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GEOLOGY OF THE McCALLS FERRY-QUARRYVILLE DISTRICT PENNSYLVANIA

BY

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AND

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OUTLINE OF THE REPORT

The McCalls Ferry-Quarryville district, as the term is used in this report, includes the McCalls Ferry and Quarryville quadrangles, in southeastern Pennsylvania. It comprises parts of Lancaster, Chester, and York Counties and lies within a belt of metamorphosed pre-Cambrian and Paleozoic rocks that extends southwestward from Trenton, N. J., through eastern Pennsylvania and Maryland into the Southern States of the eastern Atlantic slope.

Topographically the area consists of a well-dissected upland and a gently rolling lowland. The country is fertile and is well adapted to agriculture, which is the most important industry. The mining importance belongs to the last century, when the district furnished a considerable amount of iron ore and for several years yielded, from the Gap nickel mine, about one-sixth of the world's annual output of nickel.

The geologic interest of the district centers in the nature and origin of its highly metamorphosed rocks and in the bearing of its geologic history on the interpretation of the stratigraphy and structure of the Piedmont province as a whole. Study of this district has furnished a clue for the solution of puzzling stratigraphic problems in the region on the northeast, but the interpretation of the stratigraphy and structure of the pre-Cambrian rocks in this area itself is dependent on previous work by the writers in Maryland, where the base of the stratigraphic column was established. Therefore, in this report the district has been treated in its relation to the geology of the Piedmont province as a whole.

The northeast corner of the Quarryville quadrangle is occupied by the southern spur of a wide anticline underlying the Honeybrook upland, which extends northeastward as far as Schuylkill River. The core of this anticlinal spur, which is known as the Mine Ridge anticline, is made up of the Baltimore gneiss of early pre-Cambrian age. This formation comprises intensely metamorphosed graphitic schists and gneisses with a few calcareous lentils and numerous interlaminated layers of hornblende schist, which are probably of igneous origin. The whole series has been invaded and injected by igneous rocks ranging in composition from granite to gabbro. The invading igneous rocks are presumably pre-Cambrian, because they have not been found to cut the intensely metamorphosed Paleozoic schists and quartzites that flank the uplift. Moreover, the fact that they have been altered by a strong dynamic metamorphism that was apparently contemporaneous with the first metamorphism of the Paleozoic rocks suggests that they were intruded during a period of igneous activity that preceded Cambrian sedimentation.

Lower Cambrian arenaceous strata unconformably overlie the pre-Cambrian core of the anticline and are about 2,200 feet thick. They are overlain by 5,000 feet of calcareous and dolomitic rocks ranging in age from Lower Cambrian to Ordovician. These rocks were gently folded and lifted above sea level early in Ordovician time, and across their eroded edges the argillaceous Conestoga limestone, of probable Chazy age, was laid down. After the deposition of the Conestoga limestone the district was again folded with moderate intensity, and probably during a subsequent deformation a block of the earth's crust that originally lay farther south was driven northward and westward across the folded Paleozoic strata. In the McCalls Ferry and Quarryville quadrangles the overriding rocks comprise the Cockeysville marble, Wissahickon formation, Peters Creek schist, Cardiff conglomerate, and Peach Bottom slate, all formations of the Glenarm series, which was studied by the writers in Maryland, where it was determined to be of later pre-Cambrian age and younger than the Baltimore gneiss, which it overlies unconformably. The total thickness of the Glenarm series is probably between 8,000 and 10,000 feet, although no accurate estimate can be made, for the middle formations have been repeated by close folding.

The basal member of the series is the Setters formation, which is not exposed in the McCalls Ferry and Quarryville quadrangles but which occurs about 5 miles southeast of the district in an anticline between Avondale and Doe Run, where its thickness ranges from a few feet to 1,000 feet. It is composed of mica gneiss and mica schist with an intercalated quartzite member. Overlying the Setters formation in this anticline is the Cockeysville marble, a dolomitic and calcitic marble several hundred feet thick. In the McCalls Ferry quadrangle the Cockeysville marble crops out only in a small valley, where it is overlain by the Wissahickon albite-chlorite schist.

The Wissahickon formation occurs in two metamorphic facies—an oligoclase-mica schist and an albite-chlorite schist. The oligoclase-mica schist, which surrounds several anticlines north and northwest of Baltimore, Md., forms a belt along the southeastern border of the Piedmont province from Baltimore to Trenton, N. J. Lithologically it resembles the Setters formation, and it shows the same intense metamorphism. In the McCalls Ferry-Quarryville district it is largely replaced by the less intensely metamorphosed facies of the Wissahickon, the albite-chlorite schist, which is a heterogeneous formation comprising both feldspathic and chloritic schists with a subordinate amount of quartzite or quartzose gneiss. The difference in mineral constituents between the two facies of the Wissahickon appears to have resulted from a difference in conditions of metamorphism rather than from difference in stratigraphic position or in chemical composition. The thickness of the Wissahickon formation where it has been cut across by Susquehanna River must amount to several thousand feet, even after allowance has been made for the close folding. The two uppermost formations of the Glenarm series, the Cardiff conglomerate and the overlying Peach Bottom slate, aggregate between 1,500 and 2,000 feet in thickness. They have been included with the pre-Cambrian rather than with the Paleozoic section because it has been impossible to prove an unconformity at the base of the conglomerate, whereas the base of the Cambrian shows a distinct angular unconformity.

