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SHORTER CONTRIBUTIONS TO GENERAL GEOLOGY, 1930

LITHOLOGIC STUDIES OF FINE-GRAINED UPPER CRETACEOUS SEDIMENTARY ROCKS OF THE BLACK HILLS REGION

By WILLIAM W. RUBEY

ABSTRACT

More than nine-tenths of the Upper Cretaceous rocks in northeastern Wyoming are fine-grained shales, mudstones, and calcareous marls. A comparative study of the mineralogy, chemical and mechanical composition, density and porosity, fissility, and lamination of samples of these rocks discloses several relations that throw light on the geologic history and structural deformation of the region, and perhaps on its oil and gas possibilities.

Microscopic examination of thin sections and rock powders shows that platy crystals of a clay mineral (probably beidellite) form the chief constituent of the argillaceous rocks. Most of these platy crystals lie essentially parallel to the bedding and give the rocks an aggregate optical orientation. Calcium carbonate occurs in varying amounts in many of the samples as shells, crystals, microcrystalline lenses, and spherulites. Opaque organic matter, fine quartz sand or silt, and small nodules of pyrite are minor constituents of most of the samples. A pyritic limestone 4 feet thick from the Greenhorn formation consists of calcite, pyrite (and either pyrrhotite or ferrous sulphide), and gypsum, with smaller amounts of iron oxide, organic matter, and bone phosphate. The gypsum in this limestone is secondary, but the sulphide, which makes up one-fourth of the rock, is probably almost, if not quite, syngenetic. Field relations indicate that the sedimentation by which this pyritic limestone was formed was not unusually slow.

Partial chemical analyses indicate that the black shales contain less organic matter than the light-colored calcareous marls and that there is less chloroform-soluble bitumen in the formations of Benton age than in the overlying Niobrara and Pierre formations. The proportion of chloroform-soluble bitumen varies also, with either the dip or the porosity of the rocks; it is not possible to say which. This indicated relationship to the dip of the rocks is of economic interest, for it suggests the possibility that deformation has converted organic matter into petroleum, and that undrilled areas of steep dips and faulting may contain oil. In general, the percentages of calcium carbonate, organic matter, and pyrite vary together; the older formations contain least, the Pierre shale more, and the Niobrara and Greenhorn formations most. A consideration of the conditions favorable for the formation and preservation of these three constituents and of the abundance of remains of bottom-living mollusks in rocks near one end of this series and the absence of fossils in rocks near the other end, together with other facts, leads to the conclusion that the more calcareous, organic, and pyritic sediments probably accumulated more rapidly and in shallower water than the others. In fact, the entire series of sediments seems to be very similar to the blue muds accumulating near present coasts.

In choosing a method for determining the mechanical composition of these fine-grained rocks, many methods used in other sciences were reviewed. A mechanical analysis consists of three parts—preparation of the samples, measurement, and presentation of results. The attainable accuracy of each step in all parts of the analysis, as well as the uncertain relation of the present to the original size of the particles, must be considered in choosing the particular methods best adapted for the purpose in view. Before most fine-grained sedimentary rocks can be mechanically analyzed they must be disintegrated by chemical treatment and mechanical agitation. In this paper different methods of preparation are discussed, and a long period of soaking in water is suggested as an additional treatment. The samples may be measured by sieving, microscopic counts, or settling in a fluid. The methods that depend upon the different settling velocities of large and small particles include elutriation, decantation, increasing weight of fallen sediment, and decreasing density or turbidity of the suspension. The merits and demerits of many modifications of the different methods are considered, and the conclusion is reached that some methods give much more information in less time than others. Pyramidal diagrams and cumulative curves, plotted logarithmically, seem to present the results of measurement most satisfactorily; but the conversion of the data into these diagrams presents a different problem for each method of measurement. A simple graphic solution saves much work in converting the data obtained by many settling or "sedimentation" methods.

As a result of this review, rock samples and insoluble residues of some calcareous marls were disintegrated by two months' soaking in slightly ammoniacal water, with occasional shaking and rubbing. Organic matter retarded disintegration and dissolved slowly. The samples were measured by frequent weighing of the sediment that accumulated on a pan which hung in a thoroughly mixed suspension. An attempt to use an aluminum pan proved unsuccessful, and a platinum pan was substituted. The unexpectedly high porosity of the settled sediment caused some difficulties. It was found that the weight of sediment increased as the logarithm of the settling time and that sufficiently accurate estimates of the final weight could be made from a series of successive determinations of weight made in less than one hour. Somewhat less sediment accumulated on the pan than was expected from the size of the original sample, but investigation indicated that this loss was due chiefly to solution of organic matter, interstitial water, and fine particles, and to slumping of sediment off the edges of the pan. The successive determinations of weight and settling time were corrected for admixed small particles and the temperature of the water and thus converted into percentages and settling velocities. The exact relation between settling velocities and diameters of irregular particles is unknown, and examination of samples pipetted off during set-

tling suggested that the flat shape and large surface area of the clay minerals may be the chief source of this uncertainty. Similarly, the size terms "sand," "silt," and "clay" are not defined by diameter limits that are universally accepted. Therefore, the settling velocities were not recomputed into theoretical diameters but were plotted directly alongside the corresponding size terms, which can be correlated arbitrarily with settling velocities by a study of published measurements. Several methods of checking indicated that the final graphs are accurate within about 2 to 5 per cent. These graphs show that the samples consist almost entirely of grains smaller than very fine sand and that they are essentially unsorted mixtures of silt and clay. If this lack of sorting is due to flocculation of the particles at the time of deposition, it may be a criterion of origin in saline or marine water. The grains of calcium carbonate in the more calcareous samples seem to be better sorted than the associated detrital material.

The density and porosity of these fine-grained rocks were found to vary greatly with the temperature to which the samples had been heated. Below 200° C. this change is due almost entirely to loss of water, and consequently all determinations were made on samples previously heated to this temperature. The grain or mineral density varies with the impurities in the rock and with the size of the grains. The lump or rock density is about 2.0 or less, indicating that the positive anomalies found at some gravity stations in the Black Hills region are not caused by unusually heavy rocks near the surface. The porosity averages about 30 per cent, and it decreases apparently by quantitative relationships with increase in the dip of the beds and in the original depth of burial. The size of grain seems to be another factor affecting porosity, the finer-grained samples having been most compacted. The relation of porosity to dip and to proximity of faults is significant, for it indicates that these rocks were deformed internally by horizontal compression, that their apparent stratigraphic thickness depends upon the degree of folding, and that large volumes of water were squeezed out of them during deformation.

Field evidence, such as the greater abundance of fissile shales in stratigraphically lower beds and the slight discordances between fissility and bedding noted here and there, suggests that the fissility or shaly structure is secondary. Microscopic examination and chemical and mechanical analyses neither confirm nor refute this suggestion. Some microscopic cracks that may represent planes of fissility are slightly inclined to the bedding, but the aggregate optical orientation (probably incipient fissility) is essentially parallel to the bedding. In the samples studied fissility varied inversely with the content of calcium carbonate but was unrelated to the size of the grains.

Many of the samples exhibit more or less distinct laminations of several types that may be annual layers. The climate and other physical conditions were probably favorable for the formation and preservation of annual layers in these rocks. The average thickness of pairs of the laminations is roughly the same as the observed thickness of annual layers deposited elsewhere and the expected thickness of annual layers in these rocks as estimated by several independent methods. More detailed comparison with the probable rate of transgression of the Upper Cretaceous sea and with the supposed length of Upper Cretaceous time suggests that each pair of laminations may represent several years' deposits, but these discrepancies may be the result of many undetected disconformities or diastems in the stratigraphic sequence. The distinctness and thickness of the laminations in the different samples indicate that the shales accumulated slowly in rather deep water and that the sandier rocks and calcareous marls accumulated more rapidly and in shallower water. The discussion of these bedding laminations is summarized more fully in the report (pp. 52–53).

Three of the broader conclusions are based upon scattered evidence. The chemical composition, fossil content, bedding laminations, lack of sorting, and comparison with present-day blue muds indicate that, in general, the finer-grained noncalcareous shales accumulated more slowly and in deeper water than the calcareous, organic, and pyritic rocks. The variations in the percentage of pore space and soluble bitumens, in the fissility and aggregate optical orientation, and in the thickness of formations indicate that these fine-grained rocks lost volume and were deformed internally by loading and tilting. The percentage of soluble bitumens, probably a measure of the oil content, seems to vary with the tilt or shear to which the rocks have been subjected but not with the depth or load. Although this relation may be fortuitous, it suggests the possibility that folding may convert the organic matter in fine-grained sedimentary rocks into oil and then force it into adjacent sandstone beds.

INTRODUCTION

Fine-grained argillaceous rocks constitute the most abundant type of sedimentary rocks, and their peculiar properties make them of special interest in geologic studies. The conditions of their deposition are very different from those of the coarser-grained sandstones; they yield to deformation by compaction and plastic distortion in a manner quite unlike the coarser rocks; and, to mention only one example of their economic importance, they are commonly thought to be the source beds of petroleum. Consequently, from both the scientific and the practical points of view, they deserve careful study. Yet, though they have been widely studied and the literature describing them is voluminous, they probably are the least understood of the sedimentary rocks. This perplexity is due in part to the difficulty of studying them microscopically, owing to their incoherence and the small size of their constituent particles, and in part to their extremely diverse chemical composition. However, recent advances in microscopic technique and in the mineralogy of clays and improved methods of chemical and physical examination give promise of placing the study of argillaceous rocks on a much more satisfactory basis.

More than nine-tenths of the Upper Cretaceous rocks exposed on the northwest flank of the Black Hills in Wyoming and Montana are shales, mudstones, and calcareous marls. Consequently, in a somewhat detailed study [1] of these Upper Cretaceous rocks, with special reference to their geologic history, types of structural deformation, and oil and gas possibilities,

[1] Rubey, W. W., Origin of the siliceous Mowry shale of the Black Hills region: U. S. Geol. Survey Prof. Paper 154, pp. 153–170, 1929; Cretaceous and Cenozoic formations on the northwest flank of the Black Hills: U. S. Geol. Survey Prof. Paper — [in preparation]; The oil and gas possibilities of the Black Hills rim in Wyoming and Montana: U. S. Geol. Survey Bull. — [in preparation].

it became necessary to devote particular attention to these fine-grained sedimentary rocks.

The present report gives the results of a comparative study of some of the outstanding lithologic characteristics of a group of samples chosen as representative of the rock types in the different formations of this region. Except for the theoretical discussion of the bedding laminations, this paper is essentially an empirical or descriptive study of the lime-clay series of sedimentary rocks and an attempt to interpret the observed relations between the characteristics of the different rock types. The conclusions reached are merely tentative. The relationships pointed out and the explanations offered could be definitely substantiated only by a much more detailed study of the rocks within this area and by observations extending over a much wider area. However, if any part of the paper suggests fruitful problems in this or other areas or proves helpful to others who are studying similar rocks, it will have accomplished its chief purpose.

The writer is indebted to his colleagues, most of them members of the United States Geological Survey, for many helpful criticisms and suggestions in the preparation of this report. W. H. Bradley contributed to all parts of the paper by his generous discussions; C. S. Ross gave advice on the mineralogy of the rocks; the Bureau of Mines kindly made the determinations of organic matter and Taisia Stadnichenko offered valuable suggestions in their interpretation; C. K. Wentworth reviewed the section on mechanical analyses; P. G. Nutting determined the densities and porosities and helped in their interpretation; and Adolph Knopf and J. B. Reeside, jr., criticized the discussion of bedding laminations.

ROCKS STUDIED

The stratigraphic relations and areal extent of the Upper Cretaceous rocks that crop out for about 150 miles along the northwest flank of the Black Hills uplift are described in another report.[2] Their general characteristics and thickness, together with the stratigraphic names adopted in that report, are summarized in the following table:

[2] Rubey, W. W., Cretaceous and Cenozoic formations on the northwest flank of the Black Hills: U. S. Geol. Survey Prof. Paper — [in preparation].

Upper Cretaceous formations in northeastern Wyoming and southeastern Montana

Series	Group	Formation and member		Thickness (feet)	General characteristics
Eocene.		Wasatch formation.		100+	Continental deposits.
		Fort Union formation.	Tongue River member.	650–1, 050	
			Lebo shale member.		
Eocene (?).		Lance formation.	Tullock member.	1, 000–2, 250	
			Hell Creek member.		
Upper Cretaceous.	Montana.	Fox Hills sandstone.		150–250	Brownish sandy shale and siltstone with beds of sandstone and ferruginous concretions. Marine fossils. Gradational into underlying Pierre shale. Forms a prominent grassy scarp.
		Pierre shale.		150–250	Dark-gray fissile shale and mudstone with calcareous concretions. Locally contains light-buff sandy shale. Marine fossils.
			Monument Hill bentonitic member.	150±	Impure bentonite and siltstone. Some calcareous and barite concretions. Marine fossils. Commonly forms a scarp. Named from exposures at Monument Hill, sec. 32, T. 56 N., R. 68 W., Crook County, Wyo.
				500–800	Dark mudstone and shale with abundant calcareous concretions. Light gray in upper part; iron-stained in lower part. Marine fossils.
			Mitten black shale member.	150–200	Blue-black fissile shale with few iron-stained calcareous concretions. Marine fossils. Forms a prominent scarp. Named from exposures along Mitten Prong in T. 56 N., R. 68 W., Crook County, Wyo.

Upper Cretaceous formations in northeastern Wyoming and southeastern Montana—Continued

Series	Group	Formation	Member	Thickness (feet)	General characteristics
Upper Cretaceous.	Montana.	Pierre shale.	Gammon ferruginous member. Groat sandstone bed.	800–1, 000	Abundant iron-stained concretions and thin beds of siderite in light-gray mudstone and shale. Fossils scarce but consist of marine species. Commonly forms bare buttes. Named from exposures along Gammon Creek in T. 57 N., Rs. 67 and 68 W., Crook County, Wyo. Groat sandstone bed near top of member (150 feet thick in northern part of area) consists of ferruginous and glauconitic sandstone and siltstone. Named from exposures along Groat Creek in T. 7 S., R. 56 E., Carter County, Mont. Pedro bentonite bed at base of member (locally 20 feet thick but not widespread) consists of hard white massive clay and tuff. Named from exposures near Pedro, sec. 5, T. 45 N., R. 63 W., Weston County, Wyo. Possible unconformity at or near base of member.
	Colorado.	Niobrara formation.	Beaver Creek chalky member.	125–200	Chalk marl and calcareous siltstone, gray where fresh but weathering to light yellow. Marine fossils. Named from exposures along Beaver Creek in T. 46 N., R. 64 W., Weston County, Wyo.
			Sage Breaks shale member.	250–325	Gray noncalcareous mudstone and shale with many large light-gray calcareous septarian concretions. Fossils scarce but consist of marine species. Commonly forms scarps and buttes. Named from exposures in the Sage Breaks, in T. 46 N., R. 63 W., Weston County, Wyo. Included in Carlile of previous reports.
		Carlile shale.	Turner sandy member.	150–200	More or less sandy shale and siltstone with iron-stained concretions. Persistent thin beds of sandstone, locally conglomeratic and phosphatic and containing abundant shark teeth, in lower part. Marine fossils. Forms a minor scarp. A distinct faunal break and possible unconformity at base. Named from exposures along Turner Creek in Tps. 46 and 47 N., R. 64 W., Weston County, Wyo.
				75–125	Dark-gray shale with a few calcareous concretions. Marine fossils.
		Greenhorn formation.		50–350	Chalk marl, thin-bedded limestone, and light-gray sandy shale with calcareous concretions. Marine fossils. Forms a prominent scarp. Interfingers with underlying Belle Fourche shale. Limestone facies in south; thin concretionary facies in northwest; thick chalk-marl facies in northeast.
		Graneros shale.	Belle Fourche shale member.	350–1, 000	Black fissile shale and mudstone with concretions and bentonite beds in upper half and lowermost part. Upper concretions calcareous; lower ones sideritic. Fossils scarce in lower part, but all are marine species.
			Mowry siliceous shale member.	125–225	Hard siliceous claystone, dark gray where fresh, light silvery gray where weathered. Contains many fish scales. Many thin beds of bentonite. Marine fossils. Forms a scarp or ridge. Gradational into and includes at base a few feet of soft shale ("Nefsy shale member"). Clay Spur bentonite bed (1 to 4 feet thick) at top over entire area. Named from exposures near Clay Spur, sec. 30, T. 47 N., R. 63 W., Weston County, Wyo.

Upper Cretaceous formations in northeastern Wyoming and southeastern Montana—Continued.

Series	Group	Formation and member		Thickness (feet)	General characteristics
Upper Cretaceous.	Colorado.	Graneros shale.	Newcastle sandstone member.	0–75 (commonly 40).	An extremely variable unit of discontinuous beds of sandy shale, sandstone, impure lignite, bentonite, and (where thin) phosphatic nodules. Contains both continental and marine or brackish-water fossils. Local unconformities within member.
			Skull Creek shale member.	175–275	Black fissile shale with few ferruginous concretions. Thin sandy beds in lower part. Fossils scarce, but consist of marine species.
----?----					
Lower Cretaceous.	Inyan Kara.	Fall River[a] sandstone (=so-called Dakota sandstone of previous reports on this area).		150–350	An extremely variable group consisting of discontinuous beds of sandstone, sandy shale, conglomerate, lignite, and variegated siltstone. In general, though not in detail, the higher sandstones are more heavily iron stained and slabby and the lower ones lighter gray and massive. Continental fossils throughout greater part but marine fossils in upper 20 feet. Named from exposures along Inyan Kara Creek in the northeastern part of the Moorcroft quadrangle, Wyo.
		?			
		Fuson formation.			
		?			
		Lakota sandstone.			

[a] Russell, W. L., The origin of artesian pressure: Econ. Geology, vol. 23, pp. 134-136, 1928.

A group of samples (air-dried from three to four years) was selected for chemical and mechanical analyses and determinations of density and porosity. These samples are described below.

A. Black fissile shale about 50 feet above base of Skull Creek member of Graneros shale. Associated with thin sandy beds and fossiliferous ferruginous concretions. Center of sec. 28, T. 50 N., R. 65 W., Crook County, Wyo. Beds dip 5° SW.

B. Dark-gray hard siliceous claystone from uppermost part of Mowry member of Graneros shale. Associated with bentonite beds and contains many Radiolaria. Center of sec. 35, T. 57 N., R. 66 W., Crook County, Wyo. Beds dip 3° NW.

C. Dark-gray to black shale or mudstone from upper few feet of Belle Fourche member of Graneros shale. W. ½ sec. 34, T. 58 N., R. 62 W., Crook County, Wyo. Beds dip 4° NE.

D. Pyritic limestone from basal bed of Greenhorn formation, a 4-foot bed consisting of 1-inch layers of highly fossiliferous, somewhat sandy limestone. Contains mollusks, ammonites, fish bones, and teeth. Lowermost layer contains the most pyrite. Immediately overlies with sharp contrast shale like sample C. SW. ¼ sec. 11, T. 57 N., R. 62 W., Crook County, Wyo. Beds dip 6° NE.

E. Light-buff to drab calcareous marl about 60 feet below top of Greenhorn formation. Contains many small mollusks. Associated with thin beds of fossiliferous sandy limestone, pyrite nodules, and silty chalkstone concretions. NW. ¼ sec. 14, T. 57 N., R. 62 W., Crook County, Wyo. Beds dip about 1° NE.

F. Dark-gray shale about 15 feet above base of Turner member of Carlile shale. Associated with sandy shale and thin beds of coarse sandstone. W. ½ sec. 19, T. 47 N., R. 64 W., Weston County, Wyoming. Beds dip 33° SW.

G. Gray shale or mudstone from Sage Breaks member of Niobrara formation. Associated with many large calcareous septarian concretions. SW. ¼ sec. 5, T. 45 N., R. 62 W., Weston County, Wyo. Beds dip 50° S.

H. Light-gray calcareous marl (unweathered) from lower part of Beaver Creek member of Niobrara formation. SW. ¼ sec. 30, T. 46 N., R. 63 W., Weston County, Wyo. Beds dip 10° SW. and lie between two small normal faults.

I. Gray siltstone from lower part of Gammon member of Pierre shale. Associated with many thin beds of siderite and iron oxide. Collected near a fossil vertebrate skeleton. S. ½ sec. 32, T. 49 N., R. 66 W., Crook County, Wyo. Beds dip 5° SW. and lie near several normal faults.

J. Dark-brown to black shale probably from Mitten member of Pierre shale. Contains many small fragments of organic matter. SE. ¼ sec. 6, T. 44 N., R. 61 W., Weston County, Wyo. Beds dip 45° SW.

K. Gray mudstone about 100 feet below Monument Hill member of Pierre shale. Contains many small mollusks. Associated with a few calcareous concretions. NW. ¼ sec. 18, T. 44 N., R. 62 W., Weston County, Wyo. Beds dip 7° SW. and lie near plane of a normal fault.

Thin sections of these 11 samples and of 24 other specimens of shale and marl from the same beds in this region were cut, and the textural features, such as fissility and bedding laminations, were studied. The mineralogic examination included a study of these 35 thin sections, of a polished face of sample D, and of the crushed powders of these and about 10 other specimens.

MINERALOGY

The samples of shale, mudstone, marl, and limestone were examined both in thin section and with refractive-index liquids. The index liquids were used on crushed powders and on individual particles that had been disintegrated and sized in water.

SHALE, MUDSTONE, AND MARL

In every specimen examined, except those that were very calcareous, definitely crystalline micaceous clay is the chief constituent, and all the thin sections that were cut perpendicular to the bedding show a pronounced aggregate orientation and positive elongation of these clay crystals parallel to the bedding. Many of these individual crystals do not lie exactly parallel to one another, but in all thin sections their average attitude is sensibly parallel to the bedding. Therefore, as the individual crystals show parallel extinction and as the slow ray vibrates in the plane of elongation, thin sections of the shale cut perpendicular to the bedding all show an aggregate positive elongation very much as they would if they were cut from single crystals.

The clay crystals are brown to gray and, having but one pronounced cleavage, are predominantly platy. Except for this cleavage, nearly all the crystals have rather indefinite outlines. Many grains seem to consist of groups of fibers lying in one plane; some crystals are slightly bent. The crystals range from less than 1 micron to as much as 100 microns in maximum diameter, and they are approximately equidimensional when lying upon their cleavage faces. In cross section, however, their length is commonly about seven times as great as their thickness.

Not all the clay crystals in any one sample have exactly the same refractive index, but the range is slight, and in all the samples studied the refractive index of the dominant clay mineral is between 1.55 and 1.57—in some slightly less, in others slightly more than 1.56. The maximum birefringence, as seen in thin sections cut perpendicular to the elongation, is about 0.02 or 0.03.

The dominant clay mineral seems to be the same in all the samples. Many characters, such as the parallel extinction, positive elongation, and crystal habit, are identical in all the samples examined, and the refractive indices and birefringences are relatively uniform. C. S. Ross, of the United States Geological Survey, identified this mineral as belonging to the beidellite [3] type, or perhaps in the isomorphous series between beidellite and nontronite.[4] He suggests that the rather high refractive indices may be due to a high content of Fe_2O_3 (indicated also by the brown color of the mineral), to adsorbed alkalies, or to loss of adsorbed water. Some of the small grains that show very definite crystal outlines may be anauxite.[5]

All the samples contain some and a few contain much fine quartz sand or silt. These quartz grains are angular or subangular and range from 10 to 100 microns in diameter. Some of the samples contain a few flakes of muscovite and biotite.

All specimens from the Niobrara and Greenhorn formations and some specimens from the upper part of the Pierre shale contain noticeable amounts of relatively pure calcium carbonate (ω=1.66±). It occurs in several forms—(*a*) as shells of globular foraminifers from 50 to 150 microns in diameter; (*b*) as spherulites from 4 to 7 and from 30 to 50 microns in diameter; (*c*) as minute lenses of microcrystalline calcite, 100 to 300 microns long, which lie parallel to the bedding; and (*d*) as sand grains and well-developed crystals from 30 to 100 microns in diameter. Specimens from the lower part of the Pierre shale contain spherulites of siderite from 5 to 50 microns in diameter.

Nearly all the thin sections examined are dark brown with organic matter, which consists chiefly of opaque dark-brown to black structureless fragments but also contains some transparent amber-colored masses of optically inactive material. These fragments and masses range in shape and size from equidimensional grains at the limit of visibility to very thin lenses more than 1 millimeter long that lie parallel to the bedding. A fragment of sample F, from which the mineral constituents were dissolved in hydrofluoric acid, also showed that the organic matter consists of particles of many different sizes.

A few very small nodules of finely crystalline pyrite were noted in most of the samples. Doubtless many other minor constituents are present in these rocks, but the small size of the particles and the lack of complete chemical analyses make their determination very difficult.[6]

PYRITIC LIMESTONE

Sample D, pyritic limestone from the Greenhorn formation, is sufficiently different in lithologic type to merit separate description. The chief constituents of this rock are calcium carbonate, pyrite, and gypsum. (See p. 11.) The calcium carbonate is present as large and small fragments of *Inoceramus* shells and as crystals of relatively pure calcite (ω=1.665±) from 35 to 250 microns in diameter. The pyrite was determined with the reflecting microscope by C. S. Ross; it is minutely crystalline and occurs as isolated nodules, which range in size from the limit of visibility to several millimeters in diameter, and as surface coatings on some of the shell fragments. Some pyrrhotite or possibly ferrous sulphide is associated with the pyrite, for hydrogen sulphide is evolved when the sample is digested in dilute hydrochloric acid. Although usually a high-temperature mineral, pyrrho-

[3] Beidellite, according to Mr. Ross, is the most widespread of the clay minerals.

[4] Larsen, E. S., and Wherry, E. T., Beidellite, a new mineral name: Washington Acad. Sci. Jour., vol. 15, pp. 465–466, 1925. Ross, C. S., and Shannon, E. V., The chemical composition and optical properties of beidellite: Idem, pp. 467–468, 1925; The minerals of bentonite and related clays and their physical properties: Am. Ceramic Soc. Jour., vol. 9, pp. 93–96, 1926. Larsen, E. S., and Steiger, George, Dehydration and optical studies of alunogen, nontronite, and griffithite: Am. Jour. Sci., 5th ser., vol. 15, pp. 14–15, 1928.

[5] Ross, C. S., and Foshag, W. F., Anauxite, A mineral species, based on material from Bilin, Czechoslovakia: Am. Mineralogist, vol. 13, pp. 153–155, 1928. Allen, V. T., Anauxite from the Ione formation of California: Idem, pp. 145–152.

[6] See Grout, F. F., Relation of texture and composition of clays: Geol. Soc. America Bull., vol. 36, pp. 393–416, 1925.

tite not uncommonly occurs with other iron sulphides in fine-grained sedimentary rocks,[7] and it can be formed in the laboratory with other iron sulphides at low temperatures.[8] The gypsum in this limestone is microfibrous and present chiefly in subparallel veinlets. (See pl. 5, *B*.) A lesser amount occurs as replacement films around shell fragments and calcite crystals in and immediately adjacent to pyrite masses. Some of the isolated crystals of calcite in these pyrite masses have been completely altered to gypsum. The pyrite masses also contain shells of Foraminifera, from 60 to 170 microns in diameter, that have been altered to gypsum and to iron oxide. Minor constituents of the rock are reddish iron oxide (bordering the pyrite masses), brownish bone phosphate (fragments of bones, teeth, and fish scales), and dark organic matter.

The gypsum in this sample is clearly secondary and, with the iron oxide, was very probably derived from the weathering of pyrite. The pyrite, on the other hand, was probably formed at the time of deposition of the rock or shortly thereafter. Its occurrence in isolated, minutely crystalline nodules and its intimate association with unpyritized shell fragments and perfect calcite crystals of different sizes (calcium carbonate being the substance it might most conceivably have replaced) strongly suggest that it formed on the sea floor or in the bottom oozes along with the fragments of shells and bones and the organic matter and carbonates. The partial corrosion indicated by the surface coatings of pyrite on some of the shell fragments probably took place soon after deposition in the waters or muds charged with hydrogen sulphide. (See p. 13.) The fact that the group of samples studied falls into a rough series in which the proportions of pyrite, organic matter, carbonates, and clay seem to be interrelated (see pp. 11–13) affords further evidence that the pyrite in this sample was formed during or soon after the deposition of the other constituents.

This sample came from the basal bed of the Greenhorn formation, and it contains an unusually large amount of pyrite. Hence it accords with Goldman's observation [9] that sulphides, like glauconite and phosphate, commonly occur at basal contacts. However, Goldman's hypothesis that these basal sulphides accumulate during periods of unusually slow sedimentation does not seem to account for this particular occurrence of pyrite. Detailed tracing of beds and many measurements of stratigraphic sections in the general vicinity of the outcrop from which sample D was collected show conclusively that this bed, like others in the Greenhorn formation, grades laterally within a few miles into calcareous concretions and black noncalcareous shale and that the calcareous beds are not measurably thinner than their shale equivalents,[10] as the hypothesis would require. In fact, the available evidence indicates that the calcareous beds accumulated somewhat more rapidly than the shale beds. (See pp. 13, 52, 53.) It may be that waters which were escaping upward from the immediately underlying black shales as they were being compacted deposited some sulphide in the base of the calcareous muds and thus contributed to the large accumulation of pyrite in the basal bed of the limestone.

CHEMICAL COMPOSITION

ANALYSES

The percentage of organic matter in rocks that contain carbonate and hydrous minerals can not be determined by combustion nor by summing up total carbon, hydrogen, oxygen, and nitrogen. Methods specially adapted for differentiating between the organic and inorganic carbon and hydrogen in such rocks have been developed in the laboratories of the Bureau of Mines,[11] and fortunately it was possible to have the samples of Upper Cretaceous rocks from the Black Hills region analyzed in those laboratories.

[7] Hatch, F. H., and Rastall, R. H., The petrology of the sedimentary rocks, pp. 201, 217, London, 1923.

[8] Allen, E. T., Crenshaw, J. L., and Johnston, John, The mineral sulphides of iron: Am. Jour. Sci., 4th ser., vol. 33, p. 214, 1912.

[9] Goldman, M. I., Lithologic subsurface correlation in the "Bend series" of north-central Texas: U. S. Geol. Survey Prof. Paper 129, pp. 4–5, 1921; Basal glauconite and phosphate beds: Science, new ser., pp. 171–173, 1922; Mississippian formations of San Saba County, Tex.: U. S. Geol. Survey Prof. Paper 146, p. 56, 1926.

[10] Rubey, W. W., Cretaceous and Cenozoic formations on the northwest flank of the Black Hills: U. S. Geol. Survey Prof. Paper — [in preparation].

[11] Fieldner, A. C., Selvig, W. A., and Taylor, G. G., The determination of combustible matter in silicate and carbonate rocks: Bur. Mines Tech. Paper 212, 1919.

Partial chemical analyses of Upper Cretaceous rocks from Black Hills region

[Analyst, H. M. Cooper, Bureau of Mines]

	A	B	C	D	E	F	G	H	I	J	K
Organic hydrogen	0. 18	0. 14	0. 63	0. 31	0. 47	0. 56	0. 19	0. 72	0. 84	0. 54	0. 50
Organic carbon	1. 20	1. 40	2. 12	2. 29	. 74	. 81	1. 47	3. 08	1. 14	5. 07	1. 75
Nitrogen	. 09	. 05	. 11	. 04	. 11	. 09	. 11	. 11	. 11	. 29	. 14
Oxygen [a]	. 94	. 00	. 00	. 00	. 00	. 00	. 93	2. 51	. 00	2. 57	. 00
Sulphur	. 37	. 28	. 19	[b] 16. 71	. 19	. 13	. 03	. 89	. 05	. 48	. 79
Carbon dioxide	. 00	([c d])	. 00	19. 98	4. 32	. 04	. 11	27. 46	2. 41	. 04	2. 98
"Moisture"	4. 95	2. 46	2. 60	3. 54	2. 49	3. 42	2. 34	1. 00	1. 90	4. 78	3. 92
"Combined" water	5. 04	2. 54	5. 45	2. 13	4. 48	3. 92	4. 86	2. 30	4. 24	4. 98	4. 24
Ash	[e] 87. 23	93. 37	89. 57	[f] 55. 05	87. 70	91. 12	89. 96	[f] 61. 93	[g] 90. 47	[e] 81. 25	[f] 86. 08
	100. 00	100. 24	100. 67	100. 05	100. 50	100. 09	100. 00	100. 00	101. 16	100. 00	100. 40
Chloroform-soluble	. 06	. 07	. 08	. 03	. 05	. 06	2. 10	. 18	. 33	. 20	. 10
Calcium carbonate [h]	. 00	[c]. 00	. 00	45. 41	9. 82	. 09	. 25	62. 42	[i] 6. 35	. 09	6. 78
Organic matter [j]	2. 41	1. 59	2. 86	2. 64	1. 32	1. 46	2. 70	6. 42	2. 09	8. 47	2. 39
Pyrite [k]	. 69	. 52	. 36	[l] 25. 2	. 36	. 24	. 06	1. 66	. 09	. 90	1. 48

[a] Oxygen calculated by difference, probably a minimum value.
[b] In part as SO_3.
[c] CO_2 in another sample of Mowry shale=0.00 (U. S. Geol. Survey Prof. Paper 154, p. 157, 1929).
[d] Not determined.
[e] Ash minus the SO_3 found in the ash.
[f] Ash minus the SO_3 and CO_2 found in the ash.
[g] Contains oxidized iron from siderite.
[h] Calcium carbonate calculated from CO_2.
[i] Ferrous carbonate instead of calcium carbonate.
[j] Organic matter =H+C+N+O, probably a minimum value.
[k] Pyrite calculated from S.
[l] Allowing for S present as SO_3.

A. Black shale from Skull Creek member of Graneros shale.
B. Siliceous shale from Mowry member of Graneros shale.
C. Black shale from Belle Fourche member of Graneros shale.
D. Pyritic limestone from Greenhorn formation.
E. Calcareous marl from Greenhorn formation.
F. Gray shale from upper member (Turner sandy member) of Carlile shale.
G. Gray shale from lower member (Sage Breaks shale member) of Niobrara formation.
H. Calcareous marl from upper member (Beaver Creek chalky member) of Niobrara formation.
I. Ferruginous shale from lower part (Gammon ferruginous member) of Pierre shale.
J. Black shale from middle part (Mitten black shale member) of Pierre shale.
K. Gray shale from upper part of Pierre shale.

ORGANIC MATTER

Several of the hydrogen-carbon ratios are very high. Three of the samples (E, F, and I) show higher atomic ratios between these two elements than are shown by any known organic compounds, and redeterminations on these samples again gave high hydrogen contents.[12] This result suggests that inorganic hydrogen was not entirely eliminated by the method of analysis used.

The oxygen as given in these analyses was determined by difference, and as any oxidation of iron in pyrite, siderite, or clay minerals would increase the weight of the ash, the percentage of oxygen is therefore probably too low in every sample. Otherwise the low oxygen content of most of the samples would indicate that the organic matter consists largely of hydrocarbons instead of carbohydrate-like compounds,[13] as is suggested by direct microscopic examination.

The total organic matter, taken as the sum of organic hydrogen, organic carbon, nitrogen, and oxygen in the different samples, ranges from 1.3 to 8.5 per cent by weight (about 2 to 14 per cent by volume). As the oxygen percentages are probably too low, these totals are also probably minimum figures. It is of interest to note that the black shales, such as samples A and C, do not contain more than the average percentage of organic matter. Apparently the dark color is due quite as much to absence of carbonates as to abundance of organic matter.

The percentages of organic matter soluble in chloroform range from 0.03 to 2.10[14] and average about 0.3 per cent. The maximum percentage corresponds to about 5 gallons of soluble bitumen to a ton of the rock. The samples appeared to be truly comparable, for no relation between these percentages and the time that had elapsed between collection and analysis could be detected.

It is noteworthy that these percentages of chloroform-soluble organic matter seem to vary stratigraphically; formations of Benton age, below the Niobrara formation (A to F), contain less and the higher formations (G to K) contain more than 0.09 per cent. Washburne[15] and Geis[16] have suggested that the Mowry shale may be an important source of light oils, but the single sample from this forma-

[12] Selvig, W. A., personal communication.

[13] For a table of the chemical compositions of organic compounds commonly found in sedimentary rocks see White, David, The carbonaceous sediments, in Twenhofel, W. H., Treatise on sedimentation, pp. 311–313, 1926.

[14] A check determination on sample G, made by E. T. Erickson, of the United States Geological Survey, gave 2.0 per cent.

[15] Washburne, C. W., Some physical principles of the origin of petroleum: Am. Assoc. Petroleum Geologists Bull., vol. 3, pp. 357–359, 1919.

[16] Geis, W. H., The origin of light oils in the Rocky Mountain region: Idem, vol. 7, pp. 499–504, 1923.

tion that was analyzed (B) yielded only 0.07 per cent by chloroform extraction. The apparent relation to stratigraphic position suggests that either increasing depth of burial or overburden does not increase but may even decrease the percentage of chloroform-soluble organic matter, or else the conditions of deposition, alteration after deposition, or the kind of organic matter itself changed progressively so as to obscure completely any increase due to increasing overburden.

The percentages of soluble organic matter seem to be related also to the degree of deformation that the rocks have undergone, as indicated in the following table:

Apparent relation of percentage of soluble organic matter to degree of deformation

Sample	Deformation: Dip of rocks where sampled	Deformation: Proximity to faults	Percentage by weight of chloroform-soluble organic matter
	°		
G	50		2.10
J	45		.20
F	33		.06
H	10	(a)	.18
Average			.64
K	7	(a)	.10
D	6		.03
I	5	(a)	.33
A	5		.06
Average			.13
C	4		.08
B	3		.07
E	1		.05
Average			.07

a Near faults.

The exceptions to this average increase of chloroform-soluble matter with increase of dip may be caused by variations of other influencing factors, such as the stratigraphic position or depth, the amount and composition of the organic matter, and the extent to which the rock has yielded internally to the deformation to which it has been subjected. (See pp. 35, 38, 54.) Several of these factors seem to be operative, for not only is there a stratigraphic variation but the percentage of chloroform-soluble matter in a general way increases with the percentage of organic matter, and it also seems to vary with the composition of the organic matter.

This relationship, if it is a real and not merely an accidental one, suggests that the chloroform-soluble matter was generated in place or else made more soluble by the same processes that tilted the rocks. The more obvious alternative explanations, such as an inflow of liquid hydrocarbons into the rock or a decrease of inorganic constituents in the rock during folding, seem to be clearly inapplicable to the group of samples studied. Mead[17] has suggested that, inasmuch as sandstones increase in volume when deformed, fluids may move into rocks that are being folded. However, as is discussed on pages 35 and 38, very fine-grained sediments such as shales probably decrease instead of increase in volume when deformed, and in the samples studied porosity seems to decrease with increasing dip of the rocks—that is, during folding fluids probably moved out of the rocks represented by these samples, not into them. The other alternative explanation mentioned above—that folding squeezes and hence decreases the percentage of inorganic constituents in a rock—could rest only upon a failure to distinguish between percentage by volume and percentage by weight. Although compacting during folding decreases the volume of a rock (pp. 35–38, 54), it might also decrease, and certainly would not increase, the percentage by weight of chloroform-soluble organic matter in the rock.

This suggestion that the chloroform-soluble matter may have been generated during folding is one that would be expected under the theory proposed by McCoy and Trager[18] that the organic matter in a rock is progressively transformed into soluble bitumens by deformation. However, other geologists[19] have reached different conclusions. McCoy's theory conforms with the evidence (pp. 35–36, 38, 54) that the rocks represented by these samples have been deformed internally by tilting and faulting.

If the percentage of soluble organic matter is an approximate measure of the oil in a rock and if the apparent relationship should prove to be a real and general one, it is of considerable economic importance, for it suggests that oil may be formed in any of the marine shales, mudstones, or marls in this region that have been sufficiently deformed by tilting or faulting. At first thought this suggestion seems to be contradicted by the results of drilling, for in this region oil has been found only in rocks of the Colorado group and older formations. However, a closer inspection indicates that this apparently contradictory evidence from drilling is by no means conclusive. For one thing, suitable reservoir rocks for the accumulation of petroleum are unknown in the Pierre shale throughout the southern part of the region, and they have not been penetrated by the drill where present farther

[17] Mead, W. J., The geologic rôle of dilatancy: Jour. Geology, vol. 33, pp. 691, 697–698, 1925.

[18] McCoy, A. W., Notes on the principles of oil accumulation: Jour. Geology, vol. 27, pp. 252–254, 1919. Trager, E. A., Kerogen and its relation to the origin of oil: Am. Assoc. Petroleum Geologists Bull., vol. 8, pp. 301–311, 1924. McCoy, A. W., Soluble matter in oil shale: Idem, vol. 9, p. 1025, 1925.

[19] Van Tuyl, F. M., and Blackburn, C. O., The effect of rock flowage on the kerogen of oil shale: Idem, vol. 9, pp. 158–164, 1925. Mead, W. J., and Hawley, J. E., The generation of oil in rocks by shearing pressures: Am. Petroleum Inst. Bull., vol. 8, No. 54, p. 3, 1927.

north. Furthermore, in the four general districts in northeastern Wyoming (Newcastle, Osage, Thornton, and Moorcroft) where oil has been found in Upper Cretaceous rocks, its occurrence is closely related to minor faulting and steep dips.[20] Thus the results of drilling in themselves suggest that other districts of minor faulting and steep dips in the same region may be worth drilling.

The implications of this apparent increase of chloroform-soluble matter with increasing dip are so far-reaching as to indicate the desirability of examining it more critically and of determining if possible whether it is due to the causal relation assumed or to some more indirect or even fortuitous relation. As is discussed more fully on pages 35–38, the porosity as well as the content of chloroform-soluble matter in these rocks is related to their dip. In general the porosity decreases and the percentage of chloroform-soluble matter increases with increasing dip. Hence it might be expected that the percentage of chloroform-soluble matter would increase with decreasing porosity. Indeed, careful comparison shows a very rough relation of this sort, but it is not nearly as distinct as the relation of either porosity or chloroform-soluble matter to dip.

However, as the percentage of chloroform-soluble matter depends at least in part upon the percentage of total organic matter (p. 9), and as the percentage of total organic matter is different in the different samples, it may be a fairer test to use for comparison with the porosity the proportion of organic matter that is soluble in chloroform, instead of the gross percentage of chloroform-soluble matter. If this comparison is made, using the total carbon, hydrogen, and nitrogen (neglecting oxygen because of the uncertainty in its determination, p. 8) as an index of the amount of organic matter in the different samples, we find a rather distinct relationship.

Apparent relation between ratio of organic matter to chloroform-soluble matter and porosity

Sample	Total percentage by weight of organic carbon, hydrogen, and nitrogen	Percentage by weight of chloroform-soluble organic matter	Ratio of organic matter to chloroform-soluble matter	Porosity (percentage by volume)
C	2. 86	0. 08	35. 8	33. 3
J	5. 90	. 20	29. 5	35. 8
E	1. 32	. 05	26. 4	37. 6
A	1. 47	. 06	24. 5	32. 5
F	1. 46	. 06	24. 3	23. 8
K	2. 39	. 10	23. 9	25. 4
H	3. 91	. 18	21. 7	25. 4
I	2. 09	. 33	6. 3	26. 0
G	1. 77	2. 10	. 8	25. 3

[20] Rubey, W. W., The oil and gas possibilities of the Black Hills rim in Wyoming and Montana: U. S. Geol. Survey Bull. — [in preparation].

That is, the ratio of organic matter to chloroform-soluble matter seems to decrease with decreasing porosity. In other words, the more porous the rock the smaller the proportion of its organic matter that is soluble in chloroform. This apparent relation suggests that in the more porous samples a portion of the chloroform-soluble matter may have been lost by oxidation[21] or carried away by percolating ground waters, an interpretation very different from the one based on the apparent relation between chloroform-soluble matter and dip.

The data available in this study do not warrant a decision as to which of these two possible explanations best accounts for the variations in the percentage of chloroform-soluble matter in these samples. The only conclusions that can safely be drawn are that the soluble organic matter in these rocks seems to increase with increasing dip and decreasing porosity. Inasmuch as the porosity appears to be fundamentally related to the dip, it is impossible to say whether the chloroform-soluble matter has been increased by deformation of the more highly tilted rocks or lost by oxidation or flushing from the more porous samples. The problem is of sufficient economic interest to justify further study.

PYRITE

The percentages of pyrite in the different samples may be estimated from the sulphur content. A small fraction of the sulphur may, of course, be combined with the organic matter. However, in nearly all the samples the ratio of sulphur to total organic matter is much too large for the sulphur to be accounted for in this way, and as disseminated pyrite was found in many of the samples by microscopic examination it seems that in this group of samples organic sulphur is probably negligible. Another portion of the total sulphur may be combined as gypsum, as in sample D; but in view of the fact that none was found in microscopic examination of the other samples and the probability that any such sulphates were formed by weathering of the pyrite (p. 7), it still seems justifiable to take the sulphur content as a measure of the pyrite once present in the unaltered rock. The presence of small quantities of either organic sulphur or sulphates would make the estimated percentage of pyrite too high. However, a third portion of the sulphur may be combined as pyrrhotite or as ferrous sulphide (see p. 6); and as this form of combination would make the estimated percentage of iron sulphide too low, the possible errors tend to offset one another. Computed thus, the pyrite content ranges from less than 0.1 to nearly 1.7 per cent in the shale and marl

[21] Hawley, J. E., Generation of oil in rocks by shearing pressures: Am. Assoc. Petroleum Geologists Bull., vol. 13, pp. 313–314, 1929.

samples and to 25.2 per cent in the pyritic limestone (sample D).

CARBONATES

The chemical determinations of CO_2 and the microscopic determinations of the mineral species and purity of the carbonates afford the data necessary for computing the percentages of carbonate in the different samples. In all but one of the samples containing any carbonate it was found to be relatively pure calcite; in that one (sample I) the carbonate is siderite. The three samples from the Graneros shale contained no carbonate; the maximum was 62 per cent, in a sample of marl from the Niobrara formation. Next to the Niobrara and Greenhorn samples, those from the Pierre shale contain the most carbonate.

In considering these differences in the composition of the different formations, the possible importance of the large carbonate concretions in some of the shale members must not be overlooked. For example, the sample of shale from the lower or Sage Breaks member of the Niobrara formation (sample G) contains only 0.25 per cent of carbonate, but this member also contains many large concretions made up of relatively pure calcite, which were not represented in the sample. Similarly, the samples of shale from the Skull Creek and Belle Fourche members of the Graneros shale (samples A and C) contain no carbonate, but these shales are interbedded with zones of sideritic concretions. The lateral continuity or persistence of these and of other concretions in the region and the low carbonate content and relative impermeability [22] of the inclosing shale beds lead the writer to believe with Tarr [23] that, for the most part, the concretions were formed before the muds were deeply buried. If this view is correct, the total carbonates in the entire member (concretions and all—not simply the carbonates in the shale) should be considered in comparing the different formations. However, rough estimates by the writer indicate that the total volume of the concretions is not sufficiently great to alter appreciably the order of relative carbonate contents that is indicated by the rock specimens alone.

COMPOSITION OF PYRITIC LIMESTONE

The CO_2 content indicates less than 50 per cent of carbonates in sample D, but in preparing it for mechanical analysis 73 per cent was found soluble in dilute hydrochloric acid. Microscopic examination indicated that the other soluble material consisted of gypsum, bone phosphate, and iron oxide, and the evolution of H_2S in hydrochloric acid indicated some pyrrhotite. (See pp. 6–7.) With additional determinations, made by J. G. Fairchild, of the United States Geological Survey, of SO_3 (8.2 per cent) and P_2O_5 (less than 1 per cent), the probable composition of the rock was computed as follows:

Composition of pyritic limestone

		Per cent
Calcite		45.4
Pyrite	about	25.2
Gypsum		17.6
Iron oxide (computed as limonite)	less than	6.1
Organic matter	more than	2.6
Bone phosphate	about	2.0
Moisture		1.1
		100.0

This composition, as computed from the analyses, seems fairly reliable, for it is independently checked within 1 per cent by both the acid-soluble and the ash determinations. It indicates that the ash, as corrected for SO_3 and CO_2, consists of two-thirds CaO, from calcite, gypsum, and bone phosphate, and one-third Fe_2O_3, from pyrite and limonite. It also indicates by the relative amounts of the different constituents that the gypsum was in part derived from the weathering of pyrite and calcite originally in the rock and in part introduced from outside. This conclusion seems to be borne out by the observation of both sulphatized carbonate grains and secondary veinlets of gypsum. (See pl. 5, *B*, and pp. 6–7.)

CONDITIONS OF DEPOSITION SUGGESTED BY CHEMICAL COMPOSITION

It is of especial interest to find that in a general way the percentages of carbonate vary with the percentages of organic matter and pyrite and that all three vary with stratigraphic position. In the following table the samples are arranged in order of increasing percentages of total carbonates and organic matter. This arrangement makes the samples fall into general stratigraphic groups, in which the average percentages of pyrite increase with the average percentages of carbonate and organic matter.

Percentages of total carbonates plus organic matter and of pyrite

Samples from the Colorado group, not including those from the calcareous parts of the Greenhorn and Niobrara formations

Sample	Carbonates plus organic matter	Pyrite
F	1.55	0.24
B	1.59	.52
A	2.41	.69
C	2.86	.36
G	2.95	.06
Average	2.3	.4

[22] In fact, water was being squeezed out of the shale, rather than entering it, from the time of deposition until the time of uplift and erosion. (See pp. 35–38, 54.)

[23] Tarr, W. A., Syngenetic origin of concretions in shale: Geol. Soc. America Bull., vol. 32, pp. 373–384, 1921.

Percentages of total carbonates plus organic matter and of pyrite—Continued

Samples from the Pierre shale

Sample	Carbonates plus organic matter	Pyrite
I	8. 44	0. 09
J	8. 56	. 90
K	9. 17	1. 48
Average	8. 7	. 8

Samples from the calcareous parts of the Greenhorn and Niobrara formations

Sample	Carbonates plus organic matter	Pyrite
E	11. 14	0. 36
D	48. 05	25. 20
H	68. 84	1. 66
Average	42. 7	9. 1

The several interrelationships that bring about this grouping of the samples indicate that the conditions of deposition of the marine sediments in the Black Hills region during Upper Cretaceous time varied widely and that those conditions that were favorable for the preservation of carbonates were, in general, also favorable for the preservation of organic matter and pyrite.

This series into which all the samples seem to fall is probably suggestive of the range of conditions of deposition. In view of the fact that the carbonate consists partly of Foraminifera and molluscan shells, it is not surprising to find that the amount of organic matter varies with the amount of carbonate. Also the association of organic matter with iron sulphides is readily acceptable, for it has been noted commonly in recent sediments.[24] Even the association of carbonates and sulphides is not surprising, for there are theoretical reasons for believing that the stagnant waters favorable to the formation of iron sulphide may constitute one of the favorable environments for the accumulation of calcium carbonate. As demonstrated by Wells [25] and by Johnston and Williamson,[26] the solubility of calcium carbonate depends largely upon the carbon dioxide content of sea water, which in turn depends chiefly upon the temperature of the water. The warmer (that is, in general, the shallower) the water, the less calcium carbonate it can dissolve. The surface water of the ocean has been found to be essentially saturated with respect to calcium carbonate. Hence it would seem that deeper waters would be unsaturated and that calcium carbonate could accumulate in deeper, colder water only where the rate of precipitation and burial exceeds the rate of solution or where the bottom waters are so stagnant that they become and remain locally saturated with respect to the carbonate. Otherwise circulation of the bottom waters would remove the saturated layer and would also bring dissolved oxygen into contact with organic matter, thus forming carbon dioxide and causing still more solution of the carbonate. This association of carbonate, organic matter, and pyrite therefore suggests the tentative hypothesis that the samples that contain the most of these constituents accumulated under anaerobic conditions.

However, this simple hypothesis does not meet all the facts satisfactorily. Foul, stagnant muds are not a favorable habitat for bottom-living organisms. The more highly calcareous and organic rock specimens contain the most fossils of bottom-living mollusks, and the less calcareous specimens are essentially barren of all fossils—not the reverse, as this hypothesis would indicate. Also anaerobic conditions of the sort postulated might be expected to yield deposits of siderite [27] instead of a mixture of calcium carbonate and iron sulphide. However, only one of the samples (I) contains detectable amounts of ferrous carbonate, and it is the less, not the more calcareous formations in this region that contain the more highly ferruginous carbonate concretions [Skull Creek shale (A), Belle Fourche shale (C), and Carlile shale (F)]. Furthermore, the preservation of very thin laminations suggests that the beds represented by the less calcareous samples (see pp. 51–52) were deposited in relatively quiet, not in relatively agitated (and therefore well-oxidized) water. Consequently, a somewhat different explanation for the association of calcium carbonate, organic matter, and pyrite must be sought.

A modified explanation is suggested by a separate consideration of the conditions favorable for preservation of each of these three constituents. The conditions theoretically most favorable for the preservation of calcium carbonate are warm waters (shallow or tropical), rapid burial, or stagnant saturated bottom waters. As the evidence seems to preclude the possibility of stagnant waters, there is at least a suggestion that the deposits represented by the more calcareous samples were formed at relatively shallow

[24] Murray, John, and Renard, A. F., *Challenger* Rept., Deep-sea deposits, p. 253, 1891. Murray, John, and Irvine, Robert, On the chemical changes which take place in the composition of the sea water associated with blue muds on the floor of the ocean: Roy. Soc. Edinburgh Trans., vol. 37, p. 498, 1895. Harder, E. C., Iron-depositing bacteria and their geologic relations: U. S. Geol. Survey Prof. Paper 113, pp. 62–63, 1919. Goldman, M. I., "Black shale" formation in and about Chesapeake Bay: Am. Assoc. Petroleum Geologists Bull., vol. 8, pp. 195–201, 1924. Bastin, E. S., The problem of the natural reduction of sulphates: Am. Assoc. Petroleum Geologists Bull., vol. 10, pp. 1272–1280, 1926.

[25] Wells, R. C., The solubility of calcite in water in contact with the atmosphere and its variation with temperature: Washington Acad. Sci. Jour., vol. 5, pp. 617–622, 1915; New determinations of carbon dioxide in water of the Gulf of Mexico: U. S. Geol. Survey Prof. Paper 120, pp. 1–16, 1919.

[26] Johnston, John, and Williamson, E. D., The rôle of inorganic agencies in the deposition of calcium carbonate: Jour. Geology, vol. 24, pp. 729–750, 1916.

For a discussion of the application of these principles to the interpretation of the conditions of sedimentation of shales see Rubey, W. W., Origin of the siliceous Mowry shale of the Black Hills region: U. S. Geol. Survey Prof. Paper 154, pp. 164–165, 1929.

[27] Harder, E. C., op. cit., pp. 55, 72.

depths in warm, well-lighted, and well-oxidized water, where food supply and organisms were abundant and burial was rapid.

The amount of organic matter preserved in a sedimentary rock is the "product of an equation between the rate of supply of organic matter and the rate of decomposition"; [28] it depends upon the chemical composition and the rate of growth of the organic matter, the rate of burial, and the rate of decay before and after burial. In general, relatively shallow water and moderately rapid sedimentation afford both a suitable habitat and optimum conditions for the preservation of organisms.

Ferrous sulphide is readily formed, in the presence of abundant organic matter and under reducing conditions, by the bacterial reduction of sulphates or by the action of hydrogen sulphide (chiefly from decaying protein) upon ferrous salts (bicarbonates, etc.) dissolved in sea water, upon ferric hydroxides previously deposited in the muds, or upon iron in clay minerals.[29] The ferrous sulphide is said to change over eventually into the more stable form, pyrite.

Combining these conclusions indicates that the conditions most favorable for the formation and preservation of calcium carbonate, organic matter, and pyrite together would be relatively shallow water and a rapid rate of accumulation and burial of organic matter. Under these conditions much organic matter and carbonate would be preserved by burial, and iron sulphide would form abundantly in the putrefying ooze—that is, chiefly below its upper surface.

This suggested explanation is more or less confirmed by studies of the present-day conditions of deposition of marine sediments. The blue muds are the predominant type of deposit in both deep and shallow water in all partly inclosed seas and for several hundred miles seaward from the oceanic coasts.[30]

The materials of which the blue muds are principally composed are derived from the disintegration of continental land and are very complex in character. When collected this deposit is blue or slate-colored, with an upper red or brown colored layer, which had been in immediate contact with the water. The blue color is due to organic matter and sulphide of iron in a fine state of division, and these muds have, as a rule, when taken from the sounding tube or dredge, a smell of sulphureted hydrogen. The red or brown color of the thin watery upper layer is evidently due to the presence of ferric oxide or ferric hydrate, but as the deposit accumulates this oxide is transformed into sulphide and ferrous oxide in the presence of organic matter in the underlying layers. * * * Sometimes the samples are homogeneous, at other times the aspect is heterogeneous, owing to the presence of large fragments of rocks and shells and small fragments of calcareous organisms. * * * They may contain from only a trace to 35 per cent of carbonate of lime.[31]

In general, the blue muds, like other oceanic deposits, show a gradual decrease in quantity of calcium carbonate with increase of depth.[32] The amount of fine mud was found to range from 16 to 97 per cent.[33]

The presence of sulphides and sulphureted hydrogen in all harbor muds, muddy bays near land, and, indeed, in nearly all the terrigenous deposits, such as the blue muds, is a sure indication that soluble and insoluble albuminoid and other organic matters are distributed throughout these muds and are in process of decomposition. Probably sulphides are present in all deep-sea deposits, but they are most abundant in muds near land, where there is rapid accumulation and where a large quantity of organic matter is borne down from the continents. In the red clays and the other truly pelagic deposits the quantity of organic matter is much less, and, owing to the slow accumulation, the sulphides are probably oxidized as soon as formed and never make up any considerable portion of the deposit.[34]

Murray and Renard conclude that, of the different types of marine sediments, blue muds "not far removed from embouchures of large rivers" are accumulating the most rapidly.[35]

From a study of the sea water associated with blue muds, collected from depths of 1 to 2, 3 to 5, and 16 fathoms, Murray and Irvine [36] concluded that calcium sulphate in the water is reduced to sulphides by bacterial decay of abundant organic matter and recombined with ferric oxide in the surface layer and with carbon dioxide from decaying organic matter so as to make the muds rich in iron sulphide and the sea water immediately above the muds more nearly saturated with respect to calcium carbonate. Similar conclusions were drawn by Van Delden [37] from an examination of black sulphide-bearing muds (overlain by a layer of light-colored mud a few millimeters thick) from canals in Holland.

Hence the association of calcium carbonate and pyrite in sediments is not inconsistent; in fact, the conditions favorable for the preservation of organic matter and sulphides also may be favorable for the growth and preservation of lime-secreting, bottom-living organisms. (See p. 53.)

[28] Goldman, M. I., op. cit., p. 200.

[29] Murray, John, and Irvine, Robert, On the chemical changes which take place in the composition of the sea water associated with blue muds on the floor of the ocean: Roy. Soc. Edinburgh Trans., vol. 37, pp. 481–507, 1895. Allen, E. T., Crenshaw, J. L., and Johnston, John, The mineral sulphides of iron: Am. Jour. Sci., 4th ser., vol. 33, pp. 169–236, 1912. Harder, E. C., op. cit., pp. 40–44, 60–64, 74–75, 82–84. Bastin, E. S., op. cit., pp. 1270–1299. Newhouse, W. H., Some forms of iron sulphide occurring in coal and other sedimentary rocks: Jour. Geology, vol. 35, pp. 73–83, 1927.

[30] Murray, John, and Renard, A. F., *Challenger* Rept., Deep-sea deposits, pp. 229–233, 1891.

[31] Idem, p. 229.

[32] Idem, p. 230.

[33] Idem, p. 231.

[34] Idem, p. 253.

[35] Idem, p. 411.

[36] Murray, John, and Irvine, Robert, On the chemical changes which take place in the composition of the sea water associated with blue muds on the floor of the ocean: Roy. Soc. Edinburgh Trans., vol. 37, pp. 481–507, 1895.

[37] Cited by Bastin, E. S., op. cit., p. 1275.

MECHANICAL ANALYSES

The mechanical composition, or the relative amounts of particles of the size of gravel, sand, and clay in a sedimentary rock, is one of its most significant lithologic characteristics. The proportions of grains of different sizes serve as an index not only of the general appearance and behavior of the rock under common conditions but also of its economic value and geologic history. This size distribution of particles is equally important in other fields of investigation, as in agriculture, ceramics, metallurgy, chemistry, and highway engineering and in the manufacture of cement, abrasives, paint, graphite, fertilizers, rubber, photographic emulsions, and of some drugs and foodstuffs. Consequently many methods for the determination of particle size are continually being proposed, and especially within the last 10 years much has been written on the subject. As many of the methods that geologists have used for determining the mechanical composition of fine-grained sedimentary rocks are in some respects unsatisfactory, a review of the methods that seemed applicable to such rocks is presented here to indicate the basis on which a method suitable for the particular study described in this paper was chosen.

REVIEW OF PROPOSED METHODS APPLICABLE TO FINE-GRAINED SEDIMENTARY ROCKS

The mechanical analysis of a sedimentary rock involves three more or less distinct problems: The rock must be prepared for study; the sizes of the particles must be measured; and the data must be presented in some readily usable form. In general, each of these problems is more difficult for the finer than for the coarser sediments.

PREPARATION OF SAMPLE

For studies in which microscopic examination and grain counting is a sufficiently accurate method of determining the mechanical composition, hand specimens of the rock may be examined under the lens directly, thin sections may be cut,[38] or samples may be disintegrated for a more thorough study of the individual grains. But for the examination of many common types of sedimentary rocks, some more objective and mechanical method of analysis is probably more accurate. The essential preparation for these more objective analyses is a thorough disintegration of the rock—that is, a thorough separation of the individual rock particles. Some rocks disintegrate readily in water or even in air. Others, such as siliceous shales and silicified sandstones, can not be disintegrated without methods that so change the size and number of the grains that mechanical analyses are then worthless. Most fine-grained sedimentary rocks require some sort of special treatment to separate the individual particles.

The method of disintegration to be used depends largely upon the object of the analysis. If the present composition of the rock is desired for descriptive or mineralogic purposes, all secondary minerals should be retained. If, however, the mechanical composition at the time of deposition is sought, these secondary minerals must be excluded. For many geologic investigations the goal is this original composition, because from it inferences can be drawn as to the velocity of currents and other conditions at the time of sedimentation. Of course it is not always possible to determine just which minerals are secondary, and even if they are known it is commonly difficult to remove them without greatly altering the original constituents. For example, if the rock examined is a sandstone cemented with calcite, the investigator must know that the sand grains are not carbonates before he proceeds to remove the cement with acid. In clay rocks the difficulties are even greater. At the time of sedimentation most of the clay particles may have been flocculated by salts dissolved in the water[39] into aggregates which settled as units, and thus the present size of the individual particles would give no clue to the size of the original aggregates. Also the clay particles may have been very small at the time of deposition, but after burial they may have grown into much larger crystals of clay minerals.[40] Thus the present size of the clay particles may differ greatly from their original size. Furthermore, even the present composition is difficult to determine, as it is nearly impossible to disintegrate most clay rocks completely. Crushing may fracture the delicate platy crystals, heating may break down the hydrous clay compounds, and chemicals or even pure water may dissolve or cause chemical changes in the finer particles.

In view of these uncertainties and difficulties, it is clearly impossible to attain anything more than an approximation to the initial size distribution of particles in clay rocks. Some methods of preparation, measurement, and presentation give more accurate results than others, but it is useless to seek methods that are much more precise than the least accurate or most uncertain step in the entire determination.

[38] For methods particularly adapted to cutting thin sections of shale and clay, see Sayles, R. W., Microscopic sections of till and stratified clay: Geol. Soc. America Bull., vol. 32, pp. 59–62, 1921; Ross, C. S., A method of preparing thin sections of friable rock: Am. Jour. Sci., 5th ser., vol. 7, pp. 483–485, 1924; Methods of preparation of sedimentary materials for study: Econ. Geology, vol. 21, pp. 460–468, 1926.

[39] Grabau, A. W., Principles of stratigraphy, pp. 655–657, 1913.

[40] Leith, C. K., and Mead, W. J., Metamorphic geology, pp. 107–108, 1915. Lewis, J. V., Fissility of shale and its relation to petroleum: Geol. Soc. America Bull., vol. 35, p. 578, 1924. Twenhofel, W. H., Treatise on sedimentation, p. 186, 1926.

For most of the clay rocks the uncertainty about the degree of flocculation during sedimentation and the amount of secondary growth of clay particles since deposition is probably greater than the errors in some of the methods of disintegration; likewise the unavoidable errors in even the best methods of disintegration probably exceed the errors in most of the methods of measurement and presentation of results.

Many investigations have been made, especially by soil scientists, of the methods of disintegrating soils and clays, but no simple expeditious procedure that gives satisfactory results seems yet to have been found. Nearly all fine-grained rocks and soils require chemical treatment to loosen the cement or deflocculate the grains. Soil physicists recommend that carbonates and humus, which may cement the finer particles, should be removed with dilute hydrochloric acid and hydrogen peroxide or nitric acid.[41] However, geologists will probably agree that, as the carbonates and organic matter may be essential constituents of the rock, they should, if possible, be retained in the mechanical analysis. Furthermore, in the treatment with hydrochloric acid some of the finest clay material is dissolved.[42] Most geologists, soil physicists, and other investigators have agreed that the tendency of small clay particles to gather together in floccules[43] must be overcome by washing out the electrolytes in the sample with distilled water[44] or by neutralizing the water with some chemical, usually an alkali (ammonia, ammonium hydroxide, sodium carbonate, sodium hydroxide, potassium hydroxide, barium hydroxide, water glass, soap, or pyrogallol[45]). Washing out the electrolytes is a tedious process, and if it is attempted great care must be taken to avoid loss of the finer clay particles. Of the chemicals that may be added for neutralization, ammonia and sodium carbonate seem most effective, most easily handled, and least harmful to the sample. Sodium carbonate is highly effective[46] but dissolves more fine particles of silica than ammonia[47] does. Ammonia is said to cause rather than prevent flocculation, if calcium or magnesium carbonate is present in large quantities.[48] Concentrations ranging from 0.007 to 2.5 per cent of ammonia have been used by different investigators; the lower concentrations give most complete dispersions, but somewhat higher concentrations may be desirable, especially if organic matter and carbonates are present, in order to offset loss by evaporation and to maintain alkalinity throughout the analysis.[49] Puri and Keen[50] obtained most nearly complete dispersions in suspensions of 1 part soil to 100 or more parts of water.

In addition to the chemical treatment, most fine-grained rocks or soils must be subjected to mechanical agitation to separate the individual grains. This agitation may be accomplished by shaking in the hand; by some mechanical shaking, vibrating,[51] or stirring[52] device, or by boiling. Samples that do not yield to this treatment must be rubbed gently with the finger, a rubber pestle, or a brush. Violent methods such as crushing with mortar and pestle or heating to redness and disrupting by sudden immersion in cold water almost certainly defeat the purpose of the mechanical analysis. Experiments indicate that, as might be expected, differences in mechanical treatment produce widely different results in deflocculation and that the rubber-pestle and brush methods, though liable to subjective errors, are probably most effective.[53]

[41] Atterberg, A., Die rationelle Klassifikation der Sande und Kiese; Chem. Zeitung, Jahrgang 29, p. 195, 1905. Hissink, D. J., Methods of mechanical analysis of soils: Internat. Soc. Soil Sci. Proc., vol. 1, pp. 705–724, 1925. Report of the mechanical analysis subcommittee of the Agricultural Education Association—The mechanical analysis of soils; a report on the present position and recommendation for a new official method: Jour. Agr. Sci., vol. 16, pp. 123–144, 1926.

[42] Hissink, D. J., op. cit., pp. 712–713. Novák, V., and Smolík, L., Sur la quantité et la composition chimique de l'argile colloidale des sols: IV^{ème} Conférence internat. pédologie [Rome] Actes, vol. 2, pp. 128–141, 1926.

[43] Wentworth, C. K., Methods of mechanical analysis of sediments: Iowa Univ. Studies, vol. 11, pp. 42–43, 1926.

[44] Wiegner, G., Ueber den Einfluss verschiedener Vorbehandlungsmethoden auf den mit Hilfe des Schlaemmapparates von Wiegner-Gessner ermittelten Dispersitaetsgrad von Bodensuspensionen: IVème Conférence internat. pédologie [Rome] Actes, vol. 2, pp. 87–102, 1926.

[45] Briggs, L. J., Martin, F. O., and Pearce, J. R., The centrifugal method of mechanical analysis: U. S. Dept. Agr. Bur. Soils Bull. 24, pp. 22–25, 1904. Atterberg, A., op. cit. Boswell, P. G. H., The separation of the finer constituents of sedimentary rocks: Faraday Soc. Trans., vol. 18, p. 38, 1922. Holmes, A., Petrographic methods and calculations, pt. 1, p. 186, 1923. Sayles, R. W., Seasonal deposition in marine waters: Nat. Research Council Comm. on Sedimentation Rept. [Apr. 18, 1923], p. 63, 1923. Bleininger, A. V., The properties of clays: Second Colloid Symposium Mon., p. 88, Chemical Catalog Co., 1925.

[46] Puri, A. N., and Keen, B. A., Dispersions of soil in water under various conditions: Jour. Agr. Sci., vol. 15, pp. 157–158, 1925.

[47] Steiger, George, in Twenhofel, W. H., and others, Treatise on sedimentation, p. 632, 1926.

[48] Briggs, L. J., Martin, F. O., and Pearce, J. R., op. cit., p. 24.

[49] Fletcher, C. C., and Bryan, H., Modification of the method of mechanical soil analysis: U. S. Dept. Agr. Bull. 84, p. 9, 1912.

[50] Puri, A. N., and Keen, B. A., op. cit., p. 153.

[51] Whittles, C. L., The determination of the number of bacteria in soil; II, Methods for the disintegration of soil aggregates and the preparation of soil suspensions: Jour. Agr. Sci., vol. 14, pp. 346–369, 1924.

[52] Bouyoucos, G. J., Directions for determining the colloidal material of soils by the hydrometer method: Science, new ser., vol. 66, pp. 16–17, 1927.

[53] Beam, W., The mechanical analysis of arid soils: Cairo Sci. Jour., vol. 5, p. 107, 1911. Atterberg, A., Die mechanische Bodenanalyse und die Klassifikation der Mineralböden Schwedens: Internat. Mitt. Bodenkunde, Band 2, p. 314, 1912. Richter, G., Die Ausführung mechanischer und physikalischer Bodenanalysen: Idem, Band 6, pp. 193–208, 318–346, 1916. Koettgen, P., Zur Methodik der physikalischen Bodenanalyse: Idem, Band 7, pp. 205–246, 1917. Nolte, O., Der Einfluss des Kochens und des Schüttelns auf feine Mineralteilchen—ein Beitrag zur Ausführung von mechanischen Bodenanalysen: Landw. Vers. Stat., Band 93, p. 247, 1919. Odén, Sven, Bodenkundliche Forschungen an dem chemischen Universitäts-Laboratorium der Universität Upsala, 1914–1919—III, Ueber die Vorbehandlung der Bodenproben zur mechanischen Analyse: Internat. Mitt. Bodenkunde, Band 9, pp. 301–418, 1920. Hissink, D. J., Die Methode der mechanischen Bodenanalyse: Idem, Band 11, pp. 1–11, 1921. König, J., and Hasenbäumer, J., Zur Beurteilung neuer Verfahren für die Untersuchung des Bodens: Landw. Jahrb., Band 56, p. 439, 1921. Gile, P. L., Middleton, H. E., Robinson, W. O., Fry, W. H., and Anderson, M. S., Estimation of colloidal material in soils by adsorption: U. S. Dept. Agr. Bull. 1193, pp. 16–17, 1924. Puri, A. N., and Keen, B. A., op. cit., p. 159.

In nearly all these methods disintegration is sought within a few days after the clay or soil is put into water. Some facts indicate that disintegration is attained more easily by a longer period of soaking. It seems that the readiness with which a soil or clay disintegrates depends largely upon the moisture content of the sample. Air-dried samples are much harder to disintegrate than samples examined in the natural moist condition.[54] Dispersability, especially in clays, decreases with decreasing water content or with heating above the air-dried condition, and rewetting does not seem to make the dispersion easier.[55] Hilgard [56] noted that Recent sediments disintegrate much more easily then ancient sediments. Whittles [57] found it advisable to moisten soil samples slowly in order to avoid lumpiness.

These observations, like those of Terzaghi [58] and Hedberg [59] on the compacting of muds and sands, indicate that the induration of clay is due largely to loss of moisture or of pore space—that is, to compacting by drying or by pressure—and that a considerable period of time is required for a clay rock to take up again the moisture it is capable of holding at some higher humidity or lower pressure. The inference therefore seems justified that a long period of soaking in water assists in the disintegration of clay rock. This conclusion is borne out by the writer's experience in disintegrating samples of Cretaceous shale and bentonite.[60] However, a long period of soaking increases the likelihood of chemical solution of the smallest particles and may cause some hydration of the minerals.

METHODS OF MEASUREMENT

Mechanical analyses of granular mixtures, as generally understood, give the percentages of different-sized particles in a sample. In natural mixtures, such as sedimentary rocks, particles of all sizes between the largest and smallest are usually present. By this generally accepted definition, a statement of the average size of all the particles in such a rock would not be a mechanical analysis. The percentage of grains larger and smaller than some arbitrary intermediate limit might be so considered, however; and as the number of these arbitrary limits is increased the analysis approaches a complete statement of the size distribution within the sample. Obviously the more nearly complete the analysis the greater its usefulness, for thus the degree of uniformity of the sizing and any irregularities in the size distribution become more apparent. On the other hand, the more nearly complete the analysis the more laborious the determination. Fortunately, some methods give a moderately complete analysis with one manipulation, whereas others require repeated separations or a multiplication of apparatus to attain the same degree of refinement.

Aside from those methods that determine a single physical property of a granular substance, such as the average diameter [61] or the total internal surface of the particles,[62] and those methods that are especially adapted for submicroscopic particles, such as those that measure the amplitude or effectiveness of the Brownian movement,[63] there are three principal methods for the mechanical analysis of mixtures containing small particles—separation by sieves, microscopic counts, and separation based on the different settling velocities of large and small particles in a liquid. Each method measures slightly different properties of particles and is best suited for certain sizes, but it is well to check one method against the others to avoid serious errors in interpretation.

SIEVES

The size of an equidimensional hole through which a particle will pass is determined not by the maximum or minimum dimension of the particle, but chiefly by the intermediate dimension. If a mixture of particles is satisfactorily disintegrated and the sizes of the holes in several sieves are known, the proportions of grains whose intermediate dimensions are greater and less than each sieve size can be determined with relative ease and sufficient accuracy.[64] However, the method is not applicable to very small particles, for Holmes states that the smallest grains that can satisfactorily

[54] Ehrenberg, P., and Van Zyl, J. P., Weitere Untersuchungen über die Beschaffenheit der Bodenkrümmel: Internat. Mitt. Bodenkunde, Band 7, pp. 90–103, 1917; Band 8, pp. 41–49, 1918.

[55] Puri, A. N., and Keen, B. N., op. cit., pp. 150–151, 153, 156.

[56] Hilgard, E. W., On the silt analysis of soils and clays: Am. Jour. Sci., 3d ser., vol. 6, p. 335, 1873.

[57] Whittles, C. L., op. cit., p. 366.

[58] Terzaghi, Charles, Principles of soil mechanics: Eng. News-Record, vol. 95, pp. 742–746, 796–800, 832–836, 874–878, 912–915, 987–990, 1026–1029, 1064–1068, 1925.

[59] Hedberg, H. D., The effect of gravitational compaction on the structure of sedimentary rocks: Am. Assoc. Petroleum Geologists Bull., vol. 10, pp. 1035–1072, 1926.

[60] Rubey, W. W., Cretaceous and Cenozoic formations on the northwest flank of the Black Hills: U. S. Geol. Survey Prof. Paper — [in preparation].

[61] Anderson, M. S., and Mattson, Sante, Properties of colloidal soil material: U. S. Dept. Agr. Bull. 1452, p. 4, 1926. Stutz, G. F. A., and Pfund, A. H., A relative method for determining particle size of pigments: Ind. and Eng. Chemistry, vol. 19, pp. 51–53, 1927. Clark, G. L., X-rays and colloids: Colloid Symposium Mon., vol. 4, pp. 156–162, 1926.

[62] Sauramo, Matti, Studies on the Quaternary varve sediments in southern Finland: Comm. géol. Finlande Bull. 60, pp. 17–19, 1923. Zunker, F. Die Bedeutung und Bestimmung der spezifischen Oberfläche des Bodens: IVème Conférence internat. pédologie [Rome] Actes, vol. 2, pp. 238–249, 1926.

[63] Wightman, E. P., and Sheppard, S. E., The size-frequency distribution of particles of silver halide in photographic emulsions and its relation to sensitometric characteristics—II, The methods of determining size-frequency distribution: Jour. Phys. Chemistry, vol. 25, pp. 562, 569, 1921. Burton, E. F., and Reid, B. M., Determination of the size of colloidal particles by means of alternating electric fields: Philos. Mag., vol. 50, pp. 1221–1226, 1925.

[64] Wentworth, C. K., The accuracy of mechanical analyses: Am. Jour. Sci., 5th ser., vol. 13, pp. 399–408, 1927.

be separated with sieves are about 0.25 millimeter in diameter [65] ("fine sand" in most systems of classification). By using the most carefully made sieves and by long-continued shaking or by sieving under water to escape the aggregating effect of hygroscopic moisture and electrical charges, particles as small as 0.05 millimeter in diameter ("very fine sand" or "coarse silt" in most systems of classification) can perhaps be separated by this method.[66] But, as most of the particles in the common fine-grained sedimentary rocks are still smaller than this lower limit, some method other than sieving must be used for their mechanical analysis.

The principle of sieving is at least theoretically applicable to much smaller sizes by the use of filters, for the pores in ordinary and hardened filter paper approximate the dimensions [67] of the limiting diameter between silt and clay. However, little or no work seems to have been done on this possible method.

MICROSCOPIC COUNTS

The method of direct microscopic counts or estimates of the proportions of different-sized grains recommends itself to most geologists because it is free from many of the uncertainties of the more indirect methods of measurement and because it involves little work other than that necessary in routine petrographic examinations. These advantages are especially great now that a method for cutting satisfactory thin sections of soft, fine-grained rocks has been developed.[68] In certain other respects, however, microscopic counts are less satisfactory. First, the method yields only approximate results unless great pains and much time are taken in making the counts. Second, it is very difficult to resolve aggregates of silt and clay-sized particles into their constituent particles, even under high magnifications. Third, errors of sampling are much greater in a very thin fragment of the rock than in a larger specimen—that is, there is much greater likelihood of obtaining a representative sample of the rock in a 10-gram fragment than in a thin section, as one is several thousand times as large as the other. Finally, it is questionable whether or not microscopic methods measure features of as great geologic significance as some of the indirect methods. If, as seems certain, most of the grains in an argillaceous rock are irregularly shaped, the processes of sedimentation and compaction have probably caused them to lie with their flatter sides parallel to the bedding. Thus thin sections cut parallel to the bedding show the two larger dimensions, and sections cut across the bedding the minimum dimension. For most geologic studies the combined effect of all three dimensions and of the surface area and density upon the settling velocity in water or in air is of greater importance than the diameter in some one plane. Nevertheless, microscopic examinations are essential as a means of verifying and calibrating the indirect methods, and for those rocks that can not be satisfactorily disintegrated microscopic examination is the only method for determining grain size.

Udden [69] recommended counting the relative numbers of grains in each size group smaller than 0.125 millimeter in diameter, calculating their weight as if they were perfect spheres, and reducing to 100 per cent. The Rosiwal method [70] of counting those grains that fall on some random straight line is familiar to most geologists. Wentworth [71] suggests projecting the microscopic image onto a ground-glass surface for greater ease in counting. Photomicrographs or camera-lucida drawings may be measured with the planimeter [72] or by direct measurement when projected upon a screen.[73] The average size of particles in different fractions (previously separated by some other method) may be determined with blood-count chambers,[74] gelatin mounts,[75] a checker-work eyepiece micrometer,[76] or the ultramicroscope.[77]

SETTLING VELOCITIES

GENERAL FEATURES OF THE PROBLEM

A grain of sand, silt, or clay falling through water has for a moment an accelerating velocity, but, the resistance of the water quickly balancing the acceleration, the velocity becomes uniform as the grain continues its fall. Large grains, of course, fall faster than small grains and attain much greater uniform velocity. The Stokes law states that among spherical grains of equal density this uniform velocity varies as the square of the diameter. This law agrees fairly

[65] Holmes, A., Petrographic methods and calculations, pt. 1, p. 204, 1923.

[66] Nutting, P. G., and Wentworth, C. K., personal communications.

[67] Ayres, E. E. jr., Subsidence in colloidal systems, in Alexander, Jerome, Colloid chemistry, vol. 1, p. 855, Chemical Catalog Co., 1926.

[68] Sayles, R. W., op. cit. Ross, C. S., op. cit.

[69] Udden, J. A., Mechanical composition of clastic sediments: Geol. Soc. America Bull., vol. 25, pp. 658–659, 1914.

[70] Lincoln, F. C., and Rietz, H. L., The determination of the relative volumes of the components of rocks by mensuration methods: Econ. Geology, vol. 8, pp. 120–139, 1913. Johannsen, A., and Stephenson, E. A., On the accuracy of the Rosiwal method for the determination of the minerals in a rock: Jour. Geology, vol. 27, pp. 212–220, 1919. Wentworth, C. K., An improved recording micrometer for rock analysis: Idem, vol. 31, pp. 228–232, 1923. Jeffries, Z., Kline, A. H., and Zimmer, E. B., The determination of grain size in metals: Am. Inst. Min. Eng. Trans., vol. 54, p. 604, 1916.

[71] Wentworth, C. K., Methods of mechanical analysis of sediments: Iowa Univ. Studies, vol. 11, p. 38, 1926.

[72] Johannsen, A., A planimeter method for the determination of the percentage composition of rocks: Jour. Geology, vol. 27, pp. 276–285, 1919.

[73] Green, Henry, A photographic method for the determination of particle size of paint and rubber pigments: Franklin Inst. Jour., vol. 192, pp. 637–666, 1921.

[74] Wightman, E. P., and Sheppard, S. E., op. cit., pp. 571–586.

[75] Renwick, F. F., and Sease, V. B., An improved method of sedimentary analysis applied to photographic emulsions: Second Colloid Symposium Mon., pp. 37–45, Chemical Catalog Co., 1925.

[76] Fry, W. H., The microscopic estimation of colloids in soil separates: Jour. Agr. Research, vol. 24, pp. 879–883, 1923.

[77] Svedberg, T., and Nichols, J. B., Determination of size and distribution of size of particles by centrifugal methods: Am. Chem. Soc. Jour., vol. 45, pp. 2910–2917, 1923.

well with experimental results on approximately spherical grains that have densities near that of quartz and diameters greater than 0.0002 millimeter [78] and less than about 0.085 millimeter [79] or 0.2 millimeter.[80] Grains with diameters greater than about 1.5 millimeters follow another rule and fall with velocities proportional to the square root of the diameter.[81] Grains of intermediate sizes, with diameters between these two limits (about 0.2 and about 1.5 millimeters—that is, essentially " sand " grains), fall with velocities that seem to follow no simple law. Although the relation between settling velocities and diameters of grains of all sizes and different shapes can not yet be stated precisely, the very marked effect upon settling velocities of slight differences in the size of small particles affords the basis for a group of methods of sizing granular mixtures. Because of the fundamental principle involved, settling in water, these methods are especially significant in studies of sedimentary rocks.

Settling velocities depend not only upon the diameter of the grains but also upon the viscosity and density of the suspending fluid, and therefore grains that fall too rapidly in water for accurate measurement or calibration may be measured in more viscous [82] or denser liquids. Similarly, particles that fall too slowly for convenient measurement may be speeded up by using a less dense or less viscous fluid or even air,[83] or else the effect of gravitation may be increased by settling within a high-speed centrifuge.[84]

The number of particles in the suspension (the concentration) is another but related factor in the settling velocity of a particle.[85] High concentration causes interference between particles, flocculation, and vertical currents, and it appreciably alters the viscosity and density of the fluid, so that in very muddy waters the rate of fall is retarded. However, experiments indicate that this factor is negligible for concentrations of less than 1 or 2 per cent.[86]

Theoretically the simple way to determine particle sizes from the different settling velocities would be to start a mixture of different-sized particles falling from the top of a fluid simultaneously, and then by some device to measure at different times the amounts that had fallen to a certain depth. Practically, however, it is almost impossible to introduce all the particles simultaneously at the top of the fluid. Slight differences in depth of particles at the time of introduction cause considerable errors in the result, if the distance between the top and bottom of the fluid is small. If the fluid is made deeper, so as to reduce the effect of these differences in original depth, particles do not accumulate on the bottom in measurable quantities for several days, and during this period changes of temperature greatly alter the viscosity of the fluid and set up convection currents that materially affect the settling velocities of the smaller particles. (See pp. 20, 28.) Also the likelihood of contamination with dust increases with a longer settling time, and the experimental work is much less convenient because spread over a longer time.

The means by which this difficulty is met may be taken as a basis for classifying the different settling methods. (*a*) In the elutriation methods a rising current of known velocity passes through a suspension of particles and carries away those particles whose settling velocity is less than the current, regardless of their original depth in the fluid. The remaining particles are then dried and weighed. (*b*) In the decantation methods no effort is made to introduce the particles at some one height, but the mixture is repeatedly stirred, settled, and decanted, thus eventually eliminating nearly all particles whose settling velocity is less than some chosen rate.[87] Here also the remaining particles are dried and weighed. (*c*) In the so-called " sedimentation " methods particles are not washed away by a rising current or by decantation, but the rate at which the sediment accumulates at the bottom or falls out of a certain part of the suspension is noted at frequent intervals. Two totally different methods for introducing the particles into the fluid are employed. Either the sample is mixed with a small amount of fluid and this very concentrated suspension is introduced carefully into the top of the settling fluid, in an endeavor to start all particles settling from some one depth simultaneously;[88] or else the sample is thoroughly mixed with all the fluid in an endeavor to attain a uniform distribution

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78 Perrin, M. J., Brownian movement and molecular reality, pp. 34–40, London, 1910.

79 Allen, H. S., Motion of a sphere in a viscous fluid: Philos. Mag., 5th ser., vol. 50, pp. 323–338, 1900.

80 Richards, R. H., Ore dressing, vol. 3, p. 1425, 1909.

81 Rittinger, P. R. von, Aufbereitungskunde, pp. 165, 325, 1867; cited by Richards, R. H., op. cit., vol. 1, pp. 472, 476.

82 Wightman, E. P., and Sheppard, S. E., op. cit., pp. 571–586.

83 Whymper, R., Analytical elutriation methods of separating cacao husk from cocoa powder: Faraday Soc. Trans., vol. 18, p. 49, 1922.

84 Briggs, L. J., Martin, F. O., Pearce, J. R., The centrifugal method of mechanical soil analysis: U. S. Dept. Agr. Bur. Soils Bull. 24, 1904. Svedberg, T., and Nichols, J. B., Determination of size and distribution of size of particles by centrifugal methods: Am. Chem. Soc. Jour., vol. 45, pp. 2910–2917, 1923.

85 Barus, Carl, Subsidence of fine solid particles in liquids: U. S. Geol. Survey Bull. 36, pp. 19–20, 1886. Richards, R. H., op. cit., pp. 610, 626–627. Robinson, C. S., Some factors influencing sedimentation: Ind. and Eng. Chemistry, vol. 18, pp. 869–871, 1926.

86 Schloesing, T., sr., Sur l'analyse mécanique des sols: Compt. Rend., vol. 136, pp. 1608–1613, 1903. Odén, Sven, On clays as disperse systems: Faraday Soc. Trans., vol. 17, p. 334, 1922. Robinson, G. W., A new method for the mechanical analysis of soils and other dispersions: Jour. Agr. Sci., vol. 12, pp. 306–321, 1922.

87 Wentworth, C. K., Methods of mechanical analysis of sediments: Iowa Univ. Studies, vol. 11, pp. 41–42, 1926.

88 Audubert, R., and Rabaté, H., Sur une méthode de détermination de la réparation granulométrique des systèmes dispersés: Compt. Rend., vol. 180, pp. 1663–1665, 1925. Werner, Donovan, A simple method of obtaining the size distribution of particles in soils and precipitates: Faraday Soc. Trans., vol. 21, pp. 388–389, 1926. Calbeck, J. H., and Harner, H. R., Particle size and distribution by sedimentation methods: Ind. and Eng. Chemistry, vol. 19, pp. 58–61, 1927.

of particles throughout the suspension. In the latter method, as also in the decantation methods, the total sediment which in any given time has fallen to a certain depth consists of large particles which have completely settled out of the suspension and of a certain amount of smaller particles which started at intermediate depths and had not so far to fall. The amount of these smaller particles can be calculated mathematically; or, more conveniently, the true composition of the sample can be determined quickly, simply, and accurately by graphic means, as is set forth more fully on pages 24–25.

ELUTRIATION METHODS

The elutriation methods have long been used for analyzing soils and clays.[89] In one modification the grains are not dried and weighed, but their volume is read on a graduated scale at the bottom of the vessel.[90] In another modification the elutriation is accomplished through siphons of different sizes hanging in the suspension.[91] Elutriation methods possess the distinct advantage that the effects of incomplete disintegration in the preliminary preparation of the sample are reduced to a minimum by the continual agitation of the particles in the current. However, these methods also have serious disadvantages. The postulated uniform rising current is essentially unattainable. Friction along the walls of the vessel may result in a velocity at the center twice as great as the average velocity.[92] Turbulent flow, induced by the conical shape of the vessel (so shaped to prevent fine grains from escaping below the inlet of the current), may completely vitiate the results.[93] Elutriation, without some centrifugal apparatus to speed up sedimentation, is essentially inapplicable to fine clays, for the low current velocities required are attainable only with great difficulty. A minor objection is that a large sample is necessary if the sample is to be separated into many sizes. The methods require special apparatus, carefully regulated and manipulated, in order to maintain currents of known velocity rising through the suspension. Especially is this true if a battery of tubes is used to separate grains of several different sizes at one time. However, medium-grained silts can be separated into a few size fractions very satisfactorily by elutriation.

[89] Hilgard, E. W., On the silt analysis of soils and clays: Am. Jour. Sci., 3d ser., vol. 6, pp. 288–296, 333–339, 1873. Baker, H. A., On the investigation of the mechanical composition of loose arenaceous sediments by the method of elutriation, with special reference to the Thanet beds on the southern side of London Basin: Geol. Mag., vol. 57, pp. 327–329, 1920. Holmes, A., Petrographic methods and calculations, pt. 1, pp. 209–215, 1921.

[90] Lowry, T. M., A new elutriator for rapid use: Faraday Soc. Trans., vol. 18, pp. 32–33, 1922.

[91] Mieczynski, T., Über eine Methode der mechanischen Bodenanalyse und der Bodendispersion: IVème Conférence internat. pédologie [Rome] Actes, vol. 2, pp. 102–116, 1926.

[92] Baker, H. A., op. cit., p. 328.

[93] Hilgard, E. W., op. cit., p. 291.

DECANTATION METHODS

Simple decantation methods[94] are particularly useful because they require no special apparatus and because the degree of disintegration and separation attained can be easily determined by simple examination at any stage in the analysis. However, decantation is very laborious. Many decantations are necessary in order to attain a satisfactory separation even into two sizes,[95] and each additional size requires an additional series of decantations. This method was used in determining the percentage of sand-sized material in the bentonite samples from the Black Hills region,[96] but it seemed too laborious to use for determining the mechanical composition of this group of shale, mudstone, and marl samples.

"SEDIMENTATION" METHODS

In the "sedimentation" methods the rate at which particles settle out of the suspension is determined by noting (1) the increasing weight or volume of the settled sediment, (2) the decreasing density of the suspension or weight of suspended material at some depth, or (3) the decreasing turbidity of the suspension.

Weight or volume of settled sediment.—Perhaps the simplest of these methods is the one in which the volume of sediment that accumulates from a thoroughly mixed suspension at the bottom of a long graduated tube is read at frequent intervals.[97] The observed rate of accumulation is plotted, and the proportions of grains of different settling velocities are determined graphically. This simple device is free from some of the errors inherent in other "sedimentation" methods, but, in addition to being liable to some more or less preventable errors, such as the effect of the walls in a narrow tube upon settling velocities, the method is open to two serious objections. First, the bulk volume of accumulated sediment depends largely upon the closeness of packing of the grains, and this closeness of packing varies greatly with the size of the grains, the time the sediment stands, and the load of overlying sediment. (See pp. 16, 34–35.) All these variables, which affect the volume of accumulated sediment, change under the conditions of Werner's method, and the calibration of his curve is therefore exceedingly difficult. Second, the method has practical as well as theoretical difficulties. In order to obtain a

[94] Fletcher, C. C., and Bryan H., op. cit., pp. 10–11. Goldman, M. I., Petrography and genesis of the sediments of the Upper Cretaceous of Maryland: Maryland Geol. Survey, Upper Cretaceous, pp. 113–120, 1916. Sauramo, Matti, op. cit., pp. 16–17.

[95] Wentworth, C. K., Methods of mechanical analysis of sediments: Iowa Univ. Studies, vol. 11, No. 11, pp. 41–42, 1926.

[96] Rubey, W. W., Cretaceous and Cenozoic formations on the northwest flank of the Black Hills: U. S. Geol. Survey Prof. Paper — [In preparation].

[97] Werner, Donovan, A simple method of obtaining the size distribution of particles in soils and precipitates: Faraday Soc. Trans., vol. 21, pp. 381–394, 1926.

readable volume of accumulating sediment and to avoid a too concentrated suspension (p. 18), the settling tube should be 3 to 10 feet long. With so great a settling distance, even the coarsest clay particles require days to settle to the bottom of the tube, and a long period of settling increases other errors (pp. 18, 28).

Results that are probably more reliable can be obtained by measuring the weight instead of the bulk volume of the accumulated sediment. In the Odén method of continuous weighing[98] the sediment accumulates on a submerged weighing pan suspended from a balance. Elaborate devices to measure this increasing weight automatically have been superimposed upon the basically simple method.[99] Like other "sedimentation" methods, this one is least accurate for the coarser grains, which fall rapidly and set up vertical currents, and for very fine grains, whose settling velocities are disturbed by convection currents[1] due to changes in temperature. The desirability of using rather brief settling periods (see pp. 18, 28) justifies the convenience of a short settling distance.

The fundamental assumption of this method—that the particles fall vertically and therefore that those which accumulate upon the weighing pan are typical of the rest—has been questioned. It has been noted that usually only 65 to 95 per cent of the expected sediment (as calculated from the proportion of the total volume of suspension that was originally directly above the pan) actually falls upon the pan and that the higher the pan above the bottom of the settling vessel the greater this discrepancy.[2] Shaw and Winterer attribute this effect to a mutual repulsion among the deflocculated particles which causes them to move toward the walls; these writers conclude that the method is probably inaccurate for particles smaller than silt. Coutts and Crowther state that the absence of sedimentation under the pan sets up currents that deflect particles from a vertical fall.

However, as discussed more fully on page 29, this loss might possibly be explained in other ways, and some of the other possible explanations would justify the inclusion of the loss with the finest fraction. Whatever the correct explanation of the loss, Coutts and Crowther's data indicate that approximately equal quantities of coarse and fine particles are lost. Therefore the fundamental assumption that those particles which accumulate upon the pan are typical of the rest still seems justified, for if large, intermediate, and small particles are equally affected, the actual amount falling upon the pan is of less importance.

The Odén method has been criticized on the ground that its accuracy depends upon the degree of dispersion of the clay.[3] This criticism applies equally to other "sedimentation," elutriation, decantation, sieving, and even to microscopic methods. The dispersion or disintegration of the original sample is a problem of preparation rather than of methods of measurement. Although the errors arising from incomplete dispersion do not form a specific criticism of the Odén method of continuous weighing, they must of course be kept in mind in deciding upon the degree of refinement justifiable in any method of measurement.

A somewhat similar method is one in which the sediment that accumulates near the bottom of the settling vessel is continuously removed for separate weighing.[4]

Density of suspension.—The decreasing density of the suspension or weight of suspended sediment may be measured (*a*) by direct weighing of the suspended matter at different depths as sampled with a pipette, (*b*) by the buoyant effect of the suspension upon a hydrometer, and (*c*) by a manometer or by the height of a fluid column of known density that balances the weight of a certain part of the suspension. A serious objection to all these methods is that they require maximum allowable concentrations of the suspension (see p. 18) in order to afford measurable quantities of suspended matter.

Pipette method: For direct weighing the suspension may be sampled at different depths and at different times with a pipette.[5] The method requires no special apparatus, and if sufficiently large volumes of suspension (and therefore long settling times) are used,

[98] Odén, Sven, Eine neue Methode zur mechanischen Bodenanalyse: Internat. Mitt. Bodenkunde, Band 5, pp. 257–311, 1915; On the size of particles in deep-sea deposits: Roy. Soc. Edinburgh Proc., vol. 36, pp. 219–236, 1916. Vaughan, T. W., Abstract of Sven Odén's work on the determination of the effective radius of particles by their rate of settling in water: Nat Research Council Comm. on Sedimentation Rept. [Apr. 18, 1923], pp. 41–49, 1923.

[99] Svedberg, T., and Rinde, H., The determination of size particles in disperse systems: Am. Chem. Soc. Jour., vol. 45, pp. 943–954, 1923. Coutts, J. R. H., Crowther, E. M., Keen, B. A., and Odén, Sven, An automatic and continuous recording balance: Roy. Soc. London Proc., vol. 106 A, pp. 33–51, 1924. Odén, Sven, The size distribution of particles in soils and the experimental methods of obtaining them; a review: Soil Sci., vol. 19, pp. 1–36, 1925.

[1] Fisher, R. A., and Odén, Sven, The theory of the mechanical analysis of sediments by means of the automatic balance: Roy. Soc. Edinburgh Proc., vol. 44, pp. 98–115, 1924.

[2] Coutts, J. R. H., and Crowther, E. M., A source of error in the mechanical analysis of sediments by continuous weighing: Faraday Soc. Trans., vol. 21, pp. 374–380, 1926. Shaw, C. F., and Winterer, E. V., A fundamental error in mechanical analysis of soils by the sedimentation method: First Internat. Cong. Soil Sci. (Washington), Commissions I and II, Abstracts of Proc., pp. 5–9, 1927.

[3] Ormandy, W. R., discussion on "Colloidal phenomena": Faraday Soc. Trans., vol. 17, pp. 366–367, 1922. Trowbridge, A. C., discussion on "Sediments and sedimentation": Nat. Research Council, Comm. on Sedimentation, Rept. [Apr. 18, 1923], pp. 54–55, 1923. Gile, P. L., Middleton, H. E., Robinson, W. O., Fry, W. H., and Anderson, M. S., Estimation of colloidal material in soils by adsorption: U. S. Dept. Agr. Bull. 1193, pp. 7, 35, 1924.

[4] Schloesing, T., sr., Sur l'analyse mécanique des sols: Compt. Rend., vol. 136, pp. 1608–1613; vol. 137, pp. 369–374, 1903.

[5] Jennings, D. S., Thomas, M. D., and Gardner, W., A new method of mechanical analyses of soils: Soil Sci., vol. 14, pp. 485–499, 1922. Robinson, G. W., A new method for the mechanical analysis of soils and other dispersions: Jour. Agr. Sci., vol. 12, pp. 306–321, 1922. Krauss, G., Ergänzender Bericht über eine * * * neue Methode der mechanischen Bodenanalyse: Internat. Mitt. Bodenkunde, Band 13, pp. 147–160, 1923.

if samples are drawn off at frequent intervals, and if proper allowance is made for the height of removed water, a very accurate distribution curve of the weights of particles with different settling velocities can be computed. However, the method is at best slow and laborious compared with some of the other "sedimentation" methods. A modification of this method is a settling tube provided with stopcocks from which the suspension can be drawn off at different depths.[6]

Hydrometer method: Measurements of the decreasing weight of suspended matter have been made with a plummet of known volume suspended from a balance at a constant depth in the suspension [7] and with a sensitive floating hydrometer.[8] Sweeping claims have been made for these methods, but accurately measurable differences in the density of a settling suspension are attained only if the suspension is very highly concentrated, and under these conditions the settling velocities are at first abnormally slow. (See p. 18.) The floating hydrometer sinks deeper as the particles settle, and hence, as it measures the weight of particles in a column of constantly changing height, it is exceedingly difficult to calibrate. Particles accumulate on the plummet of either a floating or a suspended hydrometer and weight it down, thus giving readings that indicate erroneously rapid settling velocities. In some preliminary tests the writer obtained readings at the beginning of the settling period that were much higher than the true average specific gravity of the suspension. This discrepancy, which was also noted by Schurecht,[9] may be the result of rising currents generated by the rapidly falling larger particles. In this as in other "sedimentation" methods that start with a thoroughly mixed suspension, the data, even when corrected for other errors, must still be recalculated to allow for the amount of material that falls from intermediate depths. (See pp. 19, 24–25.) The errors in these hydrometer methods seem to be so large and numerous that interpretation of the results into a size or settling-velocity curve is nearly if not quite impossible.

Manometer method: The decreasing hydrostatic pressure in a suspension may be measured by the height to which a fluid of known density rises in a manometer or side tube attached to the settling vessel.[10] With moderate or dilute concentrations the fluid in the side tube rises only a short distance above the suspension, and so can not be read very accurately. Various modifications have been proposed to increase the accuracy of the method. For example, the vertical displacement can be read more closely if the side tube is bent over nearly horizontal,[11] and a liquid of lower specific gravity in the side tube can be read more accurately.[12] Readings may be continued for days or weeks, if necessary, without fear of damaging the instrument, and, in this respect at least, these manometer methods are superior to the methods that involve continuous weighing. In fact, preliminary tests lead the writer to believe that the general principle of manometer measurement may eventually be developed into the most thoroughly satisfactory method for mechanically analyzing fine-grained sedimentary rocks. However, present methods based on this principle are especially subject to errors caused by small variations in temperature [13] and by evaporation of the liquid in the side tube; also the initial shaking of the suspension to attain the uniform distribution of particles is difficult because of the necessary side tubes.

Turbidity of suspension.—For theoretical elegance and practical complexity no other method can compare with those in which the decreasing turbidity of the suspension is measured. The opacity of a suspension at different heights and at different times can be measured quantitatively with photometric apparatus, and from the amount of light absorbed or reflected the concentration of the suspension and the size distribution of particles can be calculated. The method has been used and verified by microscopic examinations.[14] However, calculation of particle size from the turbidity of a suspen-

[6] Trnka, R., Notes sur l'analyse mécanique du sol: IVème Conférence internat. pédologie [Rome] Actes, vol. 2, pp. 116–122, 1926.

[7] Pratolongo, U., Sull' analisi fisico-meccanica dei terreni: Staz. Sper. Agr. Ital., vol. 50, pp. 117–166, 1917; cited by Odén, Sven, Soil Sci., vol. 19, pp. 21–22, 1925. Schurecht, H. G., Sedimentation as a means of classifying extremely fine clay particles: Am. Ceramic Soc. Jour., vol. 4, pp. 812–821, 1921. Mieczynski, T., Uber eine Methode der mechanischen Bodenanalyse und der Bodendispersion: IVème Conférence internat. pédologie [Rome] Actes, vol. 2, pp. 102–116, 1926.

[8] Bouyoucos, G. J., Estimation of the colloidal material in soils: Science, new ser., vol. 64, p. 362, 1926; A rapid method for mechanical analysis of soils: Idem, vol. 65, pp. 549–551, 1927: Directions for determining the colloidal material of soils by the hydrometer method: Idem, vol. 66, pp. 16–17, 1927.

[9] Schurecht, H. G., op. cit., p. 820.

[10] Wiegner, G., Über eine neue Methode der Schlammenanalyse: Landw. Vers. Sta., Band 91, pp. 41–79, 1918.

[11] Kraemer, E. O., and Stamm, A. J., A new method for the determination of the distribution of size of particles in emulsions: Am. Chem. Soc. Jour., vol. 46, pp. 2709–2718, 1924. Kelly, W. J., Determination of distribution of particle size: Ind. and Eng. Chemistry, vol. 16, pp. 928–930, 1924.

[12] Odén, Sven, The size distribution of particles in soils and the experimental methods of obtaining them: Soil Sci., vol. 19, pp. 25–30, 1925. See also Zunker, I. F., Die Bedeutung und Bestimmung der spezifischen Oberfläche des Bodens: IVème Conférence internat. pédologie [Rome] Actes, vol. 2, pp. 238–249, 1926; Crowther, E. M., A manometric apparatus for the direct determination of summation percentage curves in mechanical analysis: First Internat. Cong. Soil Sci. (Washington), Commissions I and II, Abstracts of Proc., pp. 13–18, 1927.

[13] Odén, Sven, Soil Sci., vol. 19, pp. 23, 25, 1925.

[14] Svedberg, T., and Nichols, Z. B., Determination of size and distribution of size of particles by centrifugal methods: Am. Chem. Soc. Jour., vol. 45, pp. 2910–2917, 1923. Svedberg, T., and Rinde, H., The ultra centrifuge, a new instrument for the determination of size and distribution of size of particle in amicroscopic colloids: Idem, vol. 46, pp. 2677–2693, 1924. Stamm, A. J., and Svedberg, T., The use of scattered light in the determination of the distribution of size of particles in emulsions: Idem, vol. 47, pp. 1582–1596, 1925.

sion is exceedingly difficult, because opacity depends not only upon the number of particles in suspension and the intensity and wave length of the light used but also upon the size of the particles, the difference between the refractive indices of the particles and the liquid, the color and transparency of the particles,[15] and the features of their surface that affect the amount of light reflected. Hence, a separate calibration is necessary for each substance examined. Morison,[16] who attempted to apply the method to soils, admits that interpretation of particle diameters from determinations of turbidity is virtually impossible with our present knowledge. In view of the elaborateness of the apparatus required and the uncertainty of the results obtained, it seems fairly clear that the method is not a suitable one for the mechanical analysis of fine-grained sedimentary rocks.

PRESENTATION OF DATA

Just as there are different ways of preparing the samples and of determining the proportions of large and small grains, so there are different ways of stating the results. Mechanical analyses are not an end in themselves but are made to be used for some particular purpose, and therefore the form of presentation may be quite as important as the preparation and measurement. Some methods of presentation of the data are simple but crude; others give a more complete statement but are difficult to interpret. A sample may contain 75 per cent by weight of material finer than sand. The same sample may consist of 1 per cent of gravel, 24 per cent of sand, 60 per cent of silt, and 15 per cent of clay; and each of these groups may be further subdivided into coarse, medium, and fine fractions. As the number of subdivisions is thus increased, the statement of the composition becomes more complete but more difficult to visualize. An elaborate mathematical equation might be written to express the complete size distribution, but it would be laborious to formulate and probably more detailed than the accuracy of the data would warrant, and its significance would be difficult to grasp.

DIAGRAMS

Graphic methods of presenting the data undoubtedly give the largest amount of information in a readily understandable form. Hilgard[17] illustrated his mechanical analyses of soils by photographs of the different size separates lined up side by side in test tubes. The same effect can be obtained more easily by plotting the weights of each portion in vertical columns in the order of grain size.[18] These pyramidal diagrams, or histograms, as they have been called,[19] have been widely used by geologists. They are valuable in that they give a simple visual picture of the proportions of particles having diameters between several size limits—that is, the size and degree of sorting of a sample. However, they are unsatisfactory in that the size of the limits which happen to be used affects the form of the resulting graph greatly.[20] Increasing the number of limits and so narrowing the range of sizes between limits decreases this artificial effect and eventually gives a distribution or frequency curve[21] of the sizes that is complete, but this curve, like the mathematical equation, is tedious to derive either by experiment or by calculation.

A second type of diagram that has been much used by geologists and engineers is called the cumulative curve and shows the summed percentages of grains of different sizes.[22] Particle diameters may be so plotted as to increase toward either the right or the left of the diagram. If they increase toward the right, the diagram represents the percentages by weight of grains smaller than the given diameters; if toward the left, it shows the percentages larger than the given diameters. Plotting the larger diameters at the left and the smaller at the right seems to have been the more used and is much the easier of the two procedures if "sedimentation" methods of analysis are employed. These cumulative curves are valuable because, in representing natural granular mixtures such as sedimentary rocks and soils, they are commonly smooth curves and can be interpolated with fair accuracy from a small number of experimentally determined points.[23] That is to say, the size limits that may have happened to be used in measurement scarcely affect the result; hence the cumulative curves express the size distribution of a sample rather completely and with a minimum of effort. However, until these curves become familiar they are not as easily interpreted as the pyramidal diagrams.

In diagrams of both types the particle diameters may be plotted directly, or the logarithms of the

[15] Stutz, G. F. A., and Pfund, A. H., A relative method for determining particle size of pigments: Ind. and Eng. Chemistry, vol. 19, pp. 51–53, 1927.

[16] Morison, C. G. T., The effect of light on the settling of suspensions: Roy. Soc. London Proc., vol. 108 A, pp. 280–284, 1925.

[17] Hilgard, E. W., Soils, pp. 95–96, Macmillan Co., 1906.

[18] Goldman, M. I., Petrography and genesis of the sediments of the Upper Cretaceous of Maryland: Maryland Geol. Survey, Upper Cretaceous, pp. 122–123, 169–170, 1916.

[19] Wentworth, C. K., Methods of mechanical analysis of sediments: Iowa Univ. Studies, vol. 11, pp. 46–51, 1926.

[20] Dake, C. L., The problem of the St. Peter sandstone: Missouri Univ. School of Mines and Metallurgy Bull., vol. 6, pp. 173–174, 1921. Wentworth, C. K., op. cit., pp. 47, 52.

[21] Odén, Sven, On the size of particles in deep-sea deposits: Roy. Soc. Edinburgh Proc., vol. 36, pp. 219–236, 1916.

[22] Slichter, C. S., The motions of underground waters: U. S. Geol. Survey Water-Supply Paper 67, p. 23, 1902. Dake, C. L., op. cit., pp. 156–157. Holmes, A., Petrographic methods, pt. 1, pp. 215–226, 1921. Wentworth, C. K., op. cit., pp. 49–51.

[23] Robinson, G. W., The form of mechanical composition curves of soils, clays, and other granular substances: Jour. Agr. Sci., vol. 14, pp. 626–633, 1924.

diameters may be used.[24] If the diameters are plotted directly, poorly sorted rocks require very large diagrams, but the logarithms of the diameters can be plotted much more compactly. Furthermore, whether the settling velocity or the velocity of a current competent to move a grain is considered, the logarithms are probably of greater geologic significance than the actual diameters. The difference between diameters of 0.2 and 0.1 millimeter and between diameters of 10.2 and 10.1 millimeters is in each case 0.1 millimeter, but the proportional difference between volumes, weights, or settling velocities of the two smaller grains is tremendously greater. For these two reasons all systems of size classification of rock particles known to the writer are at least crudely logarithmic or based on ratio of sizes. For example, in Udden's classification,[25] boulders have diameters from 256 to 16 millimeters, gravel from 16 to 1 millimeter, sand from 1 to $\frac{1}{16}$ millimeter, and silt from $\frac{1}{16}$ to $\frac{1}{256}$ millimeter—that is, the ratio of size limits is 16 to 1. For the same reason most geologists have used some sort of logarithmic scale in plotting the particle diameters on graphs showing mechanical analyses.

A third type of diagram is that in which the sample is divided into three fractions, such as sand, silt, and clay, and plotted on a triangular diagram.[26] This method omits much valuable information and is normally used only for special purposes.

In the methods thus far considered the proportion by weight of grains of different sizes is used. It would be more consistent with a classification based upon particle diameters to express the percentages by volume, especially if it is true that the smaller particles in soils and rocks commonly have a lower specific gravity than the larger particles and if the significant physical properties are the result of size rather than of weight.[27] However, most of the methods of mechanical analysis that have been proposed measure the weight and not the volume of the different sizes (but see pp. 19, 20) and without separate determinations of the specific gravity of the different sizes conversion of the data into volumes is impossible.

SIZE LIMITS OF SAND, SILT, AND CLAY

It is desirable to have names for the different size fractions separated, as convenient names help greatly to classify the material in a reader's mind. A rock composed chiefly of gravel is very different from one composed chiefly of clay; even "fine sand" carries a very different picture from "coarse sand." Unfortunately, however, conceptions differ as to the precise limits that separate, let us say, clay from silt.

Of the many classifications that have been proposed,[28] there are two that are widely used. Atterberg's system[29] (boulders, more than 2,000 millimeters; blocks, 2,000 to 200 millimeters; pebbles, 200 to 20 millimeters; gravel, 20 to 2 millimeters; sand, 2 to 0.2 millimeter; powder sand, 0.2 to 0.02 millimeter; silt, 0.02 to 0.002 millimeter; clay, less than 0.002 millimeter) is based upon those diameters at which notable changes in physical behavior, such as mobility in storm waves, capillary movement of ground water, and coagulation by salt water, become noticeable. This classification system, with each group divided into three fractions by limiting diameters of 20, 10, 5, 2 millimeters, etc., and with the addition of a colloid group (less than 0.0002 millimeter), is widely used by soil scientists[30] and European geologists. The other classification is that proposed by Wentworth,[31] who ascertained the actual usage of size terms by several score geologists and as a result recommended a system very similar to that used by Udden,[32] cited above. The Atterberg and Wentworth-Udden systems are similar but by no means identical.

EXACT RELATION BETWEEN SETTLING VELOCITY AND DIAMETER OF IRREGULAR PARTICLES UNKNOWN

The Stokes law is not precise when applied to sand-sized or nonspherical particles (pp. 17–18) and therefore the settling velocities of rock particles determined by elutriation, decantation, and "sedimentation" methods can not be converted into precise particle diameters. The settling velocities of different-sized particles have been determined experimentally by many investigators, but these determinations show much variation. The Atterberg scale[33] of settling velocities (the time required for a particle to settle 10 centimeters) is widely used by soil scientists, but it differs greatly from the settling velocities found by other investigators[34] and from average diameter-velocity relations recommended by Holmes[35] and Wentworth.[36]

[24] Bøggild, O. B., Samples of the sea floor: Copenhagen Univ. Min. and Geol. Mus. Contr. Mineralogy, No. 3, p. 34, 1903.

[25] Udden, J. A., Mechanical composition of clastic sediments: Geol. Soc. America Bull., vol. 25, pp. 655–744, 1914.

[26] Holmes, A., Petrographic methods, pt. 1, pp. 226–230, 1921.

[27] Bíéler-Chatelan, T., Constitution volumétrique des sols en place: IVème Conférence internat. pédologie [Rome] Actes, vol. 2, pp. 187–208, 1926.

[28] Briggs, L. J., Martin, F. O., and Pearce, J. R., The centrifugal method of mechanical analysis: U. S. Dept. Agr. Bur. Soils Bull. 24, pp. 32–33, 1904. Wentworth, C. K., A scale of grade and class terms for clastic sediments: Jour. Geology, vol. 30, p. 384, 1922. Odén, Sven, The size distribution of particles in soils and the experimental methods of obtaining them: Soil Sci., vol. 19, p. 2, 1925.

[29] Atterberg, A., Die rationelle Klassifikation der Sande und Kiese: Chem. Zeitung, Jahrgang 29, pp. 195–198, 1905.

[30] Krauss, G., Beitrag zur mechanischen Bodenanalyse: IVème Conférence internat. pédologie [Rome] Actes, vol. 2, pp. 122–126, 1926.

[31] Wentworth, C. K., A scale of grade and class terms for clastic sediments: Jour. Geology, vol. 30, pp. 377–392, 1922.

[32] Udden, J. A., op. cit.

[33] Atterberg, A., op. cit., p. 198.

[34] For a comparison of several sets of determinations see Mohr, E. C. J., Mechanische Bodenanalyse: Dept. agr. Indes Néerlandaises Bull. 41, p. 16, Buitenzorg, 1910.

[35] Holmes, A., op. cit., p. 207.

[36] Wentworth, C. K., Methods of mechanical analysis of sediments: Iowa Univ. Studies, vol. 11, p. 41, 1926.

To avoid this uncertainty, Odén converts the observed settling velocity into an "equivalent radius"[37]—the radius of a perfect sphere that theoretically would settle at the same rate as the actual particle. Robinson[38] considers the uncertainties due to shape and the possible gel envelopes surrounding the smaller particles so great that he recommends using the settling velocities themselves (at some standard temperature) or their logarithms instead of the supposed diameters.

GRAPHIC SOLUTION OF "ACCUMULATION" AND CONCENTRATION CURVES

Many of the "sedimentation" methods of mechanical analysis start with a thoroughly mixed suspension and measure either the progressive increase in weight or volume of sediment that accumulates at the bottom of a tube or the progressive decrease in concentration of the suspension. The accumulated sediment at any given time consists of two parts—(1) large particles whose settling velocities are so great that even those particles which started falling from the very top of the suspension have had time to settle to the bottom, and (2) that portion of the smaller, more slowly falling particles which started falling from such intermediate depths in the suspension that they also have had time to settle to the bottom. The amount of the smaller particles mixed with the large ones in the accumulated sediment at any moment must be determined before the true composition of the sample is known. This correction can be calculated either mathematically[39] or graphically.[40] The necessity for this mathematical correction is a feature of most "sedimentation" methods of mechanical analyses which has greatly impressed some geologists and which, on the other hand, has been overlooked by some workers in other sciences.

The graphic method of making the correction is much simpler and quicker than the mathematical calculation, and for most data it is probably quite as accurate. It depends upon the fact that in a thoroughly mixed suspension particles of any one size, being uniformly distributed, start falling from all heights and accumulate on the bottom at a uniform rate until the last particles of this size (those that started at the very top of the suspension) have settled to the bottom; that is, if the suspension is sufficiently dilute at the outset so that changes in the specific gravity and viscosity of the liquid are negligible. If particles of two or more sizes are present in the suspension, the rate of their total or joint accumulation is a steady one until all of the largest size (*A*) have fallen to the bottom, and then the total rate of accumulation decreases abruptly. The rate of accumulation immediately after all the size *A* particles have settled is the rate at which the particles smaller than size *A* had been accumulating since the beginning—that is, the previous joint rate was faster because the particles of size *A* were also settling. When the rate of accumulation of the smaller particles is thus ascertained the total amount of the smaller particles that had settled at any time can be determined by multiplication—that is, the total amount of smaller particles that accumulated with those of size *A* is the product of the rate at which they had been accumulating since the beginning and the time required for all the particles of size *A* to settle. Similarly, the amount of still smaller particles that accumulate with those of size *B* is the product of their rate of accumulation and the time required for those of size *B* to settle, and so on.

The different rates of accumulation at the times when particles of different sizes finish falling, which are required for this computation, can be determined by rather laborious mathematics. However, plotting the time and the weight of sediment accumulated at the bottom of a suspension yields a curve that shows graphically a constantly decreasing rate of accumulation. (See pl. 1, A.) Each bend in the curve represents the time at which the last particles of some particular size reached the bottom. The slope of the curve an infinitely short time after it passes this bend gives the rate of accumulation of the smaller particles that are still settling and therefore the rate at which these smaller particles have been accumulating since the beginning.

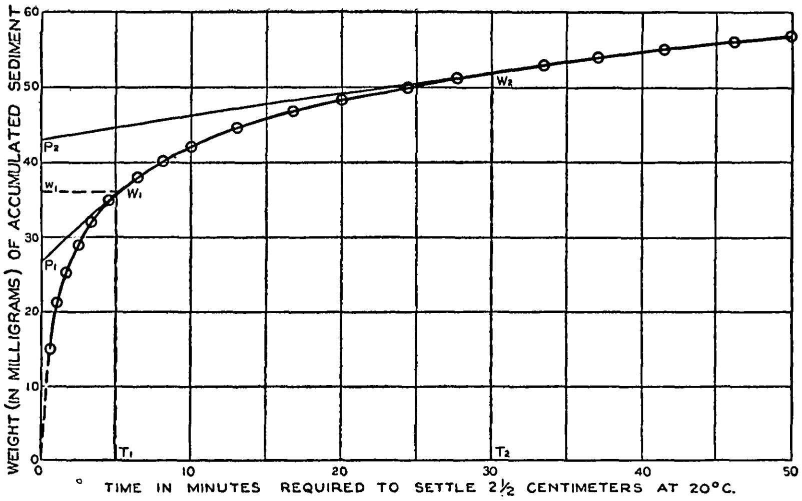
Hence, by drawing the straight line $W_1 P_1$ (pl. 1, A) tangent to one of the bends on the curve until it intercepts the vertical axis, and by putting on this axis w_1 equal to W_1, the weight of smaller particles that settled with the particular size indicated by the bend is determined. That is, this weight, represented by the line w_1P_1, equals the rate of accumulation of these particles (or, stated in mathematical terms, the derivative or slope of the curve at this point, or the tangent of the angle $w_1W_1P_1$) multiplied by the time T_1. Then P_1, the weight of larger particles that required time T_1, or less, to settle from the very top

[37] Odén, Sven, The size distribution of particles in soils and the experimental methods of obtaining them: Soil Sci., vol. 19, p. 3, 1925.

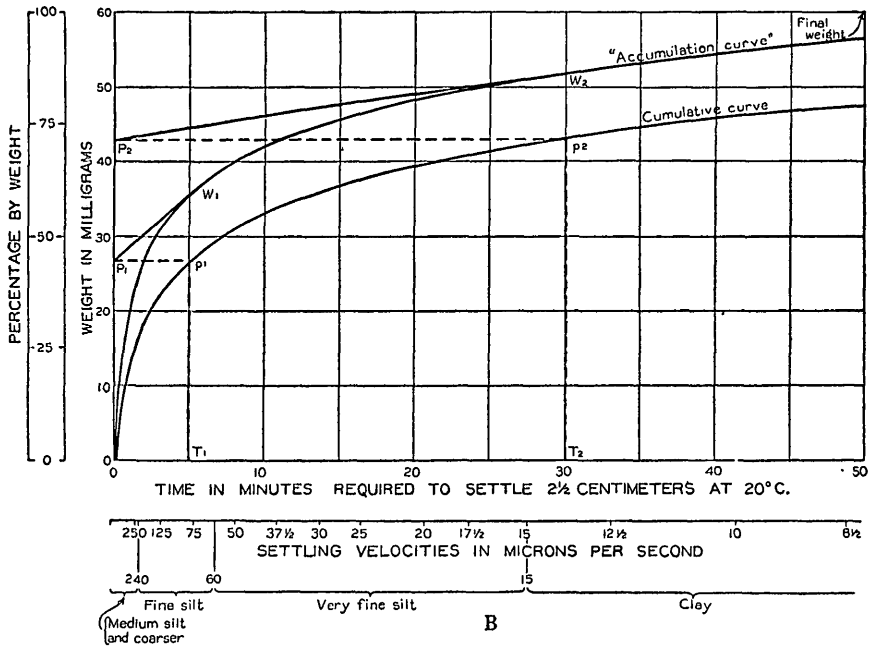
[38] Robinson, G. W., On certain regularities in the mechanical composition of soils and other granular substances: IVème Conférence internat. pédologie [Rome] Actes, vol. 2, pp. 180–187, 1926; The grouping of fractions in mechanical analysis: First Internat. Cong. Soil Sci. (Washington), Commissions I and II, Abstracts of Proc., pp. 1–2, 1927.

