Marthas Vineyard, frail as the latter appear to be. (See Pl. XCVIII.)

The most important indication pointing to the origin of these Truro sands is the apparently entire absence of fossils in the section. In the extensive outcrops which I have inspected no trace of organic matter has appeared. It seems clear that the beds were laid down under conditions which were peculiarly unfavorable for the inclusion of organic remains, or that such remains were subjected to some process which utterly removed them. Although, as before noted, these beds are considerably oxidized, it can not well be believed that fossils once present have been utterly destroyed. The Tertiary sands of Virginia and elsewhere are equally affected by decay, yet the mulluscan remains are fairly well preserved. Assuming, then, that these beds were originally formed without organic remains, the probability is fairly established that their materials were brought into the sea by glacial action. In no other way does it seem possible to account for the formation and deposition of such a mass in a marine or lacustrine area. It is to be said that this view has its difficulties, among which we may reckon the apparent absence, as before noted, of all ice-rafted blocks in the beds and the lack of clay in the greater part of the section.

Taken in connection with the seemingly nonfossiliferous clays of the Barnstable series, the Truro beds may perhaps be regarded as a stage in one of the several cycles of a glacial period. It is a well-recognized fact that the glacial flour or fine débris, which in the ordinary course of glaciation constitutes the larger part of the detritus that is formed, is normally carried much farther away from the front of the sheet than the sand, and that this in time goes farther than the pebbly matter. We may thus reckon that the Barnstable clays are the outer or relatively remote accumulations of an ordinary glacier, and that the Truro sands were laid down when the ice front was nearer the present shore line. If this view be accepted, we must then suppose that in the ice epoch which brought about the formation of this series—probably not the last Glacial epoch—the glacial sheet did not quite attain to this field, and that the land lay at a lower level than it does at present. As is clearly indicated by the extensive erosion which followed this period of deposition—erosion in which the glacier appears to have had no part—the time intervening between the formation of the Truro beds and the advance of the ice sheet which deposited the till, moraines, kames, etc., of the district must have been great. It was assuredly many times as great as that which has elapsed since the last of the ice sheets left the field.

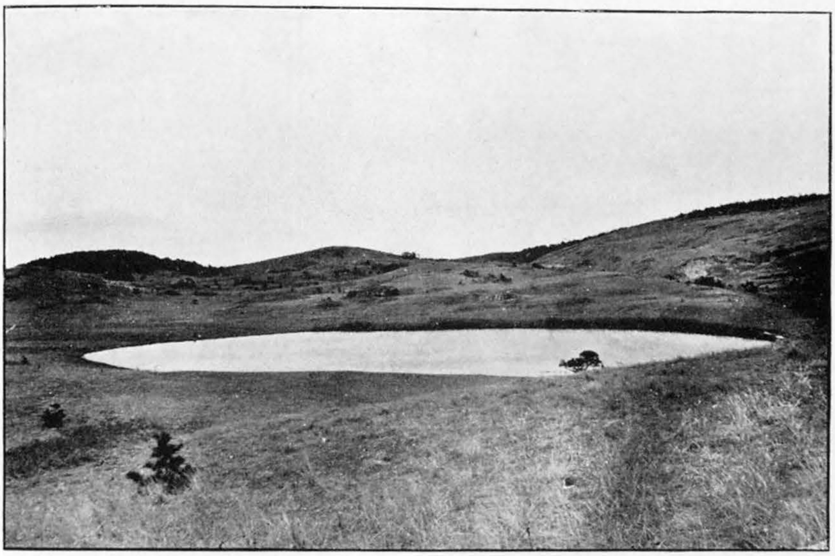
Although the matter has been before stated in a fragmentary manner, it may be well again to call attention to the accumulation of evidence which exists in this field going to show the very great length of the time which has elapsed since the close of the Pliocene. In this interval there were evidently three distinct periods of erosion, each of long duration, and an equal or greater number of widely varying changes in the position of the land in relation to the sea. As measured against the geological work which was done in these periods, that brought about since the close of the last ice advance is relatively of little account, being limited to slight changes of level and to a small amount of marine cutting, the subaerial wearing being quite insignificant, perhaps not the one-hundredth part of what was done in the earlier stages of the so-called post-Tertiary. It thus seems that, basing the measure on a vaguely assumed rate in the alteration of organic forms, we have most likely much underestimated the duration of this division of the earth's history.

## CONDITION OF THE DISTRICT AT THE BEGINNING OF THE LAST GLACIAL EPOCH.

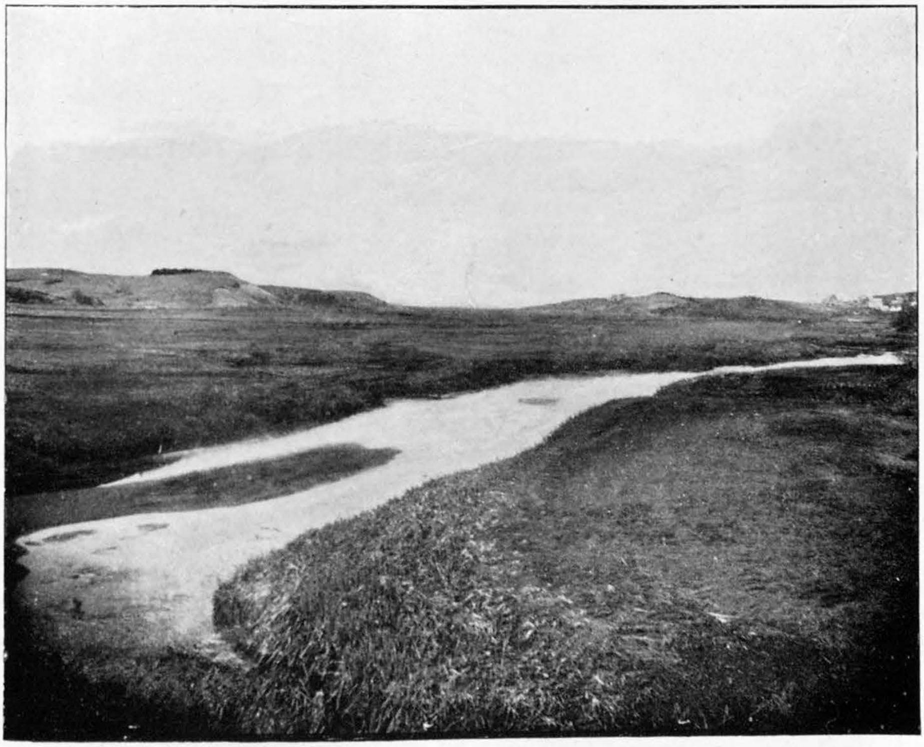
The conditions of height and shape of the land in this area immediately before the advent of the ice of the last Glacial epoch appear to be approximately determinable. It is tolerably evident that the land lay at a higher level in relation to the plane of the ocean than it does at present. This is indicated by the existence of the flooded drainage basins, which have been already in part described. These drowned valleys include not only the greater basins of Cape Cod Bay, Nantucket and Vineyard sounds, and Buzzards Bay, but many of the divisions or branches of these wide valleys where the streams tributary to the effaced rivers now enter the sea. Thus, on the islands of the south as well as on the mainland all the valleys which are now or have been in former times the seats of streams are flooded at their mouths.

On the north shore of Cape Cod we find a number of pre-Glacial channels which slope toward the bay of that name, and which evidently were at the time of their excavation the beds of streams discharging into a river that flowed northwardly to the shore line, which was farther out seaward than at present. These old valleys may be traced from Duxbury to the northern part of Truro. They point toward the central portion of the submerged trough in a normal and most suggestive manner. On the body of the cape these channels are generally much occluded by the deposits of glacial drift. They are to a considerable extent deformed by the scouring action of the streams which flowed beneath the ice sheet while it lay over the country.

The first of these channels of Cape Cod to be noted is that of Monument River, which now is a tidal stream discharging into Buzzards Bay. As it is clogged at its northern end by drift, it appears as a tributary of the ancient stream which occupied the valley of Buzzards



*A*. TOPOGRAPHY CHARACTERISTIC OF THE TRURO SERIES; VALLEY OBSTRUCTED BY GLACIAL DRIFT.



*B*. BLUFFS OF TRURO SERIES, PAMET RIVER VALLEY.

Bay. The form of the trough, which distinctly widens to the northward, suggests that it originally flowed into Cape Cod Bay, and that the ridge which originally parted it from the waters of the south was cut through by a torrent which flowed through this depression in the time when the basin to the north was occupied by the glacier. Eastward along the north shore of Yarmouth the streams appear to have been short and to have drained north and south from the highland now covered by the moraine, and which is locally known as the "Backbone of the Cape." The valleys of these streams draining northward are now but faintly traceable in the confusion of the morainal drift. On the south side of the ridge the valleys are to a great extent lowered beneath the frontal aprons of stratified materials, yet they may be indistinctly and in a general way traced by the depression in which lie the lakes and the streams which drain them.

At Bass River we have another instance in which a pre-Glacial valley (or valleys) has been enlarged and deepened by a current from the glacier. At this point there seems originally to have been two streams, one flowing northward, the other southward. The subglacial stream cut through the ridge between them, converting the trough into a broad way, which practically divides the cape, so that a trifling expenditure would suffice to make a water way from the north to the south side of the peninsula.

East of Bass River as far as Orleans the central ridge of the cape continues; the valleys become less and less blocked with till and morainal waste. This is especially the case on the north side, where the valleys in Dennis and Brewster channels, with rather obscure digitations pointing toward Cape Cod Bay, may be well traced. Beyond Orleans the ancient central watershed disappears, the sea having eaten into it from the east, and the larger valleys usually run across the width of the peninsula. This feature is best shown at Pamet River (see Pl. CIII), where one of these depressions appears—after the manner of the valleys of Bass River and Monument River—to have been depressed and widened by a glacial stream until it completely divides the peninsula, so that there is only a sand beach at the outer side to unite the farther part of the cape with the body of the area. There are, however, many lesser valleys which slope to the northward and which seem essentially river ways, though they are no longer occupied by streams. These troughs appear to be beheaded at their upper or outer ends, their conditions suggesting that their headwaters lay in the lost territory which has disappeared by recent marine erosion (see Pl. CII). In the account of the Truro beds it has been suggested that the former presence of streams on these now dry valleys may be accounted for by the above-suggested diminution of their drainage area, or perhaps in part by the diminished rainfall which appears generally to have attended the disappearance of the glacial sheet and which may have been the cause of its shrinking.

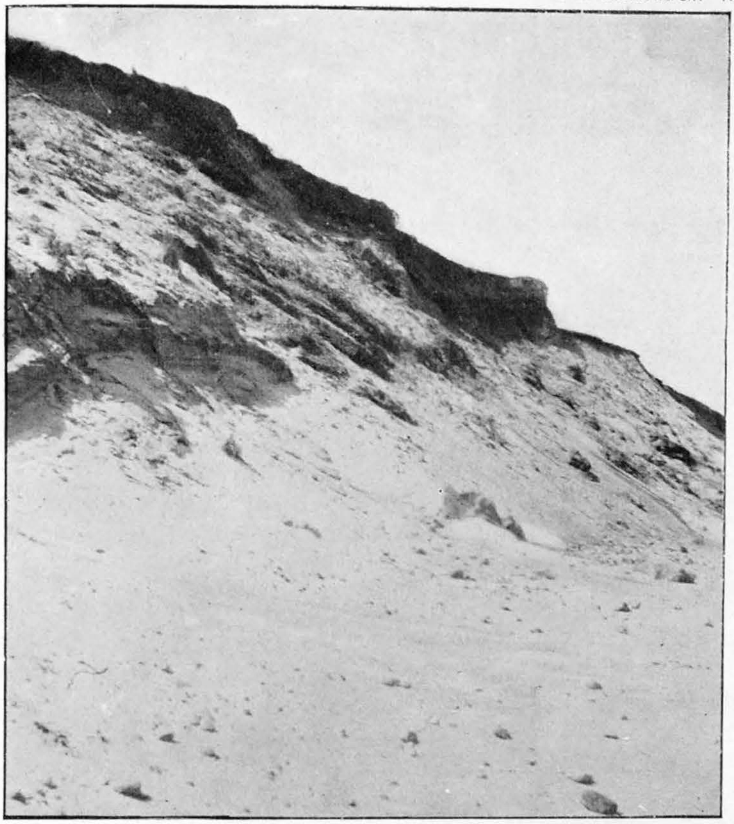
It is evident that the shape and size of Cape Cod shortly before the time when the glacier came upon it differed greatly from what we now find. In place of the narrow peninsula, in form like the flexed arm of a man, was a broad salient which extended as a connected land to some distance beyond the outer margins of Nantucket and Marthas Vineyard. At this stage the sea level probably stood about 200 feet lower than it does at present. During the time when the ice lay over the district it was depressed to a level at least 100 feet below where it now stands. This permitted the formation of the great sand plains of Marthas Vineyard, Nantucket, and the cape. When the ice departed the land in part resumed its old height, rising a little above its present elevation, and then sank, as is shown by the submerged forests which occur from point to point along the shores.

## GLACIAL HISTORY OF THE DISTRICT.

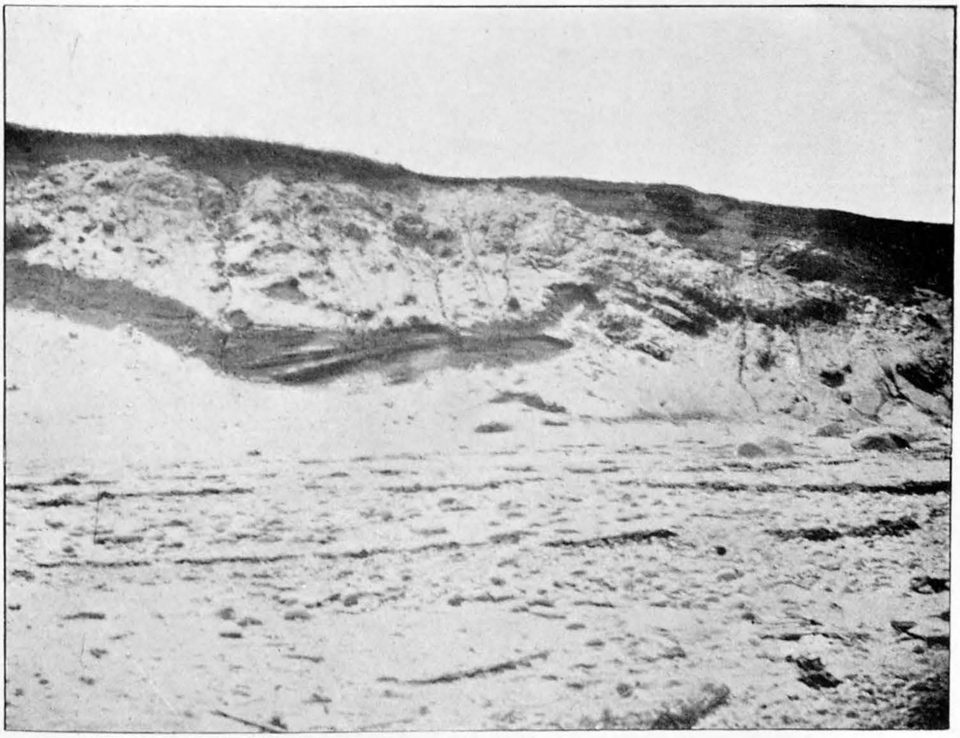
It has already been noted that the deposits contained in the strata from the Miocene to the Truro series, inclusive, suggest the existence of glacial action in this part of the world at various times since the middle Tertiary; but unless the lower Pliocene beds of Gay Head attest the actual presence of ice, there is no reason to believe that it ever rested upon this field until the last epoch. Even in that time the sojourn of the glacier was evidently brief and the work which it did of relatively slight structural or geographical importance. It has already been noted that the general character of the surface had been determined by pre-Glacial conditions The valleys and ridges existed in general about where we now find them, only now they are to a great extent filled with glacial waste.

It seems pretty clear that immediately before the advent of the glacier the surface of the cape was carved into a topography such as is likely to be formed on clays and sands by the headwaters of streams. The valleys were rather deep and steep-sided. Where the clays come to the surface these valleys appear to have had something of the sharpness of the "bad-lands" topography of the western country. This is shown by the indented character of the old surface of the Nashaquitsa beds on Marthas Vineyard, where it is revealed in the coast sections. It is indicated on Cape Cod by the sharp ridges of clay, the so-called "pounds," which occasionally appear at the surface, projecting through the thin envelope of drift. The generally slight value of glacial erosion in this district is best shown on the island of Marthas Vineyard, where, as noted in previous reports, the wearing has been so slight as to leave the pre-Glacial topography essentially undisturbed except by the filling of the valleys with detritus.

On Cape Cod the actual erosion work is little if at all greater than on the islands of the south except in the case of the valleys which were cut through by the streams flowing from beneath the glacier or



*A*. SHORE BLUFF SOUTH OF SHIP POND, PLYMOUTH, SHOWING TRURO SERIES DIPPING STEEPLY NORTHWARD.



*B*. SHORE BLUFF SOUTH OF SHIP POND, PLYMOUTH, SHOWING TRURO SERIES FOLDED AND FAULTED.

under the roof of ice. Of these the most characteristic examples are Monument and Bass rivers. The channel of Pamet River is perhaps another example of the same nature.

There seems no evident reason why the subglacial streams which were on their way to the open water of the ocean should have climbed the ridge of the cape on the south in place of turning directly to the east around its extremity, which was then some distance south of the site of Provincetown. In view of this departure from the most direct way of escape, it may be suggested that as the ice fell back to the northward it may for a time have inclosed a lake between its retreating face and the concave north shore of the cape. In this case breaches would naturally have been formed to permit the discharge of this water from the melting ice through to the sound on the south. It is, however, not certain that any part of the cape was above the level of the sea at the time when the retreat of the ice took place. The only strong point in favor of the view that these channels were glacial stream beds is the fact that they are cut down to the sea level practically throughout their whole length, and that their forms indicate the passage of a current from the northward, and in the case of Monument River there is a considerable area of stratified sands near its mouth, on Buzzards Bay, which may well be taken as the delta formed where the current poured into that basin.

## DIRECTION OF THE ICE MOVEMENT.

As the rocks of Cape Cod and the neighboring parts of southeastern Massachusetts are not of a nature to receive glacial scratches or groovings, the only indications of the direction of the ice flow are those afforded by the positions of frontal moraines and the direction in which erratics have been transported. It should be said that the moraines in this section present such discrepant evidence that conclusions drawn from their positions are not trustworthy. The transported blocks, therefore, furnish our only information, and this is in the main unsatisfactory.

In the western section of the cape, from Monument River to Orleans, the common petrographic elements of the moraine and till are granitites and the dike stones associated therewith, such as are found on the mainland. As these rocks occur along the shore from the parallel of Plymouth to Cape Ann, and may extend an indefinite distance eastward along the sea bottom, no precise evidence as to the course of the ice is to be obtained from these fragments. Eastward from the base of the cape there appears to be a constant increase in the amount of rocks of more evidently volcanic origin, such as are found sparsely about Cohasset and along the north shore of Massachusetts Bay. The deposits of this nature on the mainland are rather too limited to have afforded the large quantity of waste that appears in the cape. It seems likely that they have been derived from beds which lie beneath the sea.

So far as this evidence goes, it seems to show that the direction of the glacial movement on Cape Cod probably did not depart from the general trend indicated by the scratches observed at the nearest points on the mainland, or between north and north 30° west.

## ENERGY OF THE ICE MOVEMENT.

As has been already noted, the energy of the glacial erosion in this district appears to have been but slight. It did not suffice to wear away a rather delicate antecedent topography. On the western part of the cape the transporting power of the ice was sufficient to carry a great number of erratics, many of which are of large size, thousands of blocks each containing from 100 to 300 cubic feet being exposed on the surface of the principal moraine. In the period of its greatest extension the ice apparently passed over the ridge of the cape as far east as Orleans, crossed the valley of Nantucket Sound, and deposited on the island of Nantucket the slight morainal masses which there exist, and which perhaps mark the extreme advance of the ice on this part of the coast.

North of Orleans and thence to the end of the cape there is no distinct morainal accumulation, but occasional wide heaps of drift and the clogging of the pre-Glacial valleys show that the surface was traversed by the streams pouring forth from the glacier. The general lack of erratics other than those which may have been ice rafted, or of any accumulation which can be classed as a moraine, or even as distinct till, shows that at this point the glacier, if it actually lay on the surface, was probably so weak and thin that it had no longer any considerable abrading or transporting capacity. The conditions here resemble those found on the southern part of the highland of Marthas Vineyard, where large portions of the surface are nearly driftless.

Although the carrying power of the ice as marked by the accumulations of erratics was not great, that of the streams which flowed from beneath the glacial sheet was excelled, so far as I have found in New England, only by those which deposited the great sand plain of Marthas Vineyard. As to the extent of the deposit, that of the cape is unexampled elsewhere in southern New England. The area is probably not less than 120 square miles, but the thickness appears to be much less than that of the like mass in the island to the south.

## GLACIAL DEPOSITS.

The deposits due to the direct action of the glacial sheet are the till, the moraines, and the washed drift accumulated in the eskers and the sand plains.

The till deposits of this district are neither extensive nor characteristic. Along the north shore especially the areas are immediately underlain by the Barnstable clays. The coating is evident, but very irregular; in places it is so thick as to resemble a morainal accumulation; at others considerable tracts appear to be quite without the deposit. Toward the eastern extremity of the cape the coating becomes thinner and less recognizable. Angular erratics are rare in the section beyond Yarmouth, and beyond Orleans few erratics greater in size than those termed by Chamberlin "bowlderets" are found, and these appear to have been conveyed by floating ice.

Between Orleans and the northern portion of Truro the till becomes a mere confused mass of the materials of the local beds over which the ice has passed in its movement, with occasional erratics of moderate size which were brought from a distance. It is difficult to recognize it as a distinct element in the sections, for it is essentially wanting over large areas of the surface.

The morainal deposits of the Cape Cod district, though less extensive than those found in the central parts of the continent, are by far the most characteristic in New England, presenting phenomena which are in many ways peculiar. They deserve, therefore, the detailed consideration that will here be given them.

The moraines of southeastern Massachusetts are singularly distributed. In southern New England they lie usually in lines which are evidently almost at right angles to the direction of the ice motion, and variations from this position can usually be explained by the topography of the bed rocks over which the ice moved; but in the cape district, including the neighboring islands and the mainland, the ridges are set at curious angles to one another. There the following directions may be noted:

On the mainland the Plymouth moraine, which extends in a general southerly direction from near the harbor of that name, appears at first sight to be the largest, and is perhaps the most continuous, deposit of the kind in New England. In its northern portion, at least in Manomet Hill, it is underlain by the Truro beds, which arrangement has given a false impression as to the depth of the glacial waste. With some interruptions it is continued southward to Monument River, at the base of the cape, in an approximately meridional axis as far as Woods Hole, and thence, deflecting westward about 30°, it is continued down the Elizabeth Islands nearly to their southern extremity.

On Marthas Vineyard there are two evident morainal belts parallel to that of Falmouth—one on the north side of the island, which is characteristically developed; the other in the central section, which is faintly shown, but can be traced by scattered patches of bowlders. On Nantucket there is a small area of moraine on what is known as Sauls Hills, but the axis of the accumulation is not well indicated; it appears to be in a general east-west direction.

On Cape Cod, occupying, as before noted, the high land formed by the ancient divide of the tilted series of beds, there is a morainal mass extending in an east-west direction from Monument River to the eastern part of the town of Brewster; it may be regarded as continued in a slight form into the western part of Orleans. It is to be observed that this ridge lies at nearly a right angle to the course of the Falmouth moraine, with which it, in effect, coalesces at its western end. Although the general direction of this moraine is east and west, its shape is somewhat concentric, the curve being toward the south, the most southerly part thereof being near Bass River. It is thus evident that there are two distinct alignments of ice-morainal ridges in this district; the one, that which is clearest in its direction, being meridional in the Plymouth ridge, but deflected to a northeast-southwest course in its more southern elements; the other having essentially an east-west course.

So far the diverse positions of the moraines in the Cape Cod district have been explained by the theory of lobations in the front of the glacier, portions of the ice sheet pushing out in broad tongues, each of which made its frontal wall. These walls formed successively, intersecting one another in much the same manner as that of Falmouth intersects that of the cape. While in nowise doubting the adequacy of this explanation as applied to the interior districts of this country by Chamberlain and others, I am compelled to question its applicability to the field now under consideration, for the following reasons:

The Cape Cod district comprises no strong topographical features which could have caused the ice sheet to flow in the directions which would have to be postulated if these several moraines were formed at right angles to the axis of movement. It is unreasonable to suppose that, while the general course of the ice in the neighboring interior district was from northwest to southeast, it should have been directly southward in Massachusetts and Cape Cod bays and directly to the east in the region about Plymouth. On the contrary, the natural conditions, so far as they can be ascertained, would have led the ice in these bays to flow eastward toward the open sea and not southward toward the high ridge of the cape. I have therefore been compelled to seek another explanation of the axial order of these moraines, and have framed what seems a plausible hypothesis to account for this order without having recourse to the theory of lobation of the ice front, which has its difficulties, as just noted. This hypothesis is, in effect, that the moraines of the Cape Cod district are not of the ordinary type, but belong to a hitherto unrecognized group of hilltop drift accumulations, which, though essentially morainal in their nature, were formed under peculiar conditions, rendering them of slight value as indices of the direction of the ice movement.

It has already been incidentally noted that certain parts of the several moraines described in this report rest upon antecedently formed ridges, which, in effect, were the ancient drainage divides of the country. Let us now examine the several deposits to determine how far this peculiar character is possessed by the moraines in general. Beginning with the northernmost of these ridges, Manomet Hill, in Plymouth, it will be found that the elevation is composed mainly of stratified sands, apparently of the Truro series, as has been recently shown by the excavations made in lowering the grade of the State road, which traverses the northern end of the ridge. In other words, the mass of the ridge is of pre-Glacial age, and was probably a divide between the headwaters of the Buzzards Bay river and that which drained the basin of Cape Cod Bay. So far as can be ascertained, the same underlay of sands extends beneath the rather indistinct morainal ridge that continues the Manomet Hill deposit southward to the base of the cape. These sands are not clearly seen, in sections, to pass beneath the morainal belt, but are exposed near by in positions which make it tolerably certain that they must underlie it in the manner of a pedestal, as in the case farther north. (See fig. 92.)

The Falmouth continuation of the Plymouth ridge is by far the longest and most united mass of morainal material yet noted in New England. It extends from Monument River to Woods Hole without any breach in its distinct wall, which rises to a height of from 100 to 200 feet above the sea level throughout its length of about 18 miles.

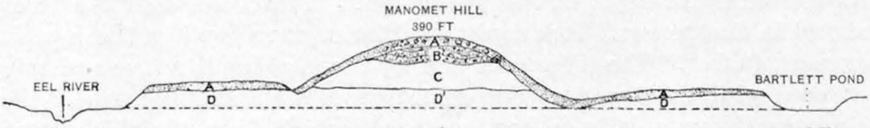



FIG. 92.—Diagrammatic section of Manomet Hill, Plymouth. A, glacial deposits; B, observed Truro deposits, 100 feet; C, supposed Truro deposits, 150 feet; D, sands and clays of unknown age, 100 feet; D', supposed continuation of D beneath the hill.

As this belt is but little traversed by roads which have been at all graded, and as its surface is covered by a dense tangle of scrubby vegetation, it is not easy to obtain sections which reveal the nature of the underlay. At Woods Hole and thence northward for about 4 miles there is abundant evidence that the moraine rests upon a ridge of older deposits. On the western face, nearly as far north as Gunning Point, the underlying clays of the older series can be traced, rising from the shore to the height of from 60 to 80 feet. Here and there along the main highway which skirts the shore to Monument River the conditions of the soil and the level of the streams also indicate that the same ridge of older rocks persists beneath the morainal cap, attaining perhaps at some points between the valleys of the brooks a height of more than 100 feet above the level of the tide. On the east side of the ridge the streams and lakes show by their levels that the ridge continues on that side of the moraine. The facts justify the conclusion that the greater part of this morainal ridge rests on the summit of a pre-Glacial divide which separated the waters of the old Buzzards Bay river from that which formed the valley that is now Vineyard Sound.

With a change of direction from north-south to northeast-southwest the Falmouth moraine is continued southward in the Elizabeth

Islands. In these isles the erratic material is in all cases but a thin overlay resting upon the crest of an ancient ridge cut in the Naushon sands, which are considered the equivalent of the Truro series. At no point, so far as I have observed, does the moraine appear to be more than 25 feet thick. In all the observed localities it evidently rests on the top of a divide formed before the advent of the ice, and it is lacking over large areas, where the stratified sands appear with-only occasional bowlders resting upon the surface. Of the mass of material composing the Elizabeth Islands above the plane of the sea, certainly not a tenth, and perhaps not a twentieth, part is of morainal nature. The rest may be of glacial origin, but if so, it was deposited far in advance of the ice and long before the advent of the glacier in this part of the field.

On Marthas Vineyard the main or northern moraine appears at first sight—and even after some inspection—to be made up of bowldery material, but on careful investigation I have found that it is a pre-Glacial ridge, the pedestal being formed of a stream divide cut in the Cretaceous and Tertiary strata. On account of the misleading appearance of the ground, I was led, in my report on this island,[1] much to overestimate the depth of this glacial wall. It has not half the mass stated in that report. It is doubtful if the average depth of the deposit exceeds 40 feet. The ridge occupied by the moraine is not completely covered with the deposit. For considerable distances the top of the elevation is essentially without materials which, from their character or distribution, may be classed as truly morainal. At other points, especially in the middle portion of the belt, on the estate known as "Seven Gates," the deposit constitutes a very characteristic morainal belt, with numerous large kettle holes and with bowlders in such abundance that the masses appear like ruined cyclopean masonry. The southernmost moraine of the island also occupies the summit of a divide, but the erratic element is small in amount and only here and there assumes a morainal character.