The overthrust block of Glenarm schists covers a little more than two-thirds of the McCalls Ferry-Quarryville district. After the thrusting the overridden and overlying blocks alike were folded along axes that were approximately parallel to the main structural lines of the preexistent folds. In some places the older structural features emerge from under the thrust block so that the preoverthrust deformation can be recognized. The last pre-Triassic deformation of the region is recorded in a cleavage that cuts across the older schistosity and folding. The folding of the Conestoga limestone, the thrusting of the pre-Cambrian block, and the folding of the thrust plane took place between the end of Middle Ordovician time and the beginning of the Triassic, but it is impossible to fix the dates of the individual deformations with more exactitude. During Triassic time the region was broken by normal faults, which accompanied or followed the intrusion of narrow dikes of diabase in this district.

The post-Triassic history is chronicled in the physiography, which gives evidence of progressive but intermittent uplifts. These successive uplifts interrupted the normal course of erosion, so that ten uncompleted cycles of erosion are recorded in the drainage history of the McCalls Ferry-Quarryville district and the adjoining region. The remnants of the old erosion surfaces that mark these partial cycles have been correlated by a study of present and preexistent drainage channels.

The main problems of the metamorphic history of the Piedmont province are embodied in the crystalline schists of the McCalls Ferry-Quarryville district. The clue to the solution of certain of these problems lies within the district itself, and the answer to some other puzzling questions has come from the surrounding region. The most striking feature of the metamorphism of the Piedmont province as a whole is its extremely variable intensity, which can not be adequately explained by any one cause alone. Some of this variation is accounted for by the determination that the metamorphism of the region occurred at more than one time and that the metamorphic effects produced at different times are to a large extent distinct. Thus the pre-Cambrian basement complex, wherever it is exposed, shows an intense alteration. Such intensity might possibly be expected on the ground that the oldest rocks have been through the metamorphic mill the greatest number of times and therefore have naturally been the most altered. However, the recent metamorphic studies of European geologists have brought out the fact that repeated metamorphism is not necessarily cumulative in its effects but is commonly retrogressive, so that a rock that was originally metamorphosed into a coarsely crystalloblastic gneiss may later become a less highly metamorphic rock—a diaphthoritic or retrometamorphic schist. The oldest pre-Cambrian gneiss shows no diaphthoritic effects and none of the superposed foliation that is so striking a feature of both the later pre-Cambrian and the Paleozoic schists. Therefore, it is probable that the earliest known crystalline schists have remained more or less inert since their formation, doubtless owing to their prolonged deep-seated position, which has preserved them from retrogressive metamorphism.

The earliest pre-Cambrian rocks (the Baltimore gneiss) may therefore be disregarded in the study of the variations in metamorphic intensity. It is evident that the metamorphic intensity of the other crystalline schists of the Piedmont province in Maryland decreases continuously from the vicinity of Baltimore toward the west and northwest, as was pointed out many years ago by Williams. But north and northeast of Baltimore, instead of a continuous decrease in intensity, there is a series of approximately parallel belts of variable intensity, starting with a belt of maximum intensity along the eastern border of the Piedmont province. Toward the northwest this maximum diminishes to a minimum metamorphism, still registered in pre-Cambrian rocks. Still farther northwest the intensity rises to a second maximum in the Paleozoic schists that flank the anticline of Mine Ridge, and north and northwest of this anticline the metamorphism in the Paleozoic phyllites and calcareous rocks once more diminishes in the direction of the Triassic basin. These metamorphic belts run in general parallel to the trend lines of the formations, but they are not genetically connected with the folding, as highly altered rocks occur in the cores of open folds while the rocks on the limbs are in the condition of phyllites or mildly altered chlorite schists. Moreover, the second maximum of intensity, which is well shown in the section along Susquehanna River and northeastward from the river, diminishes not only toward the northwest but also toward the southwest, where it gradually disappears in the decreased metamorphism between Baltimore and the Frederick Valley.

Nor does repetition of the same stratigraphic horizon by folding explain these recurrent maxima and minima of metamorphism, for some of the most highly altered rocks are the autochthonous Paleozoic schists, whereas the metamorphic intensity wanes perceptibly in the overlying pre-Cambrian fault block. Moreover, in the pre-Cambrian Glenarm series the Wissahickon formation, which was presumably metamorphosed under a uniform load, occurs in two metamorphic facies of widely differing intensity. The reason for this difference in metamorphic intensity is not obvious, but it is noteworthy that the more intensely metamorphosed facies is penetrated by numerous large igneous intrusions, which are conspicuously absent from the facies of lower intensity. The inference is that the high metamorphism of the one facies may have been caused by the thermal influence of the igneous invasion.

In the area around the Mine Ridge upland, where pre-Cambrian schists have ridden northwestward over Paleozoic rocks, the metamorphic gradient is abnormal, the maximum metamorphism occurring in the autochthonous Paleozoic strata and diminished intensity in the overthrust pre-Cambrian rocks. Such a metamorphic gradient, which is the reverse of the relation found in many places where pre-Cambrian rocks have been thrust over younger formations, is a strong argument in favor of the belief that the metamorphism was younger than the thrusting. This belief is strengthened by the fact that the somewhat meager evidence that is available regarding the sole of the thrust plane indicates that the thrusting took place under conditions that were favorable to free molecular movement rather than to freedom of molar movement.