[39] Odén, Sven, On the size of particles in deep-sea deposits: Roy. Soc. Edinburgh Proc., vol. 36, pp. 227–230, 1916.

[40] Svedberg, T., and Rinde, H., The determination of the distribution of size particles in disperse systems: Am. Chem. Soc. Jour., vol. 45, p. 947, 1923. Schramm, E., and Scripture, E. W., jr., The particle analysis of clays by sedimentation: Am. Ceramic Soc. Jour., vol. 8, pp. 243–252, 1925. Werner, Donovan, A simple method of obtaining the size distribution of particles in soils and precipitates: Faraday Soc. Trans., vol. 21, p. 342, 1926. Calbeck, J. H., and Harner, H. R., Particle size and distribution by sedimentation method: Ind. and Eng. Chemistry, vol. 19, pp. 59–60, 1927.



A



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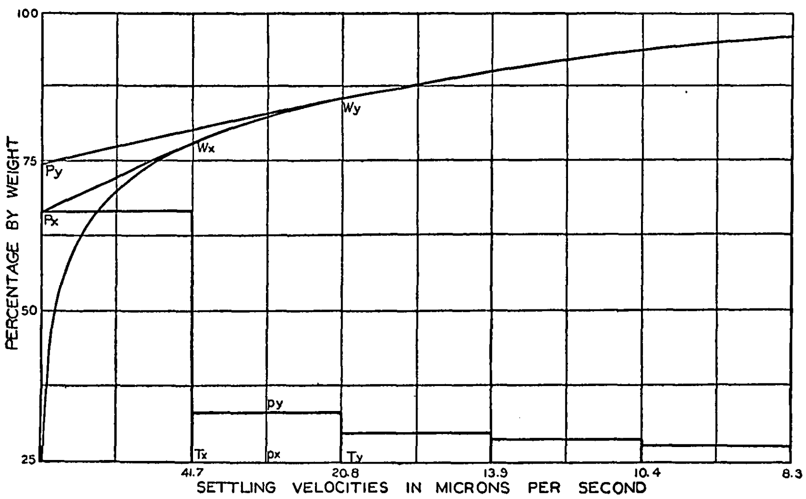


C

METHOD OF DERIVING TRUE MECHANICAL COMPOSITION OF A GRANULAR MIXTURE FROM SUCCESSIVE WEIGHTS OF ACCUMULATED SEDIMENT, WITH GRAPHIC DETERMINATION OF CUMULATIVE CURVE AND PYRAMIDAL DIAGRAM FROM "ACCUMULATION CURVE" of the suspension, equals the total weight at the time T_1 minus the weight of the smaller particles. Stated mathematically, $P = W - T \frac{dW}{dT}$. Where P=corrected weight, W=uncorrected weight, T=time, and $\frac{dW}{dT}$= change of W with respect to change of T. Similarly drawing a straight-line tangent from W_2 determines P_2, the weight of larger particles that had completely settled out of the suspension in the time T_2, and so on along the curve.

There yet remains the task of presenting the data in some readily usable form. P_2 consists of P_1 plus those smaller particles that completely settled out of the suspension in the time T_2. Now from the known total depth of the suspension the different times, T_1, T_2, etc., required for complete settling of particles of different sizes can be converted into settling velocities for the respective sizes. Also, from the final total weight of accumulated sediment the percentage of sediment with settling velocities greater than any chosen velocity can be calculated. Therefore, by changing the vertical and horizontal axes of the graph from weight and time to percentage and velocity and by plotting p_1 equal to P_1 at time T_1, p_2 equal to P_2 at time T_2, etc. (pl. 1, B), the "accumulation curve"[41] can very readily be converted into a cumulative curve showing the total percentage of particles that fall more rapidly than certain velocities. If the diameters corresponding to the different settling velocities are known, the horizontal axis of the graph can be changed from velocities to diameters, and the cumulative curve of settling velocities then becomes a cumulative curve of particle sizes or a diagram showing the percentage of particles larger than any chosen diameters.

A pyramidal diagram or histogram of the sample can be drawn with very little additional work. In section C of Plate 1, P_y minus P_x is the weight of sediment with settling times between T_x and T_y. Therefore the column p_x p_y, with a height equal to the line P_x P_y, represents the percentage of the sample between the two settling velocities or sizes indicated.

The foregoing discussion applies to the increasing weight of accumulated sediment at the bottom of a suspension. For the decreasing concentration of the suspension, determined by some method of measuring density or turbidity, the method of calculation would be the same, and the concentration curve to be used would be the "accumulation curve" inverted.

PROCEDURE ADOPTED AND RESULTS

Preparation.—After some preliminary tests upon several methods of disintegration, such as boiling in water, crushing with a wooden mallet, kneading with the fingers, digestion in dilute hydrochloric acid, and continuous agitation in water for several hours, it was decided to adopt a long period of soaking in slightly ammoniacal water, with frequent and vigorous shaking and occasional rubbing with a rubber pestle.

Clean-surfaced, nearly equidimensional fragments of the shale, weighing 1 gram at a relative humidity of 65 per cent on a balance accurate to 0.2 milligram, were placed in Florence flasks with 150 cubic centimeters of tap water and 1 cubic centimeter of 25 per cent ammonia solution. These half-filled flasks were tightly stoppered and shaken vigorously for a few seconds almost every day for eight weeks. After one month's soaking and at 1-week intervals thereafter, the fragments of mud were rubbed gently with a rubber-tipped pestle.

Under this treatment some of the samples disintegrated rapidly but others very slowly. The relative rates of disintegration of the different samples showed no relation to the content of carbonate or of water in the air-dried samples, to the porosity, or to the mechanical composition as indicated by microscopic examination and mechanical analyses. In a general way, however, the samples containing the most organic matter and the most pyrite disintegrated most slowly.

Three of the samples that disintegrated most slowly (J, H, and K) colored the water brownish; and as these samples contained relatively large percentages of organic matter, it is probable that the discoloration was caused by organic matter dissolved. Several of the suspensions formed a persistent foam when shaken. Although both the discoloration and the soapiness indicate solution of a part of the rock, neither the viscosity nor the density of the discolored or soapy liquids was changed greatly. After several months' soaking the relative viscosity of the soapiest suspension (measured in a Stormer viscosimeter) was less than 1.05 (showing an increase about equal to that produced by a drop of 1° F. in the temperature of pure water), and the density of the darkest liquid (measured in a Chainomatic specific-gravity balance) was slightly less than 1.001.

The degree of disintegration attained by this treatment was fairly satisfactory. Microscopic examination indicated that after disintegration only 5 to 25 per cent of the grains consisted of aggregates of two or more individual particles. In the final mechanical analyses corrections were made for the observed proportions of these aggregates in the different-size fractions; however, this attempted refinement may not have been worth while, for these aggregates may have settled as such when the mud accumulated on the sea floor.

In an attempt to determine the size distribution of the insoluble particles in the calcareous rocks, three of the more calcareous samples (D, E, and H) were digested in dilute hydrochloric acid (1 part HCl to

[41] Odén, Sven, Roy. Soc. Edinburgh Proc., vol. 36, pp. 227–228, 1916.

25 parts water) for one month. The acid was then decanted, after settling periods that assured the retention of all particles with settling velocities greater than 0.5 micron per second (which is much slower than the settling velocity of large clay particles), and slightly ammoniacal water was added. Microscopic examination showed, however, that this treatment did not disintegrate the insoluble particles very satisfactorily.

Measurement.—After preliminary tests of several methods of measurement, the general principle of continuous weighing was chosen as apparently one of the simplest, quickest, and most practicable ways of obtaining a moderately complete distribution curve of particle sizes. The errors inherent in this method seemed to be no greater than those due to the uncertainty about the degree of flocculation of the clay at the time of deposition, to the possible growth of clay minerals since deposition, and to the incomplete preparatory disintegration of the sample.

A disintegrated sample was washed from its flask into a beaker which was then filled with tap water to 277 cubic centimeters. The contents of the beaker were stirred thoroughly, care being taken to avoid a rotary or centrifugal motion; a platinum pan [42] was immediately introduced into the suspension, and the sediment that accumulated on this pan was weighed at frequent intervals for about 50 minutes. To check the accuracy of the method, two or more series of weighings were made on many of the samples, and on these the weighing was continued for 1½ to 18 hours.

The pan was suspended by three fine platinum wires 5 centimeters below the upper surface of the liquid and 4 centimeters above the bottom of the beaker; thus the maximum settling distance was known, and the pan was well above the sediment that accumulated on the bottom of the beaker. The diameter of the beaker was 6.3 centimeters and that of the platinum pan 3.6 centimeters, so that the formation of currents between the edges of the pan and the sides of the beaker was probably negligible. The average density of the suspension was only about 1.002 at the beginning of sedimentation and decreased to 1 as sedimentation progressed; therefore errors that might result from using highly concentrated suspensions, such as those caused by interference between particles, flocculation, vertical currents, and increased viscosity and density (see p. 18), were certainly negligible.[43] The pan was suspended from a Chainomatic balance to facilitate rapid weighing. This balance was accurate to about 0.8 milligram, and the final weight in water of the sediment that accumulated on the pan ranged from 45 to 100 milligrams, hence the readings were accurate within 2 per cent. At final weighing the sediment on the pan ranged from about 1 to 5 millimeters in thickness, indicating porosities of 95 to 99 per cent. Such thicknesses are a source of considerable error, for when the layer becomes thick fairly large quantities of sediment deposited near the edge of the pan slump off and so are not weighed.

Presentation.—The successive weight readings made during the different series of continuous weighing were plotted on large sheets of coordinate paper and corrected graphically for admixed smaller particles by the method described on pages 24–25. From the known settling distance between the top of the liquid and the pan the different periods of time on the horizontal axis of the graph were readily converted into settling velocities.

However, as a rise in temperature causes a great decrease in the viscosity of water and hence an increase in the settling velocity of particles falling through water, the observed settling velocities of different samples are not comparable unless they are reduced to some one standard temperature. The velocities on the graphs were therefore shifted by the necessary factor to reduce them to a temperature of 20°.

Decrease in viscosity of water with rising temperature [44]

	Viscosity of water		Viscosity of water
10° C	0.01309	25° C	0.00897
15° C	.01146	30° C	.00803
20° C	.01008		

From the final total weight the weights of accumulated sediment at different times were readily converted into percentages. A graph corrected in this manner is a cumulative curve of the percentages by weight of particles that fall more rapidly than certain velocities. (See pl. 1, B and C.) At the left-hand side of this graph high velocities are crowded together, but at the right low velocities are spread wide apart.

Determination of sizes of particles from settling velocities.—The exact diameter of particles can not be calculated from their settling velocities (pp. 23–24). Empirical determinations by different investigators of the diameters of particles with certain settling velocities differ greatly. However, as a check on the method used here, small samples of each of the suspensions were pipetted off at different times, and the largest of the suspended particles in each sample were measured under the microscope. Although the

[42] In accordance with the usage of some workers an aluminum pan was tried first but was found to be utterly unsatisfactory. Small gas bubbles formed on its surface in the slightly ammoniacal water, thus making accurate weighing impossible.

[43] Schloesing, T., sr., Sur l'analyse mécanique des sols: Compt. Rend., vol. 136, pp. 1608–1613, 1903. Odén, Sven, On clays as disperse systems: Faraday Soc. Trans., vol. 17, p. 334, 1922. Robinson, G. W., A new method for the mechanical analysis of soils and other dispersions: Jour. Agr. Sci., vol. 12, pp. 306–321, 1922.

[44] Original data from Poiseuille; quoted in Roth, W. A., and Scheel, K., Physikalisch-chemische Tabellen, Band 1, p. 136, Berlin, 1923.

diameters of these particles clearly decreased as the time of settling increased, the particles were so irregularly shaped that no exact relation between diameter and velocity could be discovered.

A moment's consideration shows that these discrepancies are not at all surprising. The settling velocity of a particle depends not only upon its effective weight in water but also upon the frictional resistance of the water, which in turn depends upon the shape or surface area of the particle. Among particles of a certain volume and weight but of different shapes, those that are spherical have the smallest surface area and those that are flat or rodlike have much larger areas. The spherical particles therefore fall more rapidly than the flat and fibrous ones.[45] Furthermore, microscopic measurement of the three dimensions of even those particles with known settling velocities is difficult, for if the particles are free to arrange themselves upon the glass slide, they tend to lie on their flat sides. Hence, for both of these reasons, conflicting results should be expected.

Microscopic examination of samples pipetted out from the suspensions and of thin sections of the shale cut perpendicular to bedding planes shows that in the rocks studied the constituent clay particles are platy and that their thickness is commonly only about one-seventh their length and breadth. On the assumption that the frictional resistance of water is proportional to the total surface area of the falling particles, a disk with these relative dimensions would fall at the same rate as a sphere with the same density and with a diameter only one-third as great, provided that the disk fell with its flat side always directed downward—that it never tipped up on edge and slid through the water. Obviously the platy particles do tip and fall with a zigzag path, and therefore their expected settling velocity should be greater. On the assumption, then, that the particles fall not in some one but in all possible orientations, the frictional resistance of the water would be proportional to the surface area directed vertically downward in the average of all possible orientations—that is, with an inclination of 60° ($\cos 45° \times \cos 45° = \cos 60°$) from the horizontal. On this assumption, a disk of the relative dimensions stated above would fall at the same rate as a sphere with a diameter only two-thirds as great. However, this assumption also is incorrect, as common observation of flakes of mica falling through water and of pieces of paper falling through air indicates that the average inclination is much flatter than 60° and that the clay particles probably fall with an average inclination much nearer 0° (the first assumption) than 60° (the second one). Therefore it seems probable that the settling velocity of an average clay particle from these samples of shale is the same as that of a sphere with diameter much less than two-thirds but somewhat more than one-third the maximum diameter of the clay particle. However, not all the clay particles have this average shape; some are more than 15 times as long as they are thick, and others are nearly equidimensional, so that the average relationship derived here can not be used to compute the diameters from the settling velocities.

In this discussion of the effect of shape, it has been assumed that all particles have the same density. If adsorbed films cover all the surfaces the effective density of the fine and coarse grains may not be the same, and the settling velocities would bear an even more complex relation to the diameters. Evidence that the fine grains in these samples may have a greater density than the coarse grains is given on page 33.

Because of the observed differences in shape and the possible differences in density, it is clearly impossible to calculate from the settling velocities the exact diameters of the clay particles in these samples. In fact it has been recommended[46] that the settling velocities instead of the assumed diameters be generally used in mechanical analyses. However, a statement or graph of the percentages of particles with different settling velocities does not give a very clear or easily grasped picture of the size composition of a rock, and if some of the particles are large enough to be readily measured by sieving their settling velocities are difficult to determine. Odén[47] sought to give a readily understood meaning to the different settling velocities by converting them into the "equivalent" radii of spheres having the same settling velocities.

In an attempt to avoid the uncertainty and confusion that inevitably accompanies any conversion of settling velocities into exact diameters and yet at the same time to give a readily understood interpretation of the size composition of these rock samples, the writer proposes to make a short cut from the observed settling velocities to the common names "sand," "silt," and "clay." This short cut can be made by taking advantage of the differing relations that have been reported between diameters and settling velocities and the differing concepts that have been held of just what diameters are meant by these names and more or less arbitrarily defining these names in terms of the settling velocities.

[45] Fletcher, C. C., and Bryan, H., Modification of the method of mechanical soil analysis: U. S. Dept. Agr. Bur. Soils Bull. 84, p. 13, 1912.

[46] Robinson, G. W., On certain regularities in the mechanical composition of soils and other granular substances: IVème Conférence internat. pédologie [Rome] Actes, vol. 2, pp. 180–187, 1926; The grouping of fractions in mechanical analysis; First Internat. Cong. Soil Sci. (Washington), Commissions I and II, Abstracts of Proc., pp. 1–2, 1927.

[47] Odén, Sven, The size distribution of particles in soils and the experimental methods of obtaining them: Soil Sci., vol. 19, p. 3, 1925.

If the reported settling velocities and diameters of particles are plotted on double logarithmic paper and if the size limits of the names "fine sand," "very fine sand," "coarse silt," "medium silt," "fine silt," "very fine silt," "coarse clay," and "medium clay," as defined by Atterberg, Wentworth, and Udden, are plotted to scale on the same graph, it becomes apparent that a straight line based on average settling velocity and size limit can be drawn that does not differ greatly from any of the data or size classifications. The settling velocities that delimit the different size groups on this average curve are approximately as follows:

	Settling velocity (microns per second)
Very fine sand____________more than	3, 840
Coarse silt	960–3, 840
Medium silt	240–960
Fine silt	60–240
Very fine silt	15–60
Coarse clay	3. 75–15
Medium clay	0. 9375–3. 75
Fine clay_________________less than	0. 9375

The limiting velocities between the successive size fractions decrease by the constant ratio 1 to 4. This ratio follows from the logarithmic scale of limiting diameters chosen by Atterberg, Udden, and Wentworth, and from the Stokes law (which is reliable through this range of sizes; see pp. 17–18) that the settling velocity varies as the square of the diameter.

If these somewhat arbitrary names are used for the particles with certain settling velocities, the data on the graphs showing the successive weights and the derived cumulative curves can be replotted into a more readily understood graph. On this second graph can be shown cumulative curves and pyramidal diagrams (see p. 22), also settling velocities and common names of the different fractions. (See pls. 2 and 3.) It requires slightly more than one hour to plot the readings, make the several corrections, and redraw each of these final graphs.

Accuracy of final graphs.—Duplicate and triplicate series of weighings indicate, as was to be expected from the degree of accuracy of the balance used, that for the most part the graphs are accurate within less than 2 per cent. The material represented by the part of the curve showing the most rapid velocities was tested further by sieving. After the series of weighings had been completed, several of the suspensions were washed under water through a wire sieve having square openings with a maximum or diagonal diameter of 0.09 millimeter (200-mesh). When dried and weighed, the grains of fine sand retained on this sieve were found to form only a very small proportion of every sample—a qualitative but, in view of the limits of accuracy of the measurements, not an exactly quantitative verification of the first part of the settling-velocity curves. Judged both by the repeated weighings and by these checks afforded by sieving, the curves seem to be accurate within 2 per cent. However, the effect of other possible variations that were not tested, such as the accidental choice of an unrepresentative sample and differences in degree of disintegration, may increase the limit of accuracy to 5 per cent or more.

The most uncertain part of the graphs is the weight of the final clay fraction. The final total weight of accumulated sediment on the pan might be determined in any one of three ways:

1. Weighing may be continued until all of the sediment has settled out. This complete settling may require days or weeks or even years, the time depending upon the size of the particles and the settling distance. If the settling time is long, errors creep in because of variations in temperature, which produce convection currents and changes in viscosity (see pp. 18, 20) and because of contamination with dust. Furthermore, prolonged weighing in the water-saturated atmosphere around a settling beaker is harmful to a delicate balance. Therefore it is desirable to use the shortest possible settling time that is consistent with reasonably accurate results.

2. It seems that, if the total weight of suspended sediment and the proportion of the total liquid that lay above the weighing pan were known the weight of finer particles still remaining in suspension after the coarser part had settled could be calculated by difference. It is true that a certain proportion of the rock might dissolve in the liquid and so would never settle out; but in view of the probability that the finer particles dissolve more rapidly than the larger ones and that much of the soluble material was probably adsorbed on the finer particles, it seems more or less justifiable to include whatever proportion of material has dissolved as a part of the finer fractions.[49] On the other hand, it has been noted that the amount of sediment that actually accumulates upon the pan is less than the amount that would be expected.[50] (See p. 29.) If this loss is not due simply to the solution of a portion of the original sample, the weight of the final clay fraction can not be accurately determined by difference but must be estimated by some other method. This loss in weighing is discussed more fully under the next heading.

3. An examination of the series of successive weights that extended over longer periods of time suggests very strongly that, in these particular

[49] Fletcher, C. C., and Bryan, H., op. cit., p. 13.

[50] Coutts, J. R. H., and Crowther, E. M., A source of error in the mechanical analysis of sediments by continuous weighing: Faraday Soc. Trans., vol. 21, pp. 374–380, 1926. Shaw, C. F., and Winterer, E. V., A fundamental error in mechanical analysis of soils by the sedimentation method: First Internat. Cong. Soil Sci. (Washington), Commissions I and II, Abstracts of Proc., pp. 5–9, 1927.

samples of shale, the final clay fraction can be estimated, as accurately as is warranted by the composition of the rocks and the purpose of the analysis, from a relatively short series of successive weights. This method is discussed more fully on page 30, but it may be stated here that in the writer's opinion the final weight can be estimated by this method within 5 per cent, or in any event almost certainly within less than 10 per cent of the actual weight.

Loss of sediment by the method of continuous weighing.—From the known weight and specific gravity of sediment dispersed evenly through a suspension of known volume, and from the known area and depth of a settling pan in this suspension, it seems that it might be possible to calculate the weight in water of the sediment that would eventually accumulate on the pan. However, it has been found that the amount that actually falls on the pan is usually only 65 to 95 per cent of the calculated amount. It has been suggested that this loss may be caused either by currents that deflect sideways the downward fall of particles originally above the pan[51] or by a mutual repulsion among the dispersed particles that forces them toward the walls of the settling beaker.[52]

Other possible explanations of this loss may be worth mentioning: (1) After the original weighing and during the process of disintegrating the sample, some of the sediment may be lost by adhering to the sides or stopper of the flask. (2) If the preliminary stirring is made with a rotary motion, some of the larger grains are driven centrifugally toward the walls of the settling beaker. (3) In the few seconds between the time the suspension is stirred and the time the pan is introduced into the suspension, some of the larger grains may fall below the level of the pan. (4) Some of the organic matter, moisture, alkalies, and fine particles in the original sample may become dissolved in the suspending liquid. (5) Sediment deposited near the edge of the pan may slump off, particularly if the layer of accumulated sediment becomes thick. (6) The final weighing may be made too soon, while much fine material still remains in suspension. (7) In calculating the expected weight in water of sediment, a specific gravity higher than the true one may be used inadvertently.

Most but not necessarily all of these possible sources of error would be negligible in the experiments made by Coutts and Crowther and by Shaw and Winterer. Coutts and Crowther's data indicate that the losses of silt, fine silt, and clay are about equal where the pan is appreciably smaller in diameter than the settling vessel and is more than 1 or 2 centimeters above the bottom of the vessel. These data in themselves seem to show that neither currents nor mutual repulsion are adequate explanations of the loss, because both should affect the smallest particles more than the largest. Some of the other possible explanations, which also would call for a greater loss of one size than another, are likewise inadequate.

The apparent losses in the group of shale samples studied by the writer can be estimated fairly closely. From the weight in air of the air-dried samples, the weight in water of the accumulated sediment when it seemed that all or nearly all of the suspended particles had settled, and the grain density of each particular sample, it was determined that the apparent loss ranged from 5 to 20 per cent and averaged about 10 per cent. However, the air-dried samples contained a few per cent of moisture and organic matter. The interstitial moisture would certainly not settle out of the suspension, and the discoloration of the water in which the rock was disintegrated indicates that some of the organic matter was dissolved. In fact, these percentages of apparent loss, calculated as above indicated, increased with the percentage of water and organic matter in the air-dried samples. On assuming, then, that all the contained moisture and organic matter became a part of the suspending medium and did not settle onto the pan, and correcting the specific gravity of the particles accordingly, it was found that the losses ranged from 1½ to 10 per cent and averaged about 5 per cent; that is, the calculated losses are greatly reduced but not eliminated by correcting the determination for the moisture and organic matter in the sample.

The losses over and above those caused by moisture and organic matter might have been due to any of the causes mentioned above, but, considering the surprising thicknesses of sediment that accumulated on the weighing pan (p. 26), slumping of sediment off the edges of the pan could of itself easily account for all the losses observed.

Whatever the correct explanation, the actual losses are small; and in view of the possibility or even probability that they should be included as a part of the finer fractions (see p. 28), it seems that in these samples the weight of the final clay fraction might after all be determined by difference. However, it may seem safer, and it certainly is easier, to estimate directly the final weight in water of the sediment deposited on the pan. The final weight necessary for determination by difference depends upon the weight in air, the moisture and organic content, and the grain density; hence the calculation is laborious, but this final weight can be estimated much more easily and with sufficient accuracy, as explained in the following paragraphs.

Estimation of the final weight; exponential form of "accumulation curves."—The method used here of

[51] Coutts, J. R. H., and Crowther, E. M., op. cit.
[52] Shaw, C. F., and Winterer, E. V., op. cit.

directly estimating the final weight depends upon the fact that the "accumulation curves" of the samples studied plot as straight lines on semilogarithmic paper. After some irregularities for the first few minutes, the weight of sediment was found to increase as the logarithm of the time. Obviously this rate could not continue indefinitely, but in every sample studied it held for as long as weighing was continued. (See fig. 1.)

The slopes of the different curves or the constants that relate the two variables in the different samples are not the same; but if the slope of the curve and the settling velocity of the finest clay particles are known, the final weight can be estimated well within the limit of accuracy of the rest of the mechanical analysis.

This exponential form of the "accumulation curves" means that there is a certain regularity of sequence in the proportions of the material of different settling velocities. More specifically and in geologic terms, it means that below a certain size (which coincides roughly with the lower size limit of very fine sand) there are about equal weights of the different size fractions—that is, that the shale and mudstone are essentially unsorted. (See pls. 2 and 3.)

This almost complete lack of sorting means either that after deposition the particles grew into clay minerals of many different sizes (see p. 14), or that

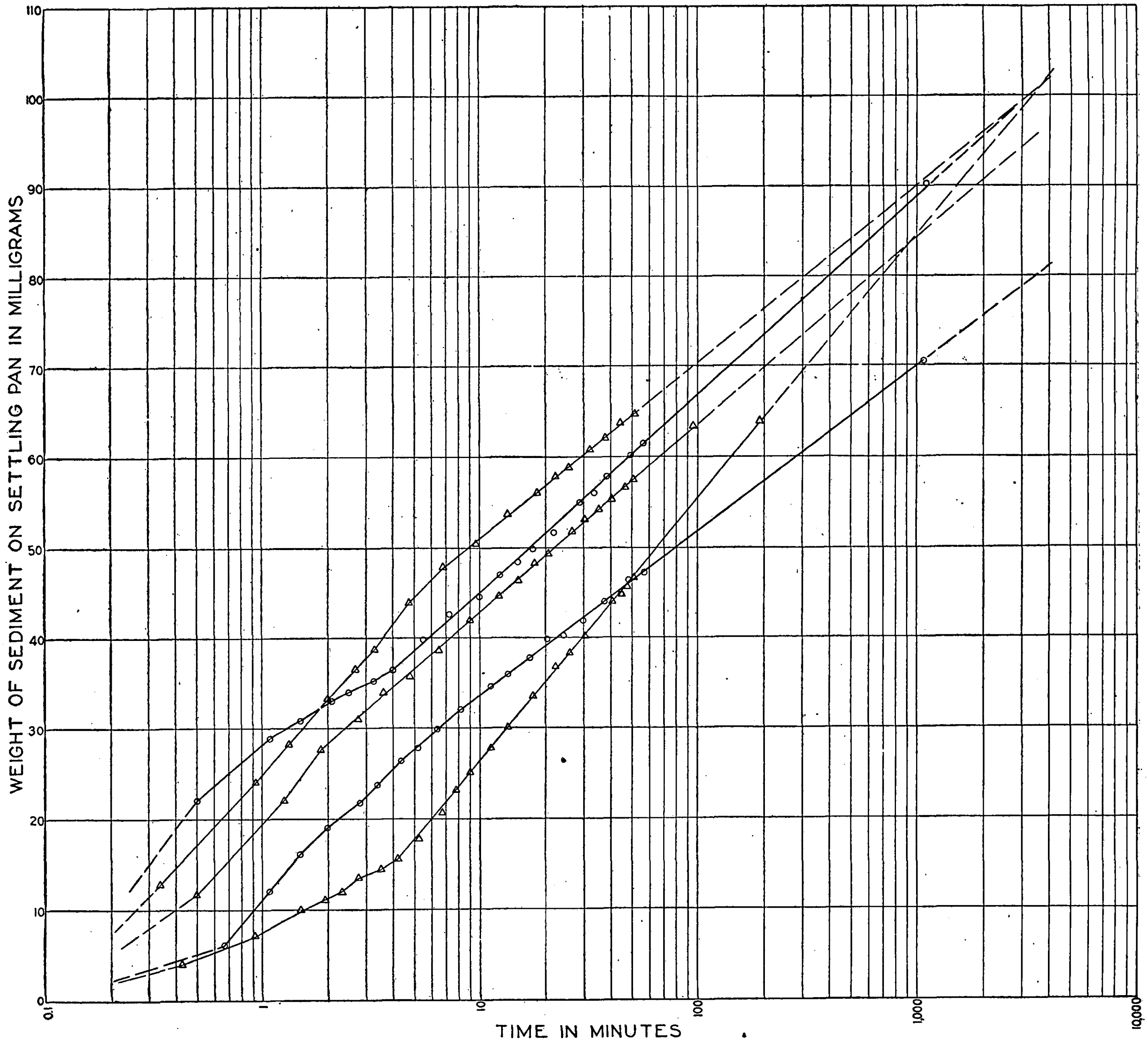
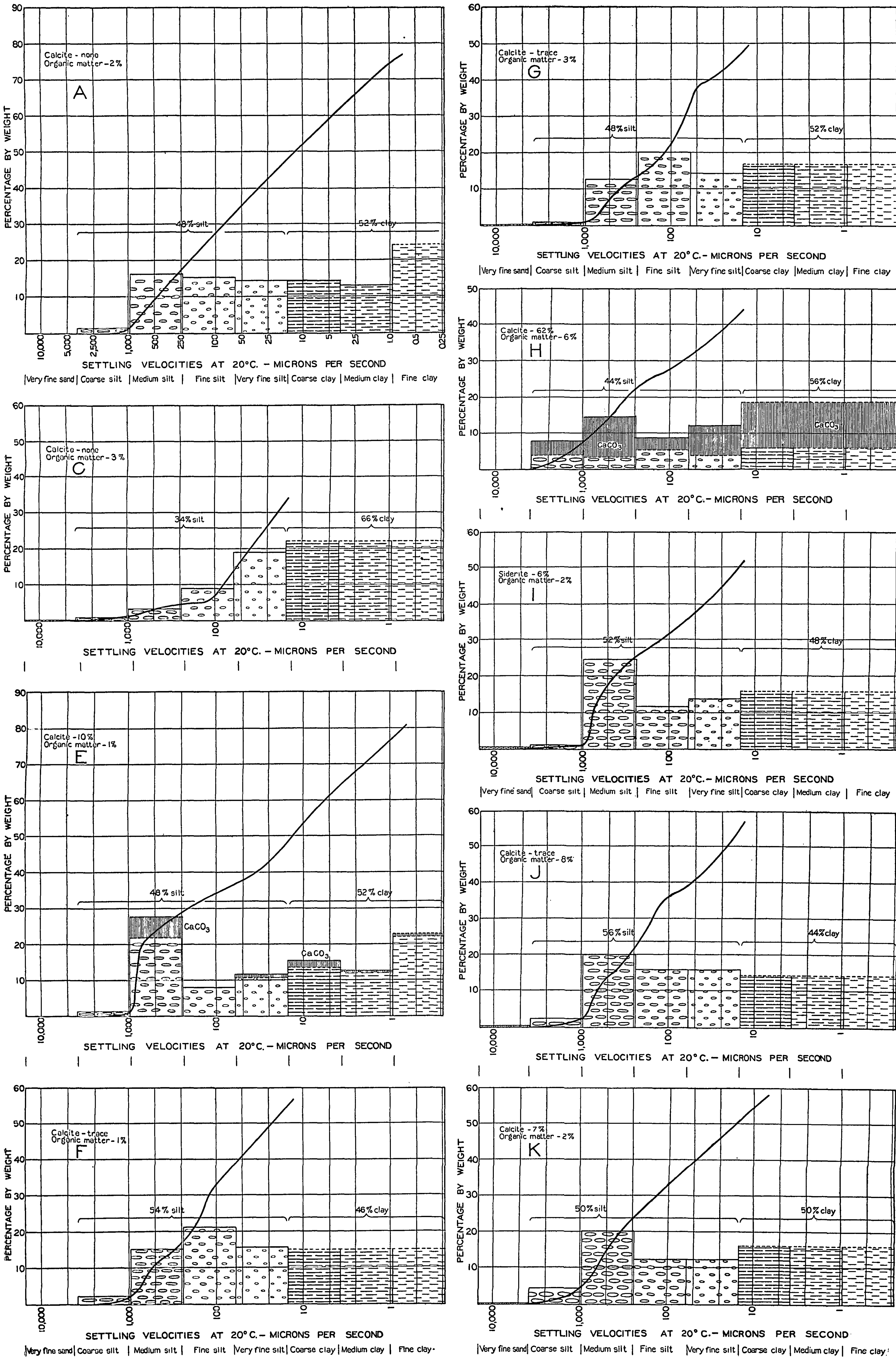


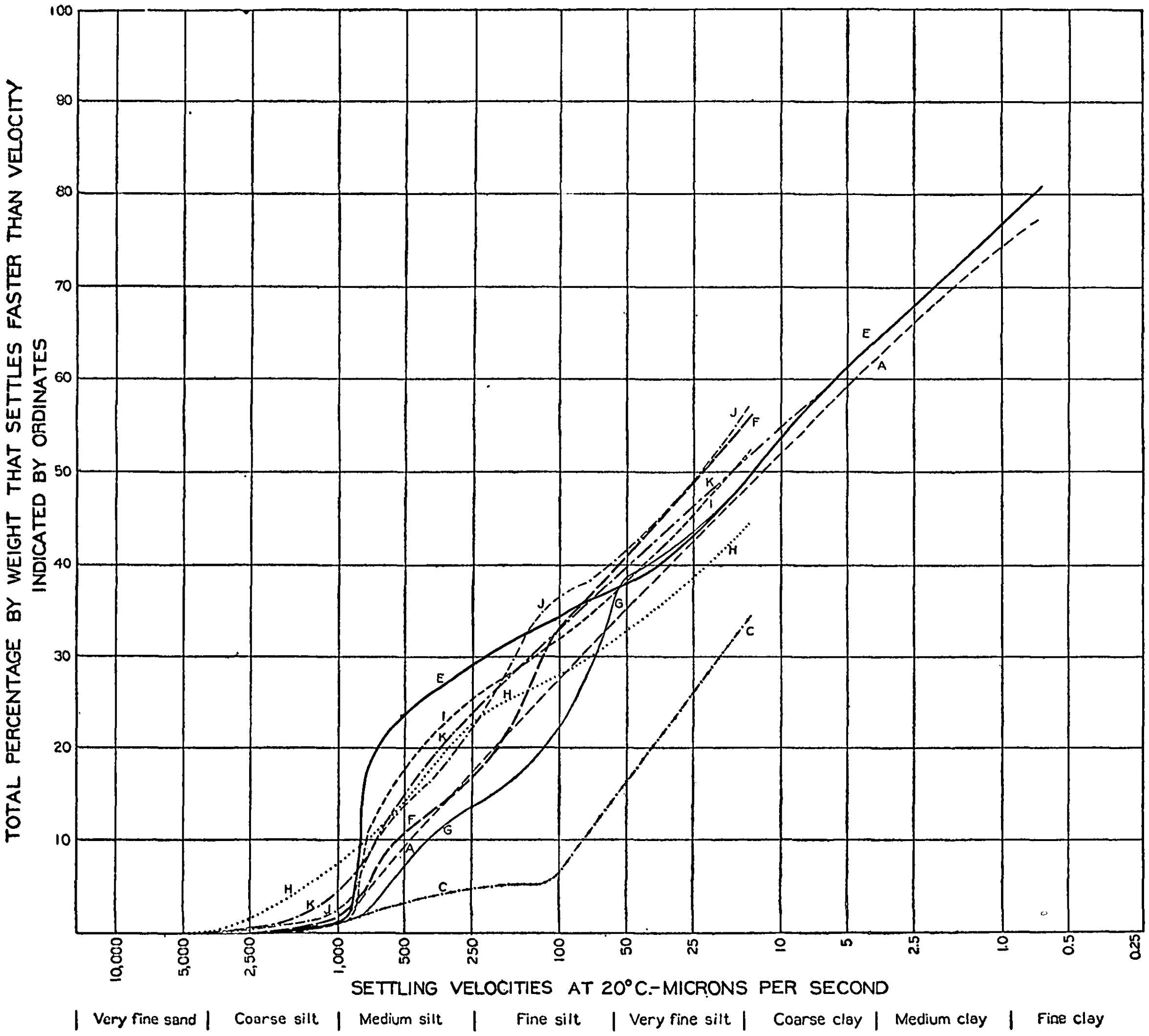
FIGURE 1.—Linear relationship between the weight of sediment on the settling pan and the logarithm of the time of settling. A graph of the actual observations made in the mechanical analysis of five rock samples from the Black Hills region. The exponential form of the "accumulation curves" here illustrated may be used to estimate the final weight of sediment



MECHANICAL ANALYSES OF UPPER CRETACEOUS SEDIMENTARY ROCKS OF THE BLACK HILLS REGION

Pyramidal diagrams and cumulative curves of each sample.

A. Black shale from Skull Creek member of Graneros shale.
C. Black shale from Belle Fourche member of Graneros shale.
E. Calcareous marl from Greenhorn formation.
F. Gray shale from upper member (Turner sandy member) of Carlile shale.
G. Gray shale from lower member (Sage Breaks shale member) of Niobrara formation.
H. Calcareous marl from upper member (Beaver Creek chalky member) of Niobrara formation.
I. Gray mudstone from lower part (Gammon ferruginous member) of Pierre shale.
J. Black shale from middle part (Mitten black shale member) of Pierre shale.
K. Gray mudstone from upper part of Pierre shale.



COMPARISON OF CUMULATIVE CURVES OF SPECIMENS OF UPPER CRETACEOUS SHALES AND MARLS FROM THE BLACK HILLS REGION

See Plate 2 for description of samples.

the detritus transported to the site of deposition and laid down there consisted of about equal amounts of silt and clay. An unsorted supply of the detritus transported to the site of deposition and a lack of sorting of the detritus during deposition might be accounted for in either of two ways: (1) Wave action might distribute unsorted mud from rivers widely over a shallow sea, and in waters that were very quiet or only gently agitated by even the most severe storms the suspended mud might settle to form an unsorted sediment or a sediment composed of alternating coarse and fine layers so thin that a small sample would appear unsorted. (2) Mud particles of many different sizes, grouped together into aggregates by flocculation in salt water, might be transported and deposited as separate aggregates, so that only those particles too large to be affected by flocculation (sand grains) would show any evidence of sorting.

Thus the lack of sorting in these rocks might be accounted for by secondary growth of clay minerals, by deposition in waters deeper than wave base, or by flocculation of different-sized particles into aggregates. It is not possible to say with any degree of assurance which of these three explanations is the most applicable to the material studied. However, the tendency toward uniform orientation and the indefinite outlines of the clay crystals (pp. 5–6), on the one hand, and the fact that the samples representing material deposited in relatively shallow water (pp. 13, 51–52, 53) are not strikingly better sorted than those representing material deposited in deeper water, on the other, suggest that neither secondary growth nor deposition below wave base adequately explains the lack of sorting. If, by elimination, flocculation is decided to be the chief factor, the possibility suggests itself that saline and fresh-water clays and shales may be distinguished by their degree of sorting.

Whatever its explanation, the lack of sorting means that the rocks consist of nearly equal amounts of silt and clay. With the single exception of sample C, which contains about twice as much clay as silt, the samples are nearly identical in their size composition, and they might with equal justification be called either "silty claystones" or "clayey siltstones."[53] For rocks as poorly sorted as these, the more inclusive terms "shale" or "mudstone" (depending upon whether or not the rock shows marked fissility[54]) are probably preferable.

Size of particles of calcium carbonate in calcareous marls.—In general, the more calcareous samples showed a slightly higher degree of sorting than the less calcareous samples. (See pl. 3.) Microscopic examination of thin sections suggested that nearly all the particles of carbonate in the marls are either of about the same diameter or of two very different diameters. It therefore seemed possible that the more perfect sorting of the calcareous samples might be caused by the abundance of carbonate grains in certain size fractions. Accordingly mechanical analyses were made of the insoluble residues of the more calcareous samples (E and H). From the known carbonate content and the mechanical analysis of the entire sample, the size distribution of the carbonate grains could then be worked out. Although the degree of disintegration of these insoluble residues was less satisfactory than that of the other samples, this comparison showed that most of the sorting of the calcareous specimens is in fact caused by the concentration of the carbonate in those size fractions that contain the greatest weight of particles; that is, the insoluble residues are essentially unsorted, but the carbonate grains tend to occur in definite sizes. (See pl. 2, E and H.)

This sorting of the carbonate grains may mean that, unlike the conditions attending the formation of the noncalcareous silt and clay, either the supply of carbonate grains was itself sorted, or that, after deposition, certain sizes grew at the expense of others, or that the carbonate grains were not flocculated but were left separate and therefore were sorted by the currents.

DENSITY AND POROSITY

DETERMINATIONS

The density and porosity of the shale samples were determined by P. G. Nutting, of the Geological Survey, who reported as follows:

SAMPLE F

Preliminary tests.—The shale works up readily in water to a mud. Even a small chip explodes violently if suddenly heated, indicating extremely minute pores (less than 0.1 micron) and considerable volatile matter. It readily dissolves in hydrofluoric acid, leaving a dark-brown submicroscopic powder. Burning a chip half an hour in a flame turned the surface layer (0.15 millimeter deep) yellowish white and the interior black, indicating an oxidation effect and organic material nonvolatile at 850° C.

	50° C.	100° C.	200° C.	280° C.	850° C.
Loss of water on heating:					
Chips of shale...per cent..	1. 65	3. 38	4. 42	5. 05	9. 65
Ground to 100 mesh.do....	1. 40	2. 65	3. 48	4. 01	7. 62
Water not recovered in 48 hours:					
Chips of shale...per cent..	. 1	. 3	1. 0	2. 1	9. 0
Ground to 100 mesh.do....	. 0	. 1	. 6	1. 4	7. 0

The powdered shale lost a somewhat smaller percentage on heating than the chips, probably because of losses on grinding

[53] Wentworth, C. K., A scale of grade and class terms for clastic sediments: Jour. Geology, vol. 30, pp. 387, 390, 1922. Twenhofel, W. H., Treatise on sedimentation, p. 186, 1926.

[54] Geikie, Archibald, Textbook of geology, vol. 1, pp. 169–170, London, 1903. Harker, A., Petrology for students, p. 216, 1919. See also footnotes on p. 39 of this report.

and sifting in an atmosphere of lower humidity (35 per cent) than the average (60 per cent).

Porosity tests, taking account of organic matter and of water in various forms: Lump volumes were determined by sand displacement, Daytona beach sand, 65–100 mesh. Grain volumes were determined by displacement of distilled turpentine, pure pinene of density 0.860 at 25° C., boiling point 158° C. This liquid adsorbs on silica and alumina much less than water. It enters pores and drives out air much more rapidly and promotes a very rapid settling of fine particles. Although free water is slightly soluble in it, adsorbed water is apparently not. The changes of density and porosity with increase in temperature to which samples had been previously heated were as follows:

Temperature (° C.)	Change		
	Grain density (gram per cubic centimeter)	Lump density (gram per cubic centimeter)	Porosity (per cent)
From 27 to 120	+0. 247	−0. 084	+8. 8
From 120 to 260	+. 072	−. 003	+1. 7
From 260 to 850	−. 405	−. 297	+1. 4

The porosity of this shale is high and dependent upon temperature. The effect of heating on the densities of lumps and of grains is considerable and significant of important internal changes.

Heating the powdered shale (finer than 100 mesh) from 27° to 120° C. increases the grain density 7 per cent, probably by removal of adsorbed water. The small change (2 per cent) from 120° to 260° is such as would be caused (in this temperature range) by dehydration of vegetable organic matter. The large decrease in density (13 per cent) between 260° and 850° C. can hardly be due to loss of material of density higher than 3, for volatile materials of such high density do not exist. The alternative conclusion is that the grains themselves lose internal water or other molecules through pores so minute that the pycnometer liquid (pinene, $C_{10}H_{16}$) can not enter them.

Heating a lump of this shale lowers its density by driving off interstitial and confined volatile matter (water?), thus decreasing its mass without appreciable change in volume. The change is 4.2 per cent between room temperature (27° C.) and 120° C., practically nil between 120° and 260° C., and 15 per cent between 260° and 850° C. The smaller change at low temperatures is probably due to loss of pore and adsorbed water. The larger change at high temperatures is probably caused by loss of combined water from hydrous aluminosilicates and organic matter. On the other hand, considerable dehydrated but oxidizable organic matter is left even at 850° C., as is shown by the dark-gray interior and by the residue on solution in hydrofluoric acid.

That is to say, as the shale is heated above room temperature its porosity increases, rapidly at first and then more and more slowly. This increase of porosity is caused partly by a slight decrease in lump density and partly by a large initial increase in grain density. Up to 260° these two changes are such that the lump density of the water-saturated specimen $\left(\text{lump density plus } \frac{\text{porosity}}{100}\right)$ is virtually constant. This indicates that the changes in grain and lump densities below 260° are due almost entirely to loss of water, which has a density of 1.0. Furthermore, the rock loses weight upon heating, but a variable part of this loss is recovered in 48 hours, and this recoverable part of the loss reaches a maximum at a temperature of about 200°. These two sets of observations indicate that the change in porosity below 200° is almost entirely caused by loss of interstitial and adsorbed water, constituents that are not an essential part of the mineral grains in the shale. Therefore 200° C. seemed to be the most satisfactory temperature at which to heat the entire group of samples before determining their porosity. The report on the full series of samples follows:

All samples were given an oven treatment of 200° for at least 16 hours to remove pore water and part of the adsorbed water, but leaving combined water to be included with grain material. Grain density was determined on material crushed to pass a 100-mesh sieve (0.15 millimeter opening), and reheated to 200° C. for at least 2 hours. Blind pores in the 100-mesh lumps are believed to be negligible when pinene is used as pycnometer fluid.

Sample	Grain density	Lump density	Pore space (per cent)
K	2. 628	1. 961	25. 4
J	2. 428	1. 559	35. 8
I	2. 755	2. 038	26. 0
H	2. 640	1. 970	25. 4
G	2. 670	1. 995	25. 3
F	2. 624	1. 999	23. 8
E	2. 795	1. 743	37. 6
C	2. 664	1. 776	33. 3
A	2. 765	1. 866	32. 5
Average	2. 66	1. 88	29. 5

The density and porosity of samples B (hard siliceous shale from the Mowry member of the Graneros shale) and D (pyritic limestone from the Greenhorn formation) were not determined.

GRAIN DENSITY

The grain density or specific gravity of the mineral fragments in the samples ranges from 2.428 to 2.795 and averages about 2.66. Comparison with the chemical analyses shows that the density depends upon the amount and kind of impurities in the rock. The sample with the lowest grain density (sample J) is the one that contains the most organic matter; the one with the highest density (sample E) is one of those that contain most calcite; and another sample with high grain density (sample I) contains much siderite.

From the chemical analyses it is possible to work backward and estimate the grain density or specific gravity of the clayey material. After taking out the proportions of moisture,[55] organic matter,[56] carbon-

[55] The moisture was determined at 105°, but the grain density at 200°. However, the resulting error in the computations is negligible. If the loss of water on heating from 105° to 200° is similar to that in sample F, the computed grain densities are too low by only a few points in the third decimal place.

[56] Density of organic matter taken as 1.5.

ates, and pyrite indicated by the chemical analyses (see p. 8), the density of the remainder of the rock can be computed. With these corrections, the computed grain density of the remaining clayey material ranges from 2.57 to 2.88. Not all these calculations are equally trustworthy, for in some of the samples the proportion of carbonates and organic matter is large, and in others it is small. If the relative validity of the different calculations is taken into account the weighted average grain density of samples that have been heated to 200° C. is about 2.72. If the grain density increases with temperature as in sample F, this average would correspond with a density of about 2.45 if the samples had not been heated above room temperature.

The corrected grain densities are not exactly (perhaps not even approximately) those of a pure clay mineral. Microscopic examination shows that some samples contain noticeable proportions of quartz sand or silt, and it is interesting to note that these sandy samples have lower density than the others. The mechanical analyses also indicate that the larger the proportion of coarse grains the lower the density.

Apparent relation between corrected grain density of the clayey material (total rock minus moisture, organic matter, carbonates, and pyrite) and percentage by weight of silt (grains coarser than clay size)

Sample	Density	Silt (per cent)
J	2. 57	56
F	2. 65	54
K	2. 66	50
C	2. 72	34
G	2. 73	48
I	2. 75	52
A	2. 82	48
E	2. 85	48
H	2. 88	44
Weighted average	2. 72	

This apparent inverse relation between the size and density of the grains is somewhat surprising, but it might be explained in any one of several ways. The determinations of grain density might involve a systematic error. For example, more inclusions of gas and liquid may be left in large particles than in small ones.[57] Or compression of the liquid film around small particles may cause the observed value of density to be greater than the true density.[58] However, neither of these possible explanations seems adequate to account for the apparent relation between size and density, and some other explanation is necessary. The clayey material (the total rock minus moisture, organic matter, carbonates, and pyrite) may consist chiefly of two ingredients—one coarse and light (perhaps quartz), the other fine and heavy[59] (possibly a clay mineral). Increasing the proportion of the fine heavy ingredient would thus give the effect noted. Or, on the other hand, the clayey material may consist of one chief ingredient, the particles of which are coated with a film of some heavier material. The finer the particle the greater its relative surface and hence the greater might be its average density. The correct explanation is not known.

A relation between the proportion of combined water and these corrected grain densities might be expected, but no such relation could be discovered.

LUMP DENSITY

The lump or rock density ranges from 1.559 to 2.038 and averages about 1.88. If the lump and grain densities vary with the temperature to which the samples have been heated, as in sample F, this average corresponds to room-temperature lump densities of about 1.96 (dry) and 2.1 (saturated with water). Corrected for the dip of the beds where sampled (as explained on pp. 35–36, the average room-temperature lump densities are about 1.7 (dry) and 2.0 (wet).

Data on the lump density of sedimentary rocks are greatly needed for estimating the weight of overburden at different depths below the surface and for interpreting the gravity anomalies, or differences between the observed gravity and that calculated by the assumption of isostatic compensation at a certain depth.[60] Particularly are these data needed near the Black Hills, as this region is one of exceptionally large and widespread positive anomalies.[61] These anomalies might conceivably have been caused by rocks heavier than the average near the surface, but these determinations show that the Upper Cretaceous shales are not heavier than the average; in fact, even without making corrections for the squeezing that some of the samples have undergone in folding (see pp. 35–38), they are unusually light. Therefore, as more than 4,000 feet of these light fine-grained sedimentary rocks underlie the gravity station at Moorcroft, Wyo., the actual positive anomaly there is even greater than the computed one. These determinations thus indicate that the basement rocks are exceptionally

[57] Sosman, R. B., The properties of silica, pp. 299–300, 304, Chem. Catalog Co., 1927.

[58] Williams, A. M., Two properties of powders: Faraday Soc. Trans., vol. 18, pp. 87–90 (especially p. 88), 1922. Harkins, W. D., and Ewing, D. T., A high pressure due to adsorption and the density and volume relations of charcoal: Am. Chem. Soc. Jour., vol. 43, pp. 1787–1802, 1921.

[59] It has been found that, among other differences in composition, the finer-sized portions of soils and clays commonly contain more of the heavy iron oxides than the portions of intermediate size. Robinson, W. O., and Holmes, R. S., The chemical composition of soil colloids: U. S. Dept. Agr. Bull. 1311, 1924. Grout, F. F., The relation of texture and composition of clays: Geol. Soc. America Bull., vol. 36, pp. 402–403, 414–415, 1925.

[60] White, David, Gravity observations from the standpoint of the local geology: Geol. Soc. America Bull., vol. 35, pp. 209–218, 1924.

[61] Idem, pp. 250–260, 275. Bowie, William, Isostatic investigations and data for gravity stations in the United States established since 1915: U. S. Coast and Geodetic Survey Special Pub. 99, fig. 7, 1924; Isostatic condition of the United States as indicated by groups of gravity stations: U. S. Coast and Geodetic Survey Serial 366, pp. 6–7, illus. opp. p. 2, 1926.

heavy or unusually near the surface in the Black Hills region, or else that the crust is overloaded there.

Sorby, Hedberg, and the writer have called attention to the fact that the porosity of argillaceous rocks commonly decreases (and hence the lump density increases) with increasing pressure or depth. The empirical relation pointed out by Sorby [62] and by Hedberg and the writer [63] can be generalized into the statement that the lump density increases as $\frac{GD+a}{D+b}$, in which G is grain density, D is depth below the present surface, and a and b are constants that depend upon the thickness of rocks eroded from above the present surface, the initial porosity of the rock, the grain density, and the extent to which the pores are filled with water.

Inasmuch as the load of the overburden increases as the integral of lump density to depth,[64] this relation means that the overburden increases as

$$GD-(Gb-a)\log\frac{D+b}{b}.$$

Argillaceous rocks are very abundant at the earth's surface, and empirical relations such as that above set forth, if they can be verified, will be useful as a means of estimating the distribution of density in the upper few miles of the earth's crust.[65] However, the lump density of a rock depends upon both the grain density and the porosity, and if the grain densities in a series of samples are not constant these possible relations can be tested most simply by an examination of the porosity.

POROSITY

The porosity, or percentage by volume of pore space, in these samples previously heated to 200° C. ranges from 23.8 to 37.6 and averages 29.5. If the porosity varies with the temperature to which the sample has been heated, as in sample F, this average corresponds to a porosity of about 20 per cent at room temperature. This large difference is due almost entirely to interstitial and adsorbed water in the shale sample (p. 32) at room temperature, and the porosity of a sample that has been heated to 200° gives a much more accurate picture of the volume composition of the rock. Corrected for the dip of the beds where sampled (see pp. 35–36), the average porosity at room temperature of the rocks before tilting is about 30 per cent.

This rather high porosity and the fact that the samples were collected from surface outcrops suggest that they may be weathered and cracked. However, this seems unlikely, for microscopic examination of thin sections cut from other fragments of the same samples disclosed neither weathering nor minute fracturing, and the fragments tested for porosity were carefully chosen. Also, Hedberg's shale samples,[66] taken from well cuttings and outcrops near wells in Kansas, have porosities that fall into a smooth curve —that is, the porosities of the outcrop samples are not disproportionately higher than those of the deeper samples.[67] This indicates that in western Kansas, at least, the porosities of the outcrop samples are as reliable as those of the deeply buried samples. Furthermore, the fact that the porosities of the samples from the Black Hills region show expectable and even quantitative relations to other variables (pp. 35–38) suggests that they are essentially correct.

In a coarse or medium-grained sandstone with 30 per cent porosity water and oil can circulate with comparative freedom, but in a shale with the same porosity the pores are much smaller, and because of surface tension and friction the liquids can scarcely move. Thus, although the shales in this region seem to have higher porosity than the oil-bearing sandstones,[68] they can still, where unfractured, act as impervious layers and confine the oil to the sandy beds. Where the shales are fractured they may serve as reservoirs for oil. The writer estimated that one-fifth of the oil produced in the Osage oil field in 1923 came from fine-grained shale beds in the Belle Fourche, Mowry, and Skull Creek members of the Graneros shale; and oil has been obtained from fractured shale beds in many oil fields in the Rocky Mountain States.[69] However, the openings from which this oil is recovered probably have no relation to the porosity of small samples of the shale.

The porosity of an argillaceous rock depends upon many factors, chief among which are the size and shape of the grains, the porosity at the time of deposition, and the amount of squeezing or compacting the rock has undergone since deposition. The presence of many small and flat grains in a rock apparently increases its original porosity. Experiments have shown that both the porosity of different samples of clay and claylike aggregates at equal pressures and the loss of the porosity of these samples with equal increments of pressure increase with increasing flat-

[62] Sorby, H. C., On the application of quantitative methods to the study of the structure and history of rocks: Geol. Soc. London Quart. Jour., vol. 64, pp. 227–231, 1908.

[63] Hedberg, H. D., The effect of gravitational compaction on the structure of sedimentary rocks: Am. Assoc. Petroleum Geologists Bull., vol. 10, pp. 1057–1058, 1926. Rubey, W. W., The effect of gravitational compaction on the structure of sedimentary rocks—a discussion: Idem, vol. 11, pp. 621–632, 1334, 1927.

[64] Nutting, P. G., The deformation of granular solids: Washington Acad. Sci. Jour., vol. 18, pp. 123–126, 1928.

[65] Williamson, E. D., and Adams, L. H., Density distribution in the earth: Washington Acad. Sci. Jour., vol. 13, pp. 413–428, 1923.

[66] Hedberg, H. D., op. cit., p. 1052.

[67] Rubey, W. W., op. cit., pp. 626–627.

[68] Collier, A. J., The Osage oil field, Weston County, Wyo.: U. S. Geol. Survey Bull. 736, pp. 96–98, 1922.

[69] Wegemann, C. H., The Salt Creek oil field., Wyo.: U. S. Geol. Survey Bull. 670, pp. 36–37, 1918. Estabrook, E. L., and Rader, C. M., History of production of Salt Creek oil field, Wyo.: Petroleum Development and Technology in 1925, pp. 209–211, Am. Inst. Min. Eng., 1925.

ness and smallness of the constituent particles.[70] The porosity at the time of deposition probably depends upon the conditions of sedimentation, the salinity of the water, and the size, shape, and degree of sorting of the constituent particles in the mud.[71] (See pp. 16, 19, 26.) After deposition, the mud is squeezed or compacted, either by folding or regional alteration[72] or by the weight of overlying rocks.[73]

The porosity of sandstones in this region seems to decrease with depth. A diamond-drill core at Osage, Wyo., through the Newcastle sandstone, Skull Creek shale, and Fall River or so-called Dakota sandstone shows porosity decreasing downward.[74]

Of these various relations, the most obvious one shown by the porosity of the shale samples studied is the relation to the degree of deformation that the rocks have undergone. In general, the more steeply dipping rocks have the lower porosity, as indicated in the following table:

Relation of porosity to degree of deformation

Sample	Degree of deformation		Porosity (per cent)
	Dip of rocks where sampled (°)	Proximity to faults	
E	1	--------	37. 6
C	4	--------	33. 3
A	5	--------	32. 5
I	5	(a)	26. 0
K	7	(a)	25. 4
H	10	(a)	25. 4
F	33	--------	23. 8
J	45	--------	35. 8
G	50	--------	25. 3

a Near faults.

This decrease of porosity with increase of dip is of considerable structural significance, for it shows that the shale beds were not simply tilted by folding but were squeezed and deformed internally. That is to say, they behaved as incompetent rocks and hence very probably formed similar, not concentric or parallel folds. This interpretation furnishes a theoretical basis for a quantitative statement of the decrease of porosity with increase of dip. The porosity and hence also the relative volume of the different samples vary roughly as the cosine of the angle of dip. Now if we

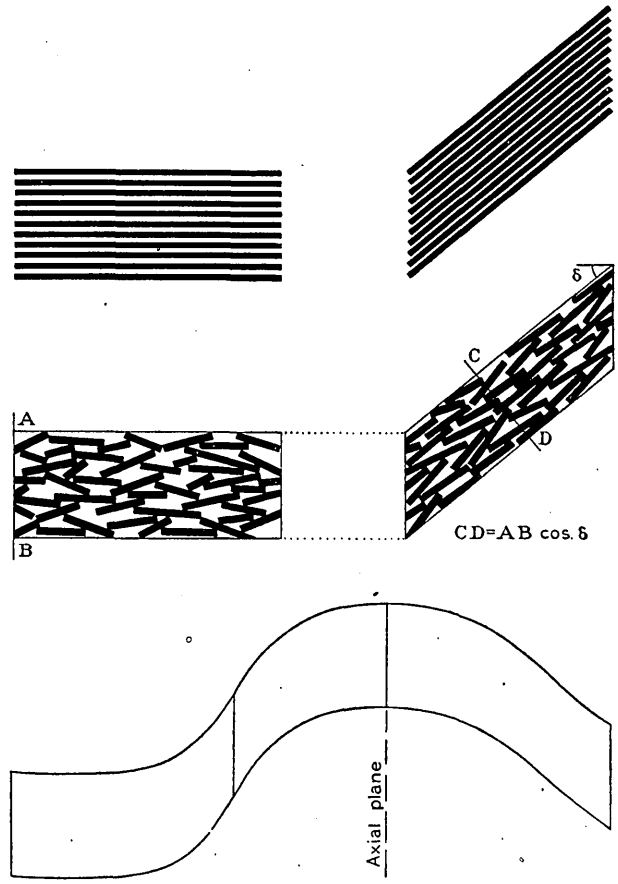


FIGURE 2.—Probable explanation of decrease of porosity with increase of dip. Whether the particles touch loosely or are held apart by cushions or envelopes of adsorbed films that cover the surface of each particle, the pores are squeezed flatter as the beds are tilted. The total volume of the pores and the stratigraphic thickness of the bed decrease, but the length of the bed and its vertical thickness remain essentially unchanged. Deformation of this type would form similar folds with vertical axial planes

assume the simplest possible type of similar folding (horizontal shortening and vertical axial planes)[75] this is exactly the relation that we would expect to find. As the beds are tilted, the spaces or pores between the clay particles are squeezed flatter, the contained fluids being squeezed out, so that, until all pores have been closed up, the thickness of the bed decreases as the cosine of the angle of dip. (See fig. 2.)

[70] Hardy, F., The physical significance of the shrinkage coefficient of clays and soils: Jour. Agr. Sci., vol. 13, pp. 243–264, 1923. Russell, J. C., and Burr, W. W., Studies on the moisture equivalent of soils: Soil Sci., vol. 19, pp. 251–266, 1925. Wintermeyer, A. M., Adaptation of Atterberg plasticity tests for subgrade soils: Public Roads, vol. 7, pp. 119–122, 1926. Terzaghi, C., Simplified soil tests for subgrades and their physical significance: Idem, pp. 153 et seq. Hedberg, H. D., op. cit., pp. 1049–1050. Rubey, W. W., op. cit., p. 624. Gilboy, Glennon, The compressibility of sand-mica mixtures: Am. Soc. Civil Eng. Proc., vol. 54, pp. 555–568, 1928. Rubey, W. W., The compressibility of sand-mica mixtures [discussion]: Idem, pp. 1936–1938. Also unpublished data furnished to the writer by E. F. Kelley and C. A. Hogentogler, of the Bureau of Public Roads, Department of Agriculture, show that a close relation exists between the porosity and total clay and silt content of soil samples.

[71] Hedberg, H. D., op. cit., pp. 1039–1042. Rubey, W. W., op. cit., p. 1334. Barus, Carl, Subsidence of fine solid particles in liquids: U. S. Geol. Survey Bull. 36, pp. 31, 35, 1886. Pickering, S. U., Flocculation: Roy. Soc. London Proc., vol. 94A, pp. 315–325, 1918.

[72] Hedberg, H. D., op. cit., pp. 1043, 1071–1072. Wilson, J. H., Lithologic character of shale as an index of metamorphism: Am. Assoc. Petroleum Geologists Bull., vol. 10, pp. 625–633, 1926. Russell, W. L., Porosity and crushing strength as indices of regional alteration: Idem, pp. 939–952.

[73] Sorby, H. C., op. cit., pp. 227–231. Terzaghi, Charles, Principles of soil mechanics: Eng. News-Record, vol. 95, pp. 742–746, 796–800, 832–836, 874–878, 912–915, 987–990, 1026–1029, 1064–1068, 1925. Hedberg, H. D., op. cit., pp. 1035–1072.

[74] Collier, A. J., The Osage oil field, Weston County, Wyo.: U. S. Geol. Survey Bull. 736, pp. 80, 97, 1922.

[75] Rubey, W. W., Determination and use of thicknesses of incompetent beds in oil-field mapping and general structural studies: Econ. Geology, vol. 21, pp. 334–339, 1926.

Although at first thought this interpretation may seem to conflict with the principle of dilatancy [76] or the expansion of granular masses when deformed, more careful consideration shows that there is no such conflict. Granular aggregates, like clay, that consist of very small particles have a much wider range of porosity than aggregates of large spheres, perhaps because of the poor sorting, angularity, and flatness of the clay particles and the greater proportionate thickness of the deformable adsorbed layer on the particles.[77] Aggregates with porosities greater than the minimum porosity or, more accurately, aggregates with the so-called open-packing porosity are deformed plastically and without increase of volume.[78] Hence clays with porosities greater than the minimum limit, which in clays is probably only a few per cent, are not dilatant but can be deformed plastically and by actual decrease instead of increase of volume.

The rate at which the rocks are deformed and the amount of overburden at the time of deformation are probably additional factors in determining whether or not rocks are dilatant. Under light load and sudden deformation even the most plastic clay is likely to fracture and shear and so increase its volume. But it is possible that under heavy load and very slow deformation even indurated and brittle shale may adjust itself internally so that its volume is decreased instead of increased.

If the proposed interpretation is correct, the porosity of the samples before they were tilted can be determined, because the porosity after tilting depends upon the total volume of pores and mineral grains, and this total volume in turn depends upon the angle of dip and the original porosity. (See fig. 2.) Let

$P_u = 100\frac{V_u}{V_u+s}$ = porosity of the untilted rock.

$P_p = 100\frac{V_p}{V_p+s}$ = present porosity after tilting.

d = present angle of dip.

s = volume of solid particles or mineral grains in rock.

V_u = volume of voids or pores between solid particles before tilting.

$V_p = \frac{P_p s}{100-P_p}$ = volume of voids after tilting.

And

$V_p + s = (V_u+s) \cos d$ = assumed volume of rock after tilting.

Then

$V_u + s = \frac{V_p+s}{\cos d}$ = original volume of rock.

$$V_u = \frac{V_p+s}{\cos d} - s.$$

$$P_u = 100\frac{V_u}{V_u+s} = 100\frac{\frac{V_p+s}{\cos d}-s}{\frac{V_p+s}{\cos d}} = 100\frac{V_p+s-\cos d s}{V_p+s}$$

$$= 100\frac{\frac{P_p s}{100-P_p}+s-\cos d s}{\frac{P_p s}{100-P_p}+s}$$

$$= 100\frac{P_p+100-P_p-\cos d\,(100-P_p)}{P_p+100-P_p}$$

$$= 100-\cos d\,(100-P_p)$$

Applying this equation shows that, as might be expected, the corrected porosities tend to decrease with depth, as indicated in the following table, and the exceptions to this regular decrease seem to be explainable as the result of other disturbing influences. Samples K, I, and H were collected near faults, and hence may be assumed to have undergone more deformation than the dip of the rocks alone would indicate. The slight departure of sample E or F from this regular decrease and other irregularities not apparent unless the porosities are plotted seem to be caused by differences in texture. (See p. 38.)

Apparent decrease of porosity with increase of stratigraphic depth

Sample	Porosity of untilted rock (present porosity corrected for dip; per cent)		Approximate stratigraphic depth below horizon of sample J (feet)
	Samples near faults	Other samples	
K	26+		
J		54.6	0
I	26+		
H	27+		
G		52.0	1,300
F		36.0	1,550
E		37.6	1,850
C		33.4	2,100
A		32.8	2,900

In other areas the porosity and depth of shale samples from wells are related to one another in such a way that it seems possible to estimate the thickness of rocks once present but now eroded from above the highest sample.[79] This method of estimation has not been adequately tested, and even if it is valid, it might not apply to the porosity of a few outcrop samples taken from a large area in which the thickness of overlying formations and the dip vary considerably. Nevertheless, when applied to the samples from the Black Hills region the method yields results that seem to be qualitatively correct. These results therefore tend to justify both the method of estimating maximum overburden and the interpretation of the relation between porosity and dip.

[76] Mead, W. J., The geologic rôle of dilatancy: Jour. Geology, vol. 33, pp. 685–698, 1925.
[77] Idem, p. 686.
[78] Idem, pp. 692, 697.
[79] Rubey, W. W., The effect of gravitational compaction on the structure of sedimentary rocks—a discussion: Am. Assoc. Petroleum Geologists Bull., vol. 11, pp. 625–628, 1927.

Calculated by this method from the corrected porosities and the depths below sample J, the thickness of overburden eroded from above the horizon of sample J may have been about 3,300 feet. This calculated thickness, compared with the average thickness of overlying formations that crop out farther west, is equivalent to that of all of the overlying marine Upper Cretaceous and continental Lance formations and part of the continental Fort Union formation.

However, the thickness of the formations that overlie the bed yielding sample J varies considerably from place to place, and hence the relative depths of the different samples below the higher formations are not the same as their relative depths below the horizon of sample J. The thickness of overburden should be calculated from the depths of the different samples below the highest bed once present in the area, but as the exact stratigraphic position of this highest bed is not known, the thickness must be calculated from depths below several of the higher formations. Under this procedure the method of estimating the thickness of overburden loses its false appearance of accuracy, for no two calculations give the same result. However, they all agree in indicating that the top of the maximum overburden lay somewhere between the base of the Lance and a horizon several thousand feet above the base of the Wasatch formation. That is to say, they all indicate that uplift and erosion began, or, more accurately, that sedimentation above the site of the samples ceased, sometime during the deposition of the continental Lance, Fort Union, or Wasatch formations or possibly somewhat later. These estimates are based, of course, upon the thickness of the formations as they crop out some distance west of the Black Hills uplift, and if before erosion these formations were thinner east of their present outcrops, the estimates would correspond with somewhat later epochs than those given.

Field evidence and regional relations indicate that, for the most part, the Black Hills rose after part of the Wasatch formation had accumulated; but uplift may have started much earlier and continued somewhat later than the time of maximum uplift. It is thus seen that the calculated thickness of overburden agrees with the field evidence in a general qualitative way but not with quantitative exactness.

A further test can be made both of the method of estimating the maximum overburden and of the interpretation of the relation between porosity and dip. If this method and this interpretation are both essentially correct, porosities calculated from them should differ from observed porosities in accordance with other disturbing influences such as variations in rock texture. The calculated porosities necessary for this comparison might be obtained by using those figures for B (eroded depth) and C (the constant) in the modified Sorby depth-porosity equation [80] that best fit the data. But a more independent method, which does not demand so much pyramiding of poorly established relations and which uses only assumptions that are based upon independent evidence, is somewhat more convincing.

By starting with the assumption that the porosities decrease with depth according to the general form of the modified Sorby depth-porosity equation (an assumption based on studies made outside of the Black Hills region) and obtaining the one constant necessary for the use of this equation from field evidence that is independent of the porosity data, all porosities can be reduced to one depth.

The constant B can be eliminated algebraically.

$$(D+B)\frac{P}{100-P}=C \qquad (1)$$

in which D is depth below the surface, B is thickness of rocks eroded from above the surface, P is porosity, and C is a constant. Then

$$P=\frac{100\,C}{B+C+D} \qquad (2)$$

$$D=\frac{100\,C}{P}-B-C \qquad (3)$$

Let P_1=porosity at depth D_1, and P_2=porosity at some greater depth, $D_2=D_1+x$.

Then

$$P_2=\frac{100C}{B+C+D_2}=\frac{100C}{B+C+D_1+x}=\frac{100C}{B+C+\left(\frac{100C}{P_1}-B-C\right)+x}=\frac{100CP_1}{100C+P_1x} \qquad (4)$$

The remaining constant, C, might be obtained from data on series of porosities from other regions. The value of this constant is relatively unimportant, for throughout a wide range the precise value chosen for it does not affect the relative order of calculated porosities. However, in order to make the figures concrete, the constant can be evaluated by introducing the assumption, based on field observations, that sedimentation above the rocks represented by these samples ceased about the time that Wasatch deposition began farther west. The average of the nine values of C obtained from this assumption is approximately 2,200, and, by putting this average value into equation (4), the theoretical porosities (uncorrected for dip) that might be expected at a depth of 7,000 feet below the base of the Wasatch formation can be computed.

Then by introducing a third assumption that the stratigraphic thickness of a bed of shale decreases as

[80] See footnotes 62 and 63.

the cosine of the angle of dip (an assumption that can be more or less justified by observed variations in the thickness of folded shale beds in this region), the effect of dip on porosity can be eliminated by the equation, $P_u=100-\cos d\ (100-P_{7000})$. See p. 36.)

The resulting porosities, now corrected for depth and dip, show a general increase with increasing percentage of silt (that is, the coarser) grains.

Apparent relation between porosity and grain size of samples

Sample	Actual porosity of samples Po	Approximate depth of sample below base of Wasatch formation (feet) Dw	Interval in feet of sample above bed 7,000 feet below base of Wasatch formation $x=7{,}000-Dw$	Computed porosity of rock 7,000 feet below base of Wasatch formation [a] $P_{7{,}000}=\frac{100CP_o}{100C+P_ox}$	Dip of rocks where sampled d	Computed porosity of untilted rock 7,000 feet below base of Wasatch formation $P_u=100-\cos d$ $(100-P_{7{,}000})$	Percentage of silt (grains coarser than clay size)
G	25. 3	5, 450	1, 550	21. 5	° 50	49. 5	48
J	35. 8	4, 200	2, 800	24. 6	45	46. 5	56
F	23. 8	5, 700	1, 300	20. 9	33	33. 7	54
A	32. 5	7, 000	0	32. 5	5	32. 8	48
E	37. 6	4, 900	2, 100	27. 7	1	27. 7	48
C	33. 3	5, 200	1, 800	26. 2	4	26. 4	34
H	25. 4	5, 250	1, 750	21. 1	[b] 10	22+	--------
I	26. 0	4, 850	2, 150	20. 7	[b] 5	21+	--------
K	25. 4	3, 550	3, 450	18. 2	[b] 7	19+	--------

[a] The constant C was taken as 2,200, the average of the values computed from the assumed general depth-porosity equation, $C=Dw\frac{Po}{100-Po}$

[b] Near faults.

In general, the finer-grained samples (those containing the smaller and flatter particles) have been compacted more than the coarser-grained samples. Inasmuch as the porosities show the relation to differences of rock texture or grain size that might have been expected on theoretical grounds, this test tends to further strengthen the proposed methods of correcting for dip and depth that were used.

The effects of differences of dip, depth, and rock texture upon porosity have been treated more or less quantitatively in the preceding paragraphs. However, the writer realizes fully that he has not established any quantitative relations by this treatment. Far more observations and experimental data than are now available would be required to determine the precise effect upon porosity of any one of the several variables involved. On the other hand, this method of treatment seems to show that the proposed explanation of a decrease of porosity with an increase of dip is not qualitatively improbable—that, within the range of porosities and dips examined, it leads to results that are of the correct order of magnitude.

If this proposed explanation is essentially correct, some interesting corollaries follow. For example, the porosity and thickness of a bed of shale decrease as the bed is tilted. This decrease of thickness represents a decrease of volume; it is not compensated by an increase in the length of the bed, for that would pull the rock particles apart and maintain the original porosity. That is, the steeply dipping beds have not been stretched by vertical uplift;[81] they have been tilted and bent into a narrower compass without important change of length. In short, they have been subjected to horizontal compression. Another corollary is that, inasmuch as the thickness measured at right angles to the bedding depends upon the angle of dip, the thickness of the bed before deformation (the thickness needed for most stratigraphic studies) is the distance across the tilted bed measured in a vertical direction.[82] (See fig. 2.) This corollary seems to be verified by field measurements of the varying thickness of shale formations in the area.[83] A third consequence of this interpretation is that large quantities of water were squeezed out of the shale as it was compressed, and this water must have escaped upward through fractures, joints, or pores, or laterally through interlaminated sandstone beds. Mead [84] has suggested that the deformation and dilatation of sediments [sandstones] may be an important factor in the movement of oil, gas, and water toward anticlines and monoclines. However, in view of the large proportion of shale interlaminated with the sandstone beds in most oil fields and the steep dips common in oil fields in the Rocky Mountain States, it seems more probable that in this region fluid movements were commonly away from rather than toward the areas of deformation.

SHALY STRUCTURE OR FISSILITY

One object of the microscopic study of the samples was to obtain some information on the nature of the

[81] Rubey, W. W., Determination and use of thicknesses of incompetent beds in oil-field mapping and general structural studies: Econ. Geology, vol. 21, pp. 338–339, fig. 2, 1926.

[82] For field methods of measuring the vertical thickness see Ickes, E. L., The determination of formation thicknesses by the method of graphical integration: Am. Assoc. Petroleum Geologists Bull., vol. 9, pp. 451–463, 1925; Rubey, W. W., op. cit., pp. 339–348.

[83] Rubey, W. W., op. cit., pp. 333–351; Cretaceous and Cenozoic formations on the northwest flank of the Black Hills: U. S. Geol. Survey Prof. Paper — [in preparation].

[84] Mead, W. J., op. cit., pp. 691, 697–698.

thin laminations or planes of fissility which constitute the essential characteristic of shale. Many geologists seem to consider these laminations simply bedding planes.[85] Others, however, think that shaly structure is a secondary fissility only approximately parallel to the bedding. Thus Dana,[86] Grabau,[87] Cole,[88] Lewis,[89] and others have explained it as the result of the rotation of original flat flakes or the distortion or fracturing of irregular particles or the growth of micaceous crystals until the grains are essentially parallel to the bedding.

Field evidence indicates that secondary processes are at least a factor in the development of shaly structure in the Upper Cretaceous rocks in the Black Hills region. The planes of parting exposed in shale pits dug for contact dip readings are generally assumed to be identical with the planes of bedding and of fissility in the weathered shale. But, as many geologists who have done detailed structural mapping in Wyoming and Montana have no doubt noticed, the observed dips and strikes in two near-by pits or on two planes in one pit are rarely identical, and the planes of parting in some shale pits are inclined 5° or even more to thin layers of clay or sandstone which may be exposed. Observations of this sort were made, for example, on the flanks of the Pump Creek and Rocky Point anticlines, in T. 48 N., R. 64 W., and T. 56 N., R. 69 W., Wyoming. Field observations such as these indicate either that fissility is not everywhere strictly parallel to bedding or that a system of joints, subparallel to bedding and not uncommon in shale, may easily be confused with true fissility or bedding. Whichever of these two explanations is correct, it seems probable that some secondary process develops planes of parting roughly parallel to bedding in shale. By inference it also seems probable that at least some of the fissility of shale is caused by this secondary process.

Other field evidence also suggests a partly secondary origin of the fissility in the Upper Cretaceous shales of this region. The generally greater abundance of shaly structure in older rocks[90] holds in this region and suggests that fissility may be produced by the weight of overburden or the lapse of time. The Skull Creek and Belle Fourche members of the Graneros shale, near the bottom of the Upper Cretaceous section, contain much the greater part of the most highly fissile shale in the Black Hills region. However, this evidence is not conclusive, for the Mitten member, near the middle of the Pierre shale, much higher in the stratigraphic section, also shows pronounced fissility.

The occurrence of some fissile shale high in the stratigraphic section suggests that shaly structure may be caused by differences in lithology. In fact, in the samples studied, for which chemical analyses are available, the fissility seems to vary inversely with the calcium carbonate content—precisely the opposite of the relation noted by Grabau.[91] On the other hand, the clay content, or grain size, seems to have little if any relation to fissility. Of the samples that were mechanically analyzed the most fissile ones show the widest range of grain size: the Belle Fourche shale contains the most clay, the fissile middle member of the Pierre about the least, and the Skull Creek an intermediate amount. That is, although the calcium carbonate content may be a factor, the effect of other lithologic differences upon fissility is not clearly shown.

Microscopic examination of the shale samples likewise yielded only negative evidence on the origin of the shaly structure. Nearly all the cracks or joints in the thin sections studied follow the bedding as marked by layers of differing grain size or composition. However, a few cracks are inclined at low angles to the bedding, and if these cracks truly represent planes of fissility, they afford evidence that at least some of the fissility is secondary.

The planes of fissility or parting presumably follow the cleavage or orientation of the mineral particles in shale. These individual mineral particles are very small, and it is difficult or impossible to determine satisfactorily the direction of elongation of each grain. However, most of the crystals of clay minerals in the thin sections of shale examined lie so nearly parallel to bedding that they give the rocks an aggregate optical orientation (pp. 5–6). The supposition that the aggregate mineral orientation represents incipient fissility is strengthened by the fact that, in general, the most fissile specimens and those collected where the rocks dip most steeply[92] show the most pronounced aggregate orientation. On the other hand, the aggregate orientation seems also to be related to the lithologic character of the rock, for it is more pronounced in the finer-grained and less calcareous specimens. The only conclusion that can be drawn from these observations with the microscope is that the shaly structure, if it follows the aggregate orientation, is essentially (but not necessarily exactly) parallel to the bedding.

[85] Geikie, Archibald, Textbook of geology, vol. 1, pp. 169–170, 1903. Chamberlin, T. C., and Salisbury, R. D., Geology, vol. 1, p. 487, 1909. Pirsson, L. V., Physical geology, p. 267, 1915; Rocks and rock minerals, p. 327, 1915. Harker, A., Petrology for students, p. 216, 1919. Hatch, F. H., and Rastall, R. H., The petrology of the sedimentary rocks, p. 200, 1923. Willis, Bailey, Geologic structures, p. 2, 1923.

[86] Dana, J. D., Manual of geology, p. 92, 1895.

[87] Grabau, A. W., Principles of stratigraphy, pp. 785–786, 1913.

[88] Cole, G. A. J., Rocks and their origins, pp. 83–84, 1922.

[89] Lewis, J. V., Fissility of shale and its relations to petroleum; Geol. Soc. America Bull., vol. 35, pp. 570–589, 1924.

[90] Lewis, J. V., op. cit., p. 581.

[91] Grabau, A. W., Principles of stratigraphy, p. 785, 1913.

[92] For evidence that the rocks have been deformed internally by tilting during earth movements see pp. 35–38, 54.

In summary, field and microscopic evidence and chemical and mechanical analyses are consistent with, but by no means conclusively prove, the interpretation that the fissility of the shale samples examined is, at least in part, a secondarily induced structure. The inclination between planes of parting and bedding observed occasionally in the field and in thin sections and the apparent effect of stratigraphic position suggest that shaly structure, although in the main parallel to bedding, may be caused by some process such as loading that acts essentially but not exactly at right angles to the bedding.[93]

BEDDING LAMINATIONS[94]

More than 30 thin sections of marine shale, selected to represent Upper Cretaceous formations in the Black Hills region, were examined. Most of these thin sections show bedding laminations marked by more or less distinct alternating differences in grain size or composition. Pairs of these laminations range from 0.05 to more than 1 millimeter in thickness, but most of them are between 0.1 and 0.4 millimeter, and they average about 0.2 millimeter (less than 0.01 inch). In parts of some thin sections the laminations are continuous and pronounced; throughout other thin sections they are discontinuous and obscure. The pairs of laminations are marked by three different kinds of alternations—(1) coarse and fine particles (silt or even fine sand and clay), (2) dark and light-colored silt layers caused by differences in the content of organic matter, and (3) calcium carbonate and silt. (See pl. 4.) The coarse and fine and the dark and light alternations are the most abundant. Examples intermediate between these two types form a gradational series, with the paired laminations made by the coarse and fine alternations the thicker and those made by the dark and light alternations the thinner.

POSSIBLE CAUSE OF LAMINATIONS

These laminations are not structural nor chemical (diffusion bands) but sedimentary features. Most of them seem to be not planes of erosion or solution[95] but simply alternations in a series of continuous deposition. The alternations were almost certainly caused by variations in the rate of supply or of deposition of the different materials. These variations might have been due to changes in the quantity of silt, clay, calcium carbonate, or organic matter in the sea water or to changes in the rate of accumulation of these materials as the result of, say, varying currents. And these changes that caused the laminations might have been cyclic, with intervals of a day, a season, or some longer period, or they might have recurred without any regularity of period.

Some of the large number of possible causes of the laminations are eliminated by the probability that at least two and perhaps all three of the different kinds of paired laminations or double layers were formed by the same general process. The gradational series of paired laminations intermediate in composition and thickness between the type marked by coarse and fine particles and that marked by much and little organic matter indicates that these two types were formed by a related or identical alternation of conditions and during approximately the same periods of time.

Storms or floods might have caused the alternations in particle size, but the paired laminations marked by varying content of organic matter and calcium carbonate seem to call for some other explanation. A single flood, for example, might wash large quantities of detritus into the sea, but it would not be likely to affect materially the amount of organic matter or calcium carbonate in sea water hundreds of miles from shore. Similarly, storm waves might disturb previously deposited mud, and in the subsequent settling the largest grains would fall most rapidly, and so a layer of sediment that was coarsest at the bottom would be formed. But storm waves would not be likely to form organic layers by thus stirring up bottom muds. It is true that settling velocities depend upon the specific gravity as well as the diameter of particles, and the specific gravity of organic matter is much less than that of silt. However, the particles of organic matter presumably were of nearly all sizes, whereas the silt particles were relatively uniform in size. Hence, in settling after disturbance by storm waves, the particles of organic matter would fall at many different velocities and probably would not become concentrated in one part of a layer.

Other and stronger evidence also indicates that occasional storms were not the chief cause of the laminations. The mere fact that very thin laminae have been preserved in these rocks shows that storm waves rarely disturbed the mud on the sea floor. Waves are caused by storms of varying intensity, and wave action that reached to the bottom would usually disturb much more than the very topmost layer of the mud. Hence the presence of these thin laminations and their relatively uniform thickness indicate that the process that caused the layers, even if it were storms, acted with great regularity of time and intensity.

A common process, which has the indicated regularity of period and intensity and which might cause alternations of coarse and fine, organic and non-

[93] Leith, C. K., and Mead, W. J., Metamorphic geology, pp. 173–179, 1915.

[94] Paper presented before the Geological Society of Washington, January 11, 1928 (Possible varves in marine Cretaceous shale in Wyoming [abstract]: Washington Acad. Sci. Jour., vol. 18, pp. 260–262, 1928).

[95] Stockdale, P. B., The stratigraphic significance of solution in rocks: Jour. Geology, vol. 34, pp. 399–414, 1926. Wepfer, E., Über die Entstehung von Schichtung [abstract]: Deutsche geol. Gesell. Zeitschr., Band 78 B, Monatsberichte, p. 57, 1926.

organic, and calcareous and noncalcareous sediments, is the yearly climatic cycle. The alternations of silt and clay might result from the varying quantity of detritus carried into a water body at different times of the year and from the varying quantity of sediment stirred up by wave action near shore at different seasons. The calcium carbonate and silt layers might be due to seasonal changes in the temperature, silt content, and salinity of the sea water. And the layers of much and little organic matter might be caused by seasonal variations in the quantity of land-derived organic matter and, especially, of planktonic life in the sea.

Shifts of marine currents every few years, such as that of the Humboldt and El Niño currents along the west coast of South America,[96] also might cause these different types of alternations in sedimentation, but in view of the relative frequency and geographic extent of pronounced climatic changes caused by the annual cycle and by shifts of marine currents throughout the world during historic times, the annual cycle seems a more probable explanation.

Annual laminations, if certainly recognized, are of considerable geologic interest, because they show the number of years required for a sedimentary deposit to accumulate and the conditions and rates of deposition of different rocks. Because it is of such interest, the possibility that the laminations are annual needs to be examined more carefully.

CONDITIONS FOR FORMATION AND PRESERVATION OF ANNUAL LAYERS

Annual laminations would form only in a region where the climate was definitely seasonal. Seasonal variations in rainfall (and perhaps in temperature also) would be essential, though they need not be great. The Upper Cretaceous climate in the Black Hills region was almost certainly seasonal. On the assumption that the region lay then as now, in the mid-latitudes (approximately 45°), seasonal differences of some sort would be expected, despite the different distribution of climates at that time. Definite evidence of seasonal climate is afforded by fossil dicotyledonous wood collected by the writer from Upper Cretaceous rocks in the Black Hills region and examined by the late F. H. Knowlton. This fossil wood, like that from a number of other places in the Northern Hemisphere,[97] has distinct annual growth rings that were caused by seasonal changes.

However, even though the climate were sufficiently seasonal to cause differences in the rate of supply or deposition of sediments, some doubt may be held that annual layers would form in marine waters. Upon entering salt water clay particles are flocculated, and the aggregates settle rapidly. Hence, as the coarse and fine particles go down together, it has been thought that laminations of differing grain size would not form in salt water.[98] However, some investigators have concluded that recognizable alternations of silt and clay do form in salt water. Kindle[99] suggested that sediments deposited in marine water would be even more sharply laminated than those deposited in fresh water. Sauramo[1] thought that annual laminations in marine or brackish-water sediments could be recognized by the imperfect separation of layers of coarse and fine grains.

Yet, whatever its effect upon the formation of laminations marked by differing grain size, flocculation probably would not prevent the formation of distinct alternations of much and little organic matter and of calcium carbonate and silt. As the organic type of laminations are as common in the shale samples examined as the silt and clay type, it seems that the supposed effect of flocculation is not a serious objection to the hypothesis that the laminations are annual.

But the hypothesis that these laminations are annual requires not only that they could be formed, but also that, once formed, they could be preserved. Very slight stirring of marine muds by waves or currents would destroy any thin laminations that might have been formed previously. Therefore, if the organic layers, for example, are annual laminations, they must have accumulated in quiet water below the depth of effective wave action. In an open sea, bottom muds seem to be disturbed by wave action at greater depth than by a crawling bottom fauna, and therefore the effect of disturbance by a bottom fauna would probably be relatively unimportant in the shales examined.[2]

The maximum depth at which ocean waves disturb bottom muds at the present time is generally estimated at about 600 feet,[3] and the depth to which effective

[96] Murphy, R. C., The oceanography of the Peruvian littoral with reference to the abundance and distribution of marine life: Geog. Rev., vol. 13, pp. 64–85, 1923; Oceanic and climatic phenomena along the west coast of South America during 1925: Idem, vol. 16, pp. 26–54 (especially pp. 53–54), 1926.

[97] Schuchert, Charles, Climates of geologic time: Carnegie Inst. Washington Pub. 192, p. 282, 1914; Evolution of geologic climates: Am. Jour. Sci., 5th ser., vol. 1, p. 324, 1921. Antevs, Ernst, The climatologic significance of annual rings in fossil woods: Am. Jour. Sci., 5th ser., vol. 9, p. 298, 1925.

[98] Sayles, R. W., Seasonal deposition in aqueo-glacial sediments: Harvard Coll. Mus. Comp. Zoology Mem., vol. 47, p. 33, 1919. Johnston, W. A., Sedimentation of the Fraser River delta: Canada Geol. Survey Mem. 125, p. 37, 1921. Sayles, R. W., Seasonal deposition in marine waters: Nat. Research Council Comm. on Sedimentation Rept. [Apr. 18, 1923], p. 61, 1923. Antevs, Ernst, Retreat of the last ice sheet in eastern Canada: Canada Geol. Survey Mem. 146, pp. 14–15, 32, 1925.

[99] Kindle, E. M., Diagnostic characteristics of marine clastics: Geol. Soc. America Bull., vol. 28, pp. 906–909, 919, 1917.

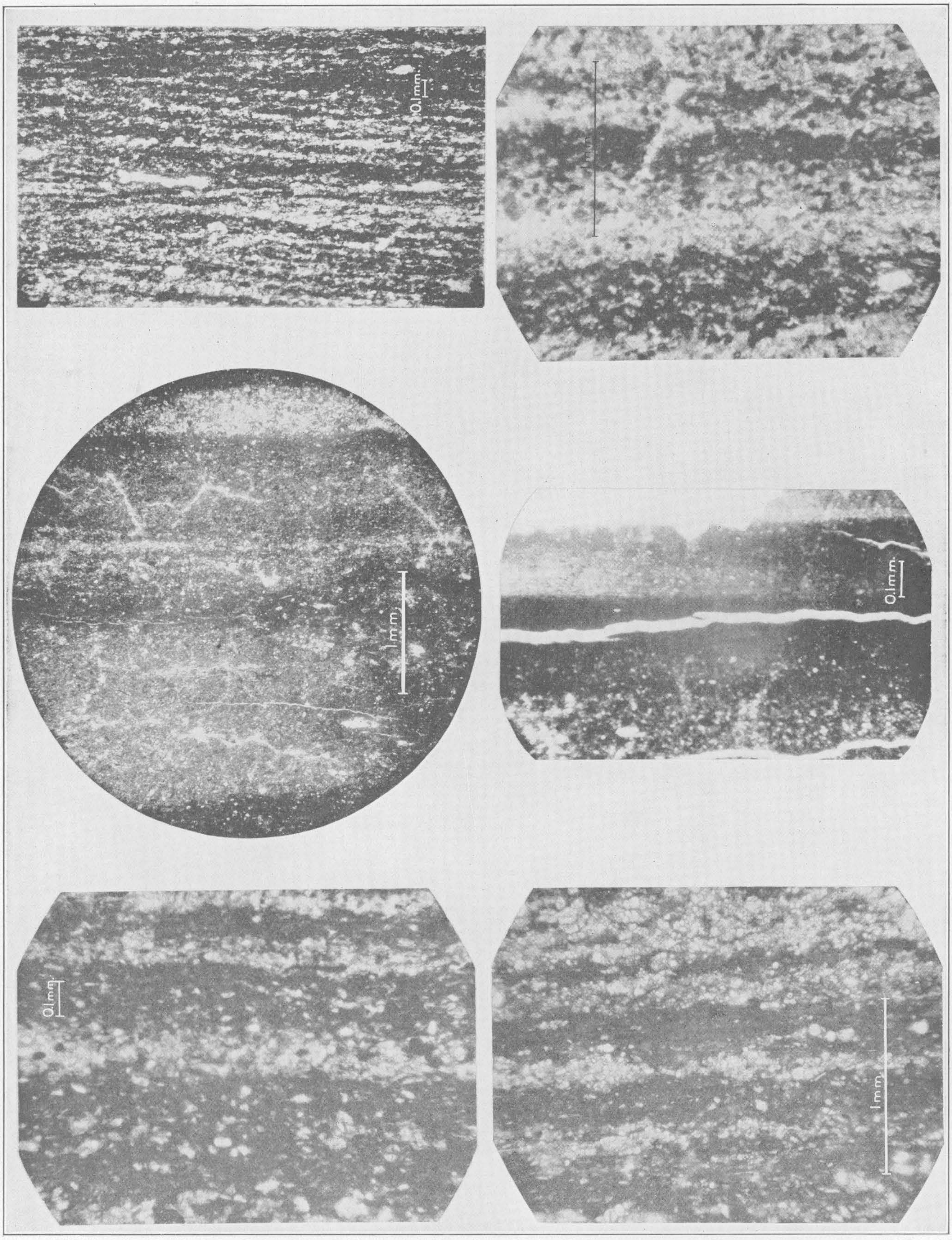
[1] Sauramo, Matti, Studies on the Quaternary varve sediments in southern Finland: Comm. géol. Finlande Bull. 60, pp. 82–83, 92–93, 98, 110, 1923.

[2] Barrell, Joseph, Criteria for the recognition of ancient delta deposits: Geol. Soc. America Bull., vol. 23, p. 426, 1912. Antevs, Ernst, Varved sediments: Nat. Research Council Comm. on Sedimentation Rept., 1925–26, p. 82, 1926; idem, 1926–27, p. 56, 1927.

[3] Johnson, D. W., Shore processes and shore-line development, pp. 76–83, New York, 1919.

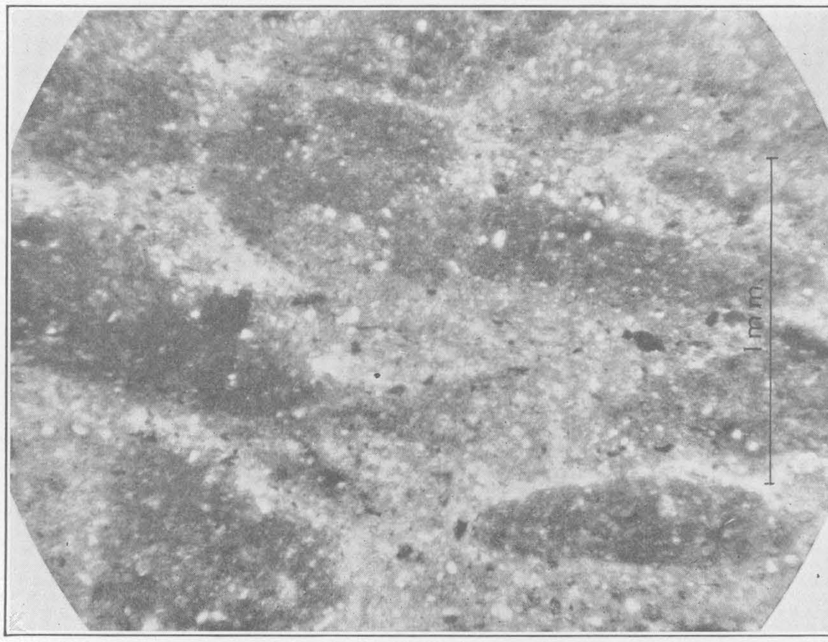
PLATE 4

1. Siliceous shale from Mowry shale member of Graneros shale in center of sec. 7, T. 48 N., 65 W., Weston County, Wyo. Laminations made by alternations of very fine light-gray silt and organic clay. Magnification shown by line 0.1 millimeter long.
2. Mudstone from lower part of Pierre shale in S. ½ sec. 32, T. 49 N., R. 66 W., Crook County, Wyo. Laminations made by alternations of (*a*) very fine quartz sand and coarse silt and (*b*) dark clay. Magnification shown by line 1 millimeter long.
3. Black shale from lower part of Belle Fourche shale member of Graneros shale in north center of sec. 11, T. 48 N., R. 66 W., Weston County, Wyo. Laminations made by alternations of quartz silt and dark organic clay. Magnification shown by line 0.1 millimeter long.
4. Gritty shale from Carlile shale in W. ½ sec. 35, T. 9 S., R. 61 E., Carter County, Mont. Laminations made by alternations of (*a*) very fine quartz sand and coarse silt and (*b*) dark clay. Magnification shown by line 1 millimeter long.
5. Black shale from upper part of Belle Fourche shale member of Graneros shale in W. ½ sec. 34, T. 58 N., R. 62 W., Crook County, Wyo. Laminations made by alternations of light and very dark silt and clay. Magnification shown by line 0.1 millimeter long.
6. Calcareous marl from Greenhorn formation in SE. ¼ sec. 31, T. 45 N., R. 61 W., Weston County, Wyo. Laminations made by alternations of crystalline aggregates of calcium carbonate and dark clay. Magnification shown by line 1 millimeter long.



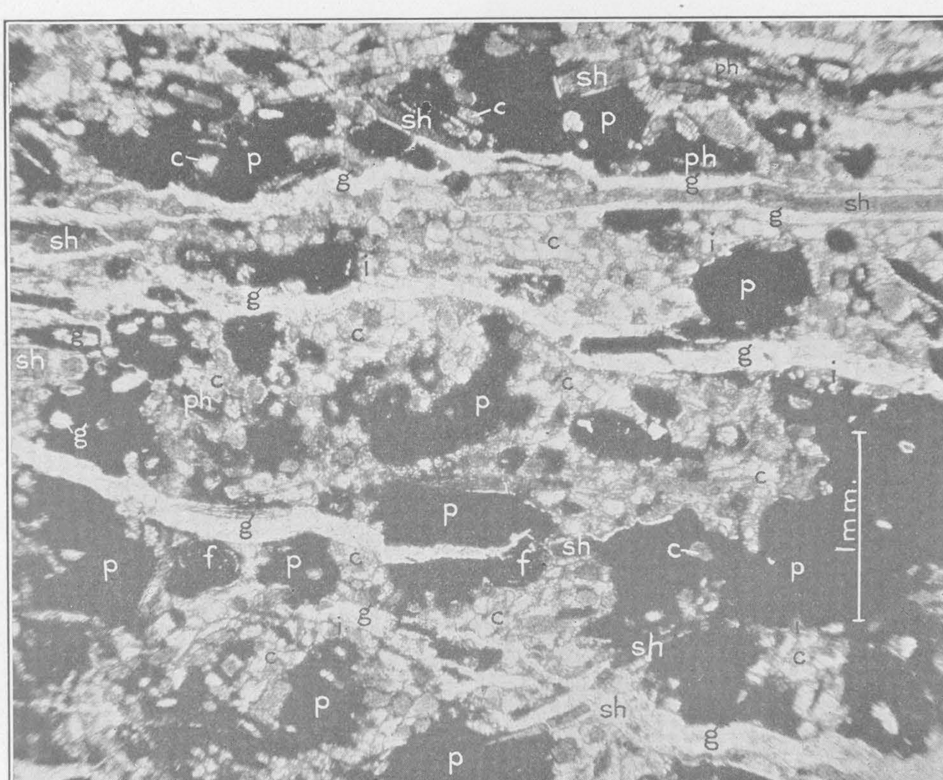
1 2 3 4 5 6

THIN SECTIONS SHOWING LAMINATIONS IN UPPER CRETACEOUS ROCKS OF THE BLACK HILLS REGION



A. MUDSTONE FROM UPPER PART OF PIERRE SHALE IN SEC. 21, T. 8 S., R. 62 E., CARTER COUNTY, MONT.

Brecciation while mud was still soft shown by minute crumpling of layers of dark clay in matrix of very fine quartz sand and coarse silt. Magnification shown by line 1 millimeter long.



B. PYRITIC LIMESTONE FROM THE BASAL BED OF THE GREENHORN FORMATION IN THE SW. ¼ SEC. 11, T. 57 N., R. 62 W., CROOK COUNTY, WYO.

Crystalline calcite (c), shells of *Inoceramus* (sh), and masses of minutely crystalline pyrite (p), cut by veinlets of microfibrous gypsum (g); shells (now gypsum) of Foraminifera (f), in the pyrite and iron oxide (i) bordering some of the pyrite masses. A few phosphatic fragments (ph) of bones and teeth. Magnification shown by line 1 millimeter long.

wave action commonly extends at 300 feet.[4] These depths depend upon the grain size of the bottom sediment, the strength of winds, the distance from the shore (the "fetch" of the wind), and other factors.[5] The Upper Cretaceous shales of the Black Hills region were deposited as fine-grained sediments in a sea that was open to the winds for many hundreds of miles; hence the conditions were favorable for a rather deep wave base. Yet ancient seas that invaded the continental platforms, such as the Upper Cretaceous interior area, are thought to have been generally rather shallow—only a few hundreds or tens of feet deep.[6] Consequently, any annual laminations that might have formed could have been preserved only if the Upper Cretaceous sea was unusually deep throughout or locally near its center in the Black Hills region or if the winds that cause waves were not as strong in Upper Cretaceous time as now.

Some evidence indicates that the Upper Cretaceous sea was relatively deep in the Black Hills region. Shallow-water deposits are very uncommon in the Upper Cretaceous rocks of that region, and whatever the origin of the thin laminations, the mere fact that they were preserved suggests that the shales were deposited below effective wave base. The Upper Cretaceous rocks become much thicker westward from the Black Hills, but this greater thickness does not mean deeper water, for as they thicken shallow-water and continental deposits become more abundant.[7] According to paleogeographic maps, the center of the Upper Cretaceous sea lay near the present Black Hills,[8] and, in view of the structural and geographic relations at that time, it seems a reasonable assumption that the water was commonly deepest near the center of the sea.

Furthermore, wave action may not have extended as deep during Upper Cretaceous time as it does to-day. The depth of wave action depends upon the intensity of winds, and as winds are caused by differences in temperature and pressure of the atmosphere, their intensity is closely related to the diversity of climate.[9] The fact that seas were widespread during Upper Cretaceous time in itself indicates that the climate then was rather moderate and geographically uniform,[10] and the geographic distribution of fossil plants, vertebrates, and invertebrates also indicates relatively widespread equable climates in Upper Cretaceous time.[11]

In summary: Annual laminae might conceivably have formed, for the Upper Cretaceous climate in the Black Hills region was probably sufficiently seasonal to cause annual laminations; and flocculation in salt water, although perhaps hindering the separation of coarse and fine layers, would not have prevented the formation of organic and lime-rich layers. Once formed, thin layers had some chances of preservation, for the sea was probably relatively deep near the present Black Hills, and wave base may have been relatively shallow at that time.

PROBABLE THICKNESS OF ANNUAL LAYERS

The hypothesis that the laminations are annual may be tested by evidence of another sort. Are the pairs of laminations too thick or too thin to be yearly deposits? The expected thickness of annual layers can not be calculated precisely, but the probable order of thickness can be estimated by several methods, and this can be compared with the observed thicknesses.

Many Pleistocene glacial-lake deposits in Europe and North America show varves (annual layers) marked by alternations of clay and silt in layers 10 millimeters or more thick.[12] Some recent and ancient lake deposits have annual laminations, and in places these laminations, which range from less than 0.1 to more than 10 millimeters in thickness, are marked by layers of organic matter and of calcium carbonate.[13] Other kinds of sediments,[14] including a few marine

[4] Barrell, Joseph, Factors in movements of the strand line and their results in the Pleistocene and post-Pleistocene: Am. Jour. Sci., 4th ser., vol. 40, pp. 6–8, 1915; Rhythms and the measurements of geologic time: Geol. Soc. America Bull., vol. 28, pp. 779–780, 1917.

[5] Barrell, Joseph, Geol. Soc. America Bull., vol. 28, p. 779, 1917.

[6] Schuchert, Charles, Paleogeography of North America: Geol. Soc. America Bull., vol. 20, p. 438, 1910. Barrell, Joseph, op. cit., pp. 768–769.

[7] Reeside, J. B., jr., personal communication.

[8] Schuchert, Charles, Sites and nature of North American geosynclines: Geol. Soc. America Bull., vol. 34, pp. 228–229, 1923.

[9] Brooks, C. E. P., Climate through the ages, pp. 47, 63–64, New York, 1926.

[10] Idem, p. 275.

[11] Schuchert, Charles, Climates of geologic time: Carnegie Inst. Washington Pub. 192, pp. 282–283, 1914. Knowlton, F. H., A fossil flora from the Frontier formation of southwestern Wyoming: U. S. Geol. Survey Prof. Paper 108, pp. 79–80, 1917; Evolution of geologic climates: Geol. Soc. America Bull., vol. 30, pp. 526–527, 1919. Stanton, T. W., Evidence of invertebrates on the question of climatic zones during Mesozoic time: Sci. Monthly, vol. 20, pp. 462–463, 1925.

[12] Antevs, Ernst, The recession of the last ice sheet in New England: Am. Geog. Soc. Research Ser. 11, 1922; Retreat of the last ice sheet in eastern Canada: Canada Geol. Survey Mem. 146, 1925; Varved sediments: Nat. Research Council Comm. on Sedimentation Rept., 1925–26, pp. 80–84, 1926; idem, 1926–27, pp. 53–59, 1927. Sauramo, Matti, Studies on the Quaternary varve sediments in southern Finland: Comm. géol. Finlande Bull. 60, 1923. Sayles, R. W., Progress of studies on seasonal deposition of sediments: Nat. Research Council Comm. on Sedimentation Rept., 1924, pp. 38–43, 1925.

[13] Whittaker, E. J., Bottom deposits of McKay Lake, Ottawa: Roy. Soc. Canada Trans., 3d ser., sec. 4, pp. 141–156, 1922. Antevs, Ernst, Retreat of the last ice sheet in eastern Canada: Canada Geol. Survey Mem. 146, pp. 5–6, 1925; Varved sediments: Nat. Research Council Comm. on Sedimentation Rept., 1925–26, p. 82, 1926; idem, 1926–27, pp. 55–56, 1927. Bradley, W. H., The varves and climate of the Green River epoch: U. S. Geol. Survey Prof. Paper 158, pp. 95–97, 1929.

[14] Antevs, Ernst, Canada Geol. Survey Mem. 146, pp. 5–7, 1925; Nat. Research Council Comm. on Sedimentation Rept., 1925–26, pp. 82–83, 1926.

deposits,[15] have also been reported as annually laminated. The laminations in the marine shales range from 1 to 30 millimeters in thickness. (See table on p. 48.)

These are the thicknesses of reported varves or annual laminations in rocks from other regions, but it is possible by three independent methods to estimate very roughly the expected thickness of annual laminations in Upper Cretaceous rocks near the Black Hills. First, the average thickness of annual layers may be obtained by dividing the thickness of beds or alternations of beds deposited during some climatic or other cycle by the number of years in the cycle. Second, the average thickness of sediments laid down annually in a basin the size of the Upper Cretaceous interior sea may be calculated roughly from the rate of erosion in present-day drainage systems. Third, this average annual increment may be estimated by dividing the total thickness of Upper Cretaceous rocks by the number of years supposed to have elapsed during Upper Cretaceous time.

ESTIMATE BASED ON SUPPOSED RECORD OF PRECESSION CYCLE

In the Greenhorn limestone and Niobrara chalk of eastern Colorado G. K. Gilbert [16] found a number of alternations of calcareous and argillaceous strata. He noted that the thickness of strata involved in these alternations increased with the proportion of shale in the rocks. As it seemed to him probable that the shale was deposited more rapidly than the lime, this suggested that the different alternations represent the deposits of some one uniform time-cycle. The precession of the equinoxes (with an average net period of about 21,000 years [17]) seemed to Gilbert to be the cycle that would best account for these alternations. He took 4 feet as the thickness of the alternations in the noncalcareous shale, which in eastern Colorado makes up most of the Upper Cretaceous rocks. If his assumptions are correct, the annual layers in this shale would average about 0.06 millimeter thick. This estimate may be applicable in a rough way to the Black Hills region, for in both regions the Upper Cretaceous rocks are very similar in lithologic character and total thickness.

[15] Heim, Albert, Einige Gedanken über Schichtung: Naturforsch. Gesell. Zürich Vierteljahrschr., vol. 54, pp. 331–332, 1909. Winkler, Artur, Unterschungen zur Geologie and Paläontologie des steirischen Tertiärs: K.-k. geol. Reichsanstalt Jahrb., vol. 63, p. 577, 1913. Sayles, R. W., The dilemma of the paleoclimatologists: Am. Jour. Sci., 5th ser., vol. 3, pp. 471–472, 1922. Sauramo, Matti, op. cit., pp. 75, 109–129. Stamp, L. D., Seasonal rhythm in the Tertiary sediments of Burma: Geol. Mag., vol. 62, pp. 515–528, 1925. Leighton, M. M., Studies of glacial sediments in 1926: Nat. Research Council Comm. on Sedimentation Rept., 1926–27, p. 44, 1927. Marr, J. E., A possible chronometric scale for the graptolite-bearing strata: Paleobiologica, Jahrgang 1, Band 1, Teil 1, pp. 161–162, 1928.

[16] Gilbert, G. K., Sedimentary measurement of Cretaceous time: Jour. Geology, vol. 3, pp. 121–127, 1895.

[17] Idem, p. 124.

ESTIMATE FROM UPPER CRETACEOUS GEOGRAPHY AND PRESENT RATE OF EROSION

If the area of land draining into a settling basin of known size and the annual rate of erosion on this land are known, the average thickness of annual deposits in the settling basin may be calculated.

The probable location and extent of land and sea in North America during Upper Cretaceous time have been graphically shown by several geologists on paleogeographic maps. These graphic interpretations of Upper Cretaceous geography are based upon the present distribution, lithologic character, and fossil content of Upper Cretaceous sedimentary rocks and the known structural history of preceding periods. Paleogeographic maps are necessarily far from accurate, but they sum up existing knowledge into a most useful regional picture. According to these maps, during Upper Cretaceous time the seas encroached upon the present North American continent along the Pacific and Atlantic coasts, and a broad arm of the sea extended northward through the interior in the present region of the Rocky Mountains and Great Plains. This interior sea (the one in which the Upper Cretaceous rocks of the Black Hills region were deposited) joined the Atlantic Ocean through an enlarged Gulf of Mexico and was separated from the Pacific Ocean by a long, narrow range of highlands.

Paleogeographic maps of the Upper Cretaceous distribution of land and sea compiled by different geologists are very similar to one another, and from them the area of the lands that then underwent erosion and the area of the interior basin in which the eroded material was deposited can be estimated. For convenience in computation only those portions of the western highlands, the interior sea, and the eastern land mass that lay within the present limits of the United States may be considered, as the ratio of land to sea in this area probably indicates approximately the ratio of areas of erosion and deposition that influenced the thickness of sediments deposited annually in the Black Hills region. On four representative paleogeographic maps[18] the area of the western highlands within the present United States is shown as approximately half a million square miles, and the areas of the interior settling basin (exclusive of the Mississippi embayment) and the eastern land mass as each slightly less than a million square miles.

How much of the areas of the western highlands and the eastern land mass drained into the interior sea is unknown, because parts of these lands sloped

[18] Willis, Bailey, and Salisbury, R. D., Outlines of geologic history, with especial reference to North America, p. 198, Chicago Univ. Press, 1910. Pirsson, L. V., and Schuchert, Charles, A textbook of geology, p. 893, 1915. Grabau, A. W., A textbook of geology, p. 704, 1921. Schuchert, Charles, Sites and nature of North American geosynclines: Geol. Soc. America Bull., vol. 34, pp. 228–229, 1923.

toward the Pacific, Atlantic, and Gulf of Mexico. The simplest assumption is that on each of these land areas the drainage divide lay midway between the bordering seas and that one-half of each land area drained into the interior sea. This assumption probably gives an unduly small drainage area tributary to the interior sea, because on the eastern land mass the divide probably lay near the Appalachian Mountains, far east of the center; and furthermore a part of the large land mass in eastern Canada probably drained southwestward into the United States. Although the probability of this error in the estimate can be recognized, it is difficult to establish quantitatively the needed correction. For that matter, as the estimate is necessarily an approximate one, such a correction is hardly worth attempting.

Another uncertainty is the average area of the interior sea during Upper Cretaceous time. It is true that the paleogeographic maps may not show the maximum extent of the seas during a period or even at any one time. On the other hand, the shore lines undoubtedly shifted constantly, and the average size of the interior sea during Upper Cretaceous time was probably much less (and the average size of the land areas correspondingly greater) than the size of the sea at the time of its maximum extent. At the beginning and end of the Upper Cretaceous invasion the interior basin was above sea level, and at times during the Upper Cretaceous epoch parts of the basin underwent erosion. However, despite the uncertainty from this constant shifting, it may be safe to assume that, on the average, the interior sea was more than one-half the size shown on the maps.

Hence, taking the areas of land and sea within the limits of the United States as shown on paleogeographic maps and assuming that one-half the drainage from these lands was tributary to the interior sea, we may estimate that the average ratio of erosion areas to deposition areas throughout Upper Cretaceous time was probably between $\frac{800,000}{1,000,000}=0.8$ and $\frac{800,000+500,000}{1,000,000-500,000}=2.6$. If, as seems likely, somewhat more than half of the adjacent land areas drained into the interior sea, both of these ratios should be increased. Although subject to considerable error because of these uncertainties, an estimate that the areas furnishing sediment averaged slightly larger than the area of deposition (perhaps $\frac{1,000,000}{750,000}=1.3\pm$) may be taken as a rough approximation to the average ratio.

Materials eroded from the land are carried away in chemical solution in stream water and by mechanical suspension and rolling along the stream bottom. The portion carried in solution may eventually be distributed throughout the ocean, but the sediment transported mechanically into marine water is nearly all deposited within a few hundred miles of the stream mouths. Of the mechanical load the material rolled along the bottom is coarse grained and deposited relatively near shore, but the suspended sediment is fine grained and makes widespread mud deposits. Consequently, the suspended sediment brought from adjacent lands each year would afford a minimum but perhaps the best available estimate of the average thickness of annual deposits over the entire interior Upper Cretaceous sea.

The volume of detritus eroded each year from a unit area of land varies from year to year and from place to place. The yearly amount and seasonal distribution of rainfall and run-off, the resistance of outcropping rocks to erosion, and the slope of the land are some of the chief factors affecting this rate of erosion. The average annual rate of erosion in many different drainage basins has been determined approximately by engineers and found to vary greatly from one basin to another. In order to estimate the erosion rate on the Upper Cretaceous land areas, drainage basins in which present conditions are roughly comparable to those of Upper Cretaceous time should be chosen; but this choice is exceedingly difficult because only the most general facts are known about the climate and topography of these Upper Cretaceous land areas.

The drainage basin of the Mississippi River is perhaps as good for comparison as any other, because, despite the fact that erosion is somewhat more rapid in this basin than in most other parts of the United States, it is very large and embraces a wide variety of the conditions that affect erosion, its rate of discharge of eroded material has been rather carefully determined, and it lies to-day in the same part of the world as the Upper Cretaceous land masses under consideration. The quantity of suspended sediment carried into the Gulf of Mexico annually by the Mississippi River has been estimated at 340,500,000 tons,[19] and the coarse material rolled along the bottom of the river at about one-tenth as much.[20] The specific gravity of the particles making up the suspended load is approximately 2.6,[21] and the drainage area of the river is about 1,265,000 square miles.[22] Therefore material equivalent to a layer of solid rock averaging about 0.0014 inch or 0.036 millimeter in thickness is eroded off the surface of the land each year and carried in suspension into the Gulf of Mexico. If, instead of being solid rock, the eroded

[19] Dole, R. B., and Stabler, Herman, Denudation: U. S. Geol. Survey Water-Supply Paper 234, p. 84, 1909.

[20] Humphreys, A. A., and Abbott, H. L., Report on physics and hydraulics of Mississippi River, p. 149, 1861.

[21] Dole, R. B., and Stabler, Herman, op. cit., p. 80.

[22] Idem, p. 84.

material contains 50 per cent of pore space, the layer removed each year averages 0.072 millimeter in thickness.

These approximate figures—the relative areas of land and sea during Upper Cretaceous time and the thickness of mud-forming detritus eroded annually—afford a basis for estimating the average thickness of annual shale deposits in the Upper Cretaceous interior sea. In nine samples of Upper Cretaceous shale from the Black Hills region the porosity ranges from 24 to 38 per cent and averages 30 per cent. (See pp. 32, 34.) The probable drainage and deposition areas indicate that an annual layer of solid silt about 0.05 millimeter thick, or, if the material has a porosity of 30 per cent, a layer 0.07 millimeter thick would be expected.

This estimate is of course only approximate; it means that the thickness of annual layers is probably to be measured in tenths or hundreths of millimeters, not in millimeters or centimeters, nor, on the other hand, in microns or millimicrons. Even should 0.07 millimeter be the correct average thickness for the entire sea, the thickness of the layers would certainly vary from place to place. Mud deposits near shore might be thicker and those farthest from land thinner than the average. However, as applied to the Black Hills region, many of the possible errors in the estimate tend to compensate one another. The possibly excessive annual thickness estimated by using the Mississippi River rather than some other drainage system as a basis of computation is probably more than offset by using too small a drainage area tributary to the interior sea and by disregarding the coarser material brought in by streams. Similarly the great distance of the Black Hills region from the shores of the Cretaceous sea and the consequent thinness of deposits that might be expected there is perhaps offset by the ignored organic and chemical deposits, such as calcium carbonate and bases adsorbed on the surfaces of fine clay particles—deposits that might be thicker in the middle of the settling basin than near shore. Furthermore, this expected thinness due to the central position of the Black Hills would be largely offset by the reworking of shore deposits. That is, present outcrops show that sediments did not long accumulate in excessive thickness near the eastern shore of the Upper Cretaceous sea. Therefore, as the floor of the eastern half of this settling basin was probably gently shelving, any temporarily excessive deposits near shore must have been reworked by waves and distributed westward.

ESTIMATE FROM TOTAL THICKNESS OF ROCKS AND SUPPOSED DURATION OF UPPER CRETACEOUS TIME

If the total thickness of rocks that accumulated during Upper Cretaceous time and the number of years required for this accumulation are known, the average thickness of annual layers may be estimated. The writer's measurements of outcropping rocks and his correlations of well logs indicate that the total thickness of Upper Cretaceous rocks (excluding the Eocene (?) Lance formation) on the western flank of the Black Hills is about 4,000 or 4,500 feet, of which all but a few hundred feet are marine shales, mudstones, and marls.[23] In southwestern Wyoming, 350 miles from the Black Hills, the maximum thickness of rocks of the same age is more than 20,000 feet,[24] but there a large proportion of the total is made up of rather coarse continental deposits, which very probably accumulated more rapidly.

The time required for the deposition of these rocks is much more difficult to determine than their thickness. Almost all estimates that have been made of the duration of Upper Cretaceous time have been calculated by allotting to the Upper Cretaceous epoch a proportionate part of the time since the earliest geologic record. The basis for the proportionate allotment is indicated below. Estimates of the number of years that have elapsed since the earth reached essentially its present condition have been made by widely different methods, and the estimates themselves have also varied widely. The principal criteria have been the time required for the development of life forms, for the loss of heat by the earth and sun, for the denudation of lands and accumulation of sediments, for the number of mountain-building revolutions recorded in the rocks, for the concentration of salt in the sea, and for the disintegration of radioactive minerals.

Of these criteria the two that are generally thought to afford most definite estimates are those based on salt in the ocean and on lead-uranium ratios. But these two, though most definite, give very different results. The quantity of sodium in the ocean divided by the quantity now being added each year (with some minor corrections) gives somewhat less than 100,000,000 years [25] as the age of the ocean. The ratio of the quantity of the unstable element uranium to the quantity of its disintegration product, lead, with which it is found associated, and the observed rate of radioactive disintegration of uranium indicate that the oldest rocks exposed on the earth's surface are more than 1,000,000,000 years old.[26] Thus, though it

[23] Rubey, W. W., Cretaceous and Cenozoic formations on the northwest flank of the Black Hills: U. S. Geol. Survey Prof. Paper — [in preparation]. (See also pp. 3–5 of this report.)

[24] Wilmarth, M. G., The geologic time classification of the United States Geological Survey compared with other classifications: U. S. Geol. Survey Bull. 769, p. 6, 1925; Tentative correlation of geologic formations in Wyoming (mimeographed chart), U. S. Geol. Survey, 1925.

[25] Clarke, F. W., The data of geochemistry, 5th ed.: U. S. Geol. Survey Bull. 770, pp. 150–155, 1924.

[26] Holmes, Arthur, and Lawson, R. W., Factors involved in the calculation of the ages of radioactive minerals: Am. Jour. Sci., 5th ser., vol. 13, p. 342, 1927.

seems that these two estimates should be approximately the same, one is more than ten times as great as the other.

The relative merits of these and other estimates have been much discussed, and efforts have been made to reconcile them. Barrell [27] pointed out that because of the exceptional height and area of continents to-day the rate at which sodium is now being added to the ocean may be many times greater than it has been throughout the geologic past. Chamberlin [28] called attention to the acceleration of erosion in drainage basins occupied by human beings, to the deposition of some oceanic sodium by adsorption or base exchange on clay particles, and to reasons for considering chlorine, the accumulation rate of which indicates a greater age than sodium, as the better fitted of the two elements to serve as a criterion of age. On the other hand, Joly [29] contends that several lines of evidence indicate that uranium disintegrated more rapidly in the past than it does to-day. Critical discussion of these and other estimates is outside the scope of this paper. For the purpose of this investigation the estimates based on the sodium and uranium methods may suffice to indicate low and high figures for the approximate duration of geologic time.