In the territory between the two moraines of Marthas Vineyard there are, as before noted, many fields which are so far free from glacial waste that they may fairly be termed driftless. It is not easy to find any material on them that may be classed even as till. Rarely is there a foot, in depth, of this deposit. This driftless character of surface is so complete that the plowshare will turn up Tertiary or Cretaceous beds containing no trace of erratics. Fields of this driftless soil some acres in extent lie within 2,000 feet of the wall-front moraines. On the north side of the principal moraine the same phenomenon of nearly driftless fields is observable, but in a less distinct manner. The areas without erratics are small, and those which are quite without till are at no point, so far as I have observed, more than an acre or two in extent. They occur at the foot of the slope on which the moraine lies, usually

[1] Seventh Ann. Rept. U. S. Geol. Survey, 1886, p. 312.

quite near the sea level. This feature of small driftless areas over which the morainal matter must have passed on its way to the glacial front indicates that the conditions which determined the deposition of the detritus were peculiar.

There is another peculiarity of the Marthas Vineyard moraines which appears to throw some light on the conditions of their formation. While commonly the ridge of detrital material is placed on the very crest of a divide, adding distinctly to its height, it not infrequently is deposited as a sheet on the southerly or outer face (outer in relation to the glacial movement) of the ridge on which it was formed. The effect is as if the materials had been pushed up the northern slope and had fallen into the attitude in which they are found.

In the case of the Cape Cod moraine the evidence is sufficient to show that the mass occupies the considerably elevated surface of a ridge which was formed before the advent of the ice. This ridge continues with no complete interruption from the base of the cape to Orleans, except for the rather deep and wide break at Bass River. So far as I have been able to determine, the moraine is gathered mainly on the southern side of this ancient divide, though it generally rises somewhat above the crest line. The feature noted on Marthas Vineyard, of considerable areas of the more ancient deposits without glacial waste, is noticeable to the north of this moraine, but is less extensively developed than on that island.

On Nantucket the moraine appears to crown the summit of an elevation composed of the Sankaty beds, which probably belong to a time immediately preceding the deposition of the Nashaquitsa series. The conditions are not clearly indicated, but there is little reason to doubt that this relatively unimportant accumulation is placed as are the others above noted.

The facts above described warrant the statement that all the characteristic morainal accumulations of this district are placed in singularly close relations to the crests of ridges which existed before the ice sheet invaded this district. The few apparent exceptions prove on examination really to be not such. Of these the most striking is the case of the northern moraine of Marthas Vineyard, where the bowldery deposit descends into a valley about the headwaters of Witch Brook. In this relatively low place, which still is about 70 feet above the sea, the moraine becomes somewhat scattered. It is, in effect, a rather flat, very stony field, in place of the well-defined accumulation exhibited on the higher ground on either side. So, too, the same morainal, detached hills which lie here and there on the north slope of Cape Cod and Marthas Vineyard appear on inspection to be small elevations of the Barnstable series, which bear some coarse drift or else masses dropped from a stranded iceberg.

The relations of the moraines in this district can not be explained on the supposition that these deposits are revealed only on the highlands, being elsewhere covered by later accumulations of washed sands. This suggestion would not be entertained for a moment by any student of the district who approached the problem without a decided preconception; although, as I know from experience in making the examination, it requires rather careful observation to avoid the mistake of supposing that the whole mass of these several ridges is of a morainal nature.

So far as I am aware, none of the moraines of the central and western portions of the country are placed on the crests of divides in the manner shown to be the rule in this shore-land district; nor am I aware that any of the accumulations in the interior parts of New England occupy the crests of ridges, except when their bases may accidentally coincide with those elevations. It therefore may be fairly assumed that there has been some peculiarity in the condition of the glacier in this part of the country which has served to bring about the curious result. I have not been able to determine the precise cause of this occupation of the preexisting divides by glaciers, but the possible explanations appear to be but two, and these will be briefly stated.

The first hypothesis is that the ridges may have served in forming a moraine by arresting the flow of the ice, already languid in its movement, for the reason that it had become attenuated at its margin. Hanging on these crests its front may have been retained in one position for a considerable time, which permitted the accumulation of the morainal deposit. The difficulty with this view is that it does not explain the absence of drift in the fields near the well-developed morainal lines. The second hypothesis is that, the region being depressed beneath the sea to a considerable but unknown depth, the ice, while remaining as a united sheet, may have floated over the valleys, grounding only upon the ridges and there depositing its contents of rock material. The portion of the ice which was shoved over the crest was probably broken into fragments which floated away. This hypothesis will explain the absence of till on much of the lowlands where, had the ice rested on the surface while it melted, it should remain to mark the decay of the sheet.

The hypothesis of the partial floating of the attenuated ice sheet finds a certain amount of support in the evidence which goes to show a considerable subsidence during the period of formation of the sand plain in front of the moraines. As I have elsewhere shown, these plains were certainly formed under water while the land lay at least 100 feet lower than its present level, and the actual depth of the submergence may have been much greater than the minimum required for the construction of the plains. The hypothesis will perhaps serve to explain also the departure of the Cape Cod moraine from the normal direction. As before remarked, it is very difficult to see how a glacier moving under ordinary conditions would have a path parallel to the shore; but if the sheet be conceived as floating in the sea, though its course might be generally eastward, its margin might be pressed continuously or from time to time against the submerged ridge on the south, on which would be dropped, as the ice sheet shattered or broke up into bergs, a portion of the contained débris. The main objection to this hypothesis is that evidence as to the actual floating of the ice over the valleys is lacking; yet it demands only conditions which must exist wherever a glacial sheet enters the sea, pushing out into water so deep that the mass leaves the lower-lying parts of the bottom. Somewhat in its favor is the fact that over the nearly driftless fields near the Marthas Vineyard moraine occasional large solitary erratics are found, and sometimes heaps of coarse débris in positions which suggest that the materials have fallen from the base of a floating glacier or from an iceberg.

It must be confessed that both hypotheses present serious difficulties; but in view of all the facts, the one last stated is more satisfactory than that of marginal lobes producing interlocking moraines—a hypothesis which does not seem applicable to this field.

### RELATIVE AGE OF THE MORAINES.

The relative age of the moraines in this district affords an interesting field for inquiry. The only criterion which appears to be accessible is that which may be derived from the comparative amount of decay of rocks of apparently the same measure of resistance to such change. Judged by this test, the moraine of Cape Cod is to be regarded as rather newer than that on Marthas Vineyard; the bowlders of like petrographical species are less broken up, and interstitial decay has penetrated to a much less depth. These determinations are based on mere inspection, but the impression thus obtained by many successive visits to each field at short intervals is clear.

Much evidence as to the petrographical nature of the hidden rocks of southeastern Massachusetts and the neighboring sea bottom can doubtless be obtained from a careful study of the materials in these moraines. This task has not been formally undertaken in the preparation of this report, but incidentally certain points have been noted, of which only one need be mentioned here, viz: On Marthas Vineyard the drift abounds, in a remarkable manner, in masses of chalcedony, some of which are a foot or more in diameter. The pebbles are so numerous that many tons could be gathered on a mile of beach on the north shore of the island. This material is not found on the mainland, nor is it known on the moraine of Cape Cod. It is therefore probable that it was riven from the area now covered by the sea.

### CLAY BOWLDERS IN TILL.

The till of Cape Cod, especially where it occurs in the moraines, or has a morainal aspect, occasionally contains large masses of clay which evidently were brought to their resting places in the manner of other erratics. These masses can be found scantily at several places on the cape. They were most clearly shown in an artificial escarpment which for some years existed near the steamboat wharf at Woods Hole, the site of which is now occupied by the Nobska House. In the excavation of a low drumloid hill which existed at the place just mentioned, a dozen or more of these clay bowlders, varying in diameter from a few inches to 6 feet, were noted by me in the course of three or four visits to the locality. Traces of the same bowlders have been seen on the northern slope of the cape and in the district about "Quisset" Harbor. As such bowlders have not been found on Marthas Vineyard in the numerous sections of similar drift materials, it is desirable to seek an explanation of the peculiar limitation of their occurrence. These conditions seem to have been as follows:

The clay of which these bowlders was formed was of a tenacious, uniform quality. Although much oxidized, it was seen to be of the same general character as that found in the brick-clay pits of the region about North Barnstable. That the clay was rather soft when it was moved is indicated by the fact that the surfaces of the masses were crowded with pebbles, in the manner in which lumps of clay, made by the waves on the seashore from the waste of clay cliffs, are coated with a layer of pebbles which have been pressed into the mass. As the glacier evidently slipped over the surface of such clays wherever that surface was of a continuously sloping form, as it now is on the northern versant of the cape, it seems likely that these bowlders were riven from areas where the ground was cut into deep ravines after the manner of the so-called bad-land topography—conditions which would favor the formation of erratics. That such irregularities existed in the cape area is sufficiently shown by the irregular "noses" or projections of clay—the so-called clay "pounds" which have been dug into here and there to obtain materials for bettering the sand roads of the district. So far as I am aware, clay bowlders as large as those at Woods Hole have not been found in other regions. Their rare occurrence is perhaps attributable in part to the fact that a deeply indented topography, formed in soft clays, has rarely been so eroded by an ice sheet. (See Pl. XCIX.)

## LENTICULAR HILLS.

The class of drift deposits known as lenticular hills or drumlins is practically wanting in the southeastern portion of Massachusetts. There are no instances in which elongated arches of till are sufficiently well developed to merit a place in this group. Here and there, however, are morainal hills which show distinct traces of the action which gives rise to these regular forms. The ridges north of Woods Hole, between that village and Quamquisset Harbor, closely approach in shape what would be termed drumlins of the lowest order if they lay in the central part of Massachusetts. So, too, on the north side of the cape, some of the drift hills in the town of Bourne show traces of a like regularity of outline. On Marthas Vineyard several of the morainal ridges in West Tisbury and Chilmark, especially that known as Prospect Hill, are distinctly of a drumloid form.

The arched hills of the cape district are, so far as I have observed, limited to the higher ground, and they approach more closely a symmetrical form as the altitude above sea level increases. This is not the case in the more northern parts of the coast of Massachusetts, for about Boston Harbor and in Ipswich very perfect specimens of the type are formed rising from the sea level. In any discussion as to the origin of these curious topographical forms this peculiarity of their distribution must be considered.

It is noteworthy that in this district the drift hills are more shapely on the side against which the ice moved than on that which was turned away from the stream, and also that the ridges of the pre-Glacial topography cut in the sands, gravels, and clays of the Cretaceous, Miocene, and Pleistocene formations have been in many cases rounded into drumloid forms. This is particularly the case on the western side of Marthas Vineyard, though instances of the same nature exist on the central and western parts of Cape Cod. Some of the hills near the Chatham Harbor margin of the Truro beds have a distinctly drumloid outline. (See Pl. XCIX.) These facts clearly point to the conclusion that whatever may have been the cause which led to the local deposition of the deep sections of till composing characteristic drumlins, the final shaping of these forms was due to the action of the ice as it passed over them.

## WASHED DRIFT.

In this field, as elsewhere in New England, the washed drift may be divided into three tolerably distinguishable groups: eskers (nearly absent here), pitted plains or kames (rare), and sand plains or morainal aprons. It is to be noted that the materials composing these deposits differ less from those of the ordinary till than is usual in the more northern parts of Massachusetts. Here the till itself is always very sandy, its pebbles are much rounded, and the clay element, as compared with the more northern localities, is relatively small in quantity. This feature is probably due to the diminished cutting power of the glacier on its outer margin and to the extent to which its detritus was worked over by the water which flowed beneath the ice on its way to the front of the sheet. The result of these actions was to diminish the total amount of the till and to make the remaining portion much sandier here than elsewhere. As a consequence of this, it is often difficult to distinguish between the drift which has been deposited in water and that which has been left upon the surface by the melting of the ice in which it was contained.

One of the eminent peculiarities of this district is the general absence of eskers. So far I have not been able to find any characteristic examples of these structures on Cape Cod. In the region from Bass River to Orleans there are certain ridges extending in a north-south direction which may possibly belong to this group of deposits, but I suspect that they are the remains of the ancient topography cut in sands of the Truro series. I am the more inclined to this view for the reason that the one ridge on Marthas Vineyard which I identified in my report on that island as an esker has since been proved by a section to be an old pre-Glacial feature, slightly modified by a coating of washed drift or very sandy till.

The probable absence in this field of eskers of the molds of the caverns in which flowed the subglacial streams goes to support the hypothesis that the ice in this section did not generally rest upon the surface, but came in contact with it only on the higher parts of the ground. Thus floating, there would be no chance for the development of the ice-roofed channels, the shapes of which became elsewhere molded in the débris with which they were in time filled. As these eskers descend to the level of the sea in the district about Boston, and are found in the region north of a line stretching from Boston to Narragansett Bay, it may, perhaps, be inferred that the conditions which are indicated in the Cape Cod district were of a rather local nature.

Pitted plains of the type so common in the districts where eskers exist are not often found on Cape Cod or in the islands to the south. The only good examples are on the frontal apron south of the moraine, where ice remnants, icebergs, or ground ice left by the retreating glacier appear to have been partly buried in rapidly accumulating sands, leaving where they melted depressions to indicate the positions they occupied. A trace of the same action is found in the central part of the great plains of Marthas Vineyard, where the occurrence of a small lake with steep sides can be accounted for only on the supposition that its site marks the place where a stranded iceberg was buried in the accumulation of sands which constitute the mass of the morainal apron.

An ordinary type of kame deposits, consisting of a number of hillocks of arched form huddled together quite without definite arrangement, a type very common in the town of Plymouth, appears to be lacking in the cape and islands. This peculiar local topography of the washed drift can most readily be explained by supposing that when the ice came to the point where it ceased to rest in the bed rock and began to float, its under surface would for a time retain the form impressed upon it by the contours of the surface over which it had flowed. There would thus come to be a space between the base of the rotting ice and the sea bottom into which the débris coming from the land would naturally be crowded. If the ice had much movement the resulting shapes of the drift would of course be destroyed, but at a late stage in the retreat of the glacier its stagnation might be so far complete as to leave the molded sands and gravels as we find them. The occurrence of this remarkable kame topography on the mainland and its absence on the neighboring peninsula and islands is what we should expect on the hypothesis that the ice was in part afloat in this portion of the field it occupied.

The characteristic form of washed drift occurring on Cape Cod and the neighboring islands is the deposit of sand and gravel laid down in front of the moraines. These deposits are more extensive in this district than in any other known to me except, perhaps, on Long Island, New York. It is to be noted that these great morainal aprons differ in certain ways from the sand plains of the mainland. On that field the plains are in most cases at the end of distinct eskers, and clearly mark the places where a subglacial stream passed into the open air or open water. They are rarely, if ever, distinctly related to the axis of defined morainal ridges, though we often find bowldery tracts at about the point where the esker passes into the plain. In these morainal aprons of the cape district, on the contrary, there are no eskers leading to them, but the broad field of sand extends up to or near the wall of coarse débris. Next this wall there is commonly a shallow, wide depression, from which the apron rises to a point somewhere about a mile away, whence it declines to the sea. Such is the form of the great plains of Marthas Vineyard and Nantucket. In Cape Cod the depression or ditch is less distinct; it will, however, be remarked by an observer who has noted the feature elsewhere.

While the ordinary sand plains have their "feeding" eskers—molds of the channels through which the débris came—those that front the great moraines of the cape lack these features. Here and there are breaches or low places in the morainal walls through which currents of water appear to have flowed, as is shown by the signs of erosion in the channels in front of these breaches. The plain exhibits broad, irregular channels which lead down to the sea. These scour ways do not appear to have been at any time occupied by open-air streams, but rather to have been excavated on a water-covered surface. This feature, like the ditch in front of the moraine, is less distinct in Cape Cod than on the neighboring islands, yet it is disclosed to close inspection and partly indicated on the topographical map. In reports on the geology of Marthas Vineyard[1] and on the geology of Nantucket[2] I have given in some detail an account of the characteristics of the plains that lie in front of the moraines on those islands. The like deposit on Cape Cod differs from those noted in the papers referred to in that it is ruder in form, that it has numerous considerable lakes on its surface, and that the scour ways are generally occupied by brooks.

The peculiarities of the morainal apron on Cape Cod, taken along with the evidence of beds of clay apparently belonging to the Barnstable series, lead to the conclusion that in place of the very deep deposit

[1] Seventh Ann. Rept., U. S. Geol. Survey, 1885-86, p. 316.
[2] Bull. U. S. Geol. Survey No. 53, 1889, p. 19.

of sand which exists beneath the plain on the above-named islands we have a relatively thin layer of detritus imposed upon a preexisting topography which is cut in rather impervious beds. The numerous swamps and lakes, as before remarked, so high above the sea that their waters could not be retained by sand barriers, are probably to be in part accounted for by the supposition that they lie in valleys which originally drained northward, as appears to have been the case with nearly or quite all of the pre-Glacial streams of the cape. These streams were dammed by the moraine. In perhaps larger measure, however, these basins are to be regarded as the molds of ice remnants about which the washed sands were gathered. That such was the case is shown by the fact that the sides of the depressions are usually very steep, the detritus having slopes which it could not have assumed at the time of its deposition unless there had been some barrier, such as the walls of ice would have supplied, to keep it from being conveyed into the cavity.

The contours of the great plains of Cape Cod, like those on the islands, clearly indicate that the material was deposited under water. In aerial overwash plains, formed as detrital cones, we find necessarily a continuous down-sloping surface. In these plains in front of the glacier of southeastern Massachusetts the surface has the gently rolling character characteristic of sands that have been arranged on the bottom of a sea which was the seat of tolerably strong currents. The slope of the Cape Cod morainal apron is essentially the same as that of the similar structures in this district, the rate of the decline to the seaward being from 12 to 15 feet to the mile.

The surface of the plain of Cape Cod is prevailingly composed of rather fine, siliceous sand. This material forms a bed having a rather remarkably even thickness of from a foot to 18 inches. This usually passes downward by a rather sharp transition into a pebbly layer in which the pebbles are from the smallest sizes up to that of a cricket ball, though rarely so large. At greater depths the admixture of sand and pebbles is rather uniform, the mass having obscure stratification. Now and then a bowlder is found. These bowlders are almost always rounded and rarely exceed 2 feet in diameter; they are often found in groups associated with gravel, and they occasionally occur on the surface. Such stratification as is exhibited is not distinctly cross bedded. In these, as in most other features, except the presence of numerous lakes, the cape plain in no way differs from the like structures in the other parts of the district.

In considering the origin of these morainal aprons of southeastern New England, the fact should be noted that deposits of like nature do not, so far as I am aware, exist in front of the moraines in the interior of the country. There are there, it is true, traces of overwash plains, but they are always much less continuous; they have, in a word, more of the nature of detrital cones. Those I have seen in Ohio, Michigan, and Wisconsin also lack the depression next the moraine, the pits occupied by lakes such as occur on Cape Cod, and the scattered bowlders in the mass and on the surface of the deposit. The difference between the structure in the two districts probably arises from the fact that those in southeastern Massachusetts were formed under water, while those in the west were deposited mainly in the air. If we suppose that the sea extended up to the ice front, and that the finer materials were, at the time of melting, given into the control of tidal currents, we can well conceive that the part of the débris which could be thus transported would receive a wide distribution over the neighboring bottom; the floating ice would convey many bowlders from the front of the moraine, dropping them haphazard as they melted; the tidal currents would carve channels on the bottom as they cut them on any sands over which they may flow. In a word, the assemblage of conditions exhibited in the morainal aprons is more consistent with the supposition that they were formed on the sea floor than in any other manner.

It will be noted that a number of the peculiar features of these moraines together tend to show that this district was rather deeply submerged at the time of their formation. I have, as yet, been able to find no evidence going to show whether the submergence was so deep as to cover the tops of the morainal walls. It may be noted, however, that on Marthas Vineyard the portion of the moraine which faces on Tisbury River has no apron on its front, but rather a steep overwash plain or long detrital cone which terminates in a valley that may have carried the wash from the glacier down to the neighboring great apron. It thus seems probable that this portion of the morainal front lay above the level at which the sea was placed at the time it was formed.

## OUTER LIMITS OF THE CAPE COD ICE SHEETS.

In view of the fact that the ice sheet on this portion of the Atlantic coast was evidently thin, the question arises as to its probable extension beyond the limits to which it can be traced by the remains it has left upon the land. On Cape Cod we find in the Truro-Wellfleet district very slight evidence—if it be, indeed, evidence at all—that the ice lay upon the surface. I am quite prepared to believe that the drift in this area is due to the action of floating ice dropping the waste it carried upon the bottom. We may from the evidence fairly conclude that we are here near the eastward margin of the effective ice sheet.

On the south the extension of the glacier appears to have been to a relatively farther point than in the east. On Nantucket there is a small area of low but fairly characteristic moraine with a well-developed frontal apron. On the island of No Mans Land, south of Gay Head, we have an extensive deposit of a till-like nature, which may, however, be due to floating ice. On the southernmost of the Elizabeth Islands the glacial drift, though scanty in amount, is still sufficient to attest the presence of the ice in that part of the field. It thus appears that the glacier probably extended its action over all the district of Cape Cod, though on the extreme south and east the effects which it exercised may have been due to portions of the ice which had been broken from the united mass and which were floating in the open sea. It is hardly to be supposed that a sheet so thin as the glacier was in this part of its course could have held together for any considerable distance from the shallows, capes, and islands where we last trace it.

In closing this portion of the report attention may be called to the value of the information concerning the frontal condition of the glacier which southeastern Massachusetts affords. In no other section of the country are the data for inquiry so ample—and, it must be confessed, so difficult to interpret.

## POST-GLACIAL DEPOSITS.

This group of deposits includes the spits, hooks, and beaches, the dunes, the marine marshes, the fresh-water swamps, the soil, and, finally, the sea shoals.

One of the first-named group of constructions we have in the hook which constitutes the whole of the area of the town of Provincetown, one of the finest existing examples of such forms. There is certainly none other of its kind in this country which so well deserves attention. The history of this feature appears to have been in general as follows:

When, after the disturbances of level which attended the last Glacial epoch, the land of Cape Cod came to its present apparently stable attitude, the elevated country of Truro extended somewhat farther to the north and east than it does at present. As this last portion of the cape in the east was worn down by waves and currents in the manner in which the work is now going on, the débris was, in part at least, carried to the end of the land, there beginning the growth to the northward of the spit. As noted by Prof. W. M. Davis, the sea beach at the north end of the Truro highland marks the point where the encroachment of the sea was arrested by the beginning of the accumulation of sands which has extended to the village of Provincetown. (See Pl. CII.)

It seems likely that there was shoal water where this spit was formed; before it began to form, indeed, the erosion of the northern face of Truro, which has just been noted, may have been part of a considerable wearing that had gone on before the spit had begun to form. There is a bit of evidence on this point drawn from the results of the "driven" wells sunk in the sand at Provincetown which, though not certain, has some value on this point. These wells, which were sunk to a depth of a few feet below the level of the sea, in place of yielding the very pure water which elsewhere has been obtained from such spits, have afforded a quality which, on account of the large amount of iron it contains, is hardly fit for use. Water of the same nature is characteristic of the old sands of pre-Glacial age wherever they have been tapped in this part of Massachusetts, the defect being due to the complete oxidation of the considerable amounts of iron which they contain, and perhaps to other chemical changes, such as do not occur in the clear siliceous sand of which the spits and beaches of this region are made. It therefore seems probable that the water of the Provincetown wells is drawn, not from the beach sands, but from the lower-lying pre-Glacial deposits.

The process of growth of the Provincetown hook appears to have been mainly by successive beaches, each formed in front of the next preceding, and each projecting northward somewhat beyond its predecessor. The supply of sand seems to have come, in part at least, from the wearing of the coast line of the Truro-Wellfleet district, and in part from the sea bottom to the eastward. There has evidently been a balance of actions which of late has served to urge the sand to the northward toward the present end of the cape. It is evident that for a time no distinct hook existed in the end of this spit; its form was somewhat like that of Monomoy, but, probably for the reason that the water deepened beyond the shallow on which it at first grew, the end near its present stage of growth turned westward to form the hook with which it now terminates.

At first the Provincetown spit was evidently narrower than it is at present, but with the carriage of sands northward along the shore the water on the side of the open sea was shallowed by the formation of a broad shelf which enabled a succession of beaches to form, each somewhat farther out than its predecessor, and in this manner the spit has been considerably widened. This process of growth appears not to be continuous. From time to time, with the varying direction and energy of the waves and of the currents which they induce, the beach works in, again to be built out with the resumption of the carriage of waste from the shore southward.

Along with the carriage of sand by the sea there has gone a considerable movement of materials by the wind. This has taken place mainly in a westerly direction from the outside beaches. When the tide is out and the air dry, even a moderate wind will move the finer parts of the material almost as easily as though it were snow, and in great storms quartz pebbles up to the size of peas may be observed to fly along at the height of some feet above the earth. As the wind loses a part of its speed in passing over the surface of the ground, the particles of sand and gravel which it bears soon fall into the eddies of the current, there forming the beginnings of dunes. As soon as these dunes form they begin to march before the wind; the bits slip up the exposed side and pass over the crest into the sheltered sea, where they remain at rest until the whole mass has been shifted forward in the same manner. In this way, by the process of constantly moving the windward layer to the leeward side, the dune slowly marches inland.

Various influences tend temporarily to arrest the march of these Provincetown dunes, as they do all such masses of wind-blown sand. As the bits journey they decay, so that they naturally cement together. Moreover, certain species of plants, such as the beach grasses, have developed the capacity to grow in the arid soil of these ridges. This they do so effectively that their roots and leaves make a mat which deprives the wind of access to the heaps.

In the present state of the Provincetown spit hook the structure appears, as a whole, to be in a tolerably balanced state, a condition into which such structures are apt to come at a certain stage of their growth. The cape does not appear to be extending northward, unless it be very slowly, the tidal currents from the bay interfering with this growth. The hooked extremity, which is made up of detritus that washed around the end of the cape, does not appear to have gained in extent in a material way during this century. The only change which menaces the established order of this unstable new land is the present inward movement of the beach on the eastern side, near Moon Pond. There we have a well recognized danger that the sea may break through, with the result that the valuable harbor of Provincetown would be endangered. It certainly would become shallower, and it might be so far changed as to lose its present great importance as a port of refuge.

The erosion of the sea on the eastern face of the cape from the northern end of Truro to the central part of Eastham has provided not only the sand which has gone to construct the Provincetown area, but also that which, moving to the southward, has built the large and beautiful line of barrier sand beaches that extends from opposite Chatham Center to the end of Monomoy Island. Although this isle is at present separated from the mainland of the cape by a shallow water way, it is, in its structure, a spit of the same general character as that at Provincetown, only less far developed. Already at its southern end it has begun to form the hook, which is the appropriate finish of such spits.

The amounts of débris which have gone both ways from the erosion district of the eastern face of Cape Cod appear to be nearly equal. The reason why the Provincetown spit is so much longer than that of Monomoy is, that the greater part of the sand which moved southward has been used in constructing the extensive barrier beaches that lie on the sea side of Orleans and Chatham; for these long and broad walls of sand probably contain rather more material than is held in the much more conspicuous spit hook at Provincetown. At present the Monomoy spit appears to be growing more rapidly than its northern equivalent, so that in time these two geographical growths may become even more alike than they are at present.

On the northern shore of Cape Cod, although there are no parts of the shore which are undergoing erosion, there is an interesting system of barrier beaches, which has been constructed since the land assumed its present level in relation to the sea. The material for these beaches has evidently been derived from the shallow bottom of the adjacent bay, it being dragged in to the shore by the action of the waves. It will be observed that while on the eastern and southern sides of the cape these beaches are always drawn near the shore, so that the lagoon they shut in is quite narrow, those on the western and northern shores depart widely from the coast, so that they inclose broad fields of water, such as Wellfleet Bay now presents or such as were found at Barnstable before the harbor was narrowed by the extensive growth of marine marshes. The reason for this more remote position of the barrier beaches in relation to the shores is, that the water on the north side of the cape appears to have been, in the beginning of the present conditions, as it is at present, shallower and with a more gently declining bottom than it had on the south. In fact, the old river basin, which is now Cape Cod Bay, had evidently a much more gradual slope than had the neighboring basins on the south. Thus the ancient form of the basin has served to qualify the shapes of the existing shores.

On the southern side of the cape the evidence of coastal erosion is somewhat the same as it is on the eastern part of the area. In certain places along this shore there are evidences of considerable but variable localized coastal erosion, the waste from which is distributed along the shore and accumulated in slight barrier beaches and hooks. Of these the most interesting is that known as Point Gammon, at the mouth of Lewis Bay. At certain places, as, for instance, at Chatham lights, observations show that for a number of years the recession of the shore went on in a singularly rapid manner, at the rate, it is said, of 10 feet per annum. It is evident, however, that this was a local adjustment of the shore, caused, it is now asserted, by the development of the beach which extends to Point Gammon, and the consequent change in the distribution of the wave action. The amount of erosion on this southern shore has probably been but a fraction of that which has gone from the eastern face of the cape, where, in Truro and Wellfleet, an extensive salient, probably amounting in area to not less than 30 square miles, has been cut away to afford the débris which has been distributed on the beaches, spits, and hooks on the north and south. That the erosion on the eastern face of the cape diminishes in a westerly direction is shown by the unembarrassed outlets of the streams which enter the sound along this shore. If there had been any considerable amount of erosion here, the sands therefrom would have been gathered in adherent and barrier beaches and spits, such as exist along this coast wherever the sea has been supplied with materials from which to make these constructions.