At first sight it might appear possible to explain the abnormal and localized intensity of metamorphism in the autochthonous Paleozoic schists around the Mine Ridge upland as static metamorphism, or as dynamic metamorphism under the normal stratigraphic load augmented by the weight of the thrust block. It is perfectly evident that the metamorphism was not produced under the causal influence of the normal stratigraphic load alone, as the Lower Cambrian sediments in the McCalls Ferry-Quarryville district were nearer to the Paleozoic shore line and therefore under lighter load than the sediments of the same age in the deeper part of the geosyncline 10 or 20 miles farther northwest. Nor does a dying out of the folding explain this difference in metamorphism, because the slaty Conestoga limestone near Lancaster is folded just as strongly as its metamorphosed equivalent near Quarryville. The additional weight of 10,000 feet of rock thrust over the normal stratigraphic column would produce a rise in geothermal temperature of only 100° C., which would not be adequate to explain the development of garnetiferous biotite schists and micaceous marbles near Quarryville in place of phyllites and slaty limestones 20 miles farther northwest. Both the autochthonous and the overriding rocks in Carroll County, Md., are more feebly metamorphosed than the corresponding rocks in Pennsylvania, although the folding is equally close in both regions. Thus if the weight of the overriding fault block were the dominant factor in the intense metamorphism in Pennsylvania this would necessitate the conclusion that the thickness of the fault block was appreciably less in Maryland than in Pennsylvania, so that folding and the attendant metamorphism would have taken place at a lesser depth and under conditions of less intensity.

Such a localized rise in geothermal temperature as is indicated by the Paleozoic schists that flank the Mine Ridge anticline suggests still another causal factor in folding—namely, the influence of underlying igneous magma.

Folding that took place under such thermal influence, combined with the locally augmented load of the overriding fault block, would doubtless account for the intense metamorphism in the autochthonous Paleozoic block and for the upward waning of intensity in the pre-Cambrian rocks.

This hypothesis of regional alteration under the influence of an underlying batholith is attractive in that it satisfactorily explains the peculiar localization of metamorphic intensity, but unfortunately it is virtually insusceptible of proof, except in so far as the widespread tourmalinization and pegmatitization of the Paleozoic rocks indicate the passage of magmatic emanations from an underlying igneous mass. This evidence is so conclusive that the presence of the underlying batholith is certain, but whether the influence of the batholith is sufficient to account for the intense metamorphism in the belt of highly altered Paleozoic rocks is speculative in the light of present evidence. In fact, from our present knowledge of the metamorphic history of the region it is impossible to ascribe the localized intensity of the metamorphism in the vicinity of the Mine Ridge upland to one rather than to the other of the two possible causes that have been outlined. It can only be said that both possibilities answer the requirements of working hypotheses, although the first hypothesis seems inadequate to account for the extremely local rise in metamorphism. It remains for these hypotheses to be later eliminated or substantiated by more detailed and more extended work on the recurrent metamorphism of the Piedmont province.

GEOLOGY OF THE McCALLS FERRY-QUARRYVILLE DISTRICT, PENNSYLVANIA

By Eleanora Bliss Knopf and Anna I. Jonas

INTRODUCTION

LOCATION OF THE DISTRICT

The McCalls Ferry and Quarryville quadrangles mapped by the United States Geological Survey are in southeastern Pennsylvania, in York, Lancaster, and Chester Counties, between parallels 39° 45′ and 40° and meridians 76° and 76° 30′. (See fig. 1.) The geology of these quadrangles is so intimately related to the geology of the adjoining portions of Pennsylvania and Maryland in which the writers have worked that this report describes not only the area included in the quadrangles but also a considerable part of the adjoining region. The McCalls Ferry quadrangle is crossed from north to south by Susquehanna River, whose right bank forms the boundary line between York County on the west and Lancaster County on the east. Fully two-thirds of the two quadrangles lies in Lancaster County, and the greater part of the remainder is in York County, only the extreme southeast corner belonging to Chester County. There are no large towns in the district. Quarryville, a borough of several hundred inhabitants, is 12 miles southeast of Lancaster and 20 miles southwest of Coatesville.

Four through railroads and a small branch line enter the McCalls Ferry and Quarryville quadrangles. The northeast corner is crossed for a distance of 6 miles by the Philadelphia division of the Pennsylvania Railroad, which is a part of the main line between Philadelphia and Pittsburgh. The central part of the district is crossed from east to west by the Atglen & Susquehanna branch of the Pennsylvania Railroad, which turns north at Shenks Ferry station and follows the east bank of the Susquehanna to Harrisburg. This line, which carries only freight, is locally known as the "low-grade freight line," because it maintains an approximately uniform grade of about 5 feet to the mile between Philadelphia and Harrisburg. The Columbia & Port Deposit branch of the Pennsylvania Railroad runs along the east bank of Susquehanna River. Its northern terminus is Columbia, and it connects at Perryville with the Philadelphia, Wilmington & Baltimore division of the Pennsylvania Railroad. The Maryland & Pennsylvania Railroad follows the winding course of Muddy Creek in the southwestern part of the district. It starts in Baltimore, Md., and its northern terminus is York, Pa. The Strasburg Railroad is a branch line of the Pennsylvania Railroad which runs from the main Philadelphia division at Leaman Place 4 miles southwest to Strasburg. A narrow-gage road