There is much closer agreement among different writers about the proportional distribution of geologic time among the different eras and periods. The proportions of total time represented by the several divisions have been estimated chiefly from the volumes and maximum thicknesses of sediments of the respective ages. The percentage of Paleozoic and later time assigned to the Upper Cretaceous in a group of estimates summarized by Barrell [30] ranges between 4½ and 9 per cent. A recent compilation [31] indicates that the maximum thickness of the Upper Cretaceous is 6.6 per cent of the total of the maximum thicknesses of Paleozoic and later rocks.

Considering both the supposed total duration of geologic time and the proportional distribution among the several periods, different geologists have estimated the duration of the Upper Cretaceous epoch at two to two and three-fourths million years on the basis of the sodium time scale, or forty to fifty million years on the basis of the uranium time scale,[32] as well as making other estimates between these figures.

The thickness of Upper Cretaceous rocks on the western flank of the Black Hills divided by two or two and three-fourths million gives about 0.5 or 0.6 millimeter as the thickness of annual layers. Dividing by forty or fifty million gives a thickness of 0.025 or 0.030 millimeter for the annual layers. The thicker Upper Cretaceous section in southwestern Wyoming would, by this method, indicate annual layers several times thicker, but there the rocks are coarser grained and presumably accumulated more rapidly than those of the Black Hills region.

This method involves an assumption that none of the Upper Cretaceous rocks are missing from the stratigraphic section in the Black Hills region, and it therefore gives a minimum estimate of the thickness of annual layers. In a general way this assumption may be justified; no large unconformities within the Upper Cretaceous series are known in this region, and the paleontologic evidence indicates that nearly every faunal zone is represented.[33] However, field evidence suggests that there were in Upper Cretaceous time at least two periods of nondeposition or erosion (probably without uplift above sea level),[34] and perhaps many smaller stratigraphic breaks that have not been found. Also the Telegraph Creek fauna [35] of Montana has not been recognized near the Black Hills, and rocks of that age may be absent there. Any such gaps in the Black Hills Upper Cretaceous section would make the estimated thickness of annual layers too small. Although for this reason the estimate is too low, no one can say whether it should be increased by one-tenth or multiplied by 10.

It is of interest to note in passing that the minimum thickness of 0.025 or 0.030 millimeter based upon radioactive disintegration is only one-half or one-third of that based upon present rates of erosion (0.07). That is, under the assumption made, erosion may have been only one-third or one-half as rapid (or, if stratigraphic breaks are important, even more) in Upper Cretaceous time as it is now—not one-tenth, one-fifteenth, or one-twentieth, as, because of the unusual height of the continents to-day, Barrell [36] thought probable for most of the geologic past.

COMPARISON OF ESTIMATES WITH OBSERVED THICKNESS OF LAMINATIONS

Three totally independent methods of estimation indicate that, if annual layers are present in the Upper Cretaceous shales of the Black Hills region, they are probably a few tenths or hundredths of a millimeter thick (see table on p. 48), and, as the observed pairs of laminations average about 0.2 millimeter thick, the hypothesis that they may be annual layers is thereby strengthened. It is true that the

[27] Barrell, Joseph, Rhythms and the measurements of geologic time: Geol. Soc. America Bull., vol. 28, pp. 834–838, 871–872, 1917.
[28] Chamberlin, T. C., The age of the earth from the geological viewpoint: Am. Philos. Soc. Proc., vol. 61, pp. 252, 266–270. 1922.
[29] Joly, John, The surface history of the earth, pp. 150–153, 1925.
[30] Barrell, Joseph, op. cit., pp. 884–885.
[31] Wilmarth, M. G., op. cit. (Bull. 769), p. 6.
[32] Barrell, Joseph, op. cit., pp. 884–885. Wilmarth, M. G., op. cit., p. 5.
[33] Reeside, J. B., jr., personal communication.
[34] Rubey, W. W., op. cit.
[35] Reeside, J. B., jr., A new fauna from the Colorado group of southern Montana: U. S. Geol. Survey Prof. Paper 132, pp. 25–31, 1923.
[36] Barrell, Joseph, op. cit., pp. 747, 761, 774–776, 891–893, 1917.

observed pairs of laminations are about three times as thick as annual layers would be expected to be from estimates based (1) on G. K. Gilbert's supposed record of precession cycles and (2) on Upper Cretaceous geography and present erosion rates. Also they are perhaps seven or eight times as thick as the minimum thickness of annual layers computed from estimates of Upper Cretaceous time based on radioactive disintegration but only about one-third as thick as the minimum thickness computed from estimates based on ocean salt. However, although most of the estimates seem to indicate that the observed pairs of laminations are slightly thicker than might be expected, yet it is doubtful if any of the estimates are more accurate than the differences between the observed and computed thicknesses.

Thickness of laminations in sedimentary rocks of Black Hills and other regions

	Thickness (millimeters)	
	Range	Average
Pairs of laminations observed in Upper Cretaceous shale of Black Hills region	0.05– 1.25	0.2
Annual laminations in sediments from other regions:		
Glacial varves	10 –100	[a]40
Lacustrine varves	0.1 – 10	[a]1
Marine varves	1 – 30	[a]2
Annual layers in Upper Cretaceous shale of eastern Colorado, computed from G. K. Gilbert's supposed record of precession cycle		[a]0.06
Annual layers in Upper Cretaceous shale of Black Hills region, estimated from—		
Upper Cretaceous geography and present rate of erosion		[a]0.07
Total thickness of Black Hills Upper Cretaceous rocks divided by supposed duration of Upper Cretaceous time:		
Cretaceous time = 2 million years (based on salt in ocean)		[b]0.60
Cretaceous time = 50 million years (based on lead-uranium ratios)		[b]0.03

[a] About. [b] More than.

The estimates certainly suggest that the thickness of the pairs of laminations is of the correct order of magnitude for annual layers. Tides and daily variations of temperature and wind, which might possibly produce recognizable alternations in the sediments, would probably make layers only hundredths or thousandths as thick. Climatic cycles with periods longer than one year also seem to be unsatisfactory explanations of the laminations. Short ones like the sunspot cycle [37] do not now and perhaps never did cause as great variations as the annual contrast between summer and winter, and so of the two the annual cycle seems to offer a much more likely explanation. The much longer cycle of the precession of the equinoxes may cause very pronounced climatic variations, but any alternations thus produced would be many thousands of times as thick as the observed laminations. Storms recurring several times a year or every few years are not eliminated by these estimates of the probable thickness of annual layers; and the type of lamination marked by alternations of grain size might be so explained. However, as discussed on pages 40–41, these laminations marked by alternations of grain size apparently were formed in much the same way as the equally common organic laminations, which were probably not storm-made; and the preservation of thin layers indicates that storms rarely disturbed the bottom muds.

RATE OF ADVANCE OF UPPER CRETACEOUS SEA

The discussion thus far has led to the conclusions that annual laminations might possibly have been formed and preserved in the Upper Cretaceous rocks of the Black Hills region, that the observed laminations in these rocks are of different kinds that can be explained most simply as annual layers, and that they are of about the right thickness to be annual layers. These conclusions, of course, do not constitute proof that they are annual layers. In fact, more detailed comparisons suggest that they may have formed every few years instead of every year, although the relatively small discrepancies found in these comparisons can be explained equally well in other ways.

If the observed laminations are annual and if their average thickness is typical of the series in the region, they furnish a rough measure of the number of years that elapsed while certain events occurred. Applied in this way, they indicate that the Upper Cretaceous sea advanced eastward from the Black Hills rather rapidly.

Eastward and southward from the Black Hills the sandstone beds at or near the base of the Upper Cretaceous, which have generally been called Dakota sandstone, seem to transgress upward in the stratigraphic section.[38] In Nebraska, Kansas,[39] eastern Colorado, and eastern Wyoming these sandstone beds, which are largely nonmarine, underlie the marine Graneros shale, which in turn underlies the Greenhorn limestone. The Greenhorn limestone was probably deposited almost simultaneously throughout this area, yet the underlying Graneros shale thins from about 1,250 feet on the northwestern flank of the

[37] Barrell, Joseph, op. cit., p. 825. Huntington, Ellsworth, and Visher, S. S., Climatic changes, their nature and causes, Yale University Press, 1922. Brooks, C. E. P., Climate through the ages, pp. 96–113, 1926.

[38] Grabau, A. W., Principles of stratigraphy, p. 739, 1913.

[39] Rubey, W. W., and Bass, N. W., The geology of Russell County, Kans.: Kansas Geol. Survey Bull. 10, pp. 62–63, 1925.

Black Hills [40] to about 40 or 60 feet in Kansas [41] and eastern Nebraska.[42] That is, the sandstone beds that underlie the Graneros shale apparently rise about 1,200 feet stratigraphically in the 450 miles between northeastern Wyoming and the vicinity of Sioux City, Iowa.

If this difference in the thickness of the Graneros shale was caused by the eastward transgression of the sandstone shore phase, so that the shale 50 feet below the Greenhorn in Wyoming was laid down far off shore at the same time as the top of the Dakota in Iowa, then the time required for the underlying 1,200 feet of Graneros to accumulate was the time required for the shore of the sea to advance the 450 miles. (See fig. 3.) If the laminations are annual and if the 1,200 feet of Graneros shale contains no stratigraphic breaks, then the sea advanced eastward across South Dakota in about 2,000,000 years (the number of feet multiplied by 1,500, the average number of laminations to the foot), or at an average rate of slightly more than 1 foot a year. Of course, even under ideally uniform conditions the rate of advance would not be constant from year to year. For example, as the sea advanced its area would increase and the area of dry land would decrease, so that the quantity of detritus spread over a unit area of the sea would become less each year. However, it does not necessarily follow that the laminations would become thinner and the rate of advance slower from year to year. If the sea floor did not sink as rapidly as detritus was delivered to it, the sea would nearly fill with sediment, and waves would sweep later detritus on out to deeper water.[43]

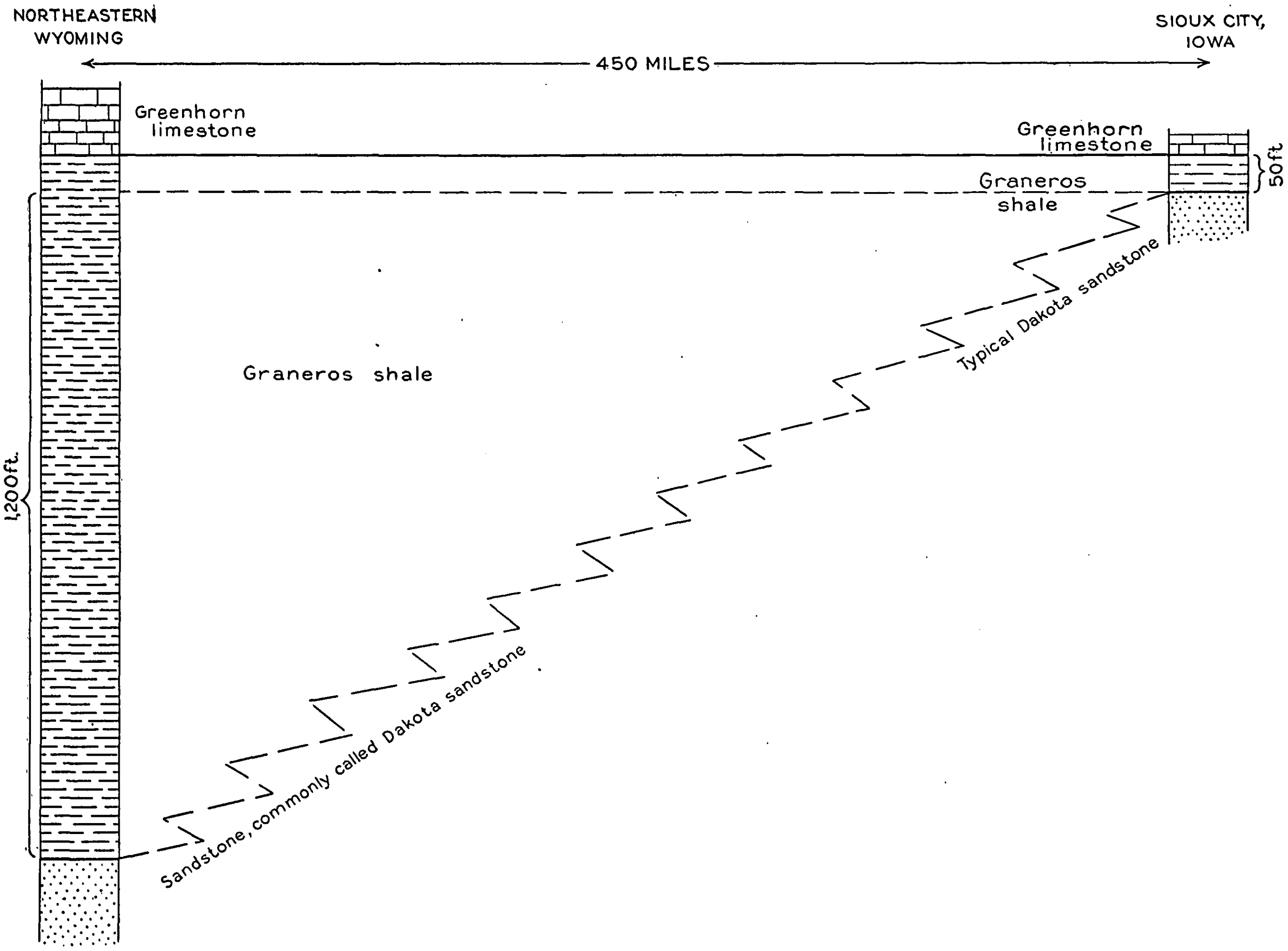


FIGURE 3.—Methods of estimating rate of eastward advance of Upper Cretaceous sea

For comparison with the foregoing estimate, the rate of advance may be computed on the same interpretation of the age relations of the Dakota sandstone but without using the laminations as a measure

[40] Rubey, W. W., Cretaceous and Cenozoic formations on the northwest flank of the Black Hills: U. S. Geol. Survey Prof. Paper — [in preparation].

[41] Rubey, W. W., and Bass, N. W., op. cit., p. 51.

[42] Condra, G. E., Geology and water resources of a portion of the Missouri River valley in northeastern Nebraska: U. S. Geol. Survey Water-Supply Paper 215, p. 11, 1908. Burchard, E. F., Geology of Dakota County, Nebr., with special reference to lignite deposits: Sioux City Acad. Sci. and Letters Proc., vol. 1, p. 150, 1904.

[43] Barrell, Joseph, op. cit., p. 748.

of time. If the estimates of geologic time based on the salt in the ocean are correct and the entire Upper Cretaceous epoch lasted only 2,000,000 years, then the 1,200 feet of Graneros shale, about one-third or one-fourth of the total thickness of Upper Cretaceous rocks (about 4,250 feet), accumulated in approximately 500,000 years, and the sea advanced about 4 or 5 feet a year. If the Upper Cretaceous epoch lasted about 50,000,000 years as indicated by lead-uranium ratios, then the sea probably advanced about 2 inches a year.

It is worth noting that the interpretation of the age relations of the Dakota sandstone on which these estimates are based permits the slowest possible advance of the sea. If, for example, the sea did not transgress eastward at a uniform rate but by a series of minor advances and retreats, then during the minor advances the shore line moved much more rapidly than the average rate. Or, again, if the difference in thickness of the Graneros shale was caused not by transgression of the sandstone eastward but by more rapid deposition of the shale toward the west—that is, if the sandstone was not transgressive but of the same age in Wyoming and Iowa—then the sea advanced so much more rapidly that the rate could not even be estimated. Or, third, the difference in thickness may have been caused partly by transgression eastward and partly by more rapid deposition westward, so that the Newcastle sandstone of Wyoming, several hundred feet above the base of the Graneros shale, may have been contemporaneous with the upper part of the Dakota sandstone of Iowa.[44] On this third interpretation the shore line moved across South Dakota during the deposition of about 200 instead of 1,200 feet of shale, and the average rate of advance must have been about six times as fast as the rates estimated from the first interpretation.

All these estimates of the rate at which the Upper Cretaceous sea spread over the land—even the one based on the lead-uranium ratios—seem rather high. The average rate of marine invasion on sinking coasts to-day is not known. In areas thought to be sinking very rapidly, such as the Netherlands, New Jersey, and northern Gulf of Mexico coasts,[45] the annual rate of vertical sinking has been estimated at as much as several millimeters, but the average rate for all sinking coasts must be much less. Some coasts are being cut back by waves several feet a year,[46] but Geikie[47] estimated that, on the average, this rate is much less than one-tenth of a foot a year. It is true that the rate of coast retreat by submergence, which depends upon the rate of sinking and the slope of the land, is not the same as the rate of cutting by waves, yet from the rather common absence of exceedingly rough topography below unconformities, it seems probable that submergence has not often greatly exceeded wave erosion and that the present rate of cutting may be taken as a rough measure of the rate of submergence in the past. The land surface over which the Upper Cretaceous sea advanced was probably rather smooth and gently sloping,[48] hence its submergence may have been somewhat more rapid than that of the more rugged, steeply sloping land surfaces of to-day. That is, the laminations suggest that the Upper Cretaceous sea may have advanced at least ten times as rapidly as the present average rate of shore advance. This may mean that the laminations are not annual but that they formed every 10 years or so. However, considering the rapid advance indicated even without using the laminations as a measure of time, it seems more likely to mean that the Upper Cretaceous sea advanced more rapidly than most present-day seas, perhaps because it came in over a relatively flat land surface.

LENGTH OF UPPER CRETACEOUS TIME

The observed thickness of pairs of laminations has been compared with the reported thicknesses of annual laminations from other regions and with the expected thicknesses, as estimated by several different methods, of annual layers in the Black Hills region. This general comparison indicated that the paired laminations were of about the right thickness to be annual layers. Hence a more detailed comparison with one of the several methods of estimation used before is justified.

If the observed layers are typical of each formation in the Black Hills region, the total duration of Upper Cretaceous time indicated by the formation thicknesses is about five or ten million years. This estimate is greater than the two or two and three-fourths million years based on the salt in the ocean and less than the forty or fifty million years based on the lead-uranium ratios. The five or ten million years is, however, a minimum estimate, for it ignores the field and paleontologic evidence of stratigraphic breaks within the Upper Cretaceous series of this region. If 1,000 feet of Upper Cretaceous rock was eroded during Upper Cretaceous time, the estimates should be increased by about one-fourth.

If, as many geologists and geophysicists believe, the lead-uranium ratios are the most accurate measure of geologic time now available, and if forty or fifty million years is therefore the most probable duration of

[44] Stanton, T. W., Some problems connected with the Dakota sandstone: Geol. Soc. America Bull., vol. 33, pp. 264–269, pl. 4, 1922.

[45] McGee, W J, The Gulf of Mexico as a measure of isostasy: Am. Jour. Sci., 3d ser., vol. 44, pp. 179–180, 187–188, 1892.

[46] Gregory, H. E., The formation and distribution of fluviatile and marine gravels: Am. Jour. Sci., 4th ser., vol. 39, p. 489, 1915. Johnson, W. D., Shore processes and shore-line development, pp. 69–71, 1919.

[47] Geikie, Archibald, Textbook of geology, vol. 1, p. 593, 1903.

[48] Lee, W. T., Reasons for regarding the Morrison an introductory Cretaceous formation: Geol. Soc. America Bull., vol. 26, pp. 311–312, 1915. Rubey, W. W., and Bass, N. W., op. cit., pp. 57–62, pl. 3.

Upper Cretaceous time, then the laminations indicate that only about one-fifth or one-tenth of total Upper Cretaceous time is represented in the rocks now present in the Black Hills region. As the Upper Cretaceous rocks in this region make a relatively complete stratigraphic section, this discrepancy leads almost certainly to one of four conclusions: (1) the laminations represent cycles longer than the present year; or (2) the average thickness of the observed laminations is not typical of the series in this region; or (3) the Upper Cretaceous epoch was much shorter than forty or fifty million years; or (4) inconspicuous stratigraphic breaks, aggregating four-fifths or nine-tenths of all Upper Cretaceous time, are distributed about evenly through the Upper Cretaceous section of the region.

It is possible that the duration of the year—the period of the earth's revolution about the sun—has varied during geologic time. It is also possible that climatic cycles of several years' length, such as the sun-spot cycle, may account for the discrepancy between the number of laminations and the estimated duration of the Upper Cretaceous epoch in years; but the contrast between summer and winter seems a more probable cause of alternations in marine sediments than the relatively slight variations during longer cycles. Again it is possible that an occasional storm destroyed the laminations of many years' accumulation, and that this disturbed sediment was redeposited after the storm as a single layer; that is, because of occasional storms, fewer layers might be preserved than the number of years required for the formation of the deposit. However, as discussed on pages 40–41, this explanation would not account for the layers of much and little organic matter, and it seems probable that the layers of different types were formed by the same process and during approximately equal intervals of time; also the thinness of the laminations and the relative uniformity of this thinness suggest that the bottom muds were never disturbed by storm waves.

The average thickness of the observed laminations may not be typical of the laminations throughout the Upper Cretaceous rocks of the Black Hills region, even though the thin sections examined were selected in the belief that they were representative. This uncertainty will remain until many more than 30 thin sections have been examined. Also, parts of the shale that show no laminations may have accumulated much more slowly than the others. However, with present information, it seems that the entire discrepancy can not be explained in this way, because in general rock types that show the thickest laminations, and hence seem to have accumulated most rapidly, show the most evidence of destruction of laminations (pl. 5, *A*) and also the largest proportion of obscurely laminated or entirely unlaminated shale, whereas the most thinly laminated rocks tend to be the most distinctly laminated. (See pp. 51–52, 53.)

Some scraps of evidence favor one of the other two explanations. The total of about 2,500,000 laminations found by Stamp [49] in the Oligocene and lower Miocene deposits of Burma, like that of the Upper Cretaceous laminations, is only about one-tenth the number of years estimated by Barrell [50] as the time during which the rocks accumulated. And Barrell, when defining diastems as minor but very common stratigraphic breaks caused by downward oscillation of wave base,[51] seems from the illustrative examples he cited to have thought that the aggregate time value of stratigraphic breaks might be five or ten times that recorded in the remaining deposits.[52] Stamp's results seem to accord with the writer's in suggesting that the estimates based on lead-uranium ratios or the proportions assigned to certain periods may be too long or that inconspicuous stratigraphic breaks may be very common. Barrell's theoretical conclusions favor the latter possibility.

CONDITIONS OF DEPOSITION INDICATED BY LAMINATIONS

Whether or not the laminations are annual, they furnish a clue to some of the conditions of deposition. Their distinctness or degree of preservation is probably a rough measure of the quietness of the water in which the sediments accumulated. Very slight stirring would destroy thin laminations in previously deposited clay; and the disturbed fine particles would be redeposited as a mixture with coarser material that came in later.[53] Distinct laminations would be formed and preserved only below the depth at which waves could move the very fine sediments; and therefore distinctly laminated sediments probably accumulated in deep water or at times of mild climate and gentle winds. The distinctness of laminations would probably depend also on the salinity of the water [54] and the climatic contrasts between the seasons of the year, but, as the Upper Cretaceous rocks of the Black Hills region are nearly all marine and as the available evidence indicates a mild but distinctly seasonal climate in the Black Hills region during Upper Cretaceous time, these conditions were fulfilled here.

On applying this test to the samples studied, it appears that the fine-grained noncalcareous shales with the organic type of laminations accumulated in deep water or at times of gentle winds. The coarser silt

[49] Stamp, L. D., Seasonal rhythm in the Tertiary sediments of Burma: Geol. Mag., vol. 62, pp. 526–527, 1925.
[50] Barrell, Joseph, op. cit., p. 884.
[51] Idem, pp. 748, 785–795.
[52] Idem, pp. 796–797, 807, fig. 5.
[53] Barrell, Joseph, Criteria for the recognition of ancient delta deposits: Geol. Soc. America Bull., vol. 23, pp. 425–427, 1912.
[54] Sauramo, Matti, op. cit. pp. 90, 92–93, 97–98, 110.

and clay layers are equally distinct, but, in view of the larger waves or shallower depths necessary to account for the disturbance of the coarser grains, it seems that the distinctly laminated finer-grained rocks probably accumulated in deeper waters than the equally well laminated coarser rocks.

This test also indicates that the mudstones in the upper part of the Pierre shale and the calcareous marls of the Niobrara formation accumulated in relatively shallow water or at times when wave base extended deeper than usual, for they show little or no lamination. Loss of laminations in the upper Pierre mudstones by wave disturbance is further indicated by the minute contemporaneous brecciation shown by several specimens examined. (See pl. 5, *A*.) The suggestion that the calcareous marls of the Niobrara formation may have accumulated in relatively shallow water accords with the greater abundance of mollusk shells in the calcareous rocks (see pp. 12, 53) and with the greater solubility of calcium carbonate in cold (deep) water than in warm (shallow) water.[55]

If the different kinds of paired laminations were formed during equal time intervals (whether or not these time intervals were years), the rocks with thicker pairs of laminations accumulated more rapidly than the rocks with thinner ones. Hence, the relative thicknesses of the pairs of laminations in different samples may indicate the relative rates of accumulation of different lithologic types and formations as a whole. However, this criterion must be used cautiously, for, as Barrell[56] pointed out, the more rapid the accumulation of individual beds the longer the alternating periods of nondeposition may be. Nevertheless, it is interesting to note that the Mowry shale, which has the thinnest pairs of laminations observed, affords independent evidence that it accumulated much more slowly[57] than other formations of the region. If the thickness of the pairs of laminations indicates the relative rates of accumulation of the different rock types and formations, the sandy shales accumulated most rapidly, the calcareous marls probably somewhat less rapidly, and the fine-grained organic shales most slowly. (See also pp. 13, 53.)

SUMMARY

Microscopic examination reveals alternate laminations that average about 0.2 millimeter in thickness in marine shale of many Upper Cretaceous formations in the Black Hills region. These laminations are of different kinds, marked by alternations of three types—(1) coarse and fine particles, (2) much and little organic matter, and (3) calcium carbonate and silt. Examples intermediate between types 1 and 2 form a gradational series in which those pairs of laminations made by alternations of particle size are the thicker and those made by alternations of organic content the thinner. This gradation indicates that at least these two kinds of alternate laminations were formed by the same or a closely related process and during approximately equal time intervals.

Storms or floods might have caused the alternations of coarse and fine particles, but the pairs of laminations marked by varying content of organic matter and of calcium carbonate seem to call for recurrent cycles of organic growth or of changes in temperature. The fact that thin laminations have been preserved indicates that sporadic storms rarely disturbed the sea floor, and the regularity of the alternations suggests that the cause, whether storms or not, recurred periodically. Either seasonal changes in temperature, rainfall, and food supply of organisms or periodic shifts of marine currents (warm and cold) probably afford the simplest explanation of all three kinds of alternations; and of these two possibilities seasonal changes appear the more likely.

It is conceivable that annual layers might have formed in the Upper Cretaceous rocks of the Black Hills region, for fossil wood and other evidence indicate that the climate was seasonal; and flocculation, which, according to some geologists, might prevent the separation of sand and clay into coarse and fine layers, probably would not prevent the formation of layers marked by alternations of either organic matter or calcium carbonate. Once formed, thin layers might have been preserved, for the deepest part of the Upper Cretaceous interior sea probably lay near the present Black Hills; also, wave action may not have extended as deep during the widespread equable climates of Upper Cretaceous time as it does to-day.

The hypothesis that the laminations are annual is tested roughly by comparing the thickness of the observed laminations with the thickness that annual layers might be expected to have. Varves in glacial deposits of other regions are commonly much thicker, but many varves in lake deposits and in some marine rocks are of about the same thickness as these laminations. Estimates of the expected thickness of annual layers in the Upper Cretaceous rocks near the Black Hills also appear to support the hypothesis. Three methods of estimation—(1) the rhythmic alternations in Upper Cretaceous rocks in eastern Colorado, which Gilbert suggested were formed during precession cycles, (2) the probable area of land draining into the

[55] Wells, R. C., The solubility of calcite in water in contact with the atmosphere and its variation with temperature: Washington Acad. Sci. Jour., vol. 5, pp. 617–622, 1915. Johnston, John, and Williamson, E. D., The rôle of inorganic agencies in the deposition of calcium carbonate: Jour. Geology, vol. 24, pp. 729–750, 1916. Wells, R. C., New determinations of carbon dioxide in water of the Gulf of Mexico: U. S. Geol. Survey Prof. Paper 120, p. 11, 1919. Rubey, W. W., Origin of the siliceous Mowry shale of the Black Hills region: U. S. Geol. Survey Prof. Paper 154, pp. 164–165, 1928.

[56] Barrell, Joseph, op. cit., p. 801.

[57] Rubey, W. W., op. cit., pp. 156, 167–169.

Upper Cretaceous sea and the present rate of erosion in the Mississippi drainage basin, (3) the total thickness of Upper Cretaceous rocks in the region divided by Barrell's estimate of the number of years in the Upper Cretaceous epoch—all indicate annual layers only slightly thinner than the observed laminations. A modification of the third method, in which the total thickness of Upper Cretaceous rocks is divided by estimates of Upper Cretaceous time based on the amount of salt in the ocean, indicates annual layers somewhat thicker than the observed laminations.

Thus the laminations are of about the right thickness to be annual; they were not caused by daily variations or by cycles several thousand years long. However, this is no proof that they are annual, for they might have formed every few months or years. In fact, more detailed comparisons (*a*) of the rate of eastward transgression of the Upper Cretaceous sea (calculated from the number of laminations and the distance of the transgression) with the probable average rate of strand-line movements to-day, and (*b*) of the length of Upper Cretaceous time indicated by the laminations with that estimated by Barrell, suggest that they may have formed every few years. Yet these two apparent discrepancies might be explained equally well (*a*) if the sea advanced rapidly then because it came in over a relatively flat surface, and (*b*) if Barrell's estimate of Upper Cretaceous time is too long or if there are many inconspicuous unconformities or diastems in the stratigraphic section.

Whether or not the laminations are annual, they suggest some of the conditions of deposition. The degree of preservation of the laminations indicates that the fine-grained organic shales accumulated in deep water or at times of gentle winds. If the laminations of different kinds formed during equal time intervals, their relative thicknesses indicate that, in general, the sandier shales and calcareous marls accumulated more rapidly and the finer grained and more organic shales more slowly.

COORDINATION OF CONCLUSIONS

Several of the conclusions reached in the separate discussions of different lithologic characteristics of these rocks overlap to a certain extent. Marshaled together, these overlapping conclusions acquire a higher degree of probability and the evidence on which they are based takes on a somewhat broader significance. Purely physical problems, such as the part played by adsorbed surface films in the behavior of argillaceous rocks, being neglected as outside the scope of this paper, the points of geologic interest, upon which several lines of evidence seem to converge, fall into three groups—the conditions of deposition of the sediments, the manner in which the rocks were later deformed, the bearing of these factors upon a proper appraisal of the oil and gas possibilities of the region.

Conditions of deposition.—Two features of the environment of marine sediments that influence to a considerable degree the nature of the sediments are the depth of water and the rate of deposition. In a general way, the temperature, agitation by waves, oxygen content, and habitability (which depends upon the availability of oxygen, light, and food) of sea water decrease with increasing depth. A consideration of the content of calcium carbonate, organic matter, pyrite, and fossils in the different specimens led to the conclusion that the more calcareous rocks probably accumulated in relatively warm water, in which many organisms lived and died. From the general relation of temperature and habitability to depth and from analogy with blue muds that are forming to-day, it seems that these calcareous rocks probably accumulated in relatively shallow water. This conclusion accords with that drawn from a study of the bedding laminations, for the noncalcareous shales are distinctly laminated, and the more calcareous rocks show evidence of loss of bedding by wave disturbance. In fact, it is possible that difference of depth and degree of wave agitation may in part account for the lack of sorting in the noncalcareous shales and the somewhat better sorting of the calcareous rocks.

The theoretical consideration of the conditions that might favor the preservation of calcium carbonate, organic matter, and pyrite together, supplemented by the analogy with present-day blue muds, and the empirical observation of the thickness of bedding laminations in the different rock types indicate that the organic and inorganic matter in the more calcareous rocks probably accumulated more rapidly than that in the fine-grained noncalcareous shales. The inverse relation between rate of deposition and depth of water that is thus suggested by these studies is not surprising. Areas where the water is shallow are commonly near the source or land from which sediment and food are carried into the sea. However, the relation is obviously not a general one that applies to all environments where marine sediments are laid down. The very fact that more detritus is available near shore means that a larger proportion of noncalcareous sediment is deposited there. And this sediment, unless the near-shore areas are sinking rapidly, is repeatedly eroded and redistributed by waves, so that the average rate of accumulation near shore is greatly reduced. It may well be that from deep water up to some unknown depth or some unknown distance from shore decreasing depth and increasing rate of sedimentation tend to give progressively more calcareous sediments, but that at still shallower depths or nearer shore this relation no longer holds.

Deformation of the rocks.—The loss of porosity with increasing depth of burial and steepness of dip, which is shown by these samples, implies a decrease of rock volume and an internal deformation of the rock mass. The fissility of the shales and the aggregate optical orientation of mineral particles also seem to show that there has been internal rearrangement of particles as the result of burial. Likewise, field evidence that shale beds have been squeezed thinner where steeply folded indicates loss of volume as the beds are tilted.

The apparent relations of the porosity to the depth of burial and to the steepness of dip, together with the partial confirmation of these relations from other lines of evidence, lead to several important corollaries. Compaction, or loss of porosity and increase of density, by burial is a continuous but decreasingly important process. Compaction by tilting indicates that horizontal compression deformed the rocks, that their apparent stratigraphic thickness depends upon the degree of folding, and that large volumes of water were squeezed out of them during folding. Although these rocks have yielded under load and pressure like incompetent members, by flattening and by bending into similar folds, they have not undergone true plastic or constant-volume deformation. Under both loading and tilting they have been thinned stratigraphically but not appreciably lengthened parallel to the bedding. However, at lesser and greater loads and pressures, these relations probably would not hold.

Occurrence of oil and gas.—Several lines of evidence seem to bear upon the oil and gas possibilities of these rocks. The percentages of chloroform-soluble organic matter in the samples apparently vary with the stratigraphic position, with the total amount of organic matter, with the composition of the organic matter, and with either the degree of deformation or the porosity of the rocks—it is impossible to say which, because the porosity depends in turn upon the degree of deformation. Perhaps all these relations might be deduced from theoretical considerations. Nevertheless it is worth noting that the suggested possibility that the deformation of the rocks may be an important factor in the generation of petroleum seems to accord with the results of drilling, for nearly all occurrences of oil in the Black Hills region are closely associated with steep dips and faulting. Porosity seems related both to the angle of dip or the shear and to the depth of burial or the load, varying inversely with the deformation. The percentage of soluble bitumens also seems to increase with the dip or shear; but, on the other hand, it seems to decrease with the depth or load. This observation suggests the possibility that there is a fundamental difference between the effects upon organic matter of straight compression and of sliding shear—in fact, this difference is a possible explanation of the contradictory results of published laboratory experiments.

Whether or not petroleum is derived from organic matter by folding, it must still, in order to form oil pools, migrate from the source rocks and accumulate in adjacent reservoir rocks or sandstones. The very fact that soluble bitumens are more abundant in steeply dipping than in horizontal beds might mean that the soluble bitumens remain where they are formed and do not migrate into interlaminated sandstones. However, the decrease of pore space with increase of dip implies that fluid was squeezed out of the rock, either upward along fissures or laterally through interlaminated sandstone beds, and presumably some of the more liquid hydrocarbons would be carried away with this fluid; that is, folding might not only form the petroleum but cause some of it to migrate from the source rocks into adjacent sandstones.

This suggestion, based on the study of a few specimens, is, of course, merely a possibility that must be tested by other investigations. This assumed effect of deformation on the generation and migration of petroleum and the more definitely observed stratigraphic distribution of chloroform-soluble organic matter in the rocks suggest possible applications or guides in the search for new oil fields in the Black Hills region.

A FLORA OF GREEN RIVER AGE IN THE WIND RIVER BASIN OF WYOMING

By Edward Wilber Berry

INTRODUCTION

The Wind River Basin, in west-central Wyoming, was explored by F. V. Hayden in 1859 and 1860 and by Meek and Hayden [1] in 1862. The geology of the basin was commented upon at greater or less length by Hayden [2] in 1869, by Meek [3] and Peale [4] in 1876, and by St. John [5] in 1883. J. L. Wortman collected Mammalia from the basin for Cope [6] in 1880 and spent the field seasons of 1891 and 1896 in the region under the auspices of the American Museum of Natural History,[7] which sent F. B. Loomis into the region in 1904, Walter Granger in 1905, and W. J. Sinclair and Walter Granger in 1910.[8] Parties in charge of Woodruff,[9] Hares,[10] Collier,[11] and Ziegler,[12] engaged in economic work, have visited parts of the basin and made brief contributions to its geology. Their interests were largely in reconnaissance work for oil, and they gave little attention to the Wind River and overlying Eocene formations.

Cope long ago correlated the Wind River formation with the Green River formation [13] on the basis of their respective positions in the Eocene succession of the western Tertiary, although there have been no means for direct faunal or floral comparisons, as the biota known from the Green River comprised fishes, insects, and plants, and that from the Wind River was made up, except for a few indecisive fresh-water mollusks, entirely of mammals.

Osborn,[14] basing his conclusions on the earlier work of Hayden and of Loomis, has, in several publications, discussed the Wind River section. He assigned to it a thickness of 1,200 to 1,400 feet and divided it into two faunal zones—the *Lambdotherium* and *Bathyopsis* zones. Osborn correlated the Wind River formation with the upper part of the Wasatch of the Big Horn Basin and with the lower part of the Huerfano formation of Colorado. As no fossil plants are known from the Huerfano or Bridger, and as those collected from the Wasatch have been confused with the Fort Union, it is impossible to check these correlations by means of the fossil plants.

Except for the forms listed below, identified by F. H. Knowlton and reported by Collier,[15] no fossil plants have been recorded from the Wind River Basin.

Lygodium kaulfussii Heer.
Dryopteris sp., new?
Cyperacites or Typha sp.
Myrica microphylla Newberry.
Quercus hardingeri Lesquereux.
Platanus or Liquidambar fruit.

It is quite possible that some specimens may have been collected by vertebrate paleontologists during their work in the Wind River Basin, and the coal and oil surveys of the United States Geological Survey may also have yielded specimens. If so these will be found in the American Museum of Natural History and the United States National Museum, respectively.

The material that forms the basis of the present preliminary report was collected for me by Mr. N. H. Brown, of Lander, Wyo., to whom I am indebted for many courtesies. The collection contains a few specimens from Crow Heart Butte and Lenore, but the bulk came from the vicinity of Tipperary, Fremont County, Wyo., in the SW. ¼ sec. 18, T. 6 N., R. 4 W. Wind River meridian. The plants occur in hard shale, whose position in the section is shown by the

[1] Meek, F. B., and Hayden, F. V., Descriptions of * * * fossils collected in Nebraska Territory: Acad. Nat. Sci. Philadelphia Proc., vol. 13, p. 433, 1862.

[2] Hayden, F. V., Geological report of the exploration of the Yellowstone and Missouri Rivers, p. 97, 1869.

[3] Meek, F. B., A report on the invertebrate Cretaceous and Tertiary fossils of the upper Missouri country: U. S. Geol. and Geog. Survey Terr. Rept., vol. 9, p. lxi, 1876.

[4] Peale, A. C., U. S. Geol. and Geog. Survey Terr. Eighth Ann. Rept., p. 148, 1876.

[5] St. John, Orestes, Report on the geology of the Wind River district: U. S. Geol. and Geog. Survey Terr. Twelfth Ann. Rept., p. 175, 1883.

[6] Cope, E. D., The badlands of the Wind River and their fauna: Am. Naturalist, vol. 14, pp. 745–748, 1880.

[7] Osborn, H. F., and Wortman, J. L., Fossil mammals of the Wasatch and Wind River beds: Am. Mus. Nat. Hist. Bull., vol. 4, pp. 135–144, 1892.

[8] Sinclair, W. J., and Granger, Walter, idem, vol. 30, pp. 83–117, 1911.

[9] Woodruff, E. G., The Lander oil field: U. S. Geol. Survey Bull. 452, pp. 3–36, 1911.

[10] Hares, C. J., Anticlines in central Wyoming: U. S. Geol. Survey Bull. 641, pp. 233–279, 1916.

[11] Collier, A. J., Gas in the Big Sand Draw anticline, Fremont County, Wyo.: U. S. Geol. Survey Bull. 711, p. 78, 1920.

[12] Ziegler, Victor, The Pilot Butte oil field, Fremont County: Wyoming State Geologist's Office Bull. 13, pp. 139–178, 1916.

[13] Scott, W. B., Memoir of Edward D. Cope: Geol. Soc. America Bull., vol. 9, p. 404, 1898.

[14] Osborn, H. F., Cenozoic mammal horizons of western North America: U. S. Geol. Survey Bull. 361, pp. 43–44, 1909.

[15] Collier, A. J., Gas in the Big Sand Draw anticline, Fremont County, Wyo.: U. S. Geol. Survey Bull. 711, p. 78, 1920.

following notes. Above the typical red-banded beds of the Wind River formation is about 25 feet of dark sandy clay, overlain by 40 to 50 feet of coarse cross-bedded buff sandstone. Above the sandstone is a thin bed of plant-bearing shale, overlain by 15 feet of thin-bedded sand, a second plant-bearing shale, and soft thin-bedded sandstone. So far as I know the Bridger has never been recognized in this area, but the possibility should not be lost sight of that the plant-bearing beds mentioned may correspond to a horizon in the lower Bridger. In fact, I am of the opinion that the plants described in the present paper are stratigraphically in what Sinclair and Granger [16] termed "so-called Bridger," and this conclusion is confirmed by Sinclair.[17]

The Green River formation, as developed in the Green River Basin, is a thick series of deposits showing a variety of lithologic phases of sedimentation, for the most part lacustrine, and it is quite possible, as Cope thought, that the typical banded beds of the Wind River formation, which underlie the "so-called Bridger" of Sinclair and Granger in the Wind River Basin, may also be of Green River age. This seems even probable, but I have no data for expressing an opinion on this point.

COMPOSITION OF THE FLORA

The flora described in this preliminary contribution consists of 41 identified species. These include 1 arthrophyte—a large *Equisetum*, which, except for its size, is of modern aspect; 6 pteridophytes, including 5 ferns and a water fern (*Salvinia*). The flowering plants are represented by 34 species, of which 6 are monocotyledons and the balance dicotyledons. The monocotyledons comprise a cat-tail (*Typha*), a large bur reed (*Sparganium*), three palms—two of which are fan and the third a feather palm, and a large-leaved monocotyledon of unknown botanic relationship. The dicotyledons represent 19 genera in 15 families and 13 orders. In addition to the forms definitely placed botanically there are 2 species of fruits referred to the genera *Nordenskiöldia* and *Carpites*, 2 to *Carpolithus*, and 3 flowers referred to the form genus *Antholithes*, in which the botanic position remains in doubt. The orders recognized are Juglandales, Myricales, Salicales, Fagales, Urticales, Geraniales, Sapindales, Rhamnales, Malvales, Laurales, Umbellales, Ebenales, and Rubiales (?)—all but the last two members of the more primitive choripetalous division of the Dicotyledonae. Only the Sapindales and Rhamnales have more than a single family represented, and each of these has two families. Of the 15 families all are represented by a single genus and 11 by a single species. The walnuts (*Juglans*) and figs (*Ficus*) are represented by 2 species each, and the soapberries (*Sapindus*) and aralias (*Aralia*) by 3 each. The most abundant types in the flora as a whole are *Salvinia*, *Lygodium*, *Sparganium*, *Sabalites*, *Juglans*, *Sapindus*, *Zizyphus*, and *Aralia*.

It is quite possible, perhaps I should say probable, that future more extensive collections would greatly alter the relative representation of these types. The plants thus far known fall into possibly two general associations, those that are aquatic or live on a wet substratum and those that may have inhabited dry ground. This phase will be discussed in the section devoted to a consideration of the ecologic conditions indicated.

The appended table gives the occurrences of the flora in other areas. The following are not known elsewhere:

- Equisetum tipperarense.
- Asplenium serraforme.
- Typha sp.
- Dryophyllum wyomingense.
- Fagara wyomingensis.
- Negundo fremontensis.
- Zizyphus wyomingianus.
- Laurus fremontensis.
- Aralia browni.
- Diospyros mira.
- Antholithes anceps.
- Antholithes browni.
- Antholithes fremontensis.
- Carpolithus browni.
- Carpolithus bridgerensis.

These 15 peculiar species constitute about 37 per cent of the known flora, or slightly more than one-third. Of the 26 species with an outside distribution 21, or 81 per cent, are common to the Green River formation. Two others, *Asplenium eolignitieum* and *Salvinia preauriculata*, that are not found in the Green River Basin occur in the Wilcox group of the Gulf Coastal Plain, and both are more common in the Wind River Basin, at what I believe to be a later horizon, than they are in the Coastal Plain. One or possibly two other species also occur in the Wilcox group, 3 are common to the Claiborne group, and 3 to the Jackson group of the Gulf Coastal Plain. Four are recorded from the Fort Union formation of the West, 1 from the Clarno formation and 1 from the Puget group of the Pacific coast region, and 1 from the Arctic Tertiary.

The most common species and the one that is most widely distributed is the climbing fern of the genus *Lygodium*, which in the Atlantic Coastal Plain section characterizes the late middle and upper Eocene Claiborne and Jackson. It would therefore not be inappropriate to name the horizon the *Lygodium* floral zone.

[16] See Sinclair, W. J., and Granger, Walter, Eocene and Oligocene of the Wind River and Big Horn Basins: Am. Mus. Nat. Hist. Bull., vol. 30, art. 7, 1911.

[17] Sinclair, W. J., personal communication January 30, 1924.

The Green River flora of the Green River Basin, as revised recently by Knowlton,[18] consists of about 80 species, of which 21 occur in the deposits in the Wind River Basin. It is legitimate to suppose that if the Wind River region were more thoroughly known, a great many more identical species would be brought to light. The only really conspicuous Green River plant not discovered in the present small collections from the Wind River Basin is *Acrostichum hesperium* Newberry, and this is found in the upper Claiborne of the Mississippi embayment.

AGE

Of the 21 species common to the Wind River Basin and the Green River Basin, 15, or about three-fourths, are not known to occur anywhere else than in these two areas, which are not geographically remote. The conclusion seems to be reasonably well established that these deposits in the Wind River Basin are of very nearly the same age as those containing the Green River flora in the Green River Basin.

Although it is thus fairly clear that this flora in the Wind River Basin is overwhelmingly Green River in its facies, it should be pointed out that the problem is not so simple as that statement would indicate. In the type section the Green River formation is a thick series, well marked lithologically from the underlying Wasatch and the overlying Bridger. No fossil plants have ever been found, so far as I know, in either the Wasatch or the Bridger in this section.

As Bradley[19] has shown, when the Green River is traced by lithology from its area of maximum development toward the margin of its basin of deposition it is found to thin and gradually disappear, thus bringing the Wasatch and Bridger into contact. It seems to be a legitimate supposition that deposition contemporaneous with that which formed the typical Green River was taking place in the Wind River Basin and elsewhere in Wyoming, and that at least the Wind River formation in the Wind River Basin, a part of the "so-called Bridger" of Sinclair and Granger, and perhaps the upper part of the Wasatch of the Wind River Basin represent the time equivalent of the maximum Green River of the Green River Basin.

As the Wasatch flora has never been studied little can be said about it, although according to Knowlton it was not markedly distinct from the Fort Union. No plants have ever been reported from beds referred to the Bridger, so that it is not possible to make any comparisons of Bridger plants with the present flora, but the considerations just mentioned make it possible that this flora may be contemporaneous with the Green River flora, and if so the differences noted may be explained as due to slightly different environments in the two regions; or it may ultimately be shown that this fossil flora in the Wind River Basin is slightly younger than the Green River flora, particularly as the bulk of the known Green River plants have come from the upper half of the Green River formation.

Many elements of both the Wind River and the Green River floras await discovery. This is indicated not only by the relatively small size of both, but also by the fact that in a contribution made since this paper was written Brown [20] has added 48 species to the Green River flora, nearly all of which came from a single locality in Colorado and a large majority of which are new species.

These questions can be definitely settled only by detailed areal work in Wyoming. Although Osborn, as already remarked, correlated the Wind River with the upper Wasatch of the Big Horn Basin and the lower Huerfano of Colorado, I believe all three are somewhat younger than the position usually assigned to them by vertebrate paleontologists, and this point comes within the province of paleobotanic correlation when the Green River section is compared with that of the Atlantic Coastal Plain and with the standard European section.

The general stage of development of the Green River flora, the stage in the evolution of Tertiary climates in North America, and a few identical species indicate that the flora is of about the same age as that of the Claiborne group of the Gulf Coastal Plain. The Claiborne, which I at first thought was of Lutetian age, is now rather definitely correlated with the subsequent Auversian stage of the European section, and I believe this to be the approximate age of the Green River flora and the flora from the Wind River Basin described in the present contribution. The Wind River formation may be Lutetian, but I much doubt if it is any older. At any rate, the Green River is well up in the middle Eocene.

The few invertebrates associated with the plants in the Wind River Basin are either Green River or Bridger species, but they are of little significance as time indicators.

ECOLOGY INDICATED

A flora of only 41 species is naturally much too small for the reconstruction of the environment of the time in which they lived, but it does afford some fairly definite indications of the probable ecology of that time. In addition to the species determined,

[18] Knowlton, F. H., U. S. Geol. Survey Prof. Paper 131, pp. 133–182, 1923.

[19] Bradley, W. H., Shore phases of the Green River formation in northern Sweetwater County, Wyo.: U. S. Geol. Survey Prof. Paper 140, pp. 121–131, 1926.

[20] Brown, R. W., Additions to the flora of the Green River formation: U. S. Geol. Survey Prof. Paper 154, pp. 279–293, 1929.

there are others too incomplete for identification and considerable fossil wood. Some of the sediments are water-laid clays that appear to have a considerable percentage of volcanic ash. Other beds are coarse sandstones and appear to represent channel deposits. Still others have every indication of being eolian—an inference I draw from the curled condition of the fossil leaves and their lack of conformity with the bedding of the matrix. It therefore seems to be a reasonable conclusion that the sediments are to a large extent river and flood-plain deposits.

In an attempt to arrive at the environmental conditions the discussion may be divided into the relations of the plants to rainfall, humidity, and ground-water conditions and their relations to temperature conditions as indicated by the distribution of their existing representatives.

At least four may be called hygrophilous. These are *Salvinia*, which is a floating plant in slack waters; *Equisetum*, not only large but with the rhizomes and attached tubers of large size embedded in the mud in which they grew; the cat-tail (*Typha*) and the large bur reed (*Sparganium*), which are distinctly palustrine types.

There are a considerable number which normally require an abundance of ground water and might be termed mesophytic. In this category I would place the representatives of the genera *Geonomites*, *Musophyllum*, *Ficus*, *Salix*, *Negundo*, *Lygodium*, *Asplenium*, *Dryophyllum*, *Fagara*, and *Dryopteris*.

The following might also be legitimately considered as belonging in the same category, because, without exception, they require deep soils and plenty of water, but I keep them separate, as less definitely requiring a distinctly wet substratum: *Sabalites*, *Juglans*, *Zizyphus*, *Ampelopsis*, *Grewiopsis*, *Laurus*, and *Diospyros*. If the *Sabalites* was like the modern cabbage palmetto in its requirements, which is by no means established, it required such conditions as I have indicated. *Zizyphus* has a somewhat restricted range in the modern flora and is chiefly developed in the Indo-Malayan region of heavy rainfall, although it is sparsely represented in the American Tropics. A modern species naturalized in the wet country of southeastern Louisiana forms extensive thickets there, and I would be inclined to think that the present country between New Orleans and Lake Charles is not unlike what Wyoming was during Green River and Bridger time, in respect not only to its general aspect but also to rainfall, humidity, and temperature extremes. If the form which I have referred to the form genus *Laurus* is really a *Sassafras* then all these seven genera except *Grewiopsis*, which is somewhat uncertain in its botanic status, are either confined to or have their greatest development in the mesophytic broad-leaved forests of the Southeastern States. From a consideration of the abundance of palms in the middle Eocene Wind River Basin and the presence of such a large-leaved monocotyledon as *Musophyllum*, this geographic comparison might be still more restricted to the sea-border States from South Carolina to Louisiana.

There remain but three genera which might be considered to be indicative of a dry substratum, and in none of the three is the evidence conclusive: I refer to *Myrica*, *Sapindus*, and *Aralia*, the last a somewhat loose attribute for forms which I believe really represent the existing genus *Oreopanax*. All three of these genera occur in modern floras in regions of abundant rainfall as well as in drier regions. Certain species of *Myrica*, for example, are found in deep swamps; others characterize barren soils and diminished ground water rather than lessened rainfall, as on sand dunes, whereas others are distinctly inhabitants of wet situations, so that *Myrica* might well represent a plant that should have been enumerated in one of the preceding categories, but I keep it separate to emphasize the possibility that it might have been a dry-soil form. The soapberries (*Sapindus*), represented by three not especially well-defined species in the fossil flora under discussion, belong to a genus with numerous living and fossil species, many of which, along with other members of the family, are found in regions with a monsoon climate. Many are coastal species in regions of strong insolation and a sandy substratum, developing evaporation-resisting coriaceous leaves, like the modern *Sapindus saponaria* Linné. This fact and the range of *Sapindus drummondi* Hooker and Arnott, from the moist clay soils of Louisiana to the dry limestone uplands of Arkansas, southern Kansas, and westward through Texas to the mountain valleys of southern New Mexico, Arizona, and northern Mexico, suggest that the fossil species may indicate either dry or wet soils and are not decisive either way. The third genus, *Aralia*, is believed to represent the existing *Oreopanax* Lecaisne and Planchon, a genus of considerable size with several species in the more or less arid uplands of Mexico and Central America, although they are by no means confined to such an environment but are rather generally distributed in tropical America.

It may be concluded that Wyoming during middle Eocene time was in general a well-watered forested region, as none of the plants enumerated in the preceding paragraphs are necessarily associated with dry soils and diminished rainfall, nearly all are forms characteristic of regions with an ample rainfall and abundant ground water, and several are aquatic types.

The evaluation of fossil floras or faunas in terms of temperature is beset with many difficulties and clouded by preconceived notions, many of which may be erroneous. Even when all personal bias is eliminated, such attempts are fraught with extreme uncertainties growing out of the complexity of factors involved.

All the genera represented in this fossil flora are old and large genera whose modern species flourish under very diverse conditions. Figs, palms, and laurels are commonly associated with the Tropics, where, at the present time, they are most abundantly represented, but figs ripen their fruits in most years as far north as Maryland and in the temperate altitudinal zone in Bolivia. Palms range in the Western Hemisphere from North Carolina southward to 34° south latitude in Chile and Argentina; and in the Old World from about 43° north latitude in Europe to about 45° south latitude in New Zealand. They exist without ripening their fruits at many places in still higher latitudes—date palms, for example, growing in cultivation out of doors in southern France.

Nearly all the Lauraceae are tropical, but several extend into our Southern States; the genus *Cinnamomum* is especially characteristic of warm-temperate rain forests, and our familiar *Sassafras* is hardy and often of large size in southern New England. The ebony family is another group the great majority of which are tropical, and yet our familiar persimmon ranges northward to Connecticut, southern Ohio, and southeastern Idaho.

Confining our attention to forms identified from the Wind River Basin and deriving our notions from the distribution of the modern relatives of the fossil species we may note that *Equisetum* is found in almost all latitudes—Arctic as well as tropical—but most abundantly in the Temperate Zone. Modern Salvinias range from the Equatorial Zone to about 45° north latitude. *Asplenium* is a very large and cosmopolitan genus, but the modern species most like the fossils are tropical. *Lygodium* is mainly tropical but extends for considerable distances into the Temperate Zone in Atlantic North America, Japan, and New Zealand. *Dryopteris* is mainly tropical but is commonly represented and widely distributed in the Temperate Zone, there being 13 species in the existing flora of Canada. *Typha* is wide ranging and of no definite climatic significance. *Sparganium* is also very wide ranging and found chiefly in temperate and cool rather than warm regions. The three palms, if their modern affinities are what they are supposed to be, are distinctly warm-temperate rather than tropical types. *Musophyllum* is of uncertain significance, and the most that can be said for it is that it indicates moisture and a fairly long growing season. *Juglans*, *Myrica*, and *Dryophyllum* are distinctly temperate types of varied habitat. *Ficus*, in this flora somewhat inconclusive as to identity, is also as much at home in a warm-temperate as in a tropical environment. *Fagara* and *Sapindus* are warm-temperate and tropical. *Negundo* is distinctly temperate. *Zizyphus* is mainly tropical but has been found fossil in all latitudes and withstands the winters of our Southern States—in Louisiana, for example. *Ampelopsis* reaches Canada in the modern flora. *Grewiopsis* is of uncertain identity and significance. If the *Laurus* represents *Sassafras*, as I believe it does, it is of temperate rather than tropical significance. *Aralia* is uncertain, but if it represents *Oreopanax*, as I have suggested, it is equally indicative of either temperate or tropical conditions, and so is *Diospyros*. The remainder of the flora, consisting of fruits and flowers, can not be evaluated, as their botanic affinities are uncertain.

The Green River formation contains an abundance of fishes at certain horizons in the Green River Basin. Other layers are packed with plant remains, and the oil shales contain innumerable bacteria, blue-green algae and fragments of other algae, saprophytic fungi, mosses and fern-spore exines, and pollen of the higher plants. Coniferous pollen has been found in the oil shales and might have been brought into the area of sedimentation by the wind from higher altitudes or even from a great distance, as such pollen which must have come from the continent of Europe, is found at the present time in Svalbard (Spitsbergen).

Knowlton,[21] from a consideration of the more extensive flora in the Green River Basin, concluded that it contained representatives of two ecologic groups—a lowland flora of warm-temperate environment and an upland flora, including *Salix*, *Myrica*, *Juglans*, *Quercus*, *Rhus*, *Acer*, and *Ilex*, that may have lived in a cooler environment.

According to Cockerell as quoted by Knowlton,[22] 296 species of insects are known from the Green River formation, and these appear to show a mixed character, the numerous lantern flies (Fulgoridae) being tropical in their affinities, whereas other groups—for example, the crane flies (Tipulidae)—show temperate affinities. Knowlton's conclusion that the Green River biota consists of certain elements that indicate tropical surroundings and others that indicate cooler, perhaps temperate conditions is one way of stating the results of comparisons with recent floras. I believe, however, that it may be somewhat misleading. Certainly the ensemble suggests a warm and genial climate, but that this borders on tropical, in any precise use of that term, or that the winter season was without frost is most doubtful. Nor is it necessary to assume that the fossils represent the mechanically mixed representatives of a lowland and an upland association.

I consider it utterly impossible to estimate the summer's heat or the winter's cold, or to attempt to give mean annual temperatures, at the remote time repre-

[21] Knowlton, F. H., U. S. Geol. Survey Prof. Paper 131, p. 147, 1923.
[22] Idem, p. 148.

sented by this flora, but a picture of temperature requirements can be given by citing a modern region where such a fossil flora would find optimum conditions for growth and reproduction. I have already suggested southern Louisiana as such a region, as regards temperature, rainfall, and humidity. This comparison might be somewhat extended, and I close my remarks on this subject with the statement that I known of no member of the Green River flora either in its restricted representation in the Wind River Basin or in its more fuller representation in the Green River Basin that would not be perfectly at home somewhere in the region between South Carolina and Louisiana in our South Atlantic and Gulf States.

Hence, if general terms are demanded, the Green River flora should be called warm-temperate and not tropical.

There are a few poorly preserved invertebrates associated with the plants at the Tipperary locality. I am indebted to Dr. Henry A. Pilsbry for the determination of these forms. They comprise *Physa bridgerensis* Meek; *Planorbis cirratus* White, or a closely related form; *Unio* sp., of the group *U. tellenoides* Hall, *U. shoshonensis* White, and *U. washakiensis* Meek, probably identical with one or the other of these; and *Oreohelix* sp. The first three are fresh-water forms, thus confirming the evidence as to environment deduced from the associated plants, and are closely comparable with forms from the Green River and Bridger formations. The fourth is a land snail of very great interest but unfortunately too incomplete for specific determination.

The statement in an earlier paragraph that the manner of preservation of some of the leaves suggests that they were buried in wind-blown sediments is not discordant with the conclusions above set forth, because locally in valleys and river beds during dry seasons such sediments are frequently observed in regions where the general climate might properly be termed "humid."

Silicified wood occurs in these beds. One section, which I have not attempted to identify specifically, is a species of *Pityoxylon* with well-marked annual rings clearly indicative of seasonal change, either hot and cold or wet and dry. It is impossible to choose between these alternatives, but as the zone of so-called summer wood is wide, the tracheids not greatly reduced in size, and their walls not inordinately thickened, I conclude that the contrast in seasons was not extreme and that the coldness or dryness, whichever it was, was not severe—in other words, that the seasonal contrasts throughout the year were not extreme. One difficulty in attempting to evaluate petrified woods is the possibility that the wood may have floated a long distance from some other environment before silicification or that it may have been reworked [illegible] older beds.

Another consideration frequently lost sight of in discussions of this sort is that the conclusions arrived at are valid only for the assemblage studied and may not be extended to include the whole formational unit. Although it seems to me that there is a very close ecologic agreement between the present flora and that of the Green River formation, I frankly admit that the present flora, although clearly a riverside, palustrine, or pond-border assemblage, may represent only one of several ecologic assemblages that may have been contemporaneous during the time of deposition of the Green River and Bridger formations.

Range of the species of the flora of Green River age found in the Wind River Basin

	Green River formation	Denver formation	Fort Union formation	Clarno formation	Puget group	Atlantic Coastal Plain: Wilcox group	Atlantic Coastal Plain: Claiborne group	Atlantic Coastal Plain: Jackson group	Arctic Tertiary	California
Equisetum tipperarense										
Salvinia preauriculata						×				
Danaea coloradensis	×									
Dryopteris weedii			×							
Asplenium serraforme										
Asplenium eolignitum						×				
Lygodium kaulfussii	×		×	×	×		×	×		
Typha sp										
Sparganium antiquum	×									
Sabalites florissanti	×									
Sabalites powellii	×									
Geonomites haydenii	×									
Musophyllum complicatum	×									
Juglans alkalina	×									
Juglans occidentalis	×		×			×				
Myrica ludwigii	×									
Salix sp	×									
Dryophyllum wyomingense										
Ficus wyomingiana	×									
Ficus ungeri	×		?				×	×		
Fagara wyomingensis										
Sapindus dentoni	×						×	×		
Sapindus obtusifolius	×									
Sapindus winchesteri	×									
Negundo fremontensis										
Zizyphus wyomingianus										
Ampelopsis tertiaria	×									
Grewiopsis wyomingensis	×									
Laurus fremontensis										
Aralia whitneyi			×							×
Aralia browni										
Aralia notata denticulata		?	?			?				
Diospyros mira										
Sambucus? winchesteri	×									
Nordenskiöldia borealis	×								×	
Carpites newberryanus	×									
Antholithes anceps										
Antholithes browni										
Antholithes fremontensis										
Carpolithus browni										
Carpolithus bridgerensis										
Sporophylls										

SYSTEMATIC ACCOUNT

Phylum ARTHROPHYTA

Class CALAMARIAE

Order EQUISETALES

Family EQUISETACEAE

Genus EQUISETUM Linné

Equisetum tipperarense Berry, n. sp.

Plate 6, Figures 4–10

This species is represented by stems, tubers, and detached sheaths. It is possibly identical with some of the nondescript Equisetums already known from the Eocene, but this can not be certainly determined, and it is therefore described as new. Nor is it demonstrated that these various classes of detached remains of *Equisetum* found abundantly in the Wind River Basin appertain to a single botanic species, as they are merely associated and not in organic connection, but the presumption is in favor of such a view. The specific name is given in allusion to the locality.

The stems have numerous flat ridges about 1 millimeter wide separated by narrow grooves, and the position of the branch scars suggests that the preservation is in the nature of internal casts. If this is so, then the stem tissue was extremely thin, as it is preserved in places as a carbonaceous residuum. These stems were of relatively large size; the smallest leaf sheath is almost a centimeter in diameter; a larger stem only slightly flattened is 2.8 centimeters in maximum and 1.5 centimeters in minimum diameter; and flattened fragments of stems are as much as 4 centimeters in width. Several nodes are preserved, but no complete internodes; the longest distance of internode preserved measures 9 centimeters. In the small nodal sheath figured the sheath is about 6 millimeters long, and the upper third of this distance is represented by the free pointed leaves. There are no traces of sheaths in connection with the stem fragments.

Immediately above the nodal diaphragm, where the vascular strands alternate, is a row of small tubercular scars about 0.75 millimeter in diameter, with hollow centers, one for each vascular strand of the stem; and these apparently represent scars of the whorl of branches that were present at the nodes. The tubers are numerous, usually detached or in whorls of five. In one whorl they are attached to a fragment of a rhizome. These tubers vary somewhat in size and form. They are all large, slightly flask shaped, and more or less crushed and in this condition are about 3 centimeters long and 1.5 centimeters in maximum diameter. They are of a type commonly referred to *Equisetum*, rather larger than in most Tertiary species but comparable in this respect to tubers from the Lance formation of Canada which I have referred to the Arctic Eocene species *Equisetum arcticum* Heer.

The species from the Wind River Basin must have rivaled in size the South American *Equisetum giganteum* Linné, which is the largest living species. The genus is an old one, extending back at least as far as the dawn of the Mesozoic, when it was represented by very large forms, one of which is found in the Triassic "Red Beds" near the locality where the fossil plants were collected in the Wind River Basin. The existing species number 25 to 30 and are widely distributed, usually growing in wet habitats. They are most abundant in the Temperate Zones but extend to the Arctic, and the largest occur in the wet Tropics. Tertiary species are numerous in both Europe and America.

The American Eocene species comprise the following: Fort Union 5, Wilcox 1, Claiborne 1, Green River 1, Clarno 1, Kenai 1, Puget 1. It should be pointed out that a number of these are vague, all are fragmentary, and beyond indicating the undoubted presence of *Equisetum* they are of slight botanic interest. The Green River species *Equisetum wyomingense* Lesquereux, with which one naturally compares the plant from the Wind River Basin, is rather vaguely characterized and figured by Lesquereux,[23] although Newberry [24] gives a better figure of what he considered the same species. This shows the Green River form to have been more slender, with stouter ribs and much longer leaf sheaths and free portions of the leaves. There can be no question with respect to the distinctness of the two.

The present species adds a characteristic and striking element to this flora, and in common with several other elements, it emphasizes the swampy habitat of at least part of the area of sedimentation. In post-Eocene time the climate of this area was apparently too dry for an abundance of *Equisetum*, even the later Eocene species coming from the humid Pacific coast region.

Occurrence: Tipperary, Wyo.

[23] Lesquereux, Leo, The Tertiary flora, p. 69, pl. 6, figs. 8–11, 1878.

[24] Newberry, J. S., U. S. Geol. Survey Mon. 35, p. 15, pl. 65, fig. 8, 1898.

Phylum **PTERIDOPHYTA**

Class **HYDROPTERIDAE**

Order **HYDROPTERALES**

Family **SALVINIACEAE**

Genus **SALVINIA** Adanson

Salvinia preauriculata Berry

Plate 6, Figures 1–3

Salvinia preauriculata Berry, Torreya, vol. 25, p. 116, fig. 4, 1925.

Remains of a species of *Salvinia* are abundant in these deposits. It may be incompletely characterized as follows: Leaves relatively thick, elliptical, with a rounded apex and a rounded or slightly cordate base, varying somewhat in size, the maximum dimensions being 16 millimeters in length and 10 millimeters in width. The midvein is well defined. The laterals are thin, nearly straight, diverging at regular intervals, very ascending in the tip, the angle of divergence decreasing regularly proximad, the basal being sometimes even slightly descending; they are connected by numerous thin, poorly preserved, oblique nervilles. The tubercles or pits lie in rows between the laterals and are usually well marked. No sporocarps have been observed. Two of the specimens, one of which is figured, show matted masses of these small characteristic floating leaves, but no traces of the submerged leaves have been detected.

This species, which is associated with fruits of *Sparganium* and unidentified sporophylls, is distinct from previously described forms but appears to be identical with undescribed material from the Wilcox group of western Tennessee.

The existing species of *Salvinia* number about a dozen, and they occur chiefly in the equatorial regions of the world, especially in South America. One species, *Salvinia natans*, ranges from southern France to India and northern China and has been reported from several localities in the United States.

The nominal fossil species also number about a dozen. The oldest of these are the present species and *Salvinia zeilleri* Fritel,[25] from the Sparnacian stage of the Paris Basin Eocene. Both of these early Eocene species are represented by unusually complete and characteristic material. A third rather well-marked species is recorded from the Puget group of Washington, which I believe to be late Eocene or Oligocene. Oligocene species are known from France, Saxony, and Bohemia, and Miocene species are recorded from Colombia, Virginia, Germany, Bohemia, Switzerland, Transylvania, Tonkin, China, and Japan. These have been reviewed recently by Florin.[26]

The present species is named from its resemblance to the existing *Salvinia auriculata* Aublet, which ranges from Cuba to Central America and Paraguay.

Occurrence: Tipperary, Wyo.

Class **LEPTOSPORANGIATAE**

Order **POLYPODIALES**

Family **POLYPODIACEAE**

Genus **ASPLENIUM** Linné

Asplenium serraforme Berry, n. sp.

Plate 8, Figure 1

Frond unknown. Pinnules of considerable size, lanceolate with an extended acuminate tip. Texture coriaceous. Margins irregularly serrate, with somewhat unequally developed, rather widely spaced and often acutely serrate small teeth. Midvein fairly stout. Laterals diverging from the midvein at very acute angles, relatively straight in their courses, usually once dichotomously forked at varying distances from the midvein, relatively closely spaced, subparallel and terminating in the margins. No traces of fructifications.

This species is unfortunately based upon very incomplete scraps and in the absence of fruiting characters is identified with some doubt. It appears to be an entirely new type in our American Tertiary floras and is moreover very similar to several Old and New World existing species of *Asplenium*. It is especially like the pinnules of *Asplenium serra* Langsdorf and Fischer, which ranges from the Antilles and Central America to Brazil. Among previously described fossil forms it is much like *Asplenium issiacense* Fritel,[27] from the Sparnacian of the Paris Basin.

Occurrence: Tipperary, Wyo.

Asplenium eolignitICUM Berry

Plate 8, Figures 2–4

Asplenium eolignitica Berry, U. S. Geol. Survey Prof. Paper 91, p. 167, pl. 11, fig. 3, 1916.

Gymnogramma haydenii Lesquereux, The Cretaceous and Tertiary floras, p. 122, pl. 19, fig. 2, 1883 (not Lesquereux, 1878).

This species was based upon rather fragmentary material from the lower Wilcox of Mississippi and the upper Wilcox of Arkansas and Texas. To it were re-

[25] Fritel, P. H., Jour. botanique, 2d ser., vol. 1, No. 8, p. 190, 1908.

[26] Florin. R., Eine Übersicht der fossilen *Salvinia*-Arten mit besonderer Berücksichtigung eines Fundes von *Salvinia formosa* Heer im Tertiär Japans: Geol. Inst. Upsala Bull., vol. 16, pp. 243–260, pl. 11, 1919.

[27] Fritel, P. H., Soc. géol. France Mém. 40, p. 16, pl. 1, figs. 1–4, 1910.

ferred specimens from the Denver formation of Colorado which Lesquereux referred to *Gymnogramma haydenii.* As I remarked in the description of the Wilcox material, it is possible that this represents the same botanic species as the rather widespread Eocene type known as *Pteris pseudopennaeformis* Lesquereux.

There is a considerable amount of material from the Wind River Basin representing this species, and as it shows several parts of the frond the description may be considerably amplified.

Fronds large pinnatifid. Pinnae large, linear lanceolate, inequilateral basally, borne rather close together in an oblique position on a stout rachis that appears to have a prominent median ridge, which may represent the vascular cylinder of the flattened specimens. The texture is coriaceous. The margins are entire below and prominently and irregularly serrate-toothed in the median region, the teeth becoming less prominent distad. The proximal margin is slightly decurrent, and the distal margin ends acutely a slight distance above the rachis. The midvein is stout, prominent, and with a central ridge. The laterals are thin, evenly spaced, diverging from the midrib at acute angles, curving outward and terminating at the margins. Nearly all the numerous laterals are dichotomously forked, usually near their base. Size variable; large specimens may have a length of 15 centimeters and a maximum width medially of 3 centimeters. The number of free pinnae on a frond is unknown; probably they are free for two-thirds of the whole length of the frond. In the upper one-third they become simplified, and the laminae become more or less confluent. No complete specimens are available, but several fragments show the features brought out in Figure 4. The margins become entire over greater distances; the teeth become dentate instead of serrate; the form may become somewhat falcate, and the size is less. The proximal basal margins are prominently decurrent on the rachis, and the distal margins connect by round or sharp sinuses with the proximal lamina of the superjacent segment. These segments diminish rapidly in size to the somewhat irregular terminal acuminate segment. No traces of fructifications have been observed.

The species was a fine, large one. It is not unlike *Asplenium issiacense* Fritel,[28] of the Sparnacian of the Paris Basin. Among previously described forms it is closest, however, to some of the forms referred to *Asplenium subcretaceum* Saporta, as recorded by Friedrich from the lower Oligocene of Saxony.[29] Among recent forms it may be compared with several species, among them *Asplenium serra* Langsdorf and Fischer, of tropical America.

Occurrence: Tipperary, Wyo.

FIGURE 4.—Restoration of the terminal part of a frond of *Asplenium eolignitícum* Berry

Genus DRYOPTERIS Adanson

Dryopteris weedii Knowlton

Plate 7, Figure 1

Dryopteris weedii Knowlton, U. S. Geol. Survey Mon. 32, pt. 2, p. 669, pl. 80, fig. 8; pl. 81, fig. 2, 1899.

This handsome species was described by Knowlton from specimens found in beds in the Yellowstone National Park which he referred to the Fort Union formation. There are four specimens in the collection from Tipperary, showing several pinnae attached to the winger rachis and two nearly complete

[28] Fritel, P. H., Soc. géol. France Mém. 40, p. 16, pl. 1, figs. 1–4, 1910.

[29] Friedrich, Paul, Beiträge zur Kenntniss der Tertiärflora der Provinz Sachsen, p. 77, pl. 8, figs. 1–4, 1883.

pinnae. These can not be differentiated from this species, although all the Tipperary specimens are sterile.

The genus *Dryopteris* as delimited by Christensen, our foremost authority, contains more than 1,000 existing species, divided into 10 groups or subgenera. The majority of these are tropical, although various species are widely distributed in the Temperate Zone. In North America, for example, 13 species are found in Canada, and 7 of these range northwestward to Alaska and 4 northeastward to Labrador. The genus in consequence of its numerous species and wide range is found in a variety of habitats, but it may be considered a distinctly mesophytic type, and the temperate species are prevailingly forms of moist woodlands.

Many fossil species have been referred to *Dryopteris*, and they are definitely recognized at least as far back as the Cretaceous.

Occurrence: Tipperary, Wyo.

Family SCHIZAEACEAE

Genus LYGODIUM Swartz

Lygodium kaulfussii Heer

Plate 7, Figures 2, 3

Lygodium kaulfussii Heer, Beiträge zur nähern Kenntnisse der sächsisch-thüringischen Braunkohle, p. 3, pl. 8, fig. 21; pl. 9, fig. 1, 1861.

There is no need to describe this splendid species of *Lygodium*. It ranges through the Eocene of North America and is found from the middle Eocene to the Oligocene of Europe. In the Mississippi embayment region it characterizes the Claiborne or middle Eocene, being absent in the lower Eocene of that region, where the genus is represented by several other species.

It is rather fully discussed in my account of the Claiborne flora,[30] and this discussion need not be repeated in the present connection. This species appears to have been abundant in the Wind River Basin, where it is represented by both sterile and fertile pinnae. It may be distinguished from *Lygodium binervatum* (Lesquereux) Berry,[31] the common Wilcox species, by its narrower, more elongated lobes, finer and more close-set branching venation, and 4 to 5 lobed form, *binervatum* having usually two short and broad lobes and stouter veins. Fertile pinnae are not rare in the material from the Wind River Basin and show 10 to 15 pairs of sporangia per pinnule.

[30] Berry, E. W., U. S. Geol. Survey Prof. Paper 92, p. 39, pl. 3, figs. 1, 5, 1924.

[31] Berry, E. W., U. S. Geol. Survey Prof. Paper 91, p. 165, pl. 10, figs. 3–8, 1916.

The genus *Lygodium* contains about 25 existing species, scandent in habit, mostly confined to the Tropics and subtropics but extending into the warmer moister parts of the Temperate Zone for considerable distances in Atlantic North America, New Zealand, and Japan.

Occurrence: Lenore and Tipperary, Wyo.

Phylum SPERMATOPHYTA

Class MONOCOTYLEDONAE

Order PANDANALES

Family TYPHACEAE

Genus TYPHA Linné

Typha sp.

Fragments of a monocotyledonous leaf of a type commonly referred to the genus *Typha* occur in the Wind River Basin. The leaves are of considerable size, with well-marked parallel linear veins connected by innumerable cross veinlets. Although this venation, which is characteristic of *Typha*, is shared by *Sparganium* and various sedges, paleobotanists have generally considered it to represent *Typha*. The genus has been recorded from scattered localities in North America and Europe from the Upper Cretaceous onward, and its occurrence in the Fort Union (Paskapoo) Eocene of western Canada is reported.

Occurrence: Tipperary, Wyo.

Family SPARGANIACEAE

Genus SPARGANIUM Linné

Sparganium antiquum (Newberry) Berry

Plate 8, Figure 5

Brasenia? antiqua Newberry, U. S. Nat. Mus. Proc., vol. 5, p. 514, 1882; U. S. Geol. Survey Mon. 35, p. 93, pl. 68, fig. 7, 1898.

Pontederites hesperia Knowlton, U. S. Geol. Survey Prof. Paper 131, p. 154, pl. 36, fig. 6, 1923.

Sparganium antiquum Berry, Bot. Gaz., vol. 78, p. 346, figs. 1–7, 1924.

Plants of considerable size, at least 40 centimeters tall; with stout, erect, parallel-veined stems; with a few alternate leaves with sheathing bases. Leaves broadly lanceolate, with numerous, essentially parallel, longitudinal veins, convergent toward the base, where they unite to form a not very conspicuous midvein. This traced upward becomes indistinguishable from the other veins in the upper half of the leaf. These veins are conspicuously connected by transverse veinlets, which die out on the leaf sheaths; the sheaths merge proximad with the stems, from which they are indistinguishable in appearance.

The inflorescence was large and branched; fragments showing three lateral branches are preserved.

These branches are at nearly right angles to the stem, rather stiff and curved, and each bears from 8 to 10 nearly sessile heads, at somewhat irregular intervals and progressively diminishing in size outward. It is not possible to determine whether or not the distal heads were staminate, as in the recent species of *Sparganium*, as the details of organization are not uniformly well preserved.

In 1882 Newberry described some objects from the Green River beds of Wyoming which he called *Brasenia? antiqua*. It should be noted that these are entirely different from the Canadian forms called *Brasenia antiqua* by Dawson, which have since been referred to the genus *Nelumbo*. The middle Eocene remains that Newberry thought represented an extinct species of *Brasenia* consisted of branched stems bearing spheroidal fruits made up of units which he described as small club-shaped pods, although he stated that the specimens were too imperfectly preserved to permit any decisiveness in their identification. No additional material similar to that studied by Newberry, which is preserved in the National Museum (No. 7018), has since been collected from the Green River beds.

In Knowlton's revision of the flora of the Green River formation, published in 1923, he described certain leaves under the name *Pontederites hesperia* and compared them with those of the existing pickerel weed, *Pontederia*. He believed that these leaves represented some Eocene member of the Pontederaceae. They lacked their basal portions but were of considerable size, with mostly parallel, longitudinal veins connected by transverse veinlets.

The present collections contain slabs of what appears to be a tuff, many of them covered with fruit heads and dissociated fruits and stem fragments that are identical with those from the Green River formation that Newberry referred to *Brasenia*. Several of these fruit heads are attached to the branches upon which they were borne in life. These heads are of various sizes, ranging from 5 to 12 millimeters in diameter, and are made up of numerous small, symmetrically beaked fruits, arranged radially on a spherical central receptacle. Some of these tiny fruits have been replaced by silica and hence show their general form with great fidelity. They are round and approximately symmetrical in cross section, about four times as long as their maximum diameter, narrowed like the neck of a carafe above the middle, and slightly expanded at their distal end. Their surface is smooth, and they show no structural details. Those that are partly eroded are seen to be hollow.

There can not be the slightest doubt that these objects are the same as those from the Green River formation that Newberry thought represented *Brasenia*, for in those of my specimens where the details are indifferently preserved the two can not be distinguished, and the supporting stems and branches are also identical. The better preserved of my specimens show that they can not represent the genus *Brasenia*, however, for in that genus the fruiting heads are made up of a much smaller number of unsymmetrical carpels. The only existing forms known to me which have fruit heads like these fossils are those of the genus *Sparganium*, and in this genus the individual fruits are also much like those of the fossil, consisting of numerous one-celled, nutlike fruits.

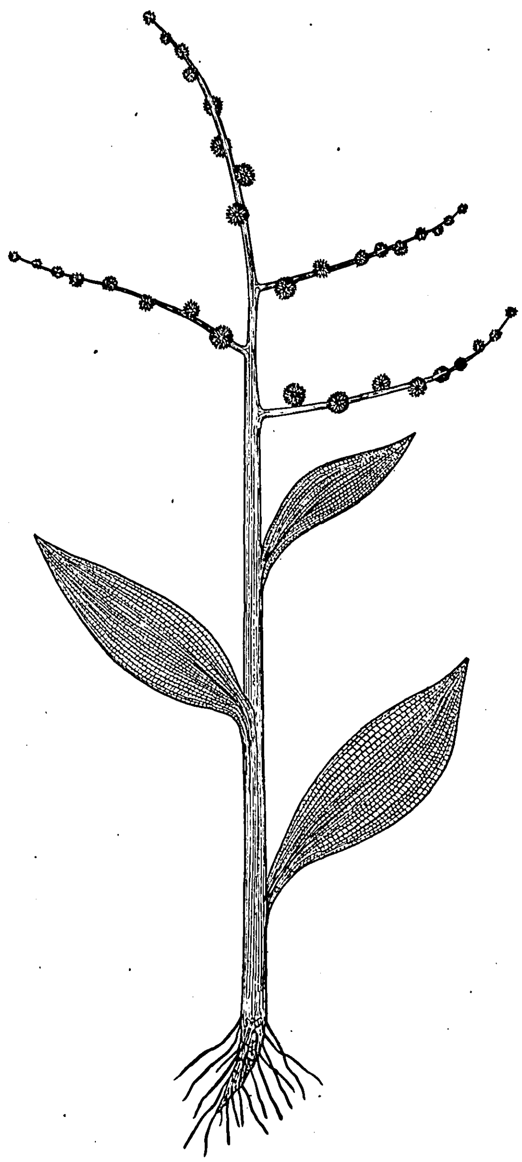


FIGURE 5.—Restoration of *Sparganium antiquum* (Newberry) Berry. One-third natural size

Associated with these *Sparganium*-like fruits in the Wind River Basin are leaves of *Pontederites hesperius* Knowlton. These are neither common nor well preserved, but they show the unmistakable parallel veining with the transverse nervilles of that form, and they are too wide and otherwise dissimilar to be confused with the associated fragments of palm foliage, nor are they at all similar to the associated fragments of the problematic monocotyledon that has been called *Musophyllum*.

In one specimen a considerable fragment of a parallel-veined stem, about 1.25 centimeters in diameter as flattened in the rock, bears a fragment of a leaf with parallel veins and transverse veinlets, exactly like *Pontederites hesperius*. This shows that the leaf was not petiolate nor auriculate at the base, as in *Pontederia*, but was cuneate-lanceolate, narrowing at the base and then expanding to form a decurrent sheath, and that these leaves were alternate in arrangement.

This leaf-bearing stem fragment is marked by parallel veins exactly like the somewhat smaller branched stems that bear the fruit heads and is identical with them in every respect, so that the conclusion is inevitable that both represent the same plant. This conclusion receives some corroboration, if such is deemed necessary, by their close association in the Wind River Basin and by the presence of identical detached leaf fragments and fruit heads in the Green River Basin, none of these classes of remains being found either together or single at any other horizons.

The family Pontederaceae has petiolate, auriculate leaves, and the fruit is a single-seeded utricle or capsule. Hence the fossil can not be referred to that family. The family Alismaceae, which has leaves with a somewhat similar venation, has these leaves petiolate and frequently auriculate, and the fruits are conspicuously different, being achenes with markedly curved seeds. The only family that appears to me to fulfill the requirements of the fossils, so far as these are known, is the Sparganiaceae.

It is true that the leaf-like spathes of the modern species of *Sparganium* have not been observed in the fossil material, probably a matter of lack of preservation of a minor detail; but in all other respects the resemblance is most exact as to stem, inflorescence, fruit heads, and individual fruits. If the fossil leaves were somewhat narrowed and elongated, a most easily accomplished modification, they would be exactly like those of the existing species of *Sparganium*.

The family Sparganiaceae comprises the single modern genus *Sparganium*, with 10 to 12 existing species, widely distributed in temperate and cool regions, growing in wet or submerged situations. They are chiefly confined to the Arctogaeic realm but occur also in Australia and New Zealand. The five or six North American species have a collective range from Newfoundland to British Columbia and southward to California, Louisiana, and Florida. The discontinuous distribution of the genus in the existing floras indicates a considerable antiquity, and this is fully borne out by the geologic record.

Fossil remains that have been referred to *Sparganium* are fairly numerous. Besides several still existing species that are represented in the Pleistocene deposits of both North America and Europe, about 15 extinct species that have been referred to this genus have been discovered.

The oldest of these are certain not very convincing fruit heads from the Upper Cretaceous Atane beds of western Greenland [32] and similar inconclusive remains from the Vermejo formation of Colorado.[33] The Eocene has furnished one or two species in addition to the one here described. These are recorded from the early Eocene Fort Union formation of Montana [34] and the late Eocene or possibly Oligocene of western Greenland,[35] Svalbard (Spitsbergen),[36] and Iceland.[37]

The genus made its appearance in Europe during Oligocene time, being reported to the extent of five species in beds of that age in France, Italy, Germany, and Alsace. Seven Miocene species have been described, from Switzerland, France, Styria, Croatia, Bohemia, Baden, and Transylvania. The Pliocene records comprise those of three species in Italy, France, Germany, and Holland.

Occurrence: Tipperary, Wyo.

Order ARECALES

Family ARECACEAE

Genus SABALITES Saporta

Sabalites florissanti (Lesquereux) Berry

Plate 9

Flabellaria florissanti Lesquereux, The Cretaceous and Tertiary floras, p. 144, pl. 24, figs. 1–2a, 1883.
Knowlton, U. S. Geol. Survey Prof. Paper 131, p. 154, 1923.

This species was described by Lesquereux from the Green River formation of Uinta County, Wyo., but as the generic term *Flabellaria* is preoccupied I have transferred it to the genus *Sabalites*. It is not at all uncommon in the Wind River Basin and represented a large fan palm not unlike the existing species of *Sabal* but not necessarily related to that genus. Knowlton records specimens of this species from the Green River formation which indicate leaves 5 feet in diameter.

[32] Heer, Oswald, Flora fossilis arctica, Band 3, p. 105, pl. 28, fig. 12, 1874.
[33] Knowlton, F. H., U. S. Geol. Survey Prof. Paper 101, p. 253, pl. 32, fig. 6, 1917 [1918].
[34] Ward, L. F., U. S. Geol. Survey Bull. 37, p. 18, pl. 3, figs. 6, 7, 1887.
[35] Heer, Oswald, op. cit., Band 1, p. 97, pl. 45, figs. 2, 13d, 1868; Band 2, p. 467, pl. 42, figs. 4, 5, 1871.
[36] Idem, Band 2, p. 51, pl. 7, fig. 3c, 1870.
[37] Idem, Band 1, pl. 145, p. 25, fig. 1, 1868.

Sabal-like palms are common in the Upper Cretaceous and Eocene of North America, ranging northward throughout our West to British Columbia. Needless to say, they denote a mild, humid environment.

Occurrence: Tipperary, Wyo.

Sabalites powellii (Newberry) Berry

Plate 10, Figures 6, 7

Sabal powellii Newberry, U. S. Nat. Mus. Proc., vol. 5, p. 504, 1883; U. S. Geol. Survey Mon. 35, p. 30, pl. 63, fig. 6; pl. 64, figs. 1, 1a, 1898.

Knowlton, U. S. Geol. Survey Prof. Paper 131, p. 153, 1923.

Several specimens of a large-leaved fan palm from the Wind River Basin are referred to *Sabalites florissanti* Lesquereux for the reason that they appear to lack the extended acumen of *Sabalites powellii* and have a relatively slender petiole, although it must always be remembered that most palm leaves lack good generic or specific characteristics. Both of these Green River species of *Sabalites* are superficially much like the leaves of the existing species of *Sabal*, and their northern range appears to afford some corroboration of such a relationship, but this is by no means established, and it appears to me to be better practice to refer such remains to the form genus *Sabalites* rather than to *Sabal.*

Both species are present in the Wind River Basin, but *Sabalites powellii* has stouter petioles and is readily distinguished from *Sabalites florissanti* by the extended acumen, as shown in the specimens here figured.

Sabalites powellii was a large-leaved fan palm, perhaps slightly smaller than the associated *Sabalites* and much like a variety of *Sabal*-like fragments of foliage described from various Eocene localities and horizons.

Occurrence: Tipperary, Wyo.

Genus GEONOMITES Visiani

Geonomites haydenii (Newberry) Knowlton

Manicaria haydenii Newberry, U. S. Nat. Mus. Proc., vol. 5, p. 504, 1883; U. S. Geol. Survey Mon. 35, p. 31, pl. 64, fig. 3, 1898.

Geonomites haydenii Knowlton, U. S. Geol. Survey Prof. Paper 131, p. 152, 1923.

This rather large-leaved feather palm was described from the Green River formation by Newberry, who referred it to the genus *Manicaria*, comparing it with the Swiss Miocene form *Manicaria formosa* Heer. He also compared it with the Raton formation species *Geonomites tenuirachis* Lesquereux.

Newberry's type, which is the only specimen extant, is a very incomplete fragment. A somewhat similar fragment of the same species is present in the collections from the Wind River Basin. I agree most heartily in Knowlton's course in transferring it from *Manicaria* to *Geonomites*. The former reference was most improbable on general grounds and inconclusive in specific features. The history of the forms referred to *Geonomites*, their resemblance to the existing species of *Geonoma*, and the distribution of *Geonoma* constitute conclusive proof, it seems to me, of the correctness of our inferences with regard to *Geonomites*.

As I have insisted in a previous publication,[38] the genus *Geonomites*, which derives its name from its resemblance to the existing genus *Geonoma* of Willdenow, is properly considered to represent the undifferentiated ancestry or the generically indistinguishable fossil representatives of the tribe Geonomeae. Of the 10 genera constituting this tribe 3, supposedly monotypic, are west African. The remaining 7 genera, including 98 per cent of the existing species, are widely distributed in America from the temperate uplands of Mexico and Central America across the Equatorial Zone. The only large genus of the tribe is *Geonoma*, which has nearly 100 existing species ranging from the Antilles to Rio de Janeiro and from southern Mexico through Central America and along the eastern slopes of the Andes to Bolivia. These palms are prevailingly small stemless or short-stemmed palms, of American origin, ascending into the altitudinal Temperate Zone, where, in Mexico and Central America, they are found in the oak forests of the uplands associated with derivatives of both the North Temperate and the tropical floras.

The genus *Geonomites* made its earliest appearance in the Upper Cretaceous Ripley formation of western Tennessee. It is found in the early Eocene in Colorado, New Mexico, and western Texas and in the middle Eocene of Texas (Mount Selman formation). In Europe it appears in the Lutetian of Italy and in Oligocene and lower Miocene deposits.

Occurrence: Tipperary, Wyo.

Palmocarpon lesquereuxi Berry, n. sp.

Amygdalus gracilis Lesquereux, U. S. Geol. and Geog. Survey Terr. Rept., vol. 8 (Cretaceous and Tertiary floras), p. 199, pl. 40, figs. 14, 15, 1883 (not figs. 12, 13, or pl. 44, fig. 6).

I have three specimens of the fruits which Lesquereux associated with the leaves from the Green River formation and described as *Amygdalus gracilis*. As Knowlton [39] has already pointed out, there was ab-

[38] Berry, E. W., U. S. Geol. Survey Prof. Paper 125, p. 6, 1919.

[39] Knowlton, F. H., U. S. Geol. Survey Prof. Paper 131, p. 164, 1923.

solutely no basis for associating these fruits with the leaves or for referring them to the genus *Amygdalus*.

The specimens from the Wind River Basin fully confirm Knowlton's opinion, which was based upon Lesquereux's figures. No affinity with *Amygdalus* is indicated. Their true botanic relationship is not so readily decided. Having in mind the genera represented by leaves in this flora, a not altogether safe procedure, I conclude that these fruits represent the dry-fleshed, berry-like fruits of one of the associated palms. Except for their larger size they are very similar to the fruits of the existing *Sabal*, although they have, of course, been greatly compressed during fossilization. As preserved they range from slightly under 1 to slightly over 2 centimeters in maximum diameter and from 8 millimeters to 1.5 centimeters in minimum diameter. In form they appear to have been prolate spheroidal to depressed spheroidal and slightly lobate, as in certain existing species of *Sabal*. The hilum is basal or subbasal and, although not large, appears to be so by the depression and wrinkling of the adjacent area and the development of what seems to be a short rhaphe. Longitudinal ridges radiate from this region, and part of the area of one of the specimens is thrown into a network of shrinkage folds exactly simulating the dried fruits of a modern *Sabal*. In addition to the longitudinal ridges corrosion of part of the surface shows a fine reticulate venation.

Occurrence: Tipperary, Wyo.

ORDER UNKNOWN

Genus MUSOPHYLLUM Goeppert

Musophyllum complicatum Lesquereux

Musophyllum complicatum Lesquereux, The Tertiary flora, p. 96, pl. 15, figs. 1–6, 1878.

Knowlton, U. S. Geol. Survey Mon. 32, pt. 2, p. 686, pl. 83, fig. 1, 1899; U. S. Geol. Survey Prof. Paper 131, p. 155, 1923.

This species is very abundant in the Wind River Basin but invariably in the same fragmentary condition as Lesquereux's type material from the Green River formation. It has also been recorded from Yellowstone Park, from beds which Knowlton called Miocene but which appear to be older.

The name *Musophyllum*, given originally to Old World forms in all probability related to *Musa*, is misleading. There is no evidence that these American Eocene forms are related to the Musaceae or that *Musa* or its ancestors were ever present in the Western Hemisphere. The fossil evidently represents a large-leaved monocotyledon, possibly related to *Heliconia* or *Canna* and at any rate indicative of mesophytic environment.

Occurrence: Tipperary, Wyo.

Monocotyledonous leaf

Iris? sp. [Lesquereux], U. S. Geol. and Geog. Survey Terr., Illustrations of Cretaceous and Tertiary plants, pl. 8, fig. 6, 1878.

Newberry, U. S. Geol. Survey Mon. 35, p. 33, pl. 22, fig. 6, 1898.

The only knowledge that we have of this form is the figure cited above, which appeared without record of horizon or locality. There is no means of checking the accuracy of this figure, nor does it offer any basis for a botanic determination.

In the collection from the Wind River Basin there are two specimens of a monocotyledonous leaf of considerable consistency, in which the size and venation agree with the figure cited and in which the degree of tapering is the same. This rather flimsy evidence suggests that the original specimen may have come from the Green River formation. On the other hand there is absolutely no basis for referring it to the genus *Iris*, although no one can deny that it might represent that genus. It seems to be too abruptly pointed to represent a ray of one of the numerous palms found at this horizon, but it might readily represent quite a variety of unrelated monocotyledonous genera.

Occurrence: Tipperary, Wyo.

Class DICOTYLEDONAE

Order JUGLANDALES

Family JUGLANDACEAE

Genus JUGLANS Linné

Juglans alkalina Lesquereux

Plate 10, Figures 1, 2

Juglans alkalina Lesquereux, The Tertiary flora, p. 288, pl. 62, figs. 6–9, 1878.

This characteristic form, which is not uncommon in the Wind River Basin was described by Lesquereux from the Green River formation of Alkali station, Wyoming, and is known only from these two occurrences.

Occurrence: Lenore and Tipperary, Wyo.

Juglans occidentalis Newberry

Juglans occidentalis Newberry, U. S. Geol. Survey Mon. 35, p. 34, pl. 65, fig. 1; pl. 66, figs. 2–4 (not fig. 1), 1898.

This species, which has been confused with *Juglans schimperi* Lesquereux, as Knowlton has shown in his recent revision of the Green River flora,[40] represents a fine large species of walnut. The genus is an old one, very common in the Eocene, and the present species, the type of which came from the Green River formation, is reported from the Raton formation of

[40] Knowlton, F. H., U. S. Geol. Survey Prof. Paper 131, pp. 159–160, 1923.

New Mexico and Colorado and from the Wilcox group of the Mississippi embayment. Leaflets identical with the type of this species are not uncommon in the Wind River Basin.

Occurrence: Tipperary, Wyo.

Order MYRICALES

Family MYRICACEAE

Genus MYRICA De Candolle

Myrica ludwigii Schimper?

Myrica ludwigii Schimper, Paléontologie végétale, vol. 2, p. 545, 1872.

Lesquereux, The Tertiary flora, p. 133, pl. 65, fig. 9, 1878.

The status of this species is most unsatisfactory, although there is no doubt that it represents a *Myrica.* It was described by Lesquereux from the Green River formation, being based on the single fragment figured, which is no longer to be found in the National Museum collections. The Tipperary specimen is even more fragmentary but undoubtedly identical with that from the Green River.

The type was from the European Miocene, and there is not the slightest doubt that the American Eocene leaves represent a different species from the European Miocene form.

Occurrence: Tipperary, Wyo.

Order SALICALES?

Family SALICACEAE?

Genus SALIX Linné?

Salix sp. Knowlton

Plate 12, Figure 5

Salix sp. Knowlton, U. S. Geol. Survey Prof. Paper 131, p. 156, pl. 37, figs. 3–5, 1923.

This form was collected from the Green River formation about 40 miles southwest of Meeker, Colo. It was described by Knowlton as follows:

> The collection made by Winchester includes a number of small leaves that appear to belong to *Salix*, though the nervation is so obscure that this assignment is not certain. They are linear-lanceolate leaves, 3.5 to about 6 centimeters long and 6 to 10 millimeters wide, and have a petiole 5 or 6 millimeters long. They are narrowed to a wedge-shaped base and are rather obtuse at the apex. The margin is perfectly entire. The nervation, with the exception of a relatively very strong midrib, is obscure but appears to consist of numerous thin secondaries at an angle of 35° or 40° that are camptodrome and arch just inside the margins; none of the finer nervation is observable.
>
> Considering the uncertainty regarding these leaves it seems hardly worth while to attempt comparisons with either living or fossil species. Small, narrow, entire willow leaves are so nondescript that it is difficult to be sure of their subsequent recognition, and for this reason the present form has not been given a specific designation.

Whatever its botanic affinity may be, and I regard this as highly conjectural, it appears to be represented in the flora found in the Wind River Basin. There is no evidence that it is not a *Salix*, and it may well represent that genus. On the other hand, there is no very satisfactory reason for referring it to *Salix.*

Occurrence: Tipperary, Wyo.

Order FAGALES

Family FAGACEAE

Genus DRYOPHYLLUM Debey

Dryophyllum wyomingense Berry, n. sp.

Plate 10, Figures 3–5

Leaves of medium size, ovate, widest at or below the middle, narrowing rather rapidly to the acute tip. Base broadly cuneate, ultimately slightly decurrent. Margins with fairly regularly spaced, prominent serrate teeth, equal or greater in number than the secondaries. Texture subcoriaceous. Length about 9 centimeters; maximum width, at or slightly below the middle, about 4 centimeters; hence the proportions are relatively shorter and broader than the majority of the late Cretaceous and early Tertiary species of the genus, approaching those of *Quercus* and departing from the typical form of *Castanea.* Fragments of leaves nearly twice the dimensions given above are contained in the collection. The petiole is not preserved, but it was evidently stout. Midrib stout, curved, prominent on the under side of the leaf. Secondaries medium stout, not especially prominent, about 12 pairs, diverging from the midrib at wide angles of over 45° and approaching 70° in the lower part of the leaf, at slightly irregular intervals curving moderately upward, subparallel, terminating at the tips of the marginal teeth. Where there are more teeth than secondaries, oblique branches from the secondaries run to their tips. Tertiary venation fine, mostly obsolete; hence the peculiarities of the tertiary venation near the marginal teeth which serve to differentiate *Dryophyllum* from *Quercus* and especially from *Castanea* can not be determined with precision. In some specimens numerous thin oblique tertiaries can be made out, as shown in the illustrations.

This species appears to be distinct from previously described forms. It is somewhat like the Green River species described by Newberry [41] as *Quercus castanoides* but differs in its shorter, wider form and more conspicuous teeth. It is also somewhat suggestive of the Arctic Tertiary species which Heer [42] called *Quercus ravniana.*

[41] Newberry, J. S., U. S. Geol. Survey Mon. 35, p. 70, pl. 65, fig. 6, 1898.

[42] Heer, Oswald, Flora fossilis arctica, Band 7, p. 90, pl. 66, fig. 3; pl. 67, fig. 7, 1883.

Among recent forms it is very like some of the leaves of our Pacific coast species of *Pasania* or the *Quercus densiflora* of Hooker and Arnott and many subsequent botanists. The genus *Pasania* of existing floras preserves for us the foliar features of the extinct genus *Dryophyllum*. In the absence of fruits it seems to me undesirable to refer this fossil from the Wind River Basin to *Pasania*, and as there is an equal or even greater doubt of its representing a true *Quercus*, I have referred it to the ancestral genus of the family Fagaceae, *Dryophyllum*, which is a most abundant type in the late Upper Cretaceous and Eocene of both America and Europe. It is true that leaves of this type have frequently been referred directly to *Quercus*, but there is grave suspicion regarding the correctness of the identification of many of the earlier forms that have been attributed to that genus.

Pasania densiflora is found on the Pacific coast from Oregon to California, its range and habitat being practically coterminous with that of the redwood (*Sequoia sempervirens*). It is therefore a mesophytic type of deep soils, abundant ground water, and heavy fogs. Although it would be hazardous to draw an environmental parallel between this Eocene form and the modern species with which it has been compared, the fact that the rather numerous known species of *Dryophyllum* are nearly all coastal species in plant associations that indicate an abundant rainfall lends considerable probability to the conclusion that such was the habitat of this species.

Occurrence: Tipperary, Wyo.

Order URTICALES

Family MORACEAE

Genus FICUS Linné

Ficus ungeri Lesquereux

Plate 12, Figure 4; Plate 14, Figure 5

Ficus ungeri Lesquereux, U. S. Geol. and Geog. Survey Terr. Ann. Rept. for 1871, suppl., p. 7, 1872; The Tertiary flora, p. 195, pl. 30, fig. 3, 1878; The Cretaceous and Tertiary floras, p. 163, pl. 44, figs. 1–3, 1883.
?Penhallow, Report on the Tertiary plants of British Columbia, p. 56, 1908.
Knowlton, U. S. Geol. Survey Mon. 32, pt. 2, p. 713, pl. 91, fig. 3, 1899; U. S. Geol. Survey Prof. Paper 131, p. 162, 1923.
Berry, U. S. Geol. Survey Prof. Paper 92, p. 58, 1924.

Several fragments in the Wind River Basin collection are referred with some hesitation to this Green River species. They represent the distal portions of a leaf, and in the type material from the Green River this part is usually broken away. The type came from Green River station above the fish beds, and the species has also been doubtfully reported from beds of Fort Union age in the Yellowstone National Park. It was reported by Penhallow from the Oligocene of British Columbia, but neither the identification nor the age determination can be relied upon. It appears to be represented, however, in the middle and upper Eocene of the Mississippi embayment.

This species appears to me to have more characters allying it to *Ficus* than the following species; but here also the botanic relationship is not conclusively established.

Occurrence: Tipperary, Wyo.

Ficus wyomingiana Lesquereux

Plate 12, Figures 1–3

Ficus wyomingiana Lesquereux, U. S. Geol. and Geog. Survey Terr. Bull. 1, p. 387, 1875; The Tertiary flora, p. 205, pl. 34, fig. 3, 1878.
Knowlton, U. S. Geol. Survey Prof. Paper 131, p. 162, 1923.

This species, based upon very imperfect material, was described by Lesquereux from the Green River formation at Green River station, Wyoming; and, according to Knowlton, the figured specimen is the only one contained in the United States National Museum collections. I have six specimens from the Wind River Basin that appear to belong to this species. These range both larger and smaller than the type, from which they differ in having the lateral primaries disappearing below the tip—that is, they are not acrodrome but subacrodrome. The new material agrees with the type in the cuneate base, the subbasilar primaries, and the character of the veinules, even to having the outer branches of the lateral primaries more ascending on one side than on the other. In outline they range from narrowly lanceolate to oval-lanceolate, in length from 2.5 to 12 centimeters, and in maximum width from less than a centimeter to 4.75 centimeters.

They are similar to various other Eocene forms that have commonly been referred to the genus *Ficus*, but this generic determination is highly problematic. I can not see any true *Ficus* characters in this material and believe that is should be referred to the Lauraceae, being probably related to the genus *Oreodaphne*, but I prefer not to make any change until such time as it may seem to be conclusively justifiable. The narrower forms are not to be distinguished from the European Tertiary *Cinnamomum lanceolatum* (Unger) Heer, but both this species and, in fact, the genus *Cinnamomum* itself have been identified with an optimism scarcely warranted by the facts, and frequently without any rational basis.

Occurrence: Tipperary, Wyo.

Ficus mississippiensis (Lesquereux) Berry mutant pseudopopulus Lesquereux

Plate 8, Figure 7

This form was described by Lesquereux [43] in 1875 as an independent species. In 1922 I presented evidence [44] to show that it was one of a series of forms to which various specific names had been given by Lesquereux, Knowlton, and Berry and that no specific limits could be made with any precision among these related forms. Their considerable range, both geologic and geographic, raises some doubt as to their representing a single botanic species, but such is my opinion.

The material from the Wind River Basin discloses a leaf of medium size, ovate, with a somewhat extended acuminate tip and a broadly rounded, ultimately decurrent base. The entire margin becomes very slightly undulate distad. The primaries are three in number and basilar, and the secondaries are typical.

There is considerable doubt as to whether this type of leaf represents a true *Ficus*, but it is certainly congeneric with many forms that are commonly regarded as fossil species of this genus, and the question can not be conclusively settled in the present state of our knowledge.

The present form is not greatly different from the associated leaves that have been referred to *Ficus wyomingiana* Lesquereux but seems to fall outside the limits of variation of that species.

Occurrence: Tipperary, Wyo.

Order GERANIALES

Family RUTACEAE

Genus FAGARA Linné

Fagara wyomingensis Berry, n. sp.

Plate 11, Figure 4

The material which forms the basis for this species is scanty in amount. It is interpreted as representing leaflets of a compound leaf, although the material does not offer any confirmation of this assumption. It resembles the leaflets of various existing Rutaceae, particularly the genera *Zanthoxylum* Linné and *Fagara* Linné, and is referred to *Fagara* for the reason that it is believed that *Zanthoxylum* was not differentiated from *Fagara* until post-Eocene time. As regards the leaflets the two can not be certainly distinguished in the fossil state, and I have advocated the hypothesis that *Zanthoxylum* represented a temperate derivative of *Fagara*, from which it is distinguished by the loss of the floral calyx.

[43] Lesquereux, Leo, U. S. Geol. Survey Terr. Bull. 1, p. 387, 1875; The Tertiary flora, p. 204, pl. 34, figs. 1a, 2, 1878.

[44] Berry, E. W., U. S. Geol. Survey Prof. Paper 131, p. 9, pls. 6, 7, 8, 1922.

This species may be described as follows: Leaflets small, ovate-elliptical, widest slightly below the middle, with a broadly rounded base, somewhat narrowed to the obtuse tip. Margins entire. Texture coriaceous. Length about 2.5 centimeters; maximum width about 1.6 centimeters. Midrib stout, very prominent, and channeled. Secondaries stout, much less prominent, 6 or 7 pairs, increasing their spacing and curvature and diverging at a smaller angle progressively from the base to the apex, camptodrome. Tertiaries thin and immersed.

This species has the same facies as various species that I have considered to represent the genus *Fagara*. These comprise 3 species from the Wilcox Eocene, 2 in the Claiborne Eocene, 2 in the Jackson Eocene, 4 in the Oligocene, and 2 in the Miocene. There are in addition a large number of fossil forms from the Tertiary in both America and Europe that have been referred to the allied genus *Zanthoxylum*.

In recent floras *Zanthoxylum* is represented by about 10 species of shrubs and small trees found in Asia and in North America as far north as Canada. The genus *Fagara* contains between 100 and 150 species of shrubs and trees with a cosmopolitan distribution in tropical and subtropical countries and extending short distances into the warmer parts of the Temperate Zone.

Occurrence: Tipperary, Wyo.

Order SAPINDALES

Family SAPINDACEAE

Genus SAPINDUS Linné

Sapindus dentoni Lesquereux

Sapindus dentoni Lesquereux, The Tertiary flora, p. 265, pl. 64, figs. 2–4, 1878.

This species of soapberry was described from the Green River formation at the mouth of White River, Utah, by Lesquereux. It is not uncommon in the Wind River Basin. It is also present in the middle and upper Claiborne and Jackson deposits of the Mississippi embayment.

Occurence: Tipperary, Wyo.

Sapindus obtusifolius Lesquereux

Sapindus obtusifolius Lesquereux, The Tertiary flora, p. 266, pl. 49, figs 8–11, 1878.

This species was originally described from the Green River formation by Lesquereux and has subsequently been collected from several localities within that formation. It has been reported from the Florissant beds and the Fort Union formation, but both of these occurrences are regarded by Knowlton as not representing this species. It is sparingly represented in the Wind River Basin collections.

Occurrence: Lenore, Wyo.

Sapindus winchesteri Knowlton

Sapindus winchesteri Knowlton, U. S. Geol. Survey Prof. Paper 131, p. 167, pl. 38, fig. 1, 1923.

This large form, described recently by Knowlton from the Green River formation of Colorado, is represented by fragmentary specimens in the Wind River Basin collections, making three supposed species of soapberry at this horizon.

Occurrence: Tipperary, Wyo.

Family ACERACEAE

Genus NEGUNDO Moench

Negundo fremontensis Berry, n. sp.

Plate 11, Figures 1–3

No complete leaves of this species have been found, but the shape of the leaflets suggests a trifoliate or odd-pinnate habit like that of the recent American species.

Terminal leaflet trilobate, with conical lobes, the median much the largest, separated by rounded sinuses; irregularly dentate above, entire below. Tips of lobes acute. Base broadly rounded, slightly decurrent. Length about 8.5 centimeters; maximum width about 5.5 centimeters. Primaries stout, prominent; the laterals diverging from the midvein near its base at angles of about 45°. Secondaries craspedodrome in part and camptodrome in part.

Lateral leaflet inequilateral, with a prominent, bluntly pointed lobe with a rounded sinus about midway between the apex and the base. General form ovate, widest medially, with an acute tip and a markedly inequilateral base, rounded on one side and decurrent on the other. Margins with irregularly spaced teeth, outwardly pointed and separated by wide, shallow, nearly symmetrically rounded sinuses. Length about 6.25 centimeters; maximum width about 3.25 centimeters. Midrib medially stout and prominent, becoming attenuated distad. Secondaries relatively stout and prominent; 6 to 8 pairs diverge from the midrib at irregular intervals, curving slightly in ascending and terminating craspedodromely in the marginal teeth except for the one immediately above the secondary on one side of the leaflet, which runs to the camptodrome lateral lobe. One secondary sends off a branch distad which terminates in a subordinate marginal tooth. Tertiaries well marked, usually anastomosing internally. Areolation fine, of small polygonal, nearly isodiametric meshes.

What I have termed terminal and lateral leaflets are found in association but not in organic union. They agree in their general facies, texture, venation, marginal characters, and areolation; and there is slight doubt as to the correctness of this interpretation. I have combined them in the accompanying restoration (fig. 6), in which the only part that is conjectural is the length of the petiolules, which are missing in all the fossil specimens.

The terminal leaflet is much like some of the forms that have been referred to *Acer aequidentatum* by Lesquereux, which came from the auriferous gravel of California. It is quite possible that some of these California specimens may represent the present species of *Negundo*, which is more like the existing California box elder than it is like our existing eastern box elder.

This characteristic form is unlike previously described forms, of which several are known. There are two Upper Cretaceous species in Colorado and Nebraska, and two early Eocene species in Colorado

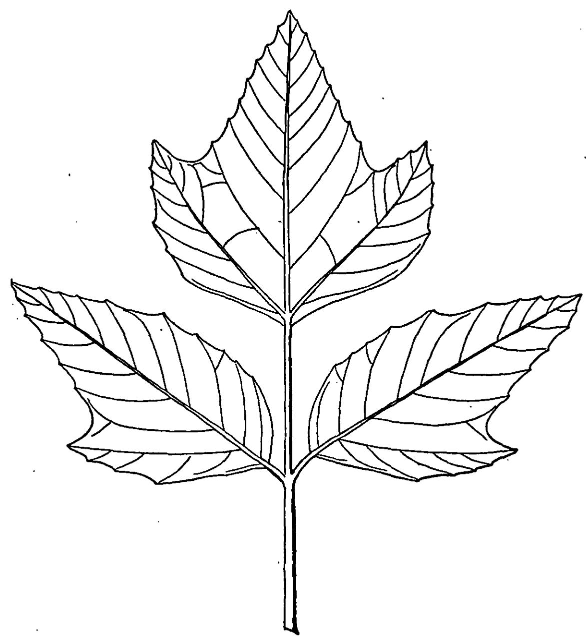


FIGURE 6.—Restoration of a complete leaf of *Negundo fremontensis* Berry

and North Dakota. The geologic history from that time to the present has heretofore been entirely unknown. The existing species are three in number. Our eastern box elder, or ash-leaved maple, ranges from central New York and western Vermont to Florida and is an inhabitant of stream and lake banks and swamp borders. It extends its range westward to New Mexico, Utah, and Arizona under somewhat drier conditions but is found only where there is a plentiful supply of ground water. There is a second existing species in Central America, and the third existing species, the California box elder, frequents stream borders and moist bottoms of the lower Sacramento and the western slopes of the San Bernardino Mountains.

Occurrence: Tipperary, Wyo.

Order RHAMNALES

Family RHAMNACEAE

Genus ZIZYPHUS Linné

Zizyphus wyomingianus Berry, n. sp.

Plate 11, Figures 6, 7

Leaves of medium size, ovate, widest below the middle, tapering upward to the acute but not produced tip. Base broadly rounded. Margins regularly finely dentate. Lenth about 8.5 centimeters; maximum width about 4 centimeters. Midrib stout and prominent, becoming thin distad. Lateral primaries nearly as stout as the midrib, from whose base they diverge at acute angles, curving upward subparallel with the margins and joining branches from the lowermost secondaries in the upper third of the leaf. Secondaries thin, two or three ascendingly curved camptodrome pairs. Tertiaries thin, transverse in direction, the marginal ones camptodrome, the internal ones percurrent or inosculating. Areolation obsolete.

This well-marked species appears to be distinct from the numerous fossil species that have been described, most resembling an undescribed species from the Eocene of the Mississippi embayment. It is also not greatly unlike *Zizyphus hyperboreus* Heer,[45] of the upper Eocene of western Greenland, which was recorded by Lesquereux from Carbon, Wyo. There are two species of *Zizyphus* known from the Green River formation, but neither of these is at all similar to this species from the Wind River Basin.

The genus *Zizyphus* contains more than 50 existing species, largely confined to the Indo-Malayan Tropics but sparingly represented in the American Tropics and naturalized to some extent in our Gulf tier of States. Its geologic history goes back to the Upper Cretaceous, there being over 50 known fossil species. The genus is particularly well represented in the Eocene of both America and Europe.

Occurrence: Tipperary, Wyo.

Family VITACEAE

Genus AMPELOPSIS Michaux

Ampelopsis tertiaria Lesquereux

Plate 11, Figure 15; Plate 14, Figure 4

Ampelopsis tertiaria Lesquereux, The Tertiary flora, p. 242, pl. 43, fig. 1, 1878.

Parthenocissus tertiaria Knowlton, U. S. Geol. Survey Prof. Paper 131, p. 170, 1923.

This striking species, first mentioned by Lesquereux in the annual report of the Hayden Survey for 1871, came from Green River, Wyo., from a horizon above the fish beds. It is represented by a somewhat larger specimen in the Wind River Basin, which obviously represents the same botanic species.

If there are any features by which fossil leaves of this type can be referred to the genus *Parthenocissus* of Planchon (1887) rather than to *Ampelopsis*, I have been unable to discover them, and I therefore can see no advantage in following Knowlton in transferring this species to *Parthenocissus*, even though it does resemble the existing *Ampelopsis quinquefolia*, which systematists now refer to *Parthenocissus*.

The present species was rather incompletely characterized by Lesquereux, and it may be redescribed as follows:

Leaves palmately compound, consisting of five leaflets from the top of a stout petiole. Leaflets broadly lanceolate, widest medially, tapering upward to the acute or acuminate tip and downward to the acuminately decurrent base. The texture is subcoriaceous. The margins, except near the base, are acutely serrate. Length 5.5 to 11 centimeters; maximum width 1.25 to 3.25 centimeters, the lateral leaflets being sometimes considerably smaller than the three central leaflets. Leaflets more or less petiolulate. Midrib stout, prominent, usually slightly curved and striate. Secondaries thin, numerous, regularly spaced, and subparallel. They diverge from the midrib at angles of about 45°, curve uniformly upward, and are camptodrome, sending off tertiary branches to the marginal teeth. Tertiaries thin, numerous, transverse, and inosculating, forming a fine areolation.

As Lesquereux pointed out, this species is very similar to the leaves of the existing Virginia creeper, *Parthenocissus quinquefolia* (Linné) Planchon, which ranges from eastern Canada to Florida, Texas, and Cuba and extends westward to Dakota, Nebraska, and Colorado. *Parthenocissus* has about 10 additional existing species in temperate eastern Asia. It is not distinguishable from *Ampelopsis* by any foliar characters. The genus *Ampelopsis* has entire, pinnate, or palmately compound leaves and about 15 existing species of temperate Asia and warmer temperate eastern North America. About half a dozen fossil species have been referred to *Ampelopsis*. These come from the early Eocene of North America and the upper Oligocene and lower Miocene of Europe.

Occurrence: Tipperary, Wyo.

Order MALVALES

Family TILIACEAE

Genus GREWIOPSIS Saporta

Grewiopsis wyomingensis Berry, n. sp.

Plate 8, Figure 6; Plate 13, Figures 1, 2

Alnus inequilateralis Lesquereux, The Tertiary flora, pl. 62, figs. 1, 2, 4 (not fig. 3), 1878.

Planera variabilis Newberry, U. S. Geol. Survey Mon. 35, fig. 5 (not figs. 6, 7), 1898.

In a preliminary examination of the Wind River Basin plants the small leaf of this species figured was

[45] Heer, Oswald, Flora fossilis arctica, Band 1, p. 123, pl. 49, fig. 2, 1868.

labeled *Alnus inequilateralis* Lesquereux, with the notation "genus doubtful," and the fragment of the larger leaf figured was labeled "like the larger leaves of *Planera variabilis* Newberry, not *Planera*."

The history of the two names given in the above synonymy is complicated and has been given by Knowlton in his recent account of the Green River flora.[46] Suffice it to say that both were composite, being partly identical and partly representing unrelated forms. Parts of each represent a species which I believe should be referred to *Grewiopsis* and which may be characterized as follows: Leaves of variable size, shortly and broadly ovate, often somewhat inequilateral, with an acute apex and a broadly rounded, slightly decurrent or cordate base. The texture is somewhat coriaceous. The margins are variably crenate, prominently so in the larger leaves, less so in the smaller. Length 4 to 9 centimeters; maximum width, at or below the middle, 3 to 7 centimeters. The petiole is invariably missing in the material from the Wind River Basin. It is stout and preserved for about a centimeter in one of the Green River specimens. The midvein is stout, prominent, and usually curved. The secondaries are mediumly stout, 4 or 5 opposite to alternate, unequally spaced pairs, diverging from the midrib at varying angles and prevailingly but not uniformly craspedodrome. The basal opposite pair diverge from the extreme base of the midvein, are more ascending and stouter than the others, and partake of the nature of lateral primaries; they give off 5 or 6 craspedodrome branches on the outside, the basal one of which may be enlarged. The upper secondaries may also give off one or two curved craspedodrome branches. The tertiaries are otherwise prevailingly percurrent.

This species lacks all the features of both *Alnus* and *Planera*. It is, however, congeneric with the numerous fossil leaves that have been referred to the genera *Grewia* or *Grewiopsis* and also greatly resembles the leaves of various existing species of the family Tiliaceae. *Grewia*, a relationship with which is implied in the name *Grewiopsis*, is an Oriental genus in the existing flora, with 75 to 100 species ranging from China and Japan to Africa and Australia. The fossil forms referred to *Grewia* or *Grewiopsis* comprise over 30 species, appearing in the Upper Cretaceous of North America and continuing in force through the Eocene, during which they appeared in the European record. They are especially common in the Fort Union formation.

Occurrence: Tipperary, Wyo.

[46] Knowlton, F. H., U. S. Geol. Survey Prof. Paper 131, p. 161, 1923.

Order LAURALES

Family LAURACEAE

Genus LAURUS of authors

Laurus fremontensis Berry, n. sp.

Plate 13, Figure 3

Leaves fairly large, obovate, widest medially, rounding upward to the abruptly obtusely pointed tip and downward to the acuminate decurrent base. Margins entire. Texture subcoriaceous. Length about 12 centimeters; maximum width about 5.5 centimeters. Petiole missing, presumably stout. Midrib stout and prominent. Secondaries stout, about 2 irregularly spaced pairs; they diverge from the midrib at various angles, wider above and narrower below, and are camptodrome. The basal pair are somewhat distinct from the others and subprimary in habit, giving off on the outside several camptodrome tertiaries, the lowermost on each side joining ascending subsecondaries from the basal region of the midrib. Tertiaries thin but well marked; the majority of the internal ones are obliquely percurrent; occasionally they anastomose, and this is especially noticeable in the region between the subprimaries and the superjacent secondaries. Along the margins the tertiaries form regular flat-arched loops.

This species is named from Fremont County. It is clearly lauraceous and somewhat suggestive of *Laurus grandis* Lesquereux,[47] with the same general form but somewhat smaller, differing especially in the venation of the basal part of the leaf.

Laurus grandis is recorded by Knowlton[48] as a Miocene form from the auriferous gravel of California and from the intermediate flora of Yellowstone National Park. There is good reason, as stated under the discussion of *Aralia browni*, for considering the California form as belonging to the Eocene Ione formation, and there is apparently some doubt regarding the age of the "intermediate flora" of Yellowstone Park.