On that portion of the western face of Cape Cod which is bordered by Buzzards Bay we find but little evidence of marine erosion. There are a few very small spits, but no barrier beaches; in fact, there are few portions of the coast south of Boston which are exposed to waves of moderate severity where the amount of work done by the sea is so small as it is here.

This glance at the shore conditions of Cape Cod shows us that only a small part of its periphery indicates any considerable amount of wasting since the land came to its present altitude. The maximum recession can not well amount to more than 3 to 4 miles. This occurred on the eastern side of the Truro-Wellfleet coast. The next most considerable loss is on the section near Hyannis. The best evidence as to the limited amount of the loss of area is afforded by the fact that the extent of the cliff shores of the area is limited; even the frail materials of the morainal aprons have not been much cut away, as is shown by the fact that their slopes are, with rare exception, prolonged down to the level of the tide. Had they been much eroded they would face the shore in steep cliffs.

Owing to a considerable local erosion which has taken place on parts of the shore of Cape Cod, there has come to be a general opinion that the peninsula is in process of rapid destruction. This view appears to be held by many well-informed residents of the peninsula. So far is this view from being true that the converse may be taken as nearer the facts. It is altogether likely that the total area of this cape country, including all the marshes, barriers, beaches, spits, and hooks that are attached thereto, is no greater than it was at the time when, by a final step of subsidence, it established its present relations of land and sea. The aggregate of this erosion is evidently much less than that which has taken place on the islands to the south.

The submarine constructions which have been made by the tidal currents in the waters about the cape are probably, in mass, much greater than are those which appear above the plane of low water. An inspection of the Coast Survey maps discloses in the soundings a curious tangle of shoals, mostly ridge like in form. As before remarked, some of these submarine elevations are probably the divides of the smaller streams which intersected the floors of the valleys at the time they were above the sea. This is clearly the case with Stone Horse Shoal, and it is most likely so with the middle ground of Vineyard Sound. Others, especially the group about the eastern entrance to Nantucket Sound and that at the north end of Muskeget Channel, are evidently due to the action of the strong and contending tidal currents which sweep through these areas of sea. It may be noted here that the absence of any signs of marine current action on the surface of the land of Cape Cod or the neighboring islands and mainland above the level of the sand plains is tolerably good evidence to show either that the Glacial submergence did not extend above that level or that, if more deeply submerged, the ice remained on the surface until the land was reelevated to about its present height.

## MARSH AND SWAMP DEPOSITS.

The marine marshes of this district are of considerable extent within the limits of the cape; their area is about 11,000 acres, the greater portion lying on the north side of the isthmus, in the Barnstable and Wellfleet reentrants. On the south and west coasts they are distributed in numerous small areas along the banks of the smaller drowned valleys and in the lagoons lying between the barrier sand reefs and the shore. As compared with the similar marshes north of Boston Harbor, these of Cape Cod exhibit a much less energy of growth. Basins which there would have been occupied by completely developed deposits are here but imperfectly covered by them. The reason for this deficiency is not to be found in any change of species, for these are the same in both districts, but probably in the fact that the amount of mud swept in by the tide is here very small as compared with what it is elsewhere; the result is that the plants are ill fed and do not attain anything like the vigor of growth which they exhibit when the water at each flooding brings much nutritious material to the plant roots. Moreover, in these sandy bays the eelgrass, which is the most effective agent in preparing the shallow water to be occupied by the marsh-making growth, does not do so well on the bottoms of drifting sand as it does on those of firmer and more supporting nature, such as are found to the north.

The fresh-water swamps of Cape Cod were originally very numerous. Though by far the greater number of them have been drained for use as cranberry plantations or converted into reservoirs to flood the vines in the proper season, some areas still remain in their natural state. In its original condition this district had a larger share of swamp grounds than any other equal area in this part of New England, and the inundated fields were more evenly distributed than elsewhere.

The reason for the great development of swamps on Cape Cod is to be found in the fact that there is a clay underlay beneath the glacial sands on the greater part of the area. Thus, the plain of the morainal apron, which in the equivalent deposits of Marthas Vineyard and Nantucket, because it is of pure sand to a great depth, is almost destitute of swamp deposits (that of the first-named island being quite without swamps), is on Cape Cod beset with lakes and with swamps which have grown in lake basins; moreover, the ridge of the old divide on which the moraine rests is wide and rather flat, which favors the development of many areas of imperfect drainage. These conditions have served to give to this region its long-continued predominance in the industry of cranberry planting. There is probably no other place where the very peculiar conditions required for this singular form of agriculture have been so well assembled.

The fresh-water swamps of this region are much better developed than are the marine marshes. The climate and soil are so dry that there is no trace of the climbing action of the bog sponge which is so common in Maine and is still notable about Boston. This general dryness somewhat arrests the growth of the peat deposits, but it favors that of many species of bushes and some trees, such as the swamp maple and the tupelo. The result is that a large part of the peaty matter of the bogs in this district is due to the leaves and stems of phænogamous plants. On this account the bog soils are evidently more fertile than are those formed by the decay of mosses alone. It is in part to this cause that we must attribute the excellence of the cranberry plantations.

It is a noteworthy fact that a very large proportion of the lakes in the cape district have escaped the action of the swamp-making agents. Many of these basins not exceeding half a mile in diameter show no trace of peaty growth about their borders. This feature is probably due in part to the very sandy character of the shores, which makes it difficult for the mosses to become implanted there—a difficulty which is the greater for the reason that the range in the level of the water is very great. In part the hindrance arises from the considerable depth of many of these ponds and the steepness of their beaches, which makes it hard for the water lilies and rushes to take root, so that the protection which their stems afford the shore from the assault of the waves is lacking. The result is that the frail beginnings of a moss plantation are likely to be broken up long before the growth has attained the strength which would enable it to resist the action of the waves.

## SOILS.

The soil of the cape differs little from that of the neighboring districts of the mainland and the islands On the north shore, from the base of the peninsula to Orleans, the general presence of the Barnstable clays or the clayey till made therefrom causes the fields to retain moisture in a way they do not in the more southern and eastern sections. On this account, rather than for any special nutritive value in the underlying material, the soil here is considerably better suited to farming than elsewhere. The vegetable matter on which the fertility of the earth so largely depends does not pass out by decay as speedily as it does in the excessively porous débris which generally underlies the surface in these fields.

Owing to the exceedingly sandy nature of the till, except, as before remarked, where it rests upon the clays of the shore, there is little difference between the soils formed on it and those formed on the sand plains lying south of the moraine; in each condition the portion of the earth which is mingled with decayed organic matter, i. e., the true soil, is rarely more than 6 inches in thickness. As the mineral matter in the drift beds of this region is exceedingly well adapted to afford the mineral elements required by vegetation, the failure of a soil to form is rather curious. The reason for the condition seems to be that the exceedingly porous nature of the earth affects plants injuriously; in the first place by causing their roots to become very dry shortly after a rain, and in the second place by permitting the speedy and complete decomposition of the decaying organic matter, so that the earth is without the necessary amount of humus. The validity of this hypothesis is shown by the fact that wherever we find a place in which the water table is retained sufficiently near the surface to permit the tilled zone to be moistened by capillary attraction from below, there we find excellent ground for tillage; moreover, wherever the plan of plowing in green crops is followed, the results show that the soil needs only suitable treatment to give excellent returns. A considerable personal experience in tilling such soils as the sandier kinds of Cape Cod enables me to say that where they can be irrigated and where they are provided with nitrogenous matter by the inexpensive plan of plowing in crops of peas, clover, or other leguminous plants, they can be made to yield profitable crops.

It is particularly desirable to have the treatment of these soils of southeastern Massachusetts made the subject of a special and well-directed inquiry. In this district we have an aggregate area which may be safely reckoned at not less than 150,000 acres whereon all efforts at tillage have ceased. The region was once fairly well wooded, but the forests have long since been cut away and their regrowth is prevented by the numerous fires which sweep over them and which still further reduce the amount of vegetable matter in the soil. These fields, when unwooded, are sold, in the rare transfers which are effected, at from 50 cents to about $3 an acre; in their present neglected condition they are really not worth any price. In view of their nearness to rail and water transportation they should invite the attention of persons who are willing to take the pains necessary to learn the most economical methods of bringing them into tillage. Sixty years ago the swamps of this district were even more unpromising fields for agriculture than these sand plains and hills, yet at the present time, in their condition as cranberry bogs, they are worth on the average more than $100 an acre over and above the expense of bringing them under cultivation.

Of the total area of Cape Cod, only about one-eighth is so occupied by morainal matter as to be untillable; about another eighth is contained in the sand spits and beaches; so that three-fourths of the whole area is, so far as the geological conditions go, fit to be made into soil, and will doubtless in time be brought under cultivation. The morainal fields afford excellent ground for the culture of forests; several species of trees do well on this bowldery earth, among which may be mentioned the white pine and the Scotch larch, both of which grow rapidly and are free from diseases. In the occasional swamps, so placed that they can not be used for cranberry culture, the swamp cedar, which affords with a rapid growth valuable timber, may be advantageously grown.

Perhaps the only land quite unfit for profitable use is that of the washed and blown sand of the beaches and spits. In the earlier conditions of our agriculture, lands such as those of Cape Cod were not worth attention. At the present time, with the increasing use of fertilizers and irrigation, these fields are likely soon to be made productive.

The facility with which water can be stored in the elevated lakes of Cape Cod invites the use of irrigation on much of its area. From a rough eye estimate (there are no maps good enough to warrant a closer study) I judge that not far from 10,000 acres of the peninsula could be effectively watered.

## HARBORS AND WATER WAYS.

In the conditions of navigation down to within sixty years of the present time the harbors of the cape were well suited to shipping. Owing, however, to the fact that these havens, with the exception of that at Provincetown, owe their basins either to flooded valleys, such as Oyster Bay, or to irregularities in the morainal fields, such as Woods Hole, they are all rather shallow and usually are shut off from the sea by bars and shoals. They are, therefore, fit only for the use of the smaller craft. Several of the ports which once sent forth many commercial ships, as, for instance, Chatham and Barnstable, do so no longer. The only port of value on the peninsula is that of Provincetown, which owes its existence to the formation of the curious beach hook which incloses its basin. The havens fit for use of pleasure boats are numerous; they are, indeed, numbered by the score. No part of the coast south of Maine so abounds in them as does the southern face of the cape. In general, these basins are susceptible of much improvement by the use of jetties, which may confine the considerable tidal water which passes through their excessively wide entrances.

More important in a general sense than the harbors of Cape Cod are the water ways which nearly traverse its width. These may be made passages by which vessels can avoid the dangerous voyage that now has to be made by all craft passing this part of the coast. A sailing vessel bound north or south of the cape has to reckon on an average of about two days' loss of time, as well as a considerable expense in the way of insurance, in making the voyage, at least during the winter half of the year. Except Cape Hatteras, there is no more dangerous portion of the Atlantic coast. The shipping which annually passes through Vineyard Sound on this voyage is said to be greater than that which traverses any like width of water in the world. From an early day there have been projects for cutting through the cape, making use of some one of the several channels—rivers so called—which nearly intersect the peninsula. Of these there are four which, with relatively slight expenditure as compared with other modern ship canals, could be opened to shipping. They are Monument River, Bass River, Town Cove in Orleans, and Pamet River in Truro.

The two last-named ways, though from an engineering point of view most practicable, are situated so far out on the cape that the worst dangers of the voyage northward would be passed before their entrances were reached; they are, therefore, not worth consideration.

The Monument River passage is, from the point of view of engineering, at least as far as opening the way is concerned, a very easy work to construct. It has, however, the peculiar disadvantage that it opens into Buzzards Bay, so that vessels must determine on passing that way from the time they start on their course around the cape. As in good weather the course can be run by a sailing vessel in twelve hours from the anchorage ground near Nobska light, in Vineyard Sound, what mariners need is a way to pass from the waters of that sound directly into Cape Cod Bay. The only passage which will afford this is that by the way of Bass River. This channel is, on the average, deeper and wider than Monument River, and it lies in a position where, at reasonable cost, vessels going northward could be provided with a safe harbor of refuge, whence, if the weather favored, they could turn the cape or could take the shorter artificial way. It is probable that the costs of these ways would not differ in any important measure. The distance from the western ports to Boston via Bass River would be longer by about 50 miles than by way of Monument River; to and from points north of Boston the additional distance would not be worth reckoning.

A considerable disadvantage of the Monument River way is that the upper part of Buzzards Bay, owing to its land-locked and currentless state, often becomes thickly packed with ice, even in winters of ordinary severity. On the other hand, the waters of Vineyard Sound, because they are the seat of strong through-running currents, are rarely thus embarrassed. It may be remarked that there is another improvement in the water ways of the cape which deserves consideration only less than a water way across the peninsula; this is, the passage through the morainal line to Woods Hole. At this point a natural breach of the moraine—one of the many which exist in the moraines of this district—affords a crooked and dangerous passage which has been much in use by vessels since the settlement of the country. A measure of benefit has been done to this way by dredging, but it remains an inadequate passage between two of the largest bays on our shore. If Monument River is to be taken as the site of the canal—and by a nearly common consent it appears to have been thus adopted—it will be more than ever necessary to provide a fit ship channel at Woods Hole, so that vessels may still have some choice as to the open route around the cape after they have entered Buzzards Bay.

Whatever is done in the way of canalizing the cape, it is clearly important that an adequate harbor of refuge should be provided on the south shore, where vessels in times of severe storm may find a safe anchorage. At present there is no such shelter fit for the use of the larger vessels which ply along the Atlantic coast between New Bedford and Provincetown. The anchorages in Vineyard Sound, with the exception of the small havens at Woods Hole and Edgartown and the breakwater at Hyannis, are all open and exposed to grave danger from the northeastern gales. The most available of these shelters—that at Vineyard Haven—is often very much crowded, so that if the outermost ships should drag their anchors a great catastrophe would be likely to occur, in which scores of vessels might be lost. A considerable number of the shipwrecks which occur where craft are on their way around Cape Cod are due to the fact that there is no perfectly safe place in the waters of Vineyard or Nantucket sounds where they can await conditions of weather which make it fit to essay the passage.

## ROAD-BUILDING MATERIALS.

The condition of the highways in Cape Cod is and always has been bad. The sandy nature of the underlay and the prevailing lack of vegetable matter in the soil make this condition inevitable unless some method of hardening the wheel way is adopted. Of these methods there are four which are more or less available at various points: The roads may be covered with oyster shells or the shells of the pecten, known locally as the scallop; they may be covered from time to time with a coating of clay; they may be graveled; and they may be macadamized.

The use of shells is in many ways to be commended where the traffic is light; but when exposed to much travel the covering is swiftly destroyed. Moreover, any general use of this material is impracticable on account of the limited sources of supply. The use of clay as a means of hardening the sands has been essayed in this district, but with poor results. The application on roads of ordinary use has to be made about once in two years, and it is so costly that in the end it is more expensive than it would be to construct a well-hardened way. Of gravel fit for road building, none is known to me except in the extreme western portion of the cape, and this is not of good quality. Thus the only satisfactory resource in this field is found in the use of broken stone, as in the well-known macadamized roads.

As there are no bed rocks attainable in the cape district which can supply material for macadamizing, it is necessary either to import the broken stone from the farther parts of Plymouth or Bristol counties or to make use of the erratics which may be had from the moraine or from the old walls composed of the smaller bowlders which have been gathered from the fields of till. As far out on the peninsula as the central part of Orleans, and within this section southward to the border of the moraine, the amount of this erratic material is great. It is, indeed, sufficient for the needs of road building in this county for all the foreseeable future. The supply of "field stone," or those which may be had from the surface of the ground, of sizes to be used in the crusher without breaking with the sledge, is limited. It is not likely to serve for more than the needs of original construction of the roads which will have to be built within the next score of years. After that the resort will have to be the pits opened in the moraine, where usually more than half the mass excavated will be large bowlders needing to be blasted in order to be made serviceable.

The petrographical character of the morainal erratics is good. They are mostly of a granite nature, with some admixture of trappean rocks. These dike materials sensibly increase in amount as we go eastward along the moraine. In this direction we find a considerable amount of volcanic débris, mostly fragments of what seem to be indurated ash beds and breccias. As before remarked, the rock masses of this morainal accumulation are not much affected by decay. The experiments made by the Massachusetts Highway Commission in the use of the field stone on the cape—of which about 12 miles of way has already been built, a portion of it having been in use for two years or more—shows that this material is excellently well adapted to building roads. It is so slightly decayed that the amount of small fragments produced is not much greater than is needed in "surfacing" the roads. When this element is excessive an adjustment can be made by sorting the stone before crushing, the product of the softer kind being used in the lower layer of the construction.

Roads made of the bowlders found on the cape can not be expected to have the endurance to traffic that is exhibited by those which have the covering layer composed of the harder traps, such as are found in the region about Boston or in the Connecticut Valley, yet for the uses they have to serve in this district they will prove very good.

The portions of the cape which are ill supplied with road-building stones are those in the southern parts, between the western part of Mashpee and the eastern part of Chatham and the towns of Eastham, Wellfleet, Truro, and Provincetown. In the first of these districts beneath the morainal apron there are, as are shown by various ditches, extensive deposits of pebbles and bowlderets which may afford local sources of supply of stone to be used in the crusher. These should be assayed in order to avoid the great cost of hauling material from the source of supply in the moraine, which is distant and accessible only by very sandy roads. These pebbly deposits seem to lie beneath the beds of the channels which extend from the moraine to the shore. They are usually covered by a thin layer of gravelly sand. The supply for the portion of the cape beyond Orleans will have to be brought by railway from the morainal district, and fortunately the railway extending to Provincetown brings all parts of this section within an average distance of about 1 mile from transportation.

## ORIGINAL EASTWARD EXTENSION OF CAPE COD.

The presence of an extensive system of drowned valleys on the Cape Cod district, including therein the islands on the south, leads to the question as to the extent to which these partly submerged lands were originally continued to the eastward. This question can not be fully answered, but some light may be thrown upon it by the facts and considerations which are noted below.

It is at a glance evident that the eastern side of the river valley into which drained the streams on the north side of the cape has been in part cut away. There is nothing to mark its former place except it be in part the shoal in which the Provincetown hook has been formed. This shoal is indistinctly continued northward, as is shown by the soundings, and the general form of the bottom of Cape Cod Bay supports the hypothesis that the valley was continued down to the depth of 100 feet or more below the present level of the sea.

On the floor of Massachusetts Bay and farther to the seaward in Georges Shoal, Cashes Ledge, and other less important elevations we have the elements of what appears to be an ancient land topography. It is possible to explain these features by the supposition that they are due to the warping of the earth or by the hypothesis that they were morainal in their nature, but neither of these views finds any definite support. Warpings of the type required to account for the facts would have to be of a type unknown in this part of the continent, at least in recent geologic periods. Moraines are contraindicated by the slight erosive power the ice evidently possessed and the thin character of the morainal accumulations on the neighboring coast.

It might seem probable that the mountain-building actions which led to the extensive dislocations of the Tertiary strata in this district were competent to bring about the formation of such ridges as we find on the sea floor in this vicinity, but the vast erosion and deposition which has taken place since these disturbances occurred would naturally have led to the destruction of any such reliefs had they been formed. Moreover, there are no indications that the stressing of these beds led to the building of sharp ridges, such as we find in these ledges and shoals. On the contrary, it is eminently probable that the region thus affected lacked at the end of the process any distinct topography except such as was given it by subaerial erosion.

It seems certain that the topography of the sea bottom in and to the east of Massachusetts Bay can not be very ancient. It evidently lies within the limits of the continental shelf—i. e., within the realm of excessive sedimentation, next the shore. Reliefs of such a sharp character would inevitably have been covered by detritus if they had long been in existence. Therefore, the probability seems to be that they are a part of the topography which, in a semisubmerged state, is preserved in the Cape Cod system of drowned valleys.

The cause of the formation of the partly or completely submerged valleys of this part of the shore may perhaps be found in the extensive dislocation which the strata have undergone in this region. If we suppose, as we well may, that the newer deposits of this part of the coast were over a large area much mountain built, the result would have been to lift what was originally a set of level and low-lying beds to a considerable height. A topography developed on such strata would be sharp, and the headwaters of the streams would be at a higher level above the sea than would be the case in the neighboring undisturbed districts. The fact that like disturbances in the region lying to the west and south have been attended by a similar preservation of the ancient topography, as in Long Island and Block Island, makes this view as to the cause of the maintenance of the Cape Cod peninsula the more probable.

In considering the conditions which have led to the formation and preservation of the Tertiary topography in the Cape Cod district, it is well to note the fact that the whole of this portion of the Atlantic coast appears to be at the present time much below the average elevation which it has recently had. This is shown by facts which indicate that the sea has not of late laid at a higher lever than about 100 feet above its present station, while the evidence from the submerged topography leads us to the conclusion that the depression below the level of most extreme elevations has amounted to at least 300 feet, and probably is much in excess of that amount. It should also be said that this submergence is not due to local causes. It is clearly a part of the very general action which has included a large portion of the shores of all the continents. The action is manifested on the eastern coast line of North America, from the mouth of the Rio Grande to the circumpolar section of the continent. It is also to be noted on the Pacific coast within the same parallels.

## ABSENCE OF SHOALS IN CAPE COD BAY AND IN BUZZARDS BAY.

The absence of shoals in Cape Cod Bay and in Buzzards Bay apparently indicates a difference in the history of these basins as compared with that of the depressions of Nantucket and Vineyard sounds. The explanation may possibly be found in the fact that the sounds, in part at least, represent a region of adjacent headwaters of several streams the cols of which were in the last great subsidence lowered beneath the sea, permitting the tidal currents freely to pass through them. These streams having a great erosive action on soft rocks, such as underlie this district, are sufficient to account for the effacement of the islands which evidently lay not long ago in these waters. It may also be remarked that the Buzzards Bay River appears to have had a steeper drainage than the other neighboring old streams on the east, which may have accounted for the more complete erosion of the divides between its branches. It seems, however, more likely that the presence of these shoals in the Nantucket and Vineyard system of sounds is partly to be attributed to the action of tidal currents in this field.

## SEAWARD CONTINUATION OF DROWNED VALLEYS.

The soundings given on the coast charts, where the water exceeds about 100 feet in depth, are not in sufficient detail to make it worth while to devote much labor to tracing the probable direction of the ancient drainage channels with a view to ascertaining how near they went to the margin of the continental shelf. We may, however, note certain of the more patent facts.

North of Cape Cod we find a deep channel between Race Point and Stellwagen Bank. The water at the western end of this channel, where it appears to connect with the Cape Cod Bay Valley, has a depth of about 35 fathoms; thence it shoals seaward to about 22 fathoms; still farther to the east it deepens rapidly to about 50 fathoms. The shallowest water in this channel is in the continuation of Cape Cod. Stellwagen Bank has a minimum of 12 fathoms of water upon it, and not much more for its length of about 20 miles; then at the deep channel leading toward Boston Harbor the bottom suddenly declines to the depth of 60 fathoms. The evident suggestion is that Stellwagen Bank is a northward continuation of the Cape Cod divide and that Race Point channel marks the position of a col on the ridge, which was some 70 feet lower than the general surface of the water shed; and also that the Cape Cod Bay River joined what we may call the Boston River near the northern end of the above-named bank.

North of the valley of Boston River another less distinct, unnamed shoal continues the line of Stellwagen Bank in such manner as to suggest that a stream corresponding in a way to that of Cape Cod Bay headed about Cape Ann and flowed southward, joining the Boston River near where that passing from the Cape Cod divides entered it, the united streams flowing on through the Stellwagen ridge to the sea. The general likeness of the outlines of these antithetic valleys, if we may use that name to designate basins of very like character whose streams flow against each other, suggests that they are carved in like materials, or, in other words, that the Cretaceous and Tertiary strata of the Cape Cod district are continued as far north at least as Cape Ann, though they do not appear above the surface. This proposition is made the more probable by the discovery by Mr. Warren Upham of fossils of possible Eocene or Cretaceous age in the drift materials near Highland light. As such remains have not been found elsewhere in the drift of the cape, they have probably been brought from beneath the level of the sea.

Evidences derived from soundings and dredgings in the Bay of Maine indicate, as is well known, the existence of Tertiary and perhaps Cretaceous rocks, at least about the Grand Banks and Georges

Bank.[1] These and other observations indicate that the shoals to the east and north of Cape Cod are probably the remains of ancient divides, and the soundings warrant, in a measure, the interpretation of ancient river valleys, but this task will not here be further essayed.

## ORIGIN OF THE CAPE DISTRICT PLATEAU.

An inspection of the contour maps will show that the pre-Glacial beds of the cape district, including the islands on the south, have their upper surfaces everywhere at about the same altitude, with a prevailing slope from the western part of Marthas Vineyard, where they rise to about 300 feet above the sea, toward the east and north, declining at Boston Harbor to the sea level, near the end of Cape Cod to a little above that level, and at Nantucket to a height of about 50 feet. The question arises as to the origin of this approximately plane surface. It may be due to either of two actions—to base-leveling or to the leveling action of the sea—or possibly to a complex of these actions. It clearly is inadmissible to suppose that the plateaulike surface is due to the survival of the original stratification surface, for, as has often been noted, the area has been greatly disturbed. Against the supposition that the approximation to horizontality just before the uplift which set at work the streams that cut the valleys of the old rivers of this district was due to base leveling, we may note that this would require us to suppose a very long period in which these much-dislocated rocks had been slowly brought to a level by atmospheric agents. As we see at present on Marthas Vineyard these rocks yield but little to such action. The streams of to-day carry away scarcely any mud; their effect is limited to a slight leaching action. To introduce such a base-leveling period of sufficient duration would call for greater lengthening of the first Pliocene time than it is reasonable to make.

The leveling of this district by the action of the sea might have been accomplished in a relatively short time. The present rate of retreat of the southern shore of Marthas Vineyard is, as before noted, about 3 feet per year; at this rate the sea would occupy 2,000 years in wearing a mile into the land. The width of this table-land, including the submerged portion, being assumed at 50 miles, the leveling process would have required about 100,000 years. Great as this time is, it is probably much less than would have been required to effect the same result by the base-leveling process alone. Here, as elsewhere along coast lines, it is likely that these two actions cooperated, the streams carrying away what they were enabled to and the waves removing the portion of the material which was not thus taken to the sea. It is not likely, however, that the time occupied in the work could have been much less than that above suggested. Here, again, we

[1] A. E. Verrill, Occurrence of fossiliferous Tertiary rocks on the Grand Bank and Georges Bank: Am. Jour. Sci., third series, Vol. XVI, p. 323.

encounter the perplexing difficulty that the history of the beds of the successive epochs in this area requires us to suppose a lapse of time since the close of the Tertiary period much greater than is commonly assumed to have occurred.

## POSITION AND CHARACTER OF DIVIDES.

The position of the several divides which mark the limits of the ancient partly drowned valleys of the Cape Cod district is such as would be expected in case the topography of the region had been developed when the surface of the country was at least 200 feet higher than it is at present. The most continuous of these crests is that of the cape itself, extending from the town of Plymouth eastward to Chatham and thence northward to the sand spit which terminates the cape. This divide may have been continued somewhat farther to the east, as will be noted hereafter. The arrangement of the valleys of the headwater streams in this section suggests their former union in two or more valleys, which declined to the north and south. North of the elbow of the cape to the Provincetown sand spit the incutting of the sea appears to have destroyed the original crest of the divide, leaving only the slope of the ridge which drained into the Cape Cod Bay river.

The divide which separates the water of the last-named stream from the upper tributaries of the Buzzards Bay river is still traceable in the long ridge which stretches in an interrupted manner from Monument River to the neighborhood of Plymouth Harbor, terminating in Manomet Hill. It is, indeed, impossible to account for the very peculiar shape of the ground in this district without supposing that it is due to the interlacing of the headwaters of adjacent but oppositely flowing streams. This western divide of the Buzzards Bay valley is continued southward in the united ridge on which the Falmouth moraine lies as far south as Woods Hole. From the harbor of Woods Hole the same divide is shown in a less united form by the line of the Elizabeth Islands to and including the island of Cuttyhunk.

On the south of Vineyard Sound we have in Marthas Vineyard the remains of the other crest of the valley of which the Elizabeth Islands form the other divide. This crest constitutes the northern range of hills of Marthas Vineyard, extending as far east as Vineyard Haven. The central and southern parts of the highlands of the island are on the headwaters of streams which seem originally to have flowed into the Muskeget River valley.

The island of Nantucket appears to be the remnant of several obscure divides, but the greater part of what is left above water is on the slope of the drainage toward the Muskeget and the Monomoy rivers. As will readily be seen, the directions of the ancient rivers in this portion of the district are by no means clear. This obscurity is mainly due to the extensive erosion by tidal currents which has taken place in this part of the field.