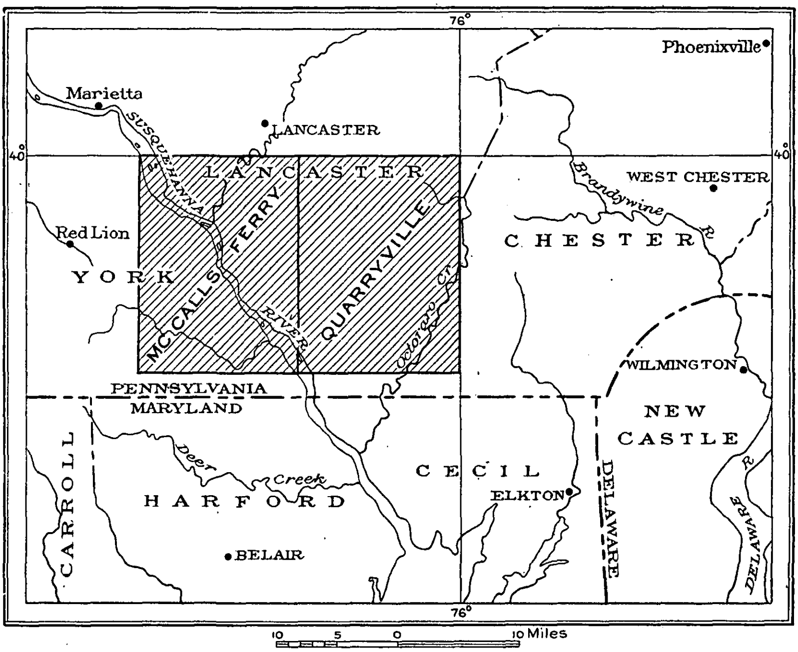

Figure 1.—Index map showing location of McCalls Ferry and Quarryville quadrangles, Pennsylvania

known as the Lancaster, Oxford & Southern Railroad, which served to connect Susquehanna River at Peach Bottom with Quarryville to the north and Oxford to the east, was abandoned in 1921. A standard-gage freight line between Quarryville and Lancaster, a branch of the Pennsylvania Railroad, is still in operation. An extensive trolley system, under the management of the Conestoga Traction Co., radiates from Lancaster, and three branches of this system terminate within the district, at Millersville, Quarryville, and Strasburg. A fourth branch of this trolley system crosses the district between Kinzers and Christiana and terminates at Coatesville, 9 miles east of Christiana. The branch of the electric railroad of the Conestoga Traction Co. that terminates at Millersville connects with Susquehanna River by the Millersville & Pequea Electric Railroad.

Susquehanna River is crossed by motor ferries at Pequea and McCalls Ferry and by a steam ferry at Peach Bottom. The river is navigable for small boats between Shenks Ferry and Holtwood, where the installation of a large dam has impounded the river water into Tucquan Lake. Below the tail race of the Holtwood dam it is navigable during the high-water season.

FIELD WORK AND ACKNOWLEDGMENTS

The field work on the area described in this report was started by the authors during the field season of 1915. It was interrupted by the World War and was resumed and almost completed during portions of the field seasons of 1919, 1920, and 1921. The final detailed mapping of the Paleozoic rocks in the northern part of the Quarryville and McCalls Ferry quadrangles was done by A. I. Jonas during part of the field seasons of 1921 and 1922. The accompanying geologic map (pl. 1) is the result of this field work.

The geological survey of the McCalls Ferry and Quarryville quadrangles was planned in 1915 by Mr. Arthur Keith, chief of the section of areal geology. In February, 1915, Dr. E. B. Mathews, to whom the area had been originally assigned before his resignation from the United States Geological Survey, turned over to the Geological Survey a reconnaissance map and some reconnaissance notes made by Mr. W. J. Miller while he was a student at Johns Hopkins University. The area was then assigned to Eleanora F. Bliss (E. B. Knopf) for mapping. The work was carried on in full cooperation with A. I. Jonas, of the Pennsylvania State Geological Survey. Mr. Sidney Paige, who was chief of the section of areal geology during the progress of the later part of the field work, inspected the work, and acknowledgments are due to him for his advice on several occasions during the preparation of the manuscript.

Acknowledgments are especially due to Mr. G. W. Stose for his collaboration on the stratigraphy of Lancaster Valley and to Mr. Adolph Knopf for his assistance on the metamorphic problems of the district.

PREVIOUS WORK

The earliest geologic work in the McCalls Ferry-Quarryville district consisted of the general investigations by the geological surveys of Pennsylvania, whose results were published in the several county reports. In 1895 Kemp studied the geology of the Gap nickel-ore deposit but confined his attention to the rocks immediately associated with the ore. In 1905 W. J. Miller made a brief reconnaissance, but the results were never published.