The only lauraceous genus of which I have a large amount of recent comparative material is *Sassafras*. I have given this genus considerable study[49] and am impressed with the possibility that this form from the Wind River Basin represents an entire leaf of an Eocene species of *Sassafras*. The points of resemblance are the general form with the obtuse tip and decurrent base, which can be matched in leaves of the

[47] Lesquereux, Leo, The Cretaceous and Tertiary floras, p. 251, pl. 58, figs. 1, 3, 1883. Knowlton, F. H., U. S. Geol. Survey Mon. 32, pt. 2, p. 725, pl. 93, fig. 3; pl. 95, fig. 1, 1899.

[48] Knowlton, F. H., U. S. Geol. Survey Bull. 696, p. 345, 1919.

[49] Berry, E. W., Bot. Gaz., vol. 34, pp. 426–450, figs. 1–4, pl. 18, 1902.

existing American species; the subprimaries with outside camptodrome branches; and especially the venation in the basal part of the leaf. I regard this conclusion of relationship as very probable, but pending proof of it I prefer to refer the species to the noncommital form genus *Laurus*.

It has always seemed remarkable that *Sassafras* should be so abundant in our Cretaceous rocks and then almost unknown in the Tertiary of North America, where plants are found in abundance and of sorts with which the *Sassafras* would be expected to be associated. Our existing species ranges from Massachusetts to Iowa and Texas in mesophytic associations, and there are two additional species in southern China—another reason for expecting it in our western Eocene.

Occurrence: Tipperary, Wyo.

Order UMBELLALES

Family ARALIACEAE

Genus ARALIA of authors

Aralia browni Berry, n. sp.

Plate 13, Figure 5

Leaves medium large, palmately lobed, presumably with a stout petiole. Length about 15 centimeters; maximum width about 18 centimeters.

The leaf is divided about two-thirds of the distance to the base into 5 lanceolate lobes, which are closely similar in size and outline and separated by rounded open sinuses. The central lobe, which is only slightly if at all larger than the other lobes, is about 10 centimeters in length and 3 centimeters in maximum width. The attitude of the lower lateral lobes is about at right angles to the central lobe, and the upper lateral lobe is slightly above the halfway or 45° position. The leaf base is cuneate and slightly decurrent. The margins are entire near the base and in the lower part of the lobes; distad they are shallowly dentate. In some large specimens they are entire nearly throughout. The primary venation is stout; the secondary venation thin and inconspicuous. Midrib stout and straight. Two very stout lateral primaries diverge from the midrib just above its base at angles of about 30°; after an interval of about 1.5 to 2 centimeters these fork to form two approximately equal branches, each of which forms the midvein of its respective lateral lobe. The secondaries are thin, numerous, regular, subparallel, and apparently camptodrome, diverging from the primaries at angles greater than 45°. Much of the secondary venation has become obliterated during fossilization, and although these leaves do not merit the term coriaceous the leaf substance was of considerable consistency. In the specimen figured the right side is broken and somewhat overlapped on the median lobe.

The species, which is common in the Wind River Basin, is named for the collector.

The most closely related fossil form is *Aralia angustiloba* Lesquereux,[50] described originally from the auriferous gravel of Nevada and Placer Counties, Calif., and doubtfully recorded from the Green River formation of Wyoming and the supposed Eocene of southwestern Oregon. The present form differs from *A. angustiloba* in its somewhat more expanded lobes and partially toothed margins. Otherwise the two are identical in all characters and may well represent the same botanic species, as a single species frequently shows variations from entire to toothed margins and vice versa. However, as this is as yet conjectural, the form from the Wind River Basin is tentatively described as an independent species. It differs from the Green River *Aralia wyomingensis* Knowlton and Cockerell in its narrower lobes, much deeper sinuses, and much less prominent marginal teeth. It may with propriety be considered an Eocene derivative of the Upper Cretaceous *Aralia saportana* Lesquereux.

The generic term *Aralia* is used in a generalized sense and not in the Linnean sense, and the present as well as most of the lobed leaves referred to *Aralia* by paleobotanists probably represent the genus *Oreopanax*, which in the modern flora is a large genus confined to the Antilles, Mexico, Central America, and South America and divided into sections according to whether the leaves are simple, lobed, or digitate. *Aralia browni* would fall in the section Lobatae, which has more than a score of existing species in the Antilles and Central and South America.

Although Knowlton considered the auriferous-gravel flora Miocene, Smith states, in a recent publication,[51] that it is for the most part proved to be of Eocene age and referable to the Ione formation.

Occurrence: Tipperary, Wyo.

Aralia notata var. denticulata Berry, n. var.

Plate 15, Figure 5

Leaves fairly large, trilobate, with a large conical central lobe and short conical lateral lobes about midway between the apex and the base. Sinus open and rounded, and lateral lobes not produced. Tips bluntly pointed. Base markedly decurrent. Margins entire below, dentate above, a tooth at the end of each secondary, separated by curved, shallow, nearly symmetrical sinuses. Texture subcoriaceous. Length

[50] Lesquereux, Leo, Harvard Coll. Mus. Comp. Zoology Mem., vol. 6, No. 2, p. 22, pl. 5, figs. 4, 5, 1878.

[51] Smith, J. P., California Acad. Sci. Proc., 4th ser., vol. 9, No. 4, p. 164, 1919.

about 19 centimeters; maximum width about 14 centimeters. Midrib stout, prominent. Lateral primaries stout, prominent, opposite, diverging from the midrib a considerable distance above its base at angles of 45° or slightly more and running to the tips of the lateral lobes. Secondaries mediumly stout, numerous, at regular intervals, slightly curved and subparallel; those in the parts of the leaf where the margins are entire are camptodrome; those in the part where the margins are toothed are craspedodrome. Tertiaries thin, curved, percurrent or anastomosing. Areolation obsolete.

In size and appearance this variety is very much like *Aralia notata*. It differs from that species in its more decurrent base, suprabasilar lateral primaries, and dentate margins.

Aralia notata was described by Lesquereux [52] from the Fort Union formation and is widespread in beds of this age in the western United States and British Columbia. It has also been recorded from the Denver and Lance formations of the West and from the Wilcox group of the Mississippi embayment. Both the species and its later variety should probably be referred to the genus *Oreopanax*.

Occurrence: Tipperary, Wyo.

Aralia whitneyi Lesquereux

Aralia whitneyi Lesquereux, Harvard Coll. Mus. Comp. Zoology Mem., vol. 6, No. 2, p. 20, pl. 5, fig. 1, 1878; U. S. Nat. Mus. Proc., vol. 11, p. 16, 1888.

Knowlton, U. S. Geol. Survey Bull. 204, p. 82, 1902; U. S. Geol. Survey Mon. 32, pt. 2, p. 748, pl. 99, fig. 3, 1899; U. S. Geol. Survey Prof. Paper 73, p. 59, 1911.

Leaves of this fine species of *Aralia* are abundant in a tuff on Crow Heart Butte. These do not differ from the type, which came from the auriferous gravel of California, except that some of the midveins of the third lateral lobes extend to the base instead of constituting a branch of the lateral primary that forms the midvein in the second pair of lateral lobes. One specimen, however, conforms exactly to the type.

Although some of the remains from Crow Heart Butte are fragmentary, there can be no doubt that they represent this species, and they therefore constitute another element tending to prove the Eocene age of part of the auriferous gravel of California.

This species has been recorded from beds in southwestern Oregon referred to the Eocene, from beds in Yellowstone National Park which Knowlton considered Miocene but which appear to be older, and doubtfully from the Mascall formation of Oregon and the Fort Union formation of Wyoming.

Occurrence: Crow Heart Butte, Wyo.

[52] Lesquereux, Leo, The Tertiary flora, p. 237, pl. 39, figs. 2–4, 1878.

Order EBENALES

Family EBENACEAE

Genus DIOSPYROS Linné

Diospyros mira Berry, n. sp.

Plate 13, Figure 4; Plate 14, Figure 7

Leaves of medium size, ovate, widest medially and narrowing about equally to the acute apex and base. Margins entire, evenly rounded. Length about 19 centimeters; maximum width about 4.5 centimeters. Petiole stout, of unknown length. Midrib stout, prominent, curved. Secondaries about 6 alternate pairs, fairly equally spaced, diverging from the midrib at angles of about 45°, curving upward and camptodrome. Tertiaries mostly obsolete.

This species, which appears to be distinct from previously described forms, is referred to the genus *Diospyros* with some hesitation because of the dearth of material collected.

The genus is a large and for the most part tropical one in existing floras and has numerous extinct species carrying its ancestry back to the Upper Cretaceous, authenticated by numerous fossil examples of the concrescent calices and in at least one occurrence (Eocene) by the petrified fruits. The genus is not uncommon in both earlier and later Tertiary floras of the western United States but has not, to my knowledge, been recorded in the Green River Eocene. The species from the Wind River Basin is very similar to the widespread Upper Cretaceous species *Diospyros primaeva* Heer, and it is also much like the existing persimmon, *Diospyros virginiana* Linné, which ranges from southern New England to Florida and Texas and is a Temperate Zone extension of a prevailingly tropical genus of warm humid regions of generous rainfall.

Occurrence: Tipperary, Wyo.

POSITION UNCERTAIN

Genus NORDENSKIÖLDIA Heer

Nordenskiöldia borealis Heer

Nordenskiöldia borealis Heer, Flora fossilis arctica, Band 2, Abt. 3, p. 65, pl. 7, figs. 1–13, 1870; Band 6, Abt. 1, p. 13, pl. 6, fig. 8, 1880.

Newberry, U. S. Geol. Survey Mon. 35, p. 137, pl. 68, figs. 4–6, 1898.

Knowlton, U. S. Geol. Survey Prof. Paper 131, p. 176, 1923.

Pedunculate capsular fruits, with 10 to 12 small ligneous carpels arranged around a central axis.

These characteristic fruits were elaborately described by Heer from the Tertiary plant beds at Kings Bay, Svalbard (Spitsbergen), where they are exceedingly common. They were subsequently recorded from western Greenland by the same author. Lesquereux [53] recorded them from the Dakota sandstone, but his determination is incorrect. Dawson [54] recorded them from the Tertiary of the Mackenzie River in British Columbia, but there is some question also as to the correctness of this identification. Newberry's identification of them in the Green River formation of Wyoming has been reexamined by Knowlton, who states that "Whatever the nature of this organism may be, there seems no doubt that it is identical with that described and figured by Heer." To judge from Heer's numerous figures this seems to be so. There are several specimens of these fruits in the collections from the Wind River Basin, and these are certainly identical with those described by Newberry from the Green River formation. The material from the Wind River Basin is rather well preserved, and I believe that Heer's interpretation is essentially correct. These objects are certainly not to be compared with *Nyssa* or *Viburnum* or other stone fruits.

They were elaborately described by Heer, who compared them with the various fruits of the Tiliaceae and Cistaceae, as well as with those of the Malvaceae and also, less happily, with certain fossil forms such as *Apeibopsis* and *Cucumites*. Schenk concurred in the comparisons with the fruits of the families Tiliaceae and Malvaceae, which seem to me to be rather apt.

The most interesting feature of their presence in the Green River and Wind River Basin floras is their great northward range, which must be taken into account in a consideration of ecology and which is also important in arriving at the age of these Arctic beds.

Occurrence: Tipperary, Wyo.

Genus CARPITES Schimper

Carpites newberryanus Knowlton

Plate 14, Figures 2, 6

Juglans occidentalis Newberry, U. S. Geol. Survey Mon. 35, p. 34, pl. 66, figs. 4a–4c (not figs. 1–4), 1898.

Carpites newberryana Knowlton, U. S. Geol. Survey Prof. Paper 131, p. 174, 1923.

This fruit was described by Newberry from the Green River formation and included with certain associated leaves in the species *Juglans occidentalis*, although Newberry pointed out that they are more like *Hicoria* than they are like *Juglans*. Knowlton has redescribed the type material and concluded that the fruits were misinterpreted by Newberry and are neither *Juglans* nor *Hicoria*.

[53] Lesquereux, Leo, U. S. Geol. Survey Mon. 17, p. 219, pl. 44, fig. 6, 1892.

[54] Dawson, J. W., Roy. Soc. Canada Trans., vol. 7, sec. 4, p. 71, pl. 10, fig. 6, 1890.

Similar fruits are present in the Wind River Basin and are equally decisive in failing to show any resemblance to the fruits of the Juglandaceae. Their botanic relationship remains questionable, although they suggest to me comparisons with the genus *Luhea* Willdenow, of the family Tiliaceae.

Occurrence: Tipperary, Wyo.

Genus ANTHOLITHES Brongniart

Antholithes anceps Berry, n. sp.

Plate 15, Figure 1

These objects, which are not uncommon, are believed to represent the rather coriaceous petals of some polypetalous flower, although it is possible that they are scalelike in nature. They have a characteristic form and markings and should be readily recognizable. They are about 8 millimeters in length and are widest medially, where they are about 3.5 millimeters across. The apex is bluntly pointed, the lateral margins full and evenly rounded, the basal margins slightly incurved, and the proximal end narrowly rounded. Five or six subequal veins diverge at acute angles from the base but are soon lost in the somewhat corrugated surface, which may be the expression of a sort of areolation in a petal of considerable consistency. They are preserved in a longitudinally curved position and have a rounded median ridge, which is the result of lateral curvature but without any appreciable median thickening.

They have the appearance of petals rather than fruit scales or valves of some capsular fruit, but I can offer no surmise regarding their botanic position.

Occurrence: Tipperary, Wyo.

Antholithes browni Berry, n. sp.

Plate 15, Figure 3

This species is represented by a single specimen and its counterpart, the preservation of which is such that the essential features are obscured.

It is larger than the associated *Antholithes fremontensis*, being about 1.75 centimeters in diameter, and is sympetalous, with 5 conical petals (?) uniting to form a rotate corolla suggestive of the flowers of the Solanaceae. The central part contains the mashed remains of the essential organs, but no details can be made out.

This flower suggests comparisons with such genera as the existing *Symplocus* and *Witheringia* and the fossil genus *Solanites* of Saporta. *Solanites* is represented by species in the Wilcox group of the Mississippi embayment and the Oligocene of France which are both smaller than the present species. Among recent forms this is very similar to a flower of *Witheringia* sp., from Peru, figured by Saporta.

Its systematic position must remain uncertain until better preserved material comes to light. It is named in honor of the collector, N. H. Brown.

Occurrence: Tipperary, Wyo.

Antholithes fremontensis Berry, n. sp.

Plate 15, Figure 2

This species is represented by three specimens, in none of which is the preservation sufficiently good to permit the determination of the essential features. These flowers are fairly large, being about 1.2 centimeters in diameter, and appear to be polypetalous with 5 pointed sepals and 5 ovate petals. The central essential organs are present but mashed beyond recognition. The flower appears to have been of considerable consistency.

It seems to me to be quite useless to speculate upon the systematic position of these specimens—a question which must await the collection of more perfectly preserved material.

Occurrence: Tipperary, Wyo.

Sporophylls

Ophioglossum hastatiforme Cockerell, Torreya, vol. 24, p. 19, 1924.

Danaea coloradensis Berry, Torreya, vol. 24, p. 49, 1924 (not Knowlton).

Xantholithes hastatiformis Cockerell, Torreya, vol. 26, p. 11, 1926.

Hastate sporophylls, in what has been subsequently learned to be an incomplete state of preservation, are not uncommon in the Wind River Basin. Specimens of these having inadvertently come into the hands of Cockerell, they were promptly described as a new species of *Ophioglossum*. My own material was taken to Washington, where Knowlton and I compared it with the type of *Danaea coloradensis* Knowlton from the Green River formation, a single specimen of a detached pinna, agreeing that the Wind River Basin material was the same. After I had announced this conclusion Knowlton recalled some curious plant fossils from Montana which had lain in the collections for years and which Ward had mentioned at an early meeting of the American Association and had later described in an imperfect sort of a way under the name *Xantholithes propheticus*.[55] This geologically older material is more complete than that from the Wind River Basin and shows that the latter could not represent a *Danaea*. I then turned my material over to Knowlton for study in conjunction with the more complete material from Montana. This has been submitted to several specialists. Maxon is positive that it bears no relationship to *Ophioglossum* or *Danaea*, and Howe is inclined to see algal relationships.

On the basis of Howe's opinion Cockerell has proposed to refer the Wyoming material to the pseudogenus *Xantholithes*, for which he would erect an extinct family, presumably belonging to the red algae, to be known as the Xantholithaceae. It would be interesting to have a convincing explanation of the occurrence of such a sturdy and well-marked type of a typically marine alliance in the continental deposits of Montana and Wyoming, thousands of miles from contemporaneous marine conditions.

My impression is that these sporophylls might represent a relict form of some member of the Williamsoniales, inasmuch as somewhat similar remains occur in the Dakota sandstone of Kansas and in the Magothy formation of Maryland,[56] but as the whole question has been transferred to other hands, no attempt will be made to describe the material from Wyoming in the present paper.[57]

Occurrence: Tipperary, Wyo.

Carpolithus bridgerensis Berry, n. sp.

Plate 15, Figure 4

A small pedunculate fruit of unknown botanic affinity. The peduncle is stout and striated (possibly owing to shrinkage), about 8 millimeters long, expanded distad into a tiny, persistent, coriaceous, 5-pointed calyx and surmounted by a subspherical fruit, which might be an indehiscent capsule, a more or less ligneous drupe, or a several-seeded berry with a thin flesh. As it shows some evidence of deformation and indications of more than a single hard seed I incline to the last interpretation. It may possibly represent the genus *Sassafras*, but no satisfactory conclusion can be reached.

Occurrence: Tipperary, Wyo.

Carpolithus browni Berry, n. sp.

Plate 14, Figure 1

Seed or stone, probably a stone; oval, compressed, with a marginal border. Surface irregularly reticulate depressed, perhaps owing to shrinkage. Length

[55] Ward, L. F., Glimpses of the Cosmos, vol. 4, p. 150, 1915.

[56] Berry, E. W., Maryland Geol. Survey, Upper Cretaceous, p. 769, pl. 51, figs. 5, 6, 1916.

[57] Since the foregoing statement was written Knowlton has passed away, and the Fort Union specimens referred to above, in so far as they could be found, have been placed in my hands. The question of the botanic affinity of the Wyoming material will be determined by the decision reached with respect to that from the Fort Union, which is so much more complete, and this I hope to discuss in a separate publication.

about 7.5 millimeters; maximum width about 6 millimeters. More broadly rounded at one end than at the other.

The botanic affinity of these objects is entirely conjectural. They greatly resemble the stones of some species of *Prunus*, as well as the seeds of the berry-like fruits of *Actinidia*, an east Asian genus of Dilleniaceae, and the unopened capsules of *Michelia* (Magnoliaceae). The first of these suggestions seems to me to be the most probable but lacks real value.

Occurrence: Tipperary, Wyo.

Sambucus? winchesteri Knowlton?

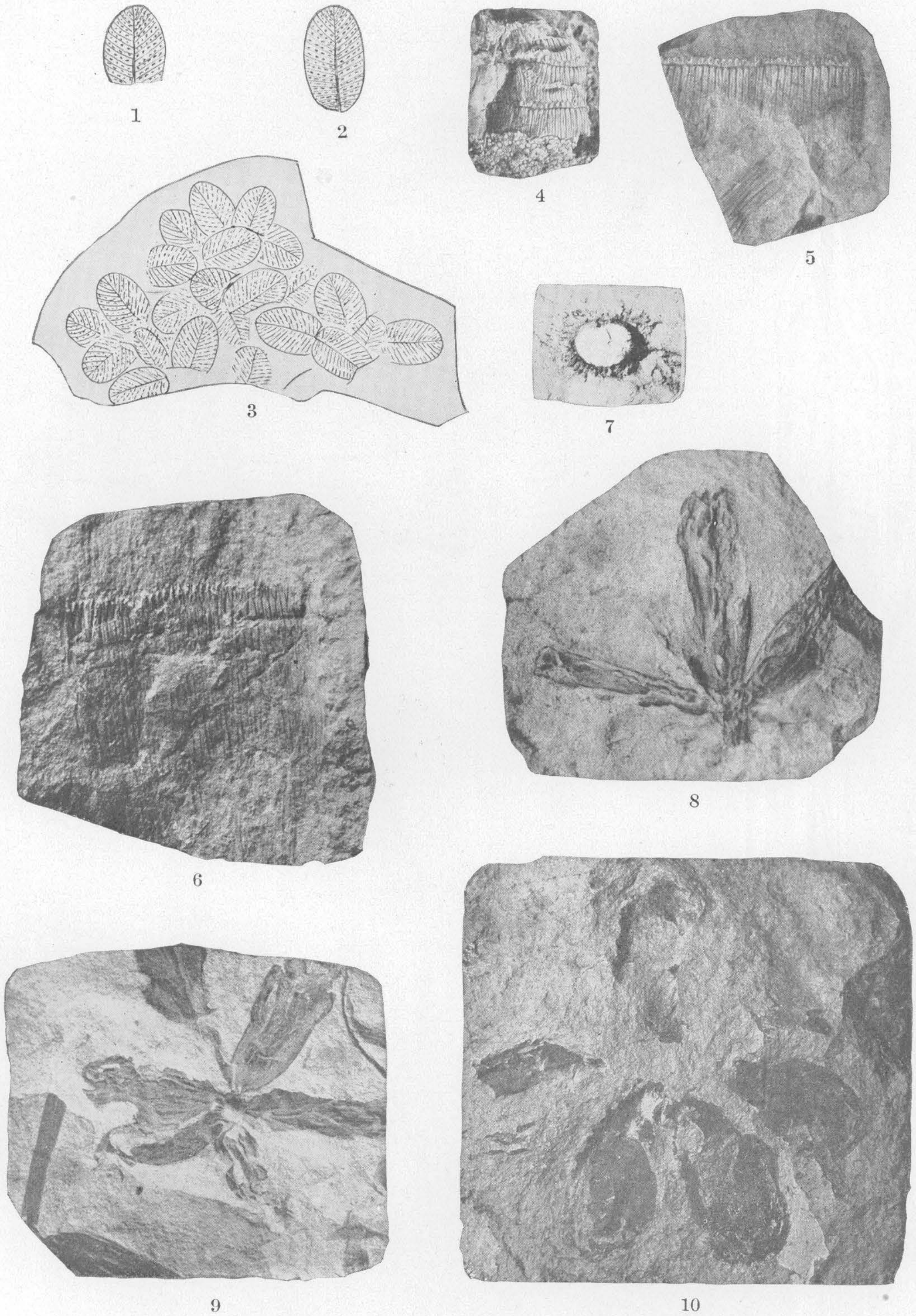
Plate 14, Figure 3

Sambucus? winchesteri Knowlton, U. S. Geol. Survey Prof. Paper 131, p. 173, pl. 40, fig. 8, 1923.

The specimen figured on the accompanying plate may represent what Knowlton tentatively referred to *Sambucus*. The present material is less well preserved that the type from the Green River formation of Colorado. It is also slightly larger, and the petals appear slightly narrower, although all are restored in the type figure, and their real width is conjectural. The petals appear to be entirely free in the specimen from the Wind River Basin, and they may be in the type, which would preclude its reference to *Sambucus*. Without suggesting any alternative relationship I have placed it under the plants of uncertain affinities, rather than in its natural place under the Gamopetalae.

A slab in the collection sent to me by Mr. Brown from an unknown locality in Wyoming contains *Lygodium kaulfussi* Heer, a leaf of *Zizyphus wyomingianus* Berry, and a flower of *Sambucus? winchesteri* Knowlton?, a Green River species, so that it presumably came from the same general horizon as the rest of the flora found in the Wind River Basin, and this is confirmed by the lithology.

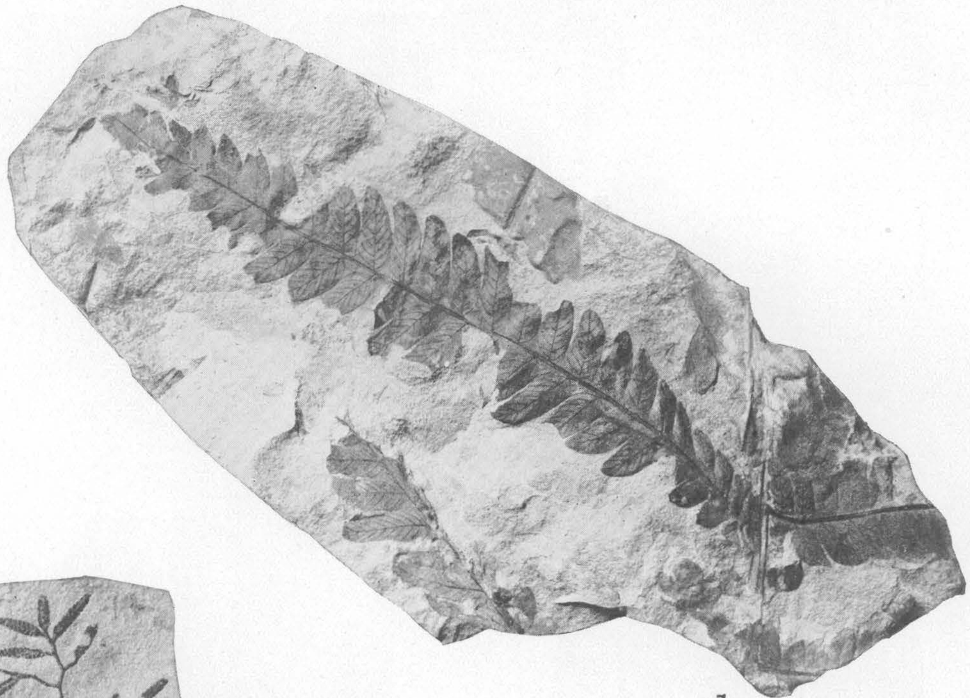
PLATES 6–15



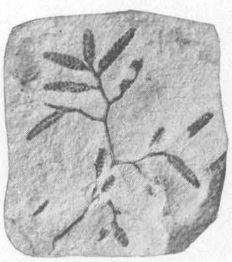
FOSSIL PLANTS FROM THE WIND RIVER BASIN

1–3. *Salvinia preauriculata* Berry. **1** and **2,** individual leaves showing the variations in form and size; **3** shows the abundance of the floating leaves in the shales.

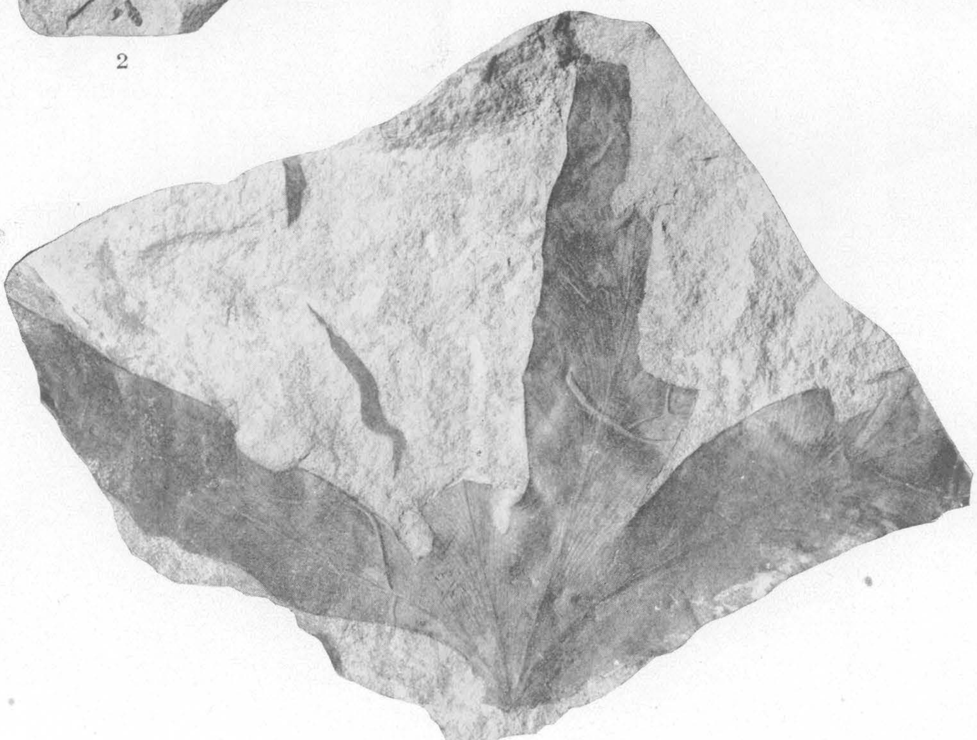
4–10. *Equisetum tipperarense* Berry, n. sp.: **4,** Fragment of a small stem with unusually short internodes; **5,** fragment of a larger stem showing branch scars; **6,** fragment of a still larger stem; **7,** detached leaf (nodal) sheath of a small stem, viewed from above; **8–10,** whorls of large tubers which are attributed to this species.



1



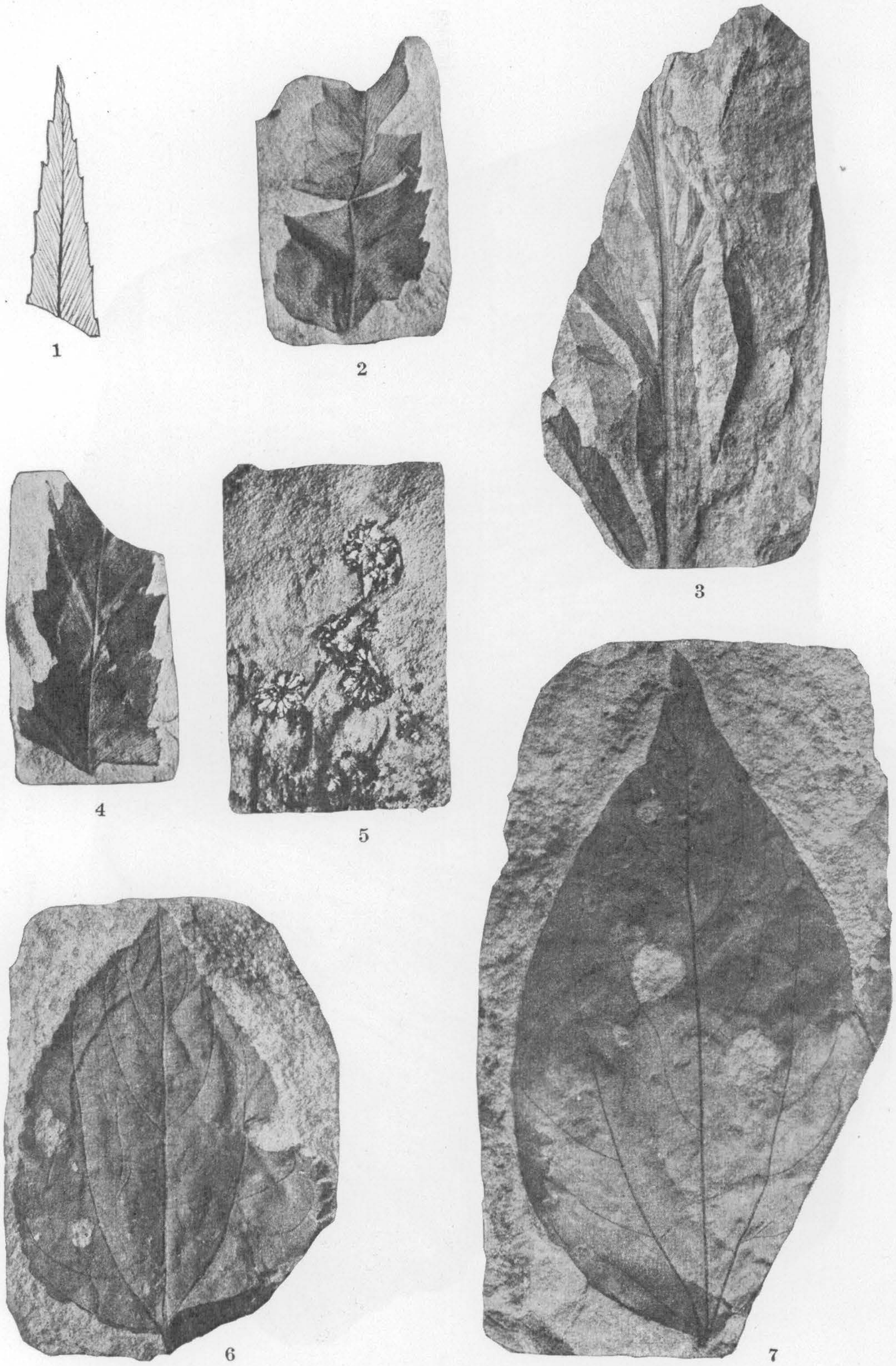
2



3

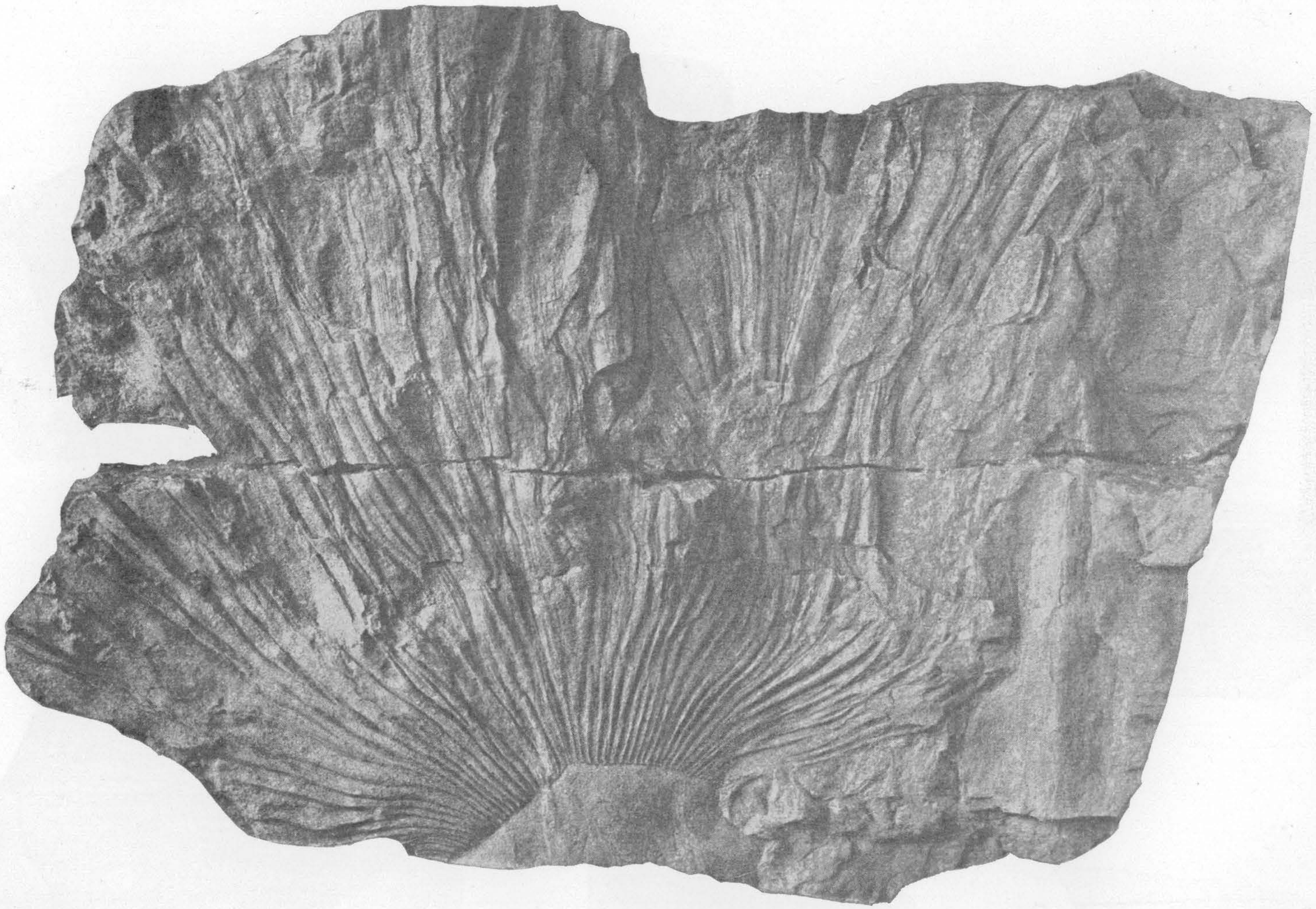
FOSSIL PLANTS FROM THE WIND RIVER BASIN

1. *Dryopteris weedii* Knowlton. Fragment of a frond from Tipperary.
2, 3. *Lygodium kaulfussii* Heer; 2, Fertile pinnae; 3, sterile pinna.



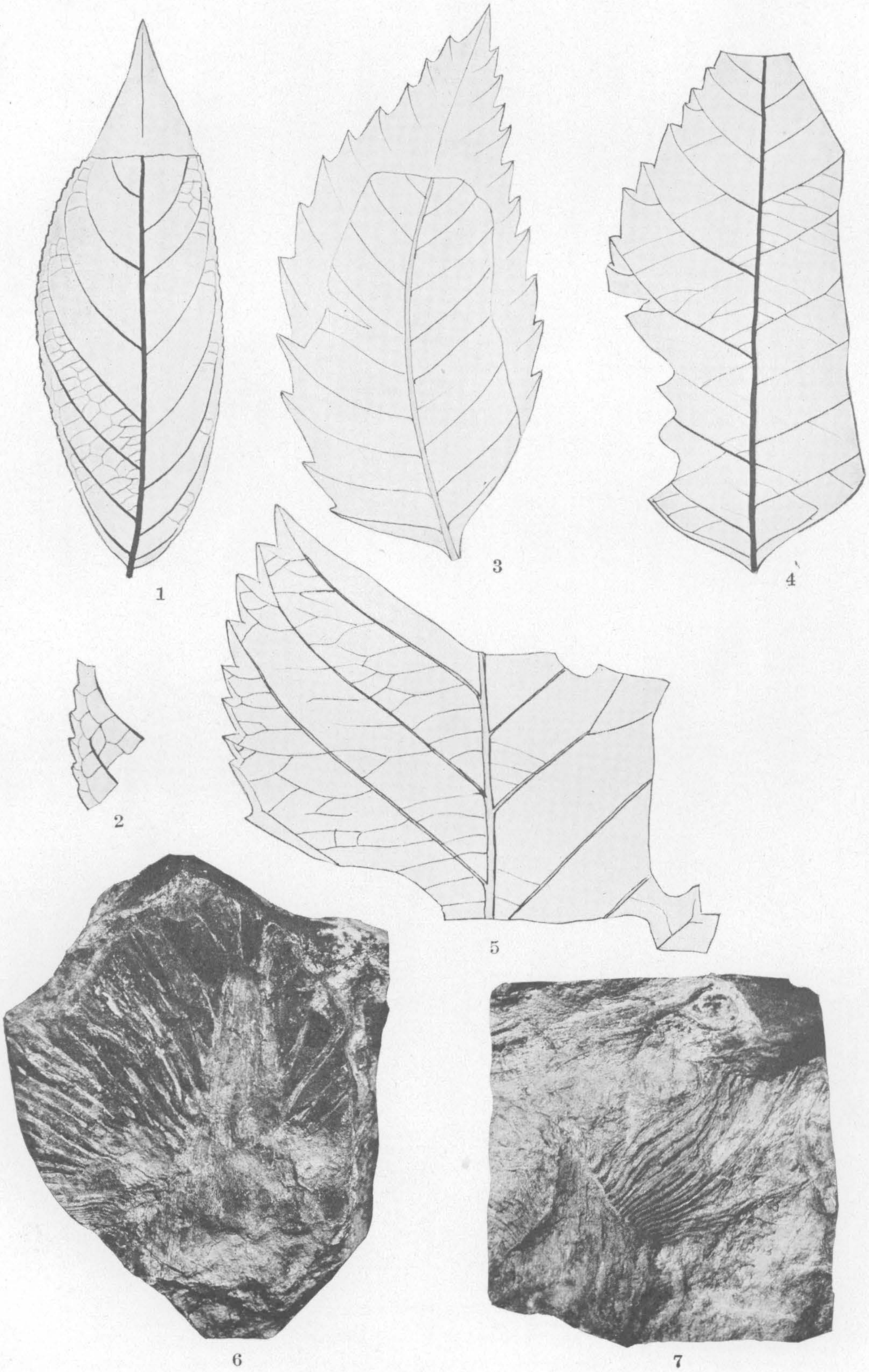
FOSSIL PLANTS FROM THE WIND RIVER BASIN

1. *Asplenium serraforme* Berry, n. sp.
2–4. *Asplenium eoligniticum* Berry: **2, 4,** Fragments from the proximal part of the frond; **3,** part of the distal part of a frond.
5. *Sparganium antiquum* (Newberry) Berry: Fruit heads, somewhat enlarged.
6. *Grewiopsis wyomingensis* Berry, n. sp.
7. *Ficus mississippiensis* (Lesquereux) Berry, mutant *pseudopopulus* Lesquereux.



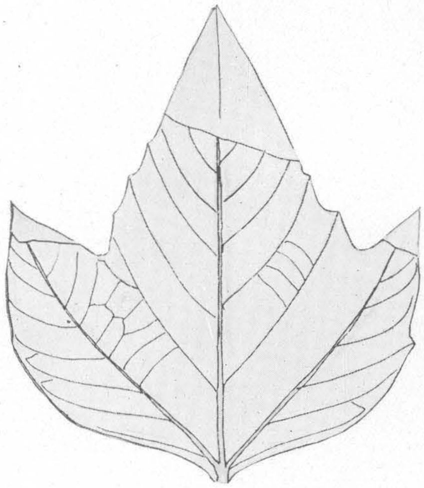
FOSSIL PLANTS FROM THE WIND RIVER BASIN

Sabalites florissanti (Lesquereux).

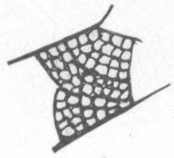


FOSSIL PLANTS FROM THE WIND RIVER BASIN

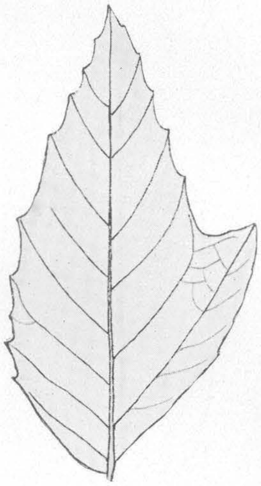
1, 2. *Juglans alkalina* Lesquereux: **1,** A leaflet; **2,** portion enlarged to show marginal venation.
3–5. *Dryophyllum wyomingense* Berry, n. sp. Showing the variations in the size and form of the leaves.
6, 7. *Sabalites powellii* (Newberry) Berry. Basal fragments of leaves showing the acumen at the top of the petiole.



1



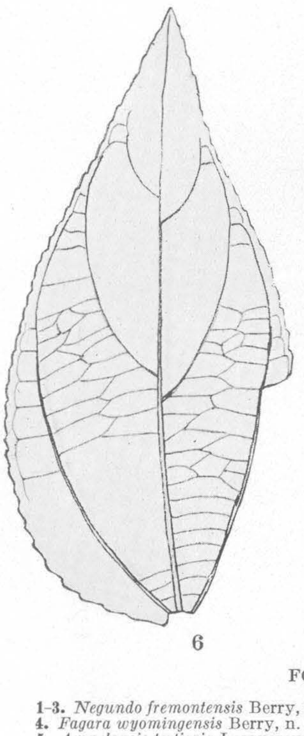
2



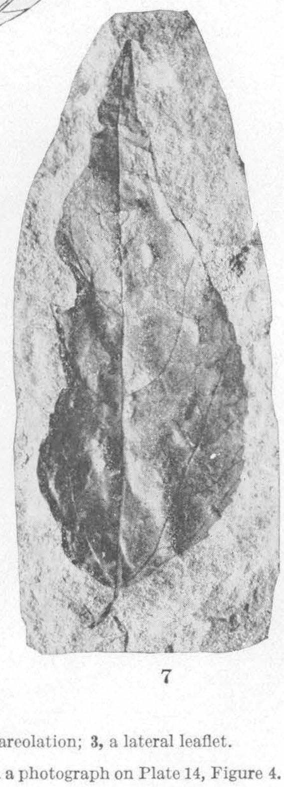
3

4

5



6



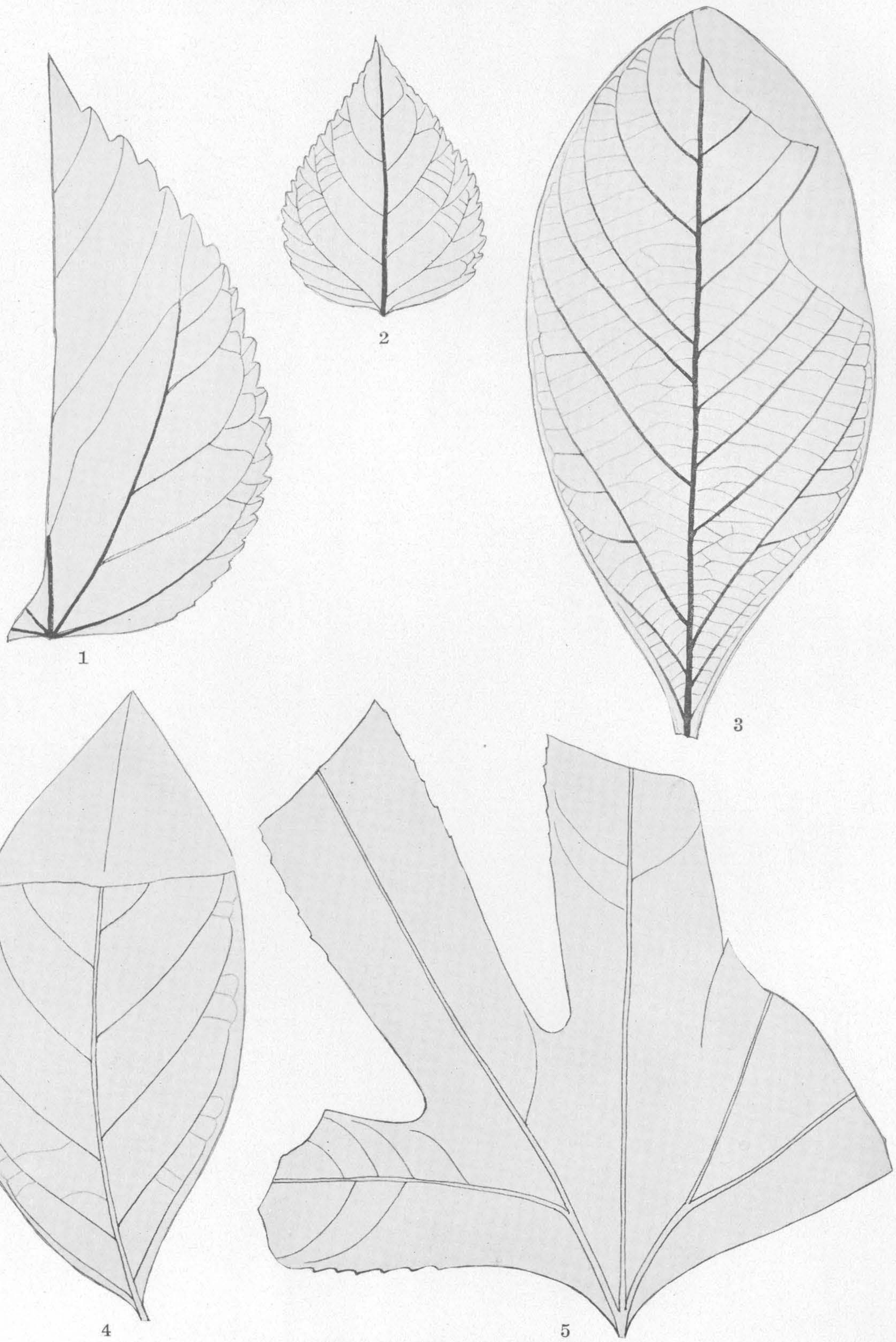
7

FOSSIL PLANTS FROM THE WIND RIVER BASIN

1-3. *Negundo fremontensis* Berry, n. sp.: **1,** A terminal leaflet; **2,** fragment enlarged to show areolation; **3,** a lateral leaflet.
4. *Fagara wyomingensis* Berry, n. sp.
5. *Ampelopsis tertiaria* Lesquereux. Drawing to show basal parts of the leaflets figured from a photograph on Plate 14, Figure 4.
6, 7. *Zizyphus wyomingianus* Berry, n. sp.

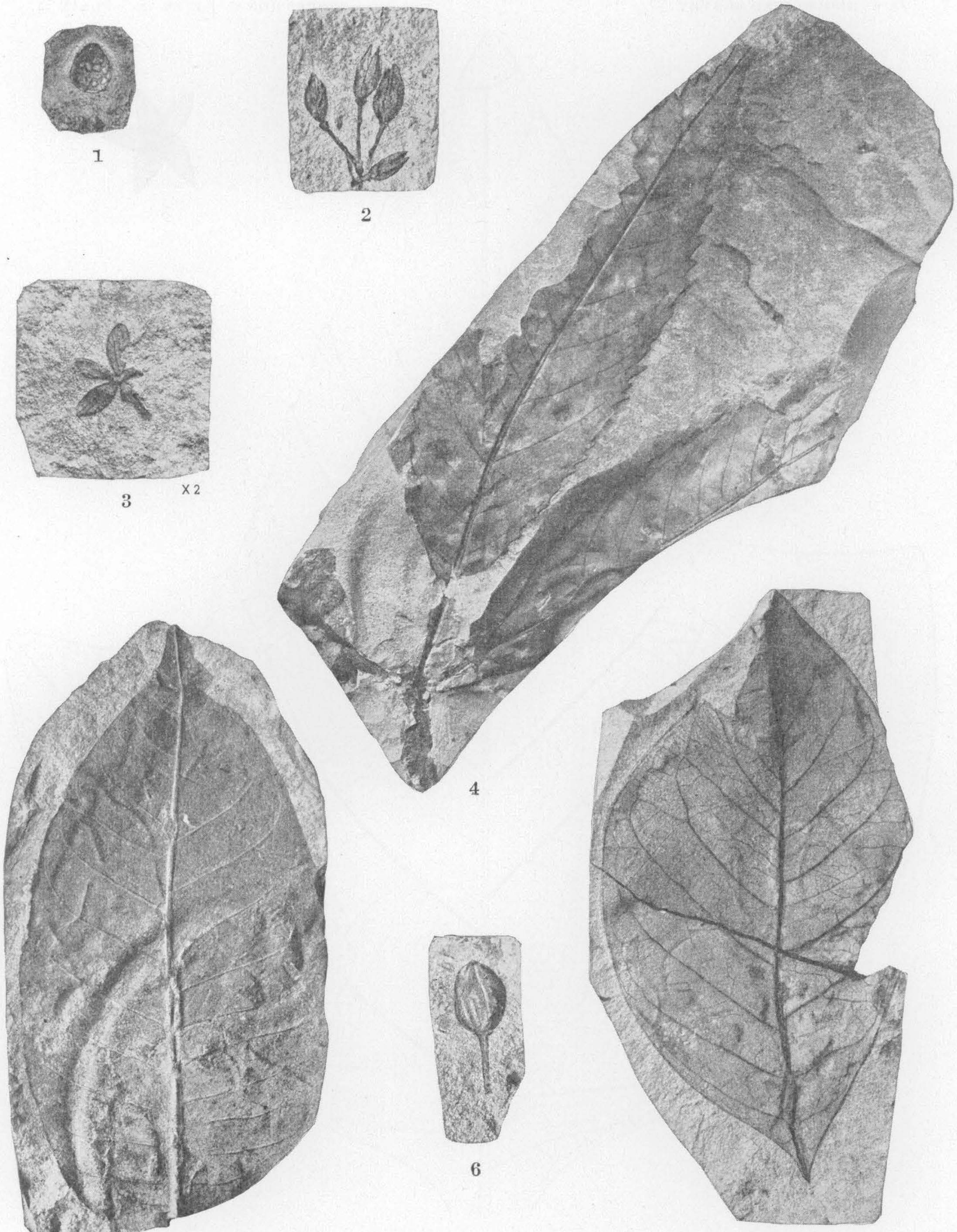
FOSSIL PLANTS FROM THE WIND RIVER BASIN

1-3. *Ficus wyomingiana* Lesquereux: **1,** A complete tiny leaf; **2,** a slightly larger leaf; **3,** maximum size seen.
4. *Ficus ungeri* Lesquereux.
5. *Salix* sp. Knowlton.
6. *Aralia notata* Lesquereux: A small leaf.



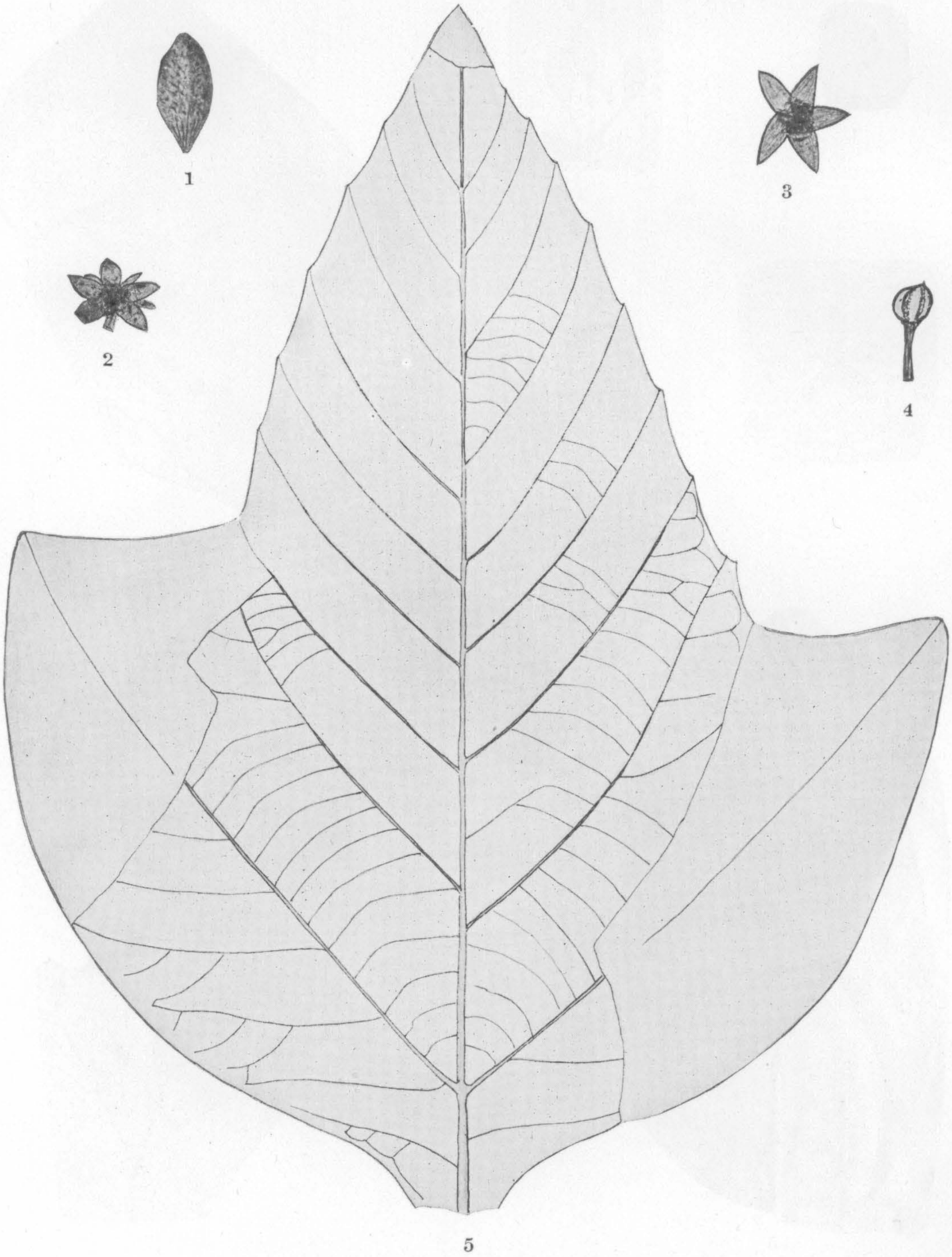
FOSSIL PLANTS FROM THE WIND RIVER BASIN

1, 2. *Grewiopsis wyomingensis* Berry, n. sp.: **1,** Large leaf of the type called by Newberry *Planera variabilis;* **2,** small leaf of the type called by Lesquereux *Alnus inequilateralis.*
3. *Laurus fremontensis* Berry, n. sp. A lauraceous form suggesting *Sassafras.*
4. *Diospyros mira* Berry, n. sp.
5. *Aralia browni* Berry, n. sp.



FOSSIL PLANTS FROM THE WIND RIVER BASIN

1. *Carpolithus browni* Berry, n. sp.
2, 6. *Carpites newberryanus* Knowlton.
3. *Sambucus? winchesteri* Knowlton (?).
4. *Ampelopsis tertiaria* Lesquereux. From photograph of specimen shown in drawing on Plate 11, Figure 5.
5. *Ficus ungeri* Lesquereux.
7. *Diospyros mira* Berry, n. sp.



FOSSIL PLANTS FROM THE WIND RIVER BASIN

1. *Antholithes anceps* Berry, n. sp.: A supposed flower petal.
2. *Antholithes fremontensis* Berry, n. sp.: A polypetalous (?) flower.
3. *Antholithes browni* Berry, n. sp.: A gamopetalous flower.
4. *Carpolithus bridgerensis* Berry, n. sp.
5. *Aralia notata denticulata* Berry, n. var.

GEOLOGY OF THE EASTERN PART OF THE SANTA MONICA MOUNTAINS, LOS ANGELES COUNTY, CALIFORNIA

By H. W. Hoots

ABSTRACT

The Santa Monica Mountains lie only a few miles northwest of the city of Los Angeles and comprise one of the prominent structural features that adjoin the Los Angeles Basin, one of the most prolific oil-producing districts of California. Even though the eastern part of these mountains may yield no oil, information concerning the rock types, structural character, and detailed geologic history of this area should be of value to petroleum geologists.

The area described in this report, which lies between Topanga Canyon on the west and the Los Angeles River on the east, presents a section of varied rock types including coarsely crystalline plutonic rocks, basic and acidic intrusive and pyroclastic rocks, metamorphic slate and schist, and a wide assortment of sedimentary rocks. The stratigraphic record is far from complete; the presence of Paleozoic rocks is very doubtful, and there is a gap in the early Tertiary record representing middle and late Eocene time and possibly part of Oligocene time. The Upper Cretaceous is well represented, and probably also the Triassic and Jurassic, although a large part and possibly all of the rocks tentatively considered to represent these two periods may be of Paleozoic age. The record of late Tertiary and Quaternary time, beginning with the lower Miocene, is fairly complete except for a fragmentary exposed Pliocene and Pleistocene record and a gap representing a considerable but unknown amount of the late middle Miocene.

Structurally the eastern part of the Santa Monica Mountains is a broad anticline whose axis lies in the extensive central area of Santa Monica slate (Triassic ?) and plunges westward from the major granitic intrusive mass just north of Hollywood. The attitude of the younger rocks, particularly those of Miocene age which cover so much of the north and south flanks of the mountains, conforms in a general way to this anticlinal structure. It is apparent from the presence of several pronounced unconformities, however, that this major fold has experienced several stages of growth and deformation. The anticlinal structure is still clearly obvious in the central part of the district; but in the eastern and western parts the original fold has been so intricately deformed by block faulting and igneous intrusion that much of the fold is either difficult to recognize as such or is down-faulted and entirely concealed beneath alluvium. Post-Topanga and pre-Modelo diastrophism produced an anticline which, to judge from its westward plunge, was complete in the district east of Topanga Canyon, although similar major uplifts of this age may have occurred farther west; post-Modelo diastrophism (pre-Pliocene and post-Pliocene), however, caused an anticlinal uplift that affected a larger area as a unit, an area which included the district west of Topanga Canyon as well as that east of it.

From the distribution of a remarkable spotted slate within the extensive area covered by the Santa Monica slate and the known relation of the spotted slate to the major exposed granitic intrusive mass, which is of probable Jurassic age, it is believed that much of the area north of the anticlinal axis is underlain by a much larger intrusive body of granitic rock. The broad character of the fold, a unique structural feature for the Coast Ranges of California, is that which might be expected to result from vertical uplift and is in marked contrast to the sharp asymmetrical folds of many Coast Range areas which appear to have undergone considerable lateral compression.

Two structural features of this area are particularly striking. One is the post-Topanga and pre-Modelo unconformity that represents the only period of folding which is comparable in importance with the deformation that occurred near the end of the Pliocene epoch. The other is the remarkably close association between faults and intrusive basalt in the pre-Modelo rocks of the Topanga and Santa Ynez Canyon district, an association which forces the conclusion that faulting and basalt intrusion had a close genetic relation during the period of post-Topanga and pre-Modelo diastrophism.

This area contains several types of rock which, to judge from the literature, are not common in California. Some of them are known elsewhere but have not yet been described in detail. The oldest of these unusual rocks is the spotted slate above mentioned, a contact-metamorphic facies of the Santa Monica slate.

Prominent reefs of white marine algal limestone from a few feet to several hundred feet thick occur in the Martinez formation (lower Eocene) and possibly also in the Chico formation (Upper Cretaceous) of some areas. These reefs commonly extend for not more than a few hundred feet and terminate abruptly. The limestone is distinctly nodular, has irregular bedding, and is characteristically spotted, owing to the abundance of nearly white irregularly shaped algae and algal colonies embedded in a limestone or argillaceous matrix of light brown or gray color. The algae appear to be of the lithothamnion type but have not yet been studied in detail.

The Modelo formation (upper Miocene) contains rock types of considerable interest. A massive bed of graywacke, dark gray because of an abundance of detrital slate fragments, occurs at the base. In some places this rock is very fossiliferous; in other places it is absent and its stratigraphic position is occupied by a 4 to 6 inch bed of oolitic phosphate, another type of rock which does not appear in the literature as a common constituent of the California Tertiary. The most distinctive feature of the Modelo formation is the abundance of hard white platy shale, a type of rock which is not at all unusual for the California Miocene but which, because of its highly siliceous character and association with beds of volcanic ash and bentonite, its microfossil content, and its remarkable banding, has proved to be worthy of study.

INTRODUCTION

This report covers a low mountain area which lies from 5 to 20 miles northwest of the city of Los Angeles and adjoins the Los Angeles Basin, one of the most prolific oil-producing districts of California. Although it appears unlikely that the eastern part of the Santa Monica Mountains will yield commercial amounts of petroleum, it seems that an understanding of the geology of these mountains will prove a practical aid to wildcat exploration for oil and gas in adjoining lowland areas of the Los Angeles Basin and San Fernando Valley.

PREVIOUS GEOLOGIC INVESTIGATIONS

W. P. Blake was apparently the first geologist to traverse a part of the Santa Monica Mountains and to record information regarding the types of rock exposed there. Arriving at San Fernando Pass on the evening of October 3, 1853,[1] Blake and his party camped by the side of a creek which they later found to be one of the tributaries of the Los Angeles River. Blake's description of their journey southward across the San Fernando Valley and the east end of the Santa Monica Mountains is both pleasing and instructive:

> Soon after leaving our camp under the fig trees, we found that we were entering a widely extended valley with a nearly level surface, without trees or verdure, and bounded on all sides by distant ranges of mountains. On turning the point of the hill, we came suddenly in sight of the mission [San Fernando] buildings, which, with the surrounding gardens, stood isolated in the seemingly desert plain and produced a most beautiful effect. The gardens were inclosed by walls, but the graceful palm rose above them, and groves of olive, lemon, and orange trees could be seen within. Outside of the walls the surface was barren and gravelly, and the fertility within is the result of irrigation. * * *
>
> Herds of cattle were seen on parts of the broad plain, feeding on dried grass or the burrs of the California clover, which covers the ground in the latter part of the summer when all the grass has disappeared. This plain doubtless presents a beautifully green surface in the winter and early summer, when watered by the rains. From the mission we passed directly [south] across the plain toward a low range of hills [the east end of the Santa Monica Mountains], which forms the boundary between it and the plain on which Los Angeles is built. The distance across the plain is about 10 miles, and the road was bordered in some places by a low growth of shrubbery and Cactaceae, which gave a peculiar aspect to the country and reminded some of the party of Mexican landscapes. The distant ranges of mountains had a peculiar barren look and in color were of various shades of brown, blue, and purple. When we reached the base of the hills we crossed a running stream, bordered by grass, which we afterwards found to be the Los Angeles River, and then the ascent of the hills immediately commenced.
>
> Range of sandstone hills between San Fernando and Los Angeles: This range appeared to extend nearly east and west, bounding the San Fernando Plain on the south and trending parallel with the Susannah Range on the north side. Like that range, this seemed formed of sedimentary strata, but they were not so well exposed; and we traveled in such haste that few observations on them were made.
>
> Toward the summit and near the roadside [probably in Cahuenga Pass] I found an outcrop of erupted rock, which was much obscured by decomposition but showed a globular character, the bank being filled with balls of various sizes, from which successive crusts of the decaying rock were scaling off. It had a dark color and contained considerable oxide of iron, indicated by the dark stains. This intrusive rock is represented on the general geological map, but subsequent observations will doubtless add many important facts to the now limited knowledge of the locality. * * *
>
> View of the Pacific Ocean: In descending from the higher parts of the range, the eye was permitted to wander over an extended area sloping gently away from the mountains toward the west. This is one of the most marked peculiarities of the landscape on the western coast; every mountain and mountain range is flanked by long, gently descending slopes, which seem like plains when passing over them, but viewed from a distance their inclination is strikingly evident. In the present instance the slopes appeared to be prolonged in a limitless plain extending to the horizon, but a more favorable point of view showed to us the broad, mirrorlike surface of the great ocean.

Dr. Thomas Antisell, geologist accompanying an exploring party under the direction of Lieut. John G. Parke, investigated the geology of parts of the Santa Monica Mountains and adjoining districts during the winter of 1854–55. Antisell mentions the occurrence of trachytic and "augitic trap" rocks and states that the range appears to be made up of strata over 500 feet in thickness which consist of hard red and yellow sandstones and "bituminous argillitic beds—soft rocks—including foraminiferous beds."[2] Three structure sections across the mountains accompany the text of his report.

In 1861 J. D. Whitney, State geologist of California, examined the eastern part of the Santa Monica Mountains and in the short time allotted to this area collected a large amount of remarkably accurate information regarding the major structure of the range and the age of some of the rocks.[3] He discovered that the core of this part of the range was composed of "dark siliceous slate" intruded and metamorphosed by granite and that an anticline existed in this old slate in the vicinity of Santa Monica Canyon.

> So that we have here one of the best possible examples of a truly anticlinal range of mountains, with a central core of granite, having all the appearance of an intrusive rock which has burst asunder and elevated the slaty strata, producing a highly metamorphic condition of the sedimentary beds along the lines of contact of the granitic mass.

Whitney also presents an instructive north-south structure section across the Santa Susana Mountains, San Fernando Valley, and Santa Monica Mountains.[4] He describes in detail the series of terraces at the mouth of Santa Monica Canyon and records the finding of fossils in the more eastern part of the range (probably in Brown Canyon or farther east), which W. M. Gabb found to be Miocene.

[1] Blake, W. P., U. S. Pacific R. R. Expl., vol. 5, pp. 73–76, 1856.

[2] U. S. Pacific R. R. Expl., vol. 7, pp. 76–78, 1857.

[3] Whitney, J. D., Geological survey of California, Geology, vol. 1, pp. 168–171, 1865.

[4] Idem, p. 121.

Prof. Jules Marcou, a geologist engaged in exploratory work for the United States Geographical Surveys West of the 100th Meridian, examined geologic features in parts of southern California in 1875–76.[5] He discusses in a general way the geology of the eastern part of the Santa Monica Mountains and apparently was particularly interested in the rocks containing fossil fish on the north flank of the mountains near the old Encino ranch—rocks which belong to the Modelo formation and which Marcou correctly classified as to age.

Watts, in the course of his pioneer work on the geology of the oil fields of California, briefly described the rocks exposed in the area west of Cahuenga Pass and also those along the shore west of Santa Monica.[6]

Eldridge [7] examined the oil-field regions of California for the United States Geological Survey about 1900 and briefly described the general geologic features of the Los Angeles district but gave little or no attention to the Santa Monica Mountains.

The only maps of the geology of the eastern part of the Santa Monica Mountains published by the Geological Survey are those by Ralph Arnold; one of them covers the southern border of the mountains east of the present site of Beverly Hills and accompanies a report on the oil-producing districts of southern California by Eldridge and Arnold.[8] Arnold's description of the general geologic features in that part of the Santa Monica Mountains covered by his report is in accord with data collected during the present investigation. Another map, largely if not entirely by Arnold, accompanies the railway guidebook for the Coast Line,[9] in which may also be found a brief description of some of the geology of this district.

In 1914 the California State Mining Bureau [10] published a map of the eastern third of the Santa Monica Mountains. This small-scale map was compiled by C. A. Waring from published reports of the United States Geological Survey and from field work done by R. N. Ferguson. It shows the general distribution of the granite, slate, basalt, and Miocene deposits; Topanga and lower Modelo sandstone and conglomerate on the north flank are classed as Monterey sandstone, the overlying Modelo of the north flank as Monterey formation, and the Modelo of the south flank as the Puente formation. Waring also gave a brief description [11] of the general stratigraphy and structure of the Los Angeles district and its oil fields.

More recently maps of the Los Angeles Basin have been published which show some of the geologic features of the eastern part of the Santa Monica Mountains.[12]

As a part of a study of the Cretaceous and Eocene deposits of southern California from 1910 to 1917, C. A. Waring [13] and his associates mapped a northwest-southeast belt across the central part of the Santa Monica Mountains, extending east and west from Topanga Canyon. This map, a part of which at least apparently is the result of work by a Stanford University geology party of which Waring was a member, shows the presence of the Chico (upper Cretaceous) and Martinez (lower Eocene) formations in the coastal belt east and west of lower Topanga Canyon.

Kew included the northwest corner of the area covered by this report in the area which he described in a recent bulletin.[14] As mentioned below, Kew also mapped the area farther east on the old topographic base but did not prepare a report before his resignation from the Geological Survey.

Schürmann [15] made two short trips into the eastern part of the Santa Monica Mountains and apparently also the Puente Hills, during which he noted the general geologic conditions and collected specimens from granitic rocks, slate, diabase, basalt, and andesite. He has described the petrographic character of these specimens and the general field relations of the rock masses from which they were collected.

FIELD WORK AND ACKNOWLEDGMENTS

W. S. W. Kew, prior to his resignation from the Geological Survey in 1923, mapped the eastern part of the Santa Monica Mountains as a part of a program to complete a geologic study of districts within and adjoining the Los Angeles Basin. His field mapping covered the eastern two-thirds of the area discussed in the present report and was done before the present large-scale and more detailed topographic maps of this region were available. In attempting to transfer Kew's geology from the old to the new maps it became apparent that additional field work would be necessary in order to adjust geologic boundaries to minor topographic features. Numerous new road cuts throughout the area were found to reveal much additional information, and it was decided to remap the area completely and to extend the mapping westward to Topanga Canyon. The writer gratefully acknowledges the valuable assistance derived from occasional field trips and discussions with Doctor Kew and from inspection of his maps, which have been at hand during

[5] Marcou, Jules, Report on the geology of southern California: U. S. Geog. Surveys W. 100th Mer., Ann. Rept., for 1876, pp. 158–160, 1876.

[6] Watts, W. L., oil and gas yielding formations of Los Angeles, Ventura, and Santa Barbara Counties: California State Min. Bur. Bull. 11, pp. 4–5.

[7] Eldridge, G. H., The petroleum fields of California: U. S. Geol. Survey Bull. 213, pp. 318–319, 1902.

[8] Eldridge, G. H., and Arnold, Ralph, The Santa Clara Valley, Puente Hills, and Los Angeles oil districts, southern California: U. S. Geol. Survey Bull. 309, 1907. For geology of the Los Angeles district, including part of the Santa Monica Mountains, see pp. 144–202.

[9] Diller, J. S., and others, Guidebook of the western United States, Part D, The Shasta Route and Coast Line: U. S. Geol. Survey Bull. 614, sheet 1A and p. 97, 1915.

[10] McLaughlin, R. P., and Waring, C. A., Petroleum industry of California; California State Min. Bur. Bull. 69, pl. 11, 1914.

[11] Idem, pp. 350–358.

[12] Kew, W. S. W., A geologic summary of California oil fields: Oil Bull., January, 1926. Eaton, J. E., A contribution to the geology of the Los Angeles Basin, California: Am. Assoc. Petroleum Geologists Bull., vol. 10, No. 8, fig. 1, p. 754, 1926. Vickery, F. P., Geology of the Los Angeles Basin: Oil Bull., April, 1928.

[13] Waring, C. A., Stratigraphic and faunal relations of the Martinez to the Chico and the Tejon of southern California: California Acad. Sci. Proc., 4th ser., vol. 7, No. 4, pp. 53–57, fig. 3, 1917.

[14] Kew, W. S. W., Geology and oil resources of a part of Los Angeles and Ventura Counties, Calif.: U. S. Geol. Survey Bull. 753, 1924.

[15] Schürmann, H. M. E., Beitrag zur Petrographie der Hollywood Hills (Santa Monica-Gebirge) bei Los Angeles: Centralbl. Mineralogie, 1928, Abt. A, pp. 7–13.

the present study. Field work for this report was done during the summer and fall of 1927.

Dr. R. D. Reed had examined many of the interesting features of the Santa Monica Mountains prior to the present investigation and has served as an unlimited source of inspiration and assistance in problems relating to stratigraphy and sedimentation.

The paleontologic work on the Bryozoa, brachiopods, mollusks, and echinoids by Dr. W. P. Woodring, on the Foraminifera by Messrs. W. D. Rankin, D. D. Hughes, and P. P. Goudkoff, on the diatoms by Mr. K. E. Lohman, and on the fossil plant material by Dr. R. W. Chaney has been indispensable to this investigation, and the conclusions of these men, presented in the following pages, add much to our knowledge of the Cretaceous and Tertiary history of the Santa Monica Mountains. The writer is particularly indebted to Doctor Woodring for his work on the macroscopic fossils, his assistance in collecting these fossils, and his suggestions as to the arrangement of the paleontologic matter in the text. Only with his assistance was it possible to place the marine Pliocene and Pleistocene rocks in their approximate stratigraphic order.

Messrs. W. D. Rankin and W. M. Smith, of Los Angeles, have assisted in the collection of samples of foraminiferal material from the Modelo formation, and Mr. Smith has kindly made several mechanical analyses of sandstone samples. Mr. Lohman, in addition to contributing valuable information regarding diatoms in the Modelo, has made the photographs of rock specimens reproduced in this report. Mr. N. A. Zaitzevsky, of Pasadena, has assisted in the preparation and examination of sandstone samples for their heavy mineral content.

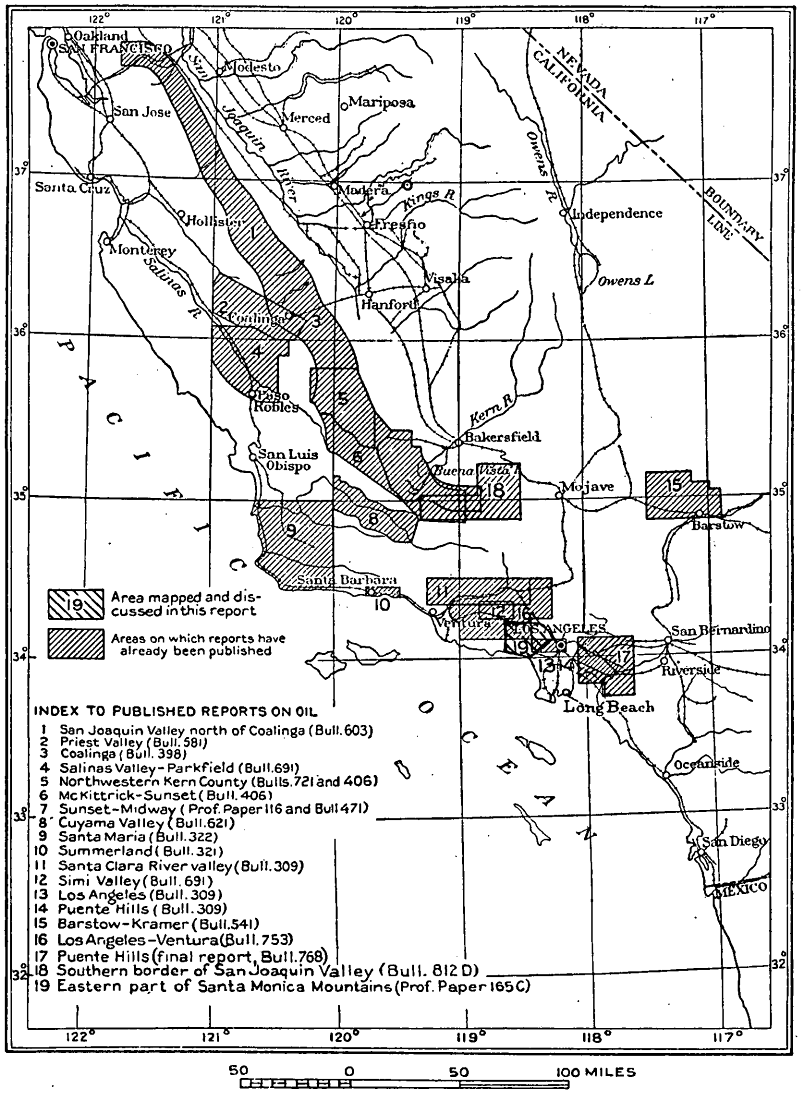


FIGURE 7.—Index map of California showing areas that include or adjoin oil-producing districts and are described in published reports of the U. S. Geological Survey

GEOGRAPHY

The Santa Monica Mountains trend in an east-west direction and form the highlands that adjoin the coast just north of the thirty-fourth parallel. The chain of the Santa Barbara Islands, directly to the west, appears to represent remnants of a westward continuation of this mountain range, much of which is now deeply submerged beneath the Pacific Ocean. The eastern part of the Santa Monica Mountains protrudes into the broad alluvial plains extending southward from the San Gabriel and Santa Susana Mountains and separates the San Fernando Valley on the north from the Los Angeles Basin lowland on the south. A minor southeastward-trending group of hills lies north of the major part of Los Angeles and, were it not for the transverse valleys of the Los Angeles and San Gabriel Rivers, would form a continuous connection between the east end of the Santa Monica Mountains and the Puente Hills. The western part of the Santa Monica Mountains merges northward into the Simi Hills, which separate the San Fernando Valley on the east from the lower part of the Santa Clara River Valley.

The area described in this report (see fig. 7 and pl. 16) comprises the eastern two-fifths of the Santa Monica Mountains and extends from Topanga and Garrapata Canyons on the west to the Los Angeles River near Glendale. It adjoins and includes along its southern border some of the finest residential districts in California, such as Beverly Hills and Hollywood, and its east end is within 5 miles northwest of the business district of Los Angeles. The city of Santa Monica is situated on the coast at the southern edge of the area, and other beach resorts—Venice, Redondo Beach, and Hermosa Beach—lie directly to the south. Burbank, Glendale, and Pasadena lie to the northeast and east—Pasadena about 8 miles from the east end of this area. The San Fernando Valley on the north, with its numerous centers of habitation, such as Van Nuys and Lankershim, and its broad expanse of intensively cultivated truck gardens, fruit and walnut groves, and farms for dairy cattle, chickens, and rabbits, extends in a broad panorama before the observer who passes along the Mulholland Highway. Practically all of this area, and much of the San Fernando Valley as well, lies within the city limits of Los Angeles. The eastern part of the mountains, east of Stone Canyon, and the southern slopes farther west, which overlook the Pacific Ocean and the Los Angeles Basin, are fast becoming high-class residential property. Some of the accessible canyons, such as Topanga Canyon, contain a fine residence here and there, but more generally only cabins for intermittent use by inhabitants of the cities.

As a result of recent developments most of the area is accessible by automobile, a fact which has greatly facilitated the completion of this project. Hollywoodland and Griffith Park, the area east of Cahuenga Pass and south of Cahuenga Peak, contain a maze of paved streets. West of Cahuenga Pass the crest of the mountains is traversed from east to west by the Mulholland Highway, which connects with the Topanga Canyon road, a north-south highway along the western border of the area. The northern base of the mountains is skirted by the Ventura Boulevard, from which several paved roads branch off to the south, cross the mountains, and lead into Hollywood and Beverly Hills.

The Los Angeles River, the major stream of the district, heads in the Simi Hills and Santa Susana Mountains, to the northwest and north, and flows southeastward across the San Fernando Valley and around the east end of the Santa Monica Mountains. It is an intermittent stream without surface water throughout much of its course during the summer. During the winter, however, this stream, in the vicinity of Glendale and Los Angeles, commonly becomes a torrent over night. Topanga Canyon and others of the major southward-draining canyons of the Santa Monica Mountains, which are fed by permanent springs, contain water throughout the year. Stone, Franklin, and Coldwater Canyons, north and northwest of Beverly Hills, do not themselves contain large permanent streams but are used as the sites for city reservoirs, which are supplied largely by water transported from the Sierra Nevada by aqueduct. Hollywood Lake, just north of Hollywood and east of Cahuenga Avenue, is another such reservoir which, like the others, adds materially to the beauty of the local landscape.

Most of the area described in this report is covered by a substantial growth of brushy vegetation, which presents a serious handicap in mapping the geology, particularly on the crest and southern slopes of the mountains in the western part of the area, where chaparral is so dense as to be practically impenetrable. Chaparral appears to favor the sandy soil derived from Cretaceous, Eocene, and Miocene formations, but the distribution and rankness of this type of vegetation in this area appear to be controlled more by moisture than by soil, as it is much thicker in the fog belt adjoining the ocean than it is farther east and is almost as thick in areas of slate as in areas of sandstone and conglomerate. Surfaces underlain by Modelo shale are commonly covered with grass and white sage and support scattered small black walnut and oak trees. The bottoms of many of the deep canyons, particularly the long ones which drain southward from the crest, are heavily wooded with oak and contain an occasional sycamore and a variety of shrublike undergrowth.

The eastern part of the Santa Monica Mountains, except for small isolated areas, has a strikingly subdued topographic form. The major streams drain southward from the main divide, which occupies an east-west line much nearer the northern than the southern edge of the mountains. The intervening ridge tops are almost flat and are remarkably concordant; for distances of 2 miles or more they extend southward from the divide across the central part of the range with little variation in height. The average altitude of the divide throughout most of the eastern part of the range is between 1,400 and 1,500 feet, although the western one-third of this part is characterized by higher areas and contains some peaks south of the divide and in the central part of the mountains which are 2,125 to 2,150 feet above sea level. The major southward-draining canyons are deep and steep walled and follow comparatively straight courses to the alluvial plain and the Pacific Ocean.

GENERAL GEOLOGY

STRATIGRAPHY

The Santa Monica Mountains present a section of varied rock types, including coarsely crystalline plutonic rocks, basic and acidic intrusive and pyroclastic rocks, metamorphic slate and schist, and a wide assortment of sedimentary rocks. The stratigraphic record is far from complete; the presence of Paleozoic rocks is very doubtful, and there is a gap in the early Tertiary record representing middle and late Eocene time and possibly part of Oligocene time. The Mesozoic appears to be fairly well represented, although Jurassic rocks may not be present and the age of the supposed Triassic deposits is not established. Except for a fragmentary exposed Pliocene and Pleistocene record and a gap representing a considerable but unknown thickness of late middle Miocene rocks the record of late Tertiary and Quaternary time, beginning with the lower Miocene, is fairly complete.

The accompanying table gives a list of the rock formations exposed in the eastern part of the Santa Monica Mountains and information regarding their probable age and general characteristics:

Rock formations exposed in the eastern part of the Santa Monica Mountains

Geologic age	Formation	Approximate thickness (feet)	Character
Recent	Alluvium	0–100	Breccia conglomerate, sandstone, and silt.
Pleistocene	Alluvial plain deposits now uplifted and deeply dissected.	0–300	Poorly sorted reddish-brown breccia conglomerate and sandstone with earthy matrix and indistinct bedding. This alluvial plain, now far above present level of drainage, has been named Santa Monica Plain.
	Marine upper Pleistocene	5–100	Fossiliferous soft gray sandstone, sandy clay, and conglomerate.
	Probable unconformity.		
	Upper Pliocene or lower Pleistocene.	100	Fossiliferous conglomerate sandstone and sandy clay.
Unconformity (gently dipping upper Pliocene or lower Pleistocene rests directly upon Pliocene dipping as much as 70°).			
Pliocene	Pliocene, upper and lower	1, 000	Soft dark-gray clay and sandstone with lenses and concretions of yellowish-gray limestone. Exposed only in coastal belt northwest of Santa Monica.
	San Diego formation (middle or upper Pliocene).	0–35	Massive soft light-gray conglomeratic sandstone.
Unconformity (some folding and erosion but extent uncertain because of scanty distribution of Pliocene).			
Upper Miocene	Modelo formation	2, 300	White punky diatomaceous and foraminiferal shale and fine sandstone, grading laterally into clay shale and sandstone.
		2, 250	Soft light-gray to brown well-bedded shale, banded hard platy siliceous shale, thin and thick massive beds of sandstone and conglomeratic sandstone, and volcanic ash. Much of shale is foraminiferal.
Unconformity (represents the most pronounced pre-Pliocene deformation, which included folding, faulting, and basalt intrusion).			
Middle Miocene	Topanga formation	4, 500–7, 500	Massive fossiliferous sandstone and conglomerate and thin-bedded shale and sandstone intercalated with intrusive and extrusive basalt and pyroclastic rocks of lower and upper Topanga age. Basal 1,000 feet of conglomerate between Stone Canyon and Cahuenga Avenue may be Vaqueros.
Lower Miocene (?) and Oligocene (?).	Vaqueros (?) and Sespe (?) formations.	3, 500–4, 000	Light-gray and red conglomerate and conglomeratic sandstone. Unfossiliferous.
Unconformity (a notable stratigraphic gap produced, at least in part, by folding and erosion of uncertain magnitude).			
Lower Eocene	Martinez formation	250+	Soft brown shale, sandy shale, and sandstone, with hard limestone concretions containing fossils. Prominent discontinuous reefs of white algal limestone.
Upper Cretaceous	Chico formation	8, 000+	Massive brown and gray conglomerate and sandstone and dark-gray shale. Fossiliferous. May also contain reefs of white algal limestone.
Unconformity (not exposed, but unquestionably present because of striking difference in metamorphism of Cretaceous rocks and older slates, exposed contacts of which are faults).			
Jurassic(?)	Granitic intrusion	------------	Granite and granodiorite. May be of Paleozoic age.
Triassic (?)	Santa Monica slate	5, 000–7, 000	Black slate, much of which has undergone contact and regional metamorphism and is locally altered to mica schist. Base not exposed.

TRIASSIC (?) ROCKS

SANTA MONICA SLATE

A large area in the central part of the district covered by this report (see pl. 16), comprising about one-fourth of the total area described, consists of dark-gray and bluish-gray to black slate which is herein named the Santa Monica slate, from its extensive exposures in the central area of the Santa Monica Mountains, east of Topanga Canyon. No fossils have been found in this formation, but in view of its similarity to the fossiliferous Triassic slate of the Santa Ana Mountains,[16] both in lithologic character and in its relations to fossiliferous Cretaceous rocks and an earlier granitic intrusion, the Santa Monica slate is considered to be of probable Triassic age. It is entirely possible, however, that this slate and the younger granitic intrusion are both Paleozoic or both Jurassic. Except for metamorphic facies and small areas of basalt intrusions the entire formation is remarkably uniform in character and consists essentially of hard dark-gray slate with only a few thin beds of equally hard light-gray siltstone and fine to coarse quartzitic sandstone. Locally some of the slate is soft and can be easily dug into with a hammer—a condition which lends encouragement to the hope that fossils once deposited in this old altered mudstone have not been entirely destroyed and may yet

[16] Mendenhall, W. C., in Willis, Bailey, and others, Index to the stratigraphy of North America: U. S. Geol. Survey Prof. Paper 71, pp. 505–506, 1912. Smith, J. P., The Middle Triassic marine invertebrate fauna of North America: U. S. Geol. Survey Prof. Paper 83, p. 145, 1914.

be found. Weathered surfaces are commonly brown, owing to the abundance of limonite deposited by percolating waters along joints and bedding or cleavage planes, and offer a rather distinct color contrast between slopes of this rock and the intrusive granite and granodiorite.

Slaty cleavage is well developed and, wherever bedding is distinct, is parallel to the original bedding planes of the rock. As a result of the fissile nature and the abundance of joints the slate commonly weathers to chips and thin slabs only a few inches across, a characteristic which tends to prevent this rock from being of commercial value for building.

Much of the slate is altered by contact metamorphism induced by the Jurassic (?) granitic intrusion. A zone 1,500 to 2,000 feet wide adjoining the major granitic mass north of Beverly Hills (see pl. 16) and a larger mass farther west between Brown and Sepulveda Canyons consists largely of mica schist and dark-gray phyllite. This rock, although commonly having a fairly uniform structure, is locally cut by numerous quartz veins and is very much distorted. (See pl. 18, *A*.) The outer boundary of the eastern schistose zone parallels, in a general way, the border of the intrusive mass and, as indicated on the geologic map, separates this zone from an extensive belt of spotted slate, the approximate distribution of which has been mapped. The spotted slate in turn appears to grade into ordinary unspotted slate through a zone in which the individual spots become progressively smaller and finally disappear.

Plate 18, *B*, is a photograph of a hand specimen of the spotted slate. Most of these spots are well-developed crystals, apparently of the mineral cordierite. Their development within this rock may be seen in thin sections to have produced a striking reorientation of the slaty cleavage within the crystal boundaries. Each spot is an individual crystal, but there is in many places a slight difference in the extinction of parts of the crystal, such as between the border or irregular areas and the remainder of the crystal. The crystals are roughly spindle-shaped, but a few have good prismatic form with parallel extinction, negative elongation, and rarely distinct transverse cleavage. The crystals are heavily charged with inclusions of biotite and a dark indeterminate substance. Zonal growth made apparent by zonal distribution of inclusions or concentric variations in the birefringence is common. The birefringence is low, being about that of quartz. The index of refraction in most crystals is only slightly less than 1.57, but in others it may be slightly greater than 1.57. The sign of the mineral is commonly negative, with 2V indeterminate, but some oblique sections suggest positive sign and moderate 2V.[17]

[17] Dr. A. O. Woodford, of Pomona College, has been very helpful in determining these properties and identifying this mineral.

The distribution of the spotted slate is irregular and widespread, and the occurrence of much of this rock has no apparent relation to exposed masses of the granitic intrusive. In view of the relation of some of the schist and spotted slate to the main granitic mass farther east and of the occurrence in foreign countries of other bodies of schist and spotted slate as the product of contact metamorphism,[18] it seems probable that the wide distribution of these two types of rock in the central and northern part of the slate-schist area is the result of metamorphism by a buried body of granite, which may well form an east-west underground connection between the widely separated granitic masses now exposed.

The writer did not find schist and spotted slate closely associated with the minor granitic masses farther west along the Mulholland Highway and can offer no reasonable explanation for their apparent absence. Additional field investigation may provide an explanation for this discrepancy.

JURASSIC (?) ROCKS

GRANITE AND GRANODIORITE

The Santa Monica slate in the central part of this area is intruded by granite, diorite and granodiorite, which, in turn, like all other rocks of this region older than upper Miocene, are cut by numerous minor intrusions of basalt. The intrusive relations of the granitic rocks with the slate are most evident in Higgins Canyon and in road cuts on the adjoining ridge to the east. At these localities stringers of slate and schist are included in the granite, and small elongate masses of granite occur in the slate. Here and elsewhere along the contact both the granite and the Santa Monica slate are cut by innumerable veins of white quartz.

The easternmost part of the mountains, in and near Griffith Park, contains several irregular masses of granitic rocks, most of which have been brought to the surface by faulting. Generally where the granite is not in fault contact with other rocks in this area and in the district west of Cahuenga Avenue it is overlain, with normal depositional contact, by lower Miocene conglomerate.

These granitic rocks are variable in character and consist of light-gray biotite granite and dark-gray diorite and granodiorite, the last consisting of green hornblende, quartz, orthoclase and plagioclase feldspar, and biotite, together with apatite, zircon, and garnet in varying proportions.[19] The granodiorite

[18] One of the best-known examples of the development of spotted slate as a result of contact metamorphism by a granite intrusion is that of the Steiger Schiefer in Alsace-Lorraine, described by Rosenbusch (Die Steiger Schiefer und ihre Contact-Zone an den Granititen von Barr-Andlau und Hochwald: Abh. zur Geol. Specialkarte von Elsass-Lothringen, Strassburg, 1877). For a description of Rosenbusch's results see Teall, J. J. H., British petrography, pp. 373-375, London, 1888, or Hatch, F. A., and Rastall, R. H., Textbook of petrology, The sedimentary rocks, pp. 251-254, London, 1923.

[19] For a more detailed petrographic description of some of the plutonic rocks see Schürmann, H. M. E., Beitrag zur Petrographie der Hollywood Hills (Santa Monica-Gebirge) bei Los Angeles: Centralbl. Mineralogie, 1928, Abt. A, pp. 7-13.

near the border of the large intrusion north of Beverly Hills is distinctly gneissic and highly micaceous, a characteristic which is well exposed in and near Franklin Canyon.

These granitic rocks are almost invariably deeply weathered and crop out as soft masses which have undergone marked disintegration and partial decomposition of mineral aggregates. These deeply weathered plutonic rocks are commonly less resistant to erosion than the lower Miocene conglomerates.

Inasmuch as there are reasons for suspecting that similar granitic masses in the Santa Lucia Mountains, farther north in the Coast Ranges, are pre-Franciscan (pre-Jurassic?), the possibility must be considered that the granite and granodiorite of the Santa Monica Mountains are also pre-Jurassic and of Triassic or Paleozoic age.

UPPER CRETACEOUS AND EOCENE ROCKS

In the western part of the area covered by this report, in the Topanga Canyon and Reseda quadrangles, a large area is covered with rocks of the Upper Cretaceous (Chico) and lower Eocene (Martinez) age, which in most of the area have not been separately mapped because of a dense covering of brush and unexposed structural complications. Separation has been possible, however, in the Reseda quadrangle, where the structure is relatively simple and the lithologic and faunal distinctions between the Chico and Martinez formations are readily determinable.

CHICO FORMATION OF THE RESEDA QUADRANGLE

The Chico formation in the Reseda quadrangle, the northwestern part of the area here described, crops out along and north of the Mulholland Highway, but has its greatest areal exposure just south of this highway. It is readily divisible into two distinct members—a lower one that consists entirely of very soft friable red conglomerate and sandstone, lying directly on the Santa Monica slate, and an upper one which consists in greater part of hard massive greenish-brown and gray conglomerate but contains intercalated units of hard dark-gray thin-bedded shale, sandstone, and limestone and thick beds of light-gray conglomeratic sandstone. (See pls. 18, *C*, and 19, *A*.) Some of the beds of hard dense limestone contain *Scaphites* and *Baculites*, and the light-gray sandstone locally yields a variety of other mollusks in addition to several species of ammonites.

The lower member of soft red conglomerate and sandstone, together with its relations to the underlying slate and the overlying upper member of the Chico, is best exposed on the Mulholland Highway just west of the area of slate shown on Plate 16. The rock, like that of the overlying member, is arkosic and contains well-rounded and polished cobbles of varicolored quartzite, dense poryphy, granite, and basalt, together with chips of black slate. Although a clean-cut contact is not exposed, it probably rests with normal depositional contact upon the Santa Monica slate, and it appears to grade upward into the overlying more highly indurated brown conglomerate. The thickness of the lower red conglomerate member varies in the direction of the strike, but along the Mulholland Highway it is approximately 750 feet. Farther south this member rapidly decreases in thickness, although the thinning may be due in part to displacement along an unmapped branch of the Temescal fault.

The upper member, although containing considerable sandstone and shale and some limestone, is composed of hard massive brown conglomerate to the extent of about 75 per cent. This conglomerate presents a compact mass of rounded cobbles which average from 3 to 5 inches in diameter, embedded in a matrix of clean micaceous sandstone having a marked greenish-brown tinge. This matrix is one of the most distinctive characteristics of much of the conglomerate of the Chico and is in striking contrast to the light-gray arkosic matrix common to conglomerates of Miocene age. This upper Chico conglomerate, although compact and well indurated is only fairly resistant to erosive agencies and commonly weathers to rounded slopes of reddish-brown color.

The total exposed thickness of the upper member of the Chico formation in the Reseda quadrangle is about 2,500 feet. In the northern border of its outcrop it is overlapped unconformably by the Modelo formation (upper Miocene), but farther south it is overlain with apparent structural conformity by fossiliferous shale and limestone of Martinez (lower Eocene) age.

Fossils collected from the lower and upper parts of the upper member of the Chico formation in this area are listed in the accompanying table under localities 16, 18, 19, and 22. W. P. Woodring materially assisted the writer in making fossil collections from the Chico and Martinez. His identifications of the Chico fossils collected from the Reseda quadrangle and also from the area farther south appear in the following table:

Fossils from the Chico formation of the eastern part of the Santa Monica Mountains

[Locality numbers plotted on Plate 16. For description of localities see pp. 123-124]

	16	17	18	19	22	23	41	46	47	48	49	54
Cephalopods:												
Eutrephoceras sp												×
Baculites sp			×									×
Pachydiscus sp. a (giant species)												×
Pachydiscus sp. b												×
Pachydiscus sp. c												×
Scaphites cf. S. gillisi Anderson	×		×									
Metaplacenticeras sanctaemonicae (Waring)		×			?	×			?	×	×	×
Metaplacenticeras californicum (Anderson)												×
Gastropods:												
Gyrodes compressa Waring						×			?			
Gyrodes sp											×	
Ampullina pseudoalveata (Packard) ?		×							×			
Turritella chicoensis Gabb										×		×
Conchothyra rotunda (Waring) ?											×	
Volutoderma averillii (Gabb)												×
Oligoptycha obliqua (Gabb)									×			×
Lamellibranchs:												
"Leda" sp									×			
Cucullaea youngi Waring											×	
Cucullaea truncata Gabb												×
Parallelodon brewerianus (Gabb)							×					×
Glycymeris veatchii (Gabb)		×			×	×		×		×	×	×
Inoceramus sp	×		×	×								
Trigonia evansana Meek		×			×	×		×	×	×	×	×
Pholadomya sp											×	
Crassatella sp. a					×							
Crassatella sp. b											×	
Protocardia sp						×						
Aphrodina varians (Gabb)		?				×			×	?	×	
Cymbophora gabbiana (Anderson)					×	×				×	×	×
Corbula sp											×	

Doctor Woodring makes the following comments regarding these Chico fossils:

Metaplacenticeras is recorded elsewhere from the basal part of the Chico formation, which is regarded as of Cenomanian age.[20] Aside from *Metaplacenticeras californicum* the rather meager ammonite fauna from the Santa Monica Mountains is not similar to that recorded by Anderson from the lower Chico of the Santa Ana Mountains. As more than one horizon is included in the lower Chico it seems probable that the beds in the Santa Monica Mountains represent a higher horizon than those in the Santa Ana Mountains carrying *Acanthoceras*. None of the specimens referred to *Metaplacenticeras sanctamonicae* and *M. californicum* has the tricarinate venter characteristic of *M. pacificum* (Smith), the type of the genus, up to a diameter of 100 millimeters.

Thirteen species of Chico fossils have already been recorded by Waring [21] from a locality in the eastern Santa Monica Mountains.

MARTINEZ FORMATION OF THE RESEDA QUADRANGLE

Near the head of Santa Ynez Canyon, in the southern part of the Reseda quadrangle, the coarse conglomerate of the Chico formation is overlain by a comparatively thin unit of shale, sandstone, and limestone which has yielded a few fossils characteristic of the Martinez formation (lower Eocene). The shale is compact and well bedded and varies in color from light brown to gray. It is sandy in part and is interbedded with thin layers of fine sandstone and with prominent but discontinuous reefs of white algal limestone.

In this locality the Martinez formation has an approximate thickness of 250 to 350 feet. Although the change at the Chico-Martinez contact from the massive upper conglomerate of the Chico to the overlying shale is abrupt and may well represent a break in sedimentation and a resulting hiatus in the stratigraphic record, there appears to be no discordance in either dip or strike of the two formations. The Martinez in this locality is, however, definitely unconformable with the overlying Vaqueros(?)-Sespe(?), there being a complete absence of much if not all of the middle and upper Eocene and a difference in strike of at least 30° and in dip of about 20° between the two formations in contact.

Fossil collections from brown shale at localities 26 and 27 have yielded several species, but the white algal limestone at locality 21 contains only unidentifiable fragments of mollusks. A giant "*Lima*" was collected from the algal limestone at locality 67C, in the Topanga Canyon quadrangle, near the limestone quarry in upper Santa Ynez Canyon. The best collection of Martinez fossils from the Reseda quadrangle came from a gray limestone concretion associated with shale and white algal limestone in a small isolated patch farther north (locality 62), apparently faulted into the basal part of the Modelo formation.

[20] Smith, J. P., California Acad. Sci. Proc., 3d ser., vol. 1, p. 204, 1900. Anderson, F. M., idem, vol. 2, pp. 27, 28, 34, 79-80, 1902. Reeside, J. B., jr., U. S. Geol. Survey Prof. Paper 147, p. 2, 1926.

[21] Waring, C. A., California Acad. Sci. Proc., 4th ser., vol. 7, pp. 56-71, 1917 (locality 3).

The algal limestone is one of the most striking and probably the most unusual rock types in the Santa Monica Mountains. It occurs in prominent white reefs from a few feet to several hundred feet thick which vary in lateral extent from only a few feet to about 4,000 feet and commonly terminate in an abrupt wall. The outcrop of a typical ledge of algal limestone is shown in Plate 19, *B*. This limestone is distinctly nodular and has irregular bedding. Although weathered outcrops are commonly white, some black or very dark gray algal limestone occurs, and fresh exposures of even the white-weathering rock have a characteristically spotted appearance due to the abundance of nearly white irregularly shaped algae and algal colonies embedded in a limestone matrix of light-brown or gray color. In some places there appear to be all grades of purity of this limestone, a condition which may be noted in single outcrops or even in hand specimens where comparatively pure white or light-brown algal limestone grades laterally or vertically into a darker-gray algal rock with a matrix that is highly argillaceous. Commonly this increase in argillaceous material is accompanied by a decrease in the number of algal growths. Well-rounded cobbles of quartzite or granite occur here and there within these limestone deposits.

Additional information regarding the character, tonnage, and plans for commercial exploitation of the Eocene limestone deposits is given on pages 133–134.

In the table below appears a list of fossils collected from the Martinez formation of the Reseda quadrangle and from beds of the same age in the Topanga Canyon quadrangle, farther south. These fossils have been identified by W. P. Woodring.

Fossils from the Martinez formation of the eastern part of the Santa Monica Mountains

[Locality numbers plotted on Plate 16. For description of localities, see pp. 123–124]

	26	27	43	44	45	59	62	67	67A	67B	67C
Gastropods:											
Euspira susanaensis (Nelson)			×	×				?		×	
Globularia sp.							×				
"Amauropsis" martinezensis (Dickerson)			×								
Turritella infragranulata infragranulata Gabb							×				
Turritella infragranulata pachecoensis Stanton	?		×	×						×	
Turritella subuvasana Nelson								×	×	×	
Priscoficus caudatus (Gabb)										×	
"Cerithium" sp.								×			
Sycum mucronatum (Gabb)			×	×							
Retipirula crassitesta (Gabb)			×	×							
"Lyria" hannibali Waring			×	×							
Pleurofusia sp.			×								
Lamellibranchs:											
"Leda"	×		×			×	×			×	
Cucullaea mathewsonii Gabb							×			×	
Glycimeris major Stanton				×			×			×	
Vulsella? sp.					?					×	
"Lima" haseltinei Dickerson											×
"Modiolus" sp.	×							×			
Pholadomya nasuta Gabb							×	?			
"Phacoides" sp.					×						
Nemocardium sp. cf. N. linteum Conrad									×		
Plagiocardium sp. cf. P. brewerii Gabb									×	×	
"Macrocallista" stantoni Waring		×		×							
"Callocardia" simiensis Nelson			×	×			×		×		
"Callocardia" sp.							×				
"Meretrix" stantoni Dickerson						×					
"Tellina" undulifera Gabb		×					×		×		
Teredo? sp.					×						
Echinoid:											
Schizaster martinezensis Kew		×									

CHICO AND MARTINEZ FORMATIONS OF THE TOPANGA CANYON QUADRANGLE

A large tract in the southwestern part of the area covered by this report, mostly in the Topanga Canyon quadrangle, is covered by steeply dipping rocks of the Chico and Martinez formations. Because of the dense covering of brush and the apparent presence of unexposed structural complications, it has not been possible to differentiate these formations in this tract, at least within the time available for this investigation. Collections of fossils characteristic of the Chico and Martinez formations, together with a rather distinct difference in lithologic character, provide, however, a basis for a description of the probable approximate distribution of these formations.

Chico fossils have been collected in this undifferentiated area from localities 17, 46, 47, 48, 49, and 54. (See pp. 123–124 and pl. 16.) Examination of the geologic map will make clear the distribution of these localities and of reefs of the white algal limestone which are known to be present in the Martinez formation farther north and which have nowhere been found closely associated with Chico fossils or typical Chico rocks. Except for a single small reef of white algal limestone in the east wall of Topanga Canyon, all such reefs are restricted to that part of the Upper Cretaceous and Eocene area which is bounded on the west by lower Miocene rocks and the lower part of Santa Ynez Canyon, on the south by a west-southwestward-trending fault across upper Pulga Canyon, on the east by the north-northwest fault that lies just west of Temescal Canyon, and on the north by a southwestward-trending fault near the head of Santa Ynez Canyon. The limestone reefs in this north-south elongated belt of dense brush and poor exposures are associated with rocks that consist largely of compact dark-gray, greenish-gray, and brown shale together with intercalated beds of greenish-gray and light-gray to brown sandstone, most of which are soft, thin, and of fine texture, although massive 50-foot beds of poorly sorted arkosic sandstone and conglomerate are present in upper Pulga Canyon. Fossils collected from this area at localities 43, 44, 45, 59, 67, 67A, 67B, and 67C,[22] and west of upper Pulga Canyon are listed in the table on page 92 and are considered by Woodring to be of Martinez age.

In the area adjoining the above-described area containing Martinez strata the Chico rocks are largely of strikingly different character, consisting in greater part of hard massive coarse brown conglomerate associated with dark gray shale and beds of hard gray sandstone which in places are highly fossiliferous. These beds are well exposed in Topanga, Santa Ynez, and Temescal Canyons. In Topanga Canyon the exposures are remarkably good, but here the Chico rocks are cut by numerous faults, so that interrupted sections of only small parts of this formation are exposed. The same is probably true to some extent of Santa Ynez Canyon, although the existence here of important faults in the Cretaceous rocks is difficult to prove. In Temescal Canyon and farther east there appears to be a continuous section of about 8,000 feet of southward-dipping Chico rocks, much the greater part of which is coarse brown cobblestone conglomerate like that along the highway in Topanga Canyon.

The cobblestone conglomerate of the Chico is the most remarkable feature of this formation. Its uniformity in the abundance and striking roundness of most of the cobbles and boulders, the existence everywhere of distinct bedding and a matrix of clean gray to greenish gray sandstone, and the presence here and there of marine fossils in associated beds of gray sandstone are features which support the belief that this conglomerate is marine, even though it and the associated beds appear to have a total thickness of 8,000 feet or more. Certainly none of it was laid down by streams on the land. A theory to account for its origin need not postulate any conditions strange to many parts of California in late geologic time; the presence of a continuously subsiding Chico basin of marine deposition adjoining a continuously rising land mass of strong relief appears to be all that is necessary.

[22] Collection 67C, consisting of a giant "*Lima*," was contributed by J. H. Gilliland, and collection 67B by J. M. Tate, both of the Los Angeles Mountain Park Co.

Conglomerate similar lithologically to the conglomerate of the Chico but containing a large fossiliferous boulder which has yielded, according to W. P. Woodring, *Mesalia martinezensis* (Gabb) and a *Turritella* of Eocene aspect crops out in the west wall of lower Topanga Canyon about 2 miles below Topanga and about 50 feet north of a spring that is used extensively by the public. It seems apparent, therefore, that much of the older conglomerate in lower Topanga Canyon is Eocene or later and that some of the massive older conglomerate elsewhere in this district may be of Eocene age. Unless unexposed faults have complicated conditions, however, the 8,000 feet of conglomerate in and east of lower Temescal Canyon is Chico, because typical Chico fossils have been collected near the top of these beds at locality 49. (See pl. 16.)

OLIGOCENE (?) AND LOWER MIOCENE ROCKS

SESPE (?) AND VAQUEROS (?) FORMATIONS

In the upper drainage area of Santa Ynez Canyon there is a thick unfossiliferous series of soft coarse red and light-gray to white arkosic sandstone and conglomerate which underlies, with apparent conformity, the massive brown fossiliferous sandstone of the lower part of the Topanga formation. Although these beds may actually be a part of the overlying *Turitella ocoyana* zone of southern California, they are herein tentatively considered to be older and equivalent to some part of the Sespe and Vaqueros formations, the latter corresponding to the *Turritella inezana* zone. It is doubtful whether rocks of Sespe age are actually present. In the adjoining area to the west, however,[23] as well as elsewhere in southern California, rocks of red color underlying fossiliferous lower or middle Miocene strata have been mapped as Sespe. The writer does not believe that red color at about this stratigraphic position warrants correlation with the red Sespe formation at its type locality, some 20 miles north-northwest of the eastern part of the Santa Monica Mountains. In a transgressing or receding basin of deposition red color resulting from the same conditions may range through a considerable thickness. However, inasmuch as such red rocks commonly constitute a mappable unit and must be named, they may, in the absence of diagnostic fossils, be considered of questionable Sespe age and mapped as Sespe(?).

The total thickness of these rocks in the area south of the Topanga fault (see pl. 16) is approximately 4,000 feet but appears to vary between 3,500 and 4,500 feet.

[23] See U. S. Geol. Survey Bull. 753, pl. 1, 1924.

Here the series is locally divisible into two members—a lower one consisting of about 2,500 feet of soft light-gray to white conglomeratic and granitic sandstone, and an upper one consisting of about 1,200 feet of equally soft conglomeratic sandstone which is distinctly red in greater part. The abrupt change in color between these two members is not, however, a consistent feature confined to one stratigraphic horizon; beds of both red and gray color pass laterally into each other, so that within a few thousand feet along the strike red conglomerates occur much lower in the section and most of the upper member is light gray.

South of the Topanga fault the contacts of these beds with adjoining Cretaceous and Eocene rocks are everywhere faulted, so that the depositional relation with the older rocks can not be determined. Two miles farther north, however, 1,800 to 2,000 feet of these beds conformably underlie the Topanga formation and overlie the Martinez formation with what appears to be considerable discordance in dip and strike. Their actual contact with the underlying Martinez shale was not observed, however, but attitudes of beds are very uniform throughout each individual formation, and the characteristic attitudes of one formation differ from those of the other by 30° in strike and 20° in amount of dip. (See pl. 16.)

Rocks similar to those described above but entirely of red color crop out in Topanga Canyon, where they are intruded by basalt and are in fault contact with the Topanga formation and with dark-gray shale of probable Chico age. (See pl. 30, *C*.) Along the southern edge of the mountains, just east of lower Topanga Canyon, these same soft red and gray conglomerates have a similar relation to overlying Topanga sandstone and are faulted against massive conglomerate of the Chico formation.

In the eastern part of the mountains, east of Stone Canyon, there is a variable thickness of unfossiliferous soft coarse gray conglomerate which rests directly on the granite and conformably beneath fossiliferous beds of the Topanga formation. Because of the absence of fossil evidence as to the age of this conglomerate and the absence of much red color to support the possible contention that it is equivalent to some part of the Sespe and Vaqueros formations, this conglomerate is herein described as a lower member of the Topanga formation.[24]

MIOCENE ROCKS

TOPANGA FORMATION (MIDDLE MIOCENE)

The Topanga formation, although far from uniform in detail features, is as a whole of fairly consistent character throughout its extent in the eastern part of the Santa Monica Mountains. The formation consists essentially of a thick, steeply dipping series of sandstone, conglomerate, and shale, together with a large amount of intrusive and extrusive basalt of Topanga age. (See pp. 95–96.) In most areas massive conglomeratic sandstone occurs in the lower part of the formation, associated with the basalt; this is generally overlain by a considerable thickness of thin-bedded shale and sandstone, locally intercalated with more massive beds of sandstone. In some areas the highest exposed part of the formation, stratigraphically above the thin-bedded shale and sandstone, is characterized by another series of massive sandstone with shale.

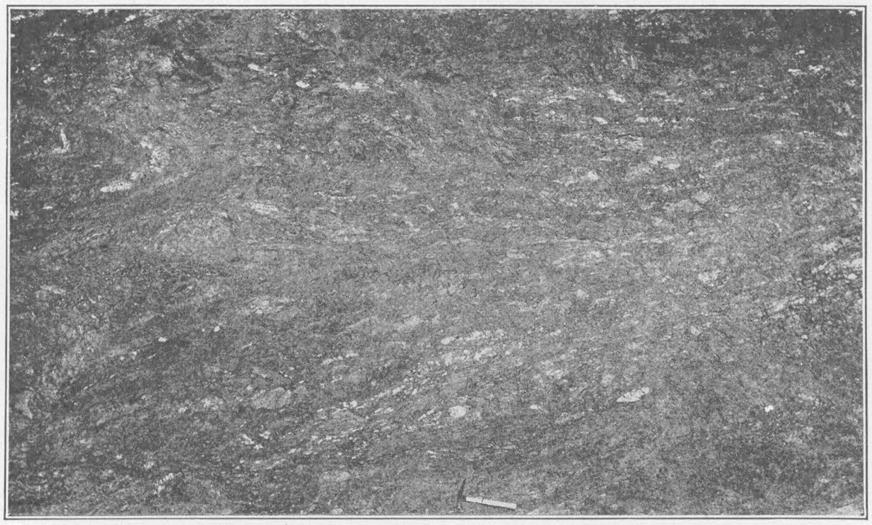
AREA BETWEEN STONE CANYON AND CAHUENGA FAULT

The Topanga formation is the most extensive largely sedimentary formation in the easternmost part of the Santa Monica Mountains, north of Hollywood and Beverly Hills. Much of the lower part of the formation, composed of massive beds of sandstone and basalt, is well exposed along Cahuenga Avenue, just south of Cahuenga Pass, and outcrops of practically the entire formation may be studied along a series of new street cuts east and west of the pass. In the area between Stone Canyon, north of Sawtelle, and the Cahuenga fault (see pl. 16) rocks herein described as belonging to the Topanga formation rest directly upon the Mesozoic granitic and metamorphic rocks that form the core of this part of the range. Freshly cut exposures of this contact at several places, together with its generally irregular character, definitely show that it is depositional throughout much of its extent. As is shown on the geologic map, however, the contact in most of the area between Brown Canyon and Laurel Canyon is faulted.

The Topanga formation in this area is divisible into three rather distinct members—a lower one that consists essentially of soft coarse gray conglomerate and loose granitic sandstone; a middle one composed of massive, more indurated beds of brown conglomerate and fossiliferous sandstone associated with minor amounts of gray shale and a thick body of basalt of both intrusive and extrusive origin; and an upper, less resistant member that consists essentially of soft thin-bedded shale and sandstone but contains a few massive ledges of brown conglomerate sandstone. These subdivisions are particularly distinct in that part of the area crossed by structure section line I–I′ (see pl. 17), and their distribution is indicated on the geologic map.

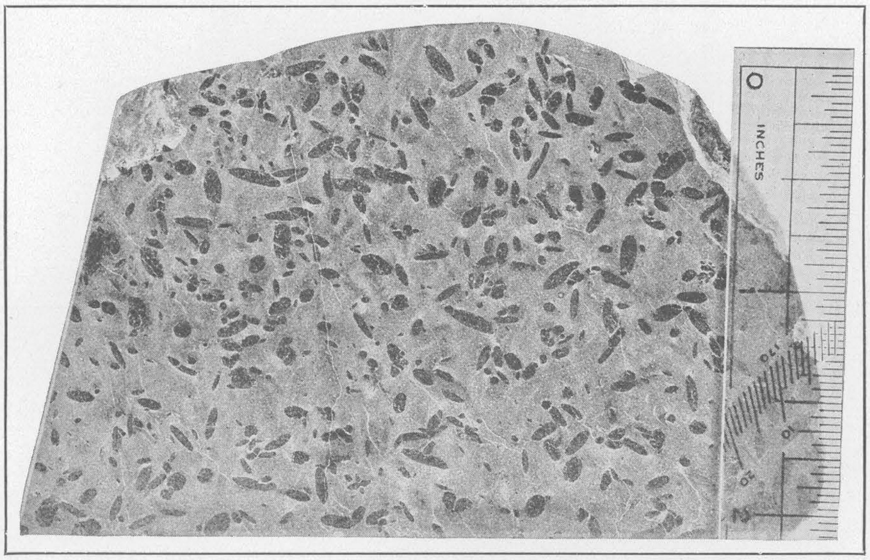
Lower conglomerate member.—The lower member, of coarse detrital character, was derived in part from the granitic and metamorphic core of the range and was deposited upon its irregular surface. Largely because of the preexisting irregularities of this old surface this conglomerate member varies considerably in thickness; east of Nichols Canyon and west of Laurel Canyon it is as much as 1,500 feet thick, but at an intervening

[24] Mr. S. A. Crouch, of the University of Southern California, has reported to the writer the finding of a perfect pelecypod cast near the head of the first canyon east of Nichols Canyon, which Dr. A. J. Tieje has identified as *Pecten andersoni*. This meager fossil evidence is indicative of middle Miocene (Topanga) age.



A. DISTORTED SCHIST, SLATE, AND PHYLLITE OF SANTA MONICA SLATE EAST OF COLDWATER CANYON, CALIF.

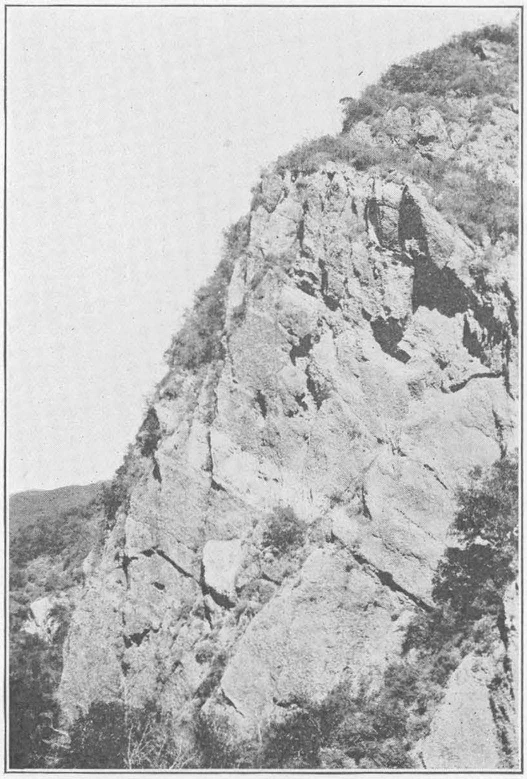
Shows the numerous veins of white quartz.



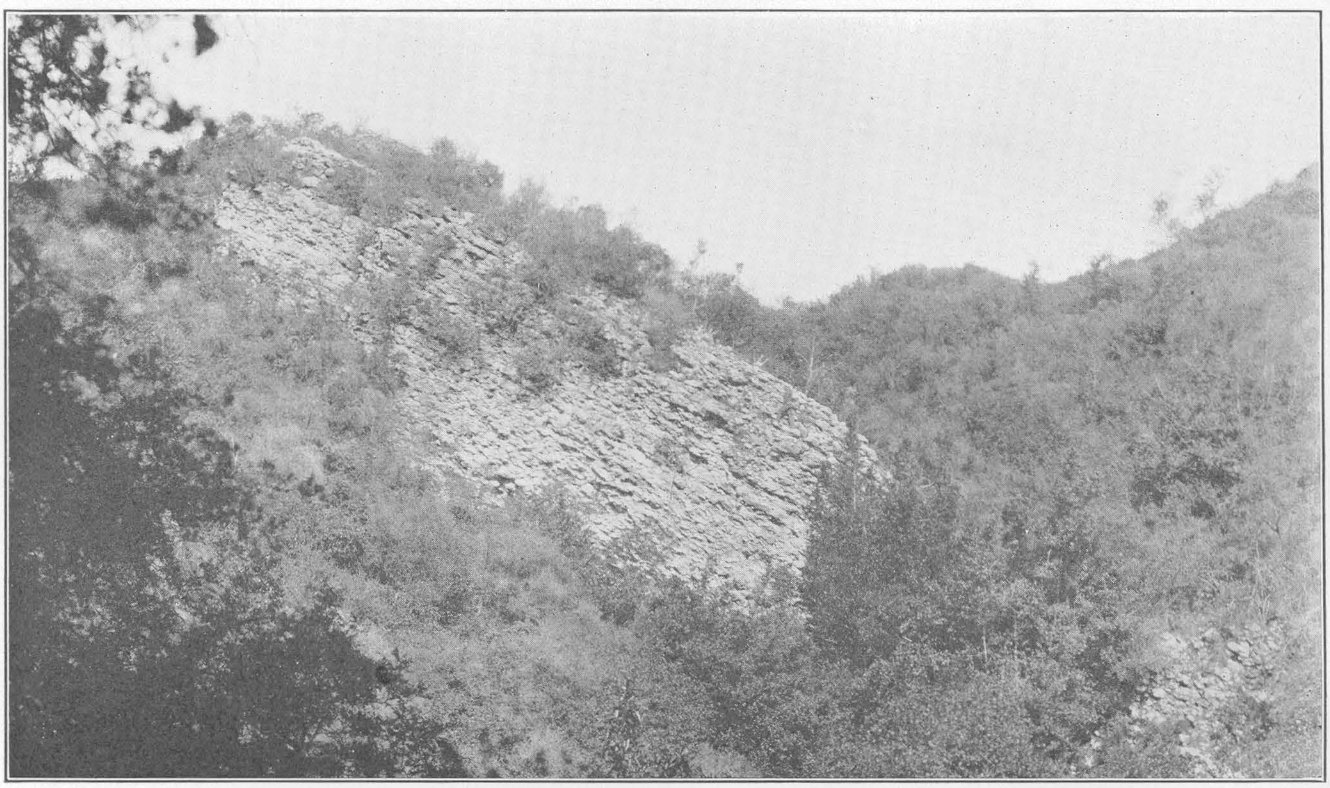
B. HAND SPECIMEN OF SPOTTED SLATE FROM SEPULVEDA CANYON, CALIF.

Photograph by K. E. Lohman.