It is to be noted that the decline in the altitude of the principal divide of this district, that which extends from Plymouth Harbor to the extremity of the Elizabeth Islands—a distance of about 45 miles—is relatively steep, being from a height of about 300 feet to that of about 100 feet above the sea, or an average of about 5 feet per mile. In view of the considerable width of the valleys in this region, this decline must be regarded as greater than is consistent with the supposition that the region is anywhere near to being completely baseleveled. There is no definite evidence as to the rate of fall of these drowned valleys, but in valleys of such width, cut in materials of so yielding a nature, it is difficult to believe that it could have exceeded 5 feet to the mile. In considering this question it should be noted that the decline of the crest line is not necessarily a true measure of the fall of the valley. In general, however, this decline is, at least in the upper part of the river's course, much less rapid than the fall of the stream. Taking the remnants of the valleys as we find them in Marthas Vineyard, we note that they support the proposition that the bottoms of the old valleys had a slope less than is indicated by their present altitudes. The valleys of the Tisbury and Tiastquan rivers below the points where they are occupied by permanent streams have a fall of about 15 feet in a mile, yet these are the upper and presumably steepest portions of the river systems to which they belong. Although there are no very certain conclusions to be drawn from this inquiry into the slopes of the old rivers of the Cape Cod district, the fact suggests that there may have been some warping movements since the topography was formed.

It is to be observed that the three best defined of the old valleys of this area, those of Cape Cod Bay, Buzzards Bay, and Vineyard Sound, show a certain measure of narrowing toward their lower parts. This feature is most evident in the case of Buzzards Bay, but it is noticeable also in the other basins. This apparently indicates stream erosion working toward the formation of circus-shaped valleys, features which are not uncommonly found in much eroded areas.

In connection with the old valleys of this area the island of No-man's-land offers matter for interesting inquiry. This bit of land, by its position and its relation to the form of the sea bottom, suggests that it is the remnant of the southern divide of a valley the stream of which drained into Vineyard Sound river near Gay Head. The shoal about this island, though it is evidently subjected to much erosion by the strong current and waves, indicates that the isle, which is rapidly wearing away, was originally of much greater extent than at present. The retreat of its shores, which appears to be going on at the rate of about 3 feet per annum, will bring about its destruction in less than a thousand years. Its place will then be for a time occupied by a shoal which, under the cutting action of the waves and tidal currents, will be planed down to a considerable depth, coming finally to the state of the shoals in Nantucket Sound.

## TRESPASSING OF RIVERS.

The position of the several divides between the ancient basins of the Cape Cod district indicates that the streams had advanced far in the development of their topography before the last great subsidence. Here and there we find evidence that the crests had been brought to rather sharp edges, having in most cases lost all trace of the original table-land character which seems to have been in some way impressed on them before the last invasion of the valleys took place. The Cape Cod crest was evidently sharp; so, too, was that of the Elizabeth Isles. That of Marthas Vineyard was of a more complicated nature, retaining much of the original table-land form.

The process of stream capture, of which there are good instances on Marthas Vineyard, which evidently took place before the last great downsinking that brought about the formation of the bays and sounds, shows that the adjustment of the topography about the heads of the streams on that island had not been anywhere near completely effected. Similar though less evident indications of such action on the mainland may be found in the valleys of Monument, Bass, and Pamet rivers, where streams had evidently in good part or altogether cut back through the divides, more or less invading the drainage of the antithetic stream. In two of the last-mentioned instances, Monument and Bass rivers, the transgression seems to have been made by streams flowing southward toward the valleys of the Nantucket Sound area. In that of Pamet River the slope seems to have been to the westward into Cape Cod Bay. The passages between the Elizabeth Isles were apparently much as we now find them before the close of the time preceding the last deep submergence. If this be the case, the head streams of either the Buzzards Bay or the Vineyard Sound river may have crossed the divides, and as these passages have been much changed by glacial action and by tidal currents it is not easy to determine in which direction the trespassing waters flowed.

As the valleys now occupied by the bays were evidently rather deep, probably at least 500 feet at their deepest part, below the higher parts of the divides, the incomplete nature of the topography on and near the crests is no good reason for supposing that their bottoms were much indented. It is a very common feature of river valleys to have their lower parts level and their upper parts deeply indented. We thus are not compelled to suppose that a great deal of filling has been necessary in order to bring about the general approximation to horizontality exhibited in the bottoms of the bays.

## AMOUNT OF SEDIMENTATION SINCE THE PRESENT LEVEL WAS ESTABLISHED.

As noted under the last heading, the fact that the bottoms of the Cape Cod system of bays and sounds are approximately level can not be taken as evidence of any great amount of sedimentation since the sea attained to about its present position at the end of the last important down-sinking. A close study of the form of these bottoms, based on the soundings of the United States Coast Survey charts, shows a multitude of slight irregularities which can not well be attributed to the differential deposition of sediments due to tidal currents, but can best be explained by supposing that the layer of imposed detritus is not yet thick enough to completely mask the preexisting ridges and valleys of these submerged areas.

If there had been a great amount of deposition on the sea floor in the Cape Cod bays we should expect to find all the original irregularities of surface due to their erosion quite effaced, in the manner in which it appears to have been destroyed on the great southern plain or on a lesser scale on the morainal plain of Marthas Vineyard; but while there are traces of such depositional shelves near the shore, as on the southern coast of the last-named isle and along the south and east shores of Cape Cod, these shelves are narrow and flat. In form they evidently are quite unlike the bottoms of the water areas at a distance of 2 or 3 miles from the coast line. The evidence from soundings goes to show that the migrations of sand in these areas are locally considerable, but they appear to occur only in the paths of relatively strong tidal currents, such as sweep through the sounds and in the bays; it is to the effect that, notwithstanding all the coastal erosion which is going on, the contribution of sands to the bottoms of the bays at the distance of a mile or more from the shores is very slight in amount. These facts lead me to doubt whether as much as an average of 50 feet in depth of detritus has been accumulated on the floors of the bays since they last came below the level of the sea.

On the south side of Marthas Vineyard and of Nantucket and on the east side of Cape Cod the considerable invasion of the land by the sea has doubtless done much to contribute material for sedimentation, but in Cape Cod Bay and Buzzards Bay the erosion of the shores has not been sufficient to supply more than a few feet of detritus over the floors of the basins. The waste of organic life deposited in these basins is relatively small in amount, being much less than in those parts of the shores and shallows to the northward, where mollusks and seaweeds are more abundant.

Owing to the situation of the Cape Cod salient it is not in a position to receive detrital materials from a distance in the manner in which they are accumulated along the shore south of New York. North of the cape the deep trough passing outward from Boston Harbor intercepts the current and wave driven waste coming from the northern shore. On the southwest there is no set of currents driving such material toward the waters of the sounds and bays. Some contribution may have been had from the wreckage of islands now reduced to shoals, but it probably has not been in large amount.

## DEPTH BENEATH SEA LEVEL AND NATURE OF THE CRYSTALLINE ROCKS IN THE CAPE COD DISTRICT.

At no point in the district of Cape Cod are the ancient crystalline rocks exposed to view, nor does the drift covering on any part of the land indicate by its character that these deposits are on the land areas, at least near the surface. In passing from the mainland toward the cape we find the nearest localities of the crystalline rocks at Plymouth Harbor and on the west shore of Buzzards Bay. Throughout the southeastern section of Massachusetts these rocks exhibit a gentle and tolerably uniform slope toward the base of the cape at the rate of about 20 feet to the mile. This would place the old granite series at the level of 100 or 200 feet below the sea level on the eastern shore of Buzzards Bay and at a depth below that plane of about 1,000 feet at Chatham Harbor. Inasmuch, however, as all this region has been greatly disturbed, and as the disturbances most probably included extensive movements of the ancient rocks as well as of the Mesozoic and Cenozoic strata which rest upon them, no great value can be given to these estimates.

On Marthas Vineyard and scantily on the shores of the southwestern portion of Cape Cod, particularly along the southeastern side of Vineyard Sound, as before remarked in the account of the glacial drift, a great quantity of chalcedonic quartz pebbles are found. So abundant are these coarse agates that hundreds of tons of the material could at times be gathered along the shore. It is evident that these fragments have been glacially transported, and, as is indicated by the character of the layer of erratics as well as by the glacial scratches, the movement of the ice was from the northwest, if indeed it was not from a point nearer the west. Nowhere on the mainland are rocks of this nature known either in situ or in the drift. Along with these chalcedonic erratics go great quantities of pebbles of white-vein quartz of an aspect quite different from any known on the shores to the northward. A close comparison of the pebbly materials on the islands of the Cape Cod district with the rocks on the mainland will undoubtedly show other cases of this kind. On the body of Cape Cod, in increasing proportion as we go from Monument River eastward, we find groups of vein and volcanic materials differing in nature from those on the neighboring islands. In this last-mentioned district the pebbles of volcanic rocks are very abundant, especially in and beyond Orleans.

The evidence afforded by the erratic materials of this district shows that just to the seaward of the shore line there is both to the north and the south of Cape Cod a belt of rocks which have been the seat of great volcanic and solfataric action, and that these deposits have been much metamorphosed. In this connection it may be noted that a belt of disturbance of the nature indicated on the sea bottom about Cape Cod begins in the region about the Bay of Fundy and extends parallel to and partly within the shore along the coast of Maine. I have noted the occurrence of such conditions in the published accounts of work done for the Survey in the districts about Passamaquoddy, Cobscook, and Orange bays, and in the district of Mount Desert and on Cape Ann. In the two first-named fields the presence of distinctly volcanic deposits is well proved. At Cape Ann evidence of true volcanic action is lacking, but the extraordinary amount of dike injections, which evidently increases as the shore is approached, shows the effect of the same system of disturbances. It thus appears probable that the coast line of this continent, from the head of the Bay of Fundy at least as far to the southward as the mouth of Buzzards Bay, lies upon the inner margin of a tract which has been greatly subjected to volcanic and solfataric action. It is not improbable that the indentation of the first-named bay may be due to the subsidences connected with these disturbances and that the position of the coast line on this part of the shoreland is to be accounted for either by the downward movement or by the relative ease with which the coastal fringe of rather incoherent deposits were eroded by the sea. Interesting as these questions are, they can not be followed further here except to suggest that the volcanic areas of the Boston Basin, of the Connecticut Valley area, and of the region in and near the lower part of the Hudson Valley, where there is reason to suspect that the volcanic rocks are newer than the Newark beds, are within the limits of this marginal fringe of such deposits. no' hue

## FOSSILS DREDGED FROM SEA FLOOR NEAR CAPE COD.

From time to time there have been reports that fossils of a character which would indicate that they came from beds such as are found on Marthas Vineyard or at Marshfield existed on the sea floor to the north and east of the cape. These specimens have been brought up on anchors or by the dredge. It seems likely that in all these instances the materials have been derived from the drift deposits on the floor of the sea, though it may be possible that, as about Georges Bank, the considerable energy of the tidal currents may scour away the soft parts of the bottom, leaving the harder fragments as a coating on the sea floor.

The evidence above referred to, though fragmentary, and at best not very trustworthy, is enough to show that to a considerable depth, say to 200 feet or more, the bottom is, in part at least, occupied by the fossiliferous deposits which were here and there exposed in the cape district just before the Glacial period. This serves to show that the

Shaler seems undecided as to age of Truro, Barnstable, Weyquosque series – These are accepted by all recent geologists as Pleistocene – stadial and interstadial. (DWC)



erosion which took away so large a share of the later pre-Glacial accumulations—those designated as the Truro and Weyquosque series or the Barnstable clays—affected a large extent of the sea bottom for a considerable distance out to sea. Without attaching too much importance to this obscure evidence, it may well be taken as of some value in showing that the region was for a long time elevated to the height of 200 feet or more above its present level, or, what comes effectively to the same thing, that the sea was at about that depth below its present plane. Thus the considerations going to show a recent submergence, which are derived from the topography of the coast and the drowned valleys, has some support from the evidence which these chance samples of the bottom afford.

It has been suggested that these fossils brought up by dredging may be from the drift deposits which are presumed to exist for some distance to the east beyond the shore. Against this hypothesis may be set the fact that fossils of the Cretaceous and Tertiary beds have rarely been found preserved in any glacial deposits, even where those beds lay immediately upon the strata richest in organic remains. The probability that they would thus occur in quantities sufficient to account for the numerous chance finds is so small that it may be disregarded.

## TIME RATIOS INDICATED BY POST-TERTIARY PRE-GLACIAL EROSION.

It is not possible, in the present state of our knowledge, to undertake any final essay in determining the time occupied in the erosive work done in this region since the close of the Pliocene epoch. It is, however, possible to give some general and relative indication of these durations.

The facts show that after the deposition of the Pliocene deposits of Marthas Vineyard a vast erosion occurred, which shaped the strong topography that is exhibited in the western part of that island. In the present condition of that area, though the valleys are deep and much of the surface is but slightly drift covered, the rate of stream erosion is almost nil. Even in times of heavy rain the brooks show hardly a trace of color due to other than the stain of decayed vegetation. In the period before the deposition of the drift the rate of wear was probably more rapid than it is at present, but it is impossible to estimate the value of this difference. We are therefore left to mere impressions as to the time required for the development of this topography. These inferences, however, are of some importance.

It is in the first place to be noted that the rocks of this region are, and have been from the time of their formation, very open-textured. They readily absorb the rain water, which cuts no channels on these areas, which are so nearly driftless that the Cretaceous and Tertiary deposits are essentially at the surface. This must have made this field, with any rainfall, which was not very much greater than that of to-day, one of slow erosion, for the reason that the waters entering the ground would have been discharged, as they now are, at the level of the permanent streams and in a very gradual manner. This action may now be seen in fields of great extent, as in the town of Chilmark, or near Gay Head. Nor can we suppose that the water penetrating to the depths exercises any considerable solvent action. The strata which it traverses contain very little soluble matter, and the springs—save that they sometimes exhibit the results of decomposing pyrites in the sulphureted hydrogen they give off and contain a considerable amount of alumina sulphate—are essentially pure. The facts above noted lead to the conclusion that the erosion of the Vineyard area has from the beginning been slow.

On the basis of a slow erosion we have to account for the formation of river valleys a mile in width and having a depth of from 150 to 200 feet, as well as for the renewal of an unknown section which has been worn away from the crests of the hills. Assuming that the average ablation of the area has been at the rate of 1 foot in one thousand years, a rate which must be accounted rapid—it is equaled, so far as ascertained, in no part of the world which bears a covering of natural vegetation—in that it would at this rate of cutting require somewhere near 200,000 years to carve out these valleys; but, as is easily seen, the valleys are only a part of the result of the erosion which the streams have applied to them. The elevated country between these troughs has also gone down, so that it does not seem unreasonable to assume that the total erosion of this valley-making period has required 300,000 years.

Beyond the clear evidence of a long erosion interval afforded by the valleys of Marthas Vineyard we perceive that there is an unmeasured and perhaps immeasurable time which intervened between the period when the rocks in which the depressions exist were dislocated and that when valley-making began. It seems tolerably evident that in this period the sea stood some hundred feet higher than it does at present and that the surface was gradually base-leveled until it came nearly to the plane indicated by the highest land of the island. This little-indicated period of erosion, the sole evidence of which is found in the faint yet distinct marks of an ancient plain, antedating the formation of the present drainage system, possibly represents a duration several times as great as that shown by the action of the streams which now are at work.

In a general way following the development of the valleys of Marthas Vineyard and those which were formed in the tilted deposits of Cape Cod, came a period of deposition in which the various beds of Nashaquitsa, Barnstable, and Truro series were laid down. The history of this stage or stages in the development of the cape district is not yet unraveled. These several sections may all be of the same age or they may represent, to a greater or less extent, the history of successive periods. However this may be, there can be no doubt that the time occupied in the deposition of these beds was very great, and that the detritus which was accumulated came in part, if not altogether, from areas north and west of the cape district, as is shown by the fact that the beds in question to a great extent mantle over the ancient topography and rise to near the level of its highest elevations.

Following the accumulations of the Truro-Barnstable groups came the third great period of aerial erosion in this field. During a period of elevation which brought the land to a level at least 200 feet higher than it is at present, the erosive work of the streams cleared away the deposits of stratified sands, clays, and gravels from the valleys which it encumbered, and extended the denudation of those broad and thick sheets of strata until only remnants of the original mass remained. The amount of erosion effected during this period immediately antedating the last ice epoch can not be gauged, for the reason that the greater part of the area in which it was effected is now submerged beneath the sea; but it was clearly much greater than that which was done in the time to which we owe the development of the several valleys of Marthas Vineyard or the like troughs of Cape Cod. To it we owe not only the general clearing out of those troughs, but the excavation of the river basins which are now marked by the bays and sounds of southeastern Massachusetts.

In considering the time required for the formation of the later stratified deposits of the Cape Cod district we have first to note that the accumulation of these beds indicates a long period of erosion, the record of which, as before remarked, is not found in this area for the reason that it was then beneath the sea. As the work done was of sufficient magnitude to form a broad sheet of detritus, extending from some point far inland, probably the central highlands of Massachusetts, to and beyond Truro and Nantucket, having a thickness on the average of not less than 100 feet, and perhaps several times as great, it is evident that the time occupied could not well have been less than that attained for the second erosive period—that which shaped the greater valleys of Marthas Vineyard. The last great erosion period—that which morcellated the stratified deposits which overlapped the old mountain-built beds—appears to have required more time than any of the earlier periods of wearing. The valleys were brought to a great width. Nearly all the deposits of the last formations were removed, leaving only fragments of them on or near the divides. As this work was done mainly on very permeable beds, into which the rain penetrates rapidly without developing small, superficial streams, the work of wearing could not have proceeded with great rapidity.

The facts as above presented lead to the conclusion that since the close of the Tertiary period, or perhaps from some time after the end of the Pliocene epoch and down to the advent of the ice in the last Glacial epoch, there have been four tolerably distinguishable periods of erosion in this field, each requiring a time the duration of which, even in a geological sense, must be accounted as long. Various estimates, made on the basis of the present rates of erosion, lead me to the conclusion that this interval was not less than one million years.

Without attaching any definite value to the reckonings as to the durations of the periods in which the wearing down on this district was effected, it may be claimed that no geologist who has attentively considered the problems of time ratios in erosion is likely, on a careful study of this field, very much to reduce this estimate. Although the assumption of something like a million years for the interval between the end of the Tertiary and the beginning of the last ice epoch is not in accordance with the views as to the time ratios in the later stages of geological history which were entertained down to the beginning of this decade, it is becoming evident that the old view as to the brevity of this interval was hastily taken and will have to be revised. It is in order to bring this point into debate that the previous estimates as to the time occupied by the completed succession of actions which have taken place in post-Tertiary pre-Glacial time are here submitted.

It may be remarked that the prejudice in favor of a brief time since the close of the Tertiary period has rested in part on the fact that the amount of consolidation which has taken place in the deposits of that portion of the earth's history has been in most cases small. It is, indeed, difficult to believe that beds which have been as little changed as the strata of this age usually are have been formed for millions of years. When, however, we note that in Mesozoic rocks, and even in those of the Paleozoic sections, the amount of alteration is often slight, we can well understand that these more recent deposits, which have not undergone deep burial, have survived for ages without essential change.

## SUMMARY AND CONCLUSIONS.

The results of the observations noted in the preceding pages of this report justify the following statements relating to the Cape Cod district, and particularly concerning the peninsula of that name.

After the erosion of the Cretaceous and Tertiary beds disclosed on Marthas Vineyard, several series of sedimentary deposits were laid down. The first of these deposits, which may fairly be reckoned as of early Pleistocene age, though deposited in horizontal attitudes on the disturbed older beds, were in turn somewhat stressed, the resulting dislocations being relatively much less considerable than those which affected the Cretaceous and Tertiary deposits. These dislocations are evident not only on Marthas Vineyard but throughout the area of Cape Cod between Woods Hole, Bourne, and Highland Light. They also extend up the coast at least as far as Plymouth Harbor; their western extension is not yet determined.

The post-Tertiary deposits just referred to appear to be divisible into two groups, the lower of which is exhibited on Marthas Vineyard and in Cape Cod as far north as Monument River and eastward to Orleans. This series is characterized by the presence of red clays and sands, which appear to owe their origin to the decomposition of Tertiary strata such as occur at Gay Head. Apparently above these beds, which have been termed the Nashaquitsa series, occurs another series, here known as the Truro. The characteristic of these beds is that they plentifully contain a fine, white, floury, micaceous sand, which is very much decayed. These sands are combined with coarser arenaceous materials and occasionally clay beds, the clay being of the ordinary grayish or yellowish hue. So far as is ascertained, these beds contain no pebbles of compound rocks, the pebbles, indeed, being very rare and, so far as observed, of quartz, and always small. The Truro series has not disclosed any materials apparently rising from decomposition of the Tertiary beds. It is possible, but not probable, that the Nashaquitsa and Truro series should be regarded as one group. The distribution of the beds, however, is against this view.

The group of brick clays known as the Barnstable series appears to have been laid down after the Nashaquitsa and before the Truro beds had been formed and dislocated. The evidence as to this succession, however, is not perfectly clear.

During the deposition and erosion of the series above noted the Cape Cod district has been subjected to a number of movements of elevation and subsidence of which the imperfectly interpretable changes are shown in fig. 87, p. 522. These movements indicate very remarkable instability in the position of this portion of the continent from Jurassic time to the present day, but the alterations of level appear to have been limited, so far as determinable, to a range not exceeding, perhaps, 1,000 feet. It is to be noted that these accidents appear to increase in frequency as we approach the present day. This, however, is most likely due to the fact that the records are more complete and interpretable as they come toward the present time; possibly, also, as my colleague, Mr. J. B. Woodworth, has noted to me, for the reason that the later records are more coastal in their nature than are those afforded by the earlier deposits. It is to be remarked as a very significant feature that the series of deposits from and including the Nashaquitsa to the close of the Barnstable series have afforded no fossils. It is possible, but not probable, that fossils may have been contained in these beds, the remains having disappeared under the very free leaching which has occurred throughout this area, where the rocks are extraordinarily porous.

It is possible, and perhaps probable, that these beds, in part at least, represent deposits in advance of glacial sheets. Nevertheless, as we have to suppose that in part the materials were laid down in salt water, it is not easy to understand the complete absence of organic remains, which, as we know from other fields, even those near by, as on Nantucket and on the coast of Maine, contain abundant fossils. Moreover, it is difficult to believe that during the extended topographic changes which occurred while the three series above mentioned were depositing, and also during the period of dislocation, which is marked in the attitude of the beds, no ice could have remained near enough to the district to affect the character of the sediments. Still further, the absence of erratic rocks in the Nashaquitsa and Truro contraindicates the action of ice.

The condition of the deposits contained in the series of the Cape Cod district, formed after the close of the Tertiary and before the advent of the glacial sheet, indicates the rapid erosion of an area of crystalline rocks which had previously been affected by a deep decay. It is conceivable that this erosion, acting on softened materials, was due to rivers, but the general absence of vegetable matter in the deposits makes it perhaps more likely that the work was accomplished by glacial erosion occurring during the periods of subsidence which are indicated by the sections.

The facts stated in the preceding pages of this report make it clear that the post-Tertiary and pre-Glacial history of southeastern Massachusetts is much more complicated than has hitherto been supposed. The interpretation of the record which has been given must be regarded as in a great measure tentative. A further development of our understanding of the facts will doubtless be attained when the related deposits on Long Island, New York, have been explored. There is reason to hope that in that field may be found the passage from the conditions of southeastern New England to those accepted in the related Columbian beds and other recent deposits in New Jersey and the portions of the coast to the southward.

# RECENT EARTH MOVEMENT IN THE GREAT LAKES REGION.

BY

GROVE KARL GILBERT.

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# ILLUSTRATIONS.

# RECENT EARTH MOVEMENT IN THE GREAT LAKES REGION.

By G. K. Gilbert.

## INTRODUCTION.

The geologic history of the earth shows that in all parts of its surface there have been great oscillations of level. Modern history also records upward and downward movements of the land at various points. The modern movements are of small amount, but it is believed that they are of the same kind as the ancient, and that the great changes of the geologic past were effected slowly. Nearly all discoveries of modern change have been made at the seashore, but there is no reason to suppose that the land is now more stable in the interior of the continents than along their coastal borders. Observations are restricted to the coast because the sea level affords the best available datum plane for comparison. The present paper discusses the stability of the region of the Laurentian lakes, and uses the surfaces of the lakes as datum levels or planes of reference.

## OBSERVATIONS BY MR. STUNTZ.

In 1869 there was presented to the American Association for the Advancement of Science a paper by G. R. Stuntz, a land surveyor of Wisconsin, describing certain observations on Lake Superior made by him in 1852 and 1853. He states that in those years a certain mill race at the falls of St. Marys River was entirely dry. As St. Marys River is the outlet of Lake Superior, its volume and the supply of water for the mill race depend on the height of water in the lake, and he therefore inferred that at that end of the lake the water was low. He also states that a small stream at Pindle's mill, entering Lake Superior not far from the outlet, runs with swift current to the lake, and has no widening, marshes, or other indication that its valley overflows by the lake setting back into it. He then describes the strongly contrasted condition of streams entering the lake near its western end:[1]

As you go westward, the Ontonagon River exhibits a slight filling up. The valley near the mouth shows that at the time it was excavated the surface of the lake was lower than at present. The same is also apparent at the mouth of Bad River, still

[1] On some recent geological changes in northeastern Wisconsin: Proc. Am. Ass. Adv. Sci., Vol. XVIII, 1870, pp. 206-207.

farther west. At the mouth of Bois Brulé the same thing is exhibited, only to a greater extent. From this to the west end of the lake not only does the lake set back into the valleys of the streams, but the waters are making rapid encroachments on the banks. So rapidly is the filling back, that the deposits of the streams do not keep pace with the filling up. The consequence is, that there is a large marsh and pond in the mouth of the valley of Bois Brulé and Aminecan River. But nowhere is this filling up more apparent than in the bay above the mouth of the St. Louis River. In several parts submerged stumps, several feet below the present water level, are found. The numerous inlets surrounding the main bay, when we consider the nature of the soil and the formation (a tough, red clay), in all of which the water is deep, could not have been excavated in the natural course of events with the water at its present level. The testimony of the Indians also goes to strengthen the same conclusions. At the time of running the State line above mentioned, the Indians, ever jealous of their rights, called me to a council to

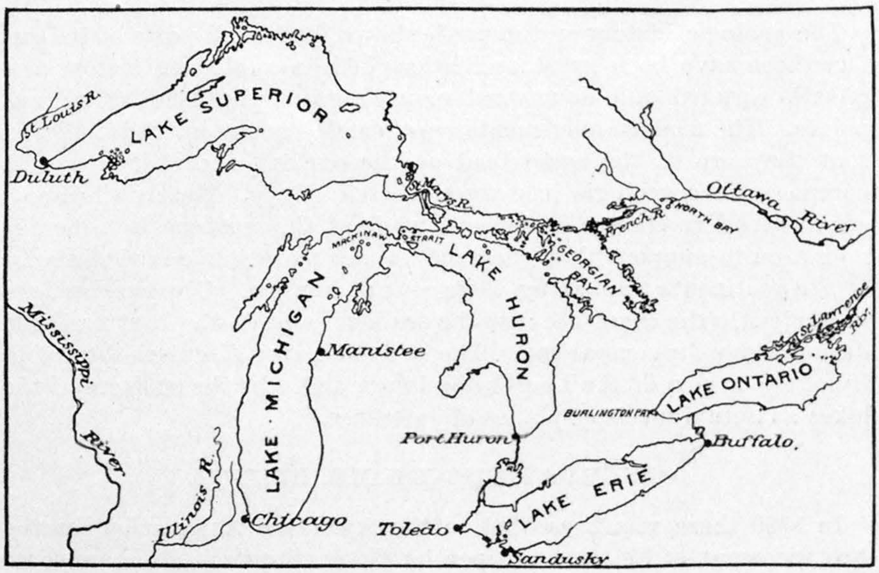



FIG. 93.—Map of Laurentian lakes.

inquire why I ran the line through Indian land. In the explanation, I gave, using the language of the law, as a starting point, the lowest rapid in the St. Louis River. The chief immediately replied that formerly there was a rapid nearly opposite the Indian village. Start, said he, from that place, and you will be near the treaty line. After he had been further questioned, I learned that it was only a few years since the river was quite rapid at the Indian village. At the time the said line was run the first rapid was about 1 mile by the stream above the village. From these facts I conclude that a change is taking place gradually in the level of this great valley.

From these data Stuntz infers "the gradual rise of water at this [west] end of the lake and the falling of the same at the east," and it is evident from the context that he refers this change of the water to a westward canting of the basin, the western part becoming lower as compared with the eastern.

So far as I am aware, this paper broaches for the first time the idea of differential elevation in the Great Lakes region, and it contains the only observations that have ever been cited as showing recent changes of that character. In later years the subject has been approached from the geologic side, and Dr. J. W. Spencer has expressed his opinion that a warping or tilting of the whole region is now in progress.

## EARTH MOVEMENTS DURING THE CLOSING EPOCHS OF THE PLEISTOCENE PERIOD.

The Great Lakes came into existence in the latest of the geologic periods, the Pleistocene. Their number and position underwent numerous and important changes during the latter part of the period, and their area and drainage systems have been greatly modified even within the time to which human history belongs. In late Pleistocene time, while the great Laurentide ice field, which just before had covered the entire lake basin, was slowly growing less through the melting away of its edges, there were a series of lakes along its southern margin. These were held in at the north by ice and on other sides by uplands, and they found outlet southward over the lowest passes of the divide between the Laurentian basin and the basins of the Mississippi, Susquehanna, and Hudson. With changes in the position of the ice barrier, individual lakes were from time to time divided or drained and separate lakes united, so that the lacustrine geography had a complex history. After the ice had wholly disappeared from the region, the drainage did not at once assume its present system, for Lake Huron, instead of overflowing to Lake Erie, discharged its surplus water over the pass at North Bay, Canada, and thence down the Mattawa and Ottawa rivers to the St. Lawrence.