Although little has been written on the McCalls Ferry-Quarryville district in particular, the geology of the belt of highly metamorphosed rocks in which the district lies has long been a subject for discussion. The tendency among the early writers, such as Rogers, Hall, and Lesley, was to place practically all the crystalline schists of the region in the pre-Cambrian. The later work of Maryland geologists placed the series that overlies the pre-Cambrian Baltimore gneiss in the Paleozoic. The tentative conclusions reached by Mathews[1] in 1905 were that the Paleozoic section probably begins with the Setters formation; that the overlying Cockeysville marble is probably "Cambro-Ordovician" and represents the same horizon as the limestone of Chester Valley in the Philadelphia area; and that the Wissahickon formation, overlying the marble, is probably Ordovician. But even before this Bascom[2] had recognized the possibility that the Wissahickon gneiss might prove to be pre-Cambrian, and in 1909,[3] at the time that the Philadelphia folio was published, she decided that the age of the Wissahickon gneiss is pre-Cambrian—first, because of certain apparent stratigraphic relations between the Wissahickon gneiss and the adjoining sedimentary rocks in Pennsylvania, and second, on the ground that it contains intrusive igneous material generally considered to be absent from the Paleozoic rocks of the region. A series of phyllites and schists, termed the "Octoraro schist," was separated by Bascom from the Wissahickon and placed in the Ordovician, because these rocks were considered to be less coarsely crystalline than the Wissahickon and to conformably overlie limestone of Beekmantown age in Chester Valley. The contact of the Wissahickon and the "Octoraro schist" was regarded as a thrust fault.

Early study of the Doe Run and Avondale district, 12 miles east of Quarryville, brought the writers[4] also to the conclusion that the Wissahickon gneiss is a different formation from the "Octoraro schist" and that it must be of pre-Cambrian age, thrust over the so-called "Cambro-Ordovician" limestone and over the "Octoraro schist," then considered to be Ordovician. The sinuous line of contact between the gneiss and the limestone and between the gneiss and the schist was considered to represent the eroded plane of the over-

[1] Mathews, E. B., Correlation of Maryland and Pennsylvania Piedmont formations: Geol. Soc. America Bull., vol. 16, pp. 329–346, 1905.

[2] Bascom, F., The geology of the crystalline rocks of Cecil County: Maryland Geol. Survey, Cecil County, pp. 107–108, 1902.

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[4] Bliss, E. F., and Jonas, A. I., Relation of the Wissahickon mica gneiss to the Shenandoah limestone and the Octoraro schist of the Doe Run and Avondale region, Chester County, Pa., private publication, February, 1914; U. S. Geol. Survey Prof. Paper 98, pp. 9–34, 1916.

thrust which had been folded in a period of deformation subsequent to the faulting.

A similar explanation of the relations of the Wissahickon gneiss to the other formations northeast of Baltimore was given by Mathews[5] in 1917. The pre-Cambrian Wissahickon gneiss is described as thrust over the Baltimore gneiss, Setters formation, and Cockeysville marble along such a fault plane as was believed to explain the anomalous position of the Wissahickon gneiss in Pennsylvania.

During the field seasons of 1919 and 1920 spent in Baltimore and Harford Counties, in Maryland, the writers were able to determine the position of the Wissahickon gneiss in the stratigraphic column and to show that the limestones of the Doe Run and Avondale district of Pennsylvania and the Cockeysville marble of Maryland are pre-Cambrian, not Shenandoah, and that they had been erroneously correlated with the limestone series of Chester Valley. The pre-Cambrian section established by the investigation is described in some detail in this report.

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GEOGRAHPY

TOPOGRAPHY

The McCalls Ferry-Quarryville district is part of the physiographic division of the eastern United States that is known as the Piedmont province. The topography of the district falls into two distinct types—a well-dissected upland, whose interstream areas are wide, flat hill summits, and a gently rolling lowland, broken here and there by long, low hills trending northeast. The lowland occupies the northern and northwestern portion of the district and widens northward into Lancaster Valley. To the south it cuts across the upland surface in a depression about 1 mile wide, near Quarryville. From Quarryville toward the northeast a long, narrow valley, known as Chester Valley, separates the upland into a northern portion, which includes the Mine Ridge upland and a southern portion known as the South Valley Hills.

Chester Valley is a well-defined topographic feature, controlled in shape and extent by the easy solubility of the underlying rock, which is limestone and dolomite. It has a linear extent of 66 miles and finally disappears north of Philadelphia, where the limestone outcrop ends on the crest of a pitching fold. The floor of the valley is flat, about a quarter of a mile in width at the narrowest part, and widening northeastward to a couple of miles. It slopes from an altitude of 580 feet at Quarryville to 100 feet near Philadelphia and is occupied by a succession of small runs, some flowing westward and some eastward. There is no continuous stream by whose agency the valley might have been formed. The bounding ridges that delimit the valley rise in steep slopes 150 to 250 feet above the valley floor.

The northern portion of the upland rises 350 feet above the level of Lancaster Valley and 140 to 200 feet above the floor of Chester Valley. To a distant observer the upland appears to be flat, but a close examination of topographic detail reveals the fact that the surface is more diversified than would at first appear. The western and northwestern rims of the upland rise appreciably above the general upland level in the narrow Mine Ridge proper. This ridge, which presents to the eye of an observer in Lancaster Valley an approximately even crest line similar to the crest of Blue Mountain, attains a maximum altitude of 920 feet and descends on both sides of this maximum to surfaces that maintain for several miles altitudes ranging from 820 to 880 feet. Similar altitudes ranging from 820 to 900 feet reappear in the upland surfaces on the west side of the Quarryville Valley and to the southwest across Susquehanna River in York County, forming the highest and flat summits of the upland country. These hills at 820 and 900 feet form the divide between northward and southward flowing streams, which is breached by Susquehanna River in a deep gorge from northwest to southeast. The steep rugged slopes of this gorge are covered with a dense growth of timber, and in the valley bottom, which ranges in width from a quarter of a mile to more than a mile, the river rushes in a deep and narrow channel past many rocky islets, which occupy most of the valley floor. Even now, to a person standing in the gorge at the mouth of one of the tributaries that empty into the main stream through ravines dense with laurel and rhododendron, the wildness of the scene recalls the days when the Susquehanna furnished to the Indian canoe a means of travel through a well-nigh impassable wilderness. But on the upland summits the atmosphere of the wilderness is lost, and the lonely wooded gorge of the river forms a sharp contrast to the fertile farming land that occupies the wide upland of comparatively slight relief.