C. THIN-BEDDED HARD BLACK SHALE AND GRAY SANDSTONE OF CHICO FORMATION IN TOPANGA CANYON, CALIF.

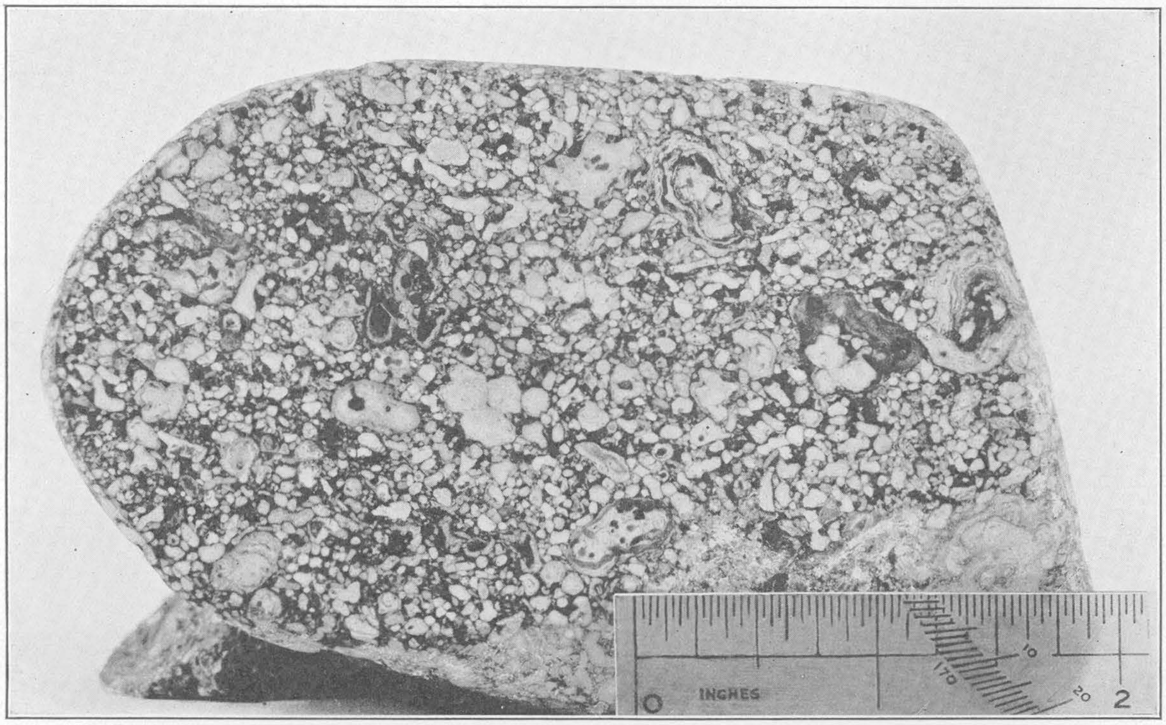


A. MASSIVE CONGLOMERATE OF CHICO FORMATION IN TOPANGA CANYON, CALIF.



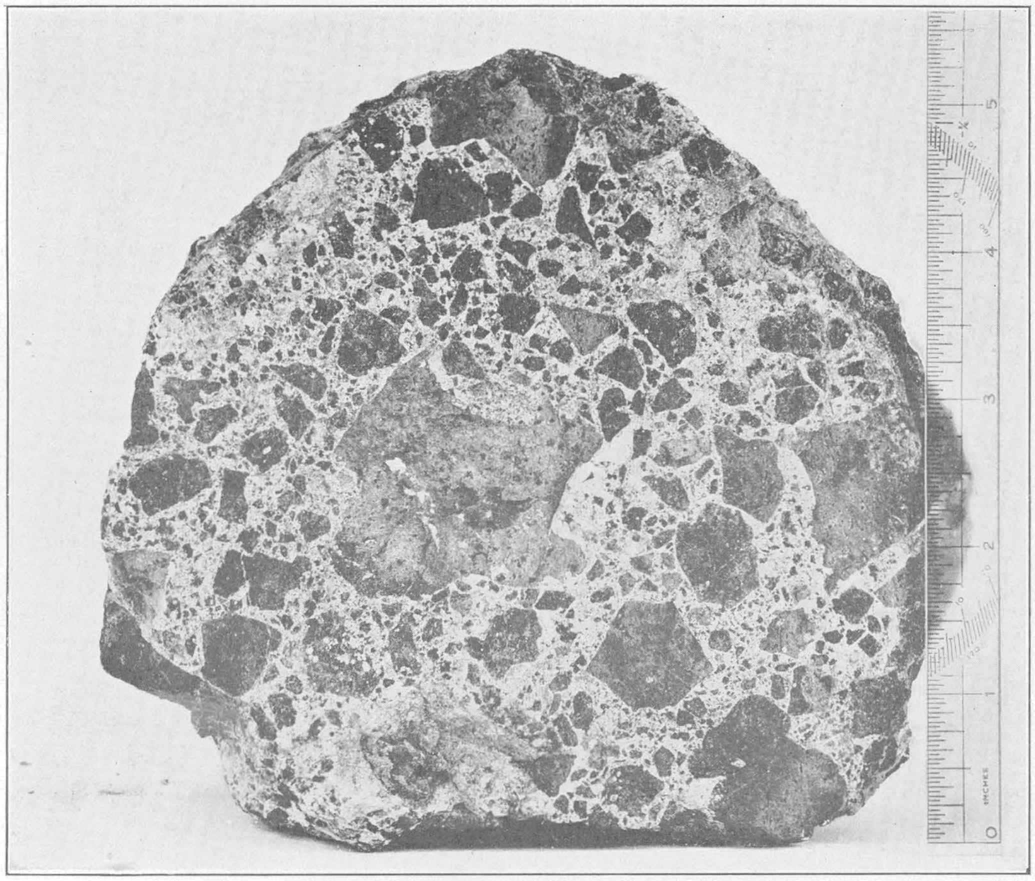
B. LEDGE OF WHITE ALGAL LIMESTONE OF MARTINEZ AGE

Limestone is about 40 feet thick. Photograph reproduced by courtesy of J. H. Gilliland, Los Angeles Mountain Park Co.



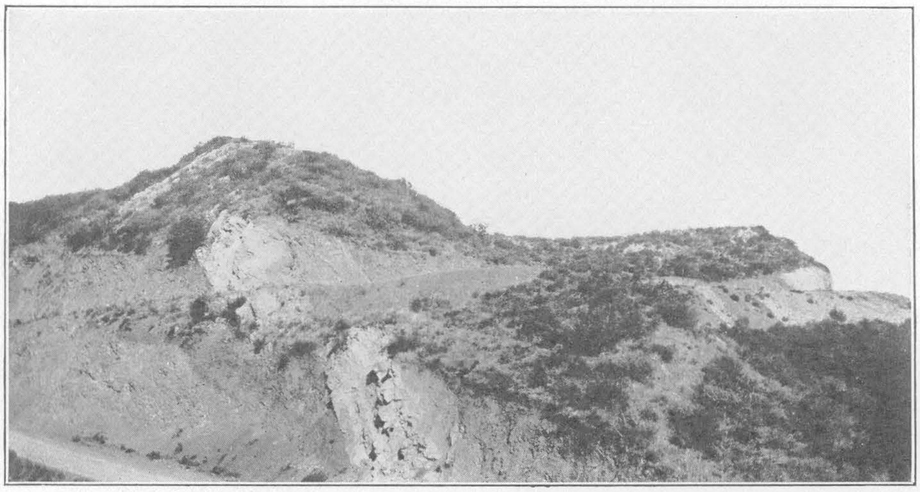
A. HAND SPECIMEN OF LIGHT-GRAY ALGAL LIMESTONE OF MARTINEZ AGE, FROM LARGE REEF AT LIMESTONE QUARRY IN UPPER SANTA YNEZ CANYON, CALIF.

Photograph by K. E. Lohman



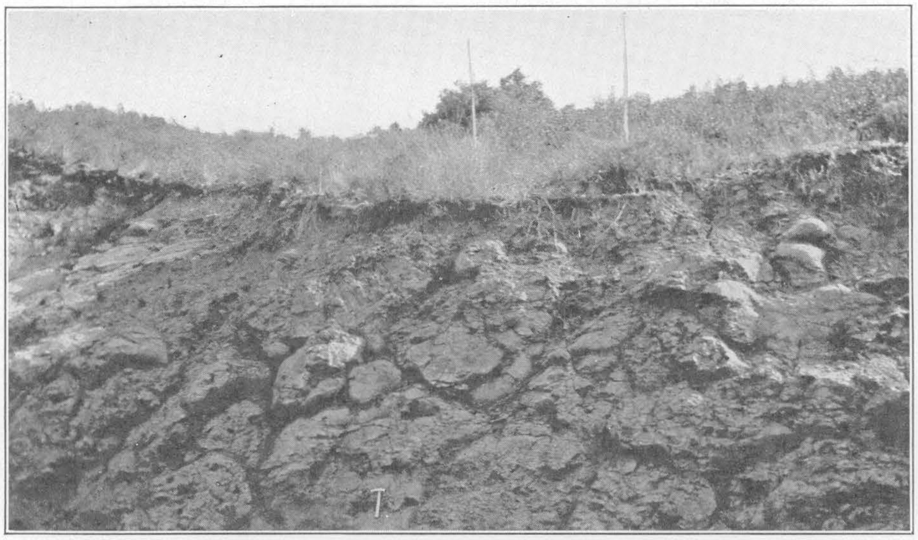
B. HAND SPECIMEN OF VOLCANIC AGGLOMERATE FROM TOPANGA FORMATION AT INTERSECTION OF MULHOLLAND HIGHWAY WITH LAUREL CANYON ROAD

Photograph by K. E. Lohman.

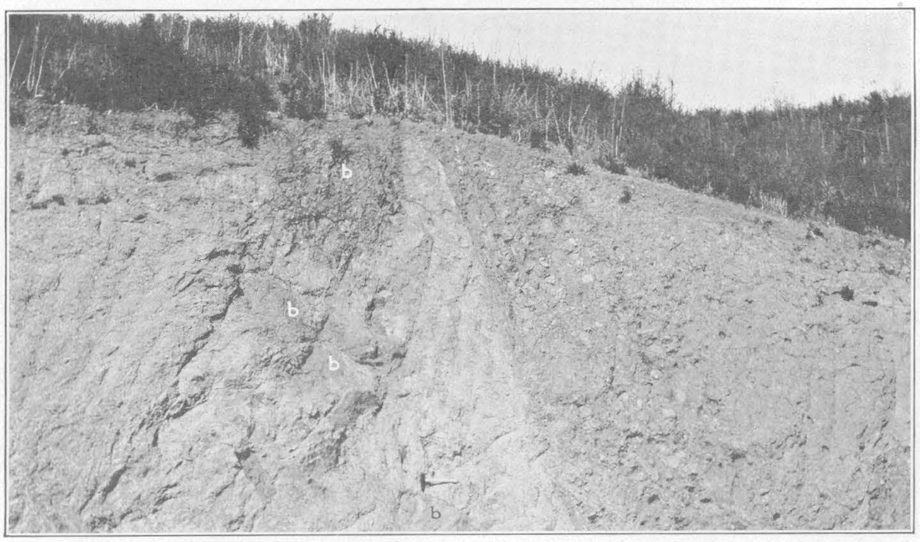


A. INTRUSIVE BASALT WITH SANDSTONE STRINGERS OF TOPANGA FORMATION

View along Mulholland Highway just west of Cahuenga Pass, Calif.



B. PILLOW STRUCTURE IN UPPER PART OF BASALT EAST OF LAUREL CANYON ROAD



C. CONGLOMERATE OF TOPANGA FORMATION

Cut near intersection of Mulholland Highway and Woodrow Wilson Drive. Shows large weathered boulders of basalt (b).

point at the head of Laurel Canyon it thins to about 250 feet. Except for the local occurrence of a basal red conglomerate which is generally only a few feet thick but which in the depression in the granite surface east of Nichols Canyon is about 200 feet thick, this member is fairly uniform throughout and consists of a mass of soft light-colored granitic and arkosic sandstone with abundant embedded cobbles and boulders as much as 1 foot in diameter. These cobbles and boulders are commonly very well rounded and have a strikingly smooth and highly polished surface. For the most part they are very hard and consist in about equal proportions of granite and gneissic granite, hard dense porphyritic basalt and andesite, and clear varicolored quartzite of fine uniform texture. The matrix of the conglomerate is similar to the associated sandstone, and, though clean and free from clayey material, is composed of angular and subangular mineral and rock fragments poorly sorted as to size. Bedding throughout this member is commonly lenticular and indistinct.

Identified heavy minerals associated with the abundant quartz and orthoclase and plagioclase feldspar in samples examined are, in order of abundance, leucoxene ilmenite, epidote, garnet, titanite, zircon, glauconite, and tourmaline. Amphiboles appear to be absent. This mineralogic character is in sharp contrast to that of the San Onofre breccia (middle Miocene) of coastal areas 20 to 40 miles farther west [25] and near San Onofre, 70 miles to the south.[26]

A prominent bed of coarse white granitic sandstone and conglomerate marks the top of this member and can be traced through practically the entire extent of this area. (See footnote 24, p. 94.)

Middle conglomeratic sandstone and basalt member.—Along structure section line I–I′ (pls. 16, 17) the middle member of massive sedimentary and basic igneous rocks has a total thickness of 3,000 feet, of which about 1,600 feet is massive brown conglomeratic sandstone with a minor amount of thin-bedded sandstone and shale. From 100 to 1,000 feet of this massive brown conglomeratic sandstone, which is locally fossiliferous, occurs at the base of this member and is overlain by a prominent elongate body of basalt which is far from uniform in thickness and which lenses out toward the west, near the head of Franklin Canyon. Basalt is to be found in small areas and narrow bands still farther west at about this same stratigraphic horizon, but its occurrence is very sporadic. Just east of the head of Nichols Canyon the basalt is overlain by about 1,300 feet of massive coarse sandstone and conglomerate, together with a subordinate amount of thin-bedded sandstone and shale. The upper part of this largely massive unit grades westward into finer and thinner bedded shale and sandstone, so that the contact between this middle member and the overlying less massive upper member (represented on the geologic map by the discontinuous dashed line) is difficult to draw in the more western area, south and southwest of Universal City. A few hundred feet above the basalt, both east and west of the Laurel Canyon road, massive conglomerate sandstone is associated with thin discontinuous beds of hard white opal shale, much like that which is so common in the lower part of the overlying Modelo formation.

One of the most significant features of the conglomeratic portion of this middle member is the striking difference between the boulders present in beds below the basalt and those characteristic of much of the 500 to 700 feet of beds which directly overlie the basalt. In the lower beds the cobbles and boulders are generally well rounded and, like those of the underlying conglomerate, consist essentially of granite, granodiorite, dense andesite and basalt porphyry, and quartzites of various color. In the beds above the basalt, the most striking feature is the abundance of deeply weathered subangular boulders of basalt as large as 4 feet across (see pl. 21, *C*); in outcrops where only 10 feet of a bed is exposed it is not uncommon to find six or eight basalt boulders that average 1 foot in diameter. The basalt composing these soft boulders is in striking contrast to the dense porphyry of the hard, highly polished boulders in beds stratigraphically lower but is identical with that of the large basalt mass directly below. Inasmuch as such boulders do not occur at any point below the basalt, their abundance in beds that directly overlie it is accepted as very good evidence that this large basalt mass in the lower part of the Topanga formation underwent erosion in some part of the mountains directly after it was formed and was the source for much of the sediment in directly overlying beds. This formation is easily accessible in road cuts near the head of Nichols Canyon and farther west where the Mulholland Highway crosses beds that directly overlie the basalt. Corroborative evidence of this relation is found in the fact that in suites of sandstone samples collected through the Topanga formation, from the base upward, green hornblende makes its first appearance in samples collected above the basalt, at the lowest horizon where basalt boulders are present.

The geologic map clearly illustrates the fact that virtually all of the basalt in the area between Stone Canyon and the Cahuenga fault is confined to one general stratigraphic position in the lower part of the Topanga formation. This relation suggests that the major part of the basalt may be extrusive and of lower Topanga age, having accumulated before, not after, the upper part of the Topanga formation. As is discussed below, this appears to be true in part for much of the basalt, particularly in the area west of the long diagonal east-northeast fault that cuts across

[25] Woodford, A. O., and Bailey, T. L., Northwestern continuation of the San Onofre breccia: California Univ. Dept. Geol. Sci. Bull., vol. 17, p. 190, 1928.

[26] Woodford, A. O., The San Onofre breccia, its nature and origin: Idem, vol. 15, pp. 159–280, 1925

Cahuenga Pass and sec. 33, T. 1 N., R. 14 W. East of this fault the tonguelike projections in the upper part of the basalt body and the presence of contact-metamorphic rock along the base of overlying sandstone prove conclusively that the upper portion of the basalt in this more eastern area is intrusive. It is believed that this upper part of the basalt east of the fault is the result of a distinctly later period of igneous activity in late Topanga or even more recent geologic time, and that this later basalt intrusion may well have caused in this area pronounced increase in the fault displacements of beds that overlie the basalt. (See on pl. 16 the relative displacements of the sandstone stringer in the midst of the basalt mass and the dashed line representing the base of the overlying upper member of the Topanga formation.) The same result could, of course, have been produced by rotational movement along this vertical fault, but the close relationship between major pre-Modelo structural deformation and basalt intrusions in the Topanga-Garrapata Canyon area, farther west, is established and may apply as well for the easternmost part of the mountains.

The lower part of the basalt is well exposed on the road just south of the Mulholland Highway in the center of the north edge of sec. 4, T. 1 S., R. 14 W., and farther west at the intersection of the Mulholland Highway with the Laurel Canyon road. At both of these localities the lower part of the basalt is distinctly pyroclastic and consists almost entirely of very angular fragments of basic volcanic rock of varying color and texture embedded in a light-gray carbonate matrix. (See pl. 20, *B*.) Most of the rock fragments are dense and finely crystalline, though some are vesicular, and others of distinctly glassy character are common. At the intersection of the Mulholland Highway with the Laurel Canyon road this pyroclastic character is associated with pronounced bedding in this extrusive rock. These characteristics are in accord with the absence of baking along the top of the underlying shale.

The definitely pyroclastic lower portion of the basalt is overlain by a nearly black dense rock which in fresh cuts shows a mass of angular blocks and gives the impression of having once been a solid rock of uniform character that is now thoroughly shattered. The angular blocks protrude from the surface of the outcrops and are separated by soft, much weathered interstitial substance, apparently of the same basaltic composition. This portion of the basalt appears to be a typical autoclastic rock which has assumed its fragmental nature subsequent to igneous activity, possibly as a result of the combined effects of cooling and structural deformation.

Much of the upper half of the basalt has a striking pillow structure (see pl. 21, *B*); other parts are vesicular and amygdaloidal, with amygdules of calcite and the zeolites. Some amygdules from basalt on the Mulholland Highway near Cahuenga Pass have been found to be of the mineral thomsonite.[33]

Locally the uppermost part of the basaltic rock contains casts of fossil shells and is of sedimentary origin, having resulted from the reworking of the upper part of the underlying basalt. Observed occurrences of this sort are found west of the diagonal fault across sec. 33, T. 1 N., R. 14 W., in the area where the base of the overlying sandstone series is not baked and where the lower few hundred feet of this overlying marine series contains abundant basalt boulders eroded from some part of the basalt mass just described.

By way of summary it may be said that much of the basalt in the area between Stone Canyon and the Cahuenga fault is extrusive and of lower Topanga age, but that some of it is definitely intrusive and considerably younger, possibly of upper Topanga age. That part of the major basalt mass which is unquestionably intrusive lies east of the diagonal fault in sec. 33 and forms the upper third of this mass. Farther west the major mass contains extrusive rock in its lower portion and has relations with overlying strata which strongly suggest that its upper part is also extrusive. Some features, however, particularly the method by which this mass terminates to the west, indicate that part of it is intrusive. It seems certain that the mass is in part extrusive and in part intrusive and that it was uplifted and subjected to erosion shortly after it was formed.

Upper thin-bedded shale and sandstone member.—The middle member of massive conglomeratic sandstone and basalt of the Topanga formation is overlain by an upper member which, along the line of structure section I–I′ (pls. 16, 17), has an exposed thickness of 2,800 to 3,000 feet. This member is well exposed in street cuts southwest and northeast of Cahuenga Avenue and is composed, to a large extent, of soft thin-bedded brown and bluish-gray shale and fine-grained sandstone, although massive beds of soft brown and gray sandstone and conglomeratic sandstone also occur in the middle and upper parts, northeast of Cahuenga Avenue. The lower 1,500 feet is made up almost entirely of thin-bedded shale and sandstone and in much of the area is in striking contrast to the massive and more resistant conglomeratic sandstone of the underlying member. The contact between these two members is represented by the discontinuous dashed line shown on the geologic map. The lower part of the upper member is broken by the diagonal fault that cuts across Cahuenga Pass, and the upper part is folded into a rather sharp syncline along and

[33] James Gilluly, of the Geological Survey, materially assisted the writer in the determination of this mineral, which, in this locality, has higher indices of refraction than those given in reference tables and textbooks, α and β both being 1.52+. For optical properties of thomsonite see Wherry, E. T., Am. Mineralogist, vol. 8, No. 7, p. 124, 1923. For other petrographic characteristics of the basalt see Schürmann, H. M. E., Beitrag zur Petrographie der Hollywood Hills (Santa Monica-Gebirge) bei Los Angeles: Centralbl. Mineralogie, 1928, Abt. A, pp. 7–13.

west of Dark Canyon and, farther east along the Cahuenga fault, abuts abruptly against massive sandstone and conglomerate of the middle member of the Topanga formation.

A single small intrusive mass of basalt occurs in the upper part of this member and crops out on one of the hilltops southeast of Universal City.

HOLLYWOODLAND-GRIFFITH PARK AREA EAST OF CAHUENGA FAULT

Because of numerous structural complications the area east of the Cahuenga fault is the most difficult area in the eastern part of the Santa Monica Mountains to interpret correctly. Irregularly distributed bodies of basalt are intruded into and faulted against granite at numerous places, and both the basalt and the granite are in turn in fault contact with sedimentary rocks that belong at various stratigraphic horizons from the bottom to the top of the Topanga formation.

Except for the widespread occurrence of Jurassic (?) granite and a single down-faulted area of the Modelo formation at the east end of the mountains, all the rocks east of the Cahuenga fault are of Topanga age. The only part of this area in which a continuous and fairly complete section of much of the Topanga formation is exposed occupies the irregular fault block due east of Cahuenga Peak and directly south of the granite mass that forms the northeastern point of the mountains. Good outcrops of much of this section may be examined along the Griffith Park Road leading north from the crest, but the ridge tops east and west of the road offer the best exposures of the massive conglomeratic sandstone in the lower part of the section. Although this section is not so thick and is without a distinct lower conglomerate member, it is much like the above-described section west of the Cahuenga fault, both as to character and as to sequence of formational members. The lower member, which corresponds to the middle member of the area farther west, is characterized by massive beds of brown and gray conglomeratic sandstone which locally contain marine fossils and are associated with basalt of both intrusive and extrusive origin.

The intrusive character of some of the basalt is evident from the presence of dikelike masses that cut across beds of sandstone. The extrusive facies of the basalt is well exposed in the upper part of the westernmost and stratigraphically highest basalt mass east of Cahuenga Peak. This mass contains about 150 feet below its top a prominent resistant ledge of brick-red color which is overlain by soft basaltic material with distinct bedding that appears to be of detrital character and to have been in part water-laid. The water-laid beds, which are commonly only a few inches thick, are associated with equally thin zones that have characteristics of flow breccias, being composed of alternating beds of dense spheroidal basalt and basalt breccia. Material of this same type occurs in the upper part of the basalt exposed along the Griffith Park Road 2,500 feet due south of the summit and at another point on the same road due east of the basalt quarry in Brush Canyon. At the former locality the overlying conglomeratic sandstone, as would be expected, is not metamorphosed and contains innumerable small fragments of basalt.

North of the summit in the fault block east of Cahuenga Peak the basalt and massive conglomeratic sandstone are overlain by a few hundred feet of soft coarse conglomerate with abundant boulders of granite, quartzite, and basalt, the basalt having been eroded from the underlying basalt of lower Topanga age. The contact of this conglomerate with the overlying thin-bedded brown and bluish-gray shale and sandstone, in this area as well as along the southern edge of the mountain, is shown on the geologic map by a dashed line. This upper member, which is very similar to the upper member of the Topanga formation west of the Cahuenga fault, is flexed into a series of sharp folds and, together with the underlying conglomerate, is faulted against the granite mass that forms the northeastern part of the hills.

The Cahuenga Peak fault block is composed of a northward-dipping series of poorly consolidated brown and gray conglomerate which grades upward into massive, fairly well cemented beds of coarse brown sandstone. The lower strikingly conglomeratic part of this series crops out on the south slope of Cahuenga Peak and in street cuts of Hollywoodland west of Brush Canyon. It is simply a mass of rounded to subangular boulders and cobbles embedded in a soft matrix of coarse arkosic sandstone. It appears to be uniform in composition throughout, the boulders and cobbles being composed of granite, granodiorite, quartzite, basic porphyry, and basalt. The boulders of basalt, which invariably are deeply weathered and very soft, are abundant at many horizons throughout this conglomeratic member and have undoubtedly been derived by subaerial erosion from the basalt series in the lower part of the Topanga formation. The presence of these abundant basalt boulders in the conglomerate of the Cahuenga Peak fault block is the only evidence found to be useful in determining the position of these rocks within the Topanga formation, but inasmuch as the occurrence of such boulders is restricted to a distinct stratigraphic zone above the basalt in other areas where more complete sections are exposed, this petrographic evidence of age is considered to be conclusive. This conglomerate is therefore regarded as younger than the basalt series of the lower part of the Topanga and equivalent, in part at least, to the similar basalt-boulder conglomerate that directly overlies the basalt series in areas east and west of Cahuenga Peak. In the Cahuenga Peak area, however, this conglomerate appears to have a thickness of

3,000 feet or more, which is in sharp contrast to the few hundred feet of this conglomerate on adjoining fault blocks east and west of Cahuenga Peak. Either the great thickness of this conglomerate south of Cahuenga Peak is more apparent than real, owing to the possible presence of unmapped strike faults that have resulted in repetitions of strata, or the Cahuenga fault block during middle Topanga time was a subsiding basin of relatively rapid deposition which, after the lower Topanga period of igneous activity, adjoined rising land areas of granite, basalt, and lower Topanga conglomeratic sandstone. The latter alternative is considered to account best for the remarkable thickness of this conglomerate. Although it is possible that faulting has increased the apparent thickness somewhat, it is unlikely that such faulting would have resulted in so notable a thickening without exposing either underlying or overlying members of the Topanga formation.

In view of the presence of poorly preserved marine fossils in the upper part of this member and the absence of clay in the clean sandstone matrix of all parts of it, this conglomerate is considered to be largely or entirely of marine origin.

Fossils collected from locality 53, between Cahuenga Peak and the Griffith Park Road, are listed on page 100.

AREA NEAR ENCINO RESERVOIR

The Encino Reservoir is on the north flank of the Santa Monica Mountains, near Encino Park, and is surrounded by coarse, poorly consolidated conglomerate and conglomeratic sandstone, which in its lower portion contains considerable poorly exposed basalt, largely of intrusive origin. The conglomerate has a maximum exposed thickness of 3,500 to 4,000 feet and is probably equivalent, in part at least, to the middle member of the Topanga formation exposed farther east, in the area between Stone Canyon and the Cahuenga fault. It is overlain unconformably by the Modelo (upper Miocene) formation, a relation which is very evident on the ridge west of the Encino Reservoir and which is well exposed in a cut along an old abandoned road 2,700 feet east of the east end of the dam. The discordance in dip of the two formations in actual contact at this road cut is 20°, beds of the Topanga formation dipping the more steeply.

The conglomerate of the Topanga formation in this area is brown and gray and contains cobbles and boulders that average 3 to 5 inches in diameter but attain a maximum size of more than 12 inches. Boulders of light-gray granite are not common and are associated with others of quartzite and varicolored porphyry, together with subangular chunks of gray shale as much as 15 inches across which have probably been eroded from the underlying Martinez or Chico formations.

Fossils collected from localities 13 and 14, east of the Encino Reservoir, are listed in the table on page 100.

TOPANGA-GARRAPATA CANYON AREA

All rocks of pre-Modelo age in the vicinity of Topanga and Garrapata Canyons, along the western edge of the area covered by this report, are steeply tilted, broken by many faults, and intruded by basalt and diabase. For the most part, rocks of the Topanga formation in this area lie north of the Topanga fault and west of the Santa Ynez fault (see pl. 16) and dip in a westerly direction away from the uplifted anticlinical area of old slate farther east.

Only a partial section of the entire formation is exposed in this area, but it includes about 3,000 feet of massive brown sandstone, conglomerate, and gray shale. Although the presence of basalt intrusions makes it difficult even to estimate normal sedimentary rock thicknesses, there appears to be 1,250 to 1,500 feet of massive hard marine conglomeratic sandstone, which occupies the lower part of the formation and rests conformably upon much softer red and gray conglomerate of the Sespe (?) and Vaqueros (?) formations. These massive lower Topanga beds contain marine fossils and have been intruded by large bodies of basic magma, now exposed as light-gray to black diabase and coarsely crystalline basalt. Directly above this lower sandstone member is several hundred feet, possibly as much as 1,000 feet, of dark-gray shale which contains scattered yellow-weathering limestone concretions and beds of soft gray sandstone of fine texture. Beds of sandstone become more abundant and are thicker and coarser toward the top of the formation; the shale member grades upward into a higher arenaceous member which is exposed along much of the Garrapata Canyon road south of Mohn Springs and consists for the most part of soft thin-bedded to massive sandstone and conglomeratic sandstone with partings and thin beds of soft gray shale.

South of Topanga post office and the Topanga fault only the lower massive conglomeratic sandstone member of the Topanga formation is exposed.

At least a part of the basalt intruded into the Topanga formation of this area appears to be younger than the basalt of the lower Topanga near Cahuenga Pass and is probably equivalent in age to the intrusive late Topanga basalt of the Cahuenga Pass area. Irregular intrusive bodies occur through practically all the exposed 3,000 feet of the Topanga formation, many of them having a distribution which apparently has been controlled by post-Topanga faulting. It seems fairly certain that at least a part of the basalt in the Topanga-Garrapata Canyon area was intruded near the end of Topanga time, probably during the period of pronounced pre-Modelo disturbance, and that faulting and basalt intrusion had a close genetic relation. (See pl. 16.) Elongate basalt bodies and at least one major fault pass beneath, not into, the base of the Modelo formation near Mohn Springs, although post-Modelo uplift of one basalt mass, which may occupy a fault zone now obscured by the basalt, apparently has been the cause of a small anticline in the overlying Modelo.

The unconformable relation between the Topanga and Modelo formations is one of the most remarkable geologic features of this area. It is best observed at fossil locality 20, on the ridge near the east edge of sec. 32, T. 1 N., R. 16 W., and on the Garrapata Canyon road at Mohn Springs Cafe. (See pl. 22, *B*.)

Fossils collected at localities 25, 28, 32, 33, 34, 56, and 63 in the Topanga formation of this area are listed in the table on page 100.

That part of the Topanga formation exposed farther south along the coast consists almost entirely of massive brown conglomeratic sandstone which overlies red conglomerate and sandstone of the Sespe (?) and Vaqueros (?) formations and which is in fault contact with rocks ranging in age from Upper Cretaceous to upper Miocene. Topanga fossils have been collected at locality 42.

AREA OF LOWER SEPULVEDA CANYON

Coarse marine conglomerate, much like that which forms the lower conglomerate member of the Topanga formation along the west side of the Stone Canyon Reservoir, is present east and west of lower Sepulveda Canyon, along the south flank of the mountains. In exposures west of Sepulveda Canyon this conglomerate contains thin beds and partings of light-gray shale which resembles that present in the Miocene elsewhere in the Santa Monica Mountains. The only identifiable fossils found were collected at locality 50 and, according to Doctor Woodring, consist of several specimens of a *Tivela;* a barnacle and a fragment of a rib bone, both of no value for age determination, were found about 500 feet northwest of this locality.

As shown on the geologic map, this conglomerate locally rests directly upon the Santa Monica slate, but elsewhere its lower exposed part is in fault contact with the Modelo. East of lower Sepulveda Canyon the Topanga conglomerate is unconformably overlain by the basal white platy shale of the Modelo with an angular discordance of 20° or more.

FOSSILS AND CORRELATION

Although the sedimentary rocks of the Topanga formation are probably all of marine origin, many of the finer-grained sandstones and some of the shales are literally filled with plant fragments, and in more than one locality fragile imprints of fragments of leaves and stems of land plants have been observed. A leaf collected from locality 8, east of Nichols Canyon, and submitted to Dr. R. W. Chaney, of the Carnegie Institution, has been identified as *Salix*, of the type *S. mixta* of the middle Miocene of Oregon.

Some of the shale stringers included in the basalt have yielded a few Foraminifera. Samples have been collected at localities 101, 102, and 103, on and near the Mulholland Highway between the Laurel Canyon road and Cahuenga Avenue. (For description of localities see p. 124.)

The sample collected at locality 102 from a shale stringer in basalt on the Mulholland Highway about 2,000 feet east of the Laurel Canyon road has been examined by D. D. Hughes and Boris Laiming, of Los Angeles. According to Mr. Hughes this sample contains rather poorly preserved calcareous Foraminifera, among which the following 10 species are identifiable:

Bolivina brevior Cushman, rare.
Bolivina cf. B. conica Cushman, very rare.
Bolivina imbricata Cushman, very rare.
Bolivina sp., partly crushed and altered, common.
Nonion costifera Cushman, very common.
Quinqueloculina cf. Q. oblonga (Montagu), common.
Quinqueloculina cf. Q. seminulum (D'Orbigny), very rare.
Valvulineria californica Cushman, rare.
Valvulineria miocenica Cushman, very common.
Virgulina cf. V. californiensis Cushman, rare.

Hughes states that this fauna occurs in the Highland School district, San Luis Obispo County, and is restricted to so-called Monterey or Salinas shale beds 400 to 600 feet above the fossiliferous Temblor sandstone described by Anderson and Martin.[34] Hughes also states that this fauna also occurs in so-called Monterey shale on the 17-mile drive near the town of Monterey three-quarters of a mile toward Pebble Beach from the tollhouse. In his opinion there is little doubt, therefore, that the sample from locality 102, collected approximately 500 feet above the base and 6,000 feet below the top of the Topanga formation of the Santa Monica Mountains, is equivalent in age to siliceous shale about 600 feet above the base of the so-called Monterey shale in the Highland School area of San Luis Obispo County and to similar shale in the vicinity of the type section of the Monterey.

The Topanga formation has yielded a fairly good macroscopic fauna of marine invertebrates, which have been collected at numerous widely scattered localities. These fossils have been identified by Dr. W. P. Woodring, of the California Institute of Technology, and are listed below, with his comments. All the fossil localities are accurately shown on Plate 16 and are described on pages 123–124.

[34] Anderson, F. M., and Martin, Bruce, Neocene record of the Temblor Basin, Calif., and Neocene deposits of the San Juan district, San Luis Obispo County, Calif.: California Acad. Sci. Proc., 4th ser., vol. 4, pp. 37–44, 1914.

Fossils from the Topanga formation of the eastern part of the Santa Monica Mountains

[Localities plotted on Plate 1. For description of localities, see pp. 123-124]

	2	5	6	7	13	14	20	25	28A	28B	32	33	34	38	42	50	53	56	60	63	66
Gastropods:																					
Tegula dalli (Arnold)							×														
Astraea topangenis (Arnold)							×														
Neverita cf. N. andersoni (Clark)	×						×		×											×	
Lunatia? sp.									×												
Calyptraea costellata (Conrad)									×					×					×	×	×
Crucibulum? sp.									×												
Turritella ocoyana Conrad								×	×	×	×	×	×	×	×					×	×
Turritella temblorensis Wiedey					×		×		×					×					×		
"Cerithium" sp.				×	×				×					×							
Tritonalia edmondi (Arnold)							×													×	
Tritonalia topangensis (Arnold)									×												
Galeodes californicus (Anderson and Martin)				?					×												
Oliva californica Anderson							×		?												
Cancellaria cf. C. dalliana Anderson																				×	
Conus owenianus Anderson						?			×												
"Drillia" sp.							×														
Bulla sp.									×								×			×	
Lamellibranchs:																					
"Leda" ochsneri (Anderson and Martin)																	×				
Anadara osmonti (Dall)				×										×					?		
Atrina sp.																	×				
"Pecten" cf. "P." andersoni (Arnold)														×							×
"Pecten" nevadanus Conrad																	×	×			
Mytilus expansus Arnold							×		×				×							×	×
Echinochoma? sp.																	×		×		
Lucinisca cf. L. nuttallii (Conrad)							×							×							
Lucinoma acutilineata (Conrad)									×												
Lucina richthofeni (Gabb)							×		×											?	
Miltha sanctaecrucis (Arnold)							×		×								×				
Diplodonta? sp.	×				×		×													×	
Trachycardium schencki (Wiedey)	×						×		×	×							×				
Trachycardium sp.																	×				
Dosinia sp. a		×				×	×		×		×			×							
Dosinia sp. b			×																	×	
Clementia conradiana (Anderson)									×								×	×			
Tivela sp.																×					
Pitar? sp.							×		×											×	
Antigona sp.							×													×	
Saxidomus cf. S. vaquerosensis Arnold							×														
Chione temblorensis Anderson							×													×	
Chione sp.				×										×					×		
Paphia? sp.									×												
Tellina? sp.							×								×						
Macoma? sp.	×											×		×							
Metis sp.								×	×			×									
Sanguinolaria sp.													×								
Siliqua sp.									×										×		
Spisula cf. S. abscissa (Gabb)																	×				
Panope sp.							×		×												
Echinoid:																					
"Scutella" norrisi Pack?								×					×								

The following comments are supplied by Woodring:

Most of the fossils in the preceding list represent species already recorded from the Topanga formation at the type locality in Topanga Canyon and in adjoining areas.[27] Whitney,[28] on the basis of Gabb's determinations, recorded six species of Miocene fossils, including *Turritella ocoyana*, from "Hancock's Canyon," an undetermined locality near the east end of the Santa Monica Mountains.

Turritella ocoyana is the most widespread species in the Topanga formation. All large specimens of this species in the collections from the eastern Santa Monica Mountains have a strong keel near the lower edge of the whorl, like those for which Wiedey [29] used the name *T. bösei* Hertlein and Jordan, which is regarded as a synonym of *T. ocoyana*, for in the type region of *T. ocoyana* in the vicinity of Bakersfield specimens that are similarly keeled and specimens with rounded whorls, which resemble each other in other features, are found at the same locality. *T. temblorensis* may be another form of *T. ocoyana*.

It is apparent that *T. ocoyana* is a polymorphic species of the group of *T. terebralis* Lamarck, as defined by Guillaume.[30] It is also apparent that in southern California and Lower California a larger proportion of specimens are strongly keeled than in the Bakersfield region, and that still farther south, in tropical America, where five additional names have been proposed, this phylum of Turritellas was represented during middle Miocene time by veritable giants.[31]

The tropical genus *Galeodes*, more familiarly known as *Melongena*, has not heretofore been recorded in southern California, but its presence farther north in San Luis Obispo County [32] indicated that it would eventually be found there.

The distribution of a small "*Cerithium*" (localities 7, 13, 28A, 38), *Galeodes californicus* (localities 7?, 28A), *Anadara osmonti* (localities 7, 38, 60?), and a small *Anomalocardia*-like *Chione* (localities 7, 38, 60) may indicate a faunal zone within the Topanga formation.

TOPANGA (?) FORMATION ON SOUTH FLANK OF MOUNTAINS

Rocks of somewhat doubtful age on the south flank of the mountains, east of the Temescal fault, rest directly upon the Santa Monica slate. They are different in character from any of the rocks exposed farther west, in the district of Upper Cretaceous and Eocene strata, and have characteristics in common with parts of the Topanga formation exposed on the north flank of the mountains. They also appear to be unconformably overlain by white siliceous shale and sandstone of the lower part of the Modelo formation, a relation which exists for Topanga and Modelo rocks along lower Sepulveda Canyon, only a short distance to the east.

The Topanga (?) formation in this area, east and west of lower Mandeville Canyon, consists largely of soft gray shale and sandstone but contains a few comparatively thin beds of conglomerate which resembles that in the lower conglomeratic member of the Topanga formation along the west side of the Stone Canyon Reservoir. At the base of the formation, resting directly upon the Santa Monica slate on the ridge west of Mandeville Canyon, there is about 50 feet of red conglomerate much like that present in other areas where either the Topanga or the Chico formation has a depositional contact with the old slate.

PROBABLE MIDDLE MIOCENE TRACHYTE

Ledges and small dikes of light-gray and cream-colored finely crystalline trachyte cut Cretaceous rocks west of the head of Rustic Canyon and Cretaceous and Eocene rocks on the Coast Highway near Castle Rock. These ledges and dikes are only a few inches thick and therefore can not be shown on the geologic map. A prominent ledge of similar trachyte, in part markedly vesicular, cuts sandstone and shale of probable Topanga age in the east wall of Mandeville Canyon three-quarters of a mile above the Beverly Boulevard. The age of this trachyte is a little uncertain, but inasmuch as some of it occurs in rocks of probable Topanga age and pebbles of similar rock are abundant in the basal gray sandstone of the Modelo on the south flank of the mountains west of Rustic Canyon, some of the trachyte is pre-Modelo (earlier than upper Miocene) and probably of about the same age as much of the basalt in the Topanga formation of the Topanga Canyon district.

MODELO FORMATION (UPPER MIOCENE)

DEFINITION

The type section of the Modelo formation is rather indefinitely located but lies north of the Santa Clara Valley in the vicinity of Modelo Canyon, 20 miles northwest of the area covered by this report. Continuous outcrops do not connect the type section with the north flank of the Santa Monica Mountains, but the similarity in character and stratigraphic position and the occurrence of lithologic features uncommon to other formations of these areas afford good evidence that the upper Miocene rocks of the Santa Monica Mountains are equivalent, in part at least, to the Modelo formation in the vicinity of the type section. It may be, however, that each of these widely separated sections contains some rocks that have no age equivalent in the other. The Modelo as mapped by Kew in and north of Santa Clara Valley contains rocks older than those which comprise the Modelo in the Santa

[27] Arnold, Ralph, New and characteristic species of fossil mollusks from the oil-bearing Tertiary formations of southern California: U. S. Nat. Mus. Proc., vol. 32, pp. 525-526, pls. 40-46, 1907. Eldridge, G. H., and Arnold, Ralph, The Santa Clara Valley, Puente Hills, and Los Angeles oil districts, southern California: U. S. Geol. Survey Bull. 309, pp. 147-148, pls. 27-33, 1907. Kew, W. S. W., Geology and oil resources of a part of Los Angeles and Ventura Counties, Calif: U. S. Geol. Survey Bull. 753, pp. 50-51, 1924. Wiedey, L. W., Notes on the Vaqueros and Temblor formations of the California Miocene, with descriptions of new species: San Diego Soc. Nat. Hist. Trans., vol. 5, No. 10, pp. 95-182, pls. 9-21, 1928.

[28] Whitney, J. D., California Geol. Survey, Geology, vol 1, p. 171, 1865.

[29] Wiedey, L. W., op. cit., pp. 117-119, pl. 10, fig. 7, pl. 11, figs. 1-3, 5-6.

[30] Guillaume, Louis, Essai sur la classification des Turritelles, ainsi que sur leur évolution et leurs migrations, depuis le début des temps tertiaires: Soc. géol. France Bull., 4th ser., vol. 24, pp. 281-311, pls. 10-11, 33 figs., 1924.

[31] Woodring, W. P., Miocene mollusks from Bowden, Jamaica, pt. 2: Carnegie Inst. Washington Pub. 385 pp. 95-98, 1928.

[32] Anderson, F. M., and Martin, Bruce, Neocene record in the Temblor Basin and Neocene deposits of the San Juan district, San Luis Obispo County, Calif.: California Acad. Sci. Proc., 4th ser., vol. 4, p. 80, pl. 4, fig. 1, 1914.

Monica Mountains.[35] At places in the Santa Clara Valley district, according to Kew,[36] the Modelo appears to rest conformably upon the Vaqueros formation (lower Miocene), with absence of any intervening strata recognized as representing the Topanga formation (middle Miocene). Kew [37] is now of the opinion, however, that this Modelo of the type district contains, in its lower part, beds of Topanga (middle Miocene) age.

Because of the striking unconformity in the Santa Monica Mountains between the Topanga and the Modelo, it is a simple task to separate the two formations, and this necessary separation has resulted in a restricted use of the name Modelo. The Modelo of this report is therefore limited to beds most if not all of which are of upper Miocene age, a usage which is in conformity with that intended and approved by the Geological Survey for the Santa Clara Valley district mapped and described by Kew in Bulletin 753. In the area described in the present report the Modelo formation consists of rocks that are all definitely younger than the Topanga formation and its correlative, the "Temblor formation" of San Joaquin Valley.

According to Hudson and Craig, the Modelo formation of the type area consists of a lower bituminous shale series, a lower massive sandstone, a middle siliceous and calcareous shale, an upper sandstone, and an upper cherty shale. Fossils indicative of middle Miocene age were found in the lower Modelo sandstone of the type section. A fauna indicative of uppermost Miocene (Santa Margarita) age was found in a sandstone bed in the upper shale member of the type section. At other localities, north of the type section, rocks that were formerly believed to be of Pico age but were found to carry fossils indicative of uppermost Miocene age rest with marked unconformity upon the underlying Modelo strata. Hudson and Craig conclude that the type section of the Modelo as mapped by Kew, however, represents continuous deposition from Vaqueros to the end of Miocene time and that the lower shale, the lower sandstone, and the middle shale are of Topanga age and the upper part of the upper shale is of Santa Margarita age. They suggest that the name Modelo be restricted to the unfossiliferous middle part of the Modelo, which includes the upper sandstone and the lower part of the upper shale.

DISTRIBUTION AND GENERAL CHARACTER

The Modelo formation is exposed on both the north and south flanks of the eastern part of the Santa Monica Mountains but has its widest distribution and greatest thickness on the north flank, west of Universal City. The greatest exposed thickness of the formation is 4,500 feet and is found at the northwestern edge of the area along structure section line A–A′ pls. 16, 17) and in the central part along structure section line E–E′, just west of Stone Canyon. The thickness, however, varies remarkably along the north flank, decreasing to 2,100 feet within a distance of 1½ miles east of the Stone Canyon locality and to only a few hundred feet near the Encino Reservoir, an area intermediate between and only 3 miles from the two thickest sections.

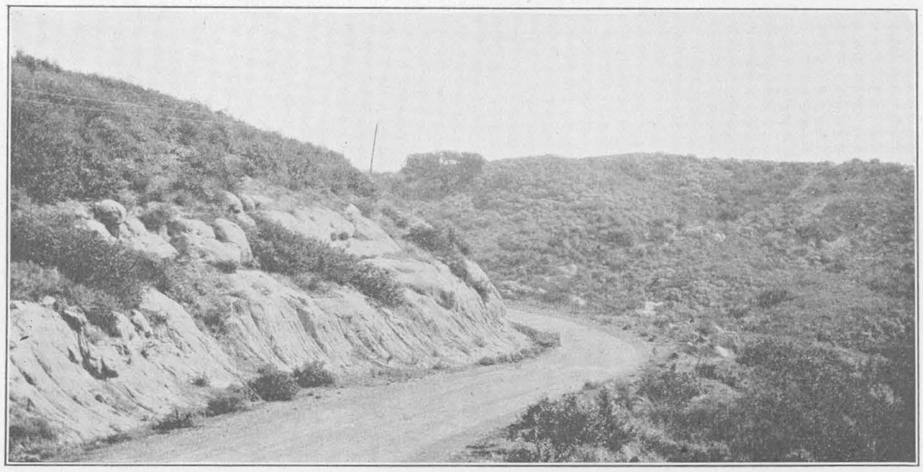
The Modelo formation overlaps every older sedimentary formation with which it is in contact with striking angular discordance. There are local exceptions—for example, at one locality in the vicinity of Will Rogers's ranch along lower Rustic Canyon, there is little or no angular discordance between the Modelo and the underlying conglomerate of the Chico formation—but on the north flank of the range Chico conglomerate passes underneath the Modelo with a strike actually normal to that common to the Modelo.

The Modelo formation is of marine origin throughout and is composed of two members which, although apparently conformable and grading into each other, are distinct in most of the district and can be separately mapped without much difficulty. A 1-inch bed containing pebbles of siliceous shale and granite embedded in a siliceous shale matrix occurs at the contact between the two members in a lower road cut a few hundred feet south of the Hollywood Country Club, 2½ miles west of Universal City. There is no discordance in attitude between the beds above and below, and no such thin conglomerate bed or other evidence of unconformity could be found in other near-by exposures. An alternation of cherty siliceous shale and punky diatomaceous shale, beds characteristic of the lower and upper members, respectively, occurs within the 2 or 3 feet of strata directly overlying the 1-inch bed of conglomerate. The writer does not believe that a noteworthy stratigraphic break occurs at this horizon. The maximum thickness of the lower member is 2,750 feet and that of the upper is 2,300 feet. In most of the area covered by this formation the exposed thickness of the lower member is greater than that of the upper and consists of units of thin-bedded shale (pl. 24, *A*, *B*) alternating with more massive units of coarse gray and brown sandstone (pl. 22, *A*), the largest of which have been individually mapped on Plate 16. Much of the shale is hard, platy, and opaline and is rich in the calcareous and siliceous remains of microscopic marine animals and plants, such as Foraminifera, Radiolaria, and diatoms. Ordinary soft earthy shale is also very common in this lower member. The upper member, throughout most of its area of exposure, consists largely of soft white punky diatomaceous shale (pl. 25, *C*) and is in striking contrast to all por-

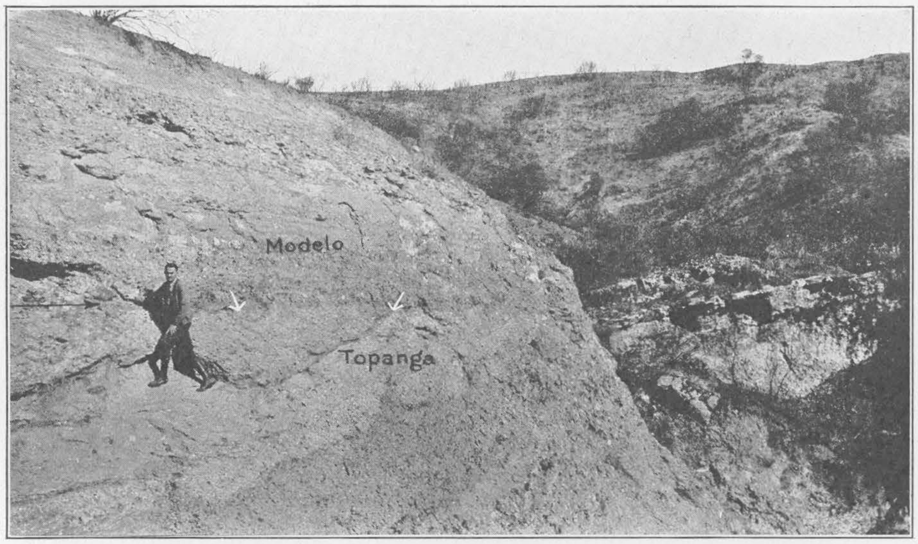
[35] Hudson, F. S., and Craig, E. K., Geologic age of the Modelo formation, California: Am. Assoc. Petroleum Geologists Bull., vol. 13, pp. 512-517, 1929.

[36] Kew, W. S. W., Geology and oil resources of a part of Los Angeles and Ventura Counties, Calif.; U. S. Geol. Survey Bull. 753, p. 58, 1924.

[37] Oral communication, March, 1928.

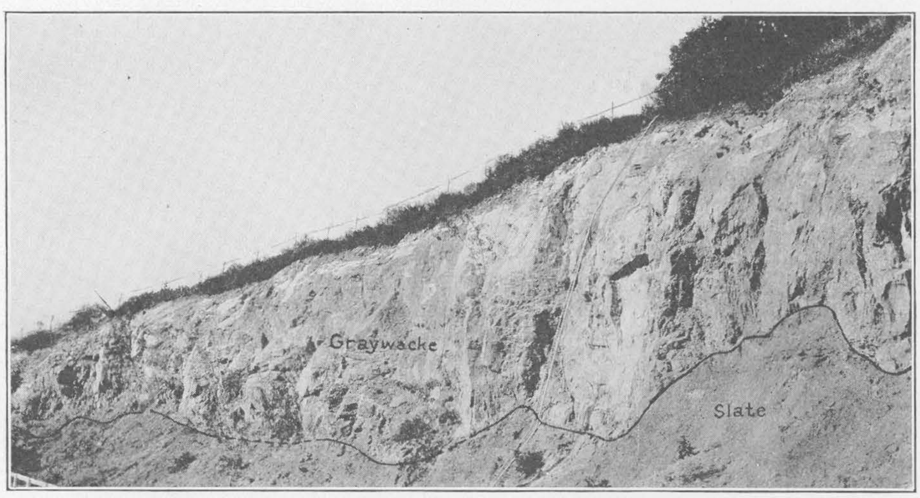


A. ONE OF THE MANY MASSIVE UNITS OF SANDSTONE IN THE LOWER MEMBER OF THE MODELO FORMATION

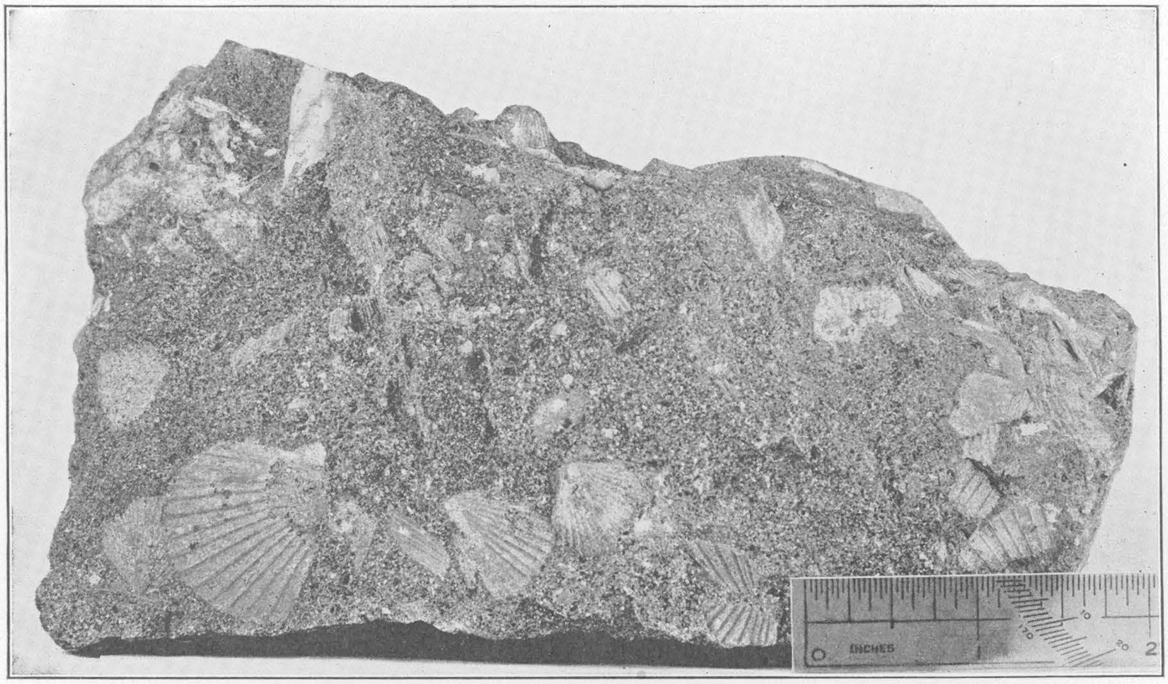


B. UNCONFORMITY BETWEEN BASAL CONGLOMERATE OF MODELO FORMATION (UPPER MIOCENE) AND CONGLOMERATE OF TOPANGA FORMATION (MIDDLE MIOCENE), NEAR MOHN SPRINGS CAFE IN GARRAPATA CANYON

Shows angular discordance; Modelo dips 10° NW., and Topanga dips 30° S.

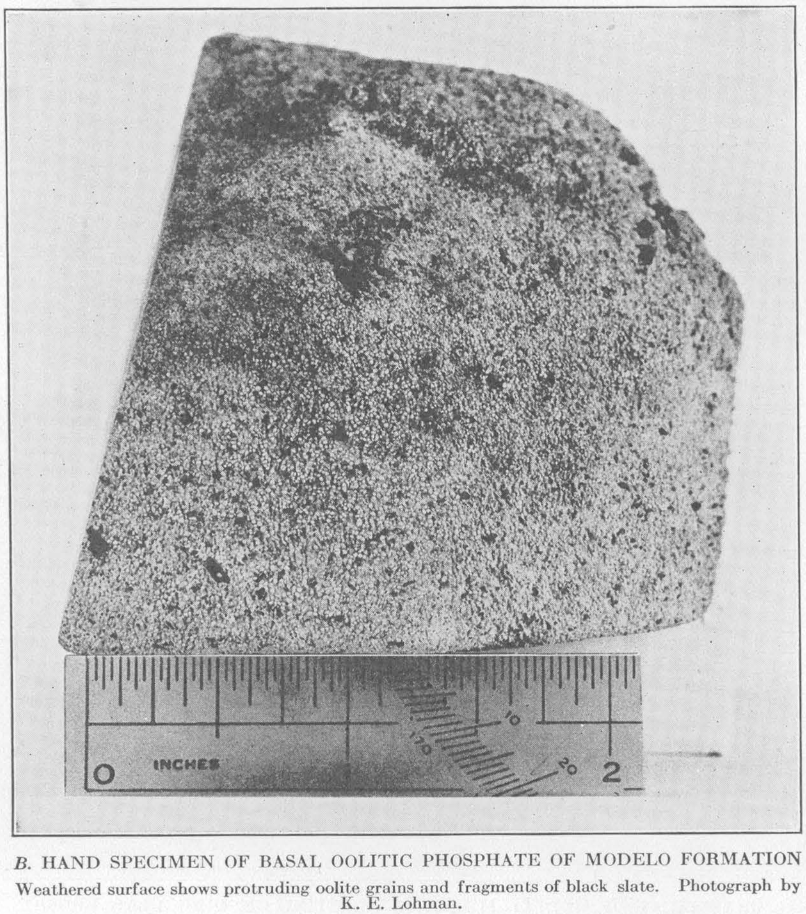


C. BASAL GRAYWACKE OF MODELO FORMATION RESTING ON IRREGULAR EROSION SURFACE OF SANTA MONICA SLATE (TRIASSIC?) ON SOUTH END OF RIDGE EAST OF BROWN CANYON



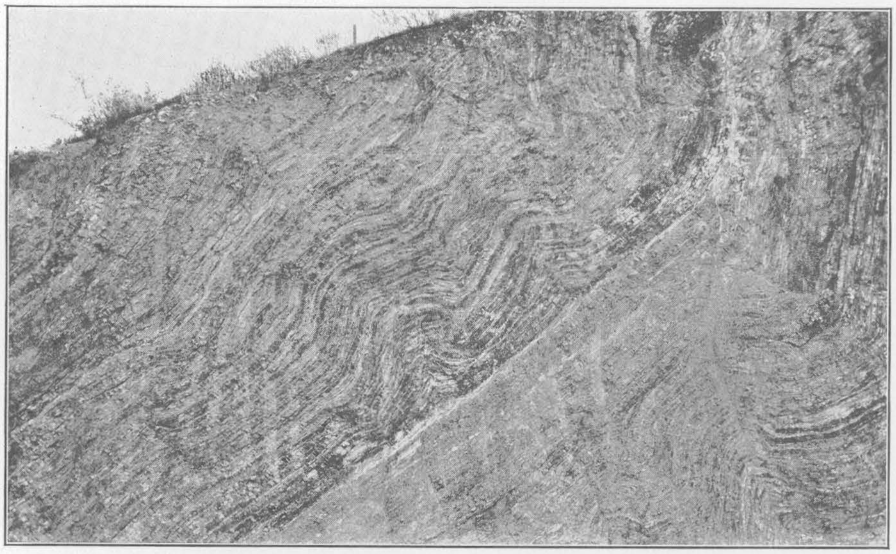
A. HAND SPECIMEN OF DARK-GRAY BASAL GRAYWACKE OF MODELO FORMATION

Shows abundant fragments of dark slate and casts of "Pecten" raymondi brionianus Trask. Photograph by K. E. Lohman.



B. HAND SPECIMEN OF BASAL OOLITIC PHOSPHATE OF MODELO FORMATION

Weathered surface shows protruding oolite grains and fragments of black slate. Photograph by K. E. Lohman.

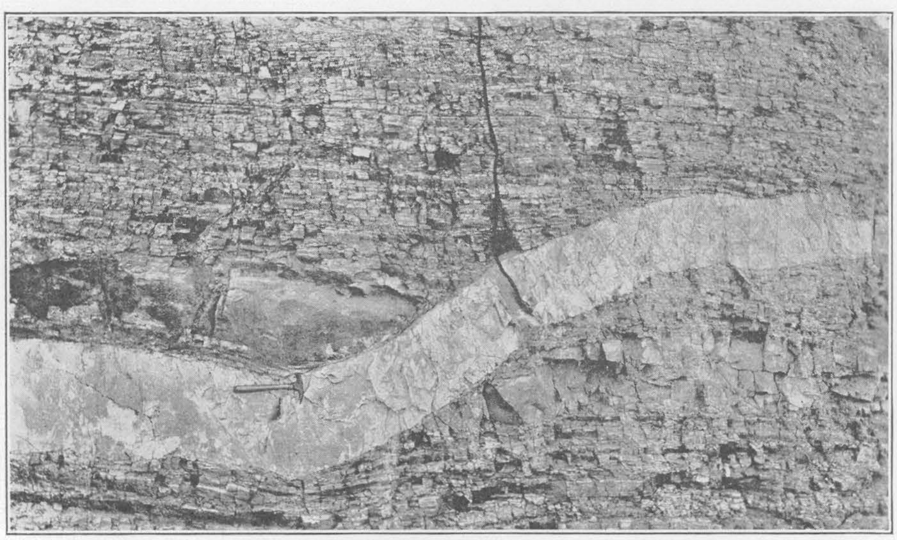


A. GRAY CHERTY SHALE IN LOWEST SHALE UNIT OF LOWER MEMBER OF MODELO FORMATION ALONG MULHOLLAND HIGHWAY

Shows folding of incompetent beds above a plane of slippage.

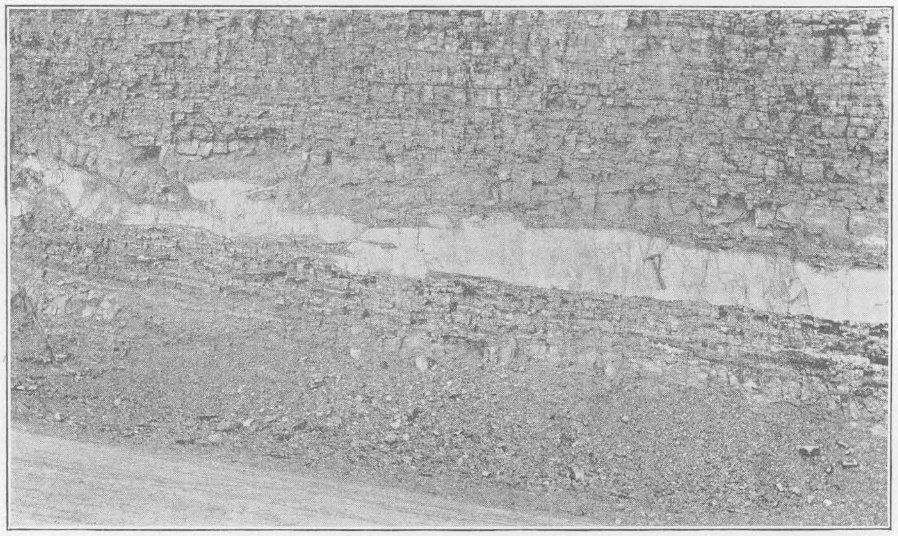
B. NONCHERTY SILICEOUS PLATY SHALE IN UPPER SHALE UNIT OF LOWER MEMBER OF MODELO FORMATION ALONG MULHOLLAND HIGHWAY

A bed of light-gray volcanic ash does not quite parallel the bedding of the underlying shale.



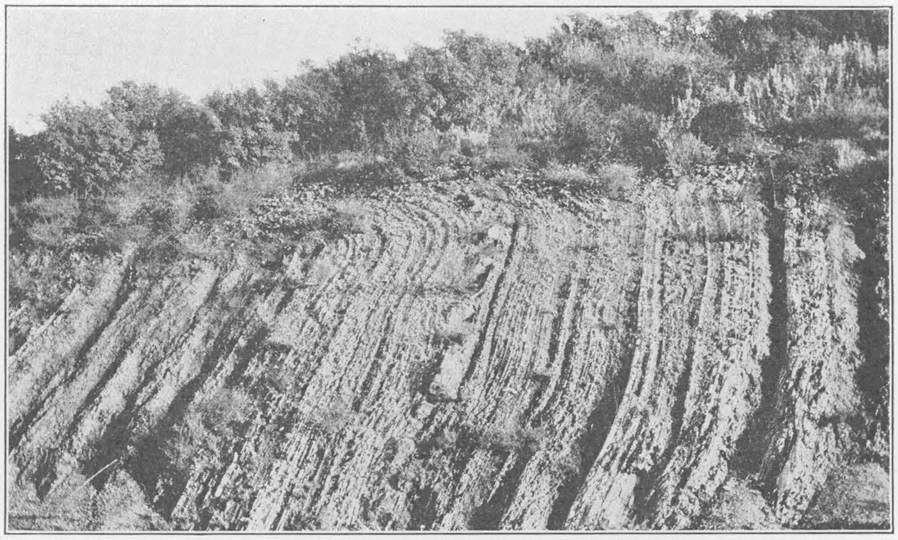
C. SANDSTONE DIKE IN GENTLY TILTED MODELO SHALE ALONG MULHOLLAND HIGHWAY

Dike passes into a sill at left and lenses out entirely. See also Plate 25, *A*.



A. SILL-LIKE INTRUSION OF SANDSTONE IN MODELO SHALE

The sandstone lenses out at left; at right it passes into the dike shown in Plate 24, *C*.

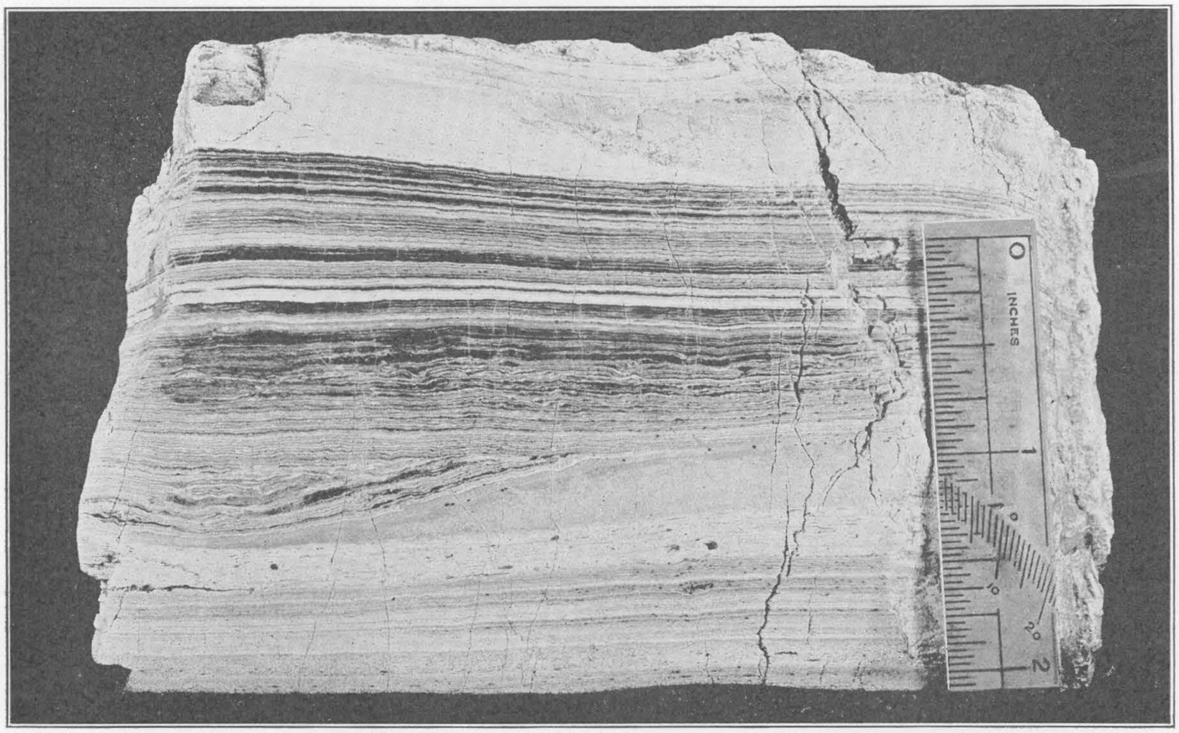


B. HARD WHITE SILICEOUS SHALE IN UPPER PART OF LOWER MEMBER OF MODELO FORMATION 1 MILE SOUTH OF GIRARD, CALIF.

Intercalated with soft clay shale, a bed of limestone, and a few thin beds of sandstone.

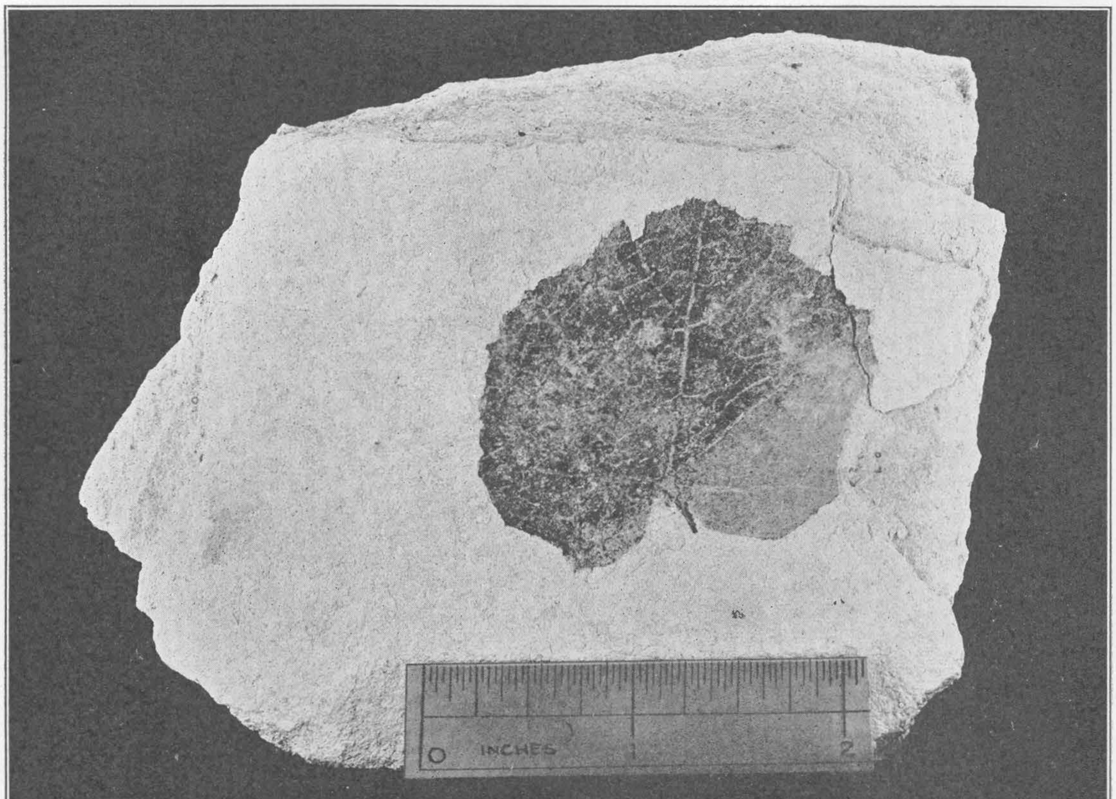
C. FINELY LAMINATED WHITE AND LIGHT-GRAY PUNKY DIATOMACEOUS SHALE

From beds just south of Ventura Boulevard and east of Girard.



A. FINELY BANDED CHERTY SHALE FROM LOWER PART OF MODELO FORMATION

Shows a diastem just below the middle. Photograph by K. E. Lohman.



B. WELL-PRESERVED IMPRINT OF LEAF FROM LAND PLANT (PLATANUS DISSECTA) IN PLATY SHALE OF LOWER MEMBER OF MODELO FORMATION

Some such imprints contain abundant marine Foraminifera. Photograph by K. E. Lohman.

tions of the lower member as well as all other stratigraphic units in the eastern part of the Santa Monica Mountains.

Only the lower part of the lower member of the Modelo is exposed along the south flank of the mountains, the remainder of the formation being concealed beneath an extensive Pleistocene alluvial plain which is crossed by numerous channels of Recent alluvium.

A section of the entire Modelo formation measured in the western edge of the area near structure section line A–A′ (pl. 16) and exposed in greater part along the Garrapata Canyon and Girard road is presented below. As the geologic map clearly shows, the lower member of the Modelo in this section includes some sandstone but contains none of the thick beds of soft sandstone present only 2 or 3 miles farther east and, for a similar reason, is radically different from the lower Modelo still farther east, in the vicinity of Stone and Brown Canyons, along structure section lines E–E′ and F–F′. Another striking difference between the lower Modelo along the Garrapata Canyon and Girard road and that of the north flank near Stone and Brown Canyons is in the character of the shale. Instead of the relatively soft shale described below, the shale near Stone and Brown Canyons is largely of the hard white platy type throughout all shale units in the lower member of the Modelo. This platy shale becomes less conspicuous as the lower member is traced farther east and appears to be partly replaced by soft bluish-gray and brown earthy shale.

Section of the Modelo formation exposed along the Garrapata Canyon and Girard highway north of Mohn Springs

Modelo formation:

19. Stratigraphically higher beds of the diatomaceous shale unit of No. 18 form the low-lying hills north of Ventura Boulevard near Girard and are associated with progressively more and finer grained soft sandstone as the top of the unit is approached. Probable approximate thickness ---------- 1, 100
18. White shale, thin bedded and becoming progressively more fissile, punky, and diatomaceous upward from base toward middle of unit. Beds of soft fine-grained brown and gray sandstone are intercalated with highly diatomaceous shale throughout lower exposed portion, where lenses and concretionary beds of hard light-gray limestone from 1 foot to 3 feet thick are common. Thickness exposed south of Ventura Boulevard on ridge just east of Girard ---------- 145
17. Compact and thin-bedded brown and light-gray shale with a few light-gray limestone concretions and concretionary beds. Free from sandstone except for a few thin beds of fine texture in upper 50 feet. Bedding planes of shale contain innumerable white granular specks and scattered black specks of carbonaceous material. Diatoms are present, but shale is not a typical diatomite- 450

Modelo formation—Continued.

16. Soft light-gray sandstone with numerous thin partings of bluish-gray shale in lower part. Sandstone ranges from fine and thin bedded in lower part to very coarse and gritty in upper part, where rounded boulders of soft bluish-gray shale from 8 to 10 inches in diameter are common. Bedding in upper, more massive part is very irregular, and there is distinct evidence of scouring and truncation of beds ---------- 155
15. Soft light-brown and grayish-brown shale with a 1-foot bed of yellow limestone at the base and light-gray slabby calcareous beds a few feet above. Shale is thin bedded and intercalated with a few ¼-inch beds of fine brown sandstone. Small but distinct rounded and oblong white specks are abundant along most bedding planes ---------- 220
14. Soft light-brown sandstone of medium-coarse texture. Some beds contain much biotite. Very little shale is present, but several thin beds of yellow-weathering limestone occur in upper part ---------- 215
13. Light bluish-gray thin-bedded shale and sandy shale intercalated with beds of soft fine gray and brown sandstone from one-fourth of an inch to 6 inches thick. Most of the shale is soft, but a 1-foot zone of platy opal shale occurs near the base, directly above a 1-foot bed of yellow limestone. Foraminifera collected from lower part ---------- 110
12. Soft gritty gray and brown arkosic sandstone in distinct beds from 2 inches to 6 feet thick separated by partings and thin beds of bluish-gray and white shale. A few irregular beds of yellow-weathering limestone occur in upper part ---------- 50
11. Hard white platy opaline shale and soft shale, poorly exposed, in lower 50 feet, apparently grading upward into beds composed largely of firm brown shale and sandy shale with thin beds of fine gray sandstone- 100±
10. Soft brown arkosic sandstone in beds from 4 inches to 3 feet thick separated by thin shale partings. Sandstone is coarse, gritty, and poorly sorted, with mineral grains as much as one-eighth of an inch across ---------- 60
9. A striking and distinctly thin-bedded series of brown and light-gray shale and fine sandstone. Fine brown sandstone comprises most of lower 100 feet or more and occurs in distinct beds from 1 inch to 12 inches thick, which are separated by thin partings or beds from one-eighth of an inch to 2 inches thick of soft bluish-gray shale and hard white platy shale. The white-weathering shale is brown when fresh and contains abundant calcareous Foraminifera. Beds of sandstone become progressively less prominent above, and north of the road crest this unit is composed largely of beautifully laminated foraminiferal shale, which ranges from dark brown to light brown and gray and from soft to hard platy opal shale. The most abundant type is brown and compact but not opaline. Thin partings of fine brown sandstone

Modelo formation—Continued.

occur throughout, and beds and lentils of yellow-weathering limestone from 1 foot to 3 feet thick are common. Two beds of yellowish-white bentonite, 4 to 6 inches thick, occur about 300 feet above the base	835
8. Thin-bedded bluish-gray and brown shale and sandy shale with several prominent beds of yellow-weathering limestone and limestone concretions. Becomes more sandy upward, with thin beds of fine brown sandstone in upper part. Top of this unit is 100 yards south of crest of road	160
7. Soft brown and dark-gray sandy shale, fine sandstone, and shale in a thin-bedded series. Beds range from less than 1 inch to 1 foot in thickness, and beds of soft brown sandstone are thicker in lower part. Beds and lenses of hard brown limestone are common	680
6. Hard white opal shale intercalated with soft brown shale, all in thin distinct beds which average less than 1 inch in thickness. Individual beds of hard white shale are as a rule beautifully banded with dark and light-gray laminae, with which are a few ⅛-inch laminae of fine well-sorted sandstone. Some of the hard platy shale is highly opaline and breaks with conchoidal fracture. A few 6-inch beds of hard brown limestone occur in this unit. Some soft brown shale contains calcareous foraminifers. Thickness approximate, owing to presence of poorly exposed anticline	100
5. Brown and bluish-gray soft shale and sandy shale with scattered concretions and lenticular beds of hard yellowish-brown limestone as much as 1½ feet thick. Columnar jointing conspicuous in lower part. White bed of bentonite, 6 inches thick, occurs 15 feet above base. Part of this unit is poorly exposed	120
4. Massive brown fine-grained sandstone; grades upward into overlying shale and sandy shale	5
3. Coarse argillaceous brown and gray sandstone, thinly but irregularly bedded and containing calcareous nodules with a few calcareous foraminifers	15
2. Brown arkosic and gritty sandstone of medium-coarse texture which grades downward into conglomerate and upward into earthy sandstone	3
1. Basal conglomerate and sandstone in distinct well-indurated beds 6 to 18 inches thick. Polished subangular and well-rounded pebbles, cobbles, and boulders 1 foot or less in diameter are very abundant; about 75 per cent of them are gray and purplish quartzite; the remaining 25 per cent consist of dark and light colored porphyries, together with light-gray granite and gneissic granite in subordinate amount. Other rock types, such as boulders of sandstone and slate, are notably scarce or absent. All this material is embedded in a coarse matrix of angular quartz and feldspar grains	12

Unconformity with angular discordance in dip of about 40°.
Topanga formation.

LOWER MEMBER

General features.—The lower member of the Modelo formation, which consists of alternating units of shale and massive sandstone and crops out over much of the north flank of the range west of Universal City and along the south flank west of Beverly Hills, is much more widespread than the upper member, which is restricted to a relatively narrow band on the north flank. (See pl. 16.) In much of the area where the Modelo rests directly upon the Santa Monica slate there is a hard massive bed of conglomeratic graywacke at the base of the formation, which is literally filled with small angular fragments of slate. (See pls. 22, *C*, and 23, *A*.) This bed is the most highly fossiliferous in the entire formation for macroscopic forms and is commonly 20 to 30 feet thick, though ranging from 50 feet down to the vanishing point. It is well exposed around the head of Sepulveda Canyon, south of the Mulholland Highway, and on the south flank between Stone and Benedict Canyons. Where this bed is not present its position at the base of the Modelo is occupied by shale and oolitic phosphate or, in areas some distance from the Santa Monica slate, as in Garrapata Canyon, by a coarse conglomerate without slate fragments and much like that common in the underlying Topanga formation.

There are two strikingly lenticular sandstone units somewhat higher in the Modelo formation; both are on the north flank of the range, one in the lower part of the formation between Sepulveda and Stone Canyons and the other slightly higher and farther west, in the area west of the Encino Reservoir. The former appears to be the easier to explain. Although this unit has a maximum thickness of 1,100 to 1,200 feet along structure section line E–E′, it lenses out entirely to the east and to the west within distances of 1 mile to 1¼ miles. (See pl. 16.) Its upper surface is parallel to the bedding in overlying units of the Modelo, but its base conforms in a remarkably close manner to the basinlike form of the old pre-Modelo topography of the underlying slate. As all of the Modelo formation is marine this lens of sandstone accumulated on the sea bottom and almost succeeded in filling a pronounced trough which crossed the present site of the Santa Monica Mountains and was a part of the old pre-Modelo surface. This sandstone is overlain, with intervening shales, by other sandstones which also lens out to the east. (See pl. 16.) The discontinuity of at least some of these sandstone units, as shown on the geologic map, is apparently due largely to the fact that most of them appear to die out and have comparatively few or no stratigraphic equivalents at points farther east along the strike; but there are other occurrences, such as the prominent sandstone lens west of the Encino Reservoir, in which the discontinuity shown is due not to the actual disappearance of a stratigraphic unit but largely to its lateral change into strata composed predominantly of shale and a relatively small amount of sandstone. It is a change in facies—not a radical change in thickness of beds at any one horizon—which accounts for the lenticularity of many of the sandstone units.

The character of the shale in the lower member of the Modelo formation also undergoes distinct lateral changes, a fact which is particularly evident when the type of shale composing most of the section in the eastern part of the area, particularly east of upper Benedict Canyon, is compared with that in the western part, along the Garrapata Canyon and Girard road. In the more eastern district the shale of all parts of the lower member, exposed in road cuts along the Mulholland Highway and subsidiary roads leading to the north and to the south, is in greater part of the hard, platy, opaline type (see pl. 24, *A*, *B*), and much of it is relatively free from intercalated partings and thin beds of sandstone. In the western edge of the area covered by this report hard, platy opaline shale in notable amount occurs only as a 100-foot unit in the lower part and a second 100-foot unit in the upper part of the lower member, most of the remainder consisting of thin-bedded soft but firm brown shale and fine sandstone.

On the south flank of the range, west and northwest of Beverly Hills, between Mandeville and Coldwater Canyons, the exposed part of the lower Modelo is similar in stratigraphic sequence to the lower Modelo directly to the north, on the north flank. The basal graywacke is very well developed on the south flank between Stone and Benedict Canyons and is overlain by 50 to 100 feet of hard platy shale, which in turn is overlain by a thick and widespread unit of soft gray and brown gritty sandstone much like that which forms the huge sandstone lens on the north flank just west of Stone Canyon.

Along the coast northwest of Santa Monica the exposed Modelo consists of a highly deformed mass of thin-bedded soft brown and hard, platy shale, some of which is highly bituminous, stained dark gray or black, and impregnated with considerable sulphur.

Basal graywacke.—The graywacke at the base of the Modelo in much of the area where this formation directly overlies the Santa Monica slate is remarkable in being so distinct, so rich in marine invertebrate fossils, and composed to so large an extent of rock fragments derived from the underlying slate. Much concerning its origin is thus readily determinable, and its good exposures in easily accessible localities make it a choice illustration of geologic processes active here prior to and at the beginning of Modelo time.

This basal graywacke is massive, is resistant to weathering, and commonly crops out in a vertical wall 20 to 30 feet high. Outcrops of this nature occur on ridges near the head of Sepulveda Canyon, just west of the Encino Reservoir, and, in somewhat subdued form, on ridge tops along the south flank of the range between Stone and Benedict Canyons. Its dark-gray color serves to identify the rock at a glance; by closer inspection this color is found to result from the abundance of angular to well-rounded fragments of black slate, which commonly average less than 1 millimeter in diameter. Subangular grains of colorless quartz are recognizable with a hand lens and have been found to be very abundant in thin sections and to be associated with grains of plagioclase and orthoclase feldspar, microcline, and quartzite. Resting upon the Santa Monica slate in some areas is a rock of light-gray speckled appearance, which is an impure microcrystalline limestone containing many fragments of black slate. At other places the graywacke contains scattered oolite grains or gives place entirely to a 6-inch bed of oolitic phosphate that is directly overlain by platy siliceous shale.

On the ridge west of the Encino Reservoir abundant slate fragments in the basal bed of the Modelo are associated with a greater abundance of cobbles of light-gray granite. This locality therefore does not offer a typical example of the basal graywacke.

The best exposure of typical graywacke and its contact with the underlying slate is to be found 2 miles northwest of Beverly Hills, in the road cut that leads up the south end of the ridge just west of Benedict Canyon. (See pl. 22, *C*.) Parts of the graywacke are very fossiliferous at this locality and on the ridge west of Brown Canyon and have yielded several species of fossil marine invertebrates, which are listed on page 110. Plate 23, *A*, shows a hand specimen of this graywacke with many casts of "*Pecten*" cf. *P. raymondi brionianus* Trask, a fossil that almost everywhere greatly outnumbers all other species combined. In the exposure shown in Plate 22, *C*, the contact between the slate and the graywacke is very uneven; irregular pockets occur on the old surface of the slate and are filled with well-rounded cobbles and boulders of schist, porphyry, basalt, and soft sandstone as much as 12 inches in diameter. Narrow cracks filled with lower Modelo sediment extend for several feet below the contact.

Oolitic phosphate.—Oolitic phosphate, consisting to a large extent of the amorphous mineral collophane, occurs as a 6-inch bed at the base of the Modelo on the first ridge top west of the Stone Canyon Reservoir and at other scattered localities as far west as Mandeville Canyon. In the west wall of Stone Canyon a 1 to 2 foot zone of oolitic phosphate occurs at the base of the bituminous shale described on page 109. Although few descriptions of oolitic rocks in the California Tertiary can be found in the literature, the presence of rock of this type is apparently not uncommon in some parts of the Miocene. Dr. R. D. Reed,[38] who has briefly described the character and occurrence of oolitic phosphate in the "Monterey shale" (Salinas

[38] Reed, R. D., Phosphate beds in the Monterey shales: Geol. Soc. America Bull., vol. 38, pp. 195–196, 1927.

shale) of the Salinas Valley region, has kindly loaned to the writer a suite of thin sections made from the Salinas Valley material. The principal difference between this material and the Modelo phosphate of the Santa Monica Mountains is its mode of occurrence and the character of the matrix surrounding the oolite grains.

The oolitic phosphate at the base of the Modelo is light brown to grayish brown and, unless examined in hand specimens, might readily be mistaken for sandstone. As is shown in Plate 23, *B*, weathered surfaces are, however, covered with rounded pellets about 0.5 millimeter in diameter which, because of their rather uniform size and roundness, are strikingly different from the ill-sorted angular mineral grains common to most Tertiary sandstones of California. These pellets are associated with small angular fragments of black slate, and, from the fact that they protrude and give a rough surface to the rock, they appear upon casual observation to be embedded in a relatively soft matrix.

Thin sections of this rock reveal a very uniform texture and mineral composition. The oolite grains range from 0.2 to 0.7 millimeters in diameter, are in contact throughout almost the entire rock, and are formed around nuclei of black slate, quartz, and quartzite fragments. The small interstitial areas remaining between the adjacent oolite grains are filled with calcite and a deep-brown substance which appears to be slightly anisotropic. The accompanying chemical analysis of the oolitic phosphate suggests that this brown substance is iron oxide (Fe_2O_3). It fills a few somewhat larger areas in the slides, and these areas contain small spots of collophane about 0.05 millimeter in diameter which have no apparent nuclei. Calcite also fills veins which individually cut across as many as 15 oolite grains and in several sections has actually replaced some of the collophane composing these grains.

Most of the oolite grains have a pale-brown color and are clouded with irregular areas and disseminated minute specks of hematite, which in reflected light have a deep reddish-brown color. Practically all the oolite grains have concentric structure, although this structure, in different grains results from different causes and varies considerably in distinctness. The most common and most distinct concentricity is made apparent by one or more rings of brown iron oxide. In other places disseminated minute brown specks of iron oxide are arranged in discontinuous concentric bands. In a large proportion of the grains concentric structure is apparent only because of distinct differences in the index of refraction of adjoining growth bands of collophane. The collophane of several fragments examined in refractive-index liquids proved, however, to be consistent in having an index of refraction between 1.61 and 1.62, a uniformity of this property which, as noted in the thin sections, is true for many individual grains. As shown by Rogers,[39] the refractive index of collophane ranges from 1.573 to 1.623.

Chemical analyses of oolitic phosphate from the base of the Modelo formation, on ridge west of Stone Canyon, and of Wyoming phosphate

	1	2
Insoluble in HCl	13.24	10.00
SiO_2	1.21	.00
Al_2O_3	1.30	.89
Fe_2O_3	8.36	.73
MgO	.10	.28
CaO	40.38	43.54
Na_2O } water-soluble alkalies	Trace.	1.10
K_2O } water-soluble alkalies	Trace.	.48
H_2O-	3.02	1.04
H_2O+	----	1.14
CO_2	9.20	6.00
P_2O_5	21.44	27.32
SO_3	None.	1.59
F	1.52	.60
Cl	Trace.	Trace.
Organic matter	.37	Not det.
	100.14	96.51
Less $O_2=F$	.64	
	99.50	

1. Phosphate from Modelo formation. J. G. Fairchild, analyst.
2. Phosphate from main phosphate bed 2½ miles east of Cokeville, Wyo. (Twenhofel, W. H., Treatise on sedimentation, p. 395, 1926).

The percentage of anhydrous phosphate ($Ca_3P_2O_8$) computed from the analysis of the sample from the Modelo formation is 46.8. On the assumption that all of the water and fluorine present is combined with this tricalcium phosphate to form collophane, the total amount of collophane is approximately 51.3 per cent. In the sample analyzed the 8.36 per cent of Fe_2O_3 (hematite) is somewhat less and the calculated 19.7 per cent of $CaCO_3$ (calcite) is greater than the amounts of these minerals suspected from examination of thin sections. The Al_2O_3 and SiO_2 are present in the rock as constituents of the detrital material that forms nuclei for the oolite grains.

Sandstone.—The distribution and lenticular nature of major sandstone units in the Modelo formation are described on page 104. These sandstones, although commonly very massive (see pl. 22, *A*), are soft and are even less resistant to erosion than the associated hard platy shale, a condition which is made apparent by the fact that many low hills, although composed mostly of sandstone and surrounded by low areas of sandstone outcrop, are capped by platy shale. Many of the major sandstone units, particularly those in the lower and middle parts of this member east of the Encino Reservoir, contain abundant sandstone concretions. Conglomerate is scarce, and, except for the basal unit already described and some fine conglomerate in a part of the prominent sandstone lens west of the Encino Reservoir, rock of this type is practically

[39] Rogers, A. F., Mineralogy and petrography of fossil bones: Geol. Soc. America Bull., vol. 35, pp. 541-547, 1924.

absent from the entire Modelo formation, although some of the major sandstone units are very coarse and gritty.

The areas mapped as shale, particularly west of the Encino Reservoir, contain a large amount of sandstone intercalated as thin beds between equally thin layers of shale. The average thickness of these beds is probably 2 or 3 inches, and the sandstone composing them is in general about as fine and as well sorted as any of the other sandstones of the district. Curve 6 in the accompanying chart (fig. 8) illustrates the texture and fairly well sorted character of these thin-bedded sandstones and may be compared with curve 3, which represents an unusually ill-sorted sandstone from this same thin-bedded series of Modelo shale and sandstone, and curve 4, which represents a much coarser but even better sorted Pleistocene marine sand.

Although some of the sandstones of this marine formation are rather poorly sorted as to size of mineral grains, all of them appear to be very clean and free from disseminated clayey material. Except in the rare conglomeratic portions, where the larger detrital rock fragments are fairly well rounded, the mineral grains of all sandstone samples examined are very angular and commonly have jagged points and show little evidence of abrasion.

Mineralogically, all these so-called sandstones of the Modelo and also those of the underlying Topanga formation are arkoses, for they contain from 30 to 60 per cent of feldspar. Most of the feldspar grains are turbid, although some are as clear as any of the quartz. Orthoclase appears to be much the most abundant feldspar, but, to judge from the indices of refraction and the occurrence of polysynthetic twinning, anorthoclase and oligoclase are very common, and all other species of plagioclase are present. The quartz and feldspars are associated with heavy minerals, which commonly form about 1 per cent or less of the entire rock. The heavy minerals identified in samples examined are, in the approximate order of abundance, leucoxene, black opaque minerals (probably ilmenite and magnetite), garnet, zircon, tourmaline, titanite, rutile, and green hornblende. No striking difference was found in the mineralogy of sandstones of the Modelo, Topanga, Sespe (?), Vaqueros (?), and Chico formations; careful study of a large number of samples from each of these formations would probably reveal some distinctive mineralogic characteristics; but inasmuch as all these formations have probably been derived directly or indirectly from the same source, it is believed that such characteristics would consist of differences in percentage composition rather than of important differences in the minerals present.

Sandstone dikes.—In several exposures along the Mulholland Highway, within 2 or 3 miles east of the Girard and Garrapata Canyon road gently dipping shale in the lower member of the Modelo formation is cut by intrusive bodies of soft light-gray sandstone. Most of these intrusions are in the form of dikes from a few inches to more than a foot thick; some of them cut the shale approximately normal to the bedding; others in 2-dimensional exposures appear to be sills. Plate 24, *C*, shows a rather unusual intrusive mass of light-gray sandstone which is clearly a dike at the right but which passes into a sill at the left and continues parallel to the bedding with remarkably uniform thickness for about 20 feet before it begins to pinch out. (See pl. 25, *A*.)

In color, hardness, texture, and mineralogy these intrusive sandstones are indistinguishable from the normal beds of sandstone that are commonly associated with shale in this part of the Modelo formation. It seems probable that they have been produced by compression of a series of shale and soft sandstone of Modelo age, by which the poorly consolidated sandstone has been forced upward along the joints and

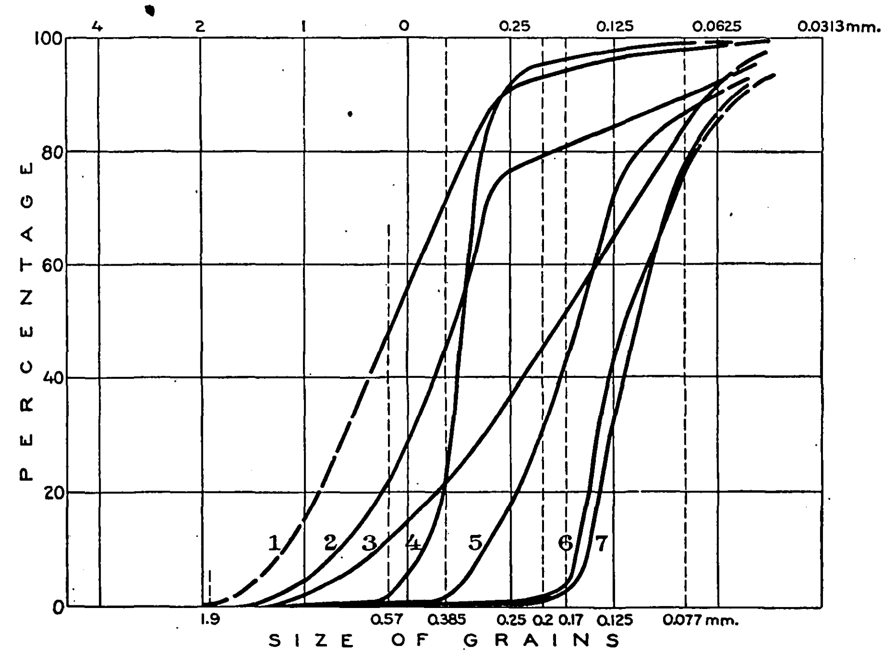


FIGURE 8.—Cumulative curves showing the texture and degree of sorting of some Miocene and Pleistocene sandstones. 1, Massive sandstone associated with conglomerate in Topanga formation 1,000 feet south of Topanga post office; 2, coarse gritty sandstone bed intercalated with punky diatomaceous shale, 100 feet below top of Modelo formation exposed on South Sherman Way; 3, coarse gritty sandstone of Modelo formation intercalated with thin partings of hard white platy siliceous shale containing abundant Foraminifera, near and similar to that shown in Plate 25, *B*; 4, Pleistocene marine beach sand collected from 200 to 300 foot terrace along coast between Topanga and Santa Ynez Canyons; 5, fine sandstone of Topanga formation associated with thin partings of shale in Topanga Canyon, near Lida Park bridge; 6, fine sandstone from 8-inch bed in Modelo formation, collected immediately below foraminiferal shale parting at crest of Girard-Garrapata Canyon road; 7, fine sandstone intercalated with richly diatomaceous shale of upper Modelo age from road cut 1¼ miles east of Girard

bedding planes of the shale. Plate 25, *A*, illustrates how such an intrusion might be expected to terminate.

Shale.—The Modelo is predominantly a shaly formation, and for this reason most of it presents a strong contrast to other formations in the eastern part of the Santa Monica Mountains. The most prominent and characteristic feature of the lower member of the Modelo is the abundance of thin-bedded hard white to light-gray and light-brown platy siliceous shale in much of the area covered by its outcrop. A small amount of shale of this type occurs as thin beds associated with sandstone and conglomerate in the underlying Topanga formation, a fact which precludes the presence of such rock from serving as an infallible guide for the identification of the Modelo; but stratigraphic units of hard platy siliceous shale more than a few feet thick (such as are shown in pl. 24) are restricted to the lower member of the Modelo in the area covered by this report.

These white or nearly white siliceous shale units are everywhere beautifully bedded and, although composed predominantly of the hard platy type, contain partings and thin beds of soft earthy shale. (See pl. 25, *B*.) The hard platy shale is uniformly cut by intersecting systems of joints and commonly weathers to thin rectangular chips only a few inches across. It varies considerably in hardness, some beds being chertlike in character, whereas others appear not to be so highly siliceous but are hard, compact, well consolidated, and just as distinctly bedded. Whether cherty or not, the shale commonly breaks along a conchoidal surface. The beds are generally only from 1 to 3 inches thick, and in addition they are almost invariably composed of distinct laminae so thin that 10 or 15 may be present in a thickness of 1 millimeter. (See pl. 26, *A*.) Shale of this type is locally rich in calcareous Foraminifera that can be readily detected with a hand lens, but the occurrence of these minute marine fossils is peculiar in that they may be and commonly are restricted to a single lamina within several feet, or perhaps tens of feet, of shale. Fish scales are abundant along many bedding planes, and even the imprints of entire fish are common in the highly fissile shale of some horizons. Isolated perfect brown casts of the leaves of land plants are locally also fairly common in foraminiferal and other beds of the lower and upper shale units. (See pl. 26, *B*.)

The hard white platy shale in the lower 100 feet of the Modelo and the softer brown shale in the upper part of the lower member at the west edge of the area contain soft light-gray beds of volcanic ash and bentonite from a few inches to 2 or 3 feet in thickness. The volcanic ash of samples examined is fairly uniform in having an index of refraction of 1.50 to 1.51, which indicates a silica content equivalent to that of a granite.

Beds, lenses, and concretions of limestone weathering yellowish brown are common throughout practically all of the Modelo formation, generally with a thickness of 1 to 3 feet. Thinner beds of hard gray laminated limestone and nodular limestone are locally intercalated with the shale, and some of them contain Foraminifera in varying abundance, even in parts of the section where all other beds are apparently barren of these minute fossils.

The thin-bedded sandstone that forms a considerable part of most of the shale units shown on the geologic map, particularly in the western part of the area covered by this report, is briefly described on page 107.

In the outcrop and hand specimen the platy shale in the lower member of the Modelo varies considerably in hardness and luster. The general appearance of the white platy shale is shown in Plate 24, *B*, and the gray, more finely banded cherty shale in Plates 24, *A*, and 26, *A*. The former, although compact and hard, weathers so that it may be readily scratched with a hammer, has a dull luster and commonly an irregular fracture, and in hand specimen does not show its composition; the latter is actually a chert which rings when struck with a hammer and always has a conchoidal fracture and a somewhat vitreous luster. Thin sections reveal the fact that both of these types of hard platy shale contain opal and cryptocrystalline silica in large amounts but in different proportions. Opal, apparently the most abundant constituent in thin sections of the cherty shale, is, in contrast, not so plentiful in the noncherty variety of platy shale; although composing a large part of this rock, it is here generally associated with a much larger amount of cryptocrystalline silica.

It seems probable that to this difference in the proportions of opal and cryptocrystalline silica may in some degree be ascribed the difference in the general appearance and hardness of these two types of shale where weathering has been active. Another factor, which has been effective in some localities at least, lies in the fact that the noncherty shale commonly contains a greater number of Foraminifera and other microscopic carbonate masses.

Analyses of shale of Modelo formation

[J. G. Fairchild, analyst]

	1	2	3
SiO_2 (total)	73. 71	73. 04	55. 80
Al_2O_3	7. 25	3. 58	4. 13
Fe_2O_3	2. 63	1. 28	1. 36
FeO	. 44	. 44	. 44
MgO	1. 47	. 45	. 50
CaO	1. 72	8. 63	18. 14
Na_2O	1. 19	. 40	. 40
K_2O	1. 00	. 55	. 75
H_2O-	2. 88	2. 82	2. 58
H_2O+	6. 94	2. 69	3. 02
TiO_2	. 50	. 30	. 25
CO_2	Trace?	5. 96	12. 81
P_2O_5	. 24	. 15	. 24
SO_3	. 16	. 05	Trace?
Organic carbon	. 00	. 10	. 16
	100. 13	100. 43	100. 58
Soluble SiO_2 (5 per cent Na_2CO_3)	4. 46	17. 22	18. 52

1. White diatomaceous shale from road near Hollywood Country Club; 40 grams.
2. Cherty shale 1 foot above volcanic ash bed on Mulholland Highway and about 50 feet above base of upper shale member of lower part of the Modelo; 150 grams.
3. Hard platy shale 1 foot below bed of volcanic ash on Mulholland Highway; 200 grams.

In addition to the large amounts of silica samples 2 and 3 contain considerable calcium carbonate—13.5 and 29.1 per cent, respectively. Much of this calcium carbonate is present as calcite in abundant shells of Foraminifera.

The distinct banding so apparent in some hand specimens and in most thin sections is due to alternations of several types of material with essentially colorless opal or cryptocrystalline silica. Most of the sections of cherty shale show alternations of opal and cryptocrystalline silica with wavy bands of two different substances, the more common one a white cloudy opaque mass and the other a speckled brown which varies considerably in intensity of color. Even with an immersion lens the character of the white opaque substance can not be satisfactorily determined. There is some suggestion that it also is largely opal and that its cloudiness is due to the presence of fine closely bunched particles of clay too small to be resolved with the ordinary petrographic microscope. The brown substance has the distribution and appearance of organic matter and is assumed to be such, there being a striking contract between it and less numerous bands of iron oxide, which, in reflected light, are reddish brown and commonly are associated with grains of magnetite.

In some thin sections of the noncherty type of shale opal bands alternate with colorless carbonate bands of about equal thickness, an association which, in some places at least, is in part the cause of the greater softness of this rock. Another type of banding is produced by the presence here and there of laminae of fine sandstone or silt composed largely of angular well-sorted detrital grains of quartz and feldspar. Such laminae are usually thicker than associated siliceous bands formed by chemical precipitation, some of them having been observed to be over 4 millimeters thick.

Measurements of several series of bands in different slides show that the average thickness for a pair of bands—that is, one opal band and an adjoining brown organic band or colorless carbonate band—is approximately 0.15 millimeter. These measurements have been taken in selected areas of the thin sections where banding is most distinct and continuous.

Both the cherty and noncherty types of hard platy siliceous shale in the lower member of the Modelo commonly contain abundant well-preserved calcareous Foraminifera. These occur embedded in bands of every character, whether opal, cryptocrystalline silica, brown organic matter, reddish-brown iron stain, or carbonate, but because of the predominance in these shales of opal and cryptocrystalline silica, these substances most commonly incase the Foraminifera. Siliceous organisms, such as Radioloria, diatoms, and sponges, are invariably scarce or totally absent from the many microscopically examined samples of all types of shale from the lower member of the Modelo. All these organisms are, however, commonly present in small numbers in the cherty and noncherty shale.

Bituminous shale.—The basal portion of the Modelo formation contains at least one unit of highly bituminous black or dark brownish-gray shale. Exposures of shale of this type at approximately the same stratigraphic horizon occur in the west wall of Stone Canyon west and northwest of the reservoir, and at a point 1½ miles from the coast about 2,000 feet up the small canyon west of Will Rogers's ranch which joins Rustic Canyon at the Beverly Boulevard. Much of the Modelo shale exposed along the Coast Highway about half a mile east of the mouth of Santa Ynez Canyon is of this same type and may also belong to the lowest few hundred feet of the formation. In 1908 Arnold and Johnson[40] investigated a reported volcano in this last-mentioned area and found that the report was due to the burning of this bituminous Miocene shale, which, they believed, had been ignited by lightning or spontaneous combustion.

This bituminous shale is thin bedded, hard, and compact and shows some resemblance to hard rubber. The dark color is apparent only in comparatively fresh exposures along cliffs; on weathered surfaces along gentle slopes the shale occurs as hard white chips scattered over the ground. In Stone Canyon shale of this type, with intercalated 1 to 3 foot beds of hard dense light-brown limestone, occupies the lower 75 to 100 feet of the Modelo formation and rests unconformably upon much more steeply dipping Topanga sandstone. The dark bituminous shale is beautifully laminated and spotted with thin light-gray and brown phosphatic streaks, which swell irregularly to form elongate phosphatic nodules parallel to the bedding.

Phosphorus determinations on representative samples of the limestone, light-gray nodules, and laminated shale showed 0.6, 14.2, and 2.14 per cent of phosphorus respectively. These percentages, when calculated as calcium phosphate, represent 3.0, 71.0, and 10.7 per cent of that compound.

In thin sections the dark bituminous shale of the lower part of the Modelo appears rich in disseminated dark-brown organic substances. Scattered detrital mineral grains, mostly quartz, are common and attain a maximum size of about 0.2 millimeter, although they constitute a relatively small part of the rock. Thin stringers of carbonate and stringers and lenticular bodies of finely crystalline chalcedony are conspicuous features of these thin sections. Light-brown transparent stringers and rounded masses of an isotropic substance, either collophane or some oil residue, are also common in sections of shale samples collected from the lower 6 feet of the Modelo formation in

[40] Arnold, Ralph, and Johnson, H. R., The so-called volcano in the Santa Monica Mountains: Science, new ser., vol. 27, pp. 553-554, 1908.

Stone Canyon. Only one diatom was noted in all the thin sections of this oil shale. Calcareous Foraminifera, however, are abundant in some of the sections but are totally absent from others.

It was noted in preliminary tests with solvents and by heating shale in test tubes that this basal bituminous shale member of the Modelo contains both free oil or oil residues and pyrobituminous substances. Consequently samples were submitted to E. T. Erickson, of the Geological Survey, for quantitative solvent tests and estimates of the amounts of pyrobitumens present, and to the Bureau of Mines for distillation tests to determine accurately the total combined amounts of all bituminous substances. Mr. Erickson has provided the following information:

Material extracted with chloroform from black shale of lower part of Modelo formation

	Per cent
Along west side of Stone Canyon Reservoir:	
5 feet above base of formation	0. 96
15 feet above base of formation	1. 50
Excavation for New Bay Club, 3,000 feet east of mouth of Santa Ynez Canyon (2 samples from different localities)	. 38

"From each of the samples the material extracted with chloroform was dark colored and solid at ordinary temperatures. When it was dissolved in a small quantity of benzene and then an excess of light petroleum ether (specific gravity 0.634) was added the characteristic precipitate was produced that is given by natural petroleum residues (particularly petroleum residues derived from asphaltic-base petroleums) when tested in like manner. Heating the precipitated material thus obtained from each sample produced the characteristic intumescence that is given by the material obtained from the treatment of asphaltic-base petroleum residues with benzene and light petroleum ether.

"For each sample the shale material remaining from the chloroform-extraction test was dried free of chloroform and then tested for pyrobituminous matter by the test-tube heating method. Each test yielded a tarry oil material such as is typically given by the pyrobituminous material in oil shale. The yield of the tarry oil thus obtained from each sample was estimated to be somewhat less than 15 gallons a ton."

The Bureau of Mines made distillation tests and submitted the following determinations of total oil yields for four representative samples:

1. Sample from 18-inch bed of shale about 6 feet above base of Modelo formation, west wall of Stone Canyon on road west of reservoir, due north of Bel Air and northwest of Beverly Hills: Oil yield, 14.5 gallons to the ton; specific gravity of oil, 0.908.
2. Sample from 15-foot bed of black shale about 6 feet above sample 1: Oil yield, 18 gallons to the ton; specific gravity of oil, 0.902.
3. Sample collected just above Coast Highway 4 miles northwest of Santa Monica and 3,000 feet east of the mouth of Santa Ynez Canyon: Oil yield, 8.6 gallons to the ton; specific gravity of oil, 0.905.
4. Sample from same locality as sample 3: Oil yield, 3.1 gallons to the ton; specific gravity of oil, 0.905.

Fossils and age.—Microscopic fossils collected from both the lower and upper members of the Modelo are listed and described on pages 112–115. Marine invertebrate fossils collected at several localities from the basal few feet of the lower member of the Modelo have been identified by W. P. Woodring as follows:

Fossils from the basal part of the lower member of the Modelo formation of the eastern part of the Santa Monica Mountains

[Localities plotted on Plate 16. For description of localities see pp. 123-124]

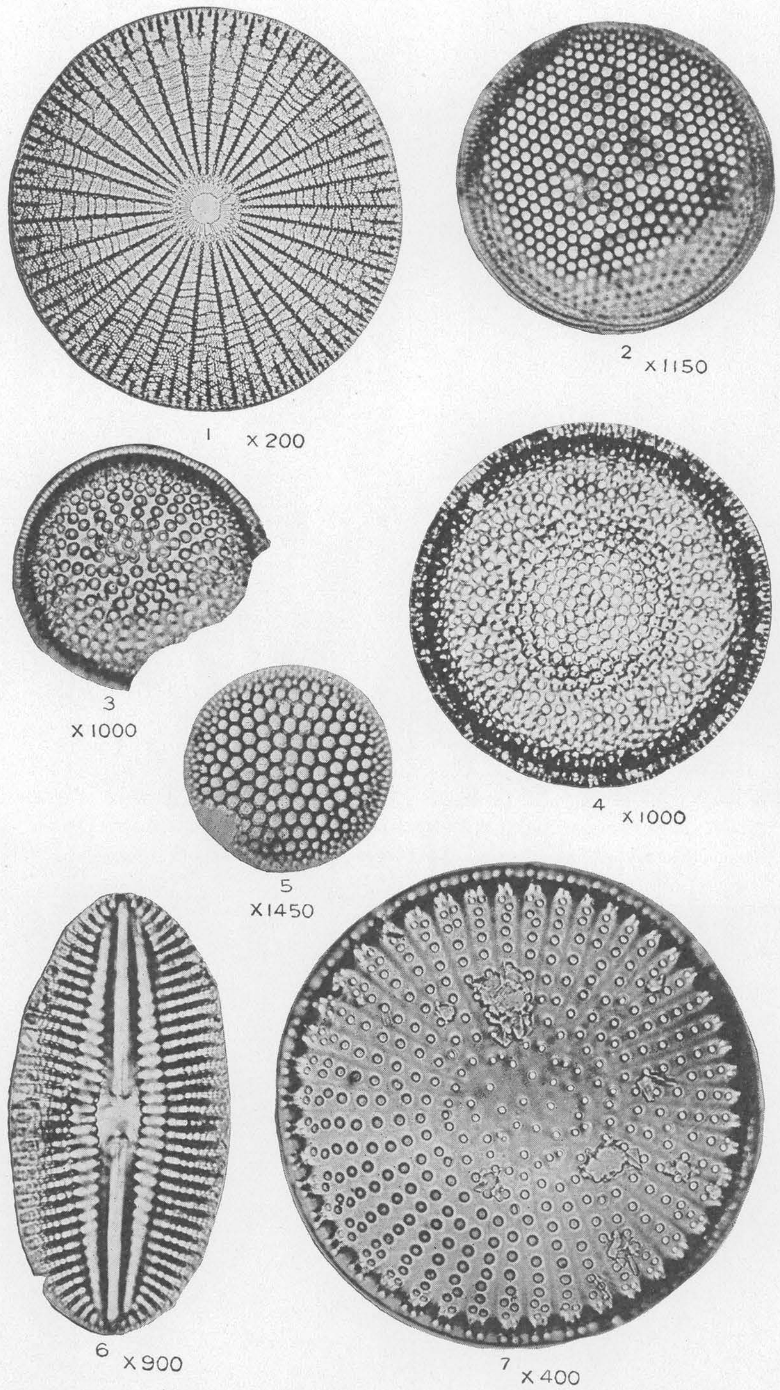
	12	15	29	30	51	52	64
Gastropods:							
Acmaea? sp							×
Haliotis palaea Woodring						×	
Tegula sp						×	
Lamellibranchs:							
"Leda" sp					×		
Navicula sp						×	
Ostrea sp		×					
"Pecten" cf. P. raymondi brionianus Trask	×	×	×	?	×	×	
Hinnites sp					×	×	
Lima sp. a (finely ribbed)						×	
Lima sp. b (coarsely ribbed)						×	
Mytilus cf. M. kewi Nomland					×	×	
Cerastoderma cf. C. nuttalli (Conrad)					×		
Cerastoderma sp						×	
Echinoids:							
Astrodapsis cf. A. brewerianus brewerianus (Rémond)				×		×	
Astrodapsis cf. A. brewerianus diabloensis Kew				×			

The following comments are made by Doctor Woodring:

The fossils from the basal part of the Modelo formation are particularly interesting, inasmuch as "*Pecten*" *raymondi*, listed above as "*Pecten*" cf. *P. raymondi brionianus*, is the only species so far recorded from these beds.[41] This "*Pecten*" is the most abundant species collected. The *Haliotis* is the first species of this genus found in Miocene beds on the Pacific coast, though it is recorded from Upper Cretaceous deposits near San Diego.[42] The sand dollar listed as *Astrodapsis* cf. *A. brewerianus brewerianus* (Rémond) is a very small primitive *Astrodapsis* that has barely raised petals. The other variety, which is slightly more

[41] Kew, W. S. W., U. S. Geol. Survey Bull. 753, p. 66, 1924.

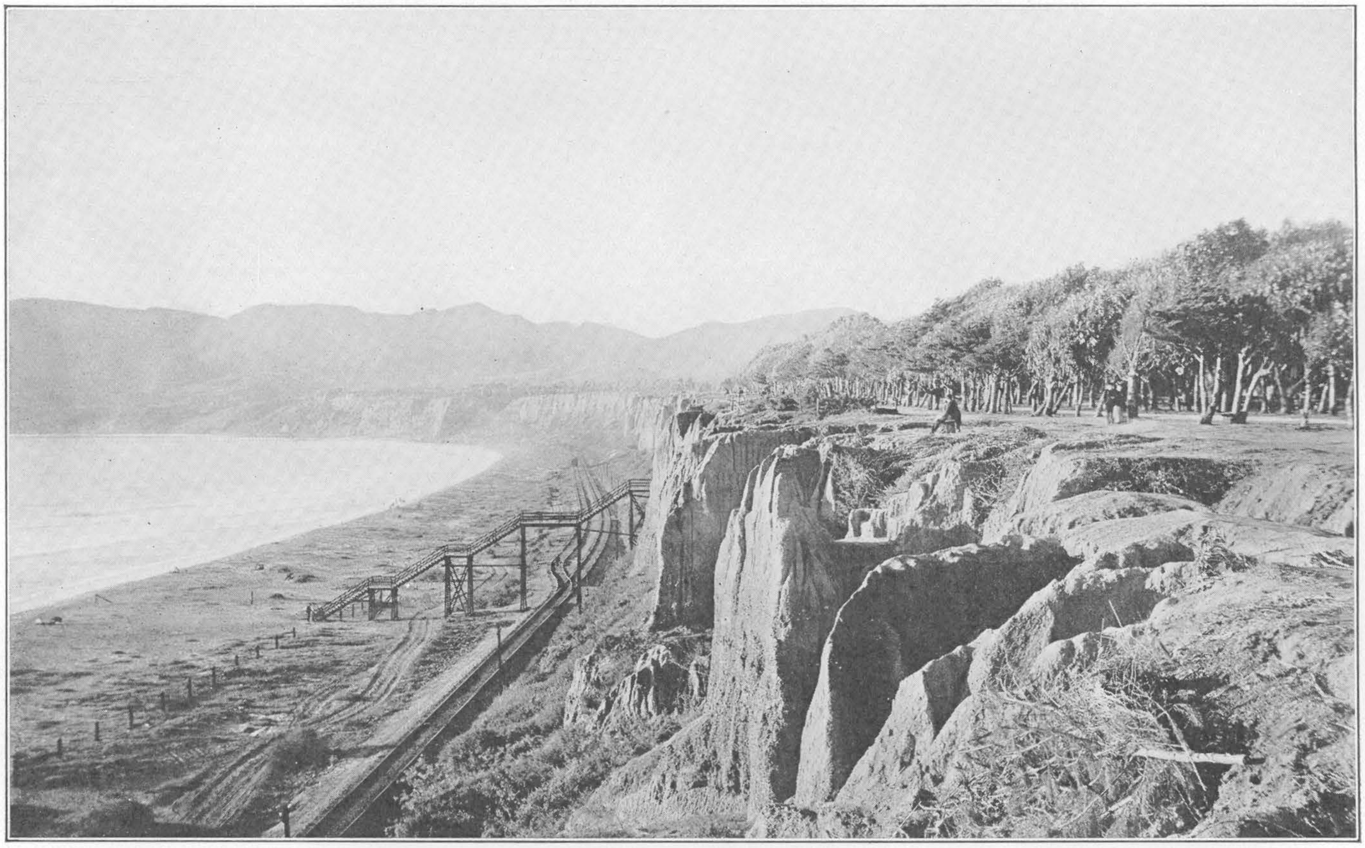
[42] Anderson, F. M., California Acad. Sci. Proc., 3d ser., vol. 2, No. 1, p. 75, pl. 9, fig. 183, 1902.



DIATOMS FROM THE UPPER MEMBER OF THE MODELO FORMATION, NORTH SLOPE OF SANTA MONICA MOUNTAINS NEAR GIRARD, LOS ANGELES COUNTY, CALIF.

Identifications and photographs by K. E. Lohman. For localities see Plate 27, *C*.

1. *Arachnoidiscus ornatus* Ehrenberg. Locality 162, about 2,900 feet above base of Modelo formation.
2. *Coscinodiscus excentricus* Ehrenberg. Locality 172, about 3,360 feet above base of Modelo formation.
3. *Coscinodiscus elegans* Greville. Locality 174, about 3,420 feet above base of Modelo formation.
4. *Coscinodiscus suboculatus* Rattray. Locality 172, about 3,360 feet above base of Modelo formation.
5. *Coscinodiscus decrescens* Grunow. Locality 172, about 3,360 feet above base of Modelo formation.
6. *Diploneis smithii* (Brebisson) Cleve. Locality 165, about 3,080 feet above base of Modelo formation.
7. *Stictodiscus californicus* Greville. Locality 160, about 2,850 feet above base of Modelo formation.



A. COASTAL BLUFFS OF PLEISTOCENE BROWN ALLUVIAL-PLAIN MATERIAL

View west from a point near mouth of Santa Monica Canyon. The bluff is about 200 feet high. Photograph by W. C. Mendenhall, 1904.

B. INDISTINCT BEDDING AND POOR SORTING COMMON IN THE PLEISTOCENE ALLUVIAL-PLAIN DEPOSITS advanced, is also very small but has slightly raised petals and a shallow notch in the posterior interambulacrum. Both forms of *Astrodapsis* are more primitive than *A. brewerianus*, and on the basis of the evidence they furnish it is concluded that the basal detrital beds of the Modelo formation are older than the Briones sandstone of the San Francisco Bay region. Inasmuch as the Briones sandstone disconformably overlies the Monterey group of the San Francisco Bay region,[43] the shales of the Modelo formation in the Santa Monica Mountains seem to be younger than the Monterey deposits near San Francisco Bay, though they may in part fall at the horizon of the hiatus between the Monterey and the Briones sandstone.

The mollusks in the detrital beds at the base of the Modelo formation represent an interesting lithologic facies, for many of them are rock clingers or nest among stones. Modern representatives of limpets (*Acmaea?*), abalones (*Haliotis*), and turbans (*Tegula*) are found on rocks. The rock scallop (*Hinnites*) attaches itself to rocks, and the arks (*Navicula*, better known as *Arca*) and file shells (*Lima*) nest among stones. These mollusks clearly lived at the foot of rock cliffs, as might be supposed from their occurrence immediately above the irregular surface of Triassic (?) slate and schist. The rarity of rock-clinging species as fossils indicates that unusual conditions are demanded for their preservation.

Doctor Woodring has also furnished the following note on mollusks from the thin-bedded shales of the Modelo formation:

"*Pecten*" *pedroanus* (Trask) (localities 4, 11, 134, 136) and an undetermined species of *Miltha?* (locality 4) are the only mollusks collected from the thin-bedded shales of the Modelo formation. "*Pecten*" *pedroanus* is found in both upper Miocene and Pliocene deposits and is closely allied to a modern species.[44] These small scallops probably lived on kelp. After death their thin shells may have remained suspended in the water for a long time before they finally settled to the bottom and thus were carried far and wide before they were buried. This mode of life and of distribution after death would account for their widespread distribution in these thin-bedded shales, in which hardly any other mollusks are found.

The vertebra of a whale and the first metacarpal from the right fore limb of a sea lion, the latter identified by Dr. Remington Kellogg, were collected at fossil locality 65 from a 10-foot sandstone at the base of the lower member of the Modelo formation near the west line of the NW. ¼ sec. 33, T. 1 N., R. 16 W. Doctor Kellogg, although expressing some uncertainty as to the specific identification of the sea-lion bone, states that it is suspiciously like the first right metacarpal of a sea lion from the "Temblor" which he has described as *Allodesmus kernensis*.

A large tooth of a shark was collected from the basal sandstone bed of the Modelo formation along the southern edge of the mountains (locality 10, on Will Rogers's ranch, just west of lower Rustic Canyon).