In the decipherment of this history much use is made of the shore lines of the vanished lakes. These consist of sand and gravel terraces that were once deltas, of cliffs and strands carved from hillsides by the waves, and of spits and beach ridges thrown up by the same agency. A number of these lines have been traced for great distances, and wherever thus traced it is found that they are no longer level, but are gently inclined. When formed they were of course horizontal, for they were made by waves generated on a water surface, and the fact that they are not now level shows that the land on which they are marked has undergone changes of relative height. The general direction of inclination of the shore lines is toward the south southwest, showing that the basin of the lakes has been canted in that direction. The amount of change has not been everywhere the same, and it is probable that the direction of the canting varies somewhat from place to place. Where several shore lines are traced on the same slope the first made are usually more steeply inclined than the last made, and hence it is inferred that the general change of relative altitude was in progress through the whole epoch of the glacial lakes. The plane of the Iroquois shore line in the basin of Lake Ontario descends toward the south-southwest at an average rate of 3½ feet a mile, the slope being steeper at the north than at the south.[1] The Oswego shore line, in the same basin, slopes in the same direction at the rate of more than 3 feet a mile. The Warren shore line, traced from Lima, New York, about the sides of the Ontario, Erie, and Huron basins to Pompeii, Michigan, is nearly level in the Maumee basin, but rises northeastward with a rate gradually increasing to 2 feet a mile. Its northward rise in Michigan is 1½ feet a mile.[2] The present southward inclination of the water plane of Lake Algonquin, which occupied the Superior, Michigan, and Huron basins, ranges from a few inches to 3 feet a mile.[3] Great Lake Nipissing, which occupied the same basins after the disappearance of the ice and had its outlet at North Bay, conformed more nearly to the present slopes, the general inclination of its water plane being about 7 inches to the mile.[4]

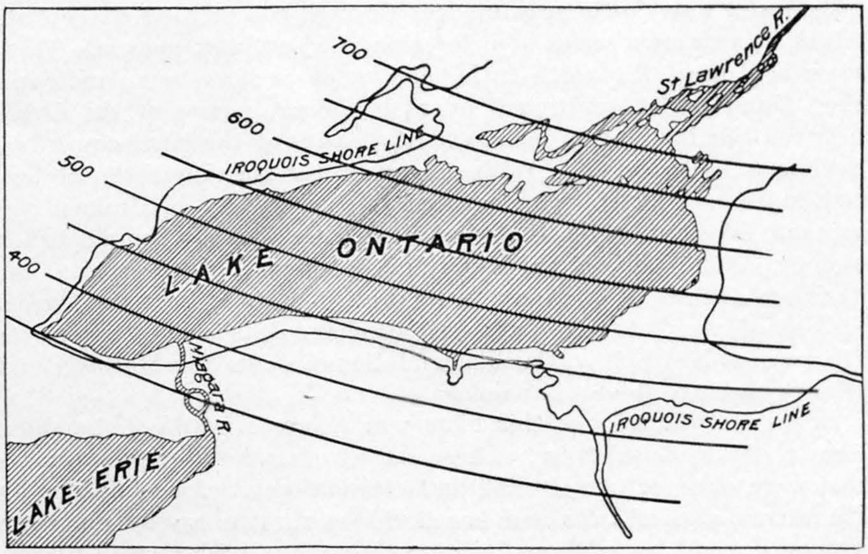



FIG. 94.—Map of the Iroquois shore line. Modern water bodies are shaded. A line shows the boundary of the ancient lake. The parallel curves are isobases.

On the accompanying maps of Lake Iroquois and Great Lake Nipissing (figs. 94 and 95) the character of the tilting is shown by means of isobases, or lines drawn at right angles to the direction of tilting. All points on one of these lines have been uplifted the same amount since the time of the corresponding lake. If we think of the plane of the water surface of one of the old lakes as having been deformed by uplift or warping, then the isobases are contours, or lines of equal present height, on the deformed plane.

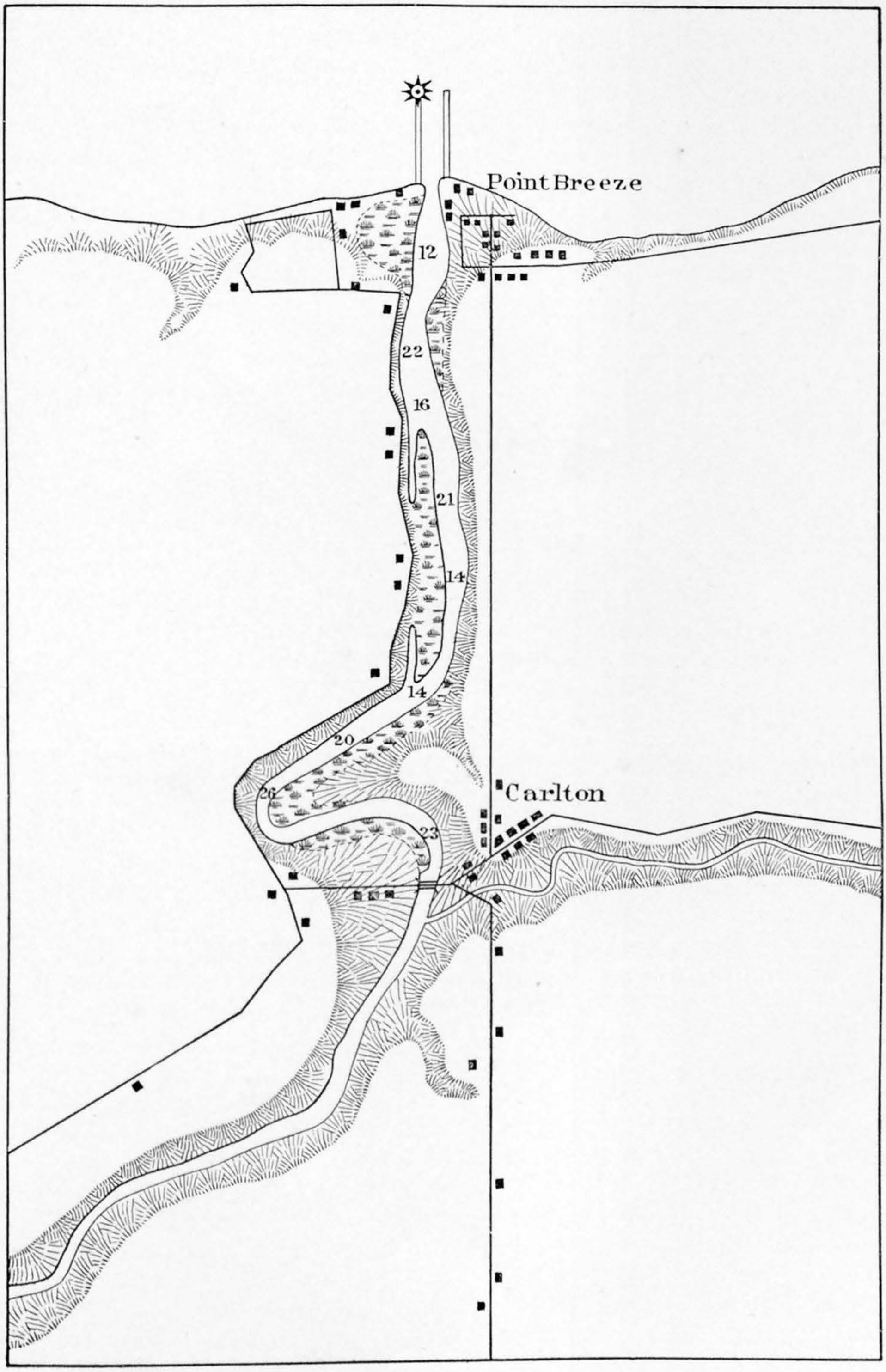
Other evidence of the tilting of the land is found in the character of

[1] J. W. Spencer, Trans. Roy. Soc. Canada, Section IV, 1889, pp. 121–134; G. K. Gilbert, Sixth Ann. Rept. Commissioners of the State Reservation at Niagara, Albany, 1890.

[2] F. B. Taylor, Bull. Geol. Soc. America, Vol. VIII, 1897, p. 55.

[3] F. B. Taylor, A Short History of the Great Lakes, Terre Haute, 1897.

[4] Ibid.




ESTUARY AT THE MOUTH OF OAK ORCHARD CREEK, ORLEANS COUNTY, NEW YORK.

Scale, 1 inch = 1,500 feet. Figures give soundings in feet.

The water ways are sharply incised in the plain. Partial refilling is shown by marshes. Slack water at lake level reaches to Carlton, above which the creeks are shallow.

stream channels as they approach lake shores. The streams reaching Lake Superior from the southwest have already been described in the quotation from Stuntz, and similar characters are found in the basins of Lake Erie and Lake Ontario. Considerable tracts of land along the southern shores and about the western ends of these lakes are smooth plains, their surfaces having been leveled by deposits of fine sediment from the Pleistocene lakes just mentioned. The creeks and rivers traversing the plains have readily cut the soft deposits, carving out narrow valleys. In the upper parts of these valleys the streams are shallow and descend with lively current, but on approaching

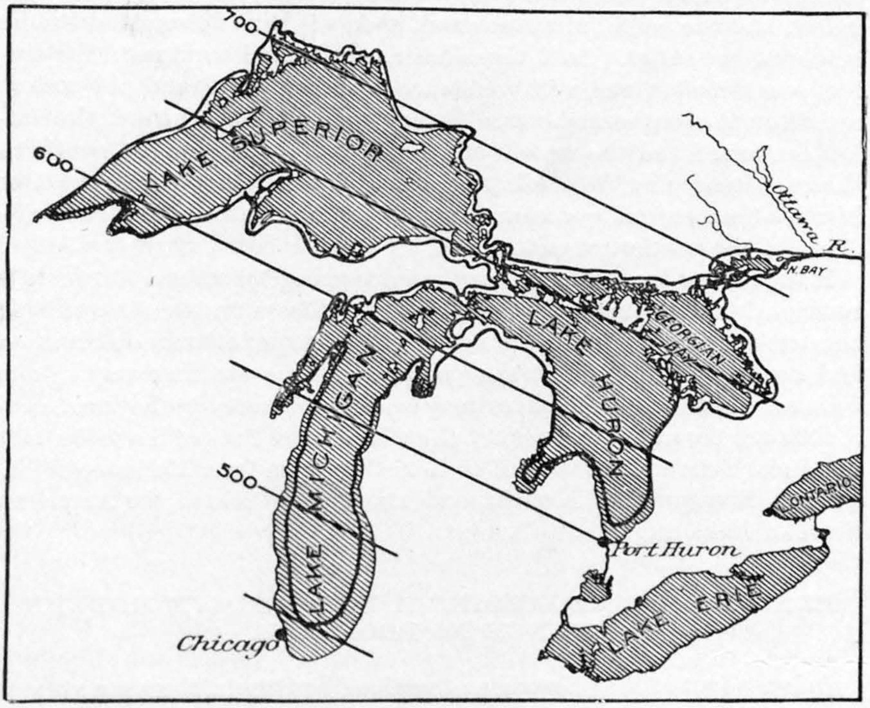



FIG. 95.—Map of the shore line of Great Lake Nipissing. Modern water bodies are shaded. A line shows the boundary of the ancient lake. The parallel lines are isobases.

the lake they become deep and sluggish, the change usually occurring several miles from the lake shore. Stated in another way, each stream, instead of debouching into the lake, enters the head of a long, narrow bay or estuary. The origin of such estuaries is well understood. They are found on all sinking coasts, and their meaning in this region is that the land has gone down or the lake level has risen, so that the waters of the lake occupy portions of the channels carved by the streams in the lowland plain. This description applies to the greater number of streams entering Lake Ontario between the Genesee and Don rivers and to those entering Lake Erie between Cuyahoga River and Maumee Bay. Individual mention may be made of Oak Orchard, Eighteenmile, and Twelvemile creeks in New York, of Twelvemile and Twentymile creeks and the Credit and Humber rivers in Ontario, and of Rocky River, Black River, Vermilion River, Old Womans Creek, Pike Creek, Turtle Creek, and Ottawa River in Ohio. Even the largest rivers of the district, including the Genesee, Niagara, Cuyahoga, and Maumee, have features indicative of the same history.[1]

By reference to the map (fig. 93, p. 602) it will be seen that the outlets of these lakes are at their northernmost points, and this fact is related to the conditions of the stream channels. The water level of a lake is maintained by the balance between inflow and outflow. It is just high enough to enable the outflowing stream to carry off the excess from inflow, and the height of water on all shores is thus determined by the height of the outlet. So if these basins are canted northward the outlets are thus lowered with reference to other parts, and the waters recede on the southern shores. If they are canted southward, the outlets are raised and the waters are made to advance on the southern shores. Reasoning from effect to cause, the fact that the lake water invades the new-made stream channels on the southern shores is evidence of the southward canting.

It should not be assumed that the "drowning" of stream channels is restricted to the tracts mentioned above. Those tracts are specified because they fall within the range of the writer's personal observation and are known to exhibit the phenomena in a striking way. It is believed that similar features may be found wherever the local conditions are favorable throughout the whole coast lines of Lake Ontario and Lake Erie, about the head of Lake Michigan from Manistee, Michigan, to Kewaunee, Wisconsin, and about the whole of the American shore of Lake Superior.

## REASONS FOR REGARDING A PROGRESSIVE MODERN CHANGE AS PROBABLE.

Independent of the phenomena described by Stuntz, there are various considerations tending to direct attention to the question of the stability or instability of the Laurentian area at the present time. The first to be mentioned is purely geologic. The epoch during which the overflow from the upper lakes followed the valleys of the Mattawa and Ottawa is definitely associated with a certain stage of the Niagara River. The cataract of Niagara is at the present time increasing the length of the Niagara gorge at a somewhat rapid rate. The formation from which the water leaps is a firm limestone 60 feet thick, and beneath this are shales which are comparatively soft and weak. The cataract, by eroding the shale, undermines the limestone, which falls away in blocks, and these blocks are in turn utilized by the water as an instrument with which to grind the shale. Whirled about by the

[1] F. B. Taylor mentions a few other localities on the same lakes: Am. Geologist, Vol. XV, 1895, pp. 174-176.

water, the blocks not only wear away the face of the shale cliff, but drill down deeply, so that beneath the cataract there is a pool nearly or quite 200 feet deep. Working in this way, the cataract has extended the gorge several hundred feet since the first accurate measurements were made, the average annual rate being between 4 and 5 feet.

With the present arrangement of the drainage system the Niagara carries the surplus water from the basins of lakes Huron, Erie, Michigan, and Superior; but when the upper lakes sent their overflow to the St. Lawrence by way of the Ottawa, the Niagara carried only the discharge from the Erie basin. Its volume was then only one-eighth of the present volume and its power was correspondingly less. It could not move the great blocks of limestone which fell from the cliff, and, instead of scooping out a deep pool, as now, it excavated a comparatively shallow channel, whose bottom was cumbered with limestone débris. Owing to this difference in method of erosion it is possible to discriminate the parts of the gorge excavated when the river was small and when it was large, and thus to determine the place of the cataract when the outlet of Lake Huron was shifted from North Bay and the Ottawa River to Port Huron and the St. Clair and Detroit rivers. That place is at the head of the Whirlpool Rapids, 11,600 feet from the present cataract. Assuming that the cataract worked at its present rate through this distance, we may compute the time consumed. At $4\frac{1}{2}$ feet a year, it would be about two thousand six hundred years. F. B. Taylor, making allowance for various qualifying factors, estimates the time to have been not less than five thousand years.[1]

When Lake Huron changed its outlet, the plane of its water surface extended from the pass at North Bay to the pass at Port Huron, but the North Bay pass now stands 140 feet higher than the Port Huron. This difference of altitude, amounting to 6 inches a mile, has, therefore, been wrought within the period of about five thousand years. In view of the gradual nature of such movements, this is not a long period to assign to the measured change, and it is natural to inquire whether the movement is not still in progress.

Dr. J. W. Spencer, who has devoted much time to the study of the Niagara gorge and the glacial lakes, is confident that change of level has not yet ceased and that it will eventually turn the water of the upper lakes southward to the Illinois and Mississippi rivers, leaving the Niagara channel dry. Addressing the American Association for the Advancement of Science in 1894, he said:[2]

> The end of the falls seems destined, if we read the future by the past, to be effected, not by the erosion expending itself on the rocks, but by terrestrial deformation turning the drainage of all the upper lakes into the Mississippi, by way of Chicago, just as the Huron waters were lately turned from the Ottawa into the Niagara drainage; and at the recent rate it would seem that about 5,000 or 6,000 years at most will be needed. The change of drainage should arrive before the cataract shall have receded to Buffalo.

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[1] A short history of the Great Lakes.

[2] Proc. Am. Ass. Adv. Sci., Vol. XLIII, 1894, p. 246.

Another consideration of the same tendency is found in the condition of the estuaries described in the last section. Most of the streams flowing into these rise in districts of unconsolidated drift and carry forward in flood time a considerable load of detritus. This is deposited in the estuaries, the coarser part making deltas at their heads, and the finer settling as mud in the deeper water. The process tends to convert the estuaries, first, to marshes, and then to dry land, but in most instances little progress in that direction has been made. There are a few creeks rising in sandy districts which have succeeded in filling their estuaries, changing them to marshes; but as a rule the delta at the head of the estuary invades it but a short distance, and the marshes which border it here and there at points sheltered from the flood currents are impassable except by boats, and have the appearance of submerged flood plains. These characters, from their close resemblance to the features observable along the subsiding parts of the Atlantic coast, give the impression that a slow flooding of the stream valleys is still in progress.

A third consideration is connected with the record of recent changes on the coasts of the continent. It has long been known that the Atlantic coast south of Connecticut is subsiding, and Prof. G. H. Cook was able to determine the rate in New Jersey as about 2 feet a century.[1] Dr. Robert Bell has recently collated a variety of facts tending to show that the land has risen in the region about Hudson and James bays,[2] and he estimates the rate at from 5 to 7 feet a century. If these two movements are parts of a general movement affecting the northeastern part of the continent, then the Great Lakes region, approximately intermediate in position between the rising and sinking areas, should be found to exhibit a southward tilting.

These various facts, all tending in one direction, are sufficient warrant for the working hypothesis that the tilting of the lake region demonstrated by the slopes of the old shore lines is still in progress; and the writer, who has for many years been interested in the problems of the Great Lakes, has made repeated efforts to secure an investigation by which the hypothesis might be tested.

The mode of investigation first suggested was the establishment of elaborate observation stations at three points—Port Huron, Chicago, and Mackinac. By a suitable series of observations at these points, the relative heights of benches might be established with high precision, the water surface being used as a leveling instrument. Then, after an interval of one or two decades the observations might be repeated and any changes in the heights of benches due to differential uplift detected. The matter was submitted in 1890 to the Superintendent of the United States Lake Survey and to the Superintendent of the United States Coast and Geodetic Survey, but, though it was received favorably by the latter officer, the work was not undertaken.

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[1] Am. Jour. Sci., 2d series, Vol. XXIV, 1857.

[2] Am. Jour. Sci., 4th series, Vol. I, 1896.

Other plans were then considered, and it was finally decided to make a study of existing records of lake level, and, if necessary, supplement them by additional observations. The results of this investigation are set forth in the following pages.

## GENERAL PLAN OF INVESTIGATION.

Variations in the height of the ocean level at any place depend chiefly on tides, winds, and atmospheric pressure. By means of long series of observations the effect of these disturbing factors can be eliminated

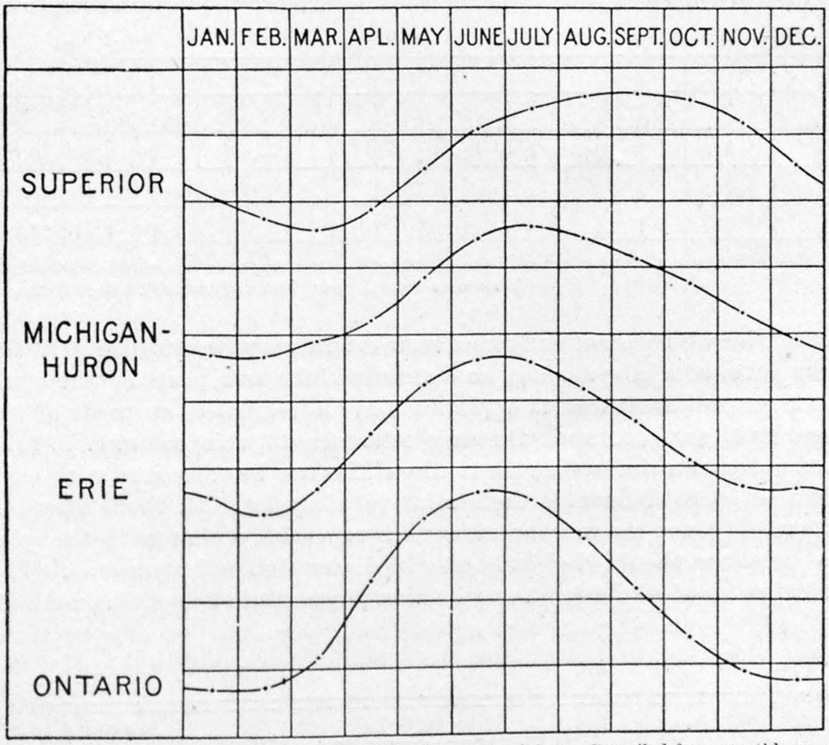



FIG. 96.—Annual oscillations of the surfaces of the Laurentian lakes. Compiled from monthly means published by the Chief of Engineers, U. S. A. Each vertical space represents six inches. The observations for Lake Superior cover the period 1862-1895; for Michigan-Huron, 1860-1895; for Erie, 1855-1895; for Ontario, 1860-1895.

and a mean level obtained which is practically uniform from year to year and decade to decade. The height of the water surface must depend also on the quantity of water in the ocean, but the actual variations of volume are so small as compared to the extent of the ocean surface that the resulting variations of level may be neglected and the mean level used as a standard for the discussion of differential movements of the earth's crust. With the Great Lakes the case is materially different. There are variations due to wind, atmospheric pressure, and tides, but when these have been eliminated by long series of observations the resulting mean level is far from constant, varying from season to season and year to year with the volume of water. In each lake there is an annual change of more than a foot, depending on the seasonal inequality between gain by precipitation and loss by evaporation (fig. 96), and there is a still greater change resulting from the cumulative effect of series of dry and series of moist years. The records show that the water surface in each lake has been several feet higher in some years than in others. (See fig. 97.)

For this reason the water surface of a lake does not afford a datum plane by reference to which the elevation or subsidence of coasts can be directly determined. Fortunately, however, there is an indirect

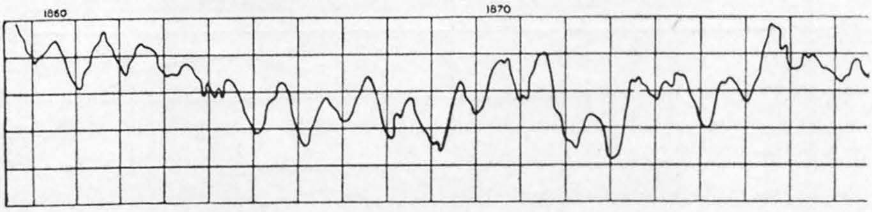



FIG. 97.—Oscillations of the surface of Lake Michigan due to changes in the volume of the lake. Wisconsin, from August, 1859, to June, 1897. Each horizontal

method by which practically the same result may be attained. If the mean level of a lake surface be determined for two parts of the coast at the same time, these two planes may be regarded as parts of the same level surface, and, through reference to this common datum, fixed objects on the land at the two localities can be compared with each other so as to determine their relative altitudes. If, then, after an interval of time, the measurements are repeated, a change in the relative height of the fixed objects may be discovered and measured. The investigation described in the following pages made use of this method.

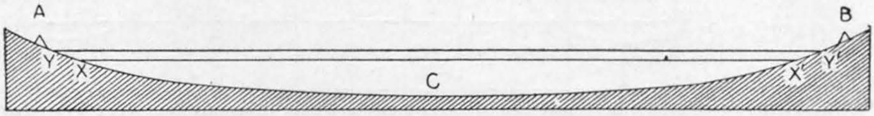
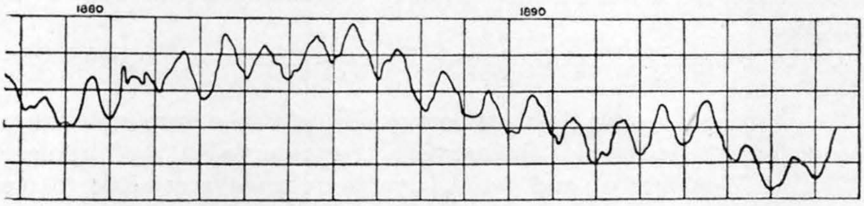



FIG. 98.—Diagram illustrating method of measuring earth movements.

The fundamental principle of the method is illustrated by the diagram, fig. 98, in which A C B is the cross profile of a lake basin. At a certain time the mean plane of the water surface occupies the position XX′. By means of the engineer's level it is ascertained that a bench mark A has a certain height above the water plane at X, and that a bench mark B has a certain height above X′. The difference between these two measurements is the difference in altitude between A and B. After an interval of years the measurements are repeated. The water plane then stands at some different level, say YY′. The height of A above Y is measured, and the height of B above Y′; the difference between the two measurements gives the relative height of A and B. If earth movements have occurred during the interval between the two sets of measurements, the second determination of the comparative height of A and B will differ from the first determination, and the amount of difference will measure the differential earth movement.

## AVAILABLE DATA.

*Gage readings.*—In order to eliminate the temporary effects of disturbing factors, it is necessary to have a series of observations of the height of the water surface at each of the localities compared. The gages by




Compiled under the direction of the Chief of Engineers, U. S. A., from gage readings at Milwaukee, space represents a calendar year; each vertical space, one foot.

means of which such observations are made are of various kinds. One of the simplest is a graduated plank, fixed vertically by attaching it to a dock or other structure, so that one end is above water and the other below. Sometimes the plank is omitted and the graduation marked upon the side of a dock or pier. The height of the water surface is ascertained by direct comparison with the lines of the graduation. Another form of gage which has been extensively used in the lakes consists of a graduated rod, not fixed, but held in the hand; with this the distance from the water surface to a fixed point is measured. Usually the fixed point chosen is above the water surface, but at one station, Port Colborne, it is the submerged sill of a canal lock. Another form of gage includes a float to which a graduated vertical rod is fixed, and the graduations of the rod are compared with a fixed point on the land; or a chain attached to the float may pass over a pulley and carry a counterpoise, in which case an index, fastened to some part of the chain or counterpoise, moves up and down past a stationary graduated scale. There are also automatic gages making periodic or continuous records.

Previous to the year 1859 records of lake level are meager, and not of such nature as to be suited to the purposes of this investigation. A general account of them is given by Col. Charles Whittlesey, in Volume XII of the Smithsonian Contributions to Knowledge, and a fuller account in the Report of the United States Deep Waterways Commission for 1896. In 1859 the investigation of lake levels was undertaken by the United States Lake Survey. Several stations were established on each lake, and at these regular observations were made, usually three times a day. From time to time stations were discontinued and others were established, and after the close of the field work of the Lake Survey the observations were, in many cases, continued by officers of the Engineer Corps in charge of harbor improvements. With reference to the present investigation, I have examined United States Lake Survey and other United States engineer records for the following stations:

On Lake Superior: Superior City, Duluth, Ontonagon, Marquette, and Sault Ste. Marie.

On Lake Michigan-Huron: Chicago, Milwaukee, Grand Haven, Lockport, Sand Beach, Port Austin, Pointe Aux Barques, Tawas, Escanaba, Thunder Bay.

On Lake Erie: Monroe, Rockwood, Cleveland, Ashtabula, Erie, Buffalo.

On Lake Ontario: Port Dalhousie, Niagara, Charlotte, Oswego, Sacketts Harbor.

These records are for the most part published in the form of monthly means, but the individual observations are preserved in the Engineer Office at Washington, and these have been made accessible to me through the courtesy of Gen. William P. Craighill, Chief of Engineers. By the Canadian Department of Railways and Canals, I have been enabled to make use of a long series of observations at Port Colborne, on Lake Erie, the head of the Welland Canal; observations at Toronto, on Lake Ontario, have been furnished me by the city engineer, and observations at Collingwood, on Lake Huron, by Mr. Frank Moberly.

*Benches.*—As gages at the water side are subject to various accidents, it is rarely possible to maintain their zeros for long periods at a constant level, and unless they are connected by leveling with bench marks of a permanent character their records have little value for purposes of comparison. Previous to 1871 such connection with benches was not made by the United States Lake Survey, or, if made, the records are lost. There were, however, certain stations, notably Chicago, Milwaukee, Cleveland, Port Colborne, Buffalo, Charlotte, and Oswego, at which this matter had received attention. The structures at Chicago on which the early bench marks were made are thought to have afterwards settled.[1] At Milwaukee the early bench marks no longer exist, and although there is reason to believe that other benches were substituted with care, my researches have not discovered a satisfactory record. The same remark applies to Buffalo; and the record of the original bench at Charlotte has been lost. At Port Colborne and Oswego the zeros of gages are permanent structures, which have probably suffered no change; and at Cleveland, although the oldest benches no longer exist, it is believed that the record of transfer is complete and satisfactory.