DRAINAGE

The district is drained by Susquehanna River and numerous tributaries. The length of the Susquehanna from its source in Lake Otsego, N. Y., to the head of Chesapeake Bay is 422 miles, and it drains an area of 27,400 square miles.[6] The portion that is included in the district here described lies 380 miles from the source and 17 miles from the mouth and has an average fall of 5.3 feet to the mile. The largest tributaries to Susquehanna River are Fishing and Muddy Creeks on the York County side and Conestoga and Pequea Creeks on the Lancaster County side. The flow of water in Susquehanna River is extremely variable. Between 1911 and 1916 the average flow below the Holtwood Dam was 20,000 second-feet, the maximum 440,000 second-feet, and the minimum 2,000 second-feet.

The amount of water carried by the river is small in comparison to the size of its valley, the river being in this sense an underfit stream. The explanation is doubtless to be found in the fact that at one time during the retreat of the ice of the glacial epoch the drainage from the Finger Lakes of New York went chiefly southward through Susquehanna River. The volume of water carried by the river at that time therefore was probably much larger than it is at present.

A curious feature of the channel is the occurrence of the "deeps" or long depressions below the level of the channel that have been described by Mathews.[7] Six of these deeps occur between Turkey Hill and Fites Eddy, where the channel is near the left bank. They range from 2 miles to slightly less than a mile in length and attain a maximum depth of 30 feet below sea level. They are confined to the eastern edge of the rock channel and are associated with large potholes. No entirely satisfactory explanation has been proposed to account for the origin of these remarkable holes. Mathews believed that they were the work of post-Talbot erosion during a time

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[6] Hollister, G. B., U. S. Geol. Survey Water-Supply Paper 108, p. 9, 1904.

[7] Mathews, E. B., Submerged "deeps" in the Susquehanna River: Geol. Soc. America Bull., vol. 28, pp. 335–346, 1917.

when the river channel was either narrower or the volume considerably greater than at present. Such increase in volume might have been caused by increased drainage through the Susquehanna, but it is difficult to explain by increased volume the localization of the deeps. Daly[8] has suggested that a more probable reason for the formation of these depressions is an increased gradient of the stream deposits on the southerly slope of a peripheral bulge of the land south of the continental ice cap. But study of the tributary drainage in the zone of the assumed peripheral upwarping of the crust shows no such trenching as must have accompanied a steepening in gradient sufficient to account for the carving of the Susquehanna deeps. Moreover, the crest of the bulge must have been at no great distance from the foot of the ice sheet, which in this region was as much as 110 miles from the Turkey Hill-Fites Eddy stretch of Susquehanna River.

CLIMATE

The mean temperatures for the eastern Piedmont of Pennsylvania are 31° F. for winter, 50° for spring, 72° for summer, and 55° for fall. The maximum summer temperature rises above 95°, and the minimum winter temperature falls below zero, especially in the month of February. There is a difference of about 3° between the climate of the western part of the Piedmont area and that of Philadelphia. In spring the last killing frost occurs from April 6 to May 27, and in fall the first killing frost from September 17 to November 14, according to the season and to the altitude of the section. Sudden melting of the ice and snow in the mountains often causes the formation of ice gorges in the Delaware and Susquehanna River valleys. The ice gorges are due to an insufficiency of water to carry out the floating ice blocks, which are penned up in narrows of the river and refrozen into a solid mass by sudden drops in temperature. The resultant ice jams have attained great size in Susquehanna River below Columbia and have caused much damage to river property, both by ice work and by flooding of the shores of the mainland and tributary streams.

The mean annual rainfall is 40 inches, as shown by records at Lancaster covering 20 years. Records at Coatesville covering a period of 32 years show a mean annual rainfall of 48 inches, with a variation from 32 to 69 inches. Records at Holtwood covering the period from 1915 to 1920 show a mean annual rainfall of 37 inches, with a maximum of 46 inches.

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[8] Daly, R. A., Oscillations of level in the belts peripheral to the Pleistocene ice caps: Geol. Soc. America Bull., vol. 31, p. 314, 1920.

GENERAL GEOLOGY

OUTLINE OF THE GEOLOGY OF THE APPALACHIAN MOUNTAINS AND THE PIEDMONT PROVINCE

The easternmost physiographic division of the Atlantic slope is the Coastal Plain, a gently sloping plain underlain by unconsolidated deposits. (See fig. 2.) Between the Coastal Plain on the east and the long parallel ridges and intermontane valleys of the Appalachian