Dr. Chester Stock has informed the writer that fragmentary fossil remains of a horse have been found in the lowest shale member of the Modelo near the head of Sepulveda Canyon. The locality is close to locality 58 shown on Plate 16. According to Doctor Stock, these remains consist of carpal bones and limb elements that are similar to those of *Merychippus* but are of little value for age determination. They were found in marine sediments associated with fossil fish and the leaves of land plants.

Leaves of land plants (pl. 26, *B*) were found in foraminiferal platy shale in the upper part of the lower member of the Modelo formation on the Mulholland Highway at locality 57. According to Dr. Ralph W. Chaney, these leaves belong to *Platanus dissecta*, the most characteristic species of the Miocene in western America and a common member of the flora of the upper part of the San Pablo formation. Doctor Chaney states that conditions in general during Miocene time were favorable for the preservation of the remains of river-border species.

UPPER MEMBER

General features.—The upper member of the Modelo formation is composed to a large extent of soft white punky diatomaceous shale. Its exposure is restricted to the northern edge of the Santa Monica Mountains west of Universal City, and all the good outcrops lie south of the Ventura Boulevard. In the eastern part of its exposure, east of the Encino Reservoir, this member is consistent in that it is made up in greater part of diatomaceous shale and contains no prominent beds of sandstone. Farther west, however, the lower part of this punky diatomaceous shale grades westward into ordinary brown earthy shale and fine brown sandstone, so that the contact between the upper and lower members of the Modelo near Girard is not drawn at the base of the diatomaceous shale but at a horizon which is equivalent to this striking change in lithology in the area farther east. A section of the upper and lower members of the Modelo has been measured near the western edge of the area near structure section line A–A′ (pl. 16) and appears on pages 103–104. From this it will be noted that the characteristic punky diatomaceous shale does not occur in the lower part of the upper member but makes its first appearance about 1,100 feet above the base and is a prominent constituent of the upper part.

Plate 25, *C*, shows an exposure of the diatomaceous shale just south of the Ventura Boulevard a quarter of a mile east of Girard. Shale of this type is remarkably well bedded and is composed of distinct laminae and thin beds of light-gray and brown color. In other exposures thin beds are not so regular and continuous and there are strong indications of scouring action, presumably by waves or currents along the sea bottom. Bedding planes generally reveal multitudes of tiny white disk-shaped tests of marine diatoms, together with fish scales and, occasionally, the small, thin, nearly transparent shells of "*Pecten*" *pedroanus* Trask, the only megascopic invertebrate fossil found in the upper member of the Modelo. Fragments of

[43] Trask, P. D., California Univ. Dept. Geol. Sci. Bull., vol. 13, No. 5, pp. 137–138, 1922.

[44] See Arnold, Ralph, U. S. Geol. Survey Prof. Paper 47, pp. 90–91, 1906.

silicified and lignitized wood are also locally present. Other than the diatoms the most abundant microscopic fossils are Radiolaria, Foraminifera, sponge spicules, and silicoflagellates. Calcareous Foraminifera are not as common as in the lower member of the Modelo, but in some thin beds or partings well-preserved forms representing several dozen species occur by the millions.

This member contains numerous thin beds of light-gray volcanic ash, similar in index of refraction to that in the lower member of the Modelo, and lenses and beds, as much as 3 feet thick, of hard limestone that weathers yellowish brown.

The contact between the lower and upper members of the Modelo is well exposed in several road cuts that connect the Mulholland Highway with the Ventura Boulevard, and except for the occurrence at one locality of a 1-inch bed of conglomerate suggestive of a local break in sedimentation and minor erosion this contact is apparently one of conformable stratigraphic relations. For instance, on the road leading north up the short ridge just east of the Hollywood Country Club punky diatomaceous shale of the upper member grades downward into hard brown platy shale of the lower member through a stratigraphic interval of 30 feet occupied by alternating thin beds of white punky diatomaceous and hard brown platy shale. The 1-inch bed of conglomerate occurs here within this gradational sequence.

Figure 8 shows the results of mechanical analyses of two sandstone samples collected from different thin beds intercalated with the finely laminated punky diatomaceous shale of this member. The finer sample, illustrated by curve 7, is more representative of the character of the sandstone associated with this shale. The presence of these intercalated sandstone beds is in itself important in providing evidence that these diatomaceous deposits accumulated on the continental shelf at depths where currents or waves were capable of moving bottom sediment. According to Barrell,[45] wave base, or the depth at which wave action ceases to be strong enough to transport bottom sediment, is generally 50 fathoms in the ocean. Currents, however, according to several investigators cited by Twenhofel,[46] are in some places sufficiently strong to transport sand at depths of 250 meters (about 137 fathoms). It seems probable that such sand would be fine and fairly well sorted as to size of grains, like the sample represented by curve 7 in Figure 8. However, the coarseness and rather poorly sorted character of the sample represented by curve 2 suggests that some of these diatomaceous deposits accumulated at depths considerably less, possibly less than 50 fathoms. The probability of a shallow-water origin for at least some diatomaceous deposits is also supported by the recent discovery of diatom epidemics off the coast of Washington.[47]

Microscopic character of shale.—In thin sections the diatomaceous shale of the upper member of the Modelo, except for the abundance of the opaline tests of diatoms and various kinds of spicules, appears much like the hard platy shale in the lower member of this formation. Cryptocrystalline silica, however, composes most of the groundmass, there commonly being a much smaller proportion of true opal. In most of the thin sections banding is not so distinct in this shale, the most common type of which is produced by the presence, locally in considerable abundance, of thin cloudy, nearly opaque bands which are white in reflected light and which alternate with more nearly transparent bands of cryptocrystalline silica. Other banding results from the presence in some sections of pale-brown bands containing a small amount of organic matter, and, less commonly, the presence of thin colorless bands of opal or of thicker bands which consist largely of angular detrital grains of quartz and plagioclase feldspar as much as 0.10 to 0.15 millimeters in diameter. Most of the diatomaceous shale contains detrital mineral grains of about this maximum size scattered throughout the groundmass of cryptocrystalline silica. Green and brown flakes of chlorite and biotite are abundant along some of the bedding planes.

Some sections are literally flooded with entire and broken tests of diatoms and spicules of sponges and Radiolaria. Radiolaria are generally common, but in none of the thin sections of the punky diatomaceous shale are they nearly as abundant as the diatoms. Small rounded or oblong bodies which appear to be sponge globules are abundant in some sections. Calcareous Foraminifera, associated with more abundant diatoms, make up a considerable part of some of the sections and have been found to be abundant in laminae of some of the shale outcrops. Other calcareous shells, invariably flattened and crushed, with elongation parallel to the bedding, are also very common and materially assist in accentuating banding in some of the sections. These shells are indeterminable, but probably are the fossil remains of pelecypods, inasmuch as such shells have been observed along bedding planes in hand specimens.

Fossils and age.—Foraminifera collected from both the lower and upper members of the Modelo, as exposed in the western part of the area along the road between Girard and Garrapata Canyon (localities 111–174), have been prepared and identified by W. D. Rankin, of Los Angeles. In order to obtain a check upon his results, Mr. Rankin submitted his prepared material to D. D. Hughes and P. P. Goudkoff for examination. The identification and distribution of the numerous species of Foraminifera are presented in the following table, prepared by Mr. Rankin:

[45] Barrell, Joseph, Rhythms and the measurements of geologic time: Geol. Soc. America Bull., vol. 28, p. 778, 1917.

[46] Twenhofel, W. H., Treatise on sedimentation, pp. 474–475, 1926.

[47] Becking, L. B., Tolman, C. F., McMillin, H. C., Field, John, and Hashimoto, T., Preliminary statement regarding the diatom "epidemics" at Copalis Beach, Washington, and an analysis of diatom oil: Econ. Geology, vol. 22, pp. 356–368, 1927.

Distribution of Foraminifera throughout section of Modelo formation exposed along road between Girard and Mohn Springs

[Prepared by W. D. Rankin. /, very rare; +, rare; ×, fairly common; ○, common; ●, abundant. For localities and stratigraphic position of samples see pl. 27, *A*, *C*.]

No.	Sample	174	173	172	155	152	151	150	148	147	146	145	144	142	139	138	136	135	134	133	132	131	130	129	128	127	126	125	124	123	122	121	120	119	117	115	114	113
1	Anomalina sp.															○	/					/		×														
2	Bolivina beyrichi Reuss var. alata (Seguenza)											/				+	+		+	×	+	+	+	×	+		/	○	+	×	○	○	+	/				
3	Bolivina brevior Cushman																												+	×	/		+				×	+
4	Bolivina decurtata Cushman															+			+	×		+	○	+	+						/							
5	Bolivina aff. B. dilatata Reuss											+		×	+	+	×	×	+		+	+	+	+	+	+	+	○					+			+	×	×
6	Bolivina hughesi Cushman					×		+			+						○	○	+	×	○	○		○	○	○	○											
7	Bolivina seminuda Cushman var. foraminata R. E. and K. C. Stewart	+	+	+			+			/				/	+																							
8	Bolivina sinuata Galloway and Wissler	●	●	●	/	○		○	+			×		○		×									/					/								
9	Bolivina sp.	○	○	○																																		
10	Bolivina aff. B. spissa Cushman	+	+								+																											
11	Bolivina sp.					+	+	+											/									+					×					
12	Bolivina sp.							+	○	×	○				×																							
13	Bulimina affinis D'Orbigny of Brady											○				/	○	×		×	+	○			+	/												
14	Bulimina inflata Seguenza																+	+				/	+					×										
15	Buliminella brevior Cushman																													+			+					×
16	Buliminella californica Cushman																○				/	+	○	+				+			/		/					
17	Buliminella curta Cushman	×	○	×																						/		+										
18	Buliminella subfusiformis Cushman	+		+															×				/															
19	Buliminella sp.	×	+	+																												+						
20	Buliminella sp.																					+		+														
21	Cassidulina aff C. californica Cushman and Hughes																				/	+	+						+	+		/						/
22	Cassidulina delicata Cushman	○	+	○																																		
23	Cassidulina quadrata Cushman and Hughes													/																								
24	Cassidulina cf. C. translucens Cushman and Hughes	+	/	+					●	○	/				×	○						+						○										
25	Cassidulina sp.								+						+	+												×										
26	Discorbis sp.				+								/																									
27	Eponides healdi R. E. and K. C. Stewart								○	○	○			/	+	×	○					/																
28	Eponides aff. E. broeckiana (Karrer)																+	×	/		/			/				+		/	+	/	/		/			/
29	Eponides tenera (Brady)								+															/				+										
30	Frondicularia foliacea Schwager													/		/						×	/	+														
31	Globigerina bulloides D'Orbigny	○	/	○	○								+						×	×	×	+	×		+	+	×	+	×	×		/			+	+	×	+
32	Globigerina cyclostoma Galloway and Wissler			/	+															+			/															
33	Globigerina quadrilatera Galloway and Wissler			+									/						×	×	+		/															
34	Globigerina cf. G. conglomerata Schwager			+																																		
35	Globobulimina pacifica Cushman		/																													/						
36	Gyroidina soldanii var. rotundimargo R. E. and K. C. Stewart							×	+	+		+		+		+	○	×				○	×	○	×		×	×			+	+			/			
37	Nodogenerina sp.							+																														
38	Nodosaria koina Schwager				+								/	/						+										/								+
39	Nodosaria tympaniplectriformis Schwager							+				+		/		/	/	+				+	×					+										
40	Nodosaria sp.													/								+	/															
41	Nodosaria sp.															+	/					+	/		/													
42	Nonion cf. N. umbilicatula (Montagu) var. pacifica Cushman																											/	+	×	/	/	×					
43	Nonion sp.													×		/	+				+	/		+	+													
44	Nonion? sp.																											/										
45	Orbulina universa D'Orbigny			/																																	+	
46	Planulina ornata (D'Orbigny)								×	×	×			×	/	/	○					+	/	+				×	/	/		/						
47	Planulina sp.												○	/							+																	
48	Polymorphina sp.											/																										
49	Pulvinulinella cf. P. bradyana Cushman	+	+	+																																		
50	Pulvinulinella sp.																														+	/	+	/		/	+	×
51	Pulvinulinella cf. P. pacifica Cushman	○	+	+																																		
52	Pullenia sp.																												/				/					
53	Quinqueloculina sp.																								/													
54	Robulus nikobarensis Schwager															/																						
55	Robulus sp.																					/																
56	Sphaeroidina variabilis Reuss									/						+								+	+			+										
57	Uvigerina sp.	+	×	+																																		
58	Uvigerina sp.			+	/				+		/					+	/	+							+		/				/			/				/
59	Uvigerina sp.				+	○			+	+		+		+		+		/				+	+		+	/	/						+					
60	Uvigerina sp.				+															/																		
61	Uvigerina sp.					+	+	×															+	/														
62	Uvigerina cf. U. striata D'Orbigny						+	+	○	○	+	+		+	+	+						×																
63	Uvigerina sp.									/		×		○		○	/			/		×	/	+														
64	Uvigerina sp.																+					+		+		+		+	+	+							+	/
65	Virgulina cf. V. californiensis Cushman																+		/	+			/								+							
66	Valvulineria araucana (D'Orbigny)				/				+							+	○				/	○	○	+				○						/	+			×

Mr. Rankin's conclusions, concurred in by Messrs. Hughes and Goudkoff, are as follows:

The foraminiferal fauna of the lower member of the Modelo is characterized by an assemblage of such species as *Bolivina beyrichi* Reuss var. *alata* (Seguenza), *Bolivina hughesi* Cushman, *Bolivina decurtata* Cushman, *Bolivina* aff. *B. dilatata* Reuss, *Bulimina affinis* D'Orbigny of Brady, *Buliminella californica* Cushman, *Buliminella* sp., and *Eponides* aff. *E. broeckiana* (Karrer). This fauna has been found in the upper shale member of the Puente formation as exposed in La Habra Canyon on the south flank of the Puente Hills.

The foraminiferal fauna of the upper member of the Modelo, as found in collections 172, 173, and 174, is characterized by an assemblage of such species as *Bolivina seminuda* Cushman var. *foraminata* R. E. and K. C. Stewart, *Bolivina sinuata* Galloway and Wissler, *Bolivina* aff. *B. spissa* Cushman, *Bolivina* sp., *Buliminella* sp., *Cassidulina delicata* Cushman, *Pulvinulinella bradyana* Cushman, and *Pulvinulina* cf. *P. pacifica* Cushman. This fauna may be the age equivalent of the upper shale member of the Modelo as mapped by Kew in Modelo Canyon, a stratigraphic zone which directly underlies shale considered by Hudson and Craig to be of Santa Margarita age. The faunas of these two widely separated areas of upper Miocene rocks have noteworthy affinities; the upper shale member of the Modelo of Modelo Canyon, however, contains arenaceous Foraminifera which were not detected in the upper Modelo of the Girard area.

Samples collected from localities 303, 304, and 107 in Potrero Canyon, along the coast near Santa Monica, came from small fault blocks of Modelo shale and, according to Goudkoff, Hughes, and Rankin, contain Foraminifera which permit them to be correlated with the middle part of the lower member of the Modelo. Samples 104 and 105, obtained near the mouth of Temescal Canyon, contain no distinctive fossils, but sample 106, from the same area, contains a fauna which Rankin says belongs to the lower member of the Modelo. The sample from locality 110, near the base of the Modelo on the south flank of the mountains half a mile north of the Beverly Hills Hotel (see pl. 16), likewise contains a foraminiferal fauna which, according to Rankin, may be correlated with the lower member of the Modelo in the Girard and Garrapata Canyon section.

A duplicate suite of samples from the Girard and Garrapata Canyon section of the Modelo was examined for diatoms by K. E. Lohman, of the California Institute of Technology. Mr. Lohman has prepared the following table showing the names and distribution of diatoms found in this section, also Plate 28, showing several of the species of diatoms.

Modelo diatoms from road leading south from Ventura Boulevard on first ridge east of Girard

[Identified by K. E. Lohman. For localities and stratigraphic position of samples 160–174 see pl. 27, *A*, *C*. /, very rare; +, rare; ×, fairly common; ○, common; ●, abundant]

		174	173	172	171	170	169	168	167	166	165	164	163	162	161	160
1	Actinoptychus undulatus Ehrenberg var.			×												
2	Arachnoidiscus ornatus Ehrenberg	/		/										/		
3	Chaetoceros incurvum Bailey			/												
4	Cladogramma californicum Grunow	+		×												
5	Cocconeis baldjikiana Grunow			+		+	+									
6	Cocconeis distans (Gregory) Grunow		/													
7	Coscinodiscus argus Ehrenberg														+	
8	Coscinodiscus asteromphalus Ehrenberg	×		×	+		+									
9	Coscinodiscus decrescens Grunow			×												
10	Coscinodiscus denarius Schmidt		×	×		×										
11	Coscinodiscus elegans Greville	×		+		+	/								/	
12	Coscinodiscus excentricus Ehrenberg	×		×										×		
13	Coscinodiscus lineatus Ehrenberg			/												
14	Coscinodiscus marginatus Ehrenberg	/		×		×					+			×		
15	Coscinodiscus oculus-iridis Ehrenberg			×											×	
16	Coscinodiscus radiatus Ehrenberg	×		+												
17	Coscinodiscus subtilis Ehrenberg			+												
18	Coscinodiscus woodwardii Eulenstein			×												
19	Cyclotella antiqua Smith?			+		+										
20	Denticula lauta Bailey	+	+	/	+				+						+	
21	Dicladia capreolus Ehrenberg			×												
22	Diploneis densistriata (Schmidt)		/	+			/									
23	Diploneis smithii (Brebisson) Cleve	/				+					/					
24	Diploneis suborbicularis (Gregory)			+		+										
25	Endictya robustus (Greville)	×		×	+			×	○	○	○	○	×	×	×	
26	Fragilaria arctica Grunow			+												
27	Grammatophora angulosa Ehrenberg			/												
28	Grammatophora merletta Hanna and Grant			/					+					/		
29	Hemidiscus simplissimus Hanna and Grant	+	+	×								+				
30	Hercotheca mammilaris Ehrenberg			+												
31	Liradiscus ovalis Greville	×		○	+	+	+					×		○		
32	Melosira exspectata Schmidt			+									+	/	+	+
33	Melosira sol (Ehrenberg) Kutzing					+						+		/		
34	Melosira sulcata (Ehrenberg) Kutzing	+	+	+		×								/		
35	Navicula lyra Ehrenberg	+		+												
36	Omphalotheca californica Hanna			×		×								+		
37	Opepora schwartzii Ehrenberg			/												
38	Periptera tetracladia Ehrenberg			×												
39	Pyxidicula cristata Ehrenberg			/												
40	Rhaphoneis archeri O'Meara	/														
41	Stephanogonia actinoptychus Ehrenberg			+												

Modelo diatoms from road leading south from Ventura Boulevard on first ridge east of Girard—Continued

		174	173	172	171	170	169	168	167	166	165	164	163	162	161	160
42	Stephanopyxis appendiculata Ehrenberg														×	
43	Stephanopyxis turris (Gregory) Ralfs														×	
44	Stictodiscus californicus Greville			/											+	
45	Syndendrium diadema Ehrenberg			○												
46	Synedra acuta Ehrenberg	+		×												
47	Synedra affinis Kutzing?													×		
48	Synedra nitzschioides Grunow			●		●			×	+	+					
51	Xanthiopyxis cingulata Ehrenberg	×		○				+								
52	Xanthiopyxis oblonga Ehrenberg													○		
53	Xanthiopyxis umbonatus Greville			×												

The following statements are made by Mr. Lohman:

No diatoms were found in the lower 2,850 feet (between samples 111 and 159). Beginning with No. 160, however, where one diatom was found, and continuing through to No. 174, the highest sample collected, diatoms were found consistently. Samples 155 to 174 all contained sponge spicules, Radiolaria, and silicoflagellates.

Endictya robustus (Greville) was the first diatom to become fairly common in going up in the section; it became common in samples 164 to 167, and then less common in succeeding samples.

In samples 163 to 168 very little diversity of form of the diatoms was noted, there being only two or three species present in any amounts above rare. Samples 169 and 170 showed a considerable increase in diversity, but sample 171 contained relatively few species and individuals. In sample 172, however, a very great and sudden increase to 45 species was observed, of which 1, *Synedra nitzschioides* Grunow, is abundant; 3, *Xanthiopyxis cingulata* Ehrenberg, *Syndendrium diadema* Ehrenberg, and *Liradiscus ovalis* Greville, are common; and 17 are fairly common. Sample 173 showed a decided decrease in number of species as well as in abundance, there being only 6 species reported. Sample 174, the last collected, although very diatomaceous, has only 17 species. This great fluctuation has two possible explanations. Either it represents a change in the actual deposition of the organisms, or it means that the samples are not truly representative.

In general the diatom flora in this section is characterized by small forms, many of which are broken. It is nothing like the flora from the "Temblor" formation of Shark Tooth Hill, nor is it like Monterey from the type locality.

In addition to the species listed in the table, there are other species, such as *Coscinodiscus suboculatus* Rattray, illustrated in Plate 28, and several species which are thought to be new.

PLIOCENE ROCKS

Rocks of Pliocene age are absent in surface outcrops over most of the area covered by this report, being exposed only in the coastal belt that lies south of the mountains and northwest of the city of Santa Monica. They are probably very extensive beneath the alluvial plains that border the mountains on the south and on the north, for they not only crop out in the Baldwin Hills and in isolated patches south of the mountains and in the City of Los Angeles, but have been penetrated by wells drilled in the Inglewood (Baldwin Hills), Beverly Hills, and Salt Lake oil fields and by scattered wildcat wells of the adjoining lowlands.

It is impossible to determine from the field relations the relative age of most of the isolated sections of Pliocene and Pleistocene rocks exposed in the coastal areas north and northwest of Santa Monica. According to W. P. Woodring's interpretation of the evidence provided by megascopic fossils from beds of these areas, the stratigraphic order for the marine Pliocene and Pleistocene of this region that contain such fossils is as follows:

Age and stratigraphic order of marine Pliocene and Pleistocene rocks

[Based entirely on fossil evidence obtained by W. P. Woodring. For fossil localities see pp. 123–124 and pl. 16]

Age	Locality	Correlation
Late Pleistocene	Overland Avenue (locality 68); upper Potrero Canyon (locality 61).	Upper San Pedro.
Lower Pleistocene or upper Pliocene.	Conglomerate at mouth of Potrero Canyon (locality 31); storm-drain ditch in Rustic Canyon (locality 311).	Cool-water fauna of Timms Point and Deadman Island.
Upper Pliocene	Upper part of Potrero Canyon section (locality 61A)	"*Arca*" *camuloensis* fauna of Los Angeles Basin and Santa Clara Valley.
Middle or upper Pliocene	Temescal Canyon (locality 55)	San Diego fauna.

All the Pliocene rocks from which megascopic fossils were collected in this area are considered by Woodring to be younger than the Pliocene beds of Elsmere Canyon (lower Pliocene) in Ventura County. According to foraminiferal evidence obtained by D. D. Hughes and P. P. Goudkoff from Potrero Canyon there are also beds in this area which are of lower Pliocene age but these beds contain no megascopic fossils.

POTRERO CANYON

Most of the exposed Pliocene section occurs in and near Potrero Canyon, near the northwestern edge of

Santa Monica. It is fairly uniform throughout and is composed of soft medium to dark gray clay shale and sandy clay shale which are rich in well-preserved shells of calcareous Foraminifera. This section has a maximum exposed thickness of about 1,000 feet. Bedding appears to be fairly distinct, but the rock lacks the firm, thin-bedded, highly fissile character common to the Miocene shales of this district. Lenses and beds of hard gray and yellowish-brown limestone are scattered throughout the clay shale, and beds of breccia have been noted which are composed of angular fragments of limestone that apparently have been derived from the underlying Miocene. These Pliocene rocks are faulted against Modelo shale at several places in Potrero Canyon.

The upper part of this section is equivalent to a part of the Pico formation at its type locality in Pico Canyon. The lower part is equivalent to foraminiferal lower Pliocene beds in Ventura and Los Angeles Counties which, in areas distant from the type section of the Pico formation, are known to underlie the Pico formation and overlie Miocene strata. The type Pico is now considered by many active workers in California stratigraphy and paleontology to be of upper Pliocene age, and the underlying Pliocene rocks to be of lower Pliocene age. For these stratigraphically lower Pliocene strata, which were not included by Kew [48] in the type section of the Pico, and, according to some paleontologists, for some additional overlying beds that occur in the lower part of the type section of the Pico, Eaton [49] has suggested the name Santa Paula formation.

Another exposure of Pliocene sandy gray clay shale occurs on the coast just west of the mouth of Potrero Canyon. This shale dips 20°–30° SE. and appears to rest directly upon vertical or nearly vertical brown Miocene shale in the bluff at the western edge of the exposure. It is possible, however, that the apparent relations have resulted from displacement along a westward extension of the fault that cuts across Potrero Canyon only a few hundred feet above its mouth. (See pl. 16.) At the mouth of this canyon almost horizontal marine upper Pliocene or lower Pleistocene beds are faulted down against tilted lower Pliocene clay shale.

The following fossils were collected at the top of the northward-dipping section of Pliocene clay shale near the head of Potrero Canyon (locality 61A) a quarter of a mile south of Pacific Palisades post office. The bed from which they came occurs directly beneath soft white flat-lying sand of Pleistocene age that also carries a rich marine invertebrate fauna. W. P. Woodring has identified these Pliocene fossils, as follows:

Bryozoon:
- Cellaria sp.

Gastropods:
- Calliostoma canaliculatum (Martyn)?
- Calliostoma costatum (Martyn).
- Homalopoma paucicostata fenestrata (Bartsch).
- Turbonilla, 3 spp.
- Odostomia sp.
- Epitonium sp.
- Tectonatica clausa (Broderip and Sowerby).
- Neverita recluziana (Deshayes).
- Crepidula onyx Sowerby.
- Alvania sp.
- Lacuna sp.
- Turritella, 2 spp.
- Bittium, 3 spp.
- Tritonalia barbarensis (Gabb).
- Tritonalia cf. T. lurida (Middendorf).
- Boreotrophon cf. B. orpheus (Gould).
- Boreotrophon sp.
- Mitrella carinata (Hinds).
- Mitrella tuberosa (Carpenter).
- Amphissa sp.
- Nassarius perpinguis (Gould)?
- Nassarius insculptus (Carpenter).
- "Cantharus" fortis (Carpenter).
- Barbarofusus traski (Dall).
- Olivella biplicata Sowerby.
- Admete sp.
- Conus californicus Hinds.
- Antiplanes perversa (Gabb).
- Antiplanes pedroana (Arnold).
- Mangilia sp.
- Philbertia sp.
- Lora sp.
- Acteon cf. A. punctocoelatus (Carpenter).

Scaphopods:
- Dentalium neohexagonum Sharp and Pilsbry.
- Dentalium sp.
- Cadulus fusiformis Pilsbry and Sharp?
- Cadulus cf. C. californicus Pilsbry and Sharp.

Lamellibranchs:
- Saccella taphria (Dall).
- Nuculana sp.
- Yoldia sp.
- "Pecten" cerrosensis Gabb.
- "Pecten" latiauritus Conrad.
- "Pseudamusium" sp.
- Hinnites sp.
- Crenella sp.
- Cyclocardia ventricosa (Gould).
- Thyasira sp.
- Thyasira gouldii (Philippi).
- Axinopsis viridis Dall.
- Lucinoma annulata (Reeve).
- Parvilucina tenuisculpta (Carpenter).
- Rochefortia sp.
- Trachycardium quadrigenarium (Conrad).
- Chione sp.
- Venerupis cf. V. staminea (Conrad).
- Psephidia ovalis Dall.
- Psephidia cymata (Dall)?
- Tellina sp.
- Macoma sp.
- Solen sicarius Gould?
- Corbula luteola Carpenter.
- Panope generosa Gould.

[48] Kew, W. S. W., Geology and oil resources of a part of Los Angeles and Ventura Counties, Calif.: U. S. Geol. Survey Bull. 753, p. 70, pl. 1, 1924.

[49] Eaton, J. E., The Fernando group of southern California, paper read before the Pacific section of Am. Assoc. Petroleum Geologists at Los Angeles Oct. 29, 1926. Eaton later used this formation name in the following publications: Ventura field controlled reservoirs: Oil and Gas Jour., Nov. 11, 1926, p. 72; The by-passing of discontinuous deposition of sedimentary materials: Am. Assoc. Petroleum Geologists Bull., vol. 13, No. 7, fig. 10, 1929.

Doctor Woodring has contributed the following statement regarding this collection:

This fauna is considerably younger than the San Diego fauna and probably is of about the same age as the *"Arca" multicostata* or *"Arca" camuloensis* fauna found in the Puente Hills,[50] in the Los Angeles Third Street tunnel,[51] and in the Santa Clara Valley.[52] Like the *"Arca" camuloensis* fauna, it embraces species of warm-water facies (*"Cantharus" fortis, "Pecten" cerrosensis, Parvilucina tenuisculpta, Trachycardium quadrigenarium, Chione, Corbula luteola*) and also a few species of cool-water aspect (*Boreotrophon, Lora, Thyasira*). It is regarded as heralding the approach of a cooler climate, but not so cool as that of the fossils found in the excavation for the Richfield Oil Co.'s building at Sixth and Flower Streets and in the bluff on Fifth Street between Grand and Flower Streets, both in Los Angeles. On an assumption of progressive lowering of temperature during the later part of Pliocene time the fauna from Potrero Canyon is older than the fossils in Los Angeles just mentioned. This age assignment is substantiated by the presence of at least one distinctive extinct San Diego species—*"Pecten" cerrosensis* Gabb (*"ashleyi* Arnold"). Others (*Ostrea "veatchii"* and *"Pecten" healeyi*) are recorded from the *"Arca" camuloensis* fauna.

P. P. Goudkoff, Boris Laiming, and D. D. Hughes have examined for Foraminifera samples collected in Potrero Canyon and in other areas in the vicinity of Santa Monica. (See pl. 27, *B*.) They have identified the species present and have prepared the following table:

Foraminifera collected from Miocene, Pliocene, and Pleistocene formations near Santa Monica, Calif.

[Identifications and age determinations by P. P. Goudkoff, Boris Laiming, and D. D. Hughes. Localities are plotted on pl. 27, *B*, and described on p. 124. a, Abundant; c, common; r, rare]

	311	312	310	309	308	307	306	305	302	301	303	304	Age
Bolivina aff. B. spissa Cushman	r												Uppermost Pliocene or Pleistocene
Cassidulina californica Cushman and Hughes	c	a	r		r			c		a	c		
Cassidulina corbyi Cushman and Hughes	c		r	r									
Cassidulina limbata Cushman and Hughes	a							r?					
Cassidulina pulchella D'Orbigny	a	r											
Cassidulina subglobosa H. B. Brady	r							r	r	c			
Cibicides aff. C. lobatus (D'Orbigny)	c												
Cibicides sp	r												
Dentalina fragment	r												
Elphidium granulosum (Galloway and Wissler)	r												
Fissurina laevigata Reuss	r	r							r	r			
Globigerina bulloides D'Orbigny	a	c			c	r.	r			r	r		
Globobulimina pacifica Cushman	r?		r										
Lagena sp.	r												
Nonion scapha (Fichtel and Moll)	r												
Nonion stelligera (D'Orbigny)	r												
Pulvinulinella pacifica Cushman	r	c	c	r	c		r		r	r			
Uvigerina sp	r												
Ostracode fragment	r												
Echinoid spines	r												
Bolivina interjuncta Cushman		a	a	a				r					Upper Pliocene
Bolivina robusta H. B. Brady		a		r	r		r	r		r			
Bolivina seminuda Cushman		r		r									
Bolivina spissa Cushman		r	a		a								
Bulimina ovata D'Orbigny		c											
Cassidulina aff. C. delicata Cushman		r	r	r			r						
Cassidulina reflexa Galloway and Wissler		r											
Cassidulina translucens Cushman and Hughes		c	r	r	c	c	c	r	a	c			
Cibicides conoideus Galloway and Wissler		c											
Frondicularia advena Cushman		c	r	r	c			r	r	r			
Gaudryina arenaria Galloway and Wissler		r	r										
Globigerina concinna Reuss		r	r	a	r	c	r			c			
Globigerina inflata D'Orbigny		r											
Globigerina quadrilatera Galloway and Wissler		c				r	r	r					
Gyroidina soldanii D'Orbigny var. rotundimargo R. E. and K. C. Stewart		r						r					
Lagena acuticostata Reuss		r											
Nodosaria arundinea Schwager		c	c							r			
Nodosaria aff. N. soluta (Reuss)		c	r	r	r								
Nodosaria tosta Schwager		r											
Orbulina universa D'Orbigny		r											
Planulina ornata D'Orbigny		r	r			r	r	r	c	c	r		
Pulvinulinella bradyana Cushman		c	r			c	r	c	r	c			
Uvigerina peregrina Cushman		a	a	a	a				r				
Uvigerina peregrina Cushman var. bradyana Cushman		c		r	c					r			
Uvigerina pygmaea D'Orbigny		r								r			
Cassidulina aff. C. tortuosa Cushman and Hughes			r										
Chilostomella ovoidea Reuss			r										
Eponides tenera (H. B. Brady)			c	r	c	r				r			
Globigerina pachyderma (Ehrenberg)			r										
Globobulimina aff. C. pyrula (D'Orbigny)			r										
Nodogenerina sp			r										
Planulina sp. A			c	r	c								
Pullenia salisburyi R. E. and K. C. Stewart			r	r	r								
Sphaeroidina bulloides D'Orbigny			r	r	r	r				r	r		
Uvigerina sp			c										
Uvigerina sp			r										
Uvigerina sp			c										

[50] Eldridge, G. H., and Arnold, Ralph, The Santa Clara Valley, Puente Hills, and Los Angeles oil districts, southern California: U. S. Geol. Survey Bull. 309, p. 107, 1907.

[51] Idem, p. 152.

[52] Idem, p. 24.

Foraminifera collected from Miocene, Pliocene, and Pleistocene formations near Santa Monica, Calif.—Continued

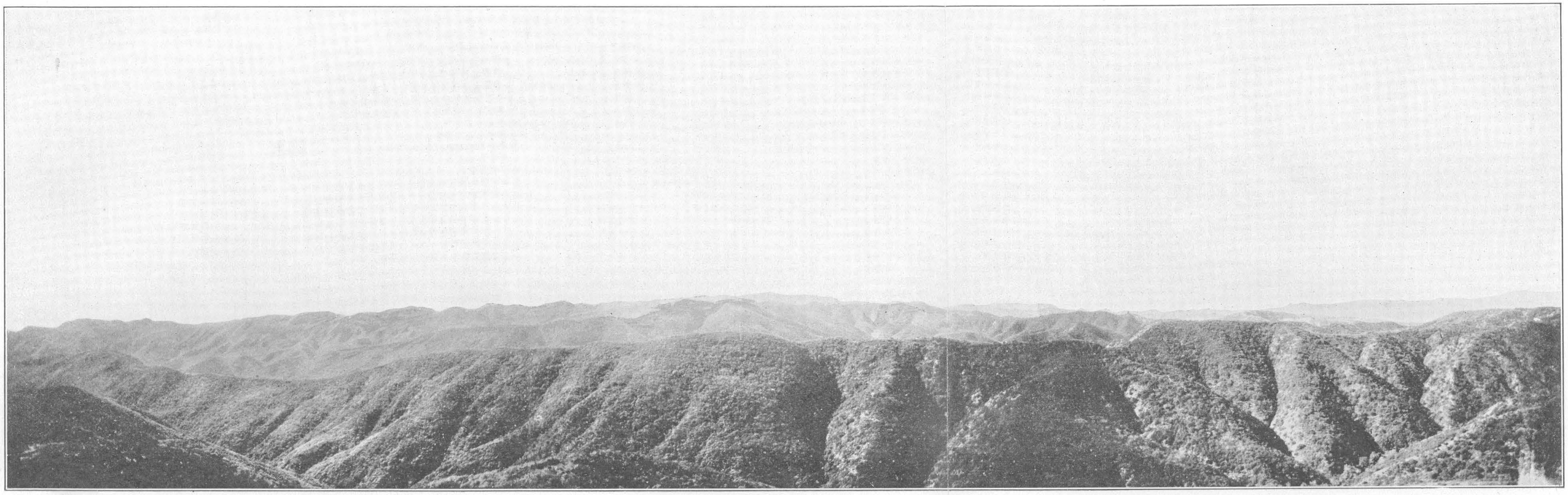
	309	308	307	306	305	302	301	303	304	Age
Bolivina aff. B. interjuncta Cushman	c		c	c	r	r				Lower Pliocene
Bulimina affinis D'Orbigny	r									
Cassidulina tortuosa Cushman and Hughes	r									
Cassidulina aff. C. laevigata D'Orbigny	r	r			r	r	r			
Cibicides?	r									
Nodosaria aff. N. arundinea Schwager	r									
Planulina aff. P. ornata (D'Orbigny)	r									
Uvigerina urnula D'Orbigny	r									
Uvigerina aff. U. pygmaea D'Orbigny	r		r	c	c	c				
Valvulineria araucana (D'Orbigny)	r		r					r	r	
Buliminella elegans (D'Orbigny) var. exilis (H. B. Brady)		r					r			
Eponides subtenera (Galloway and Wissler)		r	r							
Uvigerina peregrina Cushman var. parvula Cushman		r					r			
Bolivina aff. B. argentea Cushman			r							
Bolivina sinuata Galloway and Wissler			c	c		r		r		
Buliminella aff. B. curta Cushman var. basispinata R. E. and K. C. Stewart			c	r						
Buliminella sp			r							
Cassidulina quadrata Cushman and Hughes			r							
Nodogenerina sp			r							
Sigmolina elliptica Galloway and Wissler			r							
Uvigerina sp			c							
Nodosaria sp.?				r						
Uvigerina aff. U. urnula D'Orbigny				c						
Uvigerina sp				r						
Cibicides mckannai Galloway and Wissler					r		c			
Epistomina bradyi Galloway and Wissler					r					
Nodogenerina antillea (Cushman)					r	c	r			
Nonion pompilioides (Fichtel and Moll)					r	c	r			
Robulus rotulatus (Lamarck)					r	r	r			
Uvigerina sp					c					
Bolivina sp						r				
Bulimina inflata Seguenza						c	r		r	
Bulimina aff. B. inflata Seguenza						r				
Cibicides sp						r	c			
Cibicides aff. C. haidingeri (D'Orbigny)						c				
Gyroidina aff. G. soldanii D'Orbigny						r	r			
Pullenia sphaeroides (D'Orbigny)						r				
Radiolaria sp. a						r	r	c		
Radiolaria sp. b						r				
Fish teeth						r		c		
Bolivina argentea Cushman							r	r		
Bolivina sp							r			
Bulimina pulchella D'Orbigny							r			
Clavulina sp							r			
Fissurina sp							r			
Nodosaria parexilis Cushman and K. C. Stewart							r			
Bolivina advena Cushman								r		Upper Miocene
Bolivina aff. B. marginata Cushman								r	r	
Bolivina sp								r		
Bolivina hughesi Cushman								r	r	
Bolivina aff. B. plicata D'Orbigny								r	r	
Buliminella curta Cushman								r		
Eponides aff. E. broeckhiana (Karrer)								c	r	
Eponides sp								r	r	
Nodosaria koina Schwager								r		
Planulina sp. b								c		
Uvigerina sp								c	r	
Uvigerina sp								r		
Diatoms								r		
Radiolaria sp. c								a	r	
Radiolaria sp. d								r		
Buliminella brevior Cushman									c	
Bolivina aff. B. advena Cushman									c	
Bolivina decurtata Cushman									r	
Cassidulina sp									r	
Pulvinulinella sp									r	

Messrs. Hughes and Goudkoff have submitted the following statement concerning the Foraminifera listed in the table:

Samples 303 and 304 are of upper Miocene age and may be tentatively correlated with the middle part of the lower member of the Modelo as exposed along the road between Garrapata Canyon and Girard, or that part of the section between localities 120 and 140. (See pl. 27, *C*.)

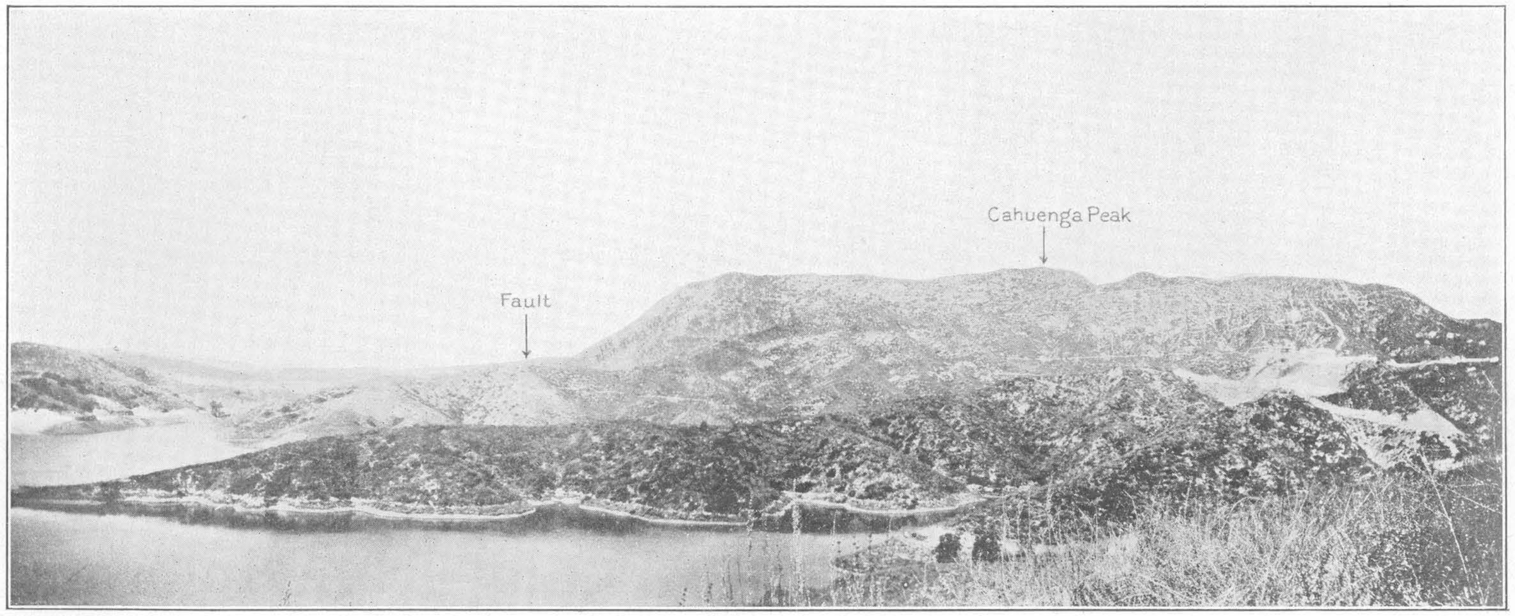
Samples 301, 302, 305, 306, and 307 are considered to be of lower Pliocene age and contain a fauna that appears to be equivalent to that found in producing wells of the Seal Beach oil field between depths of 3,000 and 4,000 feet. They also may be correlated with beds in Adams Canyon, Ventura County, that lie 12,000 to 14,000 feet stratigraphically below the so-called Saugus-Pico contact.

Samples 308, 309, and 310 are considered to be of upper Pliocene age and may be tentatively correlated with strata



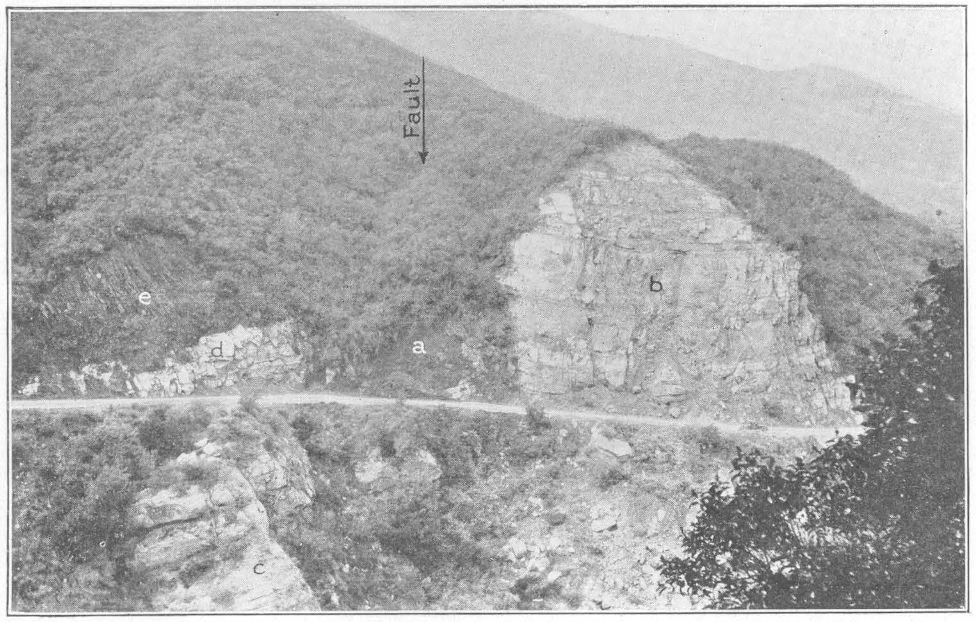
A. VIEW LOOKING WEST ALONG CREST OF SANTA MONICA MOUNTAINS, CALIF., FROM POINT ON RIDGE WEST OF MANDEVILLE CANYON

Shows the smooth, nearly horizontal sky line. The less distant ridge tops are remnants of an old Pleistocene erosion surface.



B. VIEW LOOKING NORTH-NORTHWEST ALONG CAHUENGA FAULT, CALIF.

Cahuenga Peak and the ridge east of the fault are composed of conglomerate of the Topanga formation dipping 20°–30° N. Along the fault the conglomerate abuts against younger Topanga shale.



C. VIEW LOOKING WEST ACROSS TOPANGA CANYON 1 MILE BELOW TOPANGA, CALIF.

Shows fault accompanied by a crushed zone of basalt (a), which separates brown Topanga conglomeratic sandstone (b) from red Sespe (?) or Vaqueros (?) conglomeratic sandstone (c); upper 8 feet of red conglomerate (d) is baked to purple color by diabase sill (e).

AIRPLANE VIEW TAKEN ABOVE THE CITY OF SANTA MONICA LOOKING NORTH AT SANTA MONICA MOUNTAINS, CALIF.

Photograph by Spence Airplane Photos.

AIRPLANE VIEW LOOKING NORTHWEST OVER PART OF THE RESIDENTIAL DISTRICT OF WESTERN LOS ANGELES, CALIF.

Shows the present small producing area of the Salt Lake oil field situated on the alluvial plain south of the Santa Monica Mountains. Photograph by Spence Airplane Photos.

penetrated by producing wells of the Seal Beach oil field between depths of about 2,000 and 3,000 feet. They are also equivalent to beds in Adams Canyon that lie 5,000 to 10,000 feet stratigraphically below the so-called Saugus-Pico contact.

Doctor Goudkoff's interpretation of the foraminiferal evidence provided by collections 311 and 312, taken in isolated areas, appears on pages 119–120.

SAN DIEGO FORMATION (MIDDLE OR UPPER PLIOCENE)

The San Diego formation is exposed in the coastal belt south of the Santa Monica Mountains and is represented by a massive soft light-gray conglomeratic sandstone, which crops out in lower Temescal Canyon and is best exposed along the axis of a prominent syncline in the west wall of the canyon 300 to 500 feet north of the Coast Highway. The entire thickness of sandstone at this locality is about 50 feet, but there appears to be a distinct break about 15 feet above the base which divides it into two separate units. The lower unit appears to grade downward into Modelo shale and is therefore believed to be Miocene. The upper unit is definitely Pliocene and is correlated by its megascopic fossils with the San Diego formation. Although the exposure is in a difficult place to examine satisfactorily, there is a strong suggestion that the uppermost beds of the lower unit are truncated by its upper surface. Angular blocks of *Pholas*-bored limestone, absent in the lower unit and presumably derived from the underlying Miocene, are abundant just above the contact and throughout the upper unit of sandstone. This upper sandstone is rich in countless brachiopods and Pectens, which are included in a fauna that Arnold recognized as being equivalent to that found in the San Diego formation.[53]

W. P. Woodring has identified the following fossils collected from the San Diego formation in lower Temescal Canyon, near Santa Monica (locality 55):

Brachiopod:
- Dallinella occidentalis (Dall).

Gastropod:
- Opalia varicostata Stearns.

Lamellibranchs:
- Ostrea vespertina Conrad.
- Pecten bellus hemphilli Dall?
- Pecten stearnsii Dall.
- "Pecten" healeyi Arnold.
- "Pecten" cerrosensis Gabb.
- "Pecten" purpuratus Lamarck var.
- Chlamys hastatus (Sowerby).
- Chlamys opuntia (Dall).
- Chlamys swiftii parmeleei (Dall).
- Lucinoma annulata (Reeve)?
- Miltha sp. cf. M. xantusi (Dall).

Woodring's comments are as follows:

The San Diego affinities of this small fauna were recognized by Arnold.[54] The same species are found in so-called Saugus beds north of Simi Valley.[55] The occurrence of these distinctive species so far north disposes of the idea that the warm-water San Diego fauna lived at San Diego at the same time that a cool-water fauna lived in the Los Angeles and Ventura Basins. The considerable percentage of extinct species clearly shows that this fauna is well down in the Pliocene. This is the only locality around the western border of the Los Angeles Basin where the San Diego fauna is found. It also is the only locality in southern California where fossil brachiopods are abundant. More than 200 specimens of *Dallinella occidentalis* from this locality show a wide range of variation, embracing forms like "*Terebratalia*" *smithi* Arnold. Many of the species recorded by Rivers[56] undoubtedly were collected at this locality, but more than one horizon is represented in his Pliocene.

Possibly the locality record "Tremochal Canyon, Santa Monica, Pleistocene," given by Canu and Bassler[57] for six species of Bryozoa, is an error for Temescal Canyon, but it is improbable that any but incrusting Bryozoa are to be found in the coarse sand of the San Diego formation.

PLIOCENE NEAR SAWTELLE

A small isolated exposure of late Pliocene rocks occurs on the east side of Brentwood Knoll, near the west edge of the town of Sawtelle. The rocks consist of gray clay and sandy clay and have been considerably deformed. Foraminifera from these beds are listed as collection 312 on page 117 and are discussed by Doctor Goudkoff as follows:

The microfauna found in sample 312, though rather closely related to that discovered in the so-called Pliocene beds of Deadman Island, differs from it, first, in the abundance of Bolivinas and Uvigerinas; second, in the presence of *Frondicularia advena;* third, in the absence of *Dentalina baggi, Elphidium (Themeon) crispum, Globorotalia campanulata, Globorotalia grandis, Polymorphina elongata, Polymorphina frondiculariformis,* and some others, which seem to be definitely confined to the youngest formations of southern California like those exposed in the Lomita quarry, Deadman Island, Timms Point, and at the Santa Barbara Bathhouse Beach. On the other hand, the foraminiferal association present in sample 312 shows a great similarity to that found in a lower part of the Kalorama Canyon section of Ventura County, particularly that underlying the strata mapped by Kew as Saugus. Thus I am inclined to believe that sample 312 comes from a part of the upper Pliocene.

PLIOCENE (?) IGNEOUS ROCKS

In the coastal belt there are small areas of basalt whose age is indefinite. About 1,500 feet up from the mouth of Pulga Canyon, in the small canyon directly to the west,[58] light-gray and brown shale of the Modelo formation is faulted against red sandstone and conglomerate of the Sespe (?) and Vaqueros (?) formations. Basalt has been intruded along the fault contact (see pl. 16), presumably at about the time that fault displacement occurred. As this faulting involves upper Miocene rocks, the basalt must be younger and may well be of Pliocene or post-Pliocene age.

[53] Arnold, Ralph, U. S. Nat. Mus. Proc., vol. 32, p. 527, 1907.
[54] Arnold, Ralph, U. S. Nat. Mus. Proc., vol. 32, p. 527, 1907.
[55] Woodring, W. P., California Acad. Sci. Proc., 4th ser., vol. 19, pp. 57-64, 1930.
[56] Rivers, J. J., Descriptions of some undescribed fossil shells of Pleistocene and Pliocene formations of the Santa Monica Range: Southern California Acad. Sci. Bull., vol. 3, pp. 69–72, 1904.
[57] Canu, Ferdinand, and Bassler, R. S., U. S. Nat. Mus. Bull. 125, pp. 139, 170, 179, 195, 198, 202, 1923.
[58] Since these observations were made exposures in this small canyon have become concealed by fill and excavation for a new Bay Club.

UPPER PLIOCENE OR LOWER PLEISTOCENE ROCKS

LOWER RUSTIC CANYON

A small isolated exposure of clay, silt, and soft sandstone of upper Pliocene or lower Pleistocene age occurs in a storm-drain ditch in the west wall of lower Rustic Canyon about half a mile from the coast. These beds have been tilted considerably and dip 12° S. Megascopic fossils collected from this area (locality 311) are identified and discussed by W. P. Woodring as follows:

Bryozoa:
- Cellaria fissurifera Canu and Bassler.
- Cellaria diffusa Robertson.
- Microporella sp.
- Porella collifera Robertson.
- Phidolopora pacifica (Robertson).
- Tubucellaria punctulata Gabb and Horn.
- Filisparsa clarki Canu and Bassler.
- Idmonea californica D'Orbigny.

Gastropods:
- Tricolia pulloidea (Carpenter).
- Odostomia sp.
- Alvania pedroana Bartsch.
- Bittium cf. B. rugatum Carpenter.
- Cerithiopsis sp.
- "Alabina" californica (Dall and Bartsch).
- Nassarius perpinguis (Gould).

Lamellibranchs:
- "Pecten" caurinus Gould.
- Crenella sp.
- Cyclocardia ventricosa (Gould).
- "Protocardia" centifilosa (Carpenter).
- Psephidia ovalis Dall.

(For Foraminifera collected at this locality see p. 117.)

This collection represents the cool-water horizon of the Los Angeles Basin, found at Timms Point and formerly at Deadman Island, and probably also the cool-water zone represented in the deposits that have been called Santa Barbara beds. Bryozoa are very abundant at the locality in Rustic Canyon, as at other places where this cool-water fauna is represented. Canu and Bassler [59] record 22 species of Bryozoa from Rustic Canyon, 8 species from Long Wharf Canyon (presumably Potrero Canyon), 6 species from "Tremochal" (Temescal?) Canyon, and 9 from Santa Monica without any definite locality data, making a total of 45 species of Bryozoa referred by them to the Pleistocene from localities near Santa Monica, none of which are recorded from more than one locality.

Doctor Goudkoff makes the following interpretation of the foraminiferal fauna found in collection 311 at this locality:

The microfauna present in sample 311 differs from that found in sample 312 in the absence of *Frondicularia advena*, the nearly entire absence of Bolivinas and Uvigerinas, and the presence of *Nonion scapha* and *Elphidium* (*Themeon*) *granulosus*. Such features make it possible to state, first, that sample 311 seems to be still closer to the Deadman Island beds and, second, that it falls in the part of the Kalorama section stratigraphically higher than that with which sample 312 can be correlated.

[59] Canu, Ferdinand, and Bassler, R. S., U. S. Nat. Mus. Bull. 125, 1923. Four of the species (*Cellaria fissurifera, Stephanosella biapteria, Phidolopora pacifica, Crisia serrata*) have only the indefinite locality record "Santa Monica" in the text but in the explanation of the plates are recorded from Rustic Canyon. *Porella cyclopea*, listed on page 14, seems to be a nude name.

Samples 311 and 312 may be correlated with the uppermost part of the Pico formation as mapped by Kew in the Pico Canyon district.

At the mouth of Potrero Canyon about 100 feet of gently dipping bluish-gray and brown sandstone and conglomerate rests upon and is faulted against more steeply dipping Pliocene beds and contains a fairly good fauna of marine invertebrates.

The following fossils were collected from the basal conglomeratic bed of the Pleistocene section exposed at the mouth of Potrero Canyon (locality 31). This bed crops out in the bottom of the canyon about 50 feet north of the Coast Highway. The fossils have been identified by W. P. Woodring, and his discussion follows the list.

Gastropods:
- Calliostoma costatum (Martyn).
- Calliostoma tricolor Gabb.
- Tricolia compta (Gould).
- Neverita recluziana (Deshayes).
- Turritella cooperi Carpenter.
- Bittium sp.
- Mitrella gausapata (Gould).
- Mitrella carinata (Hinds).
- Nassarius perpinguis (Hinds).
- Nassarius cooperi (Forbes).
- Olivella biplicata Sowerby.
- Olivella pedroana (Conrad).
- Conus californicus Hinds.
- Megasurcula carpenteriana (Gabb).

Lamellibranchs:
- Ostrea sp.
- Monia macroschisma (Deshayes).
- Thracia trapezoides Conrad?
- Cyclocardia ventricosa (Gould).
- Thyasira gouldii (Philippi).
- Cerastoderma nuttalli (Conrad).
- Venerupis staminea (Conrad).
- Petricola carditoides (Conrad).
- Macoma cf. M. inquinata (Deshayes).
- Gari californica Conrad?
- Spisula sp.
- Pholadidea penita (Conrad)?

In addition to many of the species in the preceding list Arnold [60] recorded the following from this locality: "*Bela*" *sanctae-monicae* (type locality), "*Pisania*" *fortis*, "*Pleuroloma*" *perversa*, and "*Trophon*" *scalariformis*. I doubt whether "*Pisania*" *fortis* is found in these beds. Aside from this record the fossils at this locality are like those at the cool-water horizon of Timms Point and Deadman Island, which lies at the top of the Pliocene or at the base of the Pleistocene, the assignment depending on where the Pliocene-Pleistocene boundary is placed. The doubtful record of *Thracia trapezoides*, a characteristic species of this cool-water horizon of the Los Angeles Basin, is based on an imperfect mold.

UPPER PLEISTOCENE ROCKS

Marine Pleistocene strata of slight thickness, probably of upper Pleistocene age, are exposed in Potrero Canyon. Similar deposits, presumably of about the same age, occupy two large isolated areas east of the city of Santa Monica, between Santa Monica and

[60] Arnold, Ralph, California Acad. Sci. Mem., vol. 3, p. 56, 1903.

Venice Boulevards. The most extensive series of Pleistocene rocks in this district, however, is the continental material that forms the broad high alluvial plain along the south flank of the mountains from Beverly Hills west to Santa Ynez Canyon—a plain which is now trenched by north-south canyons to depths of as much as 300 feet. (See pl. 31.)

MARINE PLEISTOCENE

Well-sorted light-brown sand of fine to medium-fine texture occupies the high gently sloping terraces along the coast between the mouths of Topango and Santa Ynez Canyons. A small amount of this marine sand has a distinct reddish tinge, a color which is due, no doubt, to the fact that this Pleistocene material has been derived in part from the red Sespe (?) and Vaqueros (?) formations, which are exposed in this part of the coastal belt.

Farther east, between Santa Ynez and Santa Monica Canyons, Pleistocene alluvial-plain material rests directly upon sharply deformed Tertiary rocks in much of the area, but there are localities where a thin veneer of marine Pleistocene intervenes and separates these two terranes. In upper Potrero Canyon, about a quarter of a mile south of Pacific Palisades post office, 5 feet or less of fine white and brown sand rests directly upon tilted Pliocene rocks and underlies about 50 feet of coarse brown alluvial-plain detritus. This thin bed of sand contains a well-preserved marine molluscan fauna of about 125 species (locality 61). In the lower part of this canyon about 25 feet of horizontal light-gray sand and clay of probable marine origin underlies the uppermost brown alluvial-plain material, but because of its poor exposures and uncertain character it has been mapped as a part of the overlying continental deposits.

Two large areas east of Santa Monica and between Santa Monica and Venice Boulevards (see pl. 16) are occupied by soft sand, clay, gravel, and conglomerate that are considered to be of marine origin and late Pleistocene age. Like all other marine Pleistocene deposits of this district, these sediments are not of latest Pleistocene age, for in the area of the Beverly Hills oil field they pass beneath the Pleistocene alluvial material that forms the Santa Monica Plain. In the more western area, in and just east of the city of Santa Monica, fine brown thin-bedded sand that has been washed free of all clay material is the most widespread type of rock. Near the ocean this sand occupies long, narrow ridges and bluffs which parallel the shore and which presumably represent sand bars and shore-line bluffs that were developed when the ocean stood at a higher level with relation to the land. In general, the Pleistocene of both the western and eastern areas is flat-lying, although dips as steep as 5° have been observed in the more eastern area south and southwest of the Beverly Hills oil field. This area lies along the northwestward-trending line of the Newport-Inglewood uplift and may have undergone a small amount of folding, in addition to simple uplift, in post-Pleistocene time.

The only fossils found in these two areas were collected from a conglomerate exposed in a cut bank on the west side of Overland Avenue at the northwestern edge of the town of Palms (locality 68, pl. 16).

Doctor Woodring has identified the following fossils in the collection made from the thin Pleistocene sand near the head of Potrero Canyon (locality 61). His comments follow the list.

Bryozoa:
- Cellaria sp.
- Idmonea californica D'Orbigny.

Gastropods:
- Acmaea insessa (Hinds).
- Acmaea? sp.
- Tegula ligulata (Menke).
- Tegula aureotincta (Forbes).
- Calliostoma annulatum (Martyn)?
- Calliostoma supragranosum Carpenter.
- Margarites optabilis (Carpenter).
- Astraea undosa (Wood).
- Homalopoma paucicostata (Dall).
- Tricolia pulloidea (Carpenter).
- Tricolia variegata (Carpenter).
- "Vitrinella" sp.
- Turbonilla, 5 spp.
- Odostomia, 4 spp.
- Melanella sp.
- Epitonium bellastriatum (Carpenter).
- Epitonium tinctum (Carpenter).
- Neveritia alta ("Dall") (Arnold).
- Sinum sp.
- Calyptraea contorta (Carpenter).
- Crepidula adunca Sowerby.
- Crepidula mummaria Gould.
- Crepidula lingulata Gould.
- Alvania pedroana Bartsch.
- Alvania? sp.
- Diala? sp.
- Rissoina pleistocena Bartsch.
- Lacuna unifasciata Carpenter.
- Turritella cooperi Carpenter.
- Caecum californicum Dall.
- Micranellum crebricinctum (Carpenter).
- Bittium cf. B. giganteum Bartsch.
- Bittium sp.
- Bittium attenuatum Carpenter.
- Cerithiopsis fatua Bartsch.
- Cerithiopsis fossilis Bartsch.
- Seila montereyensis Bartsch.
- Triphora, 2 spp.
- "Alabina"? sp.
- Erato columbella Menke.
- Bursa californica Hinds.
- Acanthina paucilirata (Stearns).
- "Murex" festivus Hinds.
- Tritonalia poulsoni (Carpenter).
- Tritonalia barbarensis (Gabb).
- Boreotrophon stuarti (Smith).
- Mitrella gausapata (Gould).
- Amphissa versicolor Dall.
- Nassarius californianus (Conrad).

Gastropods—Continued.
- Nassarius perpinguis (Hinds).
- Nassarius fossatus (Gould).
- Barbarofusus barbarensis (Trask).
- Marginella regularis Carpenter.
- Olivella biplicata Sowerby.
- Olivella pedroana (Conrad).
- Conus californicus Hinds.
- Clathrodrillia incisa (Carpenter).
- Elaeocyma hemphilli (Stearns).
- Megasurcula carpenteriana (Gabb).
- Mangilia cf. M. barbarensis Oldroyd.
- Mangilia, 2 spp.
- "Aesopus" cf. A. oldroydi Arnold.
- Terebra pedroana philippiana Dall.
- Acteon traskii Stearns.
- Acteon punctocoelata (Carpenter).
- Acteocina culcitella (Gould).
- Acteocina sp.
- Sulcularia sp.
- Volvula cylindrica Carpenter.
- Cylichna "alba (Brown)."
- Williamia? sp.

Scaphopods:
- Dentalium neohexagonum Sharp and Pilsbry.
- Cadulus fusiformis Pilsbry and Sharp.

Lamellibranchs:
- Nucula exigua Sowerby.
- Saccella taphria (Dall).
- Yoldia sp.
- Ostrea sp.
- "Pecten" latiauritus Conrad.
- Monia macroschisma (Deshayes) ?
- Pandora punctata Conrad.
- Pseudochama exogyra (Conrad).
- Lucinisca nuttallii (Conrad).
- Pauvilucina tenuisculpta (Carpenter).
- Erycina sp.
- Rochefortia sp.
- Sportella sp.
- Trachycardium quadrigenarium (Conrad).
- Trachycardium procerum (Sowerby).
- "Cardium" elatum Sowerby.
- "Transennella" tantilla (Gould).
- Amiantis callosa (Conrad).
- Chione undatella (Sowerby).
- Chione sp.
- Venerupis tenerrima (Carpenter).
- Cooperella sp.
- Tellina idae Dall.
- Tellina buttoni Dall.
- Tellina sp.
- Macoma yoldiformis Deshayes.
- Macoma indentata Carpenter? (more elongate than living specimens).
- Donax gouldii Dall.
- Solen sicarius Gould.
- Ensis californicus Dall.
- Siliqua lucida (Conrad).
- Mactra californica Conrad.
- Spisula hemphilli (Dall).
- Schizothaerus nuttalli (Conrad).
- Cryptomya californica (Conrad).
- Corbula luteola Carpenter.
- Saxicava arctica (Linné)
- Zirfaea gabbi Tryon?

Chiton:
- Ischnochiton conspicuus (Carpenter).

This is a distinctly warm-water fauna. *Astraea undosa, Calyptraea mammillaris, Bursa californica, Acanthina paucilirata, Terebra pedroana philippiana, Acteon traskii, Trachycardium procerum, "Cardium" elatum, Amiantis callosa, Chione undatella,* and *Spisula hemphilli* are southern species, and all except *Bursa californica* are not found farther up the coast than San Pedro, or the northern limit of their range is even farther south. *Rissoina, Erato,* and *Sulcularia* also are southern genera. None of the species found in these beds has a distinctly northern facies, except possibly the one listed as *Cylichna "alba,"* though many of them now range far up the coast.

The stratigraphic position of the sands carrying these beautifully preserved shells and the warm-water aspect of the fauna indicate that they are of upper San Pedro age. The difference in facies (fine sand as contrasted with coarse gravel) is considered sufficient to account for the difference between this fauna and the fossils of the type upper San Pedro of Deadman Island. The same sand facies and the same fauna are found several miles to the southeast at University City, along the bluff overlooking the Ballona Plain. The sands here probably correspond to the sands of the Baldwin Hills section described by Tieje [61] as the Palos Verdes sands.

W. P. Woodring has provided the following identifications of the small collection of fossils made on Overland Avenue near the northwest edge of the town of Palms (locality 68):

- Cerithidea californica (Haldemann).
- Ostrea lurida Carpenter.
- Chione undatella (Sowerby).
- Tellina meropsis Dall.

The warm-water facies is about the only clue to the age of these beds furnished by this small collection. They probably represent upper San Pedro time.

NONMARINE PLEISTOCENE

Nonmarine Pleistocene deposits of the high dissected alluvial plain west of Beverly Hills are well exposed in the sea cliff and canyon walls northwest of Santa Monica and in the many new cuts along streets and roads that cross this plain, particularly Beverly Boulevard. These deposits range in thickness from a few feet to at least 200 feet, are dark brown, and are composed of poorly sorted angular rock fragments as much as several feet across, which are embedded in a soft matrix of reddish-brown clay and sand. Bedding is characteristically indistinct and very irregular. Plate 29, *A*, shows a view of this continental material which forms the sea cliff near the mouth of Santa Monica Canyon, and Plate 29, *B*, is a closer view that shows the poor sorting and character of bedding common to these deposits.

This nonmarine material has been deposited by streams that flowed south from the Santa Monica Mountains in Pleistocene time and dumped much of their load of sediment before reaching the ocean. Since the Pleistocene epoch marine erosion has been active along this part of the coast, and as a result the shore has been cut back to a point several miles north and northeast of its earlier position. This

[61] Tieje, A. J., The Pliocene and Pleistocene history of the Baldwin Hills, Los Angeles County, Calif.: Am. Assoc. Petroleum Geologists Bull., vol. 10, No. 5, pp. 502-512, 1 fig., 1926.

marine planation has thus materially reduced the distance which the streams from the north traveled to reach base level, a condition which was an influential factor in forcing these streams to cut deep canyons in the Pleistocene alluvial plain. Other factors that have contributed to the trenching of this plain are discussed on page 130.

RECENT ALLUVIUM

Recent alluvium occurs along the courses of most of the larger streams of the mountains and covers extensive lowland areas in the adjoining San Fernando Valley and Los Angeles Basin. This material is similar to the Pleistocene alluvial-plain deposits in color, texture, sorting, and bedding, and in many places the two terranes can be distinguished only by the different relations which they bear to the present system of drainage. South of the mountains they not uncommonly merge almost imperceptibly into one another.

FOSSIL LOCALITIES

The following lists give in detail the localities where fossils were obtained in this area. The localities for macroscopic fossils are also shown by corresponding numbers on the geologic map. Localities for Foraminifera and diatoms are shown on Plate 27.

Fossil localities of Santa Monica Mountains (1927)

Macroscopic fossils

	Upper Cretaceous	Eocene (Martinez)	Topanga	Lower Modelo	Upper Modelo	Pliocene	Pleistocene
2. Brown sandstone on ridge north of Mulholland Highway, at west edge of Burbank quadrangle (triangulation station 1381)			×				
4. Platy shale near middle of Modelo formation in road cut on South Sherman Way, 200 yards south of Mulholland Way, in NW. ¼ sec. 34, T. 1 N., R. 15 W.				×			
5. Massive sandstone ledge with basalt in NW. ¼ sec. 31, T. 1 N., R. 14 W.			×				
6. Massive fine sandstone interbedded with basalt in large north-south canyon in NE. ¼ sec. 3, T. 1 S., R. 14 W.			×				
7. Brown sandstone on top of ridge north of Mulholland Highway and west of Laurel Canyon road in center of E. ½ sec. 31, T. 1 N., R. 14 W.			×				
8. Brown sandstone below basalt on road leading south from Mulholland Highway along east side of ridge in north-central part of sec. 4, T. 1 S., R. 14 W.			×				
9. Thin-bedded shale and sandstone exposed in fresh cut at rear of Riding Academy stable along west bank of the Los Angeles River, 1,000 feet south of Los Felis Boulevard				×			
10. Basal light-gray conglomeratic sandstone of Modelo formation on upper road 800 feet northeast of Will Rogers's ranch house, on mesa just north of Beverly Boulevard, west of lower Rustic Canyon				×			
11. White diatomaceous shale on minor road along north edge of hills, 2,000 feet west of South Sherman Way					×		
12. Isolated patch of basal graywacke of Modelo formation south of Mulholland Highway and just east of head of Sepulveda Canyon				×			
13. Sandstone on west side of canyon, 3,000 feet south of Encino Country Club, west edge of Van Nuys quadrangle			×				
14. Conglomeratic sandstone at top of hill at east edge of Reseda quadrangle near Encino Country Club			×				
15. Basal conglomerate of Modelo formation on ridge just west of Encino Reservoir				×			
16. Same general locality as 18 and 19	×						
17. Top of ridge west of Temescal Canyon, 700 feet north of south edge of Reseda quadrangle	×						
18. Shale on Mulholland Highway 2¾ miles east of west edge of Reseda quadrangle	×						
19. Same as 18	×						
20. Sandstone on top of hill in NE. ¼ sec. 32, T. 1 N., R. 16 W., Reseda quadrangle				×			
21. Top of ridge 1¾ miles east and 1½ miles north from southwest corner of Reseda quadrangle		×					
22. Ridge 1¾ miles east and 1¾ miles north from southwest corner of Reseda quadrangle	×						
23. Boulders in creek bed 1¼ miles north and 1¾ miles east from southwest corner of Reseda quadrangle	×						
25. Massive nodular conglomeratic sandstone on top of ridge in SW. ¼ NE. ¼ sec. 5, T. 1 S., R. 16 W.			×				
26. Shale on ridge 2 miles east and 1 mile north from southwest corner of Reseda quadrangle		×					
27. Shale on ridge 2½ miles east and three-quarters of a mile north from southwest corner of Reseda quadrangle		×					
28. Sandstone on west side of Santa Maria Ranch road in NE. ¼ sec. 32, T. 1 N., R. 16 W.; 28A is sandstone 25 feet stratigraphically above 28B			×				
29. Basal conglomerate of Modelo formation just south of 1,711-foot hill on Los Angeles city boundary and south of Mulholland Highway, in southwestern part of Reseda quadrangle				×			
30. Lower sandstone of Modelo formation in NW. ¼ sec. 32, T. 1 N., R. 16 W.				×			
31. Bed of lower Potrero Canyon, 50 feet north of Coast Highway						?	?
32. Fossil casts in sandstone and clay shale half a mile east of Trippet ranch in northwestern part of Topanga Canyon quadrangle			×				
33. Soft brown sandstone on side road just east of Topanga post office, Topanga Canyon quadrangle			×				
34. Sandstone on low ridge in fault valley half a mile northeast of Trippet ranch in northwestern part of Topanga Canyon quadrangle			×				

Fossil localities of Santa Monica Mountains (1927)—Continued

Macroscopic fossils—Continued

	Upper Cretaceous	Eocene (Martinez)	Topanga	Lower Modelo	Upper Modelo	Pliocene	Pleistocene
38. Conglomeratic sandstone near middle member of Topanga formation in Brown Canyon about 1,500 feet north of Beverly Glen			×				
41. Shale in Santa Ynez Canyon, 1,000 feet north of latitude 34° 04′	×						
42. Sandstone on top of hill just east of Santa Ynez Canyon 1 mile north of Coast Highway			×				
43. Top of ridge east of Santa Ynez Canyon 2 miles north of Coast Highway		×					
44. Same as 43		×					
45. Ridge west of upper Temescal Canyon 2 miles south of north edge of Topanga Canyon quadrangle		×					
46. Ridge west of upper Temescal Canyon, 3,000 feet south of north edge of Topanga Canyon quadrangle	×						
47. Ridge just west of Temescal Canyon near north edge of Topanga Canyon quadrangle	×						
48. Ridge just west of upper Temescal Canyon at north edge of Topanga Canyon quadrangle	×						
49. Road up small canyon three-quarters of a mile north of Beverly Boulevard half a mile west of Rustic Canyon	×						
50. Top of high ridge just east of lower Sepulveda Canyon half a mile north of Beverly Boulevard			?				
51. Basal graywacke of Modelo formation on top of ridge west of Brown Canyon, near south edge of mountains				×			
52. Basal graywacke of Modelo formation at top of south end of ridge just east of Brown Canyon				×			
53. Sandstone on southwest side of high peak on Los Angeles city boundary 1 mile east of Cahuenga Peak and west of Griffith Park Road			×				
54. Massive sandstone at quarry in Santa Ynez Canyon 1¾ miles north of Coast Highway	×						
55. Massive gray sandstone in west wall of Temescal Canyon 200 yards above mouth						×	
56. Massive conglomeratic sandstone near fault across Topanga Canyon road 2½ miles north of Coast Highway			×				
57. Base of uppermost shale of lower member of Modelo formation on Mulholland Highway half a mile east of Benedict Canyon road				×			
58. Modelo shale on top of hill just south of Mulholland Highway near head of Sepulveda Canyon, 2,000 feet east of Reseda quadrangle				×			
59. Shale and sandstone on ridge west of Temescal Canyon 2 miles north of coast		×					
60. Sandstone on Mulholland Highway a quarter of a mile east of Franklin Canyon			×				
61. Bottom of upper Potrero Canyon about a quarter of a mile south of Pacific Palisades post office							×
61A. Bottom of upper Potrero Canyon, 2 feet vertically below and 5 or 10 feet south of 61						×	
62. Limestone concretion on ridge at west edge of sec. 33, T. 1 N., R. 16 W		×					
63. Conglomeratic sandstone on ridge at west edge of sec. 33, T. 1 N., R. 16 W			×				
64. Base of Modelo formation on ridge top just west of lower Sepulveda Canyon				×			
65. Base of Modelo formation on south side of ridge at west edge of sec. 33, T. 1 N., R. 16 W				×			
66. Approximately the same locality as 38			×				
67. Sandy shale near tipple at limestone quarry in upper Santa Ynez Canyon		×					
67A. Float near west end of limestone quarry in upper Santa Ynez Canyon near 67		×					
67B. Shale underlying limestone in canyon 690 feet east of limestone quarry in upper Santa Ynez Canyon		×					
67C. Limestone in limestone quarry in upper Santa Ynez Canyon		×					
68. Cut bank on west side of Overland Avenue at northwest edge of town of Palms							×
311. Storm-drain ditch in west wall of lower Rustic Canyon half a mile from coast							×

Foraminifera [62]

Topanga formation (middle Miocene) (see geologic map for location):
- 101. Sandstone and shale stringer in basalt on Cahuenga Avenue a quarter of a mile northeast of Hollywood Bowl.
- 102. Shale stringer in basalt on Mulholland Highway 2,000 feet east of Laurel Canyon road.
- 103. Thin-bedded shale and sandstone above basalt near head of Nichols Canyon.

Modelo formation (upper Miocene) (see pl. 27, *B* and *C*, for exact locations):
- 104. Brown thin-bedded shale on road up hill from Coast Highway at Pacific Palisades beach stand.
- 105. Brown thin-bedded shale on coast at mouth of Temescal Canyon.
- 106. Same as 105.
- 107. Hard brown shale of small fault block in Potrero Canyon.
- 110. White platy shale near base of Modelo formation, half a mile north of Beverly Hills Hotel. (See geologic map for location.)
- 111–174. Section of Modelo formation between Mohn Springs, in Garrapata Canyon, and Ventura Boulevard.
- 175. Within 25 feet of top of exposed upper member of Modelo formation at north edge of mountains just west of South Sherman Way.
- 303–304. Potrero Canyon, half to three-quarters of a mile west of lower Santa Monica Canyon.

Pliocene (see pl. 27, *B*, for exact locations):
- 301–302, 305–310. Potrero Canyon, half to three-quarters of a mile west of lower Santa Monica Canyon.
- 312. From road cut in east side of Brentwood Knoll in west edge of Sawtelle just north of Wilshire Boulevard.

Upper Pliocene or lower Pleistocene:
- 311. Storm-drain ditch in west wall of lower Rustic Canyon half a mile from coast.

[62] Although these samples were collected principally for information regarding their contained Foraminifera, the presence of Radiolaria, diatoms, and other organic remains, where noted, is mentioned in the tables within the text.

STRUCTURE

The eastern part of the Santa Monica Mountains, east of Topanga Canyon, is a large anticline that has experienced several stages of growth and deformation. The anticlinal structure is still clearly obvious in the central part of the area, but in the eastern and western parts the original fold has been so intricately deformed by block faulting and igneous intrusion that much of the fold is either difficult to recognize as such, or is down-faulted and entirely concealed beneath alluvium. Pre-Modelo diastrophism produced an anticline which, to judge from the present westward plunge, was complete in the district east of Topanga Canyon, although similar major uplifts of this age probably occurred farther west; post-Modelo diastrophism, however, caused an anticlinal uplift that apparently affected a larger area as a unit, including the district west of Topanga Canyon as well as that east of it.

In order to discuss satisfactorily the character and possible origin of the Santa Monica anticline and related structural features (see pl. 16), it appears necessary to consider each individual period of pronounced diastrophism separately and to describe as accurately as possible the structural features which have resulted from each of these periods. There appear to have been at least three such periods—the Jurassic (?) granitic intrusion, the middle Miocene (post-Topanga and pre-Modelo) disturbance, and the post-Miocene disturbance. Under the last are grouped two disturbances, one near the end of Miocene time and the other near the end of Pliocene time, both of which were probably of major importance, although the latter was almost certainly the more intense. Deformation during other periods, such as that between Martinez (lower Eocene) and Sespe (?) and Vaqueros (?) (Oligocene? and lower Miocene) time, was probably also pronounced locally, but because its effects are not well known it is mentioned only occasionally in this discussion.

JURASSIC (?) GRANITIC INTRUSION

The intrusion of molten magma, which later solidified to form granite and granodiorite, into the Santa Monica slate, presumably at about the end of Jurassic time, was the first deformation of which there is record in the eastern part of the Santa Monica Mountains. Plate 16 shows the presence of a large granite mass northeast of Beverly Hills and other smaller areas of similar rocks much farther west, all of which owe their existence solely to the mechanism of igneous intrusion. Areas of granite farther east in Griffith Park and Hollywoodland are largely if not entirely up-faulted blocks. The intrusive bodies have metamorphosed the adjoining slate into phyllite and mica schist and have developed the spotted slate described on page 89.

The two major exposed intrusive bodies lie within an extensive area of Triassic (?) slate that has been folded into a broad anticline. They are separated by several miles of this anticlinal slate, but the known distribution of spotted slate, as shown on the geologic map and described on page 89, indicates that similar bodies of intrusive granite are close to the surface in intermediate areas along or slightly north of the anticlinal axis. It is considered probable, therefore, that the two major exposed bodies of intrusive granite represent merely high and relatively small portions of a much larger mass which underlies much of the area along or adjoining the axis of the Santa Monica anticline. Structure sections D–D′, E–E′, and F–F′ (pl. 17) illustrate the breadth of this fold, which plunges rapidly westward, in general away from the largest exposed granite mass. The character of this anticline and the incompetency of the Santa Monica slate to transmit horizontal stresses adequate for its development force the conclusion that the Santa Monica anticline has resulted from vertical uplift.

Although a minor amount of anticlinal warping may have resulted directly from the Jurassic (?) intrusion, inspection of the structure sections reveals the fact that the Topanga formation has been folded nearly as much as the Santa Monica slate, and that the major development of the Santa Monica anticline must therefore have occurred in post-Topanga (post-middle Miocene) time. If the original granitic intrusion occurred in or near Jurassic time, which seems likely, it is certain that the granite and surrounding slate were uplifted at a much later time, either by renewed deep-seated igneous activity or by vertically acting forces of different origin.

POST-TOPANGA AND PRE-MODELO DEFORMATION

One of the chief periods of deformation in the Santa Monica Mountain region occurred after the middle of the Miocene epoch, between Topanga time and Modelo time. This disturbance resulted in pronounced anticlinal folding of the Topanga and all older formations, large-scale faulting in many parts of the fold, and intrusions of basalt many of which followed lines of faulting (see pl. 30, *C*) and probably were genetically related to all other types of deformation during this revolutionary period. The geologic map shows the marked discordance in the attitude of the Topanga and Modelo formations in different parts of the district (see also pl. 22, *B*) and clearly illustrates the fact that the Topanga underwent as much anticlinal folding and probably more faulting and basalt intrusion in some areas during the pre-Modelo disturbance than in subsequent deformative periods. Evidence for this is to be found in the general vicinity of Beverly Glen and the head of Benedict Canyon near the Encino Reservoir, and in the Garrapata Canyon area near Mohn Springs. (See geologic map and structure sections.)

Just how much of the faulting in the western part of this district took place during middle Miocene time is uncertain. The Santa Ynez Canyon and Topanga faults are believed to have been developed at this time, and it may be that most of the other faults in this and other more eastern parts of the area were formed in the middle Miocene, prior to the Modelo epoch. Many of them have certainly experienced more recent movements, but these may have occurred along preexisting fractures of middle Miocene age. If the greater part of the movement along the Santa Ynez Canyon fault took place in middle Miocene time, which seems probable, it may be that the uplift of the block on the east was in response to pronounced anticlinal bulging in the slate and granite area still farther east. The same reasoning appears to apply even better to the Temescal fault, for the relation of this arcuate fault to the major anticlinal uplift in the slate suggests that it resulted from a concentration of vertically acting forces in the core of the range to the east. However, if the rocks mapped as Topanga (?) east of this fault near the south edge of the mountains (see pl. 16) are of Miocene age, which seems probable, then movement along the Temescal fault has not been so simple. In post-Cretaceous and pre-Topanga time, presumably during the period of deformation between Martinez and Sespe (?) or Vaqueros (?) time, there appears to have been relatively large-scale uplift of the eastern block of slate; but in post-Topanga time it seems that this block, near the southern edge of the mountains at least, subsided several thousand feet.

The Hollywood fault and the arcuate fault just west of Laurel Canyon appear to be high-angle faults and to have resulted from uplift of the large body of granite, possibly at intermittent periods throughout late Tertiary time. They may, however, have developed in post-Pliocene time: there is physiographic evidence to suggest that the last movement occurred in the Pleistocene. Little can be said regarding the age of the Cahuenga fault and associated faults of the Griffith Park-Hollywood area except that all are younger than the Topanga and the more eastern ones near Los Felis Boulevard are of post-Modelo age.

The Benedict Canyon fault that cuts diagonally across the Santa Monica Mountains northwest of Beverly Hills is unique in character to judge from the effects that apparently have resulted from displacement along it. This fault trends in a northeasterly direction and appears to be a vertical shear zone along which the displacement has been at least partly horizontal and of considerable magnitude. This fault seems to be offset by a cross fault east of Benedict Canyon. Although it is shown on Plate 16 as dying out near the Mulholland Highway, before it reaches the Topanga-Modelo contact, it may continue considerably farther but in a more easterly direction and more nearly parallel to the strike of the rocks. Movement along this fault zone appears to have resulted in a horizontal offset of several stratigraphic units for a distance of approximately 1½ miles. The axis of the Santa Monica anticline also appears to be offset, although the axis of this fold east of the fault is not very evident. This offset of the anticlinal axis indicated on the geologic map is of smaller magnitude than that of the several stratigraphic units, an apparent discrepancy which may be explained if it is assumed that considerable horizontal displacement occurred along this fault zone prior to the development of a pronounced anticlinal axis. The time at which this fault originated is uncertain. Large-scale displacement occurred in post-Topanga time, possibly before the Modelo was deposited, although certainly considerable displacement has occurred in post-Modelo time.

POST-MIOCENE DEFORMATION

Because there is little evidence in this area as to the effect of the disturbance at the end of the Miocene epoch, it is here described with post-Pliocene and post-Pleistocene disturbances under a single heading.

It seems probable that the Modelo (upper Miocene) formation was deposited over most, if not all, of the area covered by this report. As a result of post-Miocene uplift in the central part of the range the Modelo was folded to form an integral part of the Santa Monica anticline and has later been eroded from much of the crest and flanks of this large fold. Structure section E–E′ best illustrates the breadth of the anticline produced by post-Modelo folding and suggests that this last period of folding resulted merely from a rejuvenation of the vertically acting forces that had earlier produced the Santa Monica anticline. That horizontal stresses also have been active locally is evident from the apparent horizontal displacement of the Modelo along the Benedict Canyon fault.

This folding was accompanied by considerable faulting in different parts of the area, but particularly along the south flank of the mountains. The Santa Monica slate near the southern base of the range northwest of Beverly Hills is intricately distorted and appears to have undergone large-scale deformation to a much greater degree than the overlying Modelo formation, although the slate in most other areas is characterized by simple structure. The Modelo here is broken by numerous short faults, which strike approximately parallel to the base of the mountains and most of which have displacements of not more than a few hundred feet and commonly considerably less. It seems probable that these small faults have resulted from post-Miocene movement along an older buried fault which parallels the southern base of the range and, in the area west of Beverly Hills, forms a westward extension of the Hollywood fault.

The coastal area northwest of the city of Santa Monica is one of complex structural conditions and is composed of a series of small fault blocks, now largely concealed beneath Pleistocene alluvium. The faulting involved rocks of all ages from Cretaceous to Pliocene, and it appears that this belt is an integral part of the fault zone which parallels the southern base of the range and which is the result, in part at least, of deformation after Pliocene time.

The first movement along the fault which cuts across the mouth of Potrero Canyon occurred after the marine upper Pliocene or lower Pleistocene beds at the mouth of the canyon were laid down but before the overlying uppermost Pleistocene alluvial-fan material was deposited. There is a suggestion in poor exposures in the east wall of the canyon that subsequent movement has displaced the late Pleistocene alluvial-fan deposits about 25 feet. The strike of this fault is in perfect alinement with a pronounced terrace in the surface of the late Pleistocene alluvial plain east of Potrero Canyon, a terrace which may be traced eastward to Santa Monica Canyon. This fault is well exposed, is vertical, has a strike of N. 75°–80° E., and has a total vertical displacement of about 150 feet. The north block has been lifted, relatively, and may have undergone considerable horizontal movement. A minor northward-dipping thrust fault occurs south of the vertical fault.

CHARACTER OF FAULTS

Some of the faults in the Cretaceous and Eocene rocks of Topanga Canyon dip north at angles as low as 40°; whether they are normal or thrust faults is not determinable from present knowledge of the stratigraphic sequence of the strata involved or from any exposed characteristics of the faults themselves. The northeasterly fault that cuts across the Mulholland Highway near San Vicente Mountain and terminates the Modelo formation against the Santa Monica slate is a normal fault which, in present exposures, dips 40°–50° SE. With these exceptions all faults shown on the geologic map appear to be of the high-angle type and probably either are normal faults or represent vertical or nearly vertical shear planes. It seems possible that some of the smaller faults in the coastal belt northwest of the city of Santa Monica have resulted from secondary horizontal stresses developed along the fractured zone between two or more major vertically moving fault blocks. Horizontal slickensides along the planes of minor faults in other highly disturbed parts of the area indicate that minor horizontal displacements may have accompanied other major high-angle fault movements.

SUMMARY OF STRUCTURE

By way of summary it may be said that forces of uplift recurrently active in the eastern part of the Santa Monica Mountains from the Jurassic period to the present have resulted in a pronounced, somewhat asymmetrical anticline. During the first periods of deformation, from Jurassic to late middle Miocene time, uplift was concentrated in the central granite-slate area and produced a fold that plunged rapidly to the west and, to judge from the curving strike of the lower Topanga beds just west of Cahuenga Avenue, also plunged to the east. The middle Miocene uplift of the central area appears to have produced several and possibly many major high-angle tension faults along the borders of the active area. Examples are the Santa Ynez Canyon fault, the Topanga fault, and possibly also the Temescal fault and a major fault zone along the southern base of the mountains of which the Hollywood fault is a part. Some of these faults were accompanied by intrusions of basalt, a type of igneous activity which may have played an effective part in developing Miocene forces of vertical uplift. Post-Miocene deformation, including the disturbance near the end of Miocene time, one near the end of the Pliocene time, and another near the end of the Pleistocene, increased the structural relief of the Santa Monica anticline and produced many faults in upper Miocene, Pliocene, or Pleistocene strata in all parts of the area, some of which appear to have resulted from renewed movements along older buried faults.

In consequence of these recurrent periods of deformation the Santa Monica anticline is no longer a simple fold. Much of it has been disrupted by faults and basalt intrusions, and the southeastern limb of the fold is lost from view, having been dropped, relatively, along the Hollywood fault and subsequently covered with alluvium.

GEOLOGIC HISTORY

The salient events of the geologic history are set forth in the following table:

Summary of geologic history of the eastern part of the Santa Monica Mountains during Mesozoic and Tertiary time

Sedimentation	Uplift, folding, faulting	Igneous activity	Result	Known area affected
Triassic (?).			5,000–7,000 feet of dark argillaceous deposits, now the Santa Monica slate.	Whole area.
	Jurassic (?).	Yes.	Batholithic intrusion; granitic rocks; metamorphism of Triassic (?) deposits to slate, phyllite, and schist; original anticlinal bulging of intruded slate probably occurred.	Central part.
			Erosion	
Upper Cretaceous and early Eocene (Chico and Martinez).			8,000+ feet of mostly marine Chico and Martinez conglomerate, sandstone, shale, and limestone; contains small intrusions of trachyte of uncertain age.	Western part.
	Post-early Eocene.		Unconformity with angular discordance and omission of middle and late Eocene and probably most of Oligocene strata.	Western part.
			Erosion	
Oligocene (?) and early Miocene (Sespe?, Vaqueros?, Topanga).	Gentle uplift locally in early Topanga time.	Yes.	4,500–7,500 feet of conglomerate, sandstone, and shale; extrusion and intrusion (?) of basaltic material accompanied by uplift and erosion of areas of igneous activity in eastern part of this district.	Whole area except for Sespe (?).
	Late middle Miocene.	Yes.	Pronounced anticlinal folding accompanied by major normal faulting and basalt intrusions. This and subsequent erosion has resulted in a major unconformity of this district.	Eastern and western parts.
			Erosion	
Late Miocene (Modelo).		Yes.	4,500 feet of conformable foraminiferal and diatomaceous shale and sandstone (Modelo formation) with thin beds of bentonite and acidic volcanic ash throughout.	Most of area.
	End of Miocene.		Importance uncertain because of meager distribution of Pliocene rocks. Local unconformity occurs in coastal belt northwest of Santa Monica.	Coastal area +.
			Erosion	
Pliocene.		Yes.[a]	1,000+ feet of marine clay and sandstone.	Santa Monica Plain; possibly all of mountains.
	Near end of Pliocene.	(?)	Pronounced folding and faulting in coastal belt and probably one of the two chief stages in the uplift and deformation of the Santa Monica anticline; possibly accompanied by minor intrusions of basalt.	
			Erosion	
Late Pliocene or early Pleistocene.			100+ feet of fossiliferous marine conglomerate, sandstone, and sandy clay.	Coastal area near mouth of Potrero Canyon.
	Near end of Pliocene or middle of Pleistocene.		Probable minor uplift of Santa Monica Mountains and Plain with faulting, as at mouth of Potrero Canyon.	Coastal area near mouth of Potrero Canyon.
			Erosion	
Late Pleistocene (upper San Pedro).			5+ feet of marine soft white sand in coastal area near Potrero Canyon.	Coastal area, Potrero Canyon.
	Uplift, late Pleistocene.		Probable minor uplift of Santa Monica Mountains and Plain. Erosion and local removal of all of late Pleistocene marine deposits. Dissection of mature topography of mountains.	Coastal area, Potrero Canyon.
Late Pleistocene.			200± feet of alluvial-fan material.	Santa Monica Plain.
	Near end of Pleistocene.		Uplift with probable minor faulting. Dissection of Santa Monica Mountains and Plain.	Santa Monica Mountains and Plain.

[a] No definite proof of igneous activity during Pliocene time was found in this area, but beds of bentonite occur in the lower Pliocene of the Venice oil field, and thin layers of volcanic ash are present in these beds at Malaga Cove, 15 miles south of this area.

PHYSIOGRAPHY

Geologic processes active during the Quaternary period have produced some remarkable topographic forms in the eastern part of the Santa Monica Mountains and in the adjoining lowland to the south. Although no exhaustive study of the physiography of this area has been made, the following brief consideration of the origin of some of the most conspicuous of these features may serve to evaluate, in a general way, the relative importance of the various processes in the development of the present landscape.

SANTA MONICA MOUNTAINS

The Santa Monica Mountains constitute one of the features of topographic relief in southern California. Several lines of evidence, but particularly the occurrence of elephant remains in Pleistocene rocks on some of the Santa Barbara Islands, lead to the belief that these mountainous islands were connected to the mainland and formed a westward extension of the Santa Monica Mountains during comparatively late geologic time. These islands, arranged in an east-west chain, are now separated by deep submarine troughs, and this arrangement of islands and troughs is characteristic of the topographic form of the sea floor over an extensive area to the south,[63] an area which forms an unusually broad continental shelf off the west coast of the Los Angeles Basin. A study of the form and arrangement of relief on this offshore belt has led to the belief that this portion of the continental shelf, including what was once a part of the Santa Monica Mountains, has obtained much of its present form through fault displacement of a series of comparatively small more or less rectangular blocks, the up and down movements of which have resulted in a somewhat heterogeneous arrangement of islands and submarine basins and uplands. Certainly much and possibly all of this deformation by block faulting has taken place during or since Pleistocene time. These statements apply to a district outside the limits of the present investigation, a district where critical geologic data in support of these statements are largely concealed from view.

Most of the eastern part of the Santa Monica Mountains presents a remarkably subdued topographic form and, except for its contrast in altitude to the surrounding plains, can hardly be considered mountainous. Altitudes along the crest of this part of the range commonly run from 1,300 to 2,100 feet, or 800 to 1,500 feet above the adjoining plains. The crest, as shown in Plate 30, *A*, is characterized by flat-topped ridges which are strikingly concordant in general altitude. These ridge tops are remnants of an earlier physiographic surface that was developed across folded upper Miocene rocks of the Santa Monica anticline—a surface of probable early Pleistocene age which apparently had advanced to a state of old age prior to the uplift, or series of uplifts, that initiated the present drainage system.

The position of the main drainage divide of the eastern part of the range is one of the interesting physiographic features of the district. Inspection of the topography mapped on Plate 16 shows that this divide, instead of occupying a somewhat central position within the range, lies much nearer the northern edge and holds a general east-west course. The character of the geologic structure and the distribution of various types of rock throughout the mountains preclude the possibility that these factors have been influential in determining the position of this divide; it seems that the major controlling factor lies in the difference in the altitude of the two adjoining plains and in the distances which the northward and southward draining streams must flow to reach base level—distances which, of course, are inversely proportional to the cutting power of the streams.

Although the mountains are subdued and remarkably uniform in altitude over broad areas, there are higher and more abrupt minor topographic features that can, with considerable certainty, be ascribed to faulting, which, it is believed, has taken place in comparatively late geologic time, possibly during or at the end of the Pleistocene epoch. The San Vicente Mountain area, just west of upper Sepulveda Canyon, and the wedge-shaped granite mass bounded by the arcuate fault just west of upper Laurel Canyon are features of this sort. It also seems probable to the writer that the steep granite front of the southern border of the mountains north and northwest of Hollywood has resulted from major displacement along the Hollywood fault during or since the Pleistocene, a displacement which appears to have terminated on the west at the north end of the Newport-Inglewood uplift near Beverly Hills, and to have been represented farther west not by major faulting at this time but by pronounced uplift and tilting of the old Pleistocene alluvial plain, herein called the Santa Monica Plain. (See pls. 16 and 31.) Either this is true, or else the last pronounced movement along the Hollywood fault was earlier and the post-Pleistocene uplift of the Santa Monica Mountains, which caused the tilting of the Santa Monica Plain, terminated abruptly near Benedict Canyon and did not affect the area to the east, an alternative for which there appears to be no evidence, structurally or physiographically. Except for local faulting, there seems to be good reason to believe that the part of the Santa Monica Mountains covered by this report was uplifted as a unit near the end of the Pleistocene. It appears certain, in addition, (1) that the Santa Monica Mountain uplift definitely terminated the Newport-Inglewood uplift,

[63] See Willis, Bailey, A fault map of California: Seismol. Soc. America Bull., vol. 13, No. 1, supplement, March, 1923.

(2) that both of these major structural features, trending nearly at right angles to one another, were uplifted during or at the end of the Pleistocene, possibly as a result of closely related deep-seated readjustments, and (3) that this last disturbance of the Santa Monica Mountains has produced different effects along the southern border of the range, east and west of the north end of the Newport-Inglewood uplift; to the east, major displacement along the Hollywood fault has occurred; but to the west, pronounced uplift and tilting of the adjoining alluvial Santa Monica Plain was the chief result.

SANTA MONICA PLAIN

The Santa Monica Plain, now deeply incised by canyons of the present drainage system, lies north of the city of Santa Monica and extends to the east and to the west along the southern base of the Santa Monica Mountains. (See pl. 31.) It is an inclined plain formed by continental aggradation during the later part of the Pleistocene epoch, and since that time it has been uplifted and is now being mutilated by the erosive work of the present streams. The major part of the plain slopes in general accordance with the normal profile of an alluvial plain, from altitudes of 400 to 600 feet at the foot of the mountains southward to altitudes of 200 to 250 feet along the wave-cut coastal bluffs northwest of Santa Monica. Near and east of Santa Monica portions of the plain merge imperceptibly into Recent alluvium at altitudes of 175 to 200 feet.

The age of the plain is definitely late Pleistocene, because the strata of which it is composed locally rest directly upon a slight thickness of horizontal fossiliferous marine upper Pleistocene deposits.

As indicated by the existence of several levels of terraces along the deep canyons north and northwest of Santa Monica, this plain probably has not attained its present altitude by a single continuous uplift. In the neighborhood of lower Santa Monica and Temescal Canyons these terraces are strikingly developed, and there appear to be as many as four below the broad surface formed by the plain itself. Some of them are only minor features locally developed; the uppermost terrace, however, 250 to 265 feet above sea level and about 50 feet below the level of the plain, is well preserved along Temescal Canyon near Beverly Boulevard, and an equally prominent terrace occurs along the east wall of Rustic Canyon at an altitude of 400 to 425 feet, about 50 feet below the surface of the plain.

It seems possible that the development of terraces and the dissection of the Santa Monica Plain by the present streams, particularly in the area north and northwest of Santa Monica, has not been brought about entirely by uplift. A considerable amount of it, in this area at least, may be due to the part played by marine planation in destroying the southern and coastal part of the plain, thus producing a northward recession of the coast line and forcing the streams from the north to intrench their channels. Northwest of Santa Monica, west of the mouth of Santa Monica Canyon, the coast is bordered by a sea cliff 175 to 200 feet high which consists entirely of alluvial material of the Santa Monica Plain. (See pls. 29, *A*, and 31.) The possibility exists, of course, that the prominence of this sea cliff is due largely to late Pleistocene displacement along an unexposed fault that may parallel this part of the coast, and that uplift may after all have been the only effective factor in producing the cliff. The fault across the mouth of Potrero Canyon has apparently undergone about 25 feet of post-Pleistocene movement. (See p. 127.)

Of the many factors other than intermittent uplift and marine planation which may give rise to the development of stream terraces, only one—changes in climatic conditions during the Pleistocene epoch—appears to be worthy of consideration here. Notable variations in the amount and character of the rainfall are believed to have occurred during Pleistocene time, and it is entirely possible that such variations, affecting the load and cutting power of streams, played an important part in stream-terrace development.

ECONOMIC GEOLOGY

PETROLEUM

There are two producing oil fields on the plain south of the Santa Monica Mountains.[63a] These fields, the Salt Lake and Beverly Hills oil fields, are comparatively old, having been first developed in 1903 and 1908, respectively. They were among the earliest fields of the Los Angeles Basin, and although the Salt Lake oil field was one of the large producing areas of this district between 1905 and 1912, both of these fields are now relatively unimportant.

SALT LAKE OIL FIELD

The Salt Lake oil field lies on the alluvial plain about 2 miles south of the edge of the Santa Monica Mountains, in the western part of the residential district of Los Angeles. (See pl. 32.) Its discovery was due to the presence of large seeps of heavy black oil and gas on the north side of Wilshire Boulevard and south of the area which later proved productive. The asphalt around these seeps was mined in the early days for commercial use, and it is these asphalt pits that have yielded the remarkable Rancho la Brea fauna of Pleistocene vertebrates. According to some geologists these seeps mark the location at which one of the oil-producing zones of the Salt Lake field would crop out were it not for the overlying mantle of Pleistocene alluvial-plain deposits. Commercial production was

[63a] A third, the Venice oil field, was discovered in November, 1929, but details regarding the geology of this field are not available for publication.

first obtained in this field in 1903, and the bringing in of the first well marked the beginning of a period of development that proved up the last and most westerly of a string of oil fields which trend in an east-west direction through the northern portion of Los Angeles.

GEOLOGY

The Quaternary deposits that occupy the alluvial plain bordering the Santa Monica Mountains on the south cover the surface of the Salt Lake oil field. As a result, information regarding the geology of the field can be obtained only from a study of the logs of wells. In view of the present relatively small importance of this field and its certain extinction within a few years, no detailed study has been made of the subsurface geology. The geologic description given below is a compilation from an earlier description of the field by Eldridge and Arnold [64] and from an unpublished report prepared in 1917 by Joseph Jensen and made available to the writer through the courtesy of Mr. Jensen and Mr. J. A. Taff, chief geologist of the Associated Oil Co. Mr. H. J. Steiny, of the Associated Oil Co., has provided the writer with production records of the Salt Lake and Beverly Hills oil fields.

STRATIGRAPHY

According to Eldridge and Arnold, wells drilled in the Salt Lake field have revealed a stratigraphic section which consists of 50 to 100 feet of flat-lying Pleistocene clay, coarse sand, and gravel, 1,000 to 3,000 feet of folded Pliocene clayey and sandy shale of Fernando age, and an oil zone 150 to 500 feet thick, also of Fernando age, which consists of fine to coarse sand interstratified with clayey shale and "shell." Some of the sands within the oil zone appear to be lenticular, although the main oil sand, which yielded the bulk of the oil prior to 1905, has a fairly consistent thickness of 100 to 125 feet over a large part of the field.

The Salt Lake field underwent considerable development subsequent to the study by Eldridge and Arnold. Jensen concluded in 1917 that four separate oil zones had been encountered, the upper three of which were contributing materially to the total production of the field. The stratigraphically highest zone, which he termed the upper Arcturus zone, was encountered in the western part of the field at depths ranging from 650 to 1,750 feet. This zone produced oil having a Baumé gravity of 14° to 18°. The second zone, called the lower Arcturus zone, was found about 900 feet below the top of the upper Arcturus zone and produced oil of 17° to 19° Baumé. The third zone, called by Jensen the Salt Lake zone, occurs about 2,100 feet below the lower Arcturus zone and was the most prolific oil zone of the field, having yielded practically all the oil produced from the Salt Lake Oil Co.'s property and the eastern half of the Arcturus Oil Co.'s property. The oil from this zone ranged from 9° to 22° Baumé. The deepest oil zone was practically undeveloped in 1917, but definite evidence of its presence about 1,000 feet below the top of the Salt Lake oil zone was made available by the records of some of the deepest wells.

[64] Eldridge, G. H., and Arnold, Ralph, The Santa Clara Valley, Puente Hills, and Los Angeles oil districts: U. S. Geol. Survey Bull. 309, pp. 186-195, 1907.

STRUCTURE

Arnold[65] summarized his views of the general structure of the Salt Lake oil field and its relation to the near-by Los Angeles oil fields to the east as follows:

> Practically all the productive oil sands of the different Los Angeles fields lie on the southern limb of a flexure, usually a more or less well-defined anticline, whose axis extends in a westerly direction to the region approximately half a mile north of Westlake Park, where it bends about 20° to the north and extends to a point about three-fourths of a mile southeast of Colegrove and something over a mile northeast of the Salt Lake field. Here it appears to bend again to the north, probably trending about N. 60° W. In the Los Angeles city fields—that is, between the Catholic Cemetery and the Westlake Park region—the southern limb of the flexure dips normally at angles varying from 30° to 80°, while to the west, along that portion having a northwesterly trend, the dips flatten to 20° or 25°. The Salt Lake oil field is located on the northwestern flank of a minor but probably somewhat complex fold or fault, or both, developed on the comparatively low-dipping southwestern limb of the major flexure just described. * * *
>
> The exact nature of the local flexure is not known, but it is probably an anticline, more or less complicated by faults near the apex. Its axis extends in a general northeast-southwest direction. The logs of certain wells located southeast of the lagoon appear to indicate the presence of a minor anticline developed just south of the main flexure and separated from it by a fault. Still other evidence suggests a local dome-shaped structure, or quaquaversal, having its summit in the region of the lagoon.

The above description of the structure of the field is in general accordance with the conclusion of Jensen, who studied the field after its almost complete development. The accompanying structure contour map (pl. 33), based on a map prepared by Jensen, indicates the presence near the intersection of Fourth Street and La Brea Avenue of a northwestward-plunging syncline, which strikes about N. 60°-70° W. There is a strong suggestion of a southwestward-plunging anticlinal nose northeast of the syncline; southwest of the syncline and north of the brea pits and Wilshire Boulevard is a northward-dipping monocline which, in its western portion, makes a swing to the south in such a manner as to suggest that the dominant structural feature near and northwest of the brea pits is a northwestward-plunging anticlinal nose.

DEVELOPMENT

Although a few wells were drilled near the brea pits prior to 1903, the first producing well appears to have been completed in that year, and extensive develop-

[65] Eldridge, G. H., and Arnold, Ralph, op. cit., pp. 193-194.

ment continued until 1912. By 1910 a total of 120 wells, most of which were successful, had been drilled on the Salt Lake Oil Co.'s property, and by 1912, 47 wells had been drilled on the Arcturus Oil Co.'s property. These two properties yielded most of the oil produced in the Salt Lake field. Wells of the Salt Lake Oil Co. developed a single zone; those in the northeastern part of the property were 900 to 1,800 feet deep and yielded oil ranging from 9° to 15° Baumé, and those in the southern part of the property were 1,200 to 3,200 feet deep and produced oil of 13° to 21° Baumé.

Approximately 350 producing wells were drilled with cable tools over an area of about 1,000 acres. Although most of the successful wells had an initial production ranging from 350 to 500 barrels a day, the initial production of the Salt Lake well No. 61, the Arcturus No. 65, and the Arcturus No. 39 was 400, 700, and 1,200 barrels a day, respectively. The peak of production was reached in 1908, when, according to the records of the Associated Oil Co., this field produced 4,535,800 barrels of oil from 185 wells.

The table below shows the number of producing wells and the output for several years up to the end of 1927. Prior to June 1, 1928, the Salt Lake oil field produced 40,689,848 barrels of oil.

Production of Salt Lake oil field for certain years

Year	Number of producing wells	Production (barrels)	
		Total	Average per day per well
1904	5	500, 000	294
1905	30	1, 776, 768	174
1906	46	2, 465, 350	158
1907	130	2, 811, 800	64
1908	185	4, 535, 800	72
1915	287	1, 632, 160	17
1920	267	945, 637	10
1927	98	392, 028	12

During the period between 1905 and 1910 the Salt Lake oil field was one of the most productive fields of southern California. The field was discovered at a time when oil as a fuel was first being widely used. Many of the wells produced an oil of comparatively good grade, which, in a large percentage of the wells, was associated with considerable gas. Since 1908 the production has gradually declined until in 1929 less than 100 wells were producing an average of about 10 barrels a day. The decrease in production of this field has been accentuated by the fact that deeper drilling and possible extension of the field have been discouraged by the growth of the residential district of Los Angeles toward and within what was once the producing area of the field. (See pl. 32 and compare with pl. 23 of U. S. Geol. Survey Bull. 309. Also see pl. 33 and compared with pl. 16.) Only a small portion of the once extensive field was yielding oil in 1930, and it is understood that final abandonment of this field will be accomplished in the immediate future, in order to permit the continued growth of residential and business enterprises of Los Angeles.

BEVERLY HILLS OIL FIELD

The information regarding the production and subsurface geology of the Beverly Hills oil field is presented herewith through the courtesy of Messrs. J. A. Taff, Joseph Jensen, and H. J. Steiney, of the Associated Oil Co. This field like the Salt Lake oil field, is in the low-lying area south of the Santa Monica Mountains, which constitutes the northwestern border of what is commonly termed the "Los Angeles Basin." This oil field is 1 mile south of the city of Beverly Hills, about 2½ miles south of the southern border of the mountains, and about 3 miles west of the Salt Lake oil field.

The first producing well in this field was drilled in 1908, and since that time the field has provided only a comparatively small part of the oil produced in California. The largest well, No. 23 on the Wolfskill lease, produced an average of 400 barrels of oil a day during 1909. The year of peak production was 1912, when 20 wells produced a total of 246,223 barrels of oil, for a daily average per well of about 35 barrels. In 1927 10 wells produced 78,975 barrels of oil, for a daily average per well of 23 barrels. Prior to 1928 the field had produced a total of 2,444,646 barrels of oil. Oil from the Fox Hills No. 101 (Rodeo lease) tested 20.7° Baumé gravity and contained 20.6 per cent of gasoline and about 2 per cent of sulphur. Practically the entire production of this field is controlled by the Associated Oil Co.

The Beverly Hills oil field lies near the southern and eastern edges of the elevated Pleistocene alluvial surface that is herein termed the Santa Monica Plain, near the north end of the Newport-Inglewood uplift—a zone of structural deformation that crosses the Los Angeles Basin in a northwesterly direction and is represented physiographically by a discontinuous line of low hills and scarps. In general these closely associated hills and scarps are the surface expressions of anticlines and faults respectively, structural features which are commonly assumed to result from deep-seated faulting of somewhat greater magnitude. Eight producing oil fields occur at fairly regular intervals along this zone of deformation, and the Beverly Hills field is the northernmost one of the group.

According to studies made by G. D. Hanna for the Associated Oil Co., the Fox Hills No. 101 well (Rodeo lease), in the southeastern edge of the field, penetrated marine Pleistocene strata to a depth possibly as great as 900 feet, upper Pliocene from about 900 to 2,375 feet, lower Pliocene from 2,375 to 3,067 feet, strata representing the Pliocene-Miocene "transition zone" from 3,067 to 3,337 feet, and Miocene from 3,337 to 4,970 feet. A thickness of 900 feet for the Pleistocene seems excessive and is based on a sample from this depth which contained Pleistocene Mollusca but which may represent cavings. The beds penetrated by wells of this field consist almost entirely of clay shale, sand, and conglomerate. The major part of the oil is produced from a sandy zone just above the Miocene, in what Hanna calls the Pliocene-Miocene transition zone.

Structurally the Beverly Hills oil field, according to the subsurface structure contours by Jensen (pl. 34), is a pronounced asymmetrical dome of triangular shape elongated in an east-west direction. The structural closure of the fold is probably about 500 feet. Folding has been comparatively gentle on the north flank, where dips are about 15°. The south flank is much more abrupt, having dips of 45° or more.

This fold presumably lies just west of a northwestward-trending fault which is believed to be the extension of the major northwestward-trending fault of the Inglewood oil field, 4 miles to the southeast. The Fox Hills No. 101 (Rodeo lease), in the southeastern part of the field, is a deep well drilled to test the productivity of lower horizons of the Miocene. Between 3,600 and 4,000 feet cores that contained an abundance of oil were obtained. Laminations in the cores showed that drilling was parallel or almost parallel to the bedding of the Miocene shale between about 4,600 feet and the bottom of the hole. Such steep dips in the cores suggest that the well either is far from vertical or penetrated beds close to a fault zone.

Although the writer is not familiar with all the information provided by this deep test, the excellent oil showings reported between 3,600 and 4,000 feet indicate that additional deep tests located near the central part of the dome might have an excellent chance of obtaining oil. The Miocene Modelo formation, which this well appears to have entered at a depth of 3,337 feet, may be as much as 4,000 or 5,000 feet thick in this area, and the underlying sandstone of the Topanga formation may be as deep as 7,000 to 8,000 feet.

OIL POSSIBILITIES OF OTHER AREAS ADJOINING THE SANTA MONICA MOUNTAINS

None of the eastern part of the Santa Monica Mountains described in this report appears to be worthy of consideration for possible oil production. Future exploitation should be restricted to the adjoining low-lying areas south and north of the mountains.

A number of unsuccessful wells have been drilled on the plain south of the mountains between the Beverly Hills oil field and the city of Santa Monica. Several of these wells (see pl. 16), such as the Union Oil Co. Newlin No. 1, the Southland Petroleum Syndicate Garland No. 1, the Petroleum Securities Palms No. 1, and the Santa Monica-Sawtelle Oil Co. Birch No. 1, apparently were drilled to test the oil possibilities of topographically high areas, probably on the assumption that these high areas are the surface expression of anticlinal folds or up-faulted blocks. Interpretation of the structural significance of these physiographic features is difficult, and the drilling of additional wells for oil production will be hazardous. The writer offers no interpretation for these physiographic features but believes that this district west and southwest of the Beverly Hills oil field may contain areas that are capable of producing oil.[66]

Beneath the Santa Monica Plain northwest of Santa Monica there are several eastward-trending faults which cut Miocene, Pliocene, and Pleistocene rocks and whose effects are exposed in Potrero Canyon and other canyons farther west. (See pl. 16.) In view of the fact that some of these faults have displaced beds approximately equivalent to some of the oil-producing zones of the Los Angeles Basin and that some of these beds are beneath the surface and within reach of the drill, it is possible that commercial deposits of oil have accumulated along one or more of these faults northwest of Santa Monica or along their possible eastward extensions.

There is one structural feature along the northern edge of the Santa Monica Mountains that appears to have disturbed the otherwise continuous monocline which characterizes this flank of the mountains west of Cahuenga Pass. This feature is about 1½ miles north and northwest of the Encino Reservoir and appears to be a curving fault which has resulted in an abruptly discordant attitude of beds in the creek bed near the Caballero Country Club and the development of a low curving ridge of upper Modelo shale in the edge of San Fernando Valley, 1 mile to the northeast. (See pl. 16.) Little is known regarding the character of this fault. There is a bare possibility that oil may have accumulated along it, although there is no evidence that the structure of associated strata is such as to form a trap favorable for oil accumulation. In view of the scarcity of structural data and the apparent absence of seepages of petroleum or petroleum residues along the outcrops of the Modelo and underlying formations, the possibility of obtaining oil along this fault is considered to be slight.

LIMESTONE

A brief description of Eocene deposits of algal limestone in this area appears on page 92. Limestone has been quarried on a small scale in the upper part of Santa Ynez Canyon and has been used locally to surface roads. Some of these deposits, however, are of sufficient size and purity to warrant serious consideration of the use of this limestone in the manufacture of cement.

[66] In November, 1929, the Ohio Oil Co. completed a flowing well along the southern border of this area near Venice, good for 2,000 barrels of 24.3° Baumé gravity oil. Production was obtained from a zone 6,006 to 6,199 feet deep which occurs at the base of the Modelo, directly above a basement complex of Jurassic (?) schist.

The remarkable irregularity and discontinuous character of these algal limestone reefs are striking features for limestone deposits. The limestone ledge exposed in the quarry in the upper part of Santa Ynez Canyon is about 50 feet thick at the quarry but appears to thicken to about 100 feet as the outcrop is followed eastward. This ledge of limestone is continuous for a distance of only about 2,000 feet.

By far the largest deposit of limestone occurs about half a mile southeast of the limestone quarry. This deposit appears to form an integral part of the sedimentary deposits with which it is associated, being conformable with both underlying and overlying shale. It appears to have a maximum thickness of approximately 700 feet and lenses out entirely both to the southeast and northwest within 2,000 feet of its thickest portion. The Los Angeles Mountain Park Association has made a study of the advisability of mining this algal limestone for the manufacture of cement. Its present plans include the erection of a quarry and crushing plant near the west end of the deposit. The limestone would be quarried, ground fine enough to pass through a 200-mesh screen, and mixed with water so that the resulting fluid would have about the consistency of the rotary mud used in the drilling of oil wells. This fluid would then be transported by gravity through a 10 or 12 inch pipe line extending from the crushing plant for a distance of 4½ miles down Santa Ynez Canyon to an anchor buoy in the ocean, whence it would be transported by ship to a cement plant to be erected at San Pedro Harbor. It has been estimated that this deposit contains approximately 20,000,000 tons of limestone above the level of the bottom of the canyon in which the quarry and crushing plant may be located.[67] The limestone is associated with a shale of very good grade, some of which is itself highly calcareous and contains scattered nodules of algal limestone. The following chemical analyses were made by the Raymond G. Osborne Laboratory, of Los Angeles, and are inserted here through the courtesy of Mr. J. H. Gilliland, of the Los Angeles Mountain Park Association. The limestone analyzed represented a composite sample of 96 feet of limestone exposed in a shaft made in the lower part of the deposit.

[67] The writer is indebted to Mr. S. L. Gillan for general information as to the volume of limestone in this deposit.

Analyses of cement materials in Santa Ynez Canyon

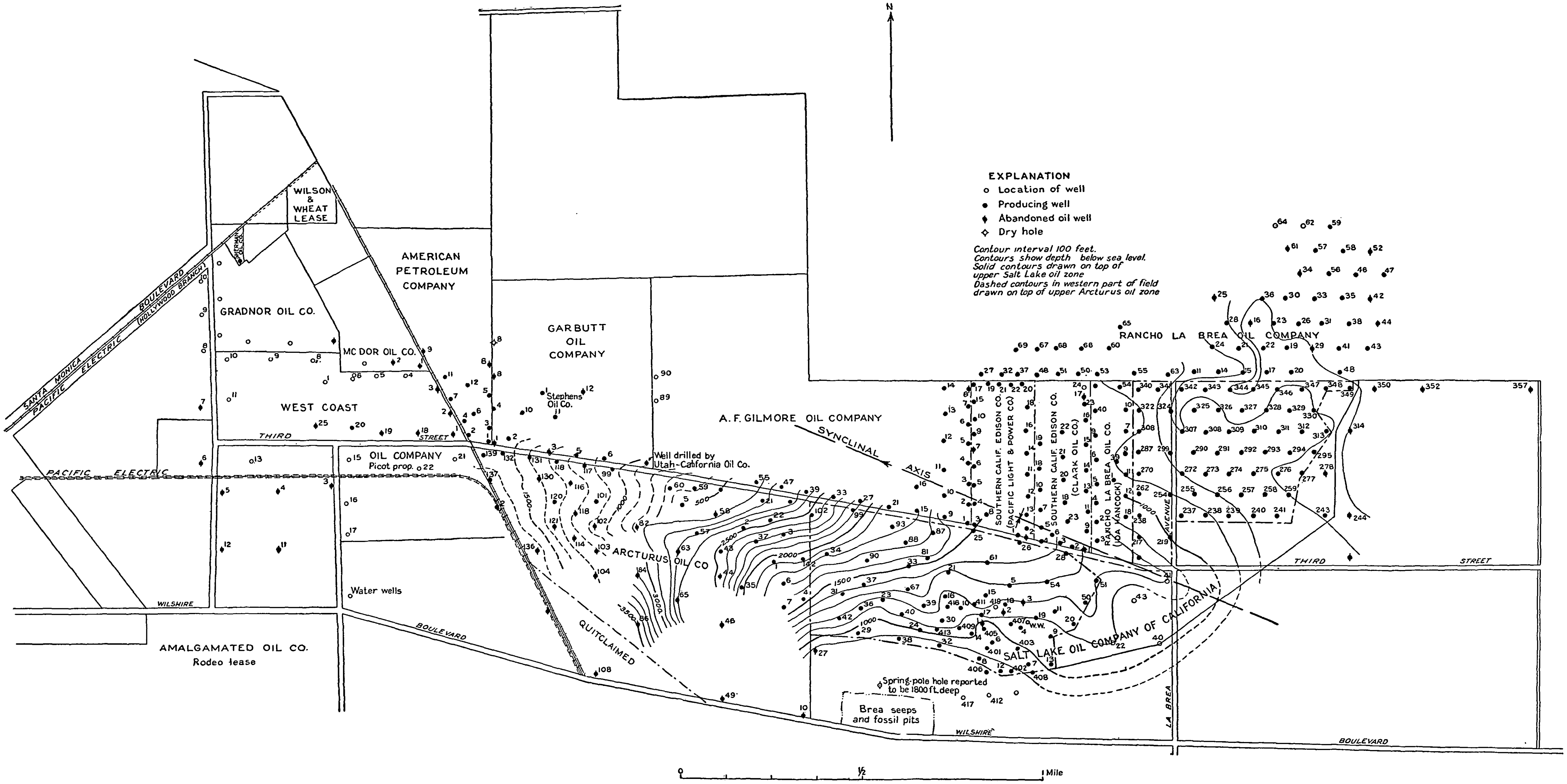
	Limestone	Shale	Raw mix [a]
Silica	3. 56	59. 26	13. 86
Alumina	. 94	15. 80	3. 69
Iron oxide	1. 86	7. 54	2. 91
Lime	50. 66	4. 10	42. 05
Magnesia	2. 13	3. 62	2. 41
Loss on ignition	40. 55	6. 58	34. 26
Sulphur trioxide	Trace.	1. 35	. 25

[a] Calculated, using 81 per cent limestone and 19 per cent shale, a mixture suggested for Portland cement.