In 1870 Gen. C. B. Comstock was placed in charge of the United States Lake Survey, and the scientific methods introduced by him

[1] Report on Chicago City Datum and City Bench Marks, by W. H. Hedges, Chicago, 1895.

included the establishment of a complete system of benches in connection with the gages. From 1872 until the completion of the field work of the Lake Survey there was an annual inspection of the gages, and the relations of their zeros to the bench marks were redetermined as often as seemed necessary. From 1871 to 1878 the supervision of gages and the reduction of records were in charge of Mr. O. B. Wheeler, and from 1879 to 1882, of Mr. A. R. Flint. The results of the present investigation are largely indebted to the care and thoroughness with which these engineers performed their work.

## SELECTION OF STATIONS AND YEARS.

Under the general method outlined above the first step was the selection of suitable pairs of stations on the shores of the various lakes. As

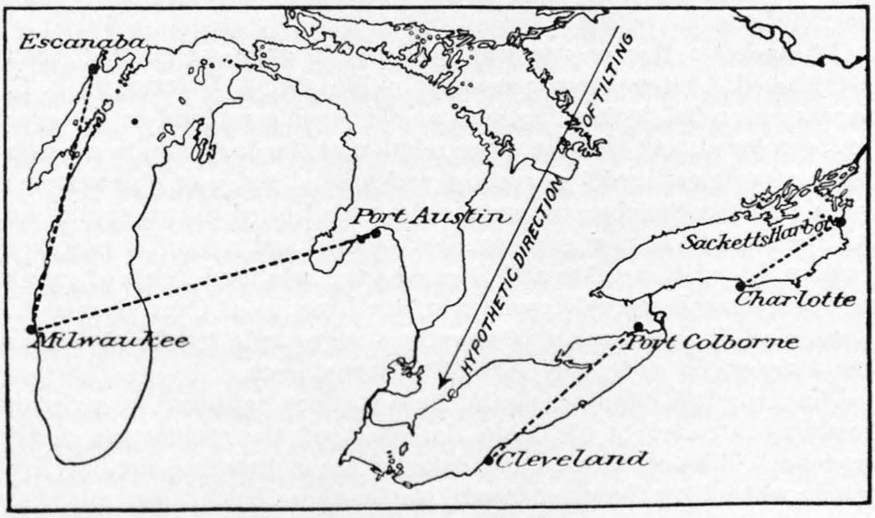



FIG. 99.—Arrangement of selected stations.

the geologic data indicated a tilting of the land toward the south-southwest, or, more precisely, in the direction S. 27° W., it was desirable to have each pair of stations separated by a long distance in that direction. As the hypothetic change was exceedingly slow, it was desirable to compare observations separated by the longest practicable time intervals. It was essential that the gage readings before and after the time intervals be accurately connected with the same benches. Consideration was also given to the fact that the results might be vitiated if use were made of observations taken during the prevalence of storms, when the water is sometimes driven by the wind so as to stand abnormally high on certain shores; and in order that the use of such observations might be avoided it was important to select years during which the force of the wind was daily recorded. With these considerations in view the available data were examined, and the following selection was made of stations (see fig. 99) and years:

For Lake Ontario, Charlotte and Sacketts Harbor, 1874 and 1896.

For Lake Erie, Cleveland and Port Colborne, 1858 and 1895.

For Lake Michigan-Huron, Milwaukee and Port Austin, 1876 and 1896, and Milwaukee and Escanaba, 1876 and 1896.

No comparison was undertaken for stations on Lake Superior.

## SPECIAL OBSERVATIONS IN 1896.

Certain of the selected stations are not now maintained by the United States engineers, and in order to complete the data it was necessary to make special observations. This was done in the summer of 1896, during the months of July, August, September, and October. The necessary attention was also given to bench marks, and provision was made for observations of a special character at the regular engineer stations.

At Sacketts Harbor use was made of a gage which had been established for temporary purposes by Maj. W. S. Stanton, U. S. E. It was connected, by leveling, with an old bench mark, and an observer was employed. At Charlotte the relation of the gage zero to a bench mark was determined by leveling, and special series of observations were made by the observer employed by the United States engineers. At Port Austin a new gage was established and a special observer employed. At Milwaukee and Escanaba the relative heights of gages and bench marks were determined under direction of Capt. George A. Zinn, U. S. E., and special observations were made by the observers regularly employed by the United States engineers.

The "special" observations at these stations consisted of series of readings intended to eliminate the effect of the oscillations called "seiches." The equilibrium of a lake surface is disturbed not only by winds, which blow the water toward the lee shore, but by inequalities of atmospheric pressure occurring during thunderstorms and during the passage of cyclones; and the impulses thus received are not quickly dissipated, but cause a long-continued swaying of the water. In large lakes these oscillations are so enduring as to cover the interval from one disturbing impulse to another, and keep the water perpetually in motion. Near the ends of the lakes and in bays with gradually converging sides the range of oscillation may be as great as 1 foot, and it ordinarily amounts at all lake stations to from 1 inch to 4 inches. For this reason a single observation may not approximate closely to the mean level of the water, and the actual mean level can be determined only by a series of observations at short intervals. In arranging the work of 1896 the observers were instructed to record the water level every five minutes for an hour each morning and evening of all days when the wind was light; and at Sacketts Harbor, where the seiche has an exceptionally long period, the length of the series was afterwards increased.

## DISCUSSION OF DATA FROM PAIRS OF STATIONS.

### SACKETTS HARBOR AND CHARLOTTE.

In 1874 the zeros of gages at these stations were points marked on docks, and readings were made by means of graduated vertical rods attached to floats. They give the distance of the water surface below the gage zeros. At the time of each observation record was also made of the direction and force of the wind. The work was under the direction of the United States Lake Survey. Mr. A. Wilder was the observer at Charlotte, and Mr. Henry Metcalf at Sacketts Harbor.

The gage at Charlotte was put in place in November, 1871, and the measurements showed its zero to be 32.7 feet below a bench mark. In January, 1873, its zero was found to be 32.959 feet below the same bench mark. On May 11, 1874, it was again compared with the bench mark, and the difference was found to have increased to 33.003 feet. It is probable that this change of .044 foot was occasioned by the settling of the dock to which the gage was attached. A manuscript report dated February 3, 1875, says: "The bank is here partly of timbers and partly of earth. The earth has been washed out and has fallen away from the timber in some places." The gage at Sacketts Harbor was also found unstable. The report of an inspection in May, 1874, states that the zero of gage "has been lowered 0.555 foot;" and a report dated February, 3, 1875, says: "This gage is fastened to the timbers of an old and unused dock. The whole structure is quite dilapidated and unstable." The instability of gages determined the selection of time for the comparison of stations. Both gages having been compared with benches in May, 1874, that at Charlotte on the 11th and that at Sacketts Harbor probably on the 14th, the computations were based on a period including these dates. Within this period selection was made of those times of observation when the wind force at both stations was less than 3 on a scale of 10. Thus treated, the observations of 54 days gave 51 comparisons.

*Computation of the height of the gage zero at Sacketts Harbor, New York, above the gage zero at Charlotte, New York, in the spring of 1874.*

| Day. | Hour. | Gage reading. | | Difference. |
|---|---|---|---|---|
| | | Sacketts Harbor. | Charlotte. | |
| 1874. | | *Feet.* | *Feet.* | *Feet.* |
| Apr. 17 | 9 p.m. | 5.43 | 3.15 | 2.28 |
| 19 | 9 p.m. | 5.33 | 3.01 | 2.32 |
| 22 | 9 p.m. | 5.23 | 2.98 | 2.25 |
| 23 | 7 a.m. | 5.16 | 2.80 | 2.36 |
| | 2 p.m. | 5.18 | 2.78 | 2.40 |
| 24 | 7 a.m. | 5.20 | 2.80 | 2.40 |
| 25 | 7 a.m. | 5.18 | 2.80 | 2.38 |
| | 2 p.m. | 5.11 | 2.81 | 2.30 |
| 27 | 7 a.m. | 5.08 | 2.78 | 2.30 |
| | 9 p.m. | 4.88 | 2.78 | 2.10 |
| 28 | 2 p.m. | 5.13 | 2.87 | 2.26 |
| | 9 p.m. | 5.06 | 2.88 | 2.18 |
| May 3 | 7 a.m. | 5.06 | 2.95 | 2.11 |
| 4 | 7 a.m. | 5.43 | 2.95 | 2.48 |
| | 2 p.m. | 5.09 | 2.94 | 2.15 |
| | 9 p.m. | 5.16 | 2.94 | 2.22 |
| 5 | 9 p.m. | 5.18 | 2.91 | 2.27 |
| 6 | 9 p.m. | 5.12 | 2.85 | 2.27 |
| 7 | 7 a.m. | 5.06 | 2.87 | 2.19 |
| 8 | 9 p.m. | 5.17 | 2.85 | 2.32 |
| 11 | 9 p.m. | 5.17 | 2.82 | 2.35 |
| 14 | 9 p.m. | 5.12 | 2.83 | 2.29 |
| 15 | 7 a.m. | 5.17 | 2.84 | 2.33 |
| | 2 p.m. | 5.25 | 2.82 | 2.43 |
| 18 | 2 p.m. | 5.13 | 2.82 | 2.31 |
| | 9 p.m. | 5.18 | 2.83 | 2.35 |
| 20 | 7 a.m. | 5.05 | 2.83 | 2.22 |
| | 2 p.m. | 5.10 | 2.82 | 2.20 |
| 21 | 7 a.m. | 5.02 | 2.81 | 2.21 |
| 22 | 9 p.m. | 5.02 | 2.82 | 2.20 |
| 24 | 7 a.m. | 5.12 | 2.95 | 2.17 |
| | 2 p.m. | 5.08 | 2.92 | 2.16 |
| | 9 p.m. | 5.08 | 2.91 | 2.17 |
| 26 | 9 p.m. | 5.08 | 2.86 | 2.22 |
| 27 | 7 a.m. | 5.00 | 2.86 | 2.14 |
| | 9 p.m. | 5.10 | 2 84 | 2.26 |
| 28 | 7 a.m. | 4.98 | 2.83 | 2.15 |
| | 9 p.m. | 5.02 | 2.83 | 2.19 |

*Computation of the height of the gage zero at Sacketts Harbor, New York, above the gage zero at Charlotte, New York, in the spring of 1874*—Continued.

| Day. | Hour. | Gage reading. | | Difference. |
|---|---|---|---|---|
| | | Sacketts Harbor. | Charlotte. | |
| 1874. | | *Feet* | *Feet.* | *Feet.* |
| May 30 | 7 a. m. | 5.00 | 2.81 | 2.19 |
| | 9 p. m. | 5.03 | 2.82 | 2.21 |
| June 1 | 9 p. m. | 4.97 | 2.82 | 2.15 |
| 2 | 7 a. m. | 5.00 | 2.83 | 2.17 |
| | 9 p. m. | 5.02 | 2.85 | 2.17 |
| 4 | 9 p. m. | 5.10 | 2.82 | 2.28 |
| 5 | 9 p. m. | 5.00 | 2.78 | 2.22 |
| 6 | 7 a. m. | 5.00 | 2.78 | 2.22 |
| | 9 p. m. | 5.09 | 2.79 | 2.30 |
| 7 | 7 a. m. | 5.10 | 2.79 | 2.31 |
| | 2 p. m. | 5.00 | 2.81 | 2.19 |
| 8 | 9 p. m. | 4.97 | 2.81 | 2.16 |
| 9 | 7 a. m. | 4.97 | 2.82 | 2.15 |
| Mean | | | | 2.247 ± .008 |

In 1896 the gage at Charlotte was a graduated plank spiked to a pile just north of the western abutment of the Rome, Watertown and Ogdensburg Railroad bridge. The readings give the distance of the water surface above the zero of the gage. At Sacketts Harbor the arrangement was similar, the gage being spiked to an unused dock. The observer at Charlotte was Mr. J. W. Preston, harbor master; at Sacketts Harbor, Mr. Wilbur S. McKee. Observations were made morning, noon, and night, the morning and evening observations being extended into series whenever the water was so little agitated by waves that the position of its surface could be determined with precision. As the times selected for these periods of observation were also comparatively free from atmospheric disturbances, and therefore favorable to a general equilibrium of lake surface, the computations were restricted to such times. In the four months of observations there were but five occasions when series were made at both stations.

*Computation of the height of the gage zero at Sacketts Harbor, New York, above the gage zero at Charlotte, New York, in the summer of 1896.*

| Date. | Hour of commencing observation. | | Number of five-minute readings. | | Mean of readings. | | Difference. |
|---|---|---|---|---|---|---|---|
| | Sacketts Harbor. | Charlotte. | Sacketts Harbor. | Charlotte. | Sacketts Harbor. | Charlotte. | Charlotte minus Sacketts Harbor. |
| 1896. | | | | | *Feet.* | *Feet.* | |
| Aug. 8 | 7. 15 a. m. | 7 a. m. | 13 | 13 | 0. 984 | 0. 962 | —0. 022 |
| 8 | 6. 30 p. m. | 6 p. m. | 13 | 12 | 0. 912 | 0. 934 | 0. 022 |
| Sept. 9 | 5. 30 a. m. | 7 a. m. | 13 | 13 | 0. 351 | 0. 428 | 0. 077 |
| 14 | 5. 00 p. m. | 6 p. m. | 13 | 13 | 0. 270 | 0. 368 | 0. 098 |
| Oct. 27 | 8. 15 a. m. | 7 a. m. | 45 | 11 | —0. 148 | —0. 048 | 0. 100 |
| Mean | | | | | | | +0. 055 ±0. 014 |

The bench at Charlotte is a mark on the upper surface of the water table of the old light-house. The walls of the building show no cracks, and there is every reason to believe the bench stable. On May 11, 1874, the zero of gage was found by Mr. E. S. Wheeler, assistant engineer United States Lake Survey, to be 33.003 feet below this bench mark. On June 30, 1896, I leveled from the zero of the present gage to the bench mark, obtaining 38.950 as the mean of two measurements. On July 11, 1897, Mr. Warner W. Gilbert obtained 38.954 feet as a mean of two measurements.

The only bench mark existing at Sacketts Harbor in 1874 and 1896 is a point on the upper outer edge of the water table at the northeast corner of the stone building known as the Masonic Temple. In May, 1874, this was determined by Mr. Wheeler to be 12.225 feet above the zero of gage. On June 28, 1896, by duplicate measurements, I found it to be 20.425 feet above the zero of the present gage. The building bearing this mark rests on a foundation of bed rock, but nevertheless has yielded to such extent that its walls are cracked. I was informed that the cracking and repairing of the walls took place some years previous to 1874, and regard it as probable that there has been no change since that date in the height of the bench mark.

These several data are combined in the following table:

*Computation of the height of the bench mark at Charlotte, New York, above the bench mark at Sacketts Harbor, New York, in 1874 and 1896.*

| | 1874. | 1896. |
|---|---|---|
| | *Feet.* | *Feet.* |
| Charlotte bench mark above Charlotte gage zero | +33.003 | +38.950 |
| Charlotte gage zero above Sacketts Harbor gage zero | — 2.247 | — 0.055 |
| Sacketts Harbor gage zero above Sacketts Harbor bench mark | —12.225 | —20.425 |
| Sum of above = Charlotte bench mark above Sacketts Harbor bench mark | +18.531 | +18.470 |
| Difference | —0.061 | |

The results of the computations indicate that the height of the Charlotte bench mark above the Sacketts Harbor bench mark has diminished in twenty-two years to the extent of 0.061 foot. This quantity is the algebraic sum of six other quantities, two measurements through water leveling and four measurements by the engineer's level. The probable errors of the water levelings are ±0.008 and ±0.014 foot; the probable errors of my own instrumental levelings were each ±0.01 foot. Assigning the same precision to the earlier levelings, we obtain for the resulting quantity (0.061 foot) a probable error of about ±.03 foot.

This probable error attempts to express only such deviations from accuracy as are exhibited by the discordance of observations; it does not include errors of the class called constant. The result may be vitiated by the instability of either bench or by river freshets in 1874, and there are qualifications related to tides and cyclonic gradient.

The data at Sacketts Harbor are not subject to errors from stream floods. The gages at Charlotte were on the bank of the Genesee River near its mouth. The channel is deep, and at ordinary river stages the current is so gentle that river level and lake level are the same, but in time of river flood the river level is somewhat higher. In 1896 no flood periods were included, but the records for 1874 are not full enough to insure freedom from flood influences. If the Charlotte data include errors due to that cause, their correction would increase the computed change of relative height.

The tides of the Great Lakes are so small as to be masked by the seiches, but they are nevertheless of sufficient magnitude to affect an investigation of this sort. Lieut. Col. J. D. Graham determined a lunar tide of Lake Michigan at Chicago amounting to 1¾ inches and a spring tide amounting to 3½ inches.[1] Gen. C. B. Comstock determined a lunar

[1] Ann. Rept. Chief of Engineers, U. S. A., for 1860, p. 296.

tide of Lake Michigan at Milwaukee of 1 inch and a solar tide of one-half inch; and a tide of 1½ inches was found at the west end of Lake Superior.[1] The tides of Lake Ontario have not been investigated, and therefore a correction for them can not be applied. It would be quite possible to eliminate their effect by making the periods of observation include complete tidal cycles; but the local conditions gave greater importance to other criteria for the selection of times. An inspection of dates with reference to tidal cycles shows that the observations are so distributed that the influence of tide can not be great.

A complete comparative discussion of lake levels should also take account of differences of atmospheric pressure. It is evident that in a condition of equilibrium the level water surface must be deformed by local inequalities of atmospheric pressure, and the effect of pressure differences of course coexists with inequalities due to other causes. In planning these computations the intention was to apply corrections for barometric gradient, but this intention was afterwards relinquished because of the difficulty of properly discussing the available barometric data. Such examination as was given to the subject led to the opinion that during the stormless periods selected for the comparison of gage readings the error arising from the neglect of the pressure correction is small.

## PORT COLBORNE AND CLEVELAND.

The character of the gage used at Cleveland in 1858 is not described in the records I have seen. Neither is the name of the observer given, but various circumstances indicate that the readings were made either by Col. Charles Whittlesey or under his immediate direction. The readings give the distance of the water surface below the high-water level of 1838, and that level was adopted by the United States Lake Survey as the plane of reference for all observations on Lake Erie. At Port Colborne the upper sill of Lock No. 27 of the Welland Canal was the zero of measurement, and the measurement was made by the lock master, Mr. John McGillivray, by thrusting a graduated pole into the water until the end rested on the lock sill. As the reference point at Cleveland was above the water surface and that at Port Colborne below, their difference in height is obtained by adding the two readings. Most of the observations at Cleveland were made at 8 a. m. and the observations at Port Colborne at noon. At Port Colborne the direction of the wind was recorded; at Cleveland, the direction and force. I do not know the scale of force employed, but the record numbers range from 0 to 5. All observations at both stations were rejected when the wind force at Cleveland was recorded as greater than 1.

The gage zero used at Cleveland in 1895 was the upper edge of a cleat nailed to a plank forming one wall of a well in a wharf. From

[1] Ann. Rept. Chief of Engineers, U. S. A., for 1872, pp. 1033, 1035, 1040; 1875, pp. 1173, 1192, 1194.

this the observer measured to the water surface in the well with a graduated rod. The gage zero was set at the level of high water in 1838, which is mentioned in the records as "the plane of reference." Three observations were made daily, at 7 a. m., 1 p. m., and 7 p. m., the work being under the direction of the United States Engineers. At Port Colborne observation was made by means of a float connected through a chain with a counterpoise, and was therefore indirect; but the readings were checked by occasional observations with a pole, after the method of 1858. An index on the counterpoise was so adjusted as to indicate on a scale the depth of water on the lock sill. I inspected the gage in 1896, finding it in close adjustment, except that an error in either direction of a fraction of an inch might arise from friction. The observer in 1895 was Mr. John Henshaw. In the following table the readings at Port Colborne, which were recorded in feet and inches, have been converted to feet and hundredths. The record of wind at the two stations was the same as in 1858, and there were also available the wind and pressure observations of the United States Weather Bureau. From an inspection of these data three periods were selected for comparison: June 28 to July 3, July 18 to 28, and August 3 to 18. These periods are so related to the tidal cycle as nearly to eliminate tidal error.

*Computation of the height of the "plane of reference" at Cleveland, Ohio, above the sill of Lock No. 27 of the Welland Canal at Port Colborne, Ontario, in 1858 and 1895.*

| Date. | Reading at Cleveland. | Reading at Port Colborne. | Sum. | Date. | Readings at Cleveland. 7 a. m. | 1 p. m. | 7 p. m. | Mean. | Reading at Port Colborne. | Sum. |
|---|---|---|---|---|---|---|---|---|---|---|
| 1858. | *Ft. in.* | *Ft. in.* | *Ft. in.* | 1895. | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* |
| Aug. 20 | 0 9.0 | 14 7 | 15 4.0 | June 28 | ........ | 3.30 | 3.50 | 3.40 | 10.92 | 14.32 |
| 24 | 0 8.0 | 14 6 | 15 2.0 | 29 | 3.55 | 3.50 | 3.53 | 3.53 | 10.75 | 14.28 |
| 25 | 0 8.5 | 14 6 | 15 2.5 | 30 | 3.55 | 3.50 | 3.53 | 3.53 | 11.08 | 14.61 |
| 28 | 0 5.0 | 14 0 | 14 5.0 | July 1 | 3.30 | 3.40 | 3.50 | 3.40 | 10.75 | 14.15 |
| 31 | 0 8.0 | 14 3 | 14 11.0 | 2 | 3.47 | 3.45 | 3.55 | 3.49 | 10.83 | 14.32 |
| Sept. 1 | 0 7.2 | 14 4 | 14 11.2 | 3 | 3.57 | 3.60 | ....... | 3.58 | 10.92 | 14.50 |
| 3 | 0 8.0 | 14 0 | 14 8.0 | 18 | 3.57 | 3.80 | 3.83 | 3.73 | 11.00 | 14.73 |
| 6 | 0 9.5 | 13 11 | 14 8.2 | 19 | 3.85 | 3.50 | 3.70 | 3.68 | 10.75 | 14.43 |
| 9 | 0 9.4 | 14 1 | 14 10.4 | 20 | 3.80 | 3.68 | 3.53 | 3.67 | 11.25 | 14.92 |
| 12 | 0 8.7 | 14 0 | 14 8.7 | 21 | 3.33 | 3.87 | 3.63 | 3.61 | 11.00 | 14.61 |
| 14 | 0 10.7 | 13 8 | 14 6.7 | 22 | 3.90 | 3.50 | 3.60 | 3.67 | 10.75 | 14.42 |
| 15 | 0 6.0 | 13 11 | 14 5.0 | 23 | 3.60 | 3.70 | 3.70 | 3.67 | 10.83 | 14.50 |
| 16 | 0 2.3 | 13 9 | 13 11.3 | 24 | 3.68 | 3.60 | 3.70 | 3.66 | 10.83 | 14.49 |
| 18 | 0 6.7 | 14 3 | 14 9.7 | 25 | 3.72 | 3.73 | 3.82 | 3.76 | 10.67 | 14.43 |
| 20 | 0 9.3 | 14 0 | 14 9.3 | 26 | 3.70 | 3.73 | 3.63 | 3.69 | 11.00 | 14.69 |
| 23 | 1 0.8 | 13 11 | 14 11.8 | 27 | 3.75 | 3.70 | 3.32 | 3.59 | 10.92 | 14.51 |
| 24 | 1 1.5 | 13 5 | 14 6.5 | 28 | 3.50 | 3.43 | 3.80 | 3.58 | 10.75 | 14.33 |
| 25 | 1 0.9 | 14 0 | 15 0.9 | Aug. 3 | ....... | 3.79 | 3.80 | 3.80 | 11.00 | 14.80 |
| 27 | 1 1.8 | 14 2 | 15 3.8 | 4 | 3.80 | 3.73 | 3.73 | 3.75 | 10.75 | 14.50 |
| 30 | 1 2.5 | 14 5 | 15 7.5 | 5 | 3.73 | 3.68 | 3.68 | 3.70 | 10.58 | 14.28 |
| Oct. 4 | 1 1.6 | 14 0 | 15 1.6 | 6 | 3.90 | 3.73 | 3.70 | 3.78 | 11.33 | 15.11 |
| 10 | 1 3.0 | 13 10 | 15 1.0 | 7 | 3.93 | 3.95 | 3.60 | 3.83 | 11.75 | 15.58 |

*Computation of the height of the "plane of reference" at Cleveland, Ohio, above the sill of Lock No. 27 of the Welland Canal at Port Colborne, Ontario, in 1858 and 1895*—Continued.

| Date. | Reading at Cleveland. | Reading at Port Colborne. | Sum. | Date. | Readings at Cleveland. | | | | Reading at Port Colborne. | Sum. |
|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | 7 a. m. | 1 p. m. | 7 p. m. | Mean. | | |
| 1858. | *Ft. in.* | *Ft. in.* | *Ft. in.* | 1895. | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* |
| Oct. 11 | 1 4.0 | 13 6 | 14 10.0 | Aug. 8 | 3.75 | 3.60 | 3.80 | 3.72 | 10.67 | 14.39 |
| 14 | 1 2.2 | 14 3 | 15 5.2 | 9 | 3.72 | 3.70 | 3.59 | 3.67 | 10.83 | 14.50 |
| 15 | 0 11.0 | 14 3 | 15 2.0 | 10 | 3.90 | 3.75 | 3.70 | 3.78 | 10.75 | 14.53 |
| 16 | 0 10.5 | 14 5 | 15 3.5 | 11 | 3.80 | 3.50 | 3.64 | 3.65 | 10.67 | 14.32 |
| 18 | 1 5.0 | 14 2 | 15 7.0 | 12 | 3.54 | 3.78 | 3.68 | 3.67 | 11.00 | 14.67 |
| 19 | 1 4.9 | 13 5 | 14 9.9 | 13 | 3.60 | 3.67 | 3.60 | 3.62 | 10.83 | 14.45 |
| 20 | 1 2.0 | 13 3 | 14 5.0 | 14 | 3.68 | 3.65 | 3.70 | 3.68 | 11.00 | 14.68 |
| 21 | 1 4.7 | 13 2 | 14 6.7 | 15 | 3.61 | 3.72 | 3.65 | 3.66 | 11.00 | 14.66 |
| 22 | 1 4.8 | 13 7 | 14 11.8 | 16 | 3.52 | 3.59 | 3.67 | 3.59 | 10.83 | 14.42 |
| 25 | 1 1.8 | 12 5 | 13 6.8 | 17 | 3.70 | 3.68 | 3.70 | 3.69 | 10.75 | 14.44 |
| 26 | 1 4.4 | 12 7 | 13 11.4 | 18 | 3.80 | 3.83 | 3.70 | 3.78 | 11.17 | 14.95 |
| 27 | 1 4.9 | 12 7 | 13 11.9 | | | | | | | |
| 28 | 1 4.3 | 12 11 | 14 3.3 | | | | | | | |
| 30 | 1 6.1 | 13 3 | 14 9.1 | | | | | | | |
| Mean (in feet) | | | 14.800 ±.057 | | | | | | | 14.561 (±.022) |

| Height of Cleveland plane of reference above Port Colborne lock-sill— | | |
|---|---|---|
| 1858 | 14.800 | ±.057 |
| 1895 | 14.561 | ±.022 |
| Difference | —0.239 | ±.06 |

The zero of gage at Port Colborne, being submerged masonry, is of unquestioned stability. The canal was constructed in 1833, and if any settling followed construction it was doubtless complete before 1858; but the appearance of the masonry above the water gives no suggestion of yielding.

The earlier work at Cleveland was connected with several bench marks, all of which have been destroyed, but before the disappearance of the last one the datum was transferred by leveling to other points. The chain on which the record depends is as follows:

1. "Top of coping of the northeast wall of the Ohio Canal lock where it joins the river." The high water of 1838 was directly compared with this bench, and Whittlesey states that it is 6.30 feet above that high-water plane.[1] As the observations in 1858 were made near the lock, and as Whittlesey, who reports them, was a civil engineer whose writings show that he appreciated the importance of precise bench marks, it is probable that the observations were properly connected with the bench. Explicit statement, however, is lacking; the record merely refers the lake level to the high water of 1838. The bench was destroyed in 1877 or 1878.

[1] Canadian Naturalist, Vol. VII, 1875, p. 412.

2. "Cross on water table, northeast corner of Johnson House block, southwest corner of Front and East River streets." On June 15, 1875 (as shown by manuscript records in the office of the Chief of Engineers, U. S. A.), Assistant Engineer T. W. Wright, United States Lake Survey, leveled from this bench mark to the canal lock coping (1), finding the difference (1 above 2) to be 3.67 feet. This bench mark is still in existence. The walls of the building are cracked in such manner as to indicate some settling of the northeast corner, and the broad flagstone on which the bench is marked stands (in 1897) 0.04 foot lower than the next stone of the water table toward the west. As the lower stone supports part of the building and the higher stone carries no load, the latter may be assumed to show the original level of the former. It is impossible to say whether this settling affects the record of water levels. The building was erected in 1842, and is therefore 55 years old; it was 33 years old in 1875 when the datum of levels was transferred to it. The datum remained with it eighteen years, until 1893. If settling has progressed at a uniform rate, the datum was affected 0.013 foot, but it is equally possible that the settling belonged to the early history of the building and that a condition of practical stability was reached prior to 1875.