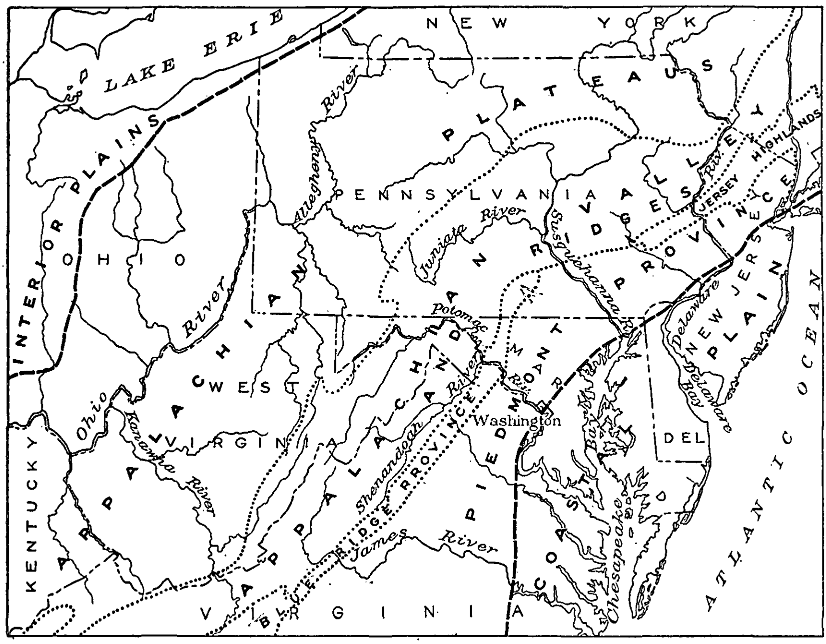

FIGURE 2.—Physiographic provinces in the Middle Atlantic States. The Jersey Highlands, including the Reading and Durham Hills, are the northern representative of the Blue Ridge province and are generally considered to extend northward into the New England upland. South Mountain in Pennsylvania forms the north end of the Blue Ridge province

Ridge and Valley province on the west lies the Piedmont province. The western boundary of the Piedmont province is formed by the Appalachian Mountains, an interrupted mountainous belt of moderate relief with maximum altitudes of about 6,000 feet.

APPALACHIAN MOUNTAINS

GENERAL FEATURES

The northern portion of the Appalachian Range comprises the Green Mountains, in Vermont and western Massachusetts, and the Hoosac Range, in southwestern Massachusetts and Connecticut. Its southwestward extension through New York crosses Hudson River between Fishkill and Peekskill and continues through New Jersey as the Highlands. This range crosses Delaware River near Easton and forms in Pennsylvania the Reading and Durham Hills, whose highest summits stand at 1,000 to 1,100 feet. The Reading and Durham Hills terminate abruptly in Penn and Neversink Mountains, which overlook Schuylkill River at Reading. In Lebanon and Berks Counties about 6 miles southwest of Reading an isolated group of hills, known as South Mountain, marks the south end of the northern Appalachians. From these hills southwestward for a distance of 50 miles there is a topographic break and disappearance of the ridge that marks the boundary between the Cumberland-Lebanon Valley of the Appalachian Ridge and Valley province on the west and the Piedmont province on the east.

South of Carlisle the Appalachian Mountains reappear in the larger South Mountain, completely disconnected from the South Mountain in Lebanon and Berks Counties, near Reading. The South Mountain of southern Pennsylvania is a wide area made up of many narrow ridges with a maximum altitude of 2,100 feet which extend southward into South Mountain and Catoctin Mountain, Maryland. The Appalachian Mountains in Virginia lie between Catoctin Mountain and the western front of the Blue Ridge. The Appalachian Mountains widen in the Southern States, forming a belt about 50 miles in width, which comprises the Unaka and Great Smoky Mountains and many others.

The topography of the Appalachian Mountains is determined by the nature of the underlying rocks. The range is composed of schists and gneisses of both sedimentary and igneous origin. In the Highlands of New Jersey and in the Reading and Durham Hills the hills that are composed of feldspathic gneisses are of rounded or irregular outlines, with more or less gently sloping sides, and the hills that consist of quartzite form ridges elongated along the strike, whose steep rugged slopes are covered with fragments of broken quartzite.

GEOLOGY OF THE READING HILLS

The writers have studied the crystalline schists of the Appalachian mountains in some detail in the Reading Hills, the southward continuation into Pennsylvania of the New York and New Jersey Highlands. The oldest rocks of sedimentary origin in these hills are graphite-bearing biotite gneisses containing intercalated layers of marble and quartzite. The graphitic gneisses are associated with hornblende schists that may belong to the sedimentary series, although it has not yet been possible to delimit them from other hornblende schists of undoubted igneous origin. The sedimentary series is lithologically similar to the Franklin limestone and interbedded graphitic schists of the New Jersey Highlands, and like those rocks it has been invaded by a series of igneous rocks ranging in composition from granite to gabbro. Numerous small isolated patches of sedimentary gneiss completely surrounded by igneous rock probably represent rifted blocks or roof pendants. In a region of moderate relief it is impossible to establish the presence of roof pendants because the depth of outcrop is not sufficient to expose the downward prolongation of the pendant into the intruding igneous material. The metamorphosed sediments are granoblastic in texture as a result of a deep-seated metamorphism that has developed various heavy minerals such as garnet, sillimanite, hornblende, and plagioclase. Undulose extinction is almost universally present and indicates that a mechanical deformation has been superimposed upon the deep-seated metamorphism. In contrast to the crystalloblastic texture of the sedimentary gneisses, the igneous gneisses show distinct igneous texture. The dynamic deformation that affected the sediments was later than the igneous intrusion, for it has left its mark upon the igneous rocks in undulatory extinction, granulation, and twisting of the mineral constituents. In certain rather well-defined zones the deformation has been so intense that the rocks involved show all stages of cataclastic action from a strong brecciation to a complete mylonitization through which the original character of the rock can be only faintly discerned. The shattering of the rock complex was accompanied by a certain amount of mineralization, as shown by the presence of crocidolitized zones,[9] and by the passage of ore-bearing solutions.