SANDSTONE AND BASALT

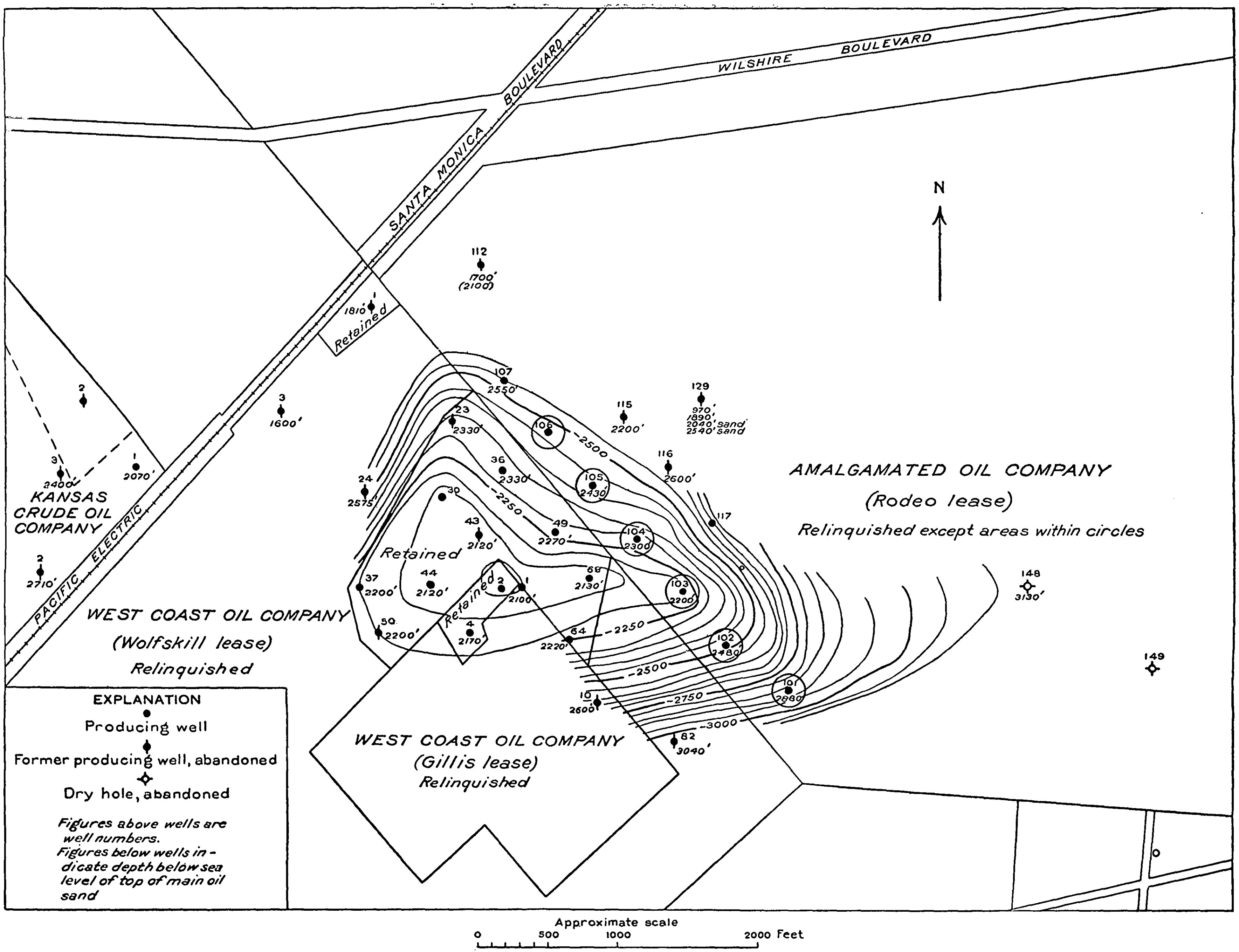
Sandstone from some parts of the Chico formation of the Topanga Canyon quadrangle and from the Topanga formation in the vicinity of Cahuenga Pass is suitable and has been used in small quantities for the construction of dwellings and retaining walls. It is commonly of medium-coarse texture, hard, and gray to greenish gray and light brownish gray.

Large quantities of deeply weathered basalt suitable for road-surfacing material are easily accessible just west of Cahuenga Pass and in Topanga Canyon. A massive hard dark-gray intrusive rock, probably diorite, is associated with hard gray Chico limestone in Santa Ynez Canyon and is being quarried for minor construction purposes.



SUBSURFACE STRUCTURE CONTOUR MAP OF THE SALT LAKE OIL FIELD, CALIFORNIA

Shows status of wells in 1917. After Joseph Jensen, Associated Oil Co.



SUBSURFACE STRUCTURE CONTOUR MAP OF THE BEVERLY HILLS OIL FIELD, CALIFORNIA

After Joseph Jensen, Associated Oil Co., 1928.

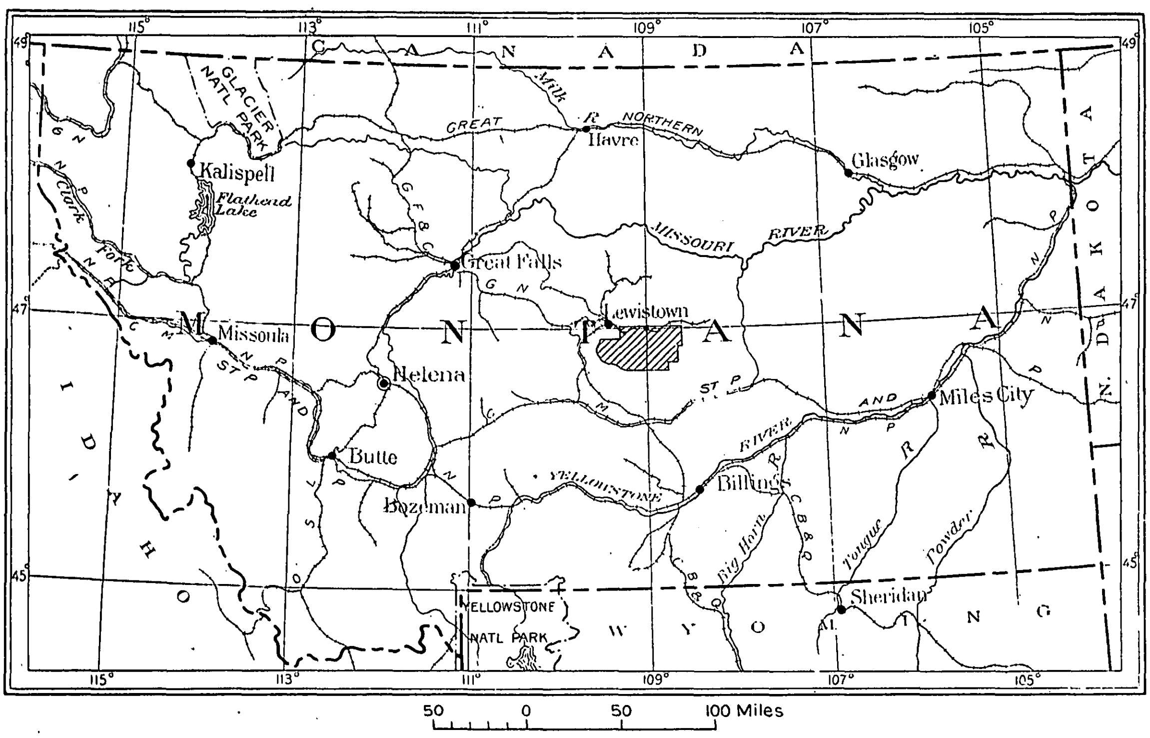
GEOLOGY OF THE BIG SNOWY MOUNTAINS, MONTANA

By Frank Reeves

INTRODUCTION

Scope of paper.—The Big Snowy Mountains form one of several isolated groups of mountains rising above the plains of central Montana. They consist entirely of sedimentary rocks that have been arched upward in a huge elliptical dome. The Big Snowy group is thus unlike the adjacent mountain groups, which consist of clusters of volcanic peaks or laccolithic domes. In its structure and stratigraphy it resembles the Black Hills of South Dakota. Such an uplift, somewhat removed from the belt of disturbance in which horizontally acting forces have apparently been dominant, constitutes an interesting problem in earth tectonics. The main purpose of the field investigations on which this paper is based was to determine the structure of the mountains. The geologic formations were therefore studied, and sufficient data were obtained to construct a combined areal and structural map. (See pl. 38.) Additional information pertaining to the character and age of the formations and the mineral resources of the area is also given.

Field work.—In 1911 W. R. Calvert made a brief geologic examination of these mountains. The writer's field work was done in May and June, 1927, with the assistance of G. E. Manger. For most of the geologic mapping, triangulation methods were used, altitudes being determined by vertical-angle readings with a telescopic alidade. The degree of accuracy of the mapping is that attained by rapid reconnaissance work, in which about four townships were mapped each week. Probably, however, detailed mapping of the area will reveal few important features not brought out in the accompanying map, except possibly in the central part of the mountains, where the difficulties offered to rapid reconnaissance by the ruggedness of the topography were accentuated by the unfavorable conditions accompanying a very cold and wet season, during which snow lay in deep drifts on the mountain ridges, and many of the trails through the canyons were impassable because of swollen streams. Three townships previously mapped by Calvert [1]—T. 14 N., Rs. 17, 18, and 19 E.—have been included in the accompanying geologic map.



(Figure 9.—Index map of Montana showing location of Big Snowy Mountains (ruled area)

[1] Calvert, W. R., Geology of the Lewistown coal field, Mont.; U. S. Geol. Survey Bull. 390, 1909.

Location and access.—The Big Snowy Mountains are approximately in the center of the State, in Fergus, Golden Valley, and Musselshell Counties. They lie about 10 miles southeast of Lewistown, the nearest railroad point, one of the largest towns in central Montana. (See fig. 9.) The greater part of the mountains is included in the Jefferson National Forest, and consequently there are no ranches in this part of the mountains and few in the adjacent foothills. No roads lead across the main range, and the only means of crossing it other than by arduous climbing on foot is afforded by the two or three bridle paths used by the forester on his inspection trips. Some of the canyons leading into the mountains can be entered on horseback, but many of them are so narrow or so blocked by fallen timber that they can be penetrated only on foot. Except on the ridges and in areas swept by recent fires, the mountains support a growth of pine and spruce.

TOPOGRAPHY

The Big Snowy Mountains proper form a southeastward-trending range about 24 miles long and 6 to 10 miles wide, which rises 3,000 to 4,000 feet above the surrounding plains. The Madison limestone, which forms the surface throughout most of the mountains, is so resistant to erosion that it has given rise to their present relief.

The backbone of the mountains is a narrow sinuous ridge about 8,000 feet above sea level, with here and there a few higher peaks. Big Snowy Peak, the highest, has an altitude of 8,533 feet. Half Moon Pass, with an altitude of 7,200 feet (see pl. 35, *A*, *C*), is the only noticeable gap in this backbone. Numerous deep canyons extend from the margin of the range into its central part. Many of these are narrow and have precipitous walls. A few of the deepest canyons on the south widen toward their heads and have the appearance of glacial cirques. The writer, however, agrees with Freeman [2] in ascribing the form of these canyons to the fact that at their heads the streams have cut down through limestone to underlying, less resistant shales, thereby causing the canyons to widen more rapidly there than farther down their course, where the streams are still cutting through limestone.

The boundary between mountain and plain is sharp on the south, where the steeply dipping Madison limestone makes a nearly continuous inclined rocky slope. (See pl. 35, *A*.) Two low tree-covered ridges formed by resistant members of the Quadrant and Kootenai formations lie about half a mile outside of this escarpment at most localities. Beyond these ridges are the treeless gravel benches of the plains, which have a noticeable slope away from the mountains.

The margin of the mountains on the north and east sides is not well defined, mainly because the formations there dip at a low angle. (See pl. 35, *B*.) Outside the main mountain slope, underlain by Madison limestone, the limestone beds in the top of the Quadrant formation weather into high bluffs that form the inner fringe of a high, fairly flat table-land extending to the northeast of the main range for 10 to 15 miles. This area is locally called the Little Snowy Mountains, but in reality it constitutes the foothills of the Big Snowy Range. Several large creeks cut deep canyons across this area, dividing it into three separate units, known as Alaska, Middle, and South Benches, which stand 5,300 to 5,500 feet above sea level. At the east end of South Bench there are two or three limestone peaks, locally known as the Durfee Hills. The surface in most parts of the Little Snowy Mountains is directly underlain by the limestone in the top of the Quadrant formation. A few sandstone outliers belonging to the overlying Ellis formation form low circular knolls above the general surface of the benches. The highest of these, which lies at the east end of South Bench, is Bald Butte. Northeast of these benches, across the valley of Flat Willow Creek, there is a large area of timbered land and grassy slopes lying slightly higher than the general level of the adjacent plains. This area owes its relief largely to the sandstones of the Kootenai formation. Button Butte, a conical hill in sec. 20, T. 14 N., R. 24 E., which has an altitude of 4,500 feet, is an outlier of these sandstones. It is perched in the center of a domed, circular area, in which older rocks crop out.

The Big Snowy Mountains and adjacent areas are drained largely by tributaries of the Musselshell River. Streams that head in the west end of the mountains, north of the backbone of the range, flow northward into the Judith River. The largest creeks that flow out of the mountains are Swimming Woman Creek, Rock Creek, and Cottonwood Creek. These creeks commonly contain running water during all except dry seasons. Half of the others probably contain running water only during unusually wet weather. Many of the streams in the central part of the mountains disappear abruptly when they reach the part of their stream beds underlain by the upper massive member of the Madison limestone, the water being carried out of the region through underground channels in this limestone. The entire volume of many of the streams, even during torrential rains, disappears into the limestone, and, to judge by the growth of vegetation and the accumulation of rock débris farther down the canyons, these conditions have long persisted.

SEDIMENTARY ROCKS

GEOLOGIC SECTION

The sedimentary rocks exposed in the Big Snowy uplift consist of approximately 11,000 feet of strata ranging in age from pre-Cambrian to Recent. The Quaternary formations are represented by the Pleis-

[2] Freeman, O. W., The origin of Swimming Woman Canyon, Big Snowy Mountains, Montana, an example of a pseudo-cirque formed by landslide sapping: Jour. Geology, vol. 33, No. 1, pp. 75–79, 1925.

tocene terrace-gravel deposits, which unconformably overlie the Tertiary and Cretaceous rocks around the mountains. The youngest of the underlying conformable series is the Lance formation, of Tertiary(?) age, which is present only in the highly tilted belt of rocks on the south side of the mountains.

Beneath the Lance is the conformable series of Upper Cretaceous formations commonly encountered in bordering uplifts in central Montana. This series of rocks is present in entirety only on the south side of the mountains, and even there it is generally concealed by the overlying terrace gravel.

The rocks lying between the Upper Cretaceous beds and the Madison limestone, of Mississippian age, crop out in the foothills of the Big Snowy Mountains. They are the Kootenai, Morrison(?), Ellis, and Quadrant formations. The Madison limestone forms the surface throughout the main part of the mountains. The underlying formations are of Cambrian and Algonkian age. The Cambrian formations, with the exception of the basal conglomerate, are exposed at the heads of several canyons in the central part of the mountains. The conglomerate and the top part of the underlying Belt rocks, of Algonkian age, are exposed only at the head of Swimming Woman Creek, south of Half Moon Pass.

The following table gives the sequence and distinctive features of the formations exposed in and adjacent to the Big Snowy Mountains:

Sedimentary formations exposed in and adjacent to the Big Snowy Mountains

Geologic age		Group and formation		Thickness (feet)	Character
Cenozoic.	Recent.	Alluvium.		0–50±	Wind-blown sand, flood-plain and alluvial-fan deposits of clay, sand, and gravel.
	Pleistocene.	Terrace gravel.		10–50	Deposits of limestone, boulders, and gravel forming flat-topped benches.
		Travertine.		10–25	Outliers of calcareous tuff deposited by warm springs.
	Unconformity?				
	Tertiary (?) (Eocene ?).	Lance formation.		900+	A brackish to fresh water sandy formation containing brown and gray sandstone, shale, and clay.
Mesozoic.	Upper Cretaceous.	Montana group.	Bearpaw shale.	1,000	Steel-gray to black marine shale containing beds of bentonite and lumpy concretions.
			Judith River formation.	200–400	Beds of fresh and brackish water origin containing sandstone and sandy shale.
			Claggett shale.	500–600	Dark-gray to brownish-black marine shale containing beds of bentonite and yellow calcareous concretions in upper part.
			Eagle sandstone.	150–250	Beds of white to buff sandstone and sandy shale.
		Colorado shale.		2,250	Dark-blue to black marine shale containing beds of bentonite, calcareous concretions, sandy shale, and sandstone.
	Lower Cretaceous.	Kootenai formation.		500±	Nonmarine red and green shale, sandstone, and nodular limestone.
	Lower Cretaceous (?).	Morrison (?) formation.		125±	Variegated shales, lenses of sandstone, and thin limestone beds.
	Upper Jurassic.	Ellis formation.		130–400	Marine sandy limestone, calcareous sandy shale, and sandstone.
	Unconformity				
Paleozoic.	Pennsylvanian.	Quadrant formation.		800–1,300	Beds of marine and nonmarine red and black shale, limestone, sandstone, and gypsum.
	Mississippian.	Madison limestone.		1,950	Massive and thin-bedded marine limestone.
	Unconformity				
	Middle Cambrian.	Meagher limestone.		300	Conglomeratic limestone with flat pebbles.
		Wolsey shale.		750	Mainly greenish micaceous shale.
		Flathead quartzite.		75	Coarse sandstone with layers of quartz conglomerate.
	Unconformity				
Proterozoic.	Algonkian (Belt series).			[a] 300	Dark limy shale.

[a] Thickness exposed.

ALLUVIUM

Deposits of Recent age, composed of sand, clay, and gravel, form the bottom lands of flood plains bordering the larger streams of the area. These deposits are best developed along Flat Willow and McDonald Creeks. Alluvial fans of Recent age are not conspicuously present. Wind-blown sand derived from sandstones of the Kootenai formation covers small areas in T. 13 N., R. 24 E. None of the alluvium in the area was mapped.

TERRACE GRAVEL

In the plains bordering the south flank of the Big Snowy Mountains and to the south and southeast of the so-called Little Snowy Mountains there are numerous terraces having an even plainsward slope of 1° to 2°. These terraces are largely underlain by deposits, 10 to 50 feet thick, of limestone boulders and gravel derived mainly from the Madison limestone and Quadrant formation. They are remnants of former coalescent alluvial fans formed by torrential streams that flowed out of the mountains and deposited their load of detrital material on the adjacent plains. These deposits of terrace gravel are most widespread on the south side of the Big Snowy Mountains, forming an almost continuous bench that conceals the underlying Cretaceous formations in a broad belt lying outside of the ridge made by the Kootenai formation. (See pl. 35, *C*.) Near the mountains these terraces rise to altitudes of 5,200 to 5,400 feet and have a plainsward slope of approximately 2°. Southeast of the Little Snowy Mountains there are also several gravel terraces at different levels, only the highest of which are shown on the map. The largest of these terraces lies directly east of the Durfee Hills along the south bank of Flat Willow Creek, which now occupies a valley 300 feet below the terrace.

The benches on the north and northeast sides of the Big Snowy and Little Snowy Mountains are not conspicuously developed. All the gravel terraces in the area, according to W. C. Alden,[3] are of Pleistocene age.

TRAVERTINE

Deposits of calcareous tuff or travertine cover two small areas a few miles north of the mountains. One of these forms Castle Butte, in sec. 14, T. 14 N., R. 19 E., and the other caps a hill in T. 14 N., R. 20 E. These are the most southerly of several deposits of travertine that Calvert[4] mapped in the Lewistown coal field. He attributed them to warm-spring activity in connection with the laccolithic intrusions of the Judith and Moccasin Mountains. (See pl. 37.) The writer accepts Calvert's conclusions as to the warm spring origin of the travertine deposits but questions whether the water owed its heat to intrusive rocks. The reasons for this difference of opinion are as follows: The conditions of structure and topography are favorable to an artesian circulation of ground water that would produce warm springs, whether or not there were intrusive rocks in the area. The pronounced doming of the formations in the main part of the Judith Mountains and minor doming in the adjacent plains make it possible for surface water that has entered porous beds at their outcrop in the mountains to pass down to considerable depths, to attain the temperature of the earth at those depths, and then to escape at the surface in warm springs. Artesian circulation in the region is attested by the fact that large flows of fresh water are commonly encountered in the Kootenai sandstones and in the top part of the Madison limestone in most wells drilled for oil, and also by the fact that in the mountains many of the streams sink into the Madison limestone. An examination of Calvert's map of the Lewistown coal field[5] shows that the travertine is grouped around the minor domes in the plains rather than around the major domes of the mountains, where the warm springs should have been the most common and lasting if they had been the result of igneous activity. The travertine overlies the truncated edges of the tilted rocks in the domes. This indicates that it was deposited so long after the intrusive activity that the small igneous masses forming the domes would probably have had time to cool off. There is now a warm spring with a temperature of about 70° F. in a small dome on the west side of the Judith Mountains, which Calvert thinks is probably due to artesian circulation. This spring is reported by Palmer[6] to be depositing travertine. Accordingly no good reason appears for attributing a different origin to the earlier warm springs in the same region. Warm springs may have been more active in the past than now, but this can reasonably be explained by supposing that decreasing rainfall and progressive peneplanation have lowered the water table in the mountains so that there is now insufficient head for the water to rise to the surface in the plains to the extent that it formerly did.

LANCE FORMATION

Rocks belonging to the Lance formation crop out in three or four localities on the south flank of the Big Snowy Mountains and probably underlie the gravel terraces there over a considerable area. The only locality, however, where more than a few feet of the formation is exposed is in sec. 9, T. 11 N., R. 17 E. Here the Lance consists of about 900 feet of sandstone and sandy shale, all standing nearly vertical, except the topmost 50 feet of beds, which lie flat. (See structure section A–B, pl. 38.) The character of the beds is shown by the following section:

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[3] Oral communication.

[4] Calvert, W. R., Geology of the Lewistown coal field, Mont.: U. S. Geol. Survey Bull. 390, pp. 34–40, 1909.

[5] Calvert, W. R., op. cit., pl. 1.

[6] Palmer, H. S., The South Moccasin Mountains, Mont. (unpublished paper in the library of the U. S. Geol. Survey), p. 120, 1923.

Section of Lance formation in sec. 9, T. 11 N., R. 17 E.

	Feet
Terrace gravel.	
Soft medium-grained, irregularly bedded gray sandstone	100
Soft coarse-grained dark-brown sandstone	300
Soft cross-bedded argillaceous sandstone in beds 5 to 10 feet thick, interbedded with yellow and greenish sandy shale	500
Dark Bearpaw shale.	

MONTANA GROUP

The Montana group consists of four conformable Upper Cretaceous formations—the Bearpaw shale, Judith River formation, Claggett shale, and Eagle sandstone. These formations are present only in the highly tilted belt of rocks lying along the south side of the Big Snowy and Little Snowy Mountains, and even there they are exposed only in a few isolated outcrops along the margin of the gravel terraces. There are sufficient exposures, however, to indicate that these formations lie beneath the gravel terraces throughout a wide area along the southern margin of the area mapped. The only locality where they are fairly well exposed is in secs. 4 and 9, T. 11 N., R. 17 E. In general the shale formations have approximately the same thickness and character as they have in the plains east of the range. The sandstone formations, however, show a slight thickening toward the southwest.

COLORADO SHALE

The Colorado shale has a wide distribution at the east end of the Little Snowy Mountains and is present in the highly tilted belt of rocks along the south side of the Big Snowy Mountains. In the greater part of this belt, however, it is concealed by terrace gravel like the other Upper Cretaceous rocks. The only locality in the area mapped where the entire thickness of the Colorado shale is well exposed is the belt of tilted rocks at the southwest side of the Big Snowy Mountains, in sec. 4, T. 11 N., R. 17 E., where it is about 2,250 feet thick. The following section of the Colorado shale, which is numbered from top to bottom, was measured in this locality:

Section of Colorado shale in sec. 4, T. 11 N., R. 17 E.

	Feet
Eagle sandstone.	
Colorado shale:	
1. Dark-blue shale with gray calcareous concretions and bentonite beds	650
2. Brown calcareous bed containing black pebbles in top part	2
3. Dark-blue shale	15
4. Dark clay shale containing black pebbles and ironstone concretions that weather into small red chips	20
5. Dark shale with bentonite beds poorly exposed	700
6. Series of sandstone beds and sandy shale, the sandstone beds varying in thickness from 6 inches to 2 feet. The thicker beds are massive and coarse grained, with some small black pebbles. The thinner beds are fine grained and contain partings of chert	50
Colorado shale—Continued.	
7. Dark shale, poorly exposed	200
8. Coarse-grained sandstone	4
9. Dark shale, poorly exposed	600
10. Well-bedded fine-grained sandstone: Beds of tan-colored sandstone, 2 to 6 inches thick, interbedded with sandy shale	20
Red Kootenai clay shales.	

Although it is not possible definitely to correlate beds in the above section with beds in the Colorado shale in other parts of Montana, the writer tentatively correlates bed 4 with the upper "red-chip zone" of the Highwood area.[7] It also appears probable that the sandstone series of No. 6 represents the Mowry shale. The sandstone of No. 10 is very persistent throughout central Montana and forms prominent escarpments and dip slopes in the eastern part of the area mapped, where it attains a thickness of 40 to 60 feet. This is the First Cat Creek sand of the Cat Creek oil field. The writer, following the earliest workers in the region, has included this sandstone in the Colorado shale, considering it to be the basal member. Bowen,[8] in his report on the area south of the Big Snowy Mountains, included it tentatively in the Kootenai formation, and it has been considered the equivalent of the Dakota sandstone, as that term has been loosely applied in the Northwest.

KOOTENAI AND MORRISON (?) FORMATIONS

Between the Colorado shale and the Ellis formation, of Upper Jurassic age, there are 600 or 700 feet of beds of nonmarine origin which, for the most part, belong to the Kootenai formation. The lower 125 feet is classed as Morrison by Calvert[9] in his study of the Lewistown coal field, but owing to the general similarity of these beds and the absence of definite criteria for subdividing them, they have been grouped together in this report. Although the beds vary markedly from place to place, the following general statement of their character may be made. At the top there is a series 150 to 300 feet thick of red and gray clay shales and lenses of white and gray sandstone, the outcrop of which usually forms a red clay soil. The middle part consists of lenticular beds of coarse-grained sandstone and interbedded red shale. To this series of beds belong the Second and Third Cat Creek sands of the Cat Creek oil field. The outcrop of these sandstones forms the low timbered ridge along the south flank of the Big Snowy Mountains. To the north of the mountains a bed of bituminous coal occurs immediately beneath the thickest and most persistent of the sandstones. The lower series consists of maroon, gray, and green shales and thin sandstones and has a thickness of about 200 feet.

[7] Reeves, Frank, Thrust faulting and oil possibilities in the plains adjacent to the Highwood Mountains, Mont.: U. S. Geol. Survey Bull. 806, p. 162, 1929.

[8] Bowen, C. F., Anticlines in a part of the Musselshell Valley, Musselshell, Meagher, and Sweetgrass Counties, Mont.: U. S. Geol. Survey Bull. 691, p. 197, 1918.

[9] Calvert, W. R., Geology of the Lewistown coal field, Mont.: U. S. Geol. Survey Bull. 390, p. 22, 1909.

ELLIS FORMATION

The Ellis formation, of Upper Jurassic age, is exposed entirely around the Big Snowy uplift. On the north side of the mountains it has a thickness of 300 to 400 feet, but on the east and south sides its thickness is only 130 to 175 feet. The formation everywhere has a top member of thin-bedded, calcareous, and glauconitic sandstone that weathers brownish yellow and commonly contains many oyster remains, which in some localities form a bed of marl. East and south of the mountains the sandstone is 40 to 50 feet thick. In some localities north of the mountains Calvert reports it to be 150 feet thick. East and south of the mountains the beds beneath the top sandstone member consist uniformly of 100 to 125 feet of sandy and limy shales which contain numerous fossils, among the most common of which are *Gryphaea calceola* and *Belemnites densus*. On the north side of the mountains this part of the formation contains a variety of beds, such as red and green shales, limestone conglomerate, and gypsum. On this side of the mountains, too, the underlying Quadrant formation abruptly decreases in thickness because of unconformity.

The following sections show the character of the Ellis formation in different localities:

Section of Ellis formation on bluff of East Fork of Big Spring Creek, sec. 2, T. 14 N., R. 19 E.[10]

	Feet
Morrison formation.	
Ellis formation:	
Partly concealed; lowest 30 feet sandy soil, sprinkled thickly with *Gryphaea calceola*, underlain by a ledge of sandstone with abundant oyster shells; lowest 20 feet is sandstone filled with oyster-shell fragments	100
Sandstone, compact, blocky, gray, weathering tan; upper 15 feet forms bold cliff; fossiliferous at top, but shells mostly fragmentary	48
Concealed; grassy slope with red soil	80
Limestone, shaly at bottom, becoming thin bedded, dove-colored, fossiliferous	10
Shale, red, sandy	4
Gypsum, white, and pure	20
Shale, dark, fissile, gypsiferous	5
Limestone, fossiliferous	3
Partly concealed but containing gypsiferous shale	45
Gypsum and shale	5
Partly concealed slope of dark soil	42
	362

Section of Ellis formation on Elk Creek, 2 miles north of Button Butte, in sec. 18, T. 14 N., R. 24 E.

	Feet
Grayish-white flaggy sandstone, weathering brownish yellow, ripple-marked in top part; some glauconite in partings	40
Dark sandy shale, only partly exposed, glauconitic at base	13
Ellis formation—Continued.	
Sandy series; basal part consists of flaggy greenish-gray sandstone members 1 to 4 feet thick, weathering dirty yellow, separated by glauconitic sands with thin clay partings which divide the sand into lentils one-eighth to 1 inch thick and a few inches long. Top of series is limy and less glauconitic	50
Dirty greenish-yellow glauconitic sandy limy shale	25
Fossil marl containing *Gryphaea calceola* var. *nebrascensis* Meek and Hayden, *Camptonectes* sp., *Cyprina? cinnabarensis* Stanton, *Pleuromya subcompressa* (Meek), *Natica* sp., *Kepplerites?* sp., *Sphaeroceras* sp.	1
Sandy limy shale	1
Quadrant formation.	
	130

[10] Calvert, W. R., op. cit., p. 20.

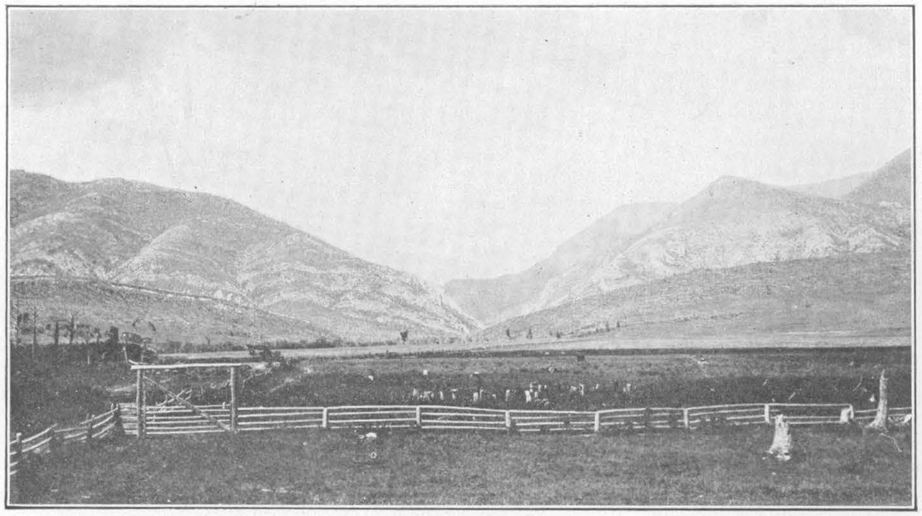
Section of Ellis formation in sec. 4, T. 10 N., R. 20 E.

	Feet
Morrison formation.	
Ellis formation:	
1. Coarse-grained tan-colored sandstone and sandy shale	65
2. Hard thin-bedded calcareous fossiliferous sandstone	10
3. Light-yellow glauconitic limy shale with numerous fossils of *Gryphaea calceola* and *Ostrea strigilecula*	15
Quadrant formation.	

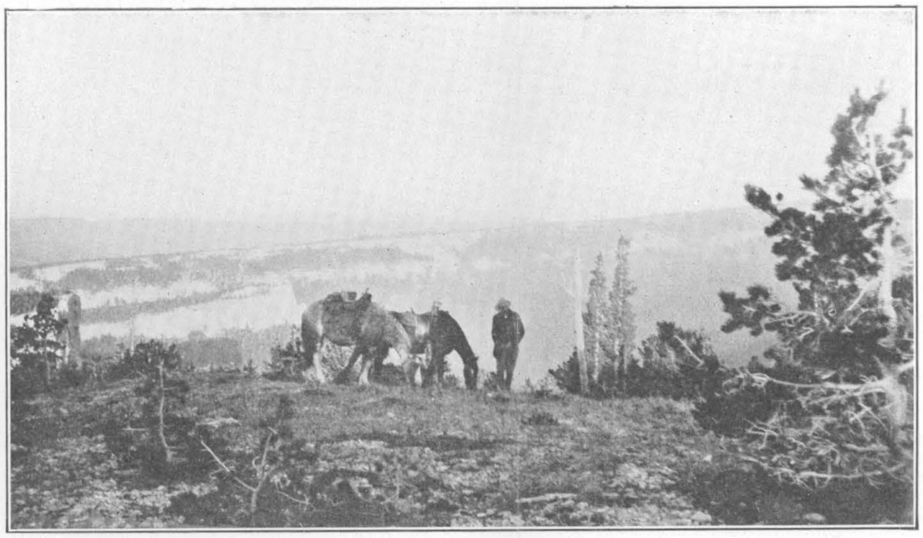
QUADRANT FORMATION

General character.—The Quadrant formation at its outcrop around the Big Snowy Mountains consists of red, gray, black, and green shales, marine limestone, brown sandstones, and gypsum beds. Its thickness is approximately 1,300 feet except along the northern margin of the area mapped, where it begins to thin rapidly in a northerly direction.

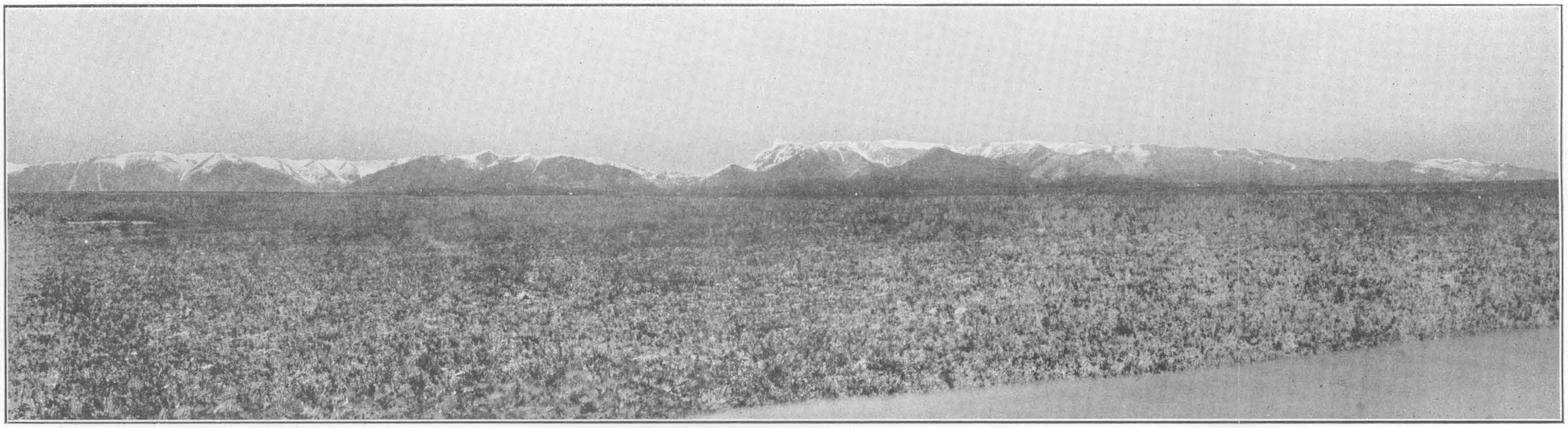
Lithologically the formation can be divided into four parts. The topmost beds throughout most of the area consist of 100 to 200 feet of thin-bedded fossiliferous limestone interbedded with red clay shale. This limestone forms a prominent ridge in the belt of steeply dipping rocks on the south flank of the mountains and weathers into red bluffs along the valleys of streams in the Little Snowy Mountains and on the north flank of Button Butte. A few miles north of the mountains this limestone series disappears, and the overlying Ellis formation rests on beds below the limestone. In one locality at the east end of the mountains along Flat Willow Creek in sec. 25, T. 13 N., R. 21 E., about 30 feet of coarse-grained white sandstone and 150 feet of red shale were found between the limestone series of the Quadrant and the limy shale at the base of the Ellis formation. Beneath the limestone series found in most areas there are 300 to 400 feet of red, brown, and black shales and sandstones. Some of the black and brown shales are carbonaceous; others are low-grade oil shale which upon distillation will yield possibly 10 gallons to the ton. The sandstones are commonly cross-bedded,



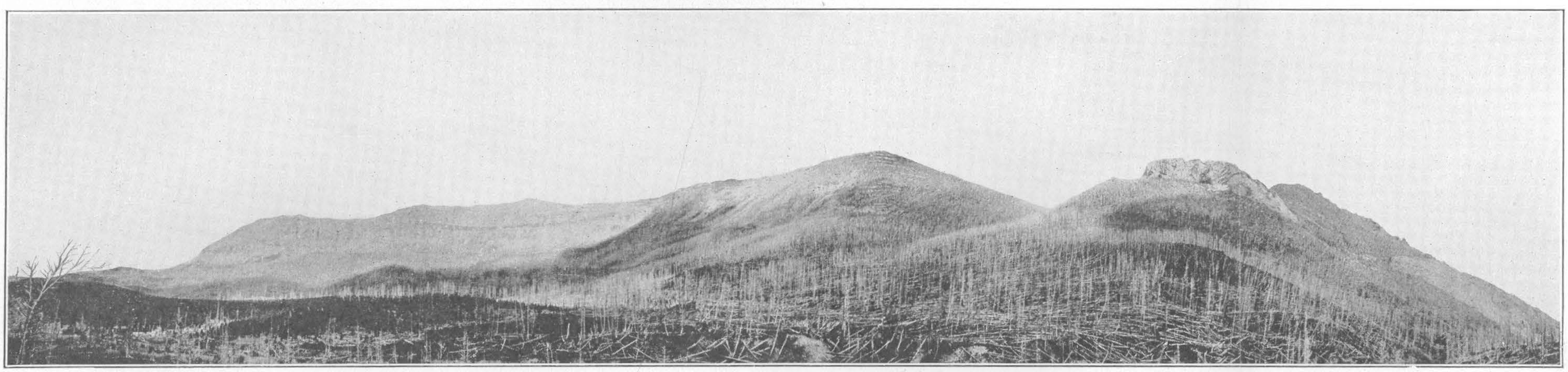
A. VIEW LOOKING NORTHWARD UP SWIMMING WOMAN CANYON TO HALF MOON PASS, MONT.



B. NORTH FLANK OF THE BIG SNOWY MOUNTAINS, MONT.
Showing the dip slope of the Madison limestone and the bluffs formed by the Quadrant formation.

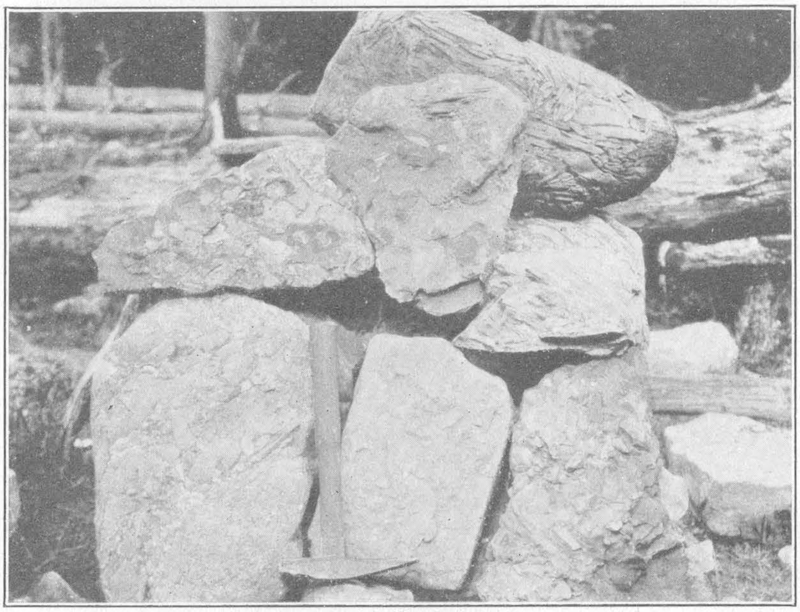


C. THE BIG SNOWY MOUNTAINS FROM THE GRAVEL TERRACES SOUTH OF THE MOUNTAINS



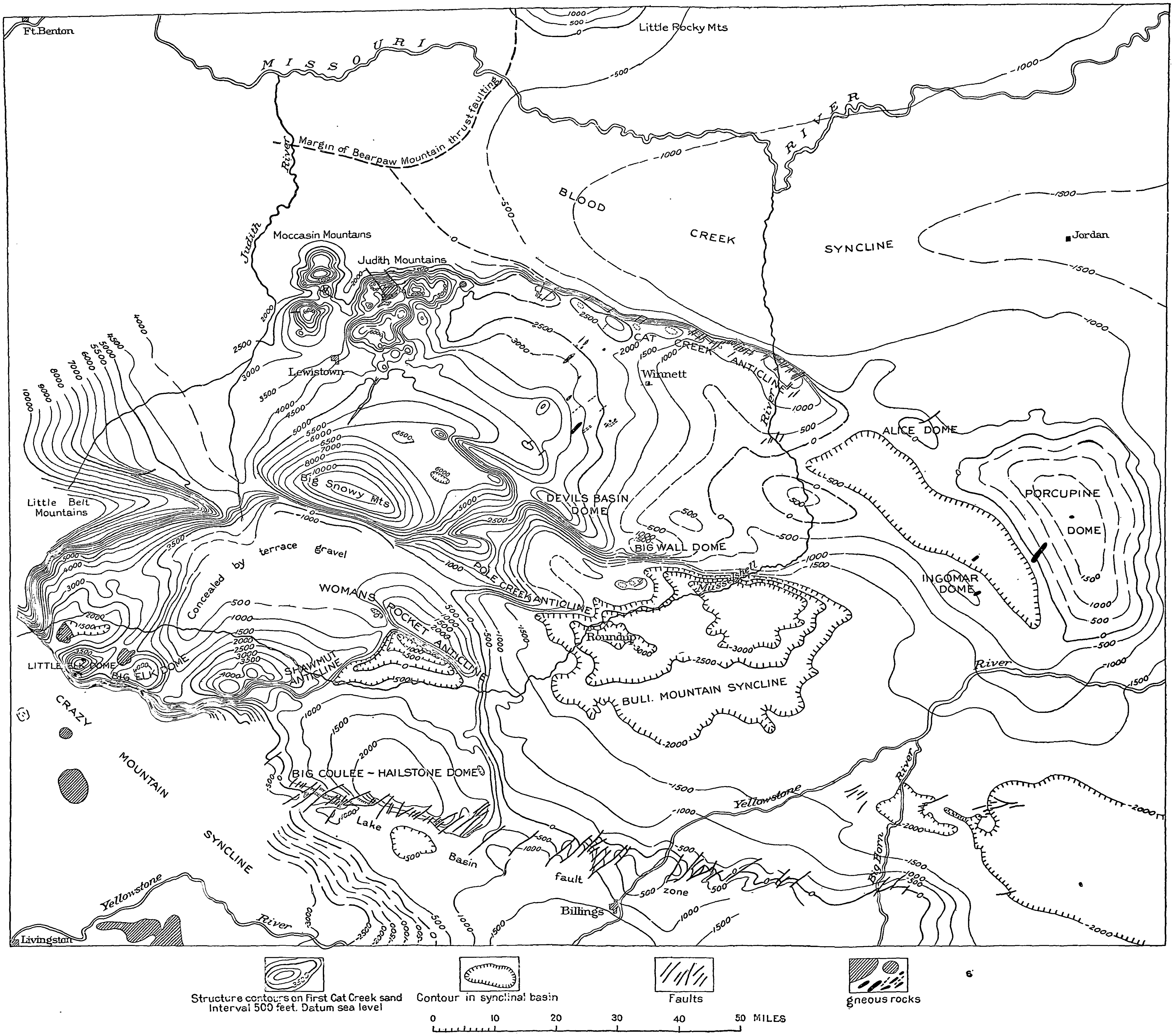
D. EAST WALL OF SWIMMING WOMAN CANYON

A



B

EDGEWISE OR INTRAFORMATIONAL CONGLOMERATES IN THE MEAGHER LIMESTONE



STRUCTURE CONTOUR MAP OF CENTRAL MONTANA

GEOLOGIC MAP AND SECTIONS OF THE BIG SNOWY MOUNTAINS AND ADJACENT AREAS, MONTANA

coarse grained and reddish brown and are 10 to 70 feet thick. In most localities there are three of these sandstones. The lowest one occurs approximately at the horizon from which the oil is obtained in the Devils Basin oil field.

Underlying the series of shales and sandstones is about 500 feet of variegated limy shale with a few thin beds of limestone. In the middle part of this series and 850 to 950 feet from the top of the formation the shale is predominantly green. On the north side of the mountains the upper part of this series consists of thin fissile calcareous shale which forms precipitous bluffs and narrow canyons. Underlying the limy shale series and occupying the basal part of the formation are 150 to 200 feet of yellow sandstone, sandy shale, and lenticular beds of gypsum.

Sections.—The following sections of the Quadrant give the characteristics of the formation in greater detail:

1. Section of upper part of Quadrant formation on north flank of Big Snowy Mountains, in sec. 17, T. 12 N., R. 20 E.[11]

[Section begins with the highest rocks exposed and probably starts near the top of the formation]

	Feet
1. Hard light-gray well-bedded limestone	15
2. Pink limy shale	8
3. Hard well-bedded light-gray fossiliferous limestone with beds of pink shale 1 to 2 feet thick occurring every 10 feet. *Composita subquadrata*[12] collected near top of series	60
4. Gray fossiliferous limestone	5
5. Pink limy shale	3
6. Massive to well-bedded gray limestone, weathering pink	30
7. Hard irregularly bedded light-gray limestone with elongated chert nodules	2
8. Thinly laminated or banded gray and pink limestone	1
9. Hard light-pink fossiliferous limestone	½
10. Red and green shale	½
11. Hard light-gray irregularly bedded limestone	3
12. Maroon and gray shale	3
13. Hard finely laminated bluish-gray limestone stained reddish brown	3
14. Maroon and green shale	8
15. Massive argillaceous and calcareous mottled sandstone	10
16. Mottled limy shale and thin-bedded limestones	10
17. Maroon and green shale with thin beds 6 inches thick of yellowish-green argillaceous limestone containing *Productus ovatus*[12]	8
18. Maroon and green shale	15
19. Hard yellowish-gray fossiliferous limestone	2
20. Maroon and green shale	3
21. Black shale	⅕
22. Grayish-yellow limestone containing ostracodes	4
23. Black petroliferous and fossiliferous shale	4
24. Hard greenish-gray calcareous clay (fossil lot 3)	2
25. Light-brown limy shale	4
26. Brittle brown petroliferous paper shale	2
27. Dark-brown shale with thin bands of red shale	25
28. Sandy shale	5
29. Thin-bedded coarse-grained sugary brownish-yellow sandstone grading upward into a sandy shale	10
30. Dark-brown shale, carbonaceous near base	30
31. Irregularly bedded to cross-bedded coarse-grained sandstone containing stems and fragments of plants and ferruginous concretions, weathering brownish yellow	15
32. Dark-blue shale changing to light gray toward top, with thin ferruginous and calcareous bands 1 to 2 inches thick near top	100
33. Massive to cross-bedded medium to coarse grained sugary sandstone, weathering brownish gray to reddish brown; lower part contains chert and ferruginous masses	70
34. Concealed	19
35. Dark shale with yellow nodular limestone at top	10
36. Grayish-green limy shale with thin flaggy beds of oolitic and petroliferous limestone and chert pebbles	?
37. Gray limy shale	35
38. White sandstone	1
39. Light-gray sandy and limy shale	50
40. Grayish-white coarse-grained sandstone	6
41. Variegated sandy and calcareous shale	20
42. Sandy argillaceous limestone with fragments of red shale	2
43. Pink limy and sandy shales, basal part concealed	70

[11] Section measured by M. I. Goldman and the writer in 1922 and published in U. S. Geol. Survey Bull. 786, pp. 53-54, 1927

[12] Determinations by G. H. Girty.

2. Section of Quadrant formation on northeast flank of Durfee Creek dome, secs. 13 and 14, T. 12 N., R. 22 E.

Ellis formation: Limy yellow shale with fossils of *Gryphaea calceola*.

Quadrant formation:	Feet
1. Thin beds of limestone and red shale (fossil lot 1)	10
2. Red clay with thin beds of limestone	70
3. Hard black to red sandstone, conglomeratic at top	5
4. Red shale	5
5. Hard red sandstone	5
6. Red clay shale and thin limestone	30
7. Gray well-bedded ridge-forming limestone	30
8. Red clay shale	15
9. Well-bedded variegated limestone and thin beds of red clay	30
10. Red limy clay	10
11. Limestone and red shale	10
12. Red shale and thin beds of limestone	30
13. Red shale	80
14. Brownish-red coarse-grained sandstone	40
15. Red shale, largely concealed	100
16. Brownish-red sandstone	30
17. Concealed; mostly shale	140
18. Thin-bedded gray and black limestone	15
19. Green shale	15
20. Hard coarsely crystalline light-brown limestone (fossil lot 7)	3
21. Green shale	20
22. Sandy yellow limestone	2
23. Green shale	40
24. Hard well-bedded black limestone	15
25. Gray and yellow limy shale	150
26. Yellow sandstone with lenses of gypsum	100
27. Sandy shale and gypsum beds	100

3. Section of Quadrant formation on south flank of Big Snowy Mountains, in sec. 36, T. 11 N., R. 20 E.

[Measured by A. A. Hammer and A. M. Lloyd [13]]

Ellis formation.	
Quadrant formation:	Feet
5. Limestone in beds 1 to 2 feet thick, gray; fossils collected at this point	42
6. Limestone with thin beds of red shale; fossils collected at this point	136
7. Red shale in lower half; upper half partly covered	91
8. Red conglomerate	3
9. Brilliant-red shale	25
10. Gray sandy shale	1
11. Light-gray shale	35
12. Probably shale, covered	16
13. Red and yellow sandstone	30
14. Dark shale	31
15. Light-brown shale	60
16. Red to gray sandstone	20
17. Shale, mostly covered, dark where exposed	41
18. Yellowish gray and pink sandstone	50
19. Black fossiliferous limestone, petroliferous	51
20. Black shale; fossils abundant	35
21. Gray to dark-gray limestone; many fossils	6
22. Dark-brown to black shale, dark gray at base	59
23. Hard gray limestone	3
24. Brown to black shale with 3-inch limestone at base	44
25. Black shale, appearing like coal	2
26. Gray limestone; remnants of *Productus*	1
27. Gray shale	29
28. Dark-gray limestone	2
29. Brown shale with thin limestone beds	53
30. Dark-gray limestone, ridge-forming	1
31. Light-gray chalklike shale	8
32. Hard limestone	4
33. Gray and green shale	15
34. Hard gray limestone	1
35. Green shale	62
36. Gray to tan limestone	6
37. Shale, mostly covered	44
38. Hard gray limestone 4 feet thick at top, with shale and limestone below	61
39. Hard gray limestone 10 feet thick, weathering white, with shale and thin limestone below	61
40. Thin series of sandstones 10 feet thick, with gray shales below	36
41. Yellow sandstone	30
42. Yellow sandstone, hard, calcareous	3
43. Gray and yellow sandstone	26
44. Gypsum, lenticular	36
45. Red limestone and shales at top, with yellow sandstone near base	64
46. Gypsum with selenite crystals	2
47. Pink to gray shale with some sandstones	19
Madison limestone.	

Quadrant-Ellis unconformity.—The contact between the Ellis and Quadrant formations is an unconformity that has a widespread extent throughout the northern Rocky Mountain region. In the area mapped this unconformity is recognizable by the fact that Upper Jurassic beds rest upon Carboniferous strata and by the lithologic features here described. In areas where the topmost bed of the Quadrant consists of the limestone series described above, the top surface of the limestone is commonly glazed and partly silicified and marked by borings one-eighth to 1 inch in diameter and one-half to 1 inch deep. It is also marked by radial cracks that are filled with clay and by potholes containing angular to well-rounded boulders of quartzite 6 inches to 2 feet thick that contain borings on their top and bottom sides, similar to those in the underlying limestone. Throughout practically all the area mapped except in the northern part and in the locality on Flat Willow Creek, heretofore mentioned, the unconformity lies between the basal limy shale of the Ellis and the limestone series of the Quadrant formation. On the north flank of the mountains, along an east-west line through the center of T. 14 N., Rs. 20, 21, and 22 E., the limestone which lies at the top of the Quadrant formation in other areas around the Big Snowy Mountains wedges out abruptly and is replaced by a few feet of conglomeratic limestone with angular pebbles. A mile or so to the north the Ellis formation rests on the Quadrant shales that elsewhere underlie the limestone series. Thence northward the Quadrant formation thins rapidly, so that at its outcrop in the Judith Mountains, according to Weed and Pirsson,[14] its thickness is only 40 feet.

In the area where the Quadrant formation begins noticeably to wedge out, the Ellis formation thickens and contains beds of red shale and gypsum. These beds occur in the lower part of the formation and appear to be an addition to the Ellis as it is found elsewhere in the Big Snowy uplift. Where they are present it is difficult to determine the contact between the Quadrant and Ellis formations, for the lithology of the contact beds is similar, and fossils are not common. The writer, however, found enough fossils to be certain that his mapping in this area is at least approximately correct. Inasmuch, however, as it does not agree with Calvert's mapping in the adjacent three townships (T. 14 N., Rs. 17, 18, and 19 E.), which are included on Plate 38 of this report, some changes have been made in Calvert's map so that the geology along the boundary of the two areas will agree.

Age of the Quadrant formation.—The term Quadrant as used in central Montana is applied to the series of strata lying between the Madison limestone, of lower Mississippian age, and the Ellis formation, of Upper Jurassic age. Although it is definitely known that these strata are of Carboniferous age, there has been some uncertainty as to the particular part of the Carboniferous to which they belong. This uncertainty has been due in part to the scarcity and unsatisfactory character of the fossil collections and in part to the peculiarities of the fauna, which, according to

[13] Am. Assoc. Petroleum Geologists Bull., vol. 10, pp. 988–989, 1926.

[14] Weed, W. H., and Pirsson, L. V., Geology and mineral resources of the Judith Mountains of Montana: U. S. Geol. Survey Eighteenth Ann. Rept., pt. 3, pp. 471–473, 1898.

Girty,[15] differs considerably from any of the other known American Carboniferous faunas. The fossil collections now at hand, however, make it possible to arrive at a fairly definite determination of the age of the Quadrant in central Montana. The following is a list of the fossils collected during the writer's field work in the Big Snowy area. The determinations were made by G. H. Girty.

Lot 1. From bed 1 of section 2, in highest bed of limestone series at top of Quadrant, in sec. 14, T. 12 N., R. 22 E.:

Orthotetes n. sp.	Nucula aff. N. illinoisensis.
Sphenotus aff. S. octocostatus.	Leptodesma aff. L. nasutum.
Sphenotus aff. S. monroensis.	Bucanopsis n. sp.
Schizodus sp.	Naticopsis aff. N. waterlooensis.
Deltopecten aff. D. monroensis.	Sphaerodoma aff. S. littonana.
Deltopecten sp.	Euomphalus? sp.
Schizodus aff. S. depressus.	

Lot 2. Near base of limestone series at top of Quadrant, in sec. 6, T. 10 N., R. 20 E.:

Productus ovatus?	Spiriferina aff. S. spinosa.
Pustula aff. P. genevievensis.	Composita trinuclea.
Camarotoechia aff. C. mutata.	Cleiothyridina sublamellosa?

Lot 3. From bed 24 of section 1, 75 feet below limestone series at top of Quadrant, in sec. 17, T. 12 N., R. 20 E.:

Enchostoma sp.	Composita aff. C. subquadrata.
Lingula sp.	Sphenotus aff. S. octocostatus.
Orbiculoidea n. sp.	Aviculipecten sp.
Schizophoria n. sp.	Modiola fontainensis?
Chonetes aff. C. sericeus.	Leptodesma? sp.
Productus ovatus var. latior.	Trepospira? sp.
Productus ovatus var. minor?	Naticopsis n. sp.
Avonia arkansana.	Meekospira? sp.
Martinia aff. M. contracta.	Cytherella? sp.

Lot 4. From limy shale about 100 feet below limestone series at top of Quadrant, in sec. 16, T. 14 N., R. 22 E.:

Lingula n. sp.	Avonia sp.
Orthotetes kaskaskiensis.	Composita? sp.
Productus ovatus?	Aviculipecten sp.
Productus aff. P. setiger.	

Lot 5. From limy shale about 300 feet below limestone series at top of Quadrant, in sec. 20, T. 14 N., R. 22 E.:

Lingula sp.	Orthotetes kaskaskiensis.

Lot 6. From limy shale about 300 feet below limestone series at top of Quadrant, in sec. 11, T. 11 N., R. 21 E.:

Amplexus n. sp.	Composita subquadrata.
Productus ovatus.	Cliothyridina sublamellosa.
Spirifer increbescens.	

Lot 7. From bed 20 of section 2, 500 feet below limestone series at top of Quadrant, in sec. 13, T. 12 N., R. 22 E.:

Rhombopora sp.	Girtyella aff. G. turgida.
Camarotoechia aff. C. mutata.	

Mr. Girty [16] states that none of these collections contains a characteristic Chester or a characteristic Pennsylvanian fauna, although the fauna of lot 1, which is by far the most varied, might be an obscure phase of the Chester or of the Pottsville. On the other hand, it contains many forms suggesting species that occur in the Ste. Genevieve fauna, of Mississippian age. The other lots from underlying strata may be Ste. Genevieve also. At any rate, the pronounced faunal change that was initiated with the horizon of lot 1 suggests that the underlying beds belong to a different geologic epoch, and they can be safely assigned to the Mississippian. Another fossil lot collected by A. A. Hammer from bed 5 of section 3, in the top part of the limestone series, in sec. 36, T. 11 N., R. 20 E., was examined by Mr. Girty, who reports that it is clearly a Pottsville fauna. From these fossil data it appears fairly certain that all the Quadrant of central Montana is of Mississippian age with the exception of the top part of the limestone series, which is probably of Pottsville age. It is evident, therefore, that the practice of correlating the Quadrant of central Montana with the Tensleep sandstone and Embar formation of southern Montana and northern Wyoming is not justified. The faunal evidence available indicates that the Tensleep and Amsden are of Pennsylvanian age, with the possibility that the lower part of the Amsden in some areas is Mississippian. It appears probable that the Amsden may be equivalent to the limestone series at the top of the Quadrant. This inference is in agreement with the faunal data and is supported by the fact that the limestone series is somewhat similar in lithology to the basal part of the Quadrant in southwestern Montana, which, on lithologic grounds, is generally considered the approximate equivalent of the Amsden. Whether or not this correlation is correct, it is evident that the greater part of the Quadrant in central Montana is older than the Quadrant of southwestern Montana and the Tensleep and Amsden of southern Montana and Wyoming.

MADISON LIMESTONE

Underlying the Quadrant formation with apparent conformity is the Madison limestone, of lower Mississippian age. This limestone forms the bulk of the mountains, and its outcrop is characterized by gray rocky surfaces which, except on the highest ridges, support a growth of pine and spruce. The upper part of the formation consists of massive beds of light-gray limestone which form the steep rocky slopes on the south flank of the mountains and the gentle slopes and narrow boxlike canyons on the north side. (See pl. 35, *A*, *B*, *D*.) The lower part of the formation is composed of thin-bedded gray to brown limestone, some beds of which are petroliferous. This part of the formation crops out in the crest of the range and forms the walls of many of the canyons on the south side of the mountains. The thickness of the Madison limestone in the Big Snowy Mountains is about 2,000 feet, or about twice the thickness that it attains in adjacent ranges. The formation is very fossiliferous, especially in its lower thin-bedded part. No fossil collections were made in the formation except two from the thin-

[15] Girty, G. H., in Calvert, W. R., U. S. Geol. Survey Bull. 390, p. 19, 1909.
[16] Personal communication.

bedded limestone which Calvert [17] regarded as comprising the basal part of the Madison limestone but which is similar in both lithology and stratigraphic position to beds in the Little Belt Mountains classed by Weed [18] as Silurian and Devonian. According to G. H. Girty, who identified these fossils, they are of Madison limestone age and confirm Calvert's conclusion that there are no beds in the Big Snowy Mountains between the Madison limestone and the Meagher limestone, of Cambrian age. The fossils are listed below.

Lot 1. Basal part of Madison limestone on Ashley Creek in sec. 13, T. 11 N., R. 19 E.:

Crania aff. C. levis.
Schuchertella chemungensis var.
Productus aff. P. burlingtonensis.
Camarotoechia aff. C. metallica.
Camarotoechia aff. C. herrickana.
Camarotoechia several indet. sp.
Girtyella sp.
Spirifer centronatus.
Spirifer mysticensis.
Spirifer n. sp.?
Reticularia cooperensis?
Martinia rostrata?
Composita?
Eumetria verneuiliana.
Bellerophon sp.

Lot 2. From lower part of Madison limestone in Half Moon Pass:

Syringopora sp.
Platycrinus sp.
Anisotrypa sp.
Schuchertella chemungensis var.
Chonetes loganensis.
Productus gallatinensis.
Spiriferina solidirostris.
Spirifer mysticensis.
Eumetria verneuiliana.
Platyceras sp.

The following fossils were collected by Calvert [19] from the Madison in the vicinity of Half Moon Pass:

Lot 3:
Menophyllum sp
Productus semireticulatus.
Productus parviformis.
Spirifer centronatus.
Lot 4:
Stenopora sp.
Schuchertella inflata.
Spirifer centronatus.
Platyceras sp.
Lot 5:
Menophyllum? sp.
Productus gallatinensis.
Lot 6:
Favosites sp.
Syringopora surcularia.
Zaphrentis sp.
Menophyllum sp.
Fenestella sp.
Schuchertella inflata.
Chonetes illinoisensis.
Camarotoechia sp.
Spirifer centronatus.
Composita sp.
Eumetria verneuiliana.
Euomphalus? sp.
Platyceras sp.
Lot 7:
Fenestella, several sp.
Pinnatopora sp.
Spirifer centronatus.
Lot 8:
Leioclema sp.
Derbya crassa.
Schizophoria n. sp.
Productus cora.
Lot 9:
Derbya crassa.
Spirifer rockymontanus.
Composita subtilita.

Girty states that

Lots 1 and 2 are of Mississippian age. Lots 4, 5, 6, and 7 contain the well-known fauna of the Madison limestone, typical in every respect. Lots 8 and 9 are without much doubt Pennsylvanian but very early Pennsylvanian, probably Pottsville. Lot 8 particularly recalls a fauna which in the Montpelier region, Idaho, comes in only a few feet above the upper Mississippian and which I am considering the lowest Pennsylvanian.

It appears doubtful whether the fossils of lots 8 and 9 were collected from the vicinity of Half Moon Pass, because all formations younger than the Madison limestone are eroded from this part of the mountains, although it is possible that they may have come from talus slopes originating from former but now eroded beds of the Quadrant formation.

MEAGHER LIMESTONE

The pre-Carboniferous formations that crop out in the Big Snowy Mountains have not been studied in detail by the writer. Calvert [20] made a brief examination of them at their outcrop in Swimming Woman Canyon in 1911 and made the subdivisions which the writer is using in this report. A more thorough study of them may result in a slightly different formational grouping, but there is probably little doubt that with the exception of the readily recognizable Algonkian Belt rocks they are of Cambrian age. The top 300 feet of the series consists of calcareous shales and thin limestone conglomerates. Calvert has correlated these beds with the Meagher limestone studied by Weed [21] in the Little Belt Mountains. This correlation is based on their lithology and stratigraphic position but is not supported by fossil evidence. That they are of Cambrian age, however, can scarcely be questioned, because they contain the limestone conglomerates that are a distinctive feature of Cambrian rocks in adjacent mountain uplifts. These limestone conglomerates occur every 25 or 50 feet in a series of limy shales and are 2 to 4 feet in thickness. The pebbles of the conglomerates are fine-grained limestone firmly cemented by a yellow calcareous cement. The pebbles are smooth and flat and range in size from that of a wheat grain to 2 inches in diameter. The larger pebbles are 0.2 to 0.3 inch thick and are angular; the smaller ones are slightly rounded. Some are pierced by cylindrical holes about 0.1 inch in diameter. Most of the larger pebbles occupy flat-lying positions parallel to the bedding planes. The smaller ones, however, commonly form a jumbled mass. (See pl. 36.) Many of the pebbles are stained light green. On fresh fracture they are bluish gray. These conglomerates are called intraformational or edgewise conglomerates and are of the type which Hadding [22] calls regressional interformational conglomerates and considers to be the result of the reworking of sediments shortly after deposition.

[17] Calvert, W. R., unpublished report on Paleozoic formations in the Big Snowy Mountains, cited by Walcott, C. D., Relations between the Cambrian and pre-Cambrian formations in the vicinity of Helena, Mont.: Smithsonian Misc. Coll., vol. 64, pp. 275-276, 1916.

[18] Weed, W. H., U. S. Geol. Survey Geol. Atlas, Little Belt Mountains folio (No. 56), p. 2, 1899.

[19] Calvert, W. R., Big Snowy Mountains and vicinity (unpublished report in files of U. S. Geol. Survey).

[20] Calvert, W. R., Smithsonian Misc. Coll., vol. 64, p. 275, 1916.

[21] Weed, W. H., U. S. Geol. Survey Geol. Atlas, Little Belt Mountains folio (No. 56), p. 2, 1899.

[22] Hadding, Assar, The pre-Quaternary sedimentary rocks of Sweden: Lunds geol.-mineral. Inst. Meddel., No. 32, p. 148, 1927.

WOLSEY SHALE

The Wolsey shale consists of about 750 feet of greenish micaceous fissile shale, with some thin calcareous beds that contain glauconite. Calvert [23] obtained sufficient fossils from the beds to fix their age as Middle Cambrian. The correlation is based on their resemblance in character and in stratigraphic position above the basal Cambrian conglomerate to the Wolsey shale in the Little Belt Mountains.

FLATHEAD QUARTZITE

The character of the Flathead quartzite is well summed up in the following statement by Calvert [24]:

> Lying unconformably on shale of the Belt series is a sandstone 75 feet thick composed mainly of pure quartz. Although indurated the sandstone is not a quartzite, as stratification is distinct and cleavage along bedding planes is marked. It is evidently a shore deposit, as cross-bedding is abundant and ripple marks are beautifully developed. The shore phase is also attested by abundant worm trails, the only evidence noted of life existing at the time of deposition of the sandstone. Layers of quartz conglomerate are of frequent occurrence in the sandstone, with pebbles generally small but occasionally attaining a diameter of 1 inch. Because of the lithologic character and stratigraphic relations of the sandstone, it is correlated with the Flathead quartzite of the Little Belt Mountains section.

BELT SERIES

At the head of Swimming Woman Canyon the Algonkian Belt rocks are exposed in a small area. These rocks consist of highly indurated dark-gray limy shale. The maximum thickness exposed is about 300 feet. This shale lies unconformably beneath the Flathead quartzite, the discordance in dip being 10° to 20°. The shale shows some signs of mineralization along joints and has been prospected at a number of places.

IGNEOUS ROCKS

In the eastern part of the area mapped and outside of the mountain uplift two dikes and two small masses of volcanic breccia were noted. Many other similar igneous rocks were mapped by the writer in the adjacent area,[25] and there may be others in the area mapped that were not observed. The dikes are composed of a fine-grained dense brick-red to reddish-yellow rock, which is so highly altered that its original mineral composition is difficult to determine. The volcanic breccias occur in small irregular masses, forming buttes that are conspicuous features of the landscape. They consist largely of angular blocks and fragments of sandstone and shale which are embedded in a fine-grained tuff. Thin sections of this tuff were examined by C. S. Ross, who reports that it is a rhyolite in which many of the phenocrysts and a large part of the groundmass are altered to calcite. These breccias are evidently not remnants of a widespread surface deposit but appear to be the filling of vents or fissures through which material may have been ejected to the surface.

[23] Calvert, W. R., Smithsonian Misc. Coll., vol. 64, p. 275, 1916.

[24] Idem, p. 274.

[25] Reeves, Frank, Geology of the Cat Creek and Devils Basin oil fields and adjacent areas in Montana: U. S. Geol. Survey Bull. 786, pl. 3, 1927.

STRUCTURE

Structure of the Big Snowy Mountains.—The structure of the Big Snowy Mountains is shown in Plate 38 by cross sections and contours. The contours are based on the top of the Kootenai formation. In the eastern half of the area the error in the contours is probably less than 25 feet, except in the belt of highly tilted strata, where it may be as much as 250 feet. In the western half of the area, owing to the lack of key beds and exact data as to the intervals between outcropping beds and the Kootenai formation, the error in the contours may be 500 feet or more. In spite of these possible errors the contour map is believed to present a fair general picture of the structure of the mountain uplift and of the domes on its northeast flank. The following is a summary of the essential features of the structure of the mountains and adjacent foothills.

The major part of the uplift is an asymmetric anticline, the axis of which is slightly sinuous but has a general trend of N. 70° W. On the south flank of the anticline the dip is very steep, the Paleozoic rocks being inclined at 45° to 60° and the Cretaceous rocks at 60° to 90°. This tilted belt is about 2 miles wide, south of which the rocks are nearly flat-lying. The difference in altitude of any bed in this area of approximately flat-lying rocks and of the same bed in the crest of the anticline is about 13,000 feet. On the north flank of the anticline the formations are but gently inclined, the prevailing dip being 8° to 10° for 6 to 8 miles. Farther to the northeast there is a broad structural terrace on which are three elliptical domes. Along the northeastern margin of this terrace is a pronounced syncline, to the northeast of which the rocks again dip northeastward except in two localities, where they are bowed upward in two domes. The axis of the high anticline to which the relief of the mountains is due plunges sharply northwestward, so that a short distance beyond the area mapped the fold disappears. It also plunges steeply southeastward and practically dies out, unless the sharp fold in the Cretaceous strata, known as the Devils Pocket anticline, may be considered a continuation of it. The minor domes of the area will not be described separately, but a few points of interest may be mentioned. All except one of these domes—namely, the Button Butte Dome, which is circular in plan—are elliptical, with axes approximately parallel to the major anticline. The Spindle Top Dome is disrupted by two normal faults similar in type to the transverse faults associated with many of the domes of the Rocky Mountain oil fields. The other domes are not noticeably faulted, and no faults were observed in the main range except some of slight throw in the Algonkian shales. On the north flank of the mountains Calvert mapped a downfaulted block which is about 6 miles long and half a mile wide.

Regional structure.—The structure of the region in which the Big Snowy Mountains lie is shown on Plate 37. The range appears as a part of an uplifted block of rectangular form which includes the Judith Mountains and an area to the east, commonly spoken of as the Cat Creek-Devils Basin uplift. The Judith Mountains consist of an eroded cluster of laccolithic domes in which are exposed strata ranging in age from Middle Cambrian to Cretaceous, together with the underlying laccolithic masses. Closely associated with the laccoliths of the Judith Mountains are those of the Moccasin Mountains, lying about 10 miles to the west. Between the Judith and Big Snowy Mountains are several low circular domes that may be of laccolithic origin. The Cat Creek-Devils Basin uplift consists of a belt of eastward-dipping strata that have been uplifted 2,000 to 4,000 feet above the nearly flat-lying strata adjacent to them. The elevation of this area has taken place chiefly along its margins, which have a trend of N. 70° W. Along these margins the strata are slightly domed, and the domes on the northern margin have an en échelon arrangement, with their axes trending N. 35°–55° W. Most of these domes are broken by short transverse faults that trend N. 50°–60° E. As the trend of the marginal belt is N. 70° W. and that of the domes N. 35°–55° W., the average trend of the faults intersects the general trend of the series of domes at 55° and the axes of the individual domes at 80°. The downthrow of most of these faults is on the southeast side. Faults of similar character and trend occur in the main part of the uplift and parallel the dikes there. Along the south margin of the uplift there are two large domes—the Devils Basin and Big Wall Domes—which are not noticeably faulted. South of this uplift there is a broad, shallow basin known as the Bull Mountain syncline. Farther south there is a second zone of faults, the Lake Basin fault zone, which closely parallels the zone of faults along the northern margin of the Cat Creek-Devils Basin uplift. These faults, however, are not conspicuously associated with domes, although there is some evidence of folding in the belt. To the west of the Big Snowy Mountains are the Little Belt Mountains, one of the front ranges of the Rocky Mountain system. From the central plateau of these mountains, which is made up of igneous rocks and sedimentary strata of Cambrian and pre-Cambrian age, a spur of folded Paleozoic rocks extends southeastward. This plunging anticline ends in the plains about 10 miles southwest of the Big Snowy Mountains. South of the Big Snowy Mountains and bordering the belt of highly tilted strata on their south flank the Cretaceous strata lie in a shallow syncline and are so widely covered by terrace gravel that little can be determined as to the structure in detail. Farther southward and southeastward there are pronounced domes and anticlines. The domes are circular in outline and are so closely associated with igneous intrusives as to suggest the probability that they are of laccolithic origin. One of the anticlines, the Shawmut, trends northeast, but all the others trend S. 60° E. and are asymmetric, the steeper dips being on the side away from the mountains. South of these domes the strata dip steeply southwestward into the Crazy Mountain syncline, in which the Cretaceous and Tertiary formations thicken markedly in a southwesterly direction. Southwest of this syncline and about 90 miles southwest of the Big Snowy Mountains are the Beartooth Mountains which, according to Bevan,[26] consist of a broad asymmetric anticline that is strongly overturned toward the Great Plains and broken along its northeast limb by a huge overthrust fault having a trend of S. 60° E.

From the above data it is apparent that the region in which the Big Snowy Mountains lie possesses a variety of structural features. An attempt to explain the origin of the Big Snowy uplift must take these features into consideration. Although it is not the province of this paper to present a lengthy analysis of the tectonics of the region, some of the writer's conclusions will be briefly stated.

Genetically the structural features may be classified into three groups—(1) circular domes, (2) transversely faulted elliptical domes and belts of en échelon faults, and (3) elongated domes and anticlines. Those of the first group are probably due to intrusive activity. The origin of the features of the second group is not definitely known, but it may be that structure of this type is produced by differential lateral movements in the earth's crust, brought about by the drag effects arising from the transfer of material in the zone of flow to preserve isostatic equilibrium. The features of the third group, in which the Big Snowy uplift is included, are probably due to the horizontally acting forces that produce most mountain systems.

GEOLOGIC HISTORY

Because of the fact that only a few hundred feet of the pre-Cambrian rocks are exposed in the Big Snowy Mountains, little information is available as to the pre-Cambrian geologic history of the area; but the record furnished by the exposed pre-Cambrian rocks of the Little Belt Mountains indicates that pre-Cambrian time was a period of widespread sedimentation, deformation, igneous activity, and erosion. Near the end of pre-Cambrian time the region was uplifted and the pre-Cambrian sediments were sub-

[26] Bevan, Arthur, Summary of the geology of the Beartooth Mountains, Mont.: Jour. Geology, vol. 31, p. 457, 1923.

jected to deformation. Erosion of these rocks ensued, resulting in their base-leveling before the incursion of the Middle Cambrian sea, in which a considerable thickness of limy shale and conglomeratic limestone was deposited. Repeated oscillations of the sea took place in late Middle Cambrian time, finally resulting in the withdrawal of the sea from this region, which remained land during all or most of the remainder of pre-Carboniferous time. At the beginning of the Carboniferous a widespread sea invaded the entire region, followed by the accumulation of a great thickness of marine limestones. In late Mississippian time conditions again became unstable, and the region was alternately the site of a shallow sea and dry land. Sometime late in the Carboniferous period the sea again withdrew from the area. A slight regional tilting ensued, followed by base-leveling. With the advent of Jurassic time the sea once more covered the region, and throughout the remainder of Jurassic and Cretaceous time and a part of Tertiary time a great mass of marine and continental sediments accumulated. Just when the movements set in that produced the structure of the present mountains and the folding and faulting in the adjacent plains can not be determined definitely, but they undoubtedly were not initiated until after Eocene time, because the Fort Union formation, of Eocene age, is included in the tilted belt on the south flank of the Devils Basin anticline. The proximity and similarity of this anticline to the Big Snowy anticline strongly suggests that they were formed contemporaneously. Since late Tertiary time crustal movements have been very slight, and the only geologic changes that have taken place in the region are those brought about by erosion, which has produced the present relief of the mountains.

ECONOMIC GEOLOGY

Mineral deposits of commercial value are not known to occur in the Big Snowy Mountains. At present the only mineral deposits worked on a commercial scale in the area mapped are gypsum beds in the Ellis formation and coal in the Kootenai formation. These deposits are mined only on the north side of the mountains, and it is doubtful whether valuable deposits will be found elsewhere within the area.

Oil and gas.—Some of the geologic conditions favorable to the occurrence of oil and gas are present in the foothills adjacent to the Big Snowy Mountains. Several of the formations possess both the organic source material and porous beds for the accumulation of hydrocarbons in commercial pools. The formations that supply both of these factors are the Colorado shale, Ellis formation, and Quadrant formation. The Kootenai formation contains several porous sandstones but no noticeable source rocks. Yet it is in sufficiently close proximity to the black Colorado shale and the limy petroliferous shales of the Ellis formation to be regarded as a potential oil-bearing formation. This is true also for the Madison limestone, especially the top part, where there are porous beds in close contact with the overlying petroliferous Ellis shales. Although petroliferous limestones are present in the lower part of the Madison, it is doubtful if porous beds occur there. Whether both source material and porous beds are supplied by the older Paleozoic formations is doubtful. The conglomerates in the Meagher limestone and Flathead quartzite may possibly be sufficiently porous to serve as reservoir rocks, but none of the Cambrian rocks are conspicuously petroliferous.

Another factor necessary for the accumulation of oil—favorable structure—is present in the foothill belt. The domes and anticlines present to the east and northeast of the main range are similar to the domes and anticlines in which many of the Rocky Mountain oil fields are found.

Notwithstanding the presence of the favorable stratigraphic and structural conditions mentioned, there is slight probability that oil or gas will be encountered here in commercial volumes, for, although but nine wells have been drilled in the area (see table) and but four of these were deep tests, these four wells were drilled in localities where the structure was favorable for the occurrence of oil, yet no commercial pools of oil were encountered. A large flow of gas was struck in the top of the Madison limestone at a depth of 990 feet in the well drilled on Alaska Bench by the Portland Syndicate (No. 1 on map). This gas, however, proved to be noninflammable and has no commercial value. An analysis of the gas by the Bureau of Mines shows the following constituents:

	Per cent
Carbon dioxide	0. 81
Oxygen	15. 52
Methane	1. 19
Ethane	2. 52
Nitrogen	79. 96

From this analysis it is evident that the gas is composed mainly of nitrogen and oxygen, the chief constituents of the atmosphere, and that these elements are present in about the same relative proportion as they occur in the atmosphere. It is reasonable, therefore, to consider that the gas is primarily of atmospheric origin. The slight dissimilarity to the gaseous mixture forming the atmosphere may readily be ascribed to the addition of the hydrocarbon gases common in sedimentary rocks and the depletion of the oxygen by the oxidation of minerals and organic débris. The accumulation of this atmospheric gas in considerable volume under pressure at a depth of approximately 1,000 feet beneath the surface is probably due to the active circulation of underground water. There is a plainsward movement of surface water in the porous beds in the Madison limestone and other formations of the area. Water entering the outcrop of these porous beds in the Big Snowy Mountains, like all surface water, contains a small amount of air in solution. As the water moves down the dip some of the dissolved air is doubtless liberated, because in sinking to lower levels the temperature of the water is raised and its ability to carry gases in solution correspondingly decreased. This liberated air doubtless forms as minute bubbles in the water and may be segregated in the crests of domes or in other traps in the same manner that natural gas and oil are segregated.

This plainsward movement of water through porous beds in the formations cropping out around the mountain uplifts in the Rocky Mountain region probably accounts for the absence of oil in the domes and anticlines near the mountains. Such an active flow of water must tend to sweep the oil and gas farther toward the plains, where the water moves more slowly through the formations and the conditions for the accumulation of oil and gas and their retention in pools throughout long geologic periods are more favorable.

Wells drilled for oil in vicinity of Big Snowy Mountains

No. on map	Company	Location				Altitude of well (feet)	Total depth (feet)	Date of completion	Formations penetrated	Remarks
		Sec.	T. N.	R. E.	Dome or locality					
1	Portland Syndicate.	17	13	20	Alaska Bench Dome.	5,550±	2,155	1925	Quadrant and Madison.	1,000,000 cubic feet of noninflammable gas in Madison at 990 feet.
2	Big Bud Syndicate.	3	13	20	Tyler Creek	(?)	300	1921	Quadrant	Shallow abandoned hole.
3	Drees Petroleum Co.	13	13	21	Potter Creek	(?)	905	1922	do	Dry hole.
4	Oklamont Oil Co.	13	13	21	do	4,850	1,840	1922	Quadrant and Madison.	Do.
5	E. G. Lewis Development Co.	29	14	24	Button Butte Dome.	(?)	235	1920	Ellis and Quadrant.	Shallow abandoned hole.
6	Mid Northern Dome Co.	27	13	23	Spindle Top Dome.	4,310	1,743	1923	Ellis, Quadrant, and Madison.	Dry hole; water in Quadrant and in Madison.
7	McElwain Development Co.	34	12	22	Sawmill Creek	4,335	105	1927	Morrison(?) and Ellis.	Shallow abandoned hole.
8	Montana Pioneer Oil Co.	19	11	23	South Willow Creek.	3,960	850	1922	Colorado and Kootenai.	Dry hole.
9	do	29	11	23	do	3,920	2,200	1924	Colorado, Kootenai, Morrison(?), Ellis, and Quadrant.	Show of oil in Kootenai. Water in all formations.

Oil shale.—Low-grade oil shales occur in the middle part of the Quadrant formation, but observation by the writer did not indicate that any of these were of sufficient richness to be of prospective commercial value. Beds a few inches thick that would yield 20 to 30 gallons to the ton may be present, but no beds a foot or more in thickness that would yield over 10 gallons to the ton were noted.

Coal.—Coal occurs in this area in the Eagle sandstone, the Kootenai formation, and Quadrant formation. Only that in the Kootenai is of commercial value. It lies 60 to 90 feet above the base of the formation and underlies a coarse-grained cross-bedded sandstone which forms the conspicuous cliffs and hogbacks characteristic of Kootenai outcrops. This coal bed is mined at several localities near Lewistown and has been mined on a small scale in the northwestern part of the area mapped, along the north bank of McDonald Creek west of Forest Grove and in the northern part of T. 14 N., Rs. 18 and 19 E. East of Forest Grove the coal thins rapidly, and it was not noted at any place where the Kootenai crops out on the east side of the mountains. No trace of the coal was noted on the south side of the mountains except in sec. 7, T. 11 N., R. 22 E., where it has a thickness of 10 to 12 inches.

Thin beds of dirty coal were noted at a few localities in the lower part of the Eagle sandstone and in the upper part of the Quadrant formation, but nowhere in the area mapped are these of workable thickness.

Building materials.—Building stone, clay for the manufacture of brick and tile, and gypsum for the manufacture of cement and other products are abundant in the Big Snowy Mountains and their foothills, but at present only the gypsum is being utilized. It is doubtful whether extensive use will be made of the other building materials, for they can be obtained more conveniently, at a lower cost, at many localities in the region along the railroad.

Beds of gypsum occur in the Ellis formation on the north flank of the Big Snowy Mountains and in the Quadrant formation at several localities around the mountains. The gypsum in the Ellis formation attains greatest thickness in T. 14 N., Rs. 18, 19, and 20 E. The United States Gypsum Products Co. operates a plant near Heath, in sec. 1, T. 14 N., R. 19 E., for mining gypsum and manufacturing wall plaster and other gypsum products. The gypsum here occurs in beds 5 to 10 feet in thickness, interbedded with limestone, limy shale, and variegated clay shale.

Metalliferous deposits.—The following statements in regard to the metalliferous deposits are based largely on data obtained for the Geological Survey by W. R. Calvert during a brief field examination in 1911. Since then little prospecting has been done. When the writer was in the area in 1927 the pits and shafts had either fallen in or were full of water.

Most of the mineral prospecting has been confined to the Algonkian shales and Cambrian quartzites along Swimming Woman Creek, but the Ellis and Quadrant formations have also been prospected in a few places at the east end of the mountains. The prospecting in the Algonkian shales and Flathead quartzites has been confined largely to brecciated zones along minor faults. These zones in the Algonkian rocks consist largely of angular fragments of indurated shale, the interstices of which are filled largely with calcite carrying very slight traces of sulphide. The openings in the Flathead quartzite, however, show a few small pockets of sulphide rich in copper and iron. The associated minerals are chalcopyrite, pyrite, galena, malachite, and azurite. Vein quartz is not present. There is no evidence of enrichment or of metamorphic action.

Although there have been consistent reports that ores found in the area have assayed from $20 to $60 to the ton, investigations of these reports by reliable mining companies have proved their falsity and confirmed the opinion held by mining engineers and geologists that further prospecting in the region is unwarranted.

THE KAOLIN MINERALS

By Clarence S. Ross and Paul F. Kerr

ABSTRACT

The clay minerals have been under investigation for several years in the laboratories of the United States Geological Survey, the National Museum, and Columbia University. The work on the minerals of kaolin has been completed, and the results are presented in this paper.

An investigation of such materials as clays presents obvious difficulties, and it has been necessary to correlate the available data by making all tests—chemical, dehydration, optical, and X ray—on identical material. The older methods have been effectively supplemented by the newer one for the study of crystalline powders by X-ray diffraction patterns. By these methods the mineralogy of clays can be completely studied.

The present work has shown that instead of one kaolin mineral, as has commonly been assumed, there are at least three distinct minerals. A large number of names have been proposed in the past, but a review of the literature indicates that only two are acceptable. Kaolinite, which has heretofore included all three minerals, is to be restricted to the most abundant one, which characterizes all or nearly all kaolin deposits; and nacrite, an old name applied to a mineral from Brand, Saxony, is to be revived for the mineral whose type came from that locality. A third mineral, originally described from the island of Anglesey, seems never to have received a distinctive name, and it is proposed to term it dickite, after Allan B. Dick, who carefully and fully described the original material.

The optical, X-ray, and chemical properties have been determined for material from many localities, and dehydration tests have been made on representative samples. The data presented in detail in this paper show that the chemical composition of the three minerals is very similar. Most of the kaolinite and all the dickite and nacrite investigated have the commonly accepted composition $2H_2O.Al_2O_3.2SiO_2$; some kaolinite varies, however, in the silica:alumina ratio, its high-silica representative, anauxite, being possibly $2H_2O.Al_2O_3.3SiO_2$.

The optical properties of kaolinite and dickite are so distinct that these two minerals can be readily distinguished where crystal grains are available for study. Nacrite is close to kaolinite in its optical properties, although there are significant differences.

The X-ray diffraction patterns are quite distinct for the three minerals and have proved indispensable in differentiating species and establishing criteria for their recognition. The dehydration curves also show characteristic differences for the three minerals. As the three minerals of kaolin thus show distinctive features in respect to three different sets of properties they can be regarded as well-established independent minerals.

The existence of three distinct kaolin minerals, whereas only one has been commonly assumed heretofore, will probably explain most of the differences of opinion about the formation of kaolin. The three minerals are stable at different temperatures and so may have been formed under dissimilar conditions. Most if not all commercial kaolin deposits that are characterized by kaolinite, the mineral stable at the lowest temperature, are formed by the weathering of feldspathic rocks; some kaolinite is also formed by the action of sulphate-bearing or thermal carbonate-bearing waters. It seems evident, however, that kaolinite is never normally formed by waters with a very high temperature.

It has been stated in the literature that a mineral formed as an intermediate product in the weathering of feldspar to kaolin is muscovite, but the observations of the writers show that this mineral is not muscovite but an unidentified mineral apparently differing from kaolinite only in having a lower water content.

Dickite, which is stable at a distinctly higher temperature, seems to have been formed by hydrothermal solutions (probably only moderately heated, however) in some deposits, but in others the action of hot solutions is excluded by the geologic evidence. Dickite seems therefore to be a hypogene mineral formed in part by hot and in part by cold solutions, and there is no known evidence to indicate that it results from purely weathering processes.

Nacrite, the mineral stable at the highest temperatures, has apparently been formed by hypogene processes in both occurrences thus far definitely known.

INTRODUCTION

The identity and properties of the minerals of clays have long interested mineralogists, persons engaged in the industrial use of clay materials, and students of soils. Although many workers have devoted time to the study of these minerals, the conclusions reached have varied widely, and no satisfactory classification has been presented. There has been a general lack of reliable criteria for the recognition of species, as is well shown by the conflicting results of different investigators. A large number of the names that have been proposed for the clay minerals have not been generally accepted, and others are in use for which no firm foundation exists. Furthermore, the inadequacy of early methods of study make it practically impossible to correlate most samples of clays with the accounts of their properties given in the literature. Descriptions of clay minerals written before 1900 or even later are with a few notable exceptions of little value judged by present-day standards, and many of the early descriptions are worthless or even misleading. This situation makes obvious the need for a resurvey of the clay minerals by modern methods.

It is only within the last 10 years that methods have been perfected which permit the detailed study of minute crystal aggregates, making it possible to obtain definite information from the optical examination of even the finely divided clay minerals. Of equal importance has been the development of the X-ray powder method for studying fine-grained crystalline materials in random orientations. Coordinated studies by optical, X-ray, chemical, and thermal methods now yield criteria for the identification of the clay minerals which are far superior to those based solely on chemical methods, as was the practice when these minerals were first described. Such criteria materially alter the conclusions reached by the former methods of study, necessitating the reinterpretation of the minerals of kaolin in the light of recent developments, and it is upon an investigation of this type that the present report is based.

CONDITIONS OF WORK AND ACKNOWLEDGMENTS

The present investigation of the clay minerals was begun in the laboratories of the United States Geological Survey in 1926. Clarence S. Ross has had general charge of the work for the Geological Survey and has examined and selected suitable materials and conducted the optical studies. Paul F. Kerr, of Columbia University, has made the X-ray studies and has cooperated in writing this report. The chemical analyses have been made largely in the laboratories of the United States National Museum, although part of the analyses and all of the dehydration tests were made in the chemical laboratory of the Geological Survey. The authorities of Columbia University have made liberal allowances from research funds to provide for the construction of equipment used in the X-ray work. The collections and other facilities of the university have also been freely employed and have been of constant assistance throughout the progress of the work. For the last two years the work has been aided by the National Research Council, which has appointed a committee on the study of the clay minerals and has contributed funds to carry on the work of chemical analysis, which has been done by Forrest A. Gonyer at the National Museum. Dr. Waldemar Lindgren has long been interested in the clay minerals associated with mineral deposits, and while chairman of the division of geology and geography of the National Research Council he aided and encouraged the present study. His interest and the funds procured through his efforts have contributed very materially toward the success of the investigation. For the last two years the work has been largely confined to the minerals of kaolin and has now reached a state where several definite conclusions as to this series of minerals can be drawn.

Any merit the present report may possess is due in a large measure to the members of the committee on clay minerals, who have so freely contributed their time and thought to the progress of the work. Thus, we have been materially aided through the cooperation of the late Dr. George P. Merrill, in freely placing at our disposal the chemical-laboratory facilities of the National Museum. Dr. Edgar T. Wherry, of the Bureau of Chemistry, has been of much assistance through his unflagging interest and constructive criticism of the manuscript. Mr. Earl V. Shannon, of the National Museum, has furnished help on the chemical side of the work. Prof. Heinrich Ries has offered many helpful suggestions and given material aid in bringing the work to a conclusion. Mr. Waldemar T. Schaller, of the Geological Survey, has contributed valued advice throughout the investigation and particularly in the consideration of the nomenclature adopted.

METHODS OF STUDY

In previous studies of the kaolin minerals the purity of the material analyzed has very commonly been open to question, and in only a few studies have complete optical determinations been made on satisfactorily analyzed material. A little X-ray work has been done, but it has not been comprehensive and has rarely if ever been properly coordinated with chemical composition and fully determined optical properties. From time to time mineralogic deductions have been based on old and faulty analyses, many of which were made on obviously impure material; on determinations of optical or X-ray properties of unanalyzed material; and on the uncritical acceptance of the labeling of museum specimens. Attempts to make short cuts to fundamental scientific conclusions through the use of uncorrelated observations of varying and doubtful reliability—the mere mulling over of old data—can never command the confidence of careful investigators. For these reasons it seemed necessary, in this study of the kaolin minerals, to make chemical analyses, optical determinations, X-ray examinations, and dehydration tests on portions of single samples whose purity had been subjected to the most rigid tests.

Most clay materials, including kaolin, commonly occur in nature as very impure mixtures that contain quartz, feldspar, micas, and in many specimens a large variety of other minerals. For this reason the first step in the study of these materials is a careful microscopic examination to determine the degree of purity and the nature of the accessory minerals. Entirely pure samples are very rare, and consequently some method of purification must usually be devised. Samples that contain large amounts of very fine grained impurities, especially those having nearly the same specific gravity as the clay, can not be successfully purified and must be rejected at the start. This necessitates the examination of a very large series of samples, before even a small number suitable for further study can be selected.

The simplest method of purification is to pick out by hand single clean crystals, one by one, under the binocular microscope. This method has been applied to a number of samples that contained crystals of unusual size, although it has required a large amount of exceedingly tedious work. It yields material, however, whose purity is undoubted. The finer-grained samples require some other method or combination of methods for their purification. Fortunately, the individual crystals of the kaolin minerals do not break down in water, and they can be washed vigorously to free them from still more finely divided associated material. The washed residue can then be run through heavy solutions to separate any remaining accessories. The kaolin minerals are as a rule only slightly lighter than the associated quartz and feldspar, making difficult the purification of fine-grained samples. For this reason heavy-solution treatments, depending on the action of gravity alone, have not been sufficient to bring about a separation, and it has usually been necessary to use in addition a powerful centrifuge. Great care to secure the correct specific gravity of the heavy liquid and repeated treatments in the centrifuge have been required, and even with all precautions it has been impossible to purify completely many promising samples. The examination of specimens and the purification of material have thus consumed a very large part of the time devoted to the study. At all stages the purity of the samples was tested by petrographic methods.

The optical properties were first determined on the relatively large crystals found in rare samples of kaolin; then, as the diagnostic characteristics became known and as experience was gained, it became possible to carry optical studies over into finer and finer grained materials. The indices of refraction and other optical properties have been determined by the standard immersion method; for much of the material, however, an oil-immersion objective has been required.

The kaolin minerals are all hydrated silicates of aluminum, and the ratios of the component oxides as determined by analysis do not differ greatly. Accordingly X-ray diffraction patterns must contain relatively large numbers of lines of reasonably sharp definition before it is possible to distinguish clearly the various species so closely related in chemical composition. Published X-ray data on the kaolins are inadequate for this purpose. The technique of X-ray diffraction photography as applied to well-crystallized substances allows the bringing out of measurable diffraction lines representing interplanar spacings well below 0.7 Ångström unit; but the physical condition of the kaolins, made up as they commonly are of extremely thin and warped flakes, limits the lines obtainable from them to those of broader spacing. Lines developed in the kaolin patterns are sufficient, nevertheless, to distinguish the minerals from one another, and of course from other clay minerals, through certain striking and consistent dissimilarities in the range between 1.00 and 1.70 Ångström units. This subject will be further discussed after the data for the individual minerals have been presented. In the course of this work it soon became evident that by working with pure materials standard diffraction patterns showing distinct differences between the various kaolin minerals could be obtained. With such standard patterns available, the identity of unknown minerals may be established by simple comparison.

The interplanar spacings in Ångström units have been computed according to the well-known formula for X-ray diffraction, $n = 2d \sin \theta$. The figures given by Duane,[1] $K\alpha_1$ 0.7078 Å. u. and $K\alpha_2$ 0.71212 Å. u. for the $K\alpha$ doublet of molybdenum, were substituted for wave length (λ) in the formula. In the pursuance of the X-ray studies, photographs were obtained in an apparatus of special design [2] adapted after the methods of Hull [3] and Debye-Scherrer.[4]

The film holders were modified by the construction of an inclined slit system set at 45° to the path of the incident rays, and the sample was placed before the final slit. This modification in the slit system, it is believed, produces diffraction lines with a sharper resolution—an important point in the study of finely divided material, such as clay. All samples were carefully checked with sodium chloride to confirm the accuracy of the computation of interplanar atomic spacing. The $MoK\beta$ and $MoK\gamma$ lines were filtered from the spectrum of molybdenum with zirconia. In view of the fact that the samples were placed in front of the final slit, measurements have been made on the outer edges of the diffraction lines, an improvement over the more frequently used line-center measurement. Many of the diffraction lines are resolved into doublets each consisting of a sharp heavy line due to the wave length $K\alpha_1$ and a less intense line due to the wave length $K\alpha_2$.

The chemical analyses were made by the standard methods and need no discussion.

The dehydration tests on the kaolins were carried out in an electric furnace, the temperature being determined with a carefully calibrated thermocouple. The sample was maintained at constant temperature until repeated weighings showed that there was no further loss of weight—that is, until the sample had reached equilibrium at that temperature. The results of the tests have been plotted as a dehydration curve with loss of water in per cent as the vertical coordinate and temperature as the horizontal coordinate.

The study of kaolins has demonstrated that it is possible to differentiate between mineral species by coordinated methods of mineralogic research, not only with materials that are very fine grained, but even with those that are seemingly amorphous. In some

[1] Duane, W., Data relating to X-ray spectra, vol. 1, No. 6, Nat. Research Council, 1920.

[2] Kerr, P. F., The determination of the opaque ore minerals by X-ray diffraction patterns: Econ. Geology, vol. 19, pp. 1–35, 1924.

[3] Hull, A. W., A new method of X-ray crystal analysis: Phys. Rev., vol. 10, pp. 661–696, 1917.

[4] Debye, P., and Scherrer, P., Interferenzen an regelos orientierten Teilchen im Röntgenlicht: Physikal. Zeitschr., vol. 17, pp. 277–282, 1916.

samples optical methods alone will identify a kaolin mineral; in many, on the other hand, optical and X-ray methods have been required, and in still others a combination of optical, X-ray, and chemical methods. It has become evident that all these methods should invariably be used in testing standards and determining criteria for future reference. As the clay minerals are probably more difficult to study than almost any other series, it seems not amiss to conclude that practically all crystalline mineral materials are amenable to the present-day methods of mineralogic research used in this series of investigations.

NOMENCLATURE

A number of names have been applied to the kaolin minerals, and diverse and conflicting opinions have been expressed about their identity and relationships, but in general they have been assigned to the single mineral species kaolinite and are included in Dana's "System of mineralogy" under that name. From time to time a new name has been proposed, merely because one worker's analysis of more or less indefinite material did not agree with that made by somebody else; and unfortunately mineralogy of this kind has not yet ceased to exist. As recently as January, 1929, the new name takizolite was proposed for a kaolin mineral on the basis of an analysis that gave an alumina-silica ratio of 1:3.5, without any optical or X-ray diffraction data as to the nature of the material analyzed being furnished.[5]

The present study has shown that the kaolin minerals can not be assigned to a single species, and evidence is presented in this paper which shows that three distinct mineral species are represented. In much of the literature on kaolin minerals so few data of diagnostic significance have been given that it is not possible to determine from the descriptions alone to what mineral species the names used apply, and furthermore the old and poorly described specimens are usually not available for restudy. For this reason no attempt has been made to discuss here all the names that have been proposed. It is important, however, to consider the names that have had wide usage and to ascertain which are applicable to the three definite mineral species that have been established by modern methods. For preliminary discussion these species will be called the "mineral of kaolin," the "Dick mineral," and "nacrite." The "mineral of kaolin" is considered to be the essential constituent of the typical kaolin of the ceramic industry. The "Dick mineral" was first described by Tookey and Dick[6] from the island of Anglesey, along the northwest coast of Wales. "Nacrite" (German nakrit) was discovered around 1800 at Brand, Saxony. Halloysite is a submicroscopically crystalline clay material that is closely related to kaolinite which will be discussed in a later paper.

In this paper the terms kaolin and kaolinite will be used repeatedly, in quite definite senses, which should be defined at this point. By kaolin is understood the rock mass which is composed essentially of a clay material that is low in iron and usually white or nearly white in color. The kaolin-forming clays are hydrous aluminum silicates of approximately the composition $2H_2O.Al_2O_3.2SiO_2$, and it is believed that other bases if present represent impurities or adsorbed materials. Kaolinite is the mineral that characterizes most kaolins, but it and the other kaolin minerals may also occur to a greater or lesser extent in clays and other rocks that are too heterogeneous to be called kaolin.

The kaolin minerals herein described are not the dominant constituents in most clays, shales, and soils, as has often been assumed. The literature is full of references to kaolinite as the typical clay mineral, although that this is an error has been pointed out by those who have given clays the closest study. Merrill[7] says:

> That kaolin is the basis of all clays is the commonly accepted opinion of most writers. The evidence upon which such an opinion is based is largely chemical, and it must be confessed unsatisfactory—so unsatisfactory, in fact, that the present writer has ventured at times to doubt the accuracy of the statement altogether.

Ries[8] says:

> All clays appear to contain a variable amount of some hydrated silicate of alumina, * * * and even though the statement is frequently made that this silicate is the mineral kaolinite, the fact is at times difficult of proof; indeed, the evidence is clearly against it in some cases. This hydrated aluminum silicate was formerly sometimes referred to as the clay substance, or clay base, on the supposition that kaolinite was the basis of all clays. It has been recognized for some years, however, that this view is incorrect, and that a clay may have very little kaolinite.

Even a casual microscopic inspection of most clays, soils, and shales shows that their clay material can not be one of the kaolin minerals, all of which have a very low birefringence, but is most commonly one of the beidellite-montmorillonite type of clay minerals, which have a moderately high birefringence. The kaolin minerals are easily differentiated from the more widespread clay minerals of the beidellite-montmorillonite type[9] by having distinctive chemical, optical, and X-ray properties. The most obvious differences are a usually lower silica and water content, a weaker birefringence, and characteristic X-ray diffraction patterns for the kaolin minerals.

[5] Iimori, Satoyasu, and Yoshimura, Jun, A pink kaolin, and ruthenium as a minor constituent of the Tanokami kaolins: Chem. Soc. Japan Bull., vol. 4, pp. 1-15, 1929.

[6] Dick, A. B., On kaolinite: Mineralog. Mag., vol. 8, pp. 15-17, 1888.

[7] Merrill, G. P., What constitutes a clay: Am. Geologist, vol. 30, p. 318, 1902.

[8] Ries, Heinrich, Clays, p. 5, 1927.

[9] Ross, C. S., and Shannon, E. V., The minerals of bentonite and related clays: Am. Ceramic Soc. Jour., vol. 9, pp. 77-96, 1926; The chemical composition and optical properties of beidellite: Washington Acad. Sci. Jour., vol. 15, pp. 467-468, 1925.

HISTORY

Nacrite.—Brongniart [10] proposed the name "nacrite" in 1807 but did not adequately present the properties of the mineral. In 1832 Breithaupt [11] described material from an ore vein near Freiberg that has since been called "nacrite" ("nakrit"). Des Cloizeaux,[12] working on material from Saxony (probably the same locality), gave "nacrite" as a synonym for "pholerite" and described the mineral as occurring in hexagonal plates. He says:

These plates are composed of six triangular sectors whose boundaries, though quite vague, nevertheless give indication of composition parallel to the faces of a right rhombic prism approximating an angle of 120° and 60°. * * * In convergent light there is given in each sector the hyperbolas which indicate two diverging optical axes whose plane is normal to the side situated upon the hexagonal contour and is consequently parallel to the principal diagonal of the base of the fundamental prism. The bisectrix is negative and evidently normal to the plane of cleavage. The dispersion of the axes is weak at 45° from the plane of polarization, as is shown by the symmetrical distribution of the colors about the two hyperbolas, and the separation of the axes is greater for the red than for the violet.

Dick[13] says of nacrite from the Einigkeit mine, near Freiberg, Saxony:

These consist of fan-shaped aggregates of pearly scales, which have the same chemical composition, hardness, specific gravity, and refractive index as the Anglesey kaolinite [that is, what is here termed the "Dick mineral"] but differ from this in being optically negative, with the acute bisectrix practically normal to the cleavage and $2E=103°$ approximately (hence $2V=60°$). The cleavage flakes usually show a division into three sectors; in the central one the emergence of the negative bisectrix is normal, whilst in the adjacent sector it is inclined slightly away from the center on either side. Through the basal cleavage of the Anglesey crystals only a diffuse negative bisectrix with the axes far outside the field is to be observed. * * * It must therefore be concluded that there are two varieties of kaolinite, one optically positive and the other optically negative.

Thus Des Cloizeaux and Dick give data that definitely differentiate nacrite from the "Dick mineral," although not so adequately from the "mineral of kaolin." Dana [14] and Doelter [15] give "nacrite" as a synonym of "kaolinite," but Mellor [16] accepts the evidence of Dick and concludes that nacrite is a distinct mineral.

The Dick mineral.—The "Dick mineral" [17] from the island of Anglesey, so far as can be determined, seems never to have received a distinctive name but has always been confused heretofore with the "mineral of kaolin." To be sure, some of the names proposed by the early mineralogists may have applied to this mineral, but if this is so it can not be determined from their inadequate descriptions. The earliest reference found that seems clearly to apply to the "Dick mineral" is a description by Knop [18] in 1859 of material from Schneckenstein. This is treated under the name "kaolin" and is distinguished from pholerite, which is described in another part of the same paper.[19] He says in speaking of the crystals of "kaolin":

They have the average length of about 0.021 and a breadth of 0.015 mm. and show the form in part of very sharply developed rhombic plates, and in part the corners which are connected by the macrodiagonal are sharply truncated in various degrees. Here and there the crystal plates exhibit rhombic prismatic aggregates and verify the assumption of the rhombic crystal systems.

The crystal forms recognized are: ∞P (110), OP (001), and $\infty \breve{P} \infty$ (010), which correspond with the principal crystal faces as determined by Dick (see p. 159), except that they are referred to the orthorhombic instead of the monoclinic system.

Kaolinite.—Johnson and Blake [20] published a paper on "kaolinite" in 1867 and probably included the "Dick mineral" under that name, although their work was in large measure based on the "mineral of kaolin" and "nacrite." In 1884 Hills [21] gave a brief description of this mineral from the National Belle mine, on Red Mountain, near Silverton, Colo. Cross and Hillebrand [22] published analyses of material from the same locality under the name "kaolinite," and in 1887 Reusch [23] briefly discussed the same material.

Johnson and Blake clearly intended the name "kaolinite" to apply to the "mineral of kaolin," and their description fits both nacrite and the "mineral of kaolin" but not the "Dick mineral." They gave, it is true, the drawing of a crystal that resembles the "Dick mineral," but the distinctive properties of this mineral were plainly not included in the data resulting from their rather careful study.

The "mineral of kaolin" has apparently never been fully characterized, although several names have been applied to what is probably this mineral. A few of these will be considered.

Pholerite.—The name "pholerite" was proposed by Guillemin [24] in 1825. He gave no adequate description but reported it from Rive-de-Gier, France, and

[10] Brongniart, A., Traité élémentaire de minéralogie, vol. 1, p. 506, 1807.

[11] Breithaupt, A., Vollständige Charakteristik des Mineral-Systems, pp. 94, 318, 1832; Berg- u. huettenm. Zeitung, vol. 24, p. 336, 1865.

[12] Des Cloizeaux, A., Supplement to Manuel de minéralogie, vol. 1, pp. 548-549, 1862.

[13] Dick, A. B., Supplementary note on the mineral kaolinite: Mineralog. Mag., vol. 15, p. 127, 1908.

[14] Dana, E. S., System of mineralogy, 6th ed., p. 685, 1892.

[15] Doelter, Carl, Handbuch der Mineralchemie, p. 125, 1917.

[16] Mellor, J. W., Do fire clays contain halloysite or clayite?: Ceramic Soc. [England] Trans., vol. 16, p. 83, 1916.

[17] Dick, A. B., On kaolinite: Mineralog. Mag., vol. 8, pp. 15-17, 1888; Supplementary note on the mineral kaolinite: Idem, vol. 15, pp. 124-127, 1908.

[18] Knop, A., Beiträge zur Kenntniss der Steinkohlen-formation und des Rothliegenden im Erzgebirgischen Bassin: Neues Jahrb., 1859, pp. 593-595.

[19] Idem, pp. 544-546.

[20] Johnson, S. W., and Blake, J. M., On kaolinite and pholerite: Am. Jour. Sci., 2d ser., vol. 43, pp. 351-361, 1867.

[21] Hills, R. C., Kaolinite from Red Mountain, Colo.: Am. Jour. Sci., 3d ser., vol. 27, p. 473, 1884.

[22] Hillebrand, W. F., Contributions to the mineralogy of the Rocky Mountains: U. S. Geol. Survey Bull. 20, p. 98, 1885.

[23] Reusch, Hans, Krystallisierter Kaolin von Denver, Colo.: Neues Jahrb., 1887, Band 2, pp. 70-72.

[24] Guillemin, A., Sur la pholérite: Annales des mines, vol. 11, p. 489, 1825.

the same material is discussed by Des Cloizeaux,[25] who states that it shows a negative bisectrix normal to the thin plates. This clearly applied to the "typical mineral of kaolin," for although Des Cloizeaux did not distinguish between nacrite and pholerite, the negative bisectrix normal to the plates excludes the "Dick mineral."

Knop [26] mentions pholerite under "Pelitische Felsittuffe" but does not give the optical properties. Some of the recorded analyses of pholerite are higher in alumina than those for the "mineral of kaolin" and for nacrite; indeed, the indicated composition from the available analyses is close to the formula $4H_2O.2Al_2O_3.3SiO_2$. This may indicate that pholerite is a member of the apparent isomorphous series discussed on page 164, which is lower in silica than $\frac{Al_2O_3}{SiO_2}=\frac{1}{2}$, but it is also possible that these old analyses were made on impure material. The mode of occurrence and the descriptions, so far as they can be interpreted, indicate that in general pholerite should be classed with the "mineral of kaolin"; less commonly the name has been applied to nacrite and to the "Dick mineral." (See p. 157.)

Anauxite.—The name anauxite was proposed for a mineral from Bilin, Czechoslovakia, by Breithaupt [27] in 1838. Smirnoff [28] has given a detailed description of it and shows photomicrographs of characteristic vermicular crystals that resemble those from the "mineral of kaolin." The name "anauxite" has also been used for the mineral from Bilin by Hauer [29] and has been retained for the mineral from that locality by Dittler and Hibsch [30] and by Ross and Foshag.[31] The name has been used by Glinka [32] for a clay from Tschawka, Russia, and Mellor [33] has expressed the opinion that anauxite is a mixture of colloidal gels of silica and clay material. Allen [34] has also described anauxite from California.

Leverrierite.—The name "leverrierite" was proposed by Termier [35] for a mineral which is abundant in the coal measures of France. Larsen and Wherry [36] described a mineral from Beidell, Colo., under the name "leverrierite" but later changed the name to "beidellite" [37] because further study showed that the mineral of Termier was distinct and not related to the beidellite-montmorillonite group of clay minerals.

Dana [38] gives leverrierite as a variety of kaolinite. Cayeux [39] considers it to be identical with kaolinite and describes its wide occurrence in the sedimentary rocks of France and the relations that lead him to believe that it has been formed in place and is not a detrital mineral.

Newtonite.—A mineral from Sneeds Creek, Newton County, Ark., was described as "newtonite" by Brackett and Williams [40] in 1891, but a restudy by Foshag [41] of apparently similar material collected at the type locality by Wherry in 1925 showed that the chemical analysis and optical investigations had been made on two entirely different lots of material—the optics on alunite and the analysis on halloysite or kaolinite.

From this brief historical summary the complexity of the problem of nomenclature is evident. Thus "pholerite" was proposed in 1825 and seems to have been applied most generally to the "mineral of kaolin" but was extended to include the material described under the older name "nacrite" [42] and was at times used for the "Dick mineral." "Kaolinite" later came to be generally accepted for the "mineral of kaolin," but it, too, included nacrite and probably the "Dick mineral." "Leverrierite" was also proposed as a name for what proves to be the "mineral of kaolin." In fact, it has been less troublesome in many instances to determine the chemical composition and optical properties of one of these minerals than to ascertain the proper name by correlation with previous inadequate descriptions.

NAMES ADOPTED

The foregoing review of the important names that have been applied to the kaolin minerals leads to the following conclusions:

Nacrite.—The name "nacrite" has been consistently applied to the mineral whose type came from Brand, near Freiberg, Saxony, and its distinctness has been recognized. Therefore there is every reason to accept and preserve the term "nacrite" and to raise the mineral thus designated to the rank of a distinct species.

Kaolinite.—The name "kaolinite" was clearly intended by Johnson and Blake to apply to the characteristic mineral of kaolin, and their description does not take account of the distinctive properties of the "Dick mineral." In common usage to-day, as well as

[25] Des Cloizeaux, A., Manuel de minéralogie, vol. 1, p. 190, 1862.
[26] Knop, A., Pelitische Felsittuffe: Neues Jahrb., 1859, pp. 544–548.
[27] Breithaupt, A., Jour. prakt. Chemie, vol. 15, p. 325, 1838.
[28] Smirnoff, W. P., Über ein krystallinisches Verwitterungsproduct des Augits: Zeitschr. Kryst. Min., vol. 43, pp. 338–346, 1907.
[29] Hauer, K. R., Zusammensetzung einiger Mineralien mit Rücksicht auf ihren Wassergehalt: K.-k. geol. Reichsanstalt Jahrb., vol. 5, p. 83, 1854.
[30] Dittler, E., and Hibsch, J. E., Über Anauxit und Cimolit von Bilin: Min. pet. Mitt., vol. 36, p. 85, 1923.
[31] Ross, C. S., and Foshag, W. F., Anauxite, a mineral species, based on material from Czechoslovakia: Am. Mineralogist, vol. 13, pp. 153–155, 1928.
[32] Glinka, K., Soc. imp. nat. St.-Pétersbourg Trans., vol. 34, pt. 5, p. 71, 1905.
[33] Mellor, J. W., Ceramic Soc. [England] Trans., vol. 11, p. 94, 1911.
[34] Allen, Victor, Anauxite from the Ione formation of California: Am. Mineralogist, vol. 13, pp. 145–152, 1928.
[35] Termier, P., Sur une phyllite nouvelle, la leverriérite, etc.: Compt. Rend., vol. 108, p. 1071, 1889; Étude sur la leverriérite: Annales des mines, vol. 17, pp. 372–398, 1890.
[36] Larsen, E. S., and Wherry, E. T., Leverrierite from Colorado: Washington Acad. Sci. Jour., vol. 7, pp. 208–217, 1917.
[37] Larsen, E. S., and Wherry, E. T., Beidellite, a new mineral name: Idem, vol. 15, pp. 465–467, 1925.
[38] Dana, E. S., System of mineralogy, p. 687, 1909.
[39] Cayeux, Lucien, Introduction à l'étude pétrographique des roches sédimentaires, pp. 230–232, pl. 9, fig. 5, 1916.
[40] Brackett, R. N., and Williams, J. F., Newtonite and rectorite, two new minerals of the kaolin group: Am. Jour. Sci., 3d ser., vol. 42, p. 11, 1891.
[41] Foshag, W. F., The identity of newtonite with alunite: Am. Mineralogist, vol. 11, pp. 33–35, 1926.
[42] Des Cloizeaux, A., Supplement to Manuel de minéralogie, vol. 1, p. 500, 1862.

when proposed, "kaolinite" is taken to mean the clay mineral of kaolin. The chief difficulty in this connection is that the optical properties of the "Dick mineral" have been the only ones that were completely described, and as the two minerals were supposed to be identical, the optical properties of the "Dick mineral" have commonly been accepted as applying to kaolinite. Thus it has come about that "kaolinite" has signified the mineral of most kaolin deposits, yet the optical properties most commonly ascribed to it have been those of the "Dick mineral," which is rare or absent in those deposits. The question accordingly arises, Which of these two minerals, the "Dick mineral" or the "mineral of kaolin," should be designated "kaolinite"? If the name "kaolinite" is applied to the "Dick mineral" the optical properties that have come to be associated with kaolinite would continue to belong to that mineral. This would mean, however, that kaolinite rarely if anywhere occurs in kaolin, and most of the references to the occurrence, origin, and properties of "kaolinite" would be incorrect. It seems much less confusing to follow the intention of Johnson and Blake, who proposed the name, and to restrict "kaolinite" to the "mineral of kaolin." This would coincide with current thought about occurrence, though not with the accepted optical properties of kaolinite. The "Dick mineral" would then require a new name.

It must be admitted that other names have been applied to the "mineral of kaolin," but of these only "pholerite" seems worthy of further consideration. "Pholerite" has priority but has never been as generally accepted as "kaolinite." The optical properties of the "mineral of kaolin" have never been adequately recorded under either name, but they were more completely described by Johnson and Blake in proposing the name "kaolinite" than they ever were under "pholerite." Most of the analyses of pholerite differ from the formula that is known to represent the pure "mineral of kaolin" (although it is possible that the samples analyzed were impure and contained bauxitic material or allophane). The propriety of using the name "pholerite" is therefore highly questionable, inasmuch as the properties of the "mineral of kaolin" were never defined under that name, the chemical analyses in general do not agree, and the name was apparently never generally accepted and has not been used by mineralogists since the proposal of the name "kaolinite" 63 years ago.[43] If "kaolinite" is adopted for the "mineral of kaolin" it will be necessary to restrict the meaning of the term, for as used by many recent authors it has included nacrite as well as the "Dick mineral." On the whole, such restriction seems preferable to proposing a new name or resurrecting the name "pholerite," which is of doubtful applicability.

It seems proper to retain the name "anauxite" for the clay mineral with an alumina-silica ratio approaching 1:3. Evidence presented on pages 164–165 indicates that this mineral and the one just discussed are members of an apparently isomorphous series, which may accordingly be termed the "kaolinite-anauxite series." Anauxite appears to have essentially the same crystal structure as the "mineral of kaolin," and the difference, so far as known, can be determined by chemical analysis alone. Its crystallography and optical properties are identical with those of the "mineral of kaolin," and in the present study it is grouped with this mineral.

Dickite.—No distinctive mineral name has apparently ever been applied to the "Dick mineral," and it seems necessary to propose a new one for it. An appropriate name is suggested by the excellent work of Dick on the material from the island of Anglesey—work so carefully done that later studies have added nothing essential to the data recorded. It therefore seems highly appropriate to use the name "dickite" for this mineral.

DESCRIPTIONS OF THE MINERALS OF KAOLIN

NACRITE

OCCURRENCE

Nacrite appears to be a rare mineral, and specimens from only two sources have been available for examination. In both these occurrences it is associated with ore deposits. The nacrite from Brand, Saxony, is commonly grown around fresh galena crystals, and that from the Eureka shaft, St. Peters Dome, Pikes Peak district, Colo., is associated with mica and cryolite where a pegmatite has been profoundly altered.[44]

CRYSTALLOGRAPHIC AND OPTICAL PROPERTIES

The nacrite from Brand, Saxony, the type locality, forms large crystal groups several millimeters in diameter that do not show good crystal faces. Radial groups of crystals are characteristic, and triplet groups united on the (110) faces are common. The optical properties are as follows: Crystal system monoclinic. Orientation of the axial plane, normal to plane of symmetry. The b axis = obtuse bisectrix Z. X = acute bisectrix (Bx_a), inclined 10° to the normal to $c(001)$; sign of inclination not determined. The extinction on $b(010)$ is therefore 10°; this is shown when the crystal plates are standing on the acute intersections of the $m(110)$ faces.

The indices of refraction are $\alpha = 1.557$, $\beta = 1.562$, $\gamma = 1.563$. Birefringence 0.006. $2V = 40° \pm 5$, $2E = 70°$. Dispersion $\rho > v$. Optical character negative.

The following crystal forms were recognized under the microscope on nacrite crystals from St. Peters

[43] On the limitations of the law of priority in mineral nomenclature see Dana, E. S., System of mineralogy, 6th ed., p. xliii, 1892.

[44] Cross, Whitman, and Hillebrand, W. F., Minerals from the neighborhood of Pikes Peak, Colo.: U. S. Geol. Survey Bull. 20, pp. 41, 42, 67–68, 1885.

Dome: *c*(001), the basal pinacoid; *m*(110), unit prism; and *b*(010), the side pinacoid. The optical properties of the nacrite from St. Peters Dome are similar to those of the Brand material. The *b* faces are absent or very slightly developed, and the crystal plates are in general elongated in the direction of the *b* axis. However, the optic axial angle ($2V$) is nearly 90°, varying slightly in both the (+) and (−) directions from that value. For this reason the optical character may be either positive or negative, and the dispersion may be $\rho > v$ or $\rho < v$.

Optical properties of nacrite

	Optical character	2V	α	β	γ	Dispersion	Extinction
1	—	40°	1.557	1.562	1.563	$\rho > v$	10°
2	+ or −	90°	1.560	1.563	1.566	$\rho > v$ / $\rho < v$	12°

1. Nacrite, Brand, Saxony. Forms transparent wedge-shaped plates with pearly luster. The larger crystals are 5 millimeters or more in diameter. Some have a pseudohexagonal outline.
2. Nacrite, St. Peters Dome, Pikes Peak district, Colo. Forms twinned groups and wedge-shaped transparent plates with pearly luster.

CHEMICAL COMPOSITION

Chemical composition of nacrite

	1	2	3
SiO_2	44.75	45.91	46.5
Al_2O_3	39.48	39.65	39.5
Fe_2O_3	.53		
MgO	.19		
CaO	.13		
CaF_2		.68	
H_2O-	.61		
H_2O+	14.40	13.77	14.0
	100.09	100.01	100.0
SiO_2: Al_2O_3	192:100	196:100	200:100

1. Brand, Saxony. F. A. Gonyer, analyst.
2. St. Peters Dome, Colo. W. F. Hillebrand, analyst.
3. Theoretical composition.