3. "Bottom of west angle iron, on bottom of north longitudinal plate girder, middle of first full-depth bent, close to stone pier, new L. S. & M. S. R. R. drawbridge, now [1893] being finished." As the bridge is symmetric and reversible, this description applies to two different points, but measurement shows that they have the same height. It was copied from manuscript records in the United States Engineers office at Cleveland, courteously placed at my service by Col. Jared A. Smith. The records show that in June, 1893, the bridge bench (3) was connected by leveling with the Johnson House bench (2) and also with the gage zero, and that the gage zero was checked by the bridge in 1896 and found correct. The gage readings in 1895 (used in our computations) are thus referred to the bridge bench. The height of the bridge bench is given as 4.34 feet above the "plane of reference," and by implication as 1.71 feet above the Johnson House bench (2). The drawbridge rests on a stone pier many years older than the present bridge, and there can be little question of its stability.

In these records of bench marks and levelings in Cleveland there is certainly much to be desired, but the presumption is nevertheless in favor of good work.

It appears from the computation that the ground at Port Colborne has risen, as compared to the ground at Cleveland, 0.239 foot, or about $2\frac{7}{8}$ inches in thirty-seven years. The probable error of this measurement, as indicated by the discordance of gage data, is three-fourths of an inch.

As a check upon this result, a third computation was made from gage readings in the summer of 1872, a year in which the gage zero at Cleveland was connected with the canal-lock bench mark by instrumental leveling. That computation gives for the height of the plane of reference at Cleveland above the lock sill at Port Colborne 14.714 feet. If we assume a gradual change from 1858 to 1895, and interpolate between 14.800 feet, the determination for 1858, and 14.561, the determination for 1895, we obtain for the summer of 1872 the value 14.710 feet, which differs from the result of that year's observations by only 0.004 foot. The observations on Lake Erie thus accord well with the theory of a progressive southward tilting of the land.

The Port Colborne gage is not so related to streams as to subject its readings to error from floods. The Cleveland gage, like the Charlotte, is on a river estuary, and the readings are subject to influence by floods. The records include no systematic account of the condition of the river, and it is therefore possible that some of the observations were made when the river level was above the lake level.

## PORT AUSTIN AND MILWAUKEE.

At each of these stations automatic gages were maintained for several years, and their tracings give the height of water level with an amount of detail permitting the complete elimination of seiches and tides; but there was, unfortunately, some uncertainty as to the position of the zeros, and the danger of thus introducing constant errors led me to avoid the automatic records and choose times when other gages were employed. The earlier period selected for the comparison was the summer of 1876, and the gages then used were floats carrying graduated vertical rods. The force and direction of the wind were recorded at Port Austin by the gage observer, and at Milwaukee by the United States Weather Bureau. From an inspection of these records, together with the Weather Bureau records of barometric gradient, selection was made of the periods July 11 to 19 and August 16 to 24, excepting only certain hours when the force of the local wind was recorded as greater than 3 in a scale of 10. This gave 46 separate comparisons, from which the difference in height of the gage zeros was computed. The chosen periods are well disposed with reference to tides. The readings at Milwaukee were made at 7 a. m., 1 p. m., and 6 p. m., by Mr. John McCabe; at Port Austin the hours were 7 a. m., 2 p. m., and 9 p. m., and the observer was Mr. J. W. Kimball. In the computations the midday observations, though one hour apart, and the evening observations, though three hours apart, were treated as simultaneous.

*Computation of height of gage zero at Port Austin, Michigan, above gage zero at Milwaukee, Wisconsin, in the summer of 1876.*

| Date. | Readings at Milwaukee. | | | Readings at Port Austin. | | | Differences. | | |
|---|---|---|---|---|---|---|---|---|---|
| | 7 a. m. | 1 p. m. | 6 p. m. | 7 a. m. | 2 p. m. | 9 p. m. | | | |
| 1876 | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* |
| July 11 | 2. 17 | 2. 23 | 2. 12 | 7. 22 | 7. 31 | 7. 47 | 5. 05 | 5. 08 | 5. 35 |
| 12 | 2. 12 | 2. 03 | 2. 16 | 7. 54 | 7. 25 | 7. 22 | 5. 42 | 5. 22 | 5. 06 |
| 13 | 2. 20 | 2. 05 | 2. 35 | 7. 35 | 7. 37 | 7, 35 | 5. 15 | 5. 32 | 5. 00 |
| 14 | 1. 95 | 2. 12 | 2. 10 | 7. 50 | 7. 41 | 7. 30 | 5. 55 | 5. 29 | 5. 20 |
| 15 | 2. 16 | 2. 15 | 2. 06 | 7. 33 | 7. 35 | 7. 30 | 5. 17 | 5. 20 | 5. 24 |
| 16 | 2. 13 | 2. 12 | 2. 11 | 7. 34 | 7. 25 | 7. 40 | 5. 21 | 5. 13 | 5. 29 |
| 17 | 2. 15 | 2. 20 | 2. 21 | 7. 37 | 7. 37 | 7. 30 | 5. 22 | 5. 17 | 5. 09 |
| 18 | 2. 20 | 2. 07 | 2. 20 | 7. 21 | 7. 25 | 7. 28 | 5. 01 | 5. 18 | 5. 08 |
| 19 | 2. 18 | ........ | ........ | 7. 34 | ........ | ........ | 5. 16 | ........ | ........ |
| Aug. 16 | 2. 29 | ........ | 2. 20 | 7. 50 | ........ | 7. 35 | 5. 21 | ........ | 5. 15 |
| 17 | 2. 19 | ........ | 2. 21 | 7. 33 | ........ | 7. 50 | 5. 14 | ........ | 5. 29 |
| 18 | 2. 23 | 2. 02 | 2. 27 | 7. 50 | 7. 35 | 7. 40 | 5. 27 | 5. 33 | 5. 13 |
| 19 | ........ | 2. 22 | ........ | ........ | 7. 45 | ........ | ........ | 5. 23 | ........ |
| 20 | ........ | 2. 18 | 2. 23 | ........ | 7. 62 | 7. 37 | ........ | 5. 44 | 5. 14 |
| 21 | 2. 25 | 2. 20 | 2. 24 | 7. 41 | 7. 34 | 7. 37 | 5. 16 | 5. 14 | 5. 18 |
| 22 | 2. 32 | 2. 18 | 2. 33 | 7. 38 | 7. 51 | 7. 40 | 5. 06 | 5. 33 | 5. 07 |
| 23 | 2. 26 | 1. 91 | 2. 10 | 7. 38 | 7. 60 | 7. 59 | 5. 12 | 5. 69 | 5. 49 |
| 24 | 2. 25 | ........ | 2. 42 | 7. 49 | ........ | 7. 49 | 5. 24 | ........ | 5. 07 |
| Mean | | | | | | | | | 5. 210 ±. 013 |

In 1896 the gage at Milwaukee consisted of a graduated rod held in the observer's hand in measuring down to the water from a fixed point or zero. At Port Austin a board, carrying a graduated scale, was spiked to the side of a timber crib, and the position of the water surface on the scale was noted by the observer. At each of these stations a series of 12 observations, at five-minute intervals, was made every morning and evening when the surface of the water was nearly smooth. The mean of a series was afterwards treated as one observation, and the computation was based on the simultaneous pairs of observations—53 in number. The selection of times was thus determined by conditions favorable for the elimination of seiches, but it appears by inspection that tidal influences also are very nearly eliminated. The observers were: At Milwaukee, Mr. John McCabe; at Port Austin, Mr. John P. Smith.

As the zero at Milwaukee was above the water, and the zero at Port Austin below, the sum of the readings gives the height of one zero above the other.

*Computation of height of gage zero at Milwaukee, Wisconsin, above gage zero at Port Austin, Michigan, in the summer of 1896.*

| Date. | Time. | Readings (means of series). | | Sum. |
|---|---|---|---|---|
| | | Milwaukee. | Port Austin. | |
| 1896. | | *Feet.* | *Feet.* | *Feet.* |
| July 20 | A. M. | 5.528 | 1.271 | 6.799 |
| 24 | A. M. | 5.483 | 1.333 | 6.816 |
| 28 | A. M. | 5.703 | 1.385 | 7.088 |
| 29 | A. M. | 5.470 | 1.354 | 6.824 |
| | P. M. | 5.560 | 1.375 | 6.935 |
| Aug. 1 | A. M. | 5.436 | 1.297 | 6.733 |
| 2 | P. M. | 5.620 | 1.425 | 7.045 |
| 7 | P. M. | 5.447 | 1.420 | 6.867 |
| 9 | A. M. | 5.519 | 1.448 | 6.967 |
| 11 | A. M. | 5.575 | 1.433 | 7.008 |
| 14 | A. M. | 5.338 | 1.455 | 6.793 |
| 20 | A. M. | 5.587 | 1.340 | 6.927 |
| 20 | P. M. | 5.571 | 1.406 | 6.977 |
| 21 | A. M. | 5.558 | 1.330 | 6.888 |
| | P. M. | 5.588 | 1.391 | 6.979 |
| 22 | A. M. | 5.505 | 1.259 | 6.764 |
| | P. M. | 5.540 | 1.229 | 6.769 |
| 24 | P. M. | 5.595 | 1.375 | 6.970 |
| 25 | A. M. | 5.721 | 1.221 | 6.942 |
| 25 | P. M. | 5.792 | 1.268 | 7.060 |
| 28 | P. M. | 5.721 | 1.279 | 7.000 |
| 30 | P. M. | 5.797 | 1.239 | 7.036 |
| Sept. 1 | P. M. | 5.725 | 1.259 | 6.984 |
| 2 | P. M. | 5.748 | 1.203 | 6.951 |
| 4 | P. M. | 5.720 | 1.248 | 6.968 |
| 5 | P. M. | 5.515 | 1.134 | 6.649 |
| 7 | P. M. | 5.739 | 1.275 | 7.014 |
| 8 | A. M. | 5.649 | 1.203 | 6.852 |
| | P. M. | 5.595 | 1.139 | 6.734 |
| 9 | P. M. | 5.585 | 1.077 | 6.662 |
| 14 | A. M. | 5.584 | 1.208 | 6.792 |
| 15 | A. M. | 5.560 | 1.181 | 6.741 |
| 20 | P. M. | 5.892 | 1.281 | 7.173 |
| 23 | A. M. | 5.791 | 1.307 | 7.098 |
| 25 | A. M. | 5.932 | 0.803 | 6.735 |
| 28 | P. M. | 5.755 | 1.167 | 6.922 |
| 29 | A. M. | 5.615 | 1.013 | 6.628 |

*Computation of height of gage zero at Milwaukee, Wisconsin, above gage zero at Port Austin, Michigan, in the summer of 1896*—Continued.

| Date. | | Time. | Readings (means of series). | | Sum. |
|---|---|---|---|---|---|
| | | | Milwaukee. | Port Austin. | |
| 1896. | | | *Feet.* | *Feet.* | *Feet.* |
| Oct. | 2 | A. M. | 5.566 | 1.250 | 6.816 |
| | 3 | A. M. | 5.594 | 1.465 | 7.059 |
| | | P. M. | 5.574 | 1.186 | 6.760 |
| | 4 | P. M. | 5.632 | 1.101 | 6.733 |
| | 5 | P. M. | 5.705 | 1.085 | 6.790 |
| | 10 | P. M. | 5.506 | 0.889 | 6.395 |
| | 15 | A. M. | 5.784 | 0.769 | 6.553 |
| | 17 | A. M. | 5.642 | 1,444 | 7.086 |
| | 18 | A. M. | 5.720 | 1,398 | 7.118 |
| | 19 | A. M. | 5.846 | 1.215 | 7.061 |
| | 22 | A. M. | 6.182 | 1.212 | 7.394 |
| | 25 | A. M. | 6.139 | 0.800 | 6.939 |
| | 26 | P. M. | 5.960 | 0.858 | 6.818 |
| | 27 | A. M. | 5.918 | 0.750 | 6.668 |
| | | P. M. | 5.809 | 0.722 | 6.531 |
| | 29 | A. M. | 5.864 | 0.724 | 6.588 |
| Mean | | | | | 6.875 ±.019 |

Milwaukee is well provided with engineer bench marks, and it is probable that thorough research would establish the connection of the gage zeros at each epoch with several of the bench marks; but after inspection of the data readily accessible, I thought it best to make use of only one bench, that called the "check point." This consists of the top of a copper bolt leaded into the north side of the center pier of the swing bridge over the river between Chestnut and Division streets. The gage observer is required at stated intervals to check the stability of the zero of his gage by means of this check point. Using two rods, with the aid of an assistant he makes a series of simultaneous measurements from the check point and from the gage zero down to the water level, and from these measurements the relation of the gage zero to the check point is determined. Their relation has also been determined by means of the engineer's level at various times, and was so determined on August 8, 1876, by Assistant Engineer L. L. Wheeler, who found the check point 0.843 foot above the gage zero. In 1896 the check observations by the observer were very thorough, series of twenty simultaneous readings being made every fortnight, and from five of these series the relation of the two points is computed as follows:

*Computation of height of Milwaukee check point above Milwaukee zero of gage in the summer of 1896.*

| | Feet. |
|---|---|
| July 12 (mean of twenty comparisons by simultaneous readings) | 1.203 |
| July 26 | 1.212 |
| August 14 | 1.200 |
| August 28 | 1.203 |
| September 16 | 1.206 |
| Mean | 1.205 ±.002 |

In response to a letter of inquiry as to the stability of the Milwaukee check point, Capt. George A. Zinn, United States engineer in charge of harbor improvements, writes as follows:

The Chestnut Street Bridge, on the center pier of which the check point is established, was built in 1872.

Mr. G. H. Benzenberg, city engineer, states that the pier rests on a pile foundation; that to his knowledge the drawbridge has never been releveled since put in place, and that if any appreciable settlement had taken place in the center pier it would have interfered with the operating of the swing bridge. He stated positively that no settlement had occurred.

The principal bench mark used in 1876 at Port Austin, called the Wisner bench mark, was a copper bolt leaded into bed rock; but in 1896 I was unable to find it, and, as at Milwaukee, I had recourse to a bench mark originally established and used as a check point. It is the top of an iron bolt driven into a vertical face of bed rock on the west side of a promontory opposite the residence of Mr. J. W. Kimball. In July, 1875, and October, 1876, Assistant Engineer T. W. Wright found the check point 7.424 feet below the Wisner bench mark; in June, 1896, I found the gage zero 5.125 feet below the check point, this quantity being the mean of two measurements.

Manuscript records in the archives of the Lake Survey state that the Port Austin gage zero was originally placed on a level with the Wisner bench mark, but that in July, 1875, it was 0.003 foot too low, and that on October 18, 1876, it was 0.040 foot too low, having settled during the interval. As the observations used fall within this interval, it was necessary to make some assumption in regard to this settling, and the assumption made was that it had been at uniform rate through the whole period. The correction interpolated for the time of observation was 0.034 foot. Combining this correction with data from leveling in 1875 and 1876, I obtained as the height of the gage zero above the check point in July and August, 1876, 7.460 feet. The various data thus described are combined in the following table.

*Computation of height of Milwaukee check point above Port Austin check point in the summers of 1876 and 1896.*

| | 1876. | 1896. |
|---|---|---|
| | *Feet.* | *Feet.* |
| Milwaukee check point above Milwaukee gage zero | 0.843 | 1.205 |
| Milwaukee gage zero above Port Austin gage zero | —5.210 | 6.875 |
| Port Austin gage zero above Port Austin check point | 7.460 | —5.125 |
| Sum of above = Milwaukee check point above Port Austin check point | 3.093 | 2.955 |
| Difference | —0.138 | |

This result indicates that the ground at Milwaukee, as compared to the ground at Port Austin, has subsided 0.138 foot in the twenty years from 1876 to 1896. It is the algebraic sum of six measurements, of which three are levelings by water surface and three by the engineer's level. The probable errors of the water-level measurements are $\pm 0.019$, $\pm 0.013$, and $\pm 0.002$. The probable errors of the Port Austin levelings in 1896, as indicated by the discordance of two independent results, is $\pm 0.008$. If the probable error of each of the other measurements was $\pm 0.010$, the probable error of the result is less than $\pm .03$ foot. There is also an uncertainty arising from the possibility that the stone pier to which the Milwaukee check mark is attached has settled, another uncertainty due to the possibility of river floods, and a third involved in the assumption that the settling of the Port Austin gage zero in 1876 was at a uniform rate. If all the settling of the Port Austin zero took place before the period of observation, the assumption makes the result too large by 0.006 foot; if all the settling took place after the observations, the assumption makes the result too small by 0.031 foot. The Port Austin record is free from stream-flood influences, but the Milwaukee gage station is on a narrow estuary, like the stations at Charlotte and Cleveland.

## ESCANABA AND MILWAUKEE.

In comparing Escanaba with Milwaukee the same general periods of observation were employed as in comparing Port Austin with Milwaukee, but the individual days, though selected in the same manner, were in part different. Fifty-one separate comparisons were made in 1876, and 52 in 1896. The selection of times was controlled by conditions favorable for the elimination of seiches, but the combination of days chosen was found to approximately eliminate tidal effects also.

The observations at Escanaba in 1876 were conducted in the same manner as at Milwaukee and Port Austin, the hours being 7 a. m., 2 p. m., and 9 p. m., and the observer Mr. George Preston. In 1896 the system was the same as at Milwaukee, the observer being Mr. Clinton

B. Oliver. The following tables give the computations for the two years:

*Computation of height of gage zero at Milwaukee, Wisconsin, above gage zero at Escanaba, Michigan, in the summer of 1876.*

| Date. | Readings at Escanaba. | | | Readings at Milwaukee. | | | Differences. | | |
|---|---|---|---|---|---|---|---|---|---|
| | 7 a. m. | 2 p. m. | 9 p. m. | 7 a. m. | 1 p. m. | 6 p. m. | | | |
| 1876. | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* |
| July 11 | 1.86 | 1.90 | 1.78 | 2.17 | 2.23 | 2.12 | 0.31 | 0.33 | 0.34 |
| 12 | 1.90 | 1.75 | 1.78 | 2.12 | 2.03 | 2.16 | .22 | .28 | .38 |
| 13 | 2.00 | 2.07 | 1.95 | 2.20 | 2.05 | 2.35 | .20 | —.02 | .40 |
| 14 | 2.10 | 2.15 | 2.05 | 1.95 | 2.12 | 2.10 | —.15 | —.03 | .05 |
| 15 | 1.96 | 1.95 | 2.00 | 2.16 | 2.15 | 2.06 | .20 | .20 | .06 |
| 16 | 1.90 | 1.95 | 1.85 | 2.13 | 2.12 | 2.11 | .23 | .17 | .26 |
| 17 | 1.89 | 1.80 | 1.85 | 2.15 | 2.20 | 2.21 | .26 | .40 | .36 |
| 18 | 2.05 | 1.85 | 1.95 | 2.20 | 2.07 | 2.20 | .15 | .22 | .25 |
| 19 | 1.75 | ........ | ........ | 2.18 | ........ | ........ | .43 | ........ | ........ |
| Aug. 16 | 2.07 | 1.94 | 1.90 | 2.29 | 2.30 | 2.20 | .22 | .36 | .30 |
| 17 | 2.00 | 1.95 | 1.95 | 2.19 | 2.13 | 2.21 | .19 | .18 | .26 |
| 18 | 1.70 | 1.93 | 2.00 | 2.23 | 2.02 | 2.27 | .53 | .09 | .27 |
| 19 | 1.78 | 2.00 | ........ | 2.13 | 2.22 | ........ | .35 | .22 | ...... |
| 20 | 1.95 | 1.90 | 1.85 | 2.23 | 2.18 | 2.23 | .28 | .28 | .38 |
| 21 | 1.85 | 2.05 | 2.10 | 2.25 | 2.20 | 2.24 | .40 | .15 | .14 |
| 22 | 1.83 | 2.05 | 1.90 | 2.32 | 2.18 | 2.33 | .49 | .13 | .43 |
| 23 | 1.91 | 1.83 | 1.75 | 2.26 | 1.91 | 2.10 | .35 | .08 | .35 |
| 24 | 1.95 | 1.85 | 1.90 | 2.25 | 2.10 | 2.42 | .30 | .25 | .52 |
| Mean | | | | | | | | | 0.255 ±.012 |

*Computation of height of gage zero at Escanaba, Michigan, above gage zero at Milwaukee, Wisconsin, in the summer of 1896.*

| Date. | Time. | Readings (means of series). | | Difference. |
|---|---|---|---|---|
| | | Milwaukee. | Escanaba. | |
| 1896. | | *Feet.* | *Feet.* | *Feet.* |
| July 2 | P. M. | 5.465 | 5.917 | 0.452 |
| 7 | A. M. | 5.434 | 5.907 | .473 |
| 8 | A. M. | 5.505 | 5.920 | .415 |
| 9 | A. M. | 5.348 | 5.837 | .489 |
| | P. M. | 5.356 | 5.765 | .409 |
| 10 | A. M. | 5.442 | 5.771 | .329 |
| | P. M. | 5.567 | 5.694 | .127 |
| 11 | P. M. | 5.576 | 5.771 | .195 |
| 13 | A. M. | 5.411 | 5.869 | .458 |
| | P. M. | 5.493 | 5.776 | .283 |
| 14 | A. M. | 5.574 | 5.750 | .176 |
| 17 | A. M. | 5.431 | 5.865 | .434 |
| 19 | A. M. | 5.524 | 6.007 | .483 |
| | P. M. | 5.496 | 5.887 | .391 |
| 20 | A. M. | 5.528 | 5.803 | .275 |
| 21 | A. M. | 5.573 | 5.973 | .400 |
| 23 | A. M. | 5.645 | 5.908 | .263 |
| 25 | A. M. | 5.601 | 5.856 | .255 |
| 28 | A. M. | 5.703 | 5.857 | .154 |
| 31 | A. M. | 5.446 | 5.938 | .492 |
| Aug. 1 | P. M. | 5.360 | 5.859 | .499 |
| 4 | A. M. | 5.654 | 5.912 | .258 |
| 8 | P. M. | 5.347 | 5.954 | .607 |
| 9 | A. M. | 5.519 | 5.658 | .139 |
| 10 | P. M. | 5.328 | 5.546 | .218 |
| 13 | A. M. | 5.273 | 5.616 | .343 |
| | P. M. | 5.378 | 5.752 | .374 |
| 14 | A. M. | 5.338 | 5.670 | .332 |
| 15 | A. M. | 5.360 | 5.730 | .370 |
| | P. M. | 5.402 | 5.710 | .308 |
| 19 | A. M. | 5.414 | 5.878 | .464 |
| 21 | A. M. | 5.558 | 5.935 | .377 |
| | P. M. | 5.588 | 5.872 | .284 |
| 22 | A. M. | 5.505 | 5.698 | .183 |
| Sept. 4 | A. M. | 5.734 | 6.028 | .294 |
| 13 | P. M. | 5.452 | 5.848 | .396 |
| 14 | A. M. | 5.584 | 5.762 | .178 |
| 16 | A. M. | 5.500 | 5.937 | .437 |

*Computation of height of gage zero at Escanaba, Michigan, above gage zero at Milwaukee, Wisconsin, in the summer of 1896*—Continued.

| Date. | Time. | Readings (means of series). | | Difference. |
|---|---|---|---|---|
| | | Milwaukee. | Escanaba. | |
| 1896. | | *Feet.* | *Feet.* | *Feet.* |
| Sept. 18 | A. M. | 5.701 | 6.047 | .346 |
| 26 | A. M. | 5.914 | 6.187 | .273 |
| 28 | P. M. | 5.755 | 6.224 | .469 |
| 29 | P. M. | 5.510 | 6.133 | .623 |
| Oct. 7 | A. M. | 5.514 | 6.164 | .650 |
| 14 | A. M. | 5.731 | 6.270 | .539 |
| 14 | P. M. | 5.813 | 6.157 | .344 |
| 17 | P. M. | 5.622 | 6.160 | .538 |
| 19 | A. M. | 5.846 | 6.287 | .441 |
| 20 | A. M. | 5.857 | 6.346 | .489 |
| 22 | A. M. | 6.182 | 6.539 | .357 |
| | P. M. | 6.148 | 6.540 | .392 |
| 26 | A. M. | 6.080 | 6.471 | .391 |
| | P. M. | 5.980 | 6.544 | .564 |
| Mean | | | | 0.374 ±.012 |

The bench employed at Milwaukee has already been described. At Escanaba there were three bench marks in good standing, as follows: No. 1, the top of the water sill on the southeast corner of the Adler building, northwest corner of Ludington street and Druseman avenue; No. 2, the top of the water sill of the Escanaba light-house at the north side of front door, against the brick wall; No. 3 is described in 1876 as the "center of a copper bolt set horizontally in the foundation of the light-house, west side, north corner, 3 feet north from steps." In a description by Capt. George A. Zinn, dated June 30, 1896, the top of the bolt is specified. I am informed by Mr. Clinton B. Oliver, the gage observer, that the diameter of the bolt is three-eighths inch. The relative heights of two or more of these bench marks have been determined in at least six different years, the measurements being made independently with the engineer's level. It is advantageous to compare these measurements, not only to learn what confidence is to be reposed in the individual benches, but for the sake of whatever light may be cast on the general precision of such data.

*Comparison of Escanaba bench marks with one another.*

| Year. | Above zero of gage. | | | Difference between bench marks. | | | Deviation from mean. | | |
|---|---|---|---|---|---|---|---|---|---|
| | No. 1. | No. 2. | No. 3. | 1—2. | 1—3. | 2—3. | 1—2. | 1—3. | 2—3. |
| | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* | *Feet.* |
| 1874.. | 4.135 | —2.100 | ....... | *a*6.235 | ....... | ....... | ........ | ........ | ........ |
| 1875.. | 7.859 | 2.382 | ....... | 5.477 | ....... | ....... | +0.003 | ........ | ........ |
| 1876.. | 7.874 | 2.409 | 1.392 | 5.465 | 6.482 | 1.017 | — .009 | +0.002 | +0.005 |
| 1880.. | 7.878 | 2.402 | 1.035 | 5.476 | *a*6.843 | *a*1.367 | + .002 | ........ | ........ |
| 1887.. | 7.753 | 2.285 | 1.280 | 5.468 | 6.473 | 1.005 | — .006 | — .011 | — .007 |
| 1896.. | 7.780 | 2.297 | 1.283 | 5.483 | 6.497 | 1.014 | + .009 | + .013 | + .002 |
| Mean. | ....... | ........ | ....... | 5.474 | 6.484 | 1.012 | ........ | ........ | ........ |

*a* Not used in computing means.

In this table the reading of the height of bench mark No. 3 in 1876 is corrected for the distance between center and top of bolt. In the first division of the table the benches are referred to zero of gage, but as the gage was not constant in position these numbers differ widely from year to year. In the second division the relations of the gages one to another are given, being deduced by subtraction from the numbers of the first division, and these figures are more accordant. It appears, however, that the difference between benches 1 and 2 in 1874 departs widely from differences found in other years, and it is therefore probable that a blunder of measurement or record was made in that year. It appears further, by inspection, that the difference between benches 1 and 3 and the difference between benches 2 and 3 in 1880 are not in accord with the differences found in other years, and it is evident that some blunder was made in the measurement or record of the height of bench 3 for that year. These figures were accordingly thrown out and not used in the computation of the means. The numbers of the third division were obtained by subtracting the means from the several numbers of the second division, and they show the deviations from mean after rejecting the records showing gross errors. Inspection of the table of deviations shows that their signs are irregularly distributed, and discovers no evidence of progressive change from year to year. It is therefore probable that all three of the benches are stable, and that the deviations of the measurements from uniformity represent ordinary errors of observation. They may accordingly be used as a rough measure of the precision, barring blunders, of the instrumental leveling on which the results of this investigation largely depend. Their mean is 0.006 foot, and the computed probable error of a single measurement is ±0.008 foot. In combining various data for the comparison of Escanaba with Milwaukee, bench mark No. 1 of Escanaba was first used.

*Computation of the height of Escanaba bench mark No. 1 above Milwaukee check point in the summers of 1876 and 1896.*

| | 1876. | 1896. |
|---|---|---|
| | *Feet.* | *Feet.* |
| Escanaba bench No. 1 above Escanaba gage zero | 7.874 | 7.780 |
| Escanaba gage zero above Milwaukee gage zero | −0.255 | .374 |
| Milwaukee gage zero above Milwaukee check point | −0.843 | −1.205 |
| Sum of above = Escanaba bench No. 1 above Milwaukee check point | 6.776 | 6.949 |
| Difference | + .173 | |

The result indicates that the ground at Escanaba, as compared with the ground at Milwaukee, has risen 0.173 foot in twenty years. This quantity is the algebraic sum of six measurements, of which three were made through water leveling and three by instrumental leveling. The probable errors of the water levelings are ±0.012, ±0.012, and ±0.002 foot; the estimated probable error of the instrumental levelings at Milwaukee is ±0.010 foot, and of the two levelings at Escanaba each ±0.008 foot. This gives as the probable error of the result ±0.022 foot.

A similar computation, using bench mark No. 2 instead of bench mark No. 1, gives 0.155 foot instead of 0.173, and a compution based on bench mark No. 3 gives 0.156. The mean of the three results is 0.161 foot, with a probable error of ±0.022 foot. The only important uncertainties to which this result is subject, besides those indicated by the discordance of measurements, arise from the possibility of the settling of the bridge pier to which the Milwaukee check point is attached and the possibility of river floods.

### DISCREPANCY NOTED BY CAPTAIN MARSHALL.

In the later work of the United States Lake Survey all determinations of lake level were referred to the high-water level of 1838, which is called the "plane of reference." That plane was directly observed by Dr. I. A. Lapham, the geologist, and with the aid of a bench mark on his house at Milwaukee was permanently recorded. For other stations on Lake Michigan-Huron its position was determined by assuming that the level of 1838 had everywhere the same height above the mean lake level as determined by long series of observations. For the determination of this plane at Escanaba use was made of observations for the period from January 1, 1860, to December 31, 1875. In 1887 Capt. W. L. Marshall, U. S. E., under whose direction the gage readings at Milwaukee and Escanaba were then made, detected a discrepancy, which he reported to the Chief of Engineers in a letter dated October 1.[1]

[1] Ann. Rept. Chief of Engineers, U. S. A., for 1887, part 3, p. 2417.