The presence of certain fine-grained and notably undeformed bostonitic dikes in the magnetite district of Rittenhouse Gap suggests the presence of an alkalic magma younger than the magma that furnished the rocks of the strongly deformed igneous complex. The Reading and Durham Hills are penetrated by dikes of Triassic diabase[10] and broken by numerous faults of Triassic age, and the possible genetic relationship of the alkalic dikes to the nephelite syenite magma of Beemerville, N. J., suggests that the considerable deformation during late Triassic time in the Reading and Durham Hills was accompanied by the intrusion of both diabasic and alkalic magma. Such an idea is in accord with the suggestion made by Wherry[11] that the hydrothermal solutions by which the gneisses were crocidolitized originated in the Triassic diabase magma. The

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[9] Wherry, E. T., and Shannon, E. V., Crocidolite from eastern Pennsylvania: Washington Acad. Sci. Jour., vol. 12, pp. 242–244, 1922.

[10] Jonas, A. I., Pre-Cambrian and Triassic diabase in eastern Pennsylvania: Am. Mus. Nat. Hist. Bull., vol. 37, pp. 173–181, 1917.

[11] Wherry, E. T., and Shannon, E. V., op. cit., p. 244.

crystalline schists of the Reading and Durham Hills and the associated igneous complex are unconformably overlain by a slightly altered series of Paleozoic sediments. Evidence of the unconformity is afforded by the discordance in structure between the Paleozoic and the pre-Cambrian rocks and by the presence in the basal Cambrian beds of reworked material derived from the underlying igneous rocks.

GEOLOGY OF CATOCTIN-SOUTH MOUNTAIN

The pre-Cambrian crystalline rocks of the New York and New Jersey Highlands and of the Reading and Durham Hills, as far south as the South Mountain in Lebanon County, Pa., form a continuous belt with similar lithologic character. There is a distinct topographic break west and southwest of this South Mountain in Lebanon-Berks County, where the rugged hills of crystalline schist are replaced by the limestone lowland and rolling shale hills of the Lebanon-Cumberland Valley. South of Carlisle, where the Appalachian Mountains reappear in the South Mountain-Catoctin Mountain ridges of southern Pennsylvania and Maryland, the pre-Cambrian core of the range is quite different in lithologic character from the crystalline core of the Reading and Durham Hills. In Catoctin Mountain and South Mountain a series of devitrified rhyolites and epidotized basaltic flows [12] intruded by granite are overlain unconformably by Cambrian sedimentary rocks that show pebbles of the underlying lava in the basal conglomerate.[13]

Up to the present time it has been impossible to establish any geologic relationship between the lavas and the plutonic intrusions of the Reading Hills. Recent work in the pre-Cambrian stratigraphy of the adjoining Piedmont Province, however, has furnished some suggestions concerning possible correlation, which will be presented in the following pages.

PIEDMONT PROVINCE

The rugged hill country of the Appalachian Mountain belt slopes rapidly downward for about 1,500 feet to the level of the Piedmont province, which extends eastward for 30 miles to the western margin of the Coastal Plain. The boundary between the Piedmont province and Coastal Plain runs through New Jersey from Perth Amboy southwestward to Trenton; south of this point it follows Delaware River through Philadelphia southwestward to Wilmington, thence

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[12] Keith, Arthur, Geology of the Catoctin belt: U. S. Geol. Survey Fourteenth Ann. Rept., pt. 2, pp. 285–395, 1894. Bascom, F., The ancient volcanic rocks of South Mountain, Pa.: U. S. Geol. Survey Bull. 136, 1896.

[13] Bascom, F., op. cit., p. 28.

through Elkton, Havre de Grace, and Baltimore to Washington. This boundary is really a border zone, which is known as the "fall line" because here the stream courses are turbulent and full of rapids. The extreme width of the Piedmont province is 120 miles; its mean width is 60 miles. Its general slope is eastward and southeastward. In New Jersey it slopes from 400 feet above sea level at the western margin to an altitude at the margin of the Coastal Plain which ranges from 100 feet on Delaware River to sea level on Newark Bay. Between Susquehanna and Delaware Rivers the higher summits of the Piedmont province attain an altitude of 1,100 feet in Welsh Mountain and slope down to about 300 feet near the fall line. The Piedmont province rises to its culminating point in South Carolina and Georgia, where it has an altitude of 1,800 to 1,900 feet along the base of the Blue Ridge.

GENERAL GEOLOGY OF THE PIEDMONT PROVINCE IN PENNSYLVANIA

The Piedmont province is divided into the Triassic Lowland and the Piedmont Upland. In Pennsylvania the Triassic deposits were laid down in a basin that now extends westward and southwestward along the northwestern edge of the province. The Triassic Lowland ranges in width from 40 miles at Delaware River to 4 miles near Cornwall. The Triassic shales and sandstones, dipping gently westward, form a rolling surface diversified by rugged hills of conglomerate and by rounded knoblike hills and long, curving ridges of diabase and indurated sediments that rise to 1,200 and 1,300 feet on both sides of Susquehanna River.

Southeast of