The table shows that the molecular ratio of silica to alumina is definitely 2, the analytical data agreeing very closely with the formula $2H_2O.Al_2O_3.2SiO_2$.

PHYSICAL PROPERTIES

The cleavage of nacrite is perfect on the basal plane (001), less perfect on the planes (010) and (110). The mineral is nonelastic, has a hardness slightly above 2.5, as it scratches muscovite with difficulty, has a specific gravity of about 2.5, and is colorless to white and transparent to translucent. It resembles muscovite but shows a more pearly luster.

Most of the crystal plates show twinning. The most common twinning law gives hexagonal forms made up of three individuals with a pair of (110) faces of one individual in contact with the corresponding faces of the other two; thus, the *b* axes of the group form an equilateral triangle. Nacrite is infusible under the blowpipe but exfoliates on heating.

X-RAY STUDY

Nacrite is more coarsely crystalline than the other kaolin minerals and hence must be finely ground in order to produce a satisfactory X-ray diffraction pattern. Specimens properly ground and screened through china silk (−300 mesh), however, give a satisfactory pattern which is distinctive from the other kaolin minerals and also from the minerals of the mica group. The measurements of the interplanar spacings of nacrite in Ångström units, together with the relative intensities of the lines based upon an estimated maximum intensity of 10, are shown in Plate 40. The table on page 169 gives the measurements of the interplanar spacings in Ångström units compared with those of dickite and kaolinite. Prints of the X-ray diffraction pattern for nacrite are pictured in Plate 39, *A*.

Rinne [45] examined nacrite from "Freiberg" (doubtless from Brand, which is near Freiberg) by means of X rays and noted that a difference existed between an X-ray pattern of "Zettlitz kaolin" and one of nacrite. The "Zettlitz kaolin" is probably kaolinite as here defined. The experimental evidence given by Rinne is in accord with our conclusions that nacrite is a mineral distinct from kaolinite and dickite. Strutinsky [46] has also examined "nacrite" from Crimea by both optical and X-ray methods, and although he gives insufficient data to draw definite conclusions, his work would seem to confirm that of Rinne.

DICKITE

OCCURRENCE

Dickite has been most commonly reported as occurring in association with metallic minerals, and this is true of the samples from Anglesey, National Belle mine, Neurode, Chihuahua, and Peru. The dickite from Greenwood, Ark., occurs in small veinlets cutting quartzite. Backbone Mountain, near Williams, Okla., is in the same folded mountain region as Greenwood, and the mineral probably has a similar origin in both localities. The dickite from Morococha, Peru, forms a coating on the interior of a small cavity in enargite and pyrite. That from Keokuk, Iowa, is a filling of the interior cavity of a geode. Dickite is a rarer

[45] Rinne, F., Röntgenographische Untersuchungen an einige feinzerteilten Mineralien, Kunstprodukten und dichten Gesteinen: Zeitschr. Kryst. Min., vol. 60, p. 63, 1924.
[46] Strutinsky, L. B., Rentgenograficheskoe issledovanie glin: Russ. Phys. Chem. Soc. Jour., vol. 58, pp. 314–325, 1926.

mineral than kaolinite and is reported to occur in large bodies only at Cusihuiriachic, Chihuahua, Mexico. In all the deposits it forms small crystals that range from a few thousandths to several tenths of a millimeter in diameter. In some occurrences its crystals are practically perfect, so that the crystal forms and optical properties can be completely determined.

CRYSTALLOGRAPHIC AND OPTICAL PROPERTIES

The almost perfect crystals from the National Belle mine, Ouray, Colo., have been the basis for the following study. Their properties are nearly identical with those of material from Greenwood, Ark., and of that from the vicinity of Amlwch, island of Anglesey, which has been completely described by Dick [47] and Miers.[48]

The optical properties of the National Belle dickite are as follows: Crystal system monoclinic. The orientation of the axial plane is normal to the plane of symmetry and inclined 16°, rear to the normal to c (001). The b axis = acute bisectrix (Bx_a) Z. Obtuse bisectrix (Bx_b) X is inclined 16° in the rear to the normal to (001). The extinction on b (010) is therefore 16°. Y inclined to a axis (trace of c (001) on b (010)) 23°. Optically positive. Indices of refraction: $\alpha = 1.560$, $\beta = 1.562$, $\gamma = 1.566$. Birefringence, 0.006. $2V = 80°$. Dispersion $\rho < v$. The angles of extinction with the cleavage (001) are distinctly greater (about 3°) in red than in violet light,[49] and under crossed nicols many of the crystals show anomalous blue interference colors.

The following crystal faces were listed by Dick [50] on the Anglesey material: b(010) = clinopinacoid; m(110) = basal pinacoid; and the presence of μ(111) = + unit pyramid was indicated by his measurements.

A clinodome (okl) was reported on crystals from both Anglesey and the National Belle mine by Dick.

Examination of material under the microscope indicated that all the above-named faces and in addition a(100) = orthopinacoid and a second pyramid of undetermined indices are present, and the presence of μ(111) = + pyramid was confirmed. The axial angle is essentially as given by Miers,[51] $\beta = 83° 11'$.

These optical properties show that the acute bisectrix is not perpendicular or nearly so to the basal plane but corresponds with the b axis and is perpendicular to the trace of the cleavage, when the crystals lie on the (010) face, and the plane of the optic axes is inclined 16° to the normal to the cleavage (001).

The most essential optical properties, as pointed out by Dick, are the positive optical character, the strongly inclined extinction, and the acute bisectrix normal to the clinopinacoid and to the trace of the cleavage.

Optical properties of dickite

[Optical character positive in all specimens]

	α	β	γ	2V	Dispersion	Extinction
1	1.560	1.562	1.566	80°	$\rho < v$	16°
2	1.560	1.561	1.566	52°	$\rho < v$	18°
3	1.561	1.562	1.567	Large.	$\rho < v$	17°
4	1.561	1.562	1.567	52°	$\rho < v$	15°
5	1.560	1.562	1.566	Large.	$\rho < v$	17°
6	1.560	1.561	1.566		$\rho < v$	18°
7	1.561	1.563	1.567			
8	Mean index 1.563			68°	$\rho < v$	15°–20°

1. Dickite from National Belle mine, Red Mountain, Ouray, Colo. (See pl. 39, *B*.) An incoherent glistening powder made up of 6-sided plates that range from 0.04 to 0.2 and average about 0.08 millimeter in diameter. Many of the crystals are in the form of curved or straight piles of plates that reach a maximum height of 0.3 millimeter. Optical properties determined by Ross.

2. Dickite from Greenwood, Sebastian County, Ark. A powder like No. 1, though buff from admixed iron oxide, but becoming white on leaching with dilute acid. Occurs in the form of small crystal plates and piles of such plates. These range from 0.03 to 0.16 millimeter in greatest diameter and are about three times as long in the direction of the a axis as in the direction of the b axis. Optical properties determined by Ross. Another lot of material from Greenwood (see analysis 2, p. 160) forms small white veins in quartzite that have a maximum width of 2 centimeters. The crystals form vermicular groups that average about 0.02 millimeter in diameter. The optical properties are like those of the more coarsely crystalline material described above. Optical properties determined by Ross.

3. Dickite (so-called pholerite) from Neurode, Silesia. Forms fine-grained massive veins of pale-greenish color. The mineral occurs in vermicular groups of crystals that have an average diameter of 0.01 millimeter. The crystals are so small that the axial angle can not be directly measured, but it is large, and the optical character is positive. Optical properties determined by Ross.

4. Dickite from Cusihuiriachic, Chihuahua, Mexico, a mining camp where the material is reported to occur in large deposits. Compact masses, pure white. The habit is vermicular, and the average diameter of the crystals is about 0.03 millimeter. Optical properties determined by Ross.

5. Dickite from Morococha, Peru. Fine-grained white material, associated with enargite. Small pseudohexagonal plates about 0.06 millimeter in diameter. This occurs in small vugs in vein material made up of coarsely crystalline enargite that is fresh and without apparent traces of alteration. Specimen furnished by the courtesy of Dr. C. P. Berkey. Optical properties determined by Ross.

6. Dickite from vicinity of Keokuk, Iowa. Fine-grained glistening white powder filling the cavity of a geode (U. S. Nat. Mus. No. 80551). Optical properties determined by Ross.

7. Dickite from Backbone Mountain, 2 miles north of Williams, Le Flore County, Okla. Glistening or fine satiny white lumps. The crystals have about the following dimensions: c axis 0.02 millimeter, b axis 0.04 millimeter, a axis about 0.06 millimeter. Optical properties determined by Schaller.[52]

[47] Dick, A. B., On kaolinite: Mineralog. Mag., vol. 8, pp. 15–27, 1888; Supplementary notes on the mineral kaolinite: Idem, vol. 15, pp. 124–127, 1908.

[48] Miers, H. A., Idem, vol. 8, p. 25, 1888; vol. 9, p. 4, 1890.

[49] Winchell, N. H., Elements of optical mineralogy, p. 239, New York, 1927.

[50] Mineralog. Mag., vol. 8, p. 25, 1888.

[51] Idem, vol. 9, p. 4, 1888.

[52] Schaller, W. T., and Bailey, R. K., Intumescent kaolinite: Washington Acad. Sci. Jour., vol. 6, pp. 67–68, 1916.

8. Dickite from Amlwch, island of Anglesey. Occurs in cavities in quartzite in hard bluish slate and associated with chalcopyrite. Optical properties compiled from descriptions by Dick[53] and Miers.[54]

CHEMICAL COMPOSITION

Chemical analyses of dickite

	1	2	3	4	5	6
SiO_2	43. 10	44. 64	45. 04	46. 35	46. 53	46. 55
Al_2O_3	40. 10	40. 42	40. 70	39. 59	38. 93	38. 90
Fe_2O_3	. 64	. 32	Faint trace.	. 11		
MnO	None.					
MgO	. 20	. 05	Trace.			
CaO	. 24	. 34	. 22			
TiO_2			Trace.			
H_2O-	1. 08	. 04	None.	13. 93	14. 54	14. 04
H_2O+	14. 82	13. 98	14. 08			
F				. 15		
O⇌F				—. 06		
	100. 18	99. 79	100. 04	100. 13	100. 00	99. 49
$SiO_2:Al_2O_3$	183:100	188:100	188:100	194:100	197:100	202:100

1. Neurode, Silesia.
2. Greenwood, Ark.
3. Cusihuiriachic, Chihuahua.
4. National Belle mine, Colo.
5. Island of Anglesey.
6. Backbone Mountain, Okla.

Analyses 1, 2 by F. A. Gonyer; 3 by J. G. Fairchild, U. S. Geol. Survey; 4 by W. F. Hillebrand, U. S. Geol. Survey Bull. 20, p. 97, 1885, and Bull. 419, p. 296, 1910; 5 by —— Tookey, Metallurgy, vol. Fuel, p. —, 1875, and Mineralog. Mag., vol. 8, p. 15, 1888; 6 by R. K. Bailey, Washington Acad. Sci. Jour., vol. 6, pp. 67–68, 1916.

The table shows that in all but analysis 6 the silica is a little below the 200:100 ratio, the accepted one for this mineral. The specimens from Neurode, Greenwood, and Chihuahua are very fine grained, and a very small proportion of bauxitelike material or allophane, which would lower the ratios, might escape detection. The other three specimens occur as coarser crystals, and their analyses are very close to the ideal formula, $2H_2O.Al_2O_3.2SiO_2$ (the same as for nacrite).

PHYSICAL PROPERTIES

The cleavage of dickite is perfect on the basal plane (001). The crystals are so small that imperfect cleavages are not evident if present. The crystal plates are probably somewhat elastic, but their small size makes exact observation difficult. The hardness is greater than 2.5, as the crystals scratch muscovite. The specific gravity is about 2.62.[55] The color in mass is white to pale cream color; massive material from Neurode is pale green. Small crystal plates are perfectly transparent. The larger crystals of dickite from the National Belle mine are 0.1 to 0.2 millimeter in greatest diameter. A few of the crystals are single crystal plates, but much of the material is in the form of crystal groups. In some of these the individual crystals are all in parallel orientation, with the a(001) faces in contact. The longest of these groups are several times as long as the diameter of the individual crystals. In other groups the crystal plates have a radial position like the leaves of a half-open book. In some groups the (001) faces are in contact but one crystal is rotated with reference to the other, and 9° is a very common angle between the b axes of the two individuals. The parallel groups give the appearance of cleavage, but each plate is an individual crystal showing its own pyramidal faces and is easily separated from its neighbor. The crystals show no cleavages until they are crushed, but then a perfect cleavage parallel to the basal plane is exhibited.

Dickite adsorbs dyes with difficulty and does not become pleochroic like kaolinite.

The crystals of the coarser-grained specimens intumesce strongly on heating.[56] Dickite is infusible under the blowpipe.

X-RAY STUDY

Dickite generally occurs in crystals sufficiently coarse to give a spotted X-ray pattern unless the mineral is ground to −300 mesh. Ground samples, however, give more satisfactory X-ray diffraction patterns than the other kaolin minerals. The lines are sharp and occur in sufficient abundance to give assurance to comparisons. There is a certain general resemblance between the patterns of dickite and those of the other two minerals of kaolin, but the differences in detail are numerous and make certain that the three are actually distinct in structure. The interplanar spacing for dickite in Ångström units is given in Figure 11, and representative X-ray diffraction patterns are shown in Plate 40.

McVay and Thompson[57] give a series of X-ray diffraction patterns and state that the diffraction patterns of the National Belle and Red Mountain "kaolinite" agree with each other and with those of the Georgia and English china clays. Study of their table of measurements,[58] however, fails to support this conclusion. For instance, in a pattern for "Georgia raw" one line at 1.62 Å.u. is noted, and in "A1 China raw" a corresponding line is given at 1.61 Å.u. In both the Red Mountain and National Belle samples, however (although the patterns are better), the line is missing. This would indicate a difference of at least one line between the two sets of patterns. Numerous

[53] Dick, A. B., On kaolinite: Mineralog. Mag., vol. 8, pp. 15–27, 1888; Supplementary note on kaolinite: Idem, vol. 25, pp. 124–127, 1908.

[54] Miers, H. A., idem, vol. 8, p. 25, 1888; vol. 9, p. 4, 1890.

[55] Dick, A. B., Mineralog. Mag., vol. 8, p. 16, 1888.

[56] Cross, Whitman, and Hillebrand, W. F., Contributions to the mineralogy of the Rocky Mountains: U. S. Geol. Survey Bull. 20, p. 97, 1885. Schaller, W. T., and Bailey, R. K., op. cit.

[57] McVay, T. N., and Thompson, C. L., Effect of heat on china clays: Am. Ceramic Soc. Jour., vol. 11, p. 829, 1928.

[58] Idem, p. 832.

other differences exist, but the measurements given by McVay and Thompson are inadequate to bring them out with certainty. It seems evident, however, that their experiments were carried out on two different minerals. The National Belle and Red Mountain patterns would correspond with those of dickite, whereas the "Georgia raw" and "A1 China raw" clays were kaolinite or anauxite, and therefore their conclusion[59] that "kaolinite" (dickite) is the predominating mineral of the clays examined is not justified by their experimental evidence.

KAOLINITE-ANAUXITE

OCCURRENCE

Kaolinite is the dominant mineral in residual kaolin and is commonly the result of profound weathering of feldspathic rocks like pegmatite and granite. It is the characteristic clay mineral in extensive sedimentary beds where it has been redeposited after erosion from areas of kaolinized rocks; less commonly it has formed in place through the profound weathering of aluminous sedimentary rocks and through the action of sulphate or carbonate bearing waters on aluminous rocks. Kaolinite is almost the only kaolin mineral that forms extensive deposits and so is the source of most if not all of the kaolin and "china clay" of commerce. It is commonly associated with halloysite, a submicroscopically crystalline clay material of similar composition and origin that will be discussed in a future paper, and more rarely with allophane.

Kaolinite most commonly forms minute vermicular crystals or very thin crystal plates like those that have been reported by many investigators.[60] The vermicular crystals vary greatly in size. The crystals in residual kaolin deposits are commonly about 0.005 to 0.01 millimeter in diameter, but in some of the sedimentary kaolins they reach a maximum diameter of 3 millimeters or more. For instance, the crystals in a sample of Cornwall kaolin were 0.002 to 0.06 millimeter in diameter, and those from Ione, Calif., 0.003 to 2.5 millimeters. The vermicular crystal groups are commonly several times as long as wide. Some of the types of crystals are shown in Plates 39, 41, and 43.

CRYSTALLOGRAPHIC AND OPTICAL PROPERTIES

The crystal faces of most kaolinite crystals are not sharp, probably because they have been washed from the inclosing material and thus somewhat rounded by attrition, and accordingly only the dominant crystal forms are known. Those recognized are $b(010)$ = clinopinacoid, $c(001)$ = basal pinacoid, and $m(110)$ = unit prism.

The kaolinite 1 mile south of Ione, Amador County, Calif., is characteristic, and its optical properties in detail are as follows: Crystal system probably monoclinic. The orientation of the axial plane is normal to the plane of symmetry and approximately normal to $c(001)$ and parallel to $a(100)$. The b axis = obtuse bisectrix Z (Bx_o). X = acute bisectrix (Bx_a), inclined 3° 30′ to the normal to the basal cleavage $c(001)$ (sign of inclination unknown). Extinction on $b(010)$ is therefore 3° 30′—that is, when the crystal lies on the side pinacoid face. Y lies nearly parallel to a axis. Optically negative. Indices of refraction $\alpha = 1.561$, $\beta = 1.565$, $\gamma = 1.566$. Birefringence, 0.005. $2V$ variable, mean value about 42°; $2E$, 68°. Dispersion $\rho > v$.

The variation of the optic angle $2V$ is shown by the following measurements made on different crystals of Ione kaolinite taken from one lot of material:

	Number of measurements		Number of measurements
57°	1	40°	1
53°	1	36°	2
49°	1	25°	2
46°	6	18°	1

The variability of the axial angle is greater in the Ione kaolinite than in most specimens, and this is caused in part, although probably not wholly, by distortion due to the warping of the flexible crystals.

A few specimens are naturally gray or brown, and these crystals show pleochroism. Absorption is $Z = Y > X$. The pleochroism in the kaolinite ("leverrierite") from St.-Étienne, France, is X = colorless, Y and Z = "cream-buff, 19″ YO-Y" of Ridgway's "Color standards," and in that from Alaska X = "cream-buff, 19″ YO-Y," Y and Z = "dark olive-buff, 21‴ O-YY." Colorless crystals that have absorbed a brown immersion oil become slightly pleochroic. When stained by a dye like malachite-green the pleochroism is very striking.

Dittler and Hibsch[61] give the following properties for anauxite from Bilin, Czechoslovakia: Crystal forms (110), (010), and (001). Angle (110): $(1\bar{1}0)$ about 60°. $\alpha = c$, $\beta = a$, $\gamma = b$. $\alpha = 1.54$, $\gamma = 1.55$, $\gamma - \alpha = 0.01$. Optical character negative. Hardness 2.5. Specific gravity 2.524.

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[59] McVay, T. N., and Thompson, C. L., op. cit., p. 840.

[60] Merrill, G. P., The nonmetallic minerals, pl. 21, 1904. Cook, G. H., and Smock, J. C., Report on the clay deposits of * * * New Jersey, p. 281, New Jersey Geol. Survey, 1878. Ries, Heinrich, Report on the clays of Maryland: Maryland Geol. Survey, vol. 4, p. 48, 1902.

[61] Dittler, E., and Hibsch, J. E., Min. pet. Mitt., vol. 36, p. 85, 1927.

Optical properties of kaolinite and anauxite

[Optical character negative and dispersion $\rho > v$ in all specimens]

		α	β	γ	2V	Extinction	Cleavage (perfect)
					°	° ′	
1, 2	Anauxite	1.559	1.564	1.565	36	1 30	(001)
3	do	1.559	1.564	1.564	37	3 30	(001)
4	do	1.561	1.566	1.567	38	3 00	(001)
5	Kaolinite	1.553	1.559	1.560	42	2 30	(001)
6	do	1.559	1.564	1.565	50	3 30	(001)
7	do	1.557	1.563	1.564	38		(001)
8, 9	do	1.561	1.565	1.566	36	3 30	(001)
10	do	1.561	1.567	1.567	23		(001)
11	do	Mean 1.563			40		
12, 14	do	1.562	1.567	1.568	40	3 00	(001)
13	do	1.566	1.561	1.562	50		(001)
15	Newtonite	1.561	1.559	1.599			
16	Leverrierite	1.559	1.565	1.566	32	3 00	(001)
17	Kaolinite	1.563	1.568	1.569	42	3 30	(001)
18	do	1.560	1.565	1.566	42	3 30	(001)
19	do	1.563	1.569	1.570	24	1 ±	(001)

1, 2. Anauxite from Bilin, Czechoslovakia.[62] An alteration product of porphyritic crystals of augite or biotite in a basalt by thermal carbonated waters. The crystals form vermicular groups, as shown in Plate 41, *A*, which range from 0.01 to 0.5 millimeter in diameter. Nearly white.

3. Anauxite from locality 1 mile west of Lancha Plana, on the banks of the Mokelumne River, Calif.,[63] in arkosic Ione sandstone. Crystals from 0.1 to 3.0 millimeters in diameter. Slightly stained with iron oxide, but those free from color were selected for analysis.

4. Anauxite from Newman pit, near Ione, Calif., Ione formation. Pale cream color; maximum diameter of crystals 1.5 millimeters.

5. Kaolinite from Sand Hill station, near Pontiac, S. C., in a well at a depth of about 175 feet. White grains 0.5 to 1.5 millimeters in diameter and 1 to 4 millimeters long. (See pl. 41, *D*.) Kaolinite crystals embedded in the halloysite matrix between the sand grains of an arkose.

6. Kaolinite from Mexia, Tex. (See pl. 43, *B*.) Embedded in the clay matrix of an arkose. The nearly white crystals have a maximum diameter of 1 millimeter.

7. Kaolinite from Roseland, Va. Small vermicular crystals in an altered pegmatite. White.

8, 9. Kaolinite from locality 1 mile south of Ione, Amador County, Calif. (See pl. 41, *C*.) White crystals 0.003 to 2.5 millimeters in diameter.

10. Kaolinite from the Abatik River, northern Alaska. Upper Cretaceous. Collected by P. S. Smith, U. S. Geological Survey. Formed by the alteration of biotite grains about 2 millimeters in diameter embedded in a bentonite (composed of the mineral montmorillonite) derived from the alteration of volcanic ash. Light gray.

11. Kaolinite from United Verde Extension mine, Jerome, Ariz. Collected by F. L. Ransome. Fine-grained massive white clay formed by the action of sulphate solutions on rhyolite.

12, 14. Kaolinite from Globe bauxite mine, just south of Bauxite, Saline County, Ark. Small vermicular crystals 0.005 to 0.4 millimeter in diameter in white veinlike masses of bauxite.

13. Kaolinite from Franklin, N. C. (See pl. 39, *A*.) In small vermicular crystals 0.002 to 0.02 millimeter in diameter, associated with halloysite and small amounts of an unnamed clay mineral (see p. 172) and residual quartz and muscovite, in a clear-white kaolin derived from a weathered pegmatite.

15. Kaolinite associated with alunite and halloysite, Newton County, Ark.[64] Collected by E. T. Wherry.

16. "Leverrierite" from St.-Étienne, Département du Gard, France. Abundant large brown crystals that reach a maximum diameter of 2.5 millimeters. (See pl. 39, *B*.)

17. Kaolinite from Hohberg, Saxony. Micaceous white crystals with a maximum diameter of 2 millimeters. Formed during the weathering of a granitic rock.

18. Kaolinite from weathered granite, Tokitsu, Gifugen, Japan.

19. Kaolinite from locality 4 miles southwest of La Plata, N. Mex., where it forms about 30 per cent of an arkosic sandstone in the McDermott formation (Upper Cretaceous?).

The optical properties in all specimens were determined by Ross.

The optical properties of these 19 samples of kaolinite and anauxite agree with one another within very narrow limits. It is particularly noteworthy that the indices of refraction of the anauxite with an alumina-silica ratio of around 1:3 do not differ perceptibly from those of kaolinite with a ratio of 1:2.

A list of 36 localities from which kaolinite has been identified through a determination of the optical properties is given below. All the identifications were made by Ross.

Dulac, Tenn	In sedimentary kaolin.
Lincoln County, Mo	Do.
Murfreesboro, Ark	Do.
Belle, Mo	Do.
Lookout, Ala	Do.
Richmond County, Va	Do.
Anna, Ill	Do.
Calhoun County, Ill	In Pennsylvanian shale (flint clay).
Quincy, Fla	In sedimentary fuller's earth.
Louisville, Miss	Associated with halloysite in arkosic sandstone.
Hirschau, near Amberg, Bavaria	In arkosic sandstone.
Elk Grove, Calif	Do.

[62] Smirnoff, W. P., Über ein krystallinisches Verwitterungsproduct des Augits: Zeitschr. Kryst. Min., vol. 43, pp. 338–346, 1907.

[63] Allen, V. T., Anauxite from the Ione formation of California: Am. Mineralogist, vol. 13, pp. 145–152, 1928.

[64] Foshag, W. F., The identity of newtonite with alunite: Am. Mineralogist, vol. 11, pp. 33–35, 1926.

Rolla, Mo	Associated with diaspore in flint clay.
Phillips, Pa	Do.
Cherokee County, Tex	Altered biotite in bentonitic clay.
Roanoke, Va	Altered product of granitic rock.
Spruce Pine, N. C	In altered granite.
Stein, near Fredeburg, Silesia	Do.
Zettlitz, Czechoslovakia	Do.
Hohberg, Saxony	Do.
Tokitsu, Gifugen, Japan	Do.
Augustusburg, Saxony	Do.
Saavan, near Breslau, Silesia	Do.
Cornwall, England	Do.
Gladys, Va	In altered pegmatite.
West Cornwall, Conn	In altered feldspathic quartzite.
Near Saragosa, Tex	In altered pegmatite.
Bärtewitz, near Mügeln, Saxony	In quartz porphyry.
Near Cambria, Wyo	In oil shale.
Eiffels, France ("lenzinite")	Associated with halloysite.
Cloudfontaine, Belgium	Do.
Near Nashville, Ark	In altered volcanic tuff.
Calhoun County, Ark	Do.
Lone Mountain, Tenn	Altered biotite in bentonite.
St. Rémi, Quebec	In replaced quartzite.
Flathead mining district, Montana.	In contact-metamorphic rock.

Hendricks [65] has recently published a significant paper on the space lattice diffraction pattern of kaolinite. The material investigated was part of that from Ione, Calif., described on page 161. This work shows that the megascopic kaolinite crystals are made up of a "micaceous crystalline mass in which the constituent crystal units diverge by as much as 5° from being parallel to the cleavage surface [of the megascopic crystal] but which are not rotated about the normal to that surface." Photographs made with the cleavage plates normal to the incident beam of X rays show a symmetrical pattern with an apparent sixfold symmetry. The data now available do not permit a complete determination of the effect on the optical properties of this divergence from parallel of the minute crystal plates, but it probably does not cause significant differences in the optical properties. This feature will be further studied and discussed in a future paper.

CHEMICAL COMPOSITION

The following table gives the chemical composition of 14 specimens of kaolinite and anauxite, arranged in the order of decreasing silica : alumina ratio

Chemical analyses of kaolinite and anauxite

[Numbers correspond to those used in the table on p. 162]

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
SiO_2	54. 32	53. 80	52. 46	48. 80	45. 56	44. 81	45. 44	44. 70	44. 74	43. 64	44. 92	44. 06	44. 26	43. 78
Al_2O_3	29. 96	32. 48	32. 20	35. 18	37. 65	37. 82	38. 52	38. 64	37. 97	38. 33	40. 22	39. 44	40. 22	40. 06
Fe_2O_3	2. 00	1. 12	1. 69	1. 24	1. 35	. 92	. 80	. 96	1. 44	1. 43	. 54	. 80	. 30	. 64
MnO	------	------	------	------	None.	None.	------	None.	None.	------	------	Trace.	None.	------
MgO	. 14	. 26	None.	None.	. 07	. 35	. 08	. 08	. 06	1. 02	. 14	. 26	. 18	. 16
CaO	. 32	. 34	. 03	. 22	. 10	. 43	. 08	. 24	. 09	1. 48	. 08	. 06	. 32	. 36
K_2O	None.	([a])	. 31	. 40	. 11	([a])	. 14	. 14	. 16	([a])	([a])	([a])	([a])	([a])
Na_2O	. 37	([a])	. 25	. 25	1. 16	([a])	. 66	. 62	. 76	([a])	([a])	([a])	([a])	([a])
TiO_2	------	------	. 55	. 61	. 19	. 37	. 16	. 22	. 27	------	------	------	None.	------
H_2O-	. 84	. 94	1. 38	1. 16	. 76	1. 10	. 60	. 64	. 58	. 60	. 08	1. 06	. 64	1. 02
H_2O+	11. 80	10. 98	12. 07	12. 81	13. 66	14. 27	13. 60	13. 88	13. 98	13. 64	14. 22	14. 16	14. 16	14. 08
	99. 75	99. 92	100. 94	100. 67	100. 61	100. 07	100. 08	100. 12	100. 05	100. 14	100. 20	99. 84	100. 08	100. 10
$SiO_2:R_2O_3$	294:100	274:100	267:100	230:100	202:100	199:100	197:100	195:100	195:100	189:100	189:100	188:100	186:100	185:100

[a] Not determined.

1. Anauxite, Bilin, Czechoslovakia. William F. Foshag, analyst. Sample consisted of altered augite phenocrysts that averaged about 1 centimeter in length; these were completely decomposed and contained no residual augite. The outer borders, which had lain in contact with the heterogeneous clay matrix (cimolite), were trimmed away and rejected. The cores proved to be composed of vermicular groups of anauxite crystals, as shown on Plate 41, *A*, with a small proportion of very fine grained interstitial clay material that is believed to be a finely crystalline form of the same mineral. The inclosing matrix is high in bases, and the low content of bases in the selected material is believed to indicate its essential purity.

2. Anauxite, Bilin, Czechoslovakia. F. A. Gonyer, analyst. Altered biotite, small cores of unaltered biotite being rejected.

3. Anauxite, Mokelumne River 1 mile west of Lancha Plana, Calif. J. G. Fairchild, United States Geological Survey, analyst. Carefully hand-picked crystals that averaged about 1 millimeter in diameter. The material is slightly stained with iron oxide, but crystals nearly free from stain were selected. This analysis agrees closely with those of the same occurrence given by Allen.

4. Anauxite, Newman pit, near Ione, Calif. J. G. Fairchild, United States Geological Survey, analyst. Cream-white hand-picked crystals that averaged about 0.5 millimeter in diameter.

5. Kaolinite from Sand Hill station, near Pontiac, S. C. F. A. Gonyer, analyst. Very pure hand-picked crystals, nearly white, average diameter about 0.5 millimeter.

6. Kaolinite associated with arkosic sand from Mexia, Tex. F. A. Gonyer, analyst. Hand-picked crystals 0.5 millimeter in diameter, nearly white.

7. Kaolinite in vermicular grains from a weathered pegmatite, Roseland, Va. F. A. Gonyer, analyst. Similar to and treated like sample 13.

8, 9. Kaolinite from locality 1 mile south of Ione, Amador County, Calif. F. A. Gonyer, analyst. Hand-picked crystals nearly 1 millimeter in diameter. Both samples from the same

[65] Hendricks, S. B., Zeitschr. Kryst. Min., vol. 71, Heft 3, pp. 269–273, 1929

lot of material. No. 8 is composed of dull-white crystals, which are the dominant type; No. 9 is composed of semitransparent crystals. The two samples thus represent the most widely dissimilar material in this occurrence, and the fact that the two analyses are so nearly identical shows that there are no essential differences in the chemical composition of the different types of crystals at this locality.

10. Kaolinite, Abatik River, northern Alaska. F. A. Gonyer, analyst. Hand-picked material that was not quite free from impurities.

11. Kaolinite, Jerome, Ariz. Collected by F. L. Ransome. F. A. Gonyer, analyst. Very fine grained massive white aggregates, the homogeneous nature of which was established by studies of thin sections.

12, 14. Kaolinite in vermicular grains from kaolin seams in bauxite, Saline County, Ark. F. A. Gonyer, analyst. Very fine grained white aggregates that were washed free from associated beidellitelike clay, which completely dispersed in water, whereas the kaolin did not. This associated substance had a dark color and a high birefringence, and so its complete removal was easily determined.

13. Kaolinite from Franklin, N. C., from a typical kaolin derived from weathered pegmatite. F. A. Gonyer, analyst. Very fine grained white vermicular crystals. The material had been washed in water to remove associated halloysite, then sized to obtain larger crystals that did not pass 200-mesh bolting silk, and finally separated by repeated treatments in heavy solution in a centrifuge. In this way all the quartz and muscovite was removed, as well as all but a small proportion of the associated mineral with vermicular habit. (See p. 172.)

For comparison an analysis [66] of the standard international sample of Zettlitz kaolin as distributed by the Ceramic Society of Czechoslovakia is given below.

Analysis of Zettlitz kaolin

SiO_2	46.90	SrO	0.006
Al_2O_3	37.40	Cr_2O_3	.015
Fe_2O_3	.65	V_2O_3	.002
P_2O_5	.08	K_2O	.84
TiO_2	.18	Na_2O	.44
ZrO_2	.007	SO_3	.03
MnO	.007	Loss on ignition	
CaO	.29	(105°–110° C.)	12.95
MgO	.27		
BaO	.02		100.087

$SiO_2 : Al_2O_3$, 2.14 : 1. Specific gravity, 2.633.

The water in the kaolinite-anauxite series varies only slightly, inversely with silica, but the alumina and silica seem to vary rather widely. For this reason the complete molecular ratios calculated from the analyses are not given, but only the significant ratios of silica to alumina. In the table on page 163 this ratio is expressed as $SiO_2:R_2O_3$, ferric oxide, which is present in small amount, being combined with alumina. The iron content of the kaolin analyses listed varies from a maximum of 2 to a minimum of 0.30 per cent, Nos. 7, 12, 13, and 14 are residual kaolins formed by the alteration of igneous rocks, and the average iron content is about 0.64 per cent of Fe_2O_3. No. 11 is associated with mineral deposits, where it was formed by the action of sulphuric acid on rhyolite, and, like the residual kaolin, is low in iron, carrying 0.54 per cent of Fe_2O_3. The other analyses listed are all of sedimentary kaolin with an average of 1.36 per cent of Fe_2O_3.

Heretofore the silica-alumina ratio of kaolinite has been regarded as the same as those in nacrite and dickite—namely, 2 : 1. The analytical data here presented indicate, however, that it is not constant. Thus, there are four samples from two localities that show a higher silica ratio than 2 : 1. The material from these four samples showed no detectable variations in optical, X-ray, or physical properties from that with a 2 : 1 ratio and contained no essential amounts of free quartz; and the complete lack of induration indicated that there was little if any cementing by opal or chalcedony. The X-ray diffraction pattern shows the absence of any crystalline form of quartz. Moreover, there was no lowering of the indices of refraction, as would occur if any significant amount of opal were present. This suggests that the higher silica content is an essential feature of these four samples, and that some sort of isomorphous series exists, although the data are insufficient to regard this as fully established. Assuming that this series includes end members with the ratios of 2 : 1 and approximately 3 : 1, it seems appropriate to call them respectively kaolinite and anauxite.

On the other hand, several analyses yield a silica-alumina ratio lower than 2 : 1, which also needs explanation. Five analyses out of 14 show a lower silica ratio than 190 : 100, although none are lower than 185 : 100. Quartz, feldspar, and muscovite are the foreign materials associated with all kaolin, but quartz and feldspar if present as impurities in the samples would raise the silica content, and muscovite would not lower it. It is possible that a small proportion of amorphous bauxite or allophane is present in a few of the samples, for this might escape detection in very fine grained material, especially if exceedingly thin films of colloidal substances formed coatings on the surfaces of the crystals. It seems improbable, however, that all of the excess alumina can be accounted for in this way, and moreover, analyses of "pholerite" (see p. 165) apparently show a still lower silica-alumina ratio, approaching 3 : 2. These facts suggest the existence of a member of the kaolinite-anauxite series that is lower in silica than the 2 : 1 ratio.

Reference to the table on page 168 shows that there is no essential difference in the optical properties of the high-silica and low-silica members of the kaolinite-anauxite series. This accordance may conceivably represent an example of "atomic isomorphism," [67] where silica and alumina proxy one another. The specific refractivity of alumina and silica [68] under these conditions is so nearly the same that no measurable

[66] Bur. Standards Tech. News Bull. 138, p. 146, 1928.

[67] Wherry, E. T., The plagioclase feldspars as a case of atomic isomorphism: Am. Mineralogist, vol. 7, pp. 113-131, 1922.

[68] Gladstone, J. H., and Dale, T. P., Researches on the refraction, dispersion, and sensitivity of liquids: Roy. Soc. London Philos. Trans., vol. 153, p. 337, 1864. Larsen, E. S., The microscopic determination of the nonopaque minerals: U. S. Geol. Survey Bull. 679, pp. 30-32, 1921.

differences in the optical properties are to be expected with varying silica-alumina ratios. An analogous situation is represented in the remarkable similarity between the optical properties of sillimanite ($Al_2O_3 \cdot SiO_2$) and mullite ($2Al_2O_3 \cdot 3SiO_2$).

The following discussion of the possible relations in this series has been furnished by Dr. Edgar T. Wherry[69]:

In recent times the view has been finding favor that isomorphism often involves equal numbers of atoms of similar dimensions irrespective of their valence. In anhydrous silicates, the valence requirements may be satisfied by varying the number of oxygen atoms, as in sillimanite and mullite, which show a sort of isomorphism when their formulas are multiplied by appropriate factors, giving respectively $Al_{16}Si_8O_{40}$ and $Al_{18}Si_6O_{39}$—that is, having the general formula $(Al+Si)_{24}O_{39-40}$.

In hydrous silicates, on the other hand, the valence requirements are more likely to be met by the number of hydrogen atoms varying, so that the series here under discussion might consist of the following end members and median members:

Name	Formula (Al+Si=24)	$SiO_2:Al_2O_3$	Theoretical composition			Recalculated analysis		
			SiO_2	Al_2O_3	H_2O	SiO_2	Al_2O_3	H_2O
Anauxite	$H_{22}Al_{10}Si_{14}O_{54}$	2.8:1	54.30	32.91	12.79	53.31	34.43	12.26
	$(H_{23}Al_{11}Si_{13}O_{54})$	2.4:1						
Kaolinite	$H_{24}Al_{12}Si_{12}O_{54}$	2 :1	46.54	39.50	13.96	46.39	39.70	13.91
Pholerite (?)	$H_{25}Al_{13}Si_{11}O_{54}$	1.7:1	42.67	42.79	14.54	44.42	41.30	14.28
	$(H_{26}Al_{14}Si_{10}O_{54})$	1.4:1						

At the right-hand side of this tabulation the data from three typical analyses of the kaolinite minerals are added for comparison, accessories having been deducted and the three essential constituents recalculated to 100 per cent. Although there is real agreement with theory only in the case of kaolinite, the deviations of the others are little greater than usual in analyses of complex minerals, suggesting the essential correctness of this interpretation of the compounds concerned.

PHYSICAL PROPERTIES

Cleavage of kaolinite is perfect on the basal plane (001), and there are apparently also cleavages corresponding to those of mica that give the six-rayed percussion figure. Cleavage flakes are somewhat flexible but nonelastic. The hardness is slightly above 2.5, as kaolinite scratches muscovite with ease, but not calcite. The specific gravity of the material from Ione, Calif., is 2.585; from Pontiac, S. C., 2.590. The color is most commonly nearly white or a very pale cream color, but a few specimens are pale brown or gray-brown. The (001) face is opaque to translucent and has a pearly luster; the other faces are adamantine. The mineral does not intumesce on heating.

Many of the large crystals of kaolinite contain small holes that have the form of "negative crystals," which are all elongated in the direction of the *c* axis of the inclosing crystals. Some of the negative crystals have a hexagonal outline, but most of them have an orthorhombic cross section like a pseudohexagonal crystal elongated in the direction of the *a* axis. The optical directions of the inclosing kaolinite crystals commonly correspond with the crystal axes of many of these, suggesting that they are in fact negative crystals of kaolinite, but a few do not correspond to the inclosing kaolinite crystal, so that their significance is not clear.

No twinning has been observed in kaolinite crystals.

[69] Personal communication.

Kaolinite adsorbs dyes readily and becomes very strongly pleochroic.

X-RAY STUDY

More X-ray diffraction patterns have been taken of samples of kaolinite-anauxite than of the other two minerals of kaolin. A representative group may be seen in Plate 42. There was no recognizable difference between the pattern of kaolinite and that of anauxite, although a large number of patterns of these two minerals were taken under varying conditions of technique which should have been adequate to bring out any existing difference. The measurements of a selected pattern are shown together with those for dickite and nacrite in Figure 11. Although the patterns of anauxite and kaolinite do not appear separable from each other, they are distinguished with ease from patterns of dickite and nacrite.

The study of the minerals of this group by means of X rays has furnished a convenient guide for other methods of investigation; for instance, the results of X-ray studies have invariably been confirmed by optical examination. Several occurrences described as independent mineral species have been found to give the kaolinite-anauxite type of X-ray pattern. Newtonite, for example, discredited as a mineral species by Foshag, has been examined by means of X rays and found to give a mixed pattern of kaolinite and alunite. Leverrierite (pl. 42) from the type locality has been found to give an X-ray pattern of the kaolinite-anauxite group. Several specimens of so-called lithomarge also give a pattern of the same type. The gradation of the kaolinite-anauxite pattern into that of the still finer grained halloysite and the analogies between this series and allophane show an interesting relationship, which will be taken up in a later paper.

X-ray patterns of the kaolinite-anauxite type have been obtained from clays originating in the following 28 localities:

Bilin, Czechoslovakia	Anauxite.
Mokelumne River, Calif	Do.
Newman pit, near Ione, Calif	Do.
Ione, Calif	Kaolinite.
Pontiac, S. C	Do.
Mexia, Tex	Do.
Roseland, Va	Do.
Bauxite, Ark	Do.
Jerome, Ariz	Do.
Murfreesboro, Ark	Do.
Belle, Mo	Do.
Richmond, Va	Kaolinite.
Near Aiken, S. C	Do.
St. Rémi, Quebec	Do.
Rochlitz, Saxony	Do.
Brooklyn, N. Y	Do.
Franklin, N. C	Do.
Cornwall, England	Do.
Hickory, N. C	Do.
St.-Étienne, Département du Gard, France	Leverrierite.
Newton County, Ark	Newtonite.
Eiffels, France	Lenzinite.
Near Nashville, Ark	Kaolin.
Lawrence County, Ind	A mixture of kaolinite leverrierite.
Hart County, Ky	Do.
Lookout, Ala	Do.
Fichtel Gebirge, Germany	Collyrite.
Freiberg, Saxony	China clay.

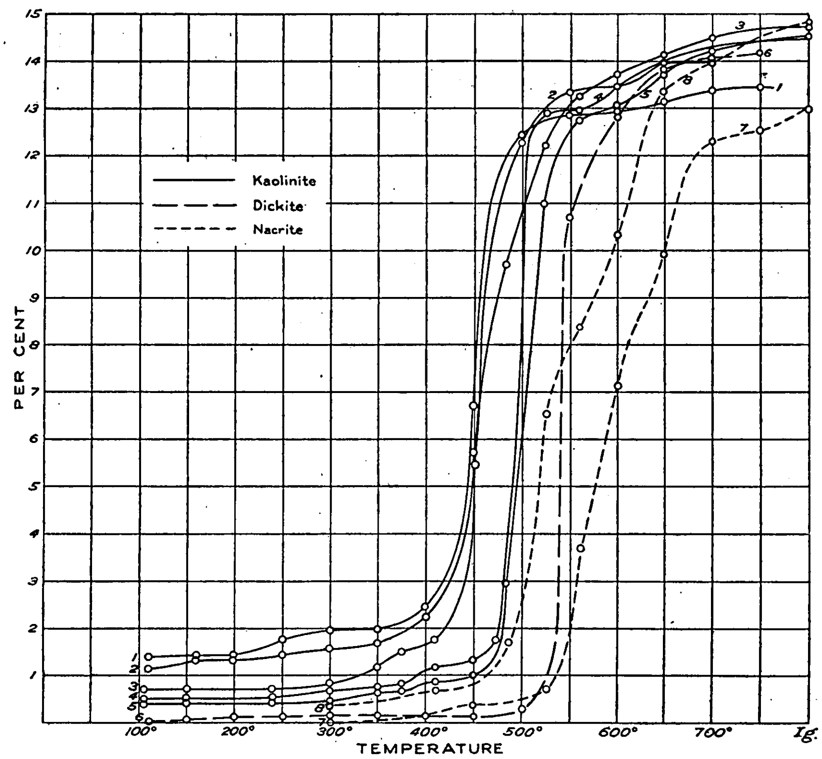


FIGURE 10.—Dehydration curves of kaolinite, dickite, and nacrite. The loss of water (vertical coordinate) is plotted against temperature (horizontal coordinate). It will be seen that kaolinite loses its water at a lower temperature than dickite and that nacrite retains part of its water to a higher temperature than either of the other kaolin minerals

DEHYDRATION STUDY OF KAOLIN MINERALS

The thermal properties and the dehydration of kaolin have been studied by a number of workers interested in the use of clays in the ceramic industries. The chemical composition of the kaolins tested has commonly been determined, but the mineral identity has usually been unknown. For these reasons it seemed advisable to make dehydration tests so that the behavior of known minerals could be compared with the results of previous work.

The dehydration of kaolin has been investigated by Le Chatelier,[70] Rieke,[71] Mellor and Holdcroft,[72] Montgomery and Brown,[73] Pieters,[74] Boege,[75] and many others.

The following table gives the total water lost at various temperatures for samples of dickite, anauxite, kaolinite, and nacrite, and Figure 10 gives the dehydration curves plotted from these data, showing that each mineral has a distinctive curve.

Dehydration temperatures of the kaolin minerals

[Percentages of water lost up to temperatures stated. Determinations by J. G. Fairchild, U. S. Geological Survey]

	110°	160°	200°	250°	300°	350°	375°	400°	450°	485°	500°	550°	600°	650°	700°	750°	Ignition
1	0. 00	0. 03	0. 10	0. 10	0. 12	0. 12	-----	0. 12	0. 12	-----	0. 28	10. 68	12. 82	13. 85	14. 05	14. 08	14. 08
2	1. 16	1. 36	1. 39	1. 45	1. 55	1. 69	-----	2. 25	5. 49	-----	12. 25	13. 35	13. 35	13. 97	13. 97	-----	13. 97
3	1. 38	1. 41	1. 41	1. 78	1. 98	1. 98	-----	2. 44	6. 70	-----	12. 42	12. 84	12. 90	13. 12	13. 38	13. 45	13. 45
	105°	**150°**	**200°**	**240°**	**300°**	**350°**	**375°**	**410°**	**450°**	**485°**	**525°**	**560°**	**600°**	**650°**	**700°**	**750°**	**Ignition**
4	0. 36	0. 40	-----	0. 40	0. 43	0. 63	0. 63	0. 87	0. 98	2. 96	10. 98	12. 74	13. 22	13. 72	14. 19	-----	14. 53
5	. 68	. 73	-----	. 73	. 86	1. 16	1. 53	1. 76	5. 69	8. 94	12. 20	13. 26	13. 70	14. 12	14. 52	-----	14. 67
6	. 47	. 47	-----	. 47	. 64	. 77	. 80	1. 17	1. 30	1. 67	12. 94	12. 94	13. 54	-----	-----	-----	14. 51
7	. 30	. 42	-----	. 42	. 51	. 78	. 78	1. 05	1. 50	1. 98	12. 93	13. 38	13. 71	-----	-----	-----	14. 73
8	-----	-----	-----	-----	. 47	-----	-----	. 67	-----	1. 74	6. 54	8. 37	10. 37	13. 37	13. 97	-----	14. 74
9	. 00	. 00	-----	. 00	. 00	. 17	. 17	. 17	. 37	. 37	. 37	3. 70	7. 10	7. 93	12. 26	12. 46	12. 93

1. Kaolinite, Cusihuiriachic, Mexico.
2. Anauxite, Newman pit, Calif.
3. Anauxite, Mokelumne River, Calif.
4. Kaolinite, Jerome, Ariz.
5. Kaolinite, Franklin, N. C.
6. Kaolinite, Ione, Calif.
7. Kaolinite, Pontiac, S. C.
8. Nacrite, Brand, Saxony.
9. Nacrite, Pikes Peak, Colo.

[70] Le Chatelier, H., Zeitschr. physikal. Chemie, vol. 1, p. 396, 1887.

[71] Rieke, R., Sprechsaal, vol. 44, pp. 637–641, 1911.

[72] Mellor, J. N., and Holdcroft, A. D., Ceramic Soc. [England] Trans., vol. 10, p. 94, 1908.

[73] Montgomery, E. T., and Brown, G. H., U. S. Bur. Standards Tech. Paper 21, 1912.

[74] Pieters, H. A. J., Dehydratatie van het kaolien, Koog-Zaandijk, 1928.

[75] Boege, Hermann, Über den Kaolingehalt von Tonen: Chemie der Erde, Band 3, Heft 2, pp. 341–369, 1927.

Kaolinite and anauxite lose very little water at the lower temperatures, and so their dehydration curves are nearly flat up to the sharp break at about 390° to 450°. There is then a very rapid loss of water till about 525°, where the curves flatten out again, with a slight continued loss of water till dehydration is complete at 700° to 750°. That is, most of the water is lost between 390°–450° and 525°. This determination agrees very closely with the careful and detailed work of Boege,[76] who gives the dehydration data for 10 kaolinitic clays. The curve given by Boege for the pure kaolin from Briesen, near Preschen, Czechoslovakia, involved the determination of 41 points and shows a very sharp break at about 410° and a sudden at 460° and the St. Peters Dome material at 550°. Both curves show a slower loss of water than those for kaolinite or dickite, and the second break, where the curve flattens again, is at 675°. The first break in the curve for nacrite is not distinctive, but the second break at the high temperature of 675° is significant.

The examination of dehydrated and partly dehydrated dickite by means of X-ray diffraction patterns (see fig. 11) gave the results shown below.

Temperature of dehydration	
550°	Dickite pattern.
600°	Two uncertain and very faint lines.
650°	No lines.
800°	No lines.

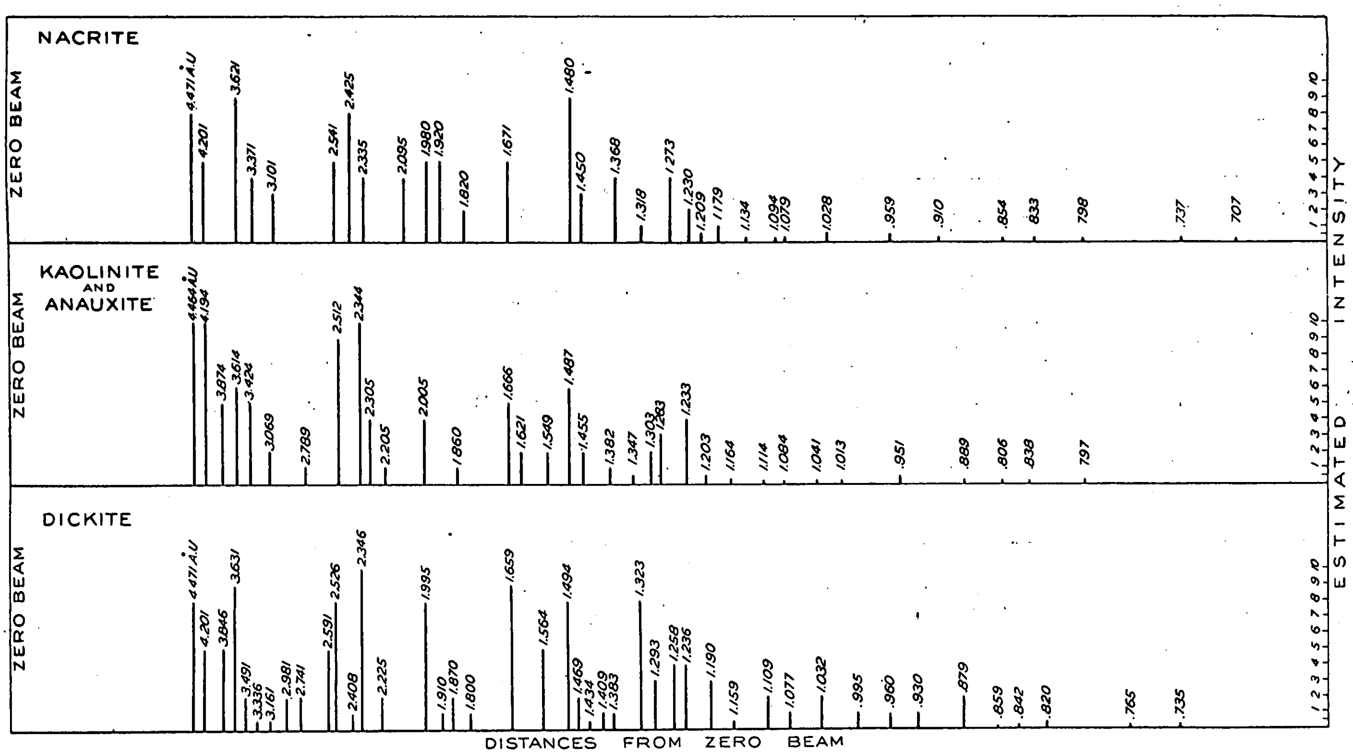


FIGURE 11.—Diagrams showing the relative character of X-ray diffraction patterns of nacrite, dickite, and kaolinite-anauxite. The height of each line represents estimated intensity; the position of the line the relative distance from the zero beam (ribbon of undeviated X rays). The figures above each line are the interplanar spacing in Ångström units (×10⁻⁸ cm.). It will be noted that although a general similarity exists between the three patterns, there are numerous differences in detail

flattening at 450°. Out of a total loss of 13.58 per cent, all but 3.5 per cent was lost between 410° and 540°. All the other determinations by Boege showed breaks at about the same points.

The curve for dickite from Cusihuiriachic, Mexico, is almost perfectly flat till just above 500°, where there is a very sharp break, and most of the water is lost between 500° and 575°, where the curve flattens out again. Thus there is a loss of water amounting to about 13.50 per cent between 510° and 575°.

The two curves of nacrite are not identical, the material from Brand, Saxony, showing its first break

[76] Boege, Hermann, op. cit., pp. 341-369.

The X-ray diffraction patterns taken alone would indicate a change from crystalline dickite to a different phase lacking crystalline structure at a temperature between 550° and 600°. This change occurs at a higher temperature than that shown by the dehydration curve for dickite given in Figure 10. There is an apparent lag in loss of crystallinity as shown by the X-ray patterns behind that to be inferred from the dehydration curve determined by weighing. The former would indicate a change at just below 600°, the latter at about 510°. If the range from 510° to 575° is taken into consideration there is a closer agreement.

Several X-ray patterns of dehydrated and partly dehydrated anauxite from the Newman pit, California, were taken with the following results:

Temperature of dehydration	
110°	Kaolinite-anauxite pattern.
400°	Kaolinite-anauxite pattern.
500°	No lines.
550°	No lines.

These patterns indicate a loss of crystallinity between 400° and 500°, approximately 100° lower than in the case of dickite.

COMPARISON OF KAOLIN MINERALS

The differences and similarities of the kaolin minerals can be best shown by the following table of their properties:

Comparison of the properties of the kaolin minerals

Similarities

	Nacrite	Dickite	Kaolinite-anauxite
Chemical composition	$2H_2O.Al_2O_3.2SiO_2$	$2H_2O.Al_2O_3.2SiO_2$	$2H_2O.Al_2O_3.2\pm SiO_2$.
Crystal system	Monoclinic	Monoclinic	Probably monoclinic.
Orientation of the axial plane A	⊥(010), nearly ‖(100)	⊥(010), nearly ‖(100)	⊥(010), nearly ‖(100).
Indices of refraction	$\alpha=1.557$ $\beta=1.562$ $\gamma=1.563$	$\alpha=1.560$ $\beta=1.562$ $\gamma=1.566$	$\alpha=1.561$. $\beta=1.565$. $\gamma=1.566$.

Dissimilarities

	Nacrite	Dickite	Kaolinite-anauxite
General appearance	Pearly luster; transparent trilled crystal plates; wedge-shaped cleavage plates.	Clear, transparent thin crystal plates.	Translucent to opaque; greatly elongated (c axis) vermicular crystals.
Orientation of bisectrices	$Bx_a=X$ ∧base normal=10°–12°, $Z=Bx_o=b$ axis.	$Bx_o=X$ ∧base normal=15°–20°, $Z=Bx_a=b$ axis	$Bx_a=X$ ∧base normal=1°–3½° $Z=Bx_o=b$ axis.
Dispersion	$\rho>v$; rarely $\rho<v$	$\rho<v$	$\rho>v$.
Optical character	Negative; rarely positive	Positive	Negative.
Angle of extinction on (010) against base.	10°–12°	15°–20°	1°–3½°.
Angle of extinction for red and blue light.	Not different	Red 3° greater than blue	Not different.
X-ray diffraction pattern	Characteristic	Characteristic	Characteristic.
Dehydration curve	Characteristic	Characteristic	Characteristic.
Staining with dyes	Not readily stained by dyes; not pleochroic.	Not strongly stained by dyes; not pleochroic.	Adsorbs dyes very strongly; becomes very pleochroic.

It will be seen from the foregoing tables that the three minerals of kaolin are similar in chemical composition, except that there is some evidence that kaolinite and anauxite constitute an isomorphous series. Dickite and nacrite clearly belong to the monoclinic crystal system, and the small angle of extinction that is usually observable in kaolinite indicates that it also is monoclinic.

Nacrite is not easily identified by its optical properties alone. The material from Brand, Saxony, has a moderate axial angle and negative optical character, with the acute bisectrix nearly perpendicular to the basal cleavage, and so is very similar to kaolinite in these properties. The nacrite from St. Peters Dome, Colorado, has a large axial angle that varies both + and − from 90°, and therefore the optical character is either positive or negative. The crystal plates are elongated in the direction of the *b* axis, whereas those of dickite are commonly elongated in the direction of the *a* axis. For these reasons the nacrite from St. Peters Dome somewhat resembles dickite of an unusual habit. The most conspicuous properties of nacrite are the decided pearly luster, the radial groups of twins, and the cleavage into wedge-shaped plates. The X-ray diffraction pattern of nacrite differs from those of the other two minerals of kaolin and is the most valuable means for its determination.

Kaolinite and dickite are easily distinguishable by their X-ray diffraction patterns and by their optical properties where crystals of sufficient size are available, as they were in nearly all the samples examined in this investigation. The acute bisectrix of kaolinite is perpendicular to the perfect basal cleavage; that of dickite is perpendicular to the edge of the crystal plates when they lie on the *b* faces. The optical character of kaolinite is negative; that of dickite is positive. The dispersion of kaolinite is $\rho<v$; that of dickite is $\rho>v$. The angle of extinction is small and is usually difficult to distinguish in kaolinite, but it is 15° to 20° in dickite. There is a marked difference in the angle of extinction for blue and red light for dickite, but kaolinite and nacrite do not show this property. Kaolinite forms elongated crystals that are commonly curved; dickite forms single thin crystal plates or booklike groups of crystal plates. In short, all the conspicuous and easily determinable optical properties of kaolinite and dickite are so different that they are easily distinguishable where crystals are available; and the elongated curved crystals of kaolinite are so characteristic that this feature alone will usually differentiate the two minerals.

The differences between dickite and kaolinite in crystal habit and general appearance are well shown in Plates 39, 41, and 43, *A*, *B*. Dickite (pl. 39, *B*) forms flat plates with very sharp crystal faces; kaolinite forms elongated crystals that are commonly curved or twisted.

The differences in the X-ray diffraction patterns of the kaolin minerals are shown in the following table, which gives the interplanar spacing in Ångström units.

Interplanar spacing for nacrite, dickite, and kaolinite, in Ångström units

[Corrected against sodium chloride (100) = 2.814 Å. u. Measurements in millimeters are given as read on film]

Nacrite			Dickite			Kaolinite		
Arc (mm.)	Spacing (Å. u.×10⁻⁸)	Estimated intensity	Arc (mm.)	Spacing (Å. u.×10⁻⁸)	Estimated intensity	Arc (mm.)	Spacing (Å. u.×10⁻⁸)	Estimated intensity
18.3	4.471	8	18.3	4.471	8	18.3	4.464	10
19.5	4.201	5	19.5	4.201	5	19.5	4.194	10
22.6	3.621	9	21.3	3.846	5	21.1	3.874	5
24.25	3.371	4	22.55	3.631	9	22.6	3.614	6
26.4	3.101	3	23.5	3.491	2	23.9	3.424	5
32.2	2.541	5	24.6	3.336	.5	25.8	3.069	2
33.8	2.425	10	25.9	3.161	.5	29.3	2.789	1
35.1	2.335	4	27.5	2.981	2	32.6	2.512	9
39.1	2.095	4	28.8	2.741	2	34.8	2.344	10
41.4	1.980	5	31.6	2.591	5	35.6	2.305	4
42.7	1.920	5	32.4	2.526	8	37.2	2.205	1
45.1	1.820	2	34.0	2.408	1	41.0	2.005	4
49.3	1.671	5	35.0	2.346	10	44.3	1.860	1
55.5	1.480	9	36.9	2.225	2	49.4	1.666	6
56.6	1.450	3	41.2	1.995	8	50.6	1.621	2
60.0	1.368	4	42.9	1.910	1	53.1	1.549	2
62.5	1.318	1	43.9	1.870	2	55.3	1.487	6
65.4	1.273	4	45.6	1.800	1	56.6	1.455	2
67.2	1.230	2	49.7	1.659	9	59.4	1.382	1
68.5	1.209	5	52.7	1.564	5	61.6	1.347	5
70.1	1.179	1	55.2	1.494	8	63.4	1.303	2
72.9	1.134	.25	56.2	1.469	2	64.4	1.283	3
75.8	1.094	.25	57.4	1.434	.5	66.9	1.233	4
76.8	1.079	.25	58.6	1.409	1	68.8	1.203	.5
80.8	1.028	.5	59.7	1.383	1	71.3	1.164	.25
87.0	.959	.5	62.4	1.323	8	74.5	1.114	.25
91.9	.910	.25	63.9	1.293	3	76.5	1.084	.25
98.1	.854	.25	65.7	1.258	4	79.8	1.041	.25
101.2	.833	.25	66.9	1.236	4	82.1	1.013	.25
106.0	.798	.25	69.4	1.190	3	87.9	.951	.5
115.7	.737	.25	71.6	1.159	.5	94.3	.889	.5
121.0	.707	.25	75.0	1.109	2	98.0	.858	.25
			77.1	1.077	1	100.7	.838	.25
			80 3.	1.032	2	106.1	.797	.25
			83.9	.995	1			
			87.0	.960	1			
			89.8	.930	1			
			94.4	.879	2			
			97.6	.859	.25			
			99.9	.842	.25			
			102.5	.820	.5			
			110.6	.765	.25			
			115.6	.735	.25			

The differences in X-ray diffraction patterns between kaolinite and dickite are not great but are ample to distinguish them and have been a great aid in building up criteria for differentiating the minerals of kaolin.

In the course of a study of the minerals of soils Stirling B. Hendricks, of the fertilizer and fixed nitrogen unit, Bureau of Chemistry and Soils, United States Department of Agriculture, has carefully checked the interplanar spacing of kaolinite, dickite, and nacrite in Ångström units, and his results agree with those given in the preceding table. He finds that the X-ray diffraction patterns give clear evidence that the three minerals are distinct. He says:[77]

> X-ray powder diffraction photographs of kaolinite, dickite, and nacrite, taken with iron and copper characteristic radiations, show that each mineral gives a characteristic pattern. A specific kaolin mineral can easily be identified by means of its X-ray diffraction pattern.

[77] Personal communication.

MODE OF ORIGIN OF KAOLIN MINERALS

The mode of formation of kaolin has important geologic bearings and has long attracted the attention of geologists and those interested in the use of clays, but there has been marked disagreement between different investigators about the genetic relations. Some have advanced a purely hydrothermal or "pneumatolytic" origin, others an origin purely through weathering processes, still others have emphasized the influence of sulphate waters, and a few the action of carbonated waters. Some have recognized only a single mode of formation, but others have granted that kaolin might be formed in different ways. It is evident that the proof of the existence of three kaolin minerals instead of one will entirely change the basis of discussion of the origin of kaolin, and it seems probable that most of the differences of opinion may be reconciled when the origin of the three minerals is considered separately.

Any review of the work of others on the origin of kaolin is handicapped by the impossibility of determining which kaolin mineral was the subject of many of the discussions. The limitations of space and lack of information prevent any complete review of previous work, but some of the most pertinent discussions may be considered. These naturally fall into two groups—those that consider the origin of kaolin in general and those that consider the origin of a particular occurrence.

REVIEW OF THE LITERATURE

Parts of a paper by Ries[78] on the origin of kaolin may well be quoted. He says:

> The writer's personal opinion is that all the workable kaolin deposits of the United States and probably many of those of central Europe are the work of surface waters, whether they entered direct from the surface or filtered through a swamp bed of peat.
>
> That kaolin may be formed by postvolcanic vapors or waters is no doubt true, as shown by the formation of this mineral below ground-water level in the wall rock of many veins and in the turquoise deposits of New Mexico, but whether any commercially valuable deposits have thus originated remains to be proven.

Lindgren[79] says:

> I have always believed that kaolin does not ordinarily form in deposits which are formed by igneous emanations or by ascending thermal waters, except possibly very close to the surface where admixture with atmospheric waters is effected. In brief, kaolin is never a high-temperature mineral but is either a product of alteration by descending waters containing sulphuric acid or carbon dioxide or of alteration by ascending weak carbonated waters close to the surface. * * *
>
> The larger part of the water is, however, given off at temperatures from 470° to 500°. A small amount of water is tenaciously held and only given off at red heat. It is held by Stremme that dissociation of Al_2O_3 and SiO_2 takes place at 500° and that at still higher temperatures and under atmospheric pressure these two compounds may reunite to sillimanite. * * *
>
> From these data it is first of all clear that kaolin can not be formed at or near the surface of the earth at a temperature of 470° or higher. * * *
>
> It is natural that only those actively interested in mining geology would fully appreciate this change of kaolin to sericite in depth. This certainly applies to all veins which have been formed in the vicinity of intrusive masses shortly after the intrusion, and it probably applies also to veins formed near the surface, although the presence here of kaolin as a primary mineral can not be denied in the same emphatic manner. The result of inductive reasoning is, then, that the agencies which produce mineral deposits—in the majority of cases heated waters under high pressure—are not capable of developing kaolin from the aluminum silicates of the rocks. * * *
>
> Potassium and lithium micas are, for instance, characteristic of tin deposits, which probably were formed at the temperature of about 500°, and it seems extremely unlikely that any kaolin could be developed simultaneously with the minerals characteristic of the tin veins.

In describing kaolin (dickite) in the Red Mountain district, Colorado, Ransome[80] says:

> As an original constituent accompanying the ores, kaolin occurs abundantly in the stock deposits of the Red Mountain district. * * * From what could be seen it appears to have accompanied the ores to the greatest depths there attained—about 1,300 feet. It was evidently derived from the country rock adjacent to the ore bodies as a product of the alteration by thermal waters.

In another paper[81] Ransome says:

> At Goldfield the intimate association of the kaolinite with the alunite, gold, and sulphides shows that all were formed at the same time and by one general process, which was anterior to and entirely independent of oxidation or weathering.

Weed[82] has described kaolinite as a product of the metasomatic action of the waters of Boulder Hot Springs, Colo. Kaolinization by the action of hydrothermal solutions has been noted by Von Inkey[83] in the dacite country rock of the Nágyág veins, and this was confirmed by Kolbeck.[84] Cross and Penrose[85] suggested that the kaolin associated with ores at Cripple Creek, Colo., was the result of hot solutions but Lindgren[86] says:

> The deep workings of the present day show that kaolin is always connected with oxidation and is not a product of the original mineralization of the district as was supposed by Penrose.

[78] Ries, Heinrich, Origin of white residual kaolin: Am. Ceramic Soc. Trans., vol. 13, pp. 51–74, 1911.

[79] Lindgren, Waldemar, The origin of kaolin: Econ. Geology, vol. 10, pp. 90, 91, 93, 1915.

[80] Ransome, F. L., Economic geology of the Silverton quadrangle, Colorado: U. S. Geol. Survey Bull. 182, p. 73, 1901.

[81] Ransome, F. L., The association of alunite with gold in the Goldfield district, Nevada: Econ. Geology, vol. 2, p. 690, 1907.

[82] Weed, W. H., Ore deposits of the Rico Mountains, Colorado: U. S. Geol. Survey Twenty-first Ann. Rept., pt. 2, p. 253, 1900.

[83] Von Inkey, Béla, Nágyág und seine Lagerstätten, p. 143, Budapest, 1885.

[84] Kolbeck, F., Oesterr. Zeitschr. Berg- u. Hüttenwesen, vol. 36, pp. 25–27, 1888.

[85] Cross, Whitman, and Penrose, R. A. F., jr., Geology and mining industries of the Cripple Creek district: U. S. Geol. Survey Sixteenth Ann. Rept., pt. 2, p. 160, 1895.

[86] Lindgren, Waldemar, Report of progress in the geological survey of the Cripple Creek district, Colorado: U. S. Geol. Survey Bull. 254, p. 21, 1904.

Kaolin that originated in zones of weathering has been described by Barnitzke[87] and Wüst[88] from Meissen and Halle, Germany; by Winkel[89] from the island of Bornholm, Denmark; and by Vogt[90] from Ekersund-Soggendal, Norway.

An elaborate study of the origin of kaolin by Rösler[91] attempts to show that the agents which caused the alteration were gaseous or gaseoaqueous in character. Rösler's untenable arguments and his inability to consider contrary evidence weaken this theory. Ries[92] has discussed the work of Rösler and pointed out its error.

In a recent paper Davison[93] has expressed the opinion that the Cornwall and Devon kaolins have been formed by the action of solutions of magmatic origin. The evidence is summarized as follows:

> The china-clay rocks are more completely kaolinized near the fissure lodes, and the completeness of kaolinization increases with depth.
>
> The china clay is associated with the occurrence of such minerals as tourmaline, gilbertite, and sericite mica.
>
> The physical characters of the west of England china clay are markedly different from those of clays of obvious meteoric origin.
>
> Clay deposits occur with a cover of unaltered granite.
>
> Though clay deposits have been worked to a depth of 350 feet, no instance is known of a deposit dying out in depth.

The origin of kaolin through weathering, with mica as an intermediate product, has been advocated by Selle,[94] Hickling,[95] and Galpin.[96] Hickling says:

> Kaolinite occurs in the form of irregular curved hexagonal prisms, showing strong transverse cleavages, or in isolated plates. The index of refraction and that of double refraction agree with those of the Anglesey kaolinite. The crystals are biaxial, and extinguish straight.
>
> The turbidity of feldspar seen in section and commonly attributed to "kaolinization" is not due to the formation of kaolinite. Muscovite is abundantly developed, and various other minerals help to cause the turbidity, but no trace of kaolin has been found in any fragment of feldspar, however decomposed.
>
> It is very probable, however, that the secondary muscovite is directly converted into kaolinite. The two minerals as they occur in the china clay are closely similar in form. * * * It is shown that micas in common clays decompose and that the product of decomposition is probably kaolinite.
>
> Atmospheric weathering is the cause of the "kaolinization" of the granite. Secondary muscovite represents the first stage of the process and is a normal product of the weathering of feldspars. * * * "Kaolinization" has taken place subsequently to the "tourmalinization" and is in no way comparable with that process.

In a few localities kaolin seems to have been formed by the action of waters impregnated with carbon dioxide. According to Gagel and Stremme[97] the granite at Karlsbad, Czechoslovakia, is altered with the production of kaolin to a depth of 40 or 50 meters around the vent of a spring containing carbon dioxide. Buddington[98] has described the occurrence of calcite and kaolinite in miarolitic cavities in a rhyolite of southeastern Alaska.

It seems quite certain that in some occurrences kaolin is formed through the action of sulphate waters that have resulted from the oxidation of sulphides. In discussing the ore deposits at Tintic, Utah, Lindgren and Loughlin[99] say:

> The field relations show with considerable certainty that limestone is replaced by kaolin and limonite and that possibly in part kaolin is replaced by limonite. A thin section of an altered limestone near the iron ore on the 400-foot level of the Dragon iron mine shows clear evidence of the replacement of calcite by kaolin.

Parsons[1] has reported kaolin formed by the action of sulphuric acid at Michipicoten, and Jaquet[2] at Broken Hill.

A review of the theories that have been advanced to explain the origin of kaolin shows that at least five modes of formation have been proposed. These are (1) weathering, (2) the action of hydrothermal solutions, (3) the effect of thermal springs bearing carbon dioxide, (4) the oxidation of sulphides into sulphuric acid and its reaction with aluminous rocks, and (5) the action of gases, either volcanic or "pneumatolytic", on large masses of country rock. All these modes of formation seem to have been verified for certain kaolin deposits except perhaps the last, which postulates the extensive action of gases. As kaolin is now known to include three distinct minerals it becomes essential to correlate the different mineral species with particular modes of formation so far as that is possible.

ORIGIN OF KAOLINITE

Feldspars constitute the chief source of kaolinite, but in a few places kaolinite or anauxite has been formed by the alteration of other minerals. At Bilin, Czechoslovakia, anauxite has been derived from augite crystals whose crystal form is still recognizable even after complete alteration[3] and also similarly from

[87] Barnitzke, J. E., Über das Vorkommen der Porzellanerde bei Meissen und Halle a. S.: Zeitschr. prakt. Geologie, vol. 17, pp. 457–473, 1909.

[88] Wüst, R., Die Entstehung der Kaolinerden der Gagend von Halle a. S: Idem, vol. 15, pp. 19–23, 1907.

[89] Winkel, H. E., Kaolinslemmeriet Rabekkegaard paa Bornholm, Denmark: Kjobenhavn. Tekn. Forenings Tidskrift, 1885.

[90] Vogt, J. H. L., The genesis of ore deposits, pp. 661–665, 1902.

[91] Rösler, H., Beiträge zur Kenntiss einiger Kaolinlagerstätten: Neues Jahrb., Beilage-Band 15, Heft 2, pp. 231–393, 1902.

[92] Ries, Heinrich, Origin of residual kaolins: Am. Ceramic Soc. Trans., vol 13 pp. 66, 67, 1911.

[93] Davison, E. H., The geology and economics of the west of England china-clay deposits: British Assoc. Adv. Sci. Rept. 96th meeting, pp. 556–557, 1928.

[94] Selle, V., Die Kaoline des thüringischen Buntsandstein: Deutsche geol. Gesell. Zeitschr., vol. 28, p. 110, 1876.

[95] Hickling, G., China clay, its nature and origin: Inst. Min. Eng. [England] Trans., vol. 36, pp. 10–32, 1908.

[96] Galpin, S. L., Studies of flint clays and their associates: Am. Ceramic Soc. Jour., vol. 14, p. 307, 1912.

[97] Gagel, C., and Stremme, H., Über einen Fall von Kaolinbildung im Granit durch einen Kalten Säuerling: Centralbl. Mineralogie, 1909, pp. 427–437, 467–475.

[98] Buddington, A. F., An association of kaolinite with miarolitic structure: Jour. Geology, vol. 31, pp. 149–151, 1923.

[99] Lindgren, Waldemar, and Loughlin, G. F., Geology and ore deposits of the Tintic mining district, Utah: U. S. Geol. Survey Prof. Paper 107, p. 262, 1919.

[1] Parsons, A. L., The productive area of the Michipicoten iron ranges: Ontario Bur. Mines Ann. Rept., vol. 24, pt. 1, p. 192, 1915.

[2] Jaquet, J. B., Geology of the Broken Hill lode, New South Wales, p. 89, 1894.

[3] Smirnoff, N. P., Über ein krystallinisches Verwitterungsproduct des Augits: Zeitschr. Kryst. Min., vol. 43, pp. 338–346, 1907.

biotite. The kaolinite described on page 162 from the Abatik River, northern Alaska, is an alteration product of biotite. Biotite phenocrysts in a bentonite derived from volcanic ash from Lone Mountain, Tennessee,[4] and Cherokee County, Tex., have been partly altered to kaolinite, as identified by Ross.[5] Vogt [6] states that kaolinite has been derived from hornblende, augite, beryl, topaz, etc., but the grounds for this statement are not given

WEATHERING PROCESSES

It seems evident that most kaolinite, including the greater part if not all of the kaolinite that forms commercial kaolin deposits, has been formed by weathering processes. They are of two types—residual kaolins, formed by the weathering in place of feldspathic rocks, and sedimentary kaolins that are transported and redeposited kaolinitic material. Residual kaolins of commercial value are most commonly the result of the profound alteration (of the katamorphic type) of pegmatites and less commonly of granitic or syenitic rocks low in ferromagnesian minerals. Feldspathic rocks that have undergone profound kaolinitic weathering are probably the dominant source of the material that after erosion, transportation, and redeposition forms sedimentary kaolin deposits. Some impure kaolinitic beds, especially those associated with shales that are far removed from feldspathic source rocks, probably have a different origin. Very profound leaching seems to result in the kaolinization of clay material that under more normal conditions would form ordinary shales. That is, the less profound weathering conditions seem very commonly to produce beidellitelike material, but where the weathering and leaching are very profound kaolinitic material may be produced. Ries[7] has given a good review of the evidence that the commercial kaolins of the United States have formed in this way, and it has been shown that the kaolins of Halle and Meissen, Germany, and certain kaolins of Denmark and Norway have been similarly formed. Pneumatolytic processes have been advocated for the kaolins of Cornwall,[8] but the recent work of Hickling seems to indicate that even the Cornwall deposits were formed by weathering. The evidence for the formation of these deposits by solutions of magmatic origin is discussed in greater detail on page 174.

[4] Nelson, W. A., Volcanic ash bed in the Ordovician of Tennessee, Kentucky, and Alabama: Geol. Soc. America Bull., vol. 33, pp. 605–616, 1922.

[5] Ross, C. S., Altered Paleozoic materials and their recognition: Am. Assoc. Petroleum Geologists Bull., vol. 12, p. 151, 1928.

[6] Vogt, J. H. L., Problems in the geology of ore deposits: Am. Inst. Min. Eng. Trans., vol. 31, p. 150, 1897.

[7] Ries, Heinrich, Origin of white residual kaolins: Am. Ceramic Soc. Trans., vol. 13, pp. 51–74, 1911.

[8] Collins, J. H., On the nature and origin of clays: Mineralog. Mag., vol. 7, pp. 205–214, 1887. Daubrée, A., Études synthétiques de géologie expérimentale, pp. 64–65, 1879.

INTERMEDIATE WEATHERING PRODUCTS

Hickling believes that feldspar does not alter directly to kaolin but that muscovite is an intermediate product. Stelle,[9] who has studied the kaolin deposits of Halle, which are derived from quartz porphyry, traces them to ordinary weathering but believes that sericite is an intermediate product. Galpin has confirmed the presence of a mineral resembling mica in certain flint clays. A micalike material with a birefringence nearly the same as that of muscovite is present in many kaolin deposits and is especially abundant in the kaolin near Franklin, N. C. This micalike material is evidently the mineral called muscovite by Hickling and sericite by Stelle and Rösling. It forms vermicular crystals like those of kaolinite in the same deposit, and many of these groups are partly altered to kaolinite. The indices of refraction are a little higher than those of kaolinite but much lower than those of muscovite. The optical properties of this substance, which may be called the "muscovitelike kaolin mineral," are as follows: Optical character negative; $2V=60° \pm 5°$; dispersion $\rho>v$; $\alpha=1.557$, $\beta=1.569$, $\gamma=1.573$; birefringence 0.014.

A sample of the muscovitelike mineral was separated from kaolinite with a lower specific gravity, described in the tables on pages 162–163, and from muscovite and quartz with a higher specific gravity, by heavy solutions in a centrifuge. Unfortunately many of the grains were partly altered to kaolinite, so the sample contained only about 60 per cent of the muscovitelike mineral proper and 40 per cent of kaolinite.

The chemical analyses of two samples of material from the kaolin at Franklin, N. C., are as follows:

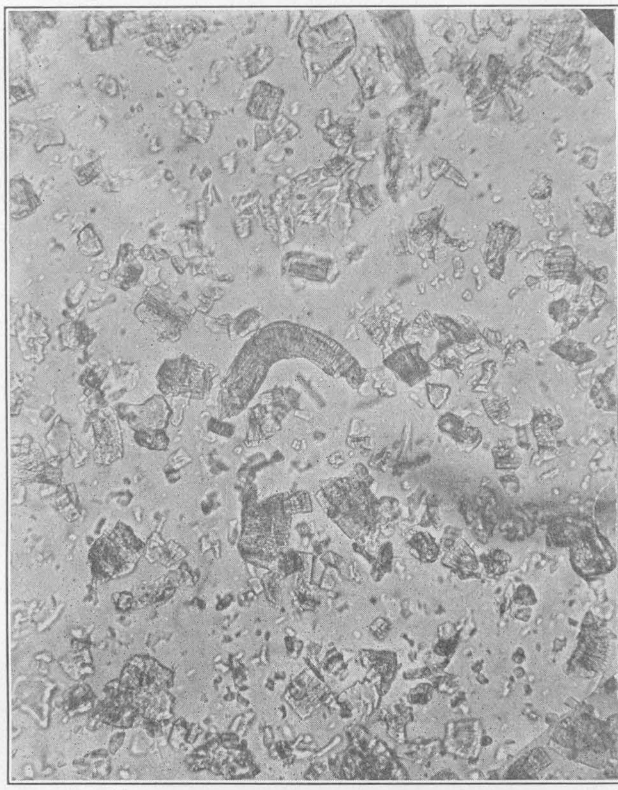
Analyses of Franklin kaolin minerals

[By F. A. Gonyer]

	Kaolinite	Muscovite-like mineral, partly altered
SiO_2	44. 26	44. 17
Al_2O_3	40. 22	40. 83
Fe_2O_3	. 30	Trace.
MgO	. 18	. 17
CaO	. 32	Trace.
K_2O		. 36
Na_2O		. 22
H_2O-	. 64	3. 96
H_2O+	14. 16	9. 90
	100. 08	99. 61

The kaolinite retains about 14 per cent of water at 105°; the muscovitelike mineral, about 40 per cent altered to kaolinite, retains 9.90 per cent of water at 110°. A simple calculation will show the amount of water that would be retained by the pure micaceous

[9] Stelle, V., Zeitschr. Naturwiss., vol. 79, pp. 321–323, 1907.



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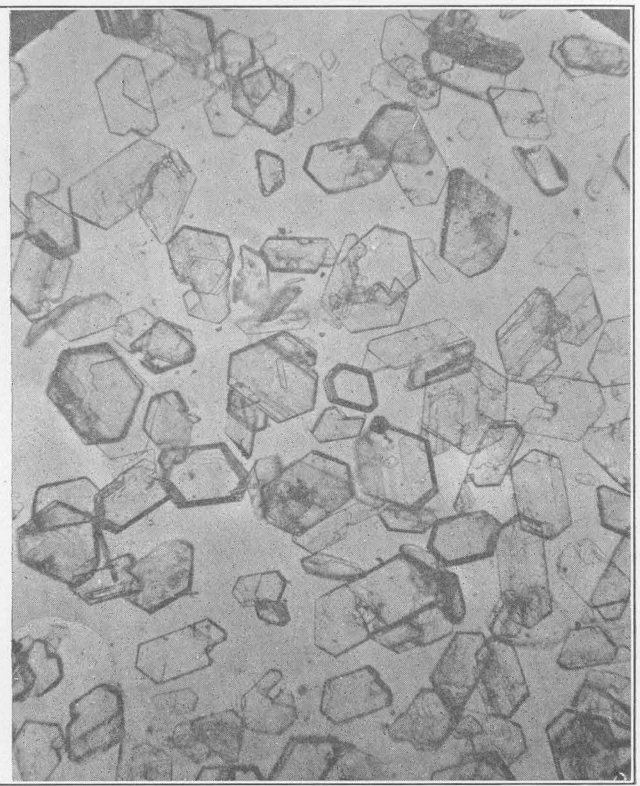

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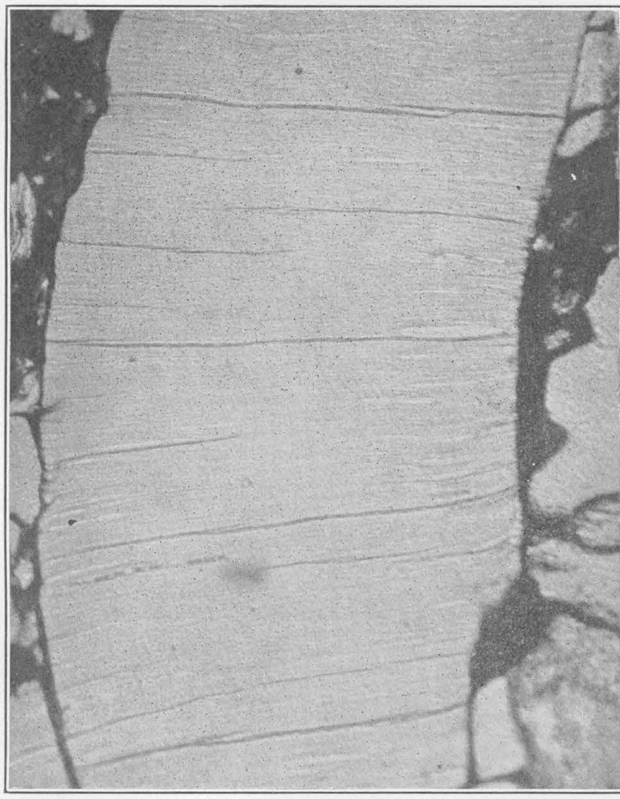

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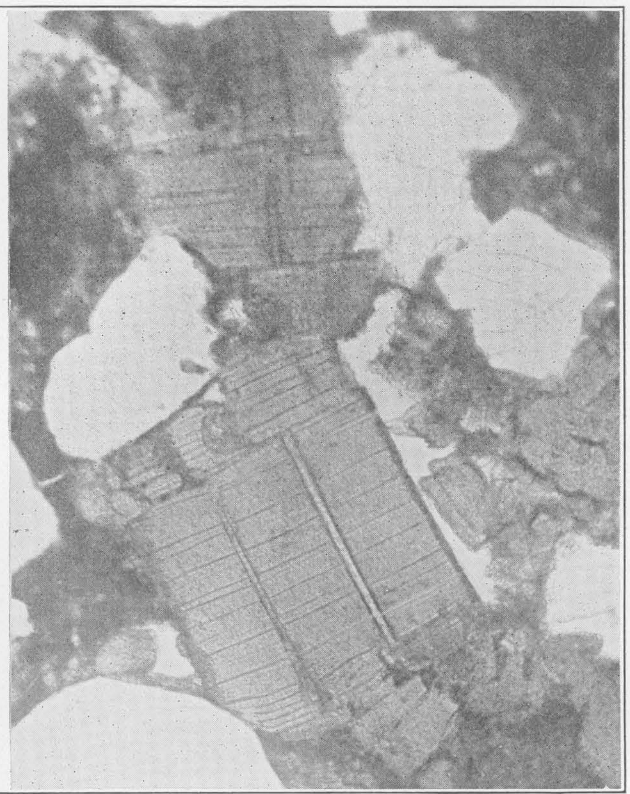
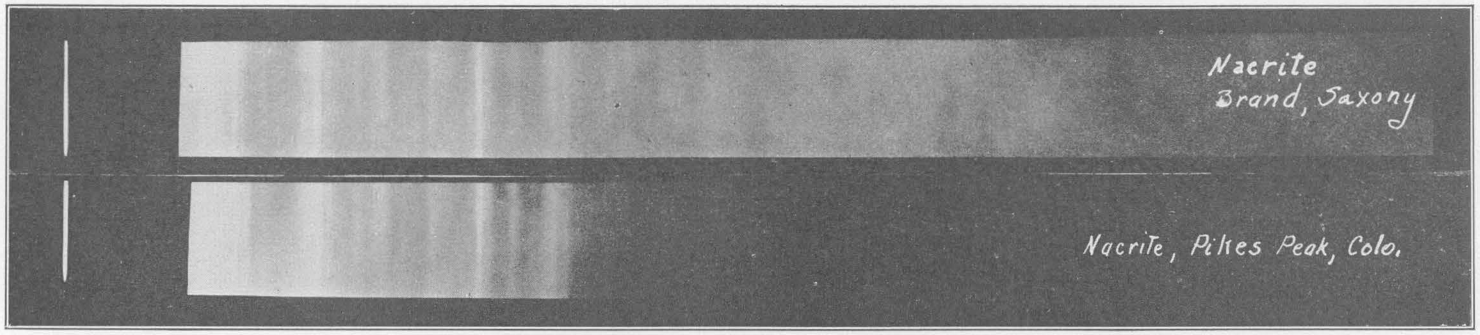


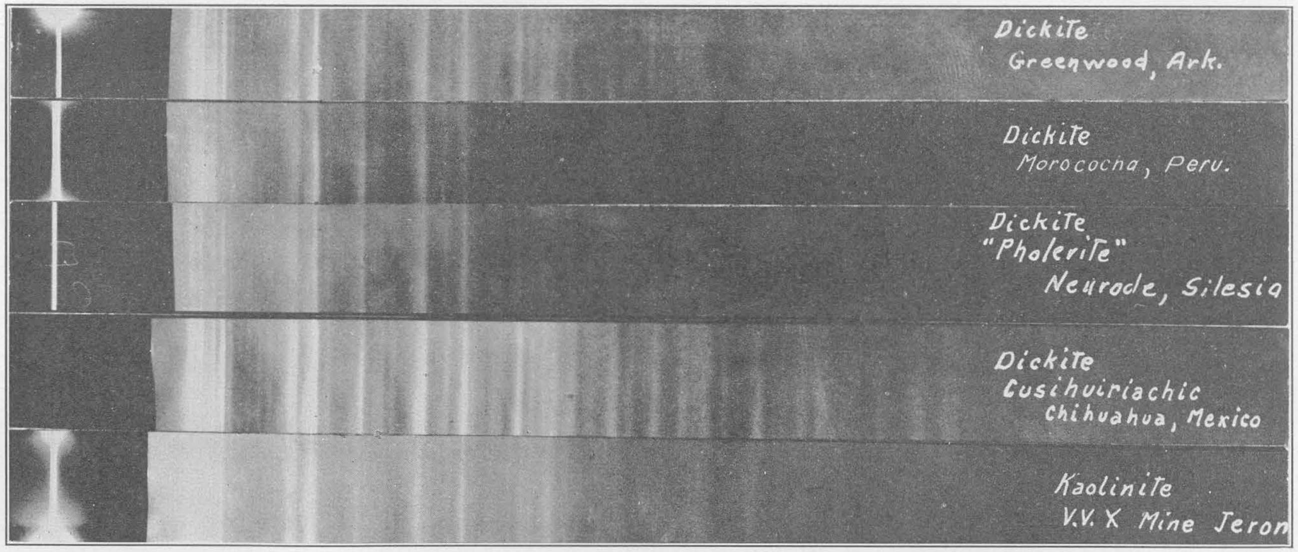
D

CRYSTALS OF KAOLINITE AND DICKITE

A. Small vermicular crystals of kaolinite, typical of those in many kaolin deposits. From Franklin, N. C. Magnification 350 diameters. *B*. Dickite crystals from National Belle mine, Red Mountain, Colo. Near the center of the figure are several crystals that show perfect pyramidal faces. Magnification 260 diameters. *C*. Kaolinite crystals (leverrierite) from Quartier Gillard near St.-Etienne, France. A secondary mineral has formed small lenslike areas in the cleavage cracks. Magnification 85 diameters. *D*. Kaolinite crystals associated with quartz grains in arkosic sandstone of the Ione formation, Mokelumne River, Calif. Kaolinite crystals show very perfect cleavage (approximately horizontal) and a parting perpendicular to the cleavage. At the left of the largest kaolinite crystal quartz has been partly replaced by kaolinite. Magnification 85 diameters.



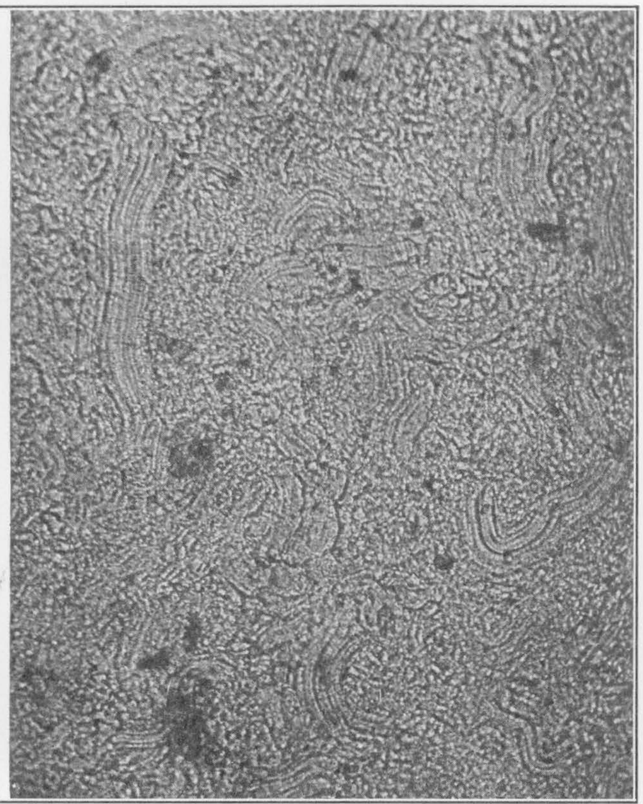
A. X-RAY DIFFRACTION PATTERNS OF NACRITE FROM PIKES PEAK, COLO., AND BRAND, SAXONY



B. X-RAY DIFFRACTION PATTERNS OF DICKITE FROM GREENWOOD, ARK.; MOROCOCHA, PERU; NEURODE, SILESIA; AND CUSIHUIRIACHIC, MEXICO, COMPARED WITH KAOLINITE FROM JEROME, ARIZ.

The dickite patterns agree with each other but differ from kaolinite.

A



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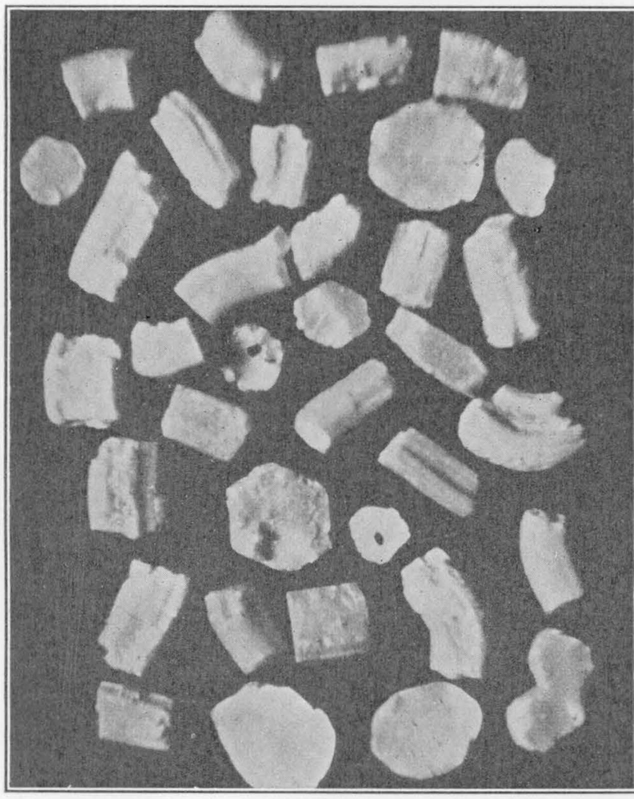

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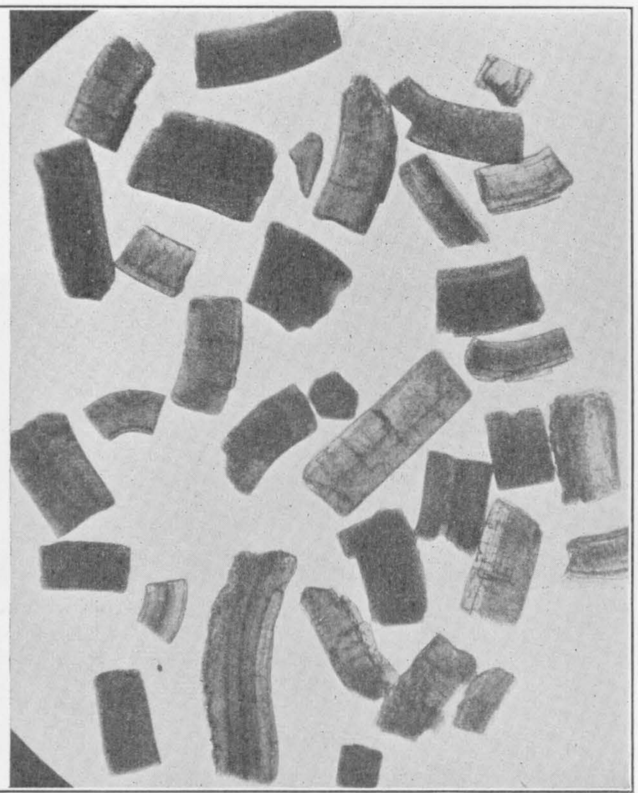
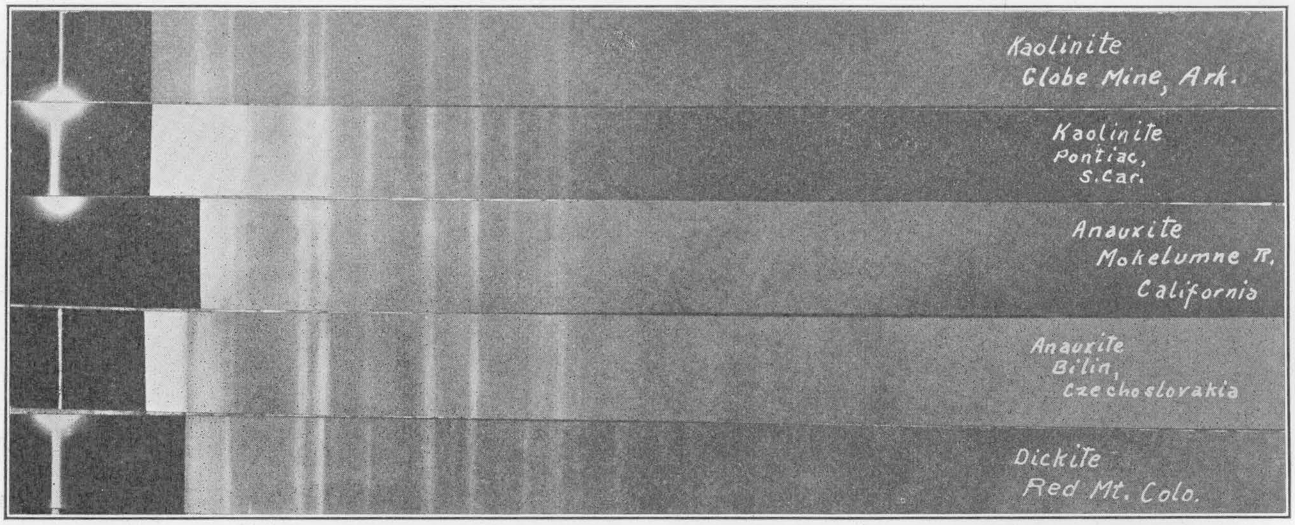


D

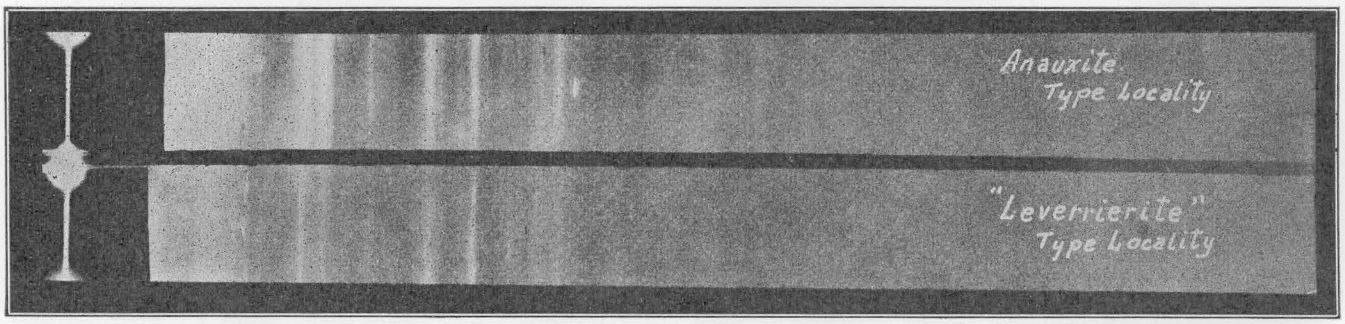
CRYSTALS OF ANAUXITE AND KAOLINITE

A. Group of crystals of anauxite developed from an augite crystal from Bilin, Czechoslovakia. Magnification 85 diameters. *B*. Group of kaolinite crystals from kaolinitic arkose, Louisville, Miss. Magnification 85 diameters. *C*. Crystals of kaolinite from Ione formation, 1 mile south of Ione, Calif. Two crystals near the middle of the figure show the cavities that resemble negative crystals; one kaolinite crystal near the lower middle shows hexagonal outline. Several of those lying on the side show characteristic curvature. Reflected light. Magnification 10 diameters. *D*. Crystals of kaolinite from locality near Pontiac, S. C. Crossed nicols. Magnification 28 diameters.



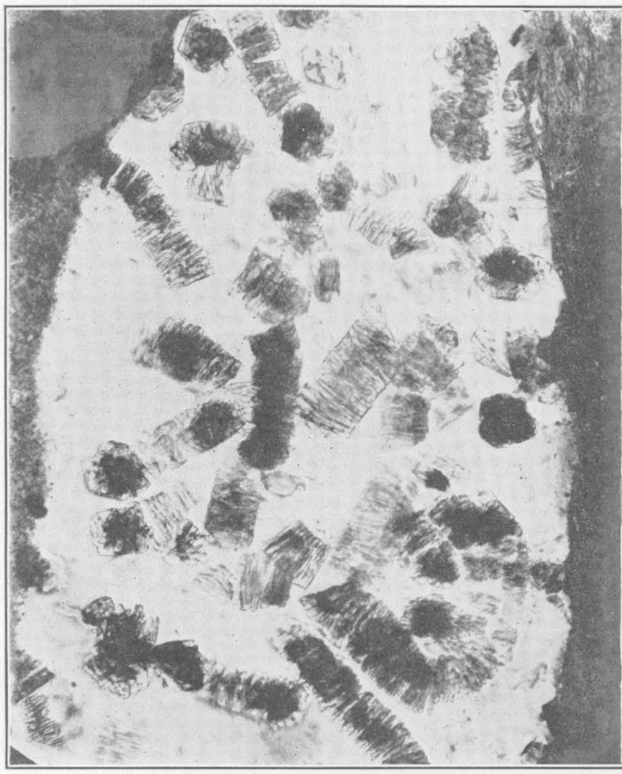
A. X-RAY DIFFRACTION PATTERNS OF KAOLINITE-ANAUXITE FROM GLOBE, ARIZ.; PONTIAC, S. C.; MOKELUMNE RIVER, CALIF.; AND BILIN, CZECHOSLOVAKIA, COMPARED WITH DICKITE FROM RED MOUNTAIN, COLO.

The kaolinite-anauxite patterns agree with each other but are different from dickite.



B. X-RAY DIFFRACTION PATTERNS OF ANAUXITE FROM THE TYPE LOCALITY (BILIN, CZECHOSLOVAKIA) AND "LEVERRIERITE" FROM THE TYPE LOCALITY (ST.-ETIENNE, FRANCE)

Aside from one or two faint lines due to inseparable mineral impurities the two are the same.



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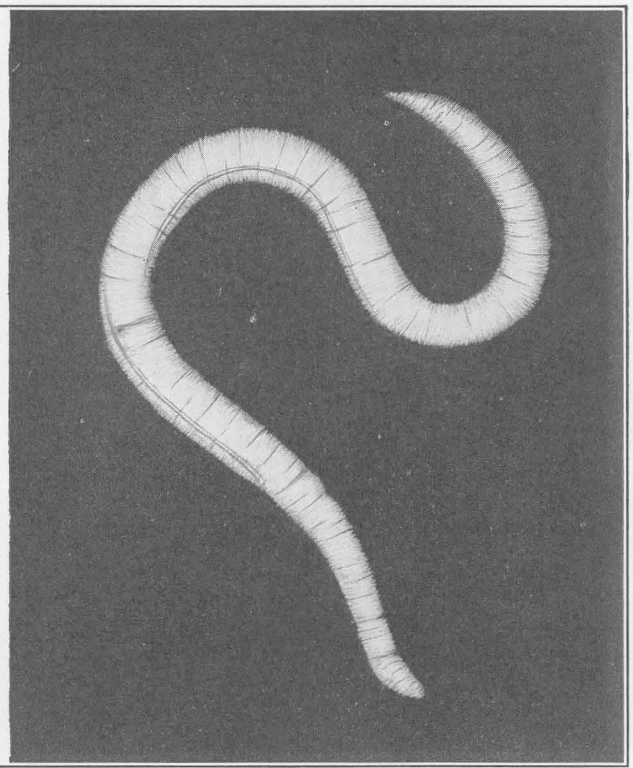

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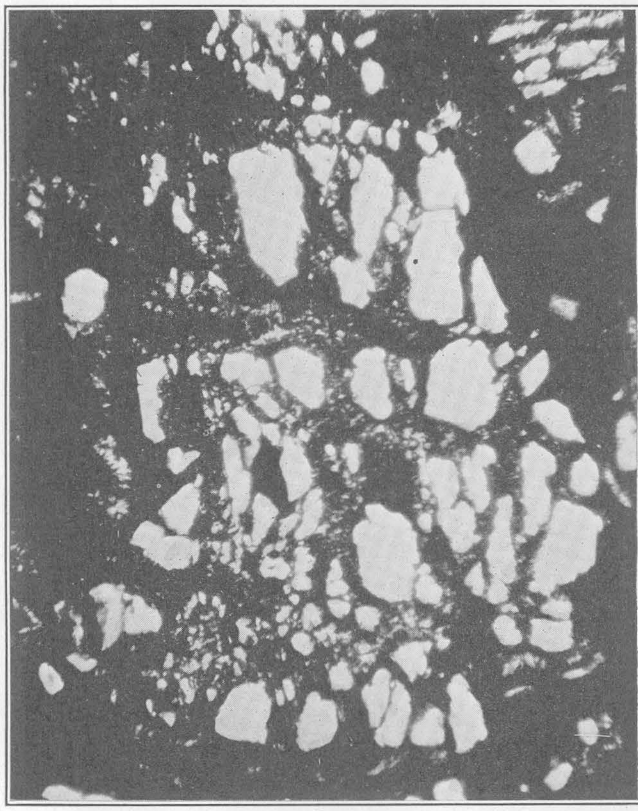

C

D

A, *B*. CRYSTALS OF KAOLINITE. *C*, *D*. EXAMPLES OF REPLACEMENT OF QUARTZ BY KAOLINITE

A. Small vermicular crystals of kaolinite from Calhoun County, Ark., typical of those in many kaolin deposits. Magnification 350 diameters. *B*. A single coiled kaolinite crystal from Mexia, Tex. It is evident that any transportation would have destroyed this fragile form, and the crystal must have developed in place—that is, it must be authigenic. Magnification 260 diameters. *C*. A single quartz crystal that has been largely replaced by clay material. The uniform illumination of the large white areas under crossed nicols shows that they are all parts of what was originally a single quartz crystal. The very small white areas are kaolinite. Magnification 85 diameters. *D*. Arkosic sandstone from Louisville, Miss., with deeply embayed areas that are the result of replacement of quartz by kaolinite. Magnification 85 diameters.

mineral, and this proves to be about 7.2 per cent. As the kaolinite and the mixed kaolinite and muscovitelike mineral contain practically the same proportions of SiO_2 and Al_2O_3, the muscovitelike mineral in pure form must have the same ratio of these constituents. The essential chemical composition of the pure muscovitelike mineral, recalculated to 100 per cent, and the molecular proportions are approximately as follows:

Chemical composition and molecular ratios of muscovitelike kaolin mineral

	Chemical composition	Molecular proportion	Molecular ratios
SiO_2	47.9	0.79	184
Al_2O_3	44.3	.43	100
High-temperature H_2O	7.8	.43	100
	100.00		

The above recalculated analysis shows that the so-called "muscovite" of Hickling is not in fact muscovite, for it contains very little K_2O, but is a clay that has within the limits of analytical error the same proportion of SiO_2 and Al_2O_3 as kaolinite. The high-temperature or essential H_2O is lower than in kaolinite, and calculations indicate that it is very close to 1 instead of 2 molecules, and so the indicated composition is $H_2O.Al_2O_3.2SiO_2$ or $HAlSiO_4$. This mineral will be further studied when pure material can be obtained.

It is therefore evident that the muscovitelike mineral that has been called secondary muscovite is not a mica at all but a clay mineral. Chemical analyses of partly purified material suggest that it is similar to kaolinite in composition but contains less water.

AUTHIGENIC CRYSTALLIZATION OF KAOLINITE

The kaolinite in some of the kaolinitic sands and arkoses presents certain very interesting relations. The large crystals derived from several such sands have been described on pages 162–163. These are best represented by material from the Ione sands of California, but good material of the same type has been isolated from Mexia, Tex., Pontiac, S. C., Bauxite, Ark., and Pike County, Ark.; and the "leverrierite" of France proves to have the same characteristics. Photomicrographs of these crystals are shown in Plates 39, 41, and 43, an unusually good one being represented in Plate 43, *B*.

Allen [10] believes that the anauxite of the Ione formation of California is the result of the alteration of biotite first to chlorite and then to anauxite. Some of the anauxite of the Ione formation is no doubt formed in this manner, but it is evident that not all of it could have been so formed. The kaolinite from a sample obtained 1 mile south of Ione, Amador County (see p. 161), contains grains that range from 0.003 to 2.5 millimeters in diameter. A large proportion of these are three to eight times as long in the direction of the *c* axis as at right angles to it, and very commonly they are curved or twisted. The elongation is just as conspicuous in the small grains a fraction of a millimeter in diameter as in the larger ones. It is improbable that crystal grains with a perfect cleavage, as in biotite or chlorite, could be weathered out of a preexisting rock, transported, and then redeposited and still retain forms like those shown in the photomicrographs. Biotite crystals of this smallest size and form are not known in igneous rocks, and furthermore, a rock or group of rocks that would yield crystals which were dominantly 0.02 to 0.004 millimeter across but reached 2.5 millimeters would be very unusual. The cross section of the anauxite or kaolinite crystals is unlike that of most biotite, as the crystals are commonly rounded and embayed.

Crystals of the same type are present in many sedimentary deposits, and those from Mexia, Tex., and Pontiac, S. C., are similar, yet in none of these deposits do they resemble biotite crystals in form. The crystal from Mexia illustrated in Plate 43, *B*, forms a perfect letter S, and it is evident that abundant fragile crystals of an easily cleavable mineral could not be transported and retain such forms. Kaolinite crystals from the Globe bauxite mine, Bauxite, Ark., are abundant in small chimneylike bodies that cut the bauxite. The original syenite from which the bauxite and kaolinite were derived contains no biotite crystals. The groups of kaolinite crystals from Calhoun County, Ark., shown in Plate 43, *A*, have formed within a single quartz crystal and so give definite evidence of authigenic crystallization. Small curved and elongated kaolinite crystals are common in kaolinized pegmatites that originally contained no biotite.

It seems evident that most of the kaolinite crystals with vermicular habit must have developed in place. This may have occurred as the feldspar or other aluminous source material of the kaolinite was being altered, or through recrystallization of amorphous or submicroscopically crystalline kaolinitic material after its transportation and redeposition, and probably it has developed in both these ways. Cayeux [11] has found that kaolinite ("leverrierite"), which is abundant in the sedimentary rocks of France, has formed in place. It is thus evident that the development of kaolinite and anauxite crystals in place is a very widespread phenomenon, seeming, in fact, to be the normal mode of formation of such crystals. Some of the kaolins of Germany and Japan that have been examined contain kaolinite crystals 1 millimeter or more

[10] Allen, V. T., Anauxite from the Ione formation, California: Am. Mineralogist, vol. 13, pp. 148-151, 1928.

[11] Cayeux, Lucien, Introduction à l'étude pétrographique des roches sédimentaires, pp. 230-232, 1916.

in diameter, but it is not evident whether these are kaolinized biotite or kaolinite that has crystallized in place.

REPLACEMENT OF QUARTZ BY KAOLINITE

Plate 43, *C* and *D*, shows a relation between kaolin and quartz which indicates that quartz is being replaced by kaolinite. Deep embayments in quartz from Louisville, Miss., are filled with kaolinite, and it seems evident that these were formed subsequent to deposition. The clay from Camden, Ala., contains quartz grains that are divided into many small segments by clay veinlets. The quartz segments remain in perfect optical orientation and indicate that the quartz has been replaced and that fragments of fractured quartz have not been pushed aside by clay material. These relations indicate that alumina has been transported in some way so that it could react with the quartz to form a clay mineral. The mode of transportation is not evident. Oxidizing pyrite may set free sulphuric acid, which reacts with aluminous minerals to form soluble aluminum sulphate. Bacteria may produce sulphates, and their action has been suggested by Logan [12] as the means of the formation of the kaolinitic rock indianaite; but this explanation is not accepted by Ries,[13] who believes that the oxidation of pyrite has formed sulphuric acid that dissolved alumina from the shale, with the production of aluminum sulphate, which attacked the quartz to form indianaite. This explains the isolated blocks of sandstone in the clay, the veins of clay in sandstone, and evidences of corrosion and replacement of sandstone by clay.

The arkosic sands of the Gulf coastal region were probably highly oxidized before deposition, and the beds, so far as known, contain little pyrite. For these reasons a source of sulphuric acid to form adequate amounts of aluminum sulphate is not evident, and the mode of transportation of alumina now present in the kaolinite of certain arkosic sands of the Gulf Coastal Plain is not clear.

FORMATION BY SULPHATE SOLUTIONS

The formation of certain deposits of kaolin (kaolinite) through the action of sulphate waters is well established and is possibly the dominant mode of formation of kaolin associated with mineral deposits. The kaolin from the United Verde Extension mine, Jerome, Ariz., was formed by the action of sulphate waters and adds another example of that mode of formation to those cited from the literature on pages 170–171. Ransome [14] says, "The mineral at Jerome is certainly due to the action of acid solutions from oxidizing sulphides on rhyolitic porphyry."

FORMATION BY HYDROTHERMAL AND "PNEUMATOLYTIC" PROCESSES

The geologic evidence seems to indicate that there are no commercial kaolin deposits in the United States in which there is clear evidence of the formation of kaolinite by hydrothermal or "penumatolytic" processes. The china-clay deposits of Cornwall have probably more often been ascribed to such processes than any others, but even there the evidence is not clear and different workers do not agree.

The origin of the china clay of western England by the action of solutions of magmatic origin is advocated by Davison,[15] but some of the arguments he advances are without real significance. The association of china clay with mineralized zones and the presence of tourmaline and other undoubted hydrothermal or "pneumatolytic" minerals have been cited as evidence, but it is just such zones that would be most permeable to downward-moving waters, and so such association can not be accepted as definite proof of hydrothermal or "pneumatolytic" origin. There is no probability that kaolinite ever forms under the high-temperature conditions that are characteristic of tourmaline.

The Cornwall china clays show certain distinctive physical properties, but their kaolin mineral is kaolinite, just as in deposits that are clearly formed by weathering. That is, they contain kaolinite, which is stable at the lowest temperature of the three kaolin minerals, and not dickite or nacrite, which are formed at the higher temperatures. It seems probable that if the Cornwall deposits contain hypogene kaolin it should be represented by the minerals with the higher temperature stability—that is, dickite or nacrite. Although the hydrothermal or "pneumatolytic" origin of the Cornwall kaolin can not be entirely rejected, it is certainly not fully established.

ORIGIN OF DICKITE

Dickite is the mineral present in most of the occurrences of kaolin that have been reported to be of hydrothermal origin. The dickite on the island of Anglesey was found in a tunnel at sea level that had been driven nearly 500 feet into fresh rock. It occurred in small cavities in a quartz vein in hard slate and was associated with fresh copper sulphide. The position and association indicate that this occurrence of dickite could not have been the result of weathering and suggests very strongly that it was produced by hypogene mineralizing solutions. The dickite from Red Mountain, Colorado, has also been reported to have been formed by the action of hypogene solutions. The mode of formation of the dickite from Neurode,

[12] Logan, W. N., Kaolin of Indiana: Indiana Dept. Conservation Div. Geology Pub. 6, pp. 35–76, 1919; Science News Service, vol. 49, p. 197, 1919.

[13] Ries, Heinrich, High-grade clays of the eastern United States: U. S. Geol. Survey Bull. 708, pp. 161–162, 1922.

[14] Personal communication.

[15] Davison, E. H., op. cit.

Saxony, and Cusihuiriachic, Chihuahua, Mexico, is not known. The specimens of dickite from Greenwood, Ark., and Williams, Okla., came from a region of folded sedimentary rocks of the Ozark Mountains. The material has not been studied in the field, and the mode of origin is unknown, but in this region the action of hypogene solutions can not be excluded. The dickite filling the cavities in a geode from the vicinity of Keokuk, Iowa, can hardly have been deposited by hot waters, but on the other hand such an occurrence could not be the result of weathering. It seems evident that dickite is commonly formed by hydrothermal solutions and occasionally by cool solutions, although nowhere, so far as known, is evidence presented that clearly shows its origin through processes of weathering.

The dehydration curves given in Figure 10 show clearly that dickite is stable at distinctly higher temperatures than kaolinite. Deposition by hot solutions is one method, though probably not the only method, by which a mineral with this higher temperature stability can be formed. Lindgren [16] says truly that kaolin [including dickite] could not have been formed at or near the surface of the earth at a temperature of 470°, the temperature at which it breaks down, but this does not preclude its formation by solutions with a moderately high temperature.

ORIGIN OF NACRITE

The association of nacrite with fresh sulphides in the material from Brand, Saxony, and with cryolite and mica that have resulted from the hydrothermal alteration of microcline at St. Peters Dome, Colo., indicates that these occurrences are the results of hydrothermal or gas-phase ("pneumatolytic") processes. The two dehydration curves given in Figure 10 do not show close agreement as to the temperature at which these samples begin to lose essential water, but both lose water more slowly than the other kaolin minerals and retain part of their water to the unusually high temperature of 675°.

This confirms the evidence of the mineral relations, which indicates that these samples of nacrite were formed by hydrothermal or gas-phase reactions, but the two occurrences of nacrite known do not give sufficient evidence for generalization on the genetic origin of this mineral.

SYNONYMY

In the past it has been very difficult or impossible to determine fully the optical properties of the clay minerals, and the impurity of most samples has caused wide variation in the results of chemical analyses. The results of different studies of clay material thus could not be correlated with one another, so that there is probably no group of minerals in which a greater number of names have been proposed than among the clays. Under these circumstances it is evident that there must be many synonyms. The task of determining these synonyms has been great, but Dana [17] accomplished it surprisingly well, considering the criteria available.

[16] Lindgren, Waldemar, The origin of kaolin: Econ. Geology, vol. 10, p. 91, 1915.

The newer mineralogic methods and especially the use of X rays for the study of fine-grained crystalline powders, make possible a more rigorous determination of identities than has been possible in the past. The following lists show the names that are believed to be synonymous so far as it is possible to ascertain without access to all the type materials. The lists are restricted to the minerals considered in this paper and do not include the names synonymous with halloysite or members of the montmorillonite-beidellite groups, which will be discussed in future papers.

Names synonymous with kaolinite and anauxite

Leverrierite= kaolinite.

Pholerite= most commonly either impure kaolinite or a member of the kaolinite-anauxite series with approximately a 2:3 alumina ratio.

Carnat= kaolinite.

"Newtonite" as analyzed= mixture of kaolinite crystals and halloysite.

"Newtonite" as determined optically= alunite.

Steinmark gives a kaolinite X-ray pattern.

Microvermiculite= kaolinite.

Talkerde, Erdiger Talk, Schuppiger Thon, Pelitische Felsittuffe = kaolinite for the most part.

Lithomarge= crystalline kaolinite and halloysite, at least in part.

Indianaite= a rock composed of halloysite and a little crystalline kaolinite.[18]

Porzellanthon, Porcellana, Terra porcellana, Argiler à porcellaine, Marge porcellana = kaolinite and halloysite for the most part.

Terra Samia, Hunterite, Collyrium, Arcilla, Tuesite = probably kaolin for the most part.

The following names seem to have been used for all three minerals of kaolin:

Pholerite = kaolinite for the most part, nacrite in part, dickite rarely.

Kaolinite = kaolinite in general, dickite and nacrite less commonly.

Kaolin = kaolinite in general, dickite and nacrite rarely.

CONCLUSIONS

Kaolin has generally been assumed to be characterized by a single clay mineral but comprises in fact at least three distinct minerals. These are:

1. A mineral from Brand, Saxony, which has previously been called nacrite and for which it seems best

[17] Dana, E. S., System of mineralogy, 6th ed., pp. 685-889, 1892.

[18] The indianaite examined contained only halloysite and kaolinite. See Ries, Heinrich, U. S. Geol. Survey Bull. 708, pp. 156-157, 1922.

to retain this name and raise it to specific rank. Composition $2H_2O.Al_2O_3.2SiO_2$.

2. A mineral first described from the island of Anglesey, which seems never to have received a distinctive name and it is proposed to call dickite, after Allan B. Dick, who described it in accurate detail. Composition $2H_2O.Al_2O_3.2SiO_2$.

3. A mineral that is characteristic of nearly all kaolin deposits, to which the name kaolinite is restricted. Composition $2H_2O.Al_2O_3.2 \pm SiO_2$.

These three minerals have distinguishing optical and physical properties and give characteristic X-ray diffraction patterns. Dehydration tests indicate that kaolinite loses the largest part of its water content at about 450° C., dickite at about 575° C., and nacrite at about 650° C.

It is probable that the three minerals originate in dissimilar ways. Kaolinite and anauxite are produced by surface weathering and by the action of sulphate or carbonate waters. Dickite is most commonly of hydrothermal origin and is not a product of weathering. Nacrite is probably formed by "pneumatolytic" or hydrothermal processes.

It has been asserted that muscovite is an intermediate product in the weathering of feldspar to kaolin, but it has been found that this intermediate material is not muscovite but a clay that is probably similar to kaolinite but with one instead of two molecules of essential water.

This investigation of the clay minerals has shown that these, among the most difficult of minerals, can be studied by the methods now available, and the way in which these methods have supplemented one another has been most satisfactory. This is especially true of the X ray and optical measurements, which seem to be the most potent present-day methods of mineralogic research.

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