In former reports the zero of Escanaba gage has been assumed as 0.76 foot above the plane of reference, but a comparison of corrected readings at Milwaukee and Escanaba shows that the determinations of the plane of reference at Milwaukee and Escanaba vary 0.187 foot, the Escanaba plane being too high or the Milwaukee determination too low.

In the light of present knowledge it seems probable that the discrepancy thus noted by Captain Marshall as an error was occasioned either wholly or in chief part by the progressive tilting of the land. This conclusion is difficult of verification, because little record survives of such checks as may have been made upon the heights of gage zeros during the period 1860–1875; but the indicated change agrees in direction, and approximately in rate, with the change deduced from the present investigation. From the middle of the period 1860-1875 to the summer of 1887 was an interval of twenty years, equal to the interval 1876–1896 here used, and the discrepancy of 0.187 foot discovered by Captain Marshall differs from the change of 0.161 foot here deduced by a quantity little greater than the probable error ascribed to the latter determination.

## SUMMARY OF RESULTS.

In the following table are assembled the numerical results as to changes in relative height of the four pairs of stations. Besides the measured changes, the table includes the periods intervening between dates of measurement and distances between the stations of each pair. The lines connecting pairs of stations have a southwesterly direction (fig. 99, p. 613), and it is the northeastern station of each pair that appears to have risen as compared to the other.

The results thus show a general agreement with the working hypothesis, that the latest change recorded by geologic data is still in progress. To make the comparison quantitative there should be substituted for the direct distances between stations the corresponding distances in the assumed direction of tilting, S. 27° W., and the measured results for various distances and various time intervals should be reduced to a common basis. In the third column of the table are the reduced distances, and in the sixth the reduced rates of change. Assuming the change to have a uniform rate and to be the same for all parts of the region, the measurements at the different pairs of stations give for a distance of 100 miles and a period of a century the quantities of the sixth column. The seventh column contains the probable errors of quantities in the sixth, and is based on the probable errors of the measured changes in pairs of stations.

*Summary of distances, time intervals, and measurements of differential earth movements.*

| Pairs of stations. | Direct distance. | Distance in direction S. 27° W. | Interval between dates of measurement. | Change in relative height. | Change per 100 miles per century. | Probable error of quantities in last column. |
|---|---|---|---|---|---|---|
| | *Miles.* | *Miles.* | *Years.* | *Feet.* | *Feet.* | *Feet.* |
| Sacketts Harbor and Charlotte | 88 | 76 | 22 | 0.061 | 0.37 | 0.18 |
| Port Colborne and Cleveland | 158 | 141 | 37 | 0.239 | 0.46 | 0.11 |
| Port Austin and Milwaukee | 259 | 176 | 20 | 0.137 | 0.39 | 0.09 |
| Escanaba and Milwaukee | 192 | 186 | 20 | 0.161 | 0.43 | 0.06 |
| Mean | | | | | 0.41 | |
| Weighted mean | | | | | 0.42 | ±0.044 |

## IS THE LAND TILTING?

With the numerical results of the investigation before us we may now recur to the main subject and ask whether the evidence warrants the conclusion that a general, gradual tilting of the basin is in progress. In the discussion of the data used in comparing the several pairs of stations it has been found that, taken at their face value, they indicate a tilting in the hypothetic direction, but it has also been found impossible to resolve all doubts as to the stability of the gages and benches and the accuracy of the measurements. By reason of these doubts the result from no single pair of stations is conclusive, but when assembled they exhibit a harmony which argues strongly for their validity. As tabulated, there are four results, but these are not all independent, since observations and measurements at Milwaukee are used twice. There are, however, three results wholly independent and a fourth partly independent. To these may be added a fifth partly independent, namely, the determination of change between Port Colborne and Cleveland for the shorter period, 1872–1895. Not only do all these results indicate a change of the same sort, but they agree fairly well as to quantity. The computed change for 100 miles in a century ranges only from 0.37 to 0.46 foot, and the greatest deviation of an individual result from the mean of four is 12 per cent. This measure of harmony appeals strongly to the judgment, and is also susceptible of approximate numerical expression. If the four determinations tabulated in the sixth column are, in fact, measures of the same quantity—that is, if the tilting has been uniform throughout, as we have assumed—then the probable error of the determined value of that quantity (0.42 foot) is less than ±0.05 foot.

The most important factors tending to throw doubt on the conclusion are the possibilities of accidental change in the various benches to which the measurements are referred. The bench at Port Austin, being a mark on bed rock, is trustworthy, and the agreement between the three benches used at Escanaba is good evidence of their stability; but the bench at Milwaukee, with which both are compared, is a pier of a bridge in daily use and may, perhaps, be slowly settling. If it is settling, the comparisons with benches at Escanaba and Port Austin may merely reveal that fact and not measure the subsidence of the land. The fact that the swing bridge on the pier has not required re-leveling is certainly favorable to the stability of the pier, especially when it is considered that a change of fully 1½ inches is to be accounted for; and there is further confirmation in the discovery of a discrepancy between Milwaukee and Escanaba by Captain Marshall, whose data are probably independent of the check mark. Of the benches on Lake Erie, the one at Port Colborne is satisfactory, but those at Cleveland may have settled at critical times, and if so their change would influence the result in the direction found. Of the benches on Lake Ontario, the one at Charlotte is eminently stable; the only practical question affects the bench at Sacketts Harbor, which is on a building that has not been wholly stable since its construction, although presumably so since the making of the bench. If the building at Sacketts Harbor settled between 1874 and 1896, the effect of the lowered bench was to produce, not such a change as appears from the measurements, but one with the opposite sign.

It seems to me that the harmony of the measurements and their agreement with prediction from geologic data make so strong a case for the hypothesis of tilting that it should be accepted as a fact, despite the doubts concerning the stability of the gages.

## RATE OF MOVEMENT.

The deduced mean rate of change—0.42 foot to the 100 miles in a century—depends on assumptions which are convenient rather than probable. These are: (1) that the whole region moves together as a unit, being tilted without internal warping, and (2) that the direction of its present tilting is identical with the direction of the total change since the epoch of the Nipissing outlet of the upper lakes. What we know of the general character of earth movements gives no warrant for such assumptions of uniformity, but no better assumptions as to this region are now available. Under the law of probabilities, the close agreement of four measurements, three of which are wholly independent, gives a good status to their mean, but there are other considerations tending to weaken this status. The probable errors of the individual measurements are rather high, ranging from 14 to 50 per cent, and this suggests the possibility that the closeness of their correspondence may be accidental. It should be remembered also that at two or three stations there was reason to believe that the gage zeros were settling during the period in which the observations were made, and the results involve the doubtful assumption that the rate of settling was uniform. There is room for doubt as to the precision of the instrumental leveling; in only a few instances is the fact of duplicate measurements recorded, and single measurements are notoriously insecure. Error was doubtless admitted by ignoring the effects of barometric gradient. River floods may have introduced errors. In the absence of flood records the records of rainfall at Rochester (near Charlotte), Cleveland, and Milwaukee were compared with the gage readings, the results showing only that if flood errors are involved they must be small. There may also be personal equations of observers, especially as the gages at pairs of stations were not in every case of the same type. For all these reasons I am disposed to ascribe only a low order of precision to the deduced rate of change, and regard it as indicating the order of magnitude rather than the actual magnitude of the differential movement.

The rate of change indicated by Stuntz's observations is more rapid. As already quoted, he states that at a time when Lake Superior was exceptionally low at its outlet, it was nevertheless so high at its western extremity as to obliterate from the St. Louis River a rapid which had been visible only a few years before. This statement involves no definite measures, but it implies that the change within the memory of individuals involves feet rather than the inches deduced from the studies in the other lakes. Similar inferences may be drawn from his statement as to submerged stumps. The recorded range of water level in Lake Superior is about 5 feet, and trees would grow little if any below high-water mark. If, then, with low stage at the east end, stumps are submerged at the west, a change of 5 feet or more would seem to have occurred during the period covered by the growth of a tree and the survival of its stump. The differences between the inferences drawn from this evidence and the result based on gage readings on the other lakes is so wide as to suggest the possibility of error in the Lake Superior observations. It is certainly important that they be verified. Unfortunately I have not been able to visit the region, and the gage records accessible to me are not so connected with bench marks as to give a satisfactory basis for computation. The United States Lake Survey made observations of lake level at Superior City from 1859 to 1871, and then transferred the station to Duluth, where it was continued for two or three years. No bench mark at Duluth is described, and the only recorded bench mark at Superior City is upon a wooden structure, Johnson & Alexander's sawmill. If this bench survives, a good test could be made by renewing the gage station at Superior City. At the other end of the lake, at Sault Ste. Marie, there are authentic benches dating from 1855.

If we assume that the rate of 0.42 foot per 100 miles per century is uniform and secular, and project it backward to the time when the drainage of Lake Huron was shifted from North Bay to Port Huron, we obtain for the period since that change about 10,000 years. From studies at Niagara, Taylor has estimated the same period as between 5,000 and 10,000 years;[1] and the comparison indicates that the rate of modern change is of such magnitude as to accord well with the idea that it merely continues the geologic change.

It is to be hoped that eventually a better measure of the rate of tilting and a surer indication of its direction may be obtained, but even with present knowledge there is interest and profit in considering the economic and geographic consequences of the tilting.

## GEOGRAPHIC CHANGES RESULTING FROM THE MOVEMENT.

Assuming that the general result of this investigation is substantially correct—that the whole lake region is being lifted on one side or depressed on the other, so that its plane is bodily canted toward the south-southwest, and that the rate of change is such that the two ends of a line 100 miles long and lying in a south-southwest direction are relatively displaced four-tenths of a foot in 100 years—certain general consequences may be stated. The waters of each lake are gradually rising on the southern and western shores or falling on the northern and eastern shores, or both. This change is not directly obvious, because masked by temporary changes due to inequalities of rainfall and evaporation and various other causes, but it affects the mean height of the lake surface. In Lake Ontario the water is advancing on all shores, the rate at any place being proportional to its distance from the isobase through the outlet (AA, fig. 100, p. 640). At Hamilton and Port Dalhousie it amounts to 6 inches in a century. The water also advances on all shores of Lake Erie, most rapidly at Toledo and Sandusky, where the change is 8 or 9 inches a century. All about Lake Huron the water is falling, most rapidly at the north and northeast, where the distance from the Port Huron isobase (CC, fig. 100) is greatest; at Mackinac the rate is 6 inches, and at the mouth of French River 10 inches, a century. On Lake Superior the isobase of the outlet (DD, fig. 100) cuts the shore at the international boundary; the water is advancing on the American shore and sinking on the Canadian. At Duluth the advance is 6 inches, and at Heron Bay the recession is 5 inches, a century. The shores of Lake Michigan are divided by the Port Huron isobase. North of Oconto and Manistee the water is falling; south of those places it is rising, the rate at Milwaukee being 5 or 6 inches a century, and at Chicago 9 or 10 inches. Eventually, unless a dam is erected to prevent, Lake Michigan will again overflow to the Illinois River, its discharge occupying the channel

[1] Bull. Geol. Soc. America, Vol. IX, 1898, p. 83.

carved by the outlet of a Pleistocene glacial lake. The summit in that channel is now 8 feet above the mean level of the lake, and the time before it will be overtopped (under the stated assumption as to rate of tilting) may be computed. Evidently the first water to overflow will be that of some high stage of the lake, and the discharge may at first be intermittent. Such high-water discharge will occur in 500 or 600 years. For the mean lake stage such discharge will begin in about 1,000 years, and after 1,500 years there will be no interruption. In about 2,000 years the Illinois River and the Niagara will carry equal portions of the surplus water of the Great Lakes. In 2,500 years the discharge of the Niagara will be intermittent, failing at low stages of the lake, and in 3,500 years there will be no Niagara. The basin of

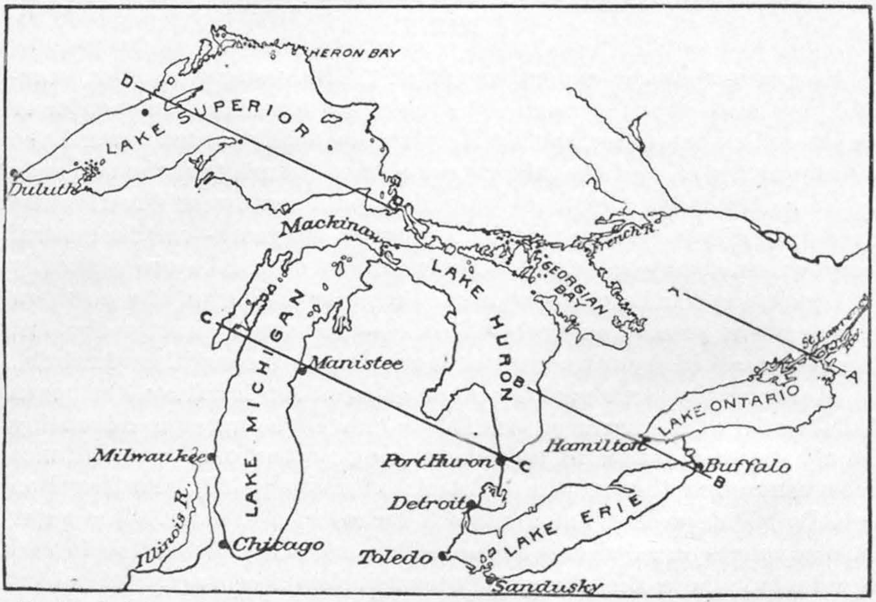



FIG. 100.—Relations of the shores of the Great Lakes to the isobases drawn through their outlets.

Lake Erie will then be tributary to Lake Huron, the current being reversed in the Detroit and St. Clair channels.

The most numerous economic bearings of this geographic change pertain to engineering works, especially for the preservation of harbors and regulation of water levels. But the modifications thus produced are so slow as compared to the growing demands of commerce for depth of water that they may have small importance. It is a matter of greater moment that cities and towns built on lowlands about Lakes Ontario, Erie, Michigan, and Superior will sooner or later feel the encroachment of the advancing water, and it is peculiarly unfortunate that Chicago, the largest city on the lakes, stands on a sinking plain that is now but little above the high-water level of Lake Michigan.

## PLANS FOR PRECISE MEASUREMENT.

While it is believed that the general fact of earth movement has been established by the present investigation, the measurement of its rate and the determination of its direction fall far short of the precision which is desirable. For the purposes of science the order of magnitude of the change is more important than its precise measurement, but there are involved great economic interests, and these demand more definite information. The account of the present investigation is therefore supplemented by an outline plan of the more elaborate investigation which appears necessary to give measurements of the precision that is desirable.

Existing data are neither full enough nor exact enough to give satisfactory measures of the small quantities sought. Doubtless a more elaborate discussion would yield better results than I have obtained, but the improvement could not be great. Observations by the Lake Survey were conducted for purposes not demanding a high order of precision, and high refinement was not attempted. The supplementary work done in 1896 attempted only to be good enough for use in combination with the work of 1874 and 1876, and can not serve as the first term of a new comparison. The problem requires a new set of high-grade observations at each station of a carefully planned system, to be followed, after an interval of at least a decade, by a second set of observations at the same stations.

Foreseeing no opportunity to undertake such a work myself, I have formulated in the following paragraphs a plan embodying the results of my experience—a plan intended to afford useful suggestions to some investigator by whom the work may be actually undertaken.

*Selection of stations.*—To measure the rate of change in any given direction, observations at two stations suffice; but to determine also the direction of change, it is necessary to use three stations grouped in the form of a triangle. The longer the sides of the triangle the better the measurement of rate, and the larger its smallest angle the better the determination of direction. A brief inspection shows that the shores of Lake Michigan and Lake Huron give the best opportunity for the planning of a well-conditioned triangle. Though the narrowness of their connecting strait has led to the giving of separate names, they are really a single lake, and the stretch of their water surface is in every direction greater than that of Lake Superior.

For the purpose in view the point of first importance is the outlet of the lake at Port Huron. This is peculiar in that the plane of mean water level has here a constant relation to the adjacent land, a relation altogether independent of the progressive deformation of the basin. This station should not be on the St. Clair River, but on the shore of the lake near by.

The second point of vantage is Chicago. As economic interests are more seriously affected by the geographic change at that point than elsewhere, it is desirable to determine directly, by comparison with Port Huron, the rate at which the lake is encroaching on the land.

A third point of prime importance is the Strait of Mackinac. Although the equilibrium levels of the surfaces of the two lakes are the same, there are considerable periods when their equilibrium is disturbed, and during such periods a current flows in one direction or the other through the strait. Only when this current is nil is the whole water body in perfect equilibrium, and it is essential to precise leveling through the water surface either that times of equilibrium be chosen or

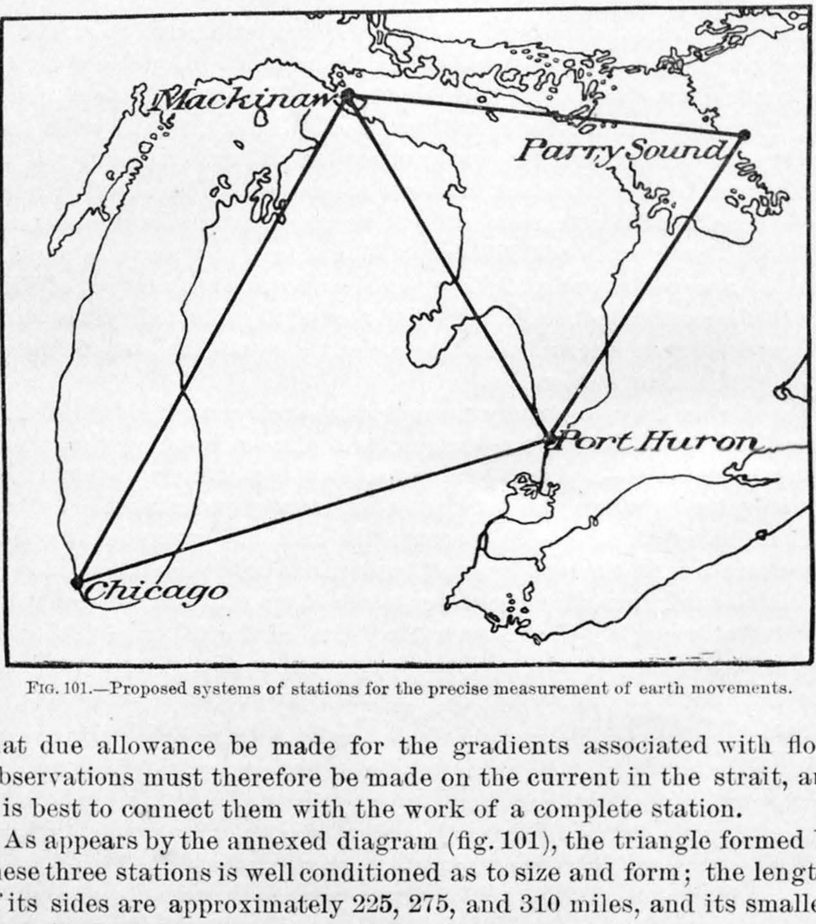



FIG. 101.—Proposed systems of stations for the precise measurement of earth movements.

that due allowance be made for the gradients associated with flow. Observations must therefore be made on the current in the strait, and it is best to connect them with the work of a complete station.

As appears by the annexed diagram (fig. 101), the triangle formed by these three stations is well conditioned as to size and form; the lengths of its sides are approximately 225, 275, and 310 miles, and its smallest angle is about 45 degrees.

While the proper use of these three stations will give answer to the questions of greatest economic and scientific importance, there will be material scientific advantage in adding a fourth station to the system. It should be placed somewhere on the north shore of Georgian Bay, and, giving consideration to accessibility as well as geographic position, it is probable that Parry Sound should be selected. By adding this station another well-conditioned triangle would be completed, and there would result an additional determination of the rate and direction of tilting. If rate and direction vary from place to place the fact will probably be brought out. There would be additional advantage in the fact that Parry Sound and Chicago are separated by the greatest practicable distance in the direction of maximum change, so that a comparatively short period of time might afford a valuable measurement. The approximate results of the present investigation indicate that the change in the relative height of Parry Sound and Chicago in ten years would be about 2 inches.

*Conditions controlling equipment.*—In order to plan intelligently the system of observations, full consideration should be given to the conditions affecting the problem, and provision should be made for all possible sources of error. Prominent among these are the various factors which modify the water level at points on the lake shore. Such factors have been considered in the preceding discussion of gage data, but they are assembled here in a more systematic way.

The lake continually receives water from streams and from rain, and continually parts with water by discharge at its outlet and by evaporation. In the long run gain and loss are equal, but for short periods they are usually unequal; so that from day to day, from season to season, and from year to year the volume of the lake and the consequent mean level of its surface are continually changing.

In bays and estuaries there are local temporary variations occasioned by the floods of tributary streams.

There are solar and lunar tides, small as compared to those of the ocean, but not so small that they may be neglected.

The wind pushes the lake water before it, piling it up on lee shores and lowering the level on weather shores. During great storms these changes have a magnitude of several feet, and the effect of light wind is distinctly appreciable. Even the land and sea breezes, set up near the shore by contrasts of surface temperature, have been found to produce measurable effects on the water level.

There is also an influence from atmospheric pressure. When the air is in equilibrium, if that ever occurs, the pressure is the same on all parts of the lake surface, and the equilibrium of the lake is not disturbed; but when the air pressure varies from point to point this variation of pressure is a factor in the equilibrium of the water surface, the surface being comparatively depressed where the air pressure is greater and elevated where it is less.

When a storm wind ceases, the water not merely flows back to its normal position but is carried by momentum beyond, and an oscillation is thus set up which continues for an indefinite period. A similar oscillation is started whenever the equilibrium is disturbed by differences of atmospheric pressure; and these swaying motions, called seiches, analogous to the swaying of water in a tub or hand basin, persist for long periods. In fact, they bridge over the intervals from impulse to impulse, so that the water of the Great Lakes never comes to rest.

Every disturbance which causes the water to rise on one shore of the lake and fall on the other interferes with the equilibrium between the two lakes at the Mackinac Strait. If a strong wind blows the water eastward, raising the level on the east shores of the lakes and lowering it on the west shores, there is high water at the west end of the strait and low water at the east, producing a current toward the east; and when the wind ceases the water that has poured from Lake Michigan to Lake Huron must return, producing a current in the opposite direction. Theoretically, analogous effects should be produced by tides and barometric gradients, and there can be little question of their detection if the phenomena at the strait are studied.

These various influences work independently but simultaneously, and their effects are blended in the actual oscillations of the water surface at any point. In using the water surface for the purpose of precise leveling, it is necessary to take account of all such factors and make provision for the avoidance or correction of the errors they tend to produce.

*Equipment.*—In view of the complexity of the phenomena to be analyzed, it is desirable that most of the instruments employed be of the automatic kind, giving continuous record. While such instruments accomplish much more than could be done by an observer alone, they do not dispense with his services. They are complex as compared to the apparatus for personal observation, and can be successfully employed only by a man of scientific training. The first essential, therefore, at each of the stations is an expert observer.

The gage employed for the determination of water height should be of some automatic type, giving a continuous record. This is necessary in order that the study of the record may furnish data for the complete elimination of errors from tides, seiches, and land and sea breezes. The gage should be protected, not only from the direct shock of waves, but from all secondary agitation of the water due to wave shock. It should be so installed as to be secure from settling. The height of its zero should be readily verifiable.

Near each station there should be at least three benches, constructed with special reference to permanence and stability. They should be independent of one another and independent of other structures.

Pressure of the air should be continuously recorded by a barograph, carefully standardized. A wind vane and anemometer should give automatic records.

At Mackinac there should also be means for securing a record of the direction and velocity of water currents.

*Treatment of observations.*—Stress having already been laid on the importance of putting the work in expert hands, it would be unwise to attempt the formulation of a code of instructions for either the making or the reduction of observations; but there may be advantage in a few suggestions based on experience acquired in the present investigation.

While it is doubtless possible to deduce from a study of currents in Mackinac Strait a theory of the relation of those currents to the equilibrium conditions of the two lakes, it will probably be found best to use the current observations chiefly for the discrimination of favorable and unfavorable times, and to compare the lake-level observations only for times when the current at Mackinac is gentle. It will also be better to avoid the use of observations during the prevalence of strong winds or high barometric gradients than to attempt the application of corrections for those factors.

The times not barred by high winds, high gradients, and currents will ordinarily not be found to have such duration and distribution that tidal effects can be eliminated by including complete tidal cycles. It will therefore be necessary to discuss the solar and lunar tides for each station and prepare tables of correction to be applied to all observations employed. The same treatment will be necessary for the effects of land and sea breezes. Barometric gradient of amount too great to be ignored nearly always exists, and this, as determined by observations at the stations themselves, should be the subject of computation and correction.

Seiches should be fully discussed for each station, and the observations finally used should be grouped in periods of sufficient length to eliminate the seiche effect.

## SUPPLEMENT.—INVESTIGATION BY MR. MOSELEY.

The main body of manuscript for this paper was prepared in June and July, 1897. An abstract was communicated in August to the American Association for the Advancement of Science, meeting in Detroit, and a fuller abstract was printed in the September number of the National Geographic Magazine.[1] As a result of this publication I became acquainted with a cognate investigation by E. L. Moseley, of Sandusky, Ohio. His data and results were communicated to the Ohio Academy of Sciences in December, 1897, and printed soon afterwards in a Sandusky newspaper. They received more permanent as well as fuller presentation in an article contributed to the Lakeside Magazine,[2] and this article reaches me while the proof sheets of the present paper are in hand. As will be readily understood from the following abstract, the data he has gathered constitute an important contribution to the subject.

[1] Modification of the Great Lakes by earth movement: Nat. Geog. Mag., September, 1897, Vol. VIII, pp. 233-247.

[2] Lake Erie enlarging; the islands separated from the mainland in recent times; by E. L. Moseley: The Lakeside Magazine, Lakeside, Ohio, April, 1898, Vol. I, pp. 14-17.

North of Sandusky Bay, near the west end of Lake Erie, is a cluster of islands, of which the five largest are each several miles in extent.[1] About them the water is shallow, and if the lake were lowered 30 to 35 feet they would all be connected with the mainland. On these islands grow many species of wild plants, and the origin of this flora is related to the geologic history of the islands. There was a time during the ice retreat when the whole basin was covered by a glacial lake. If the water was gradually lowered from the plane of the glacial lake to the present plane of Lake Erie, the islands were at first barren and were eventually occupied only by such plants as were in some way conveyed across the intervening straits, from 2½ to 3 miles wide. As Moseley points out, there are many modes of such adventitious introduction, but they could not be expected to give to the islands a flora so varied as that of the adjacent mainland.

If, on the other hand, as inferred from the slopes of the old shore lines and other data, the attitude of the land was different when the glacial lake was drained away, the original Lake Erie occupied only the eastern part of the Erie basin, and the western part, including the district of the islands, was dry land. Subsequently, from the tilting of the land, the lake waters advanced westward so as to flood the straits and convert the lowland hills into the present islands. In connection with such a geologic history the islands would have acquired their flora at the same time with the mainland, and should now present the same variety of species, so far as local conditions permit. Moseley has carefully compared the insular flora with that of the mainland, and finds that the only mainland species which do not occur on the islands are such as do not find there a congenial soil. The botanic evidence thus supports the geologic, and verifies the conclusion that the land has been tilted toward the southwest since the birth of Lake Erie.

The islands are composed largely of limestone and are surrounded by limestone cliffs. In South Bass or Put-in Bay Island there are caves opening at the water's edge and partly occupied by lake water. Exploring these, Moseley finds stalactites extending from the roof down into the water, and stalagmites lying 3 or 4 feet below the present surface of the lake. Comparing the present water level with the lowest levels known in recent times, it appears that these stalagmites have not been above water during the present century, and as stalagmites are formed only in the air, it is clear that the lake has encroached on the land since they were made.

These data show only that a change has occurred, and ascribe no date, but other phenomena observed in the neighborhood of Sandusky indicate clearly that change is now in progress. A tract of land on which hay was made in 1828 is now permanently under water. A tract of land one-half mile square, surveyed in 1809, has since become marsh, with water and mud 12 to 18 inches deep. Various parts of Sandusky

[1] For the relation of these islands to the lake and the isobase of its outlet, see fig. 100, p. 640.

Bay where rushes grew within the memory of men still living are now covered with open water. "By the high water that prevailed in 1858 to 1860 large trees were killed in many places where the waves could not reach them." "Hundreds of walnut stumps are still standing on the border of the marshes east of Sandusky where even now, although the water is lower than usual, it is too wet for walnut trees to grow. One that stood recently on ground only 6 inches above the present lake level measured 5 feet 4 inches in diameter. We may infer from this that during the life of this tree, probably over 300 years, the water was not so high as in the present century." Many stumps and prostrate trunks with roots and branches attached are found from 1 to 4 feet below the present lake level, and in one locality it is inferred that the lake during the life of the trees must have been as much as 8 feet lower than "during much of the time for the last forty years."

These various facts, and others of the same tenor enumerated by Moseley, are in complete accord with the qualitative results derived from the discussion of gage readings, but, like the data gathered by Stuntz, they suggest a more rapid rate of change than do those results.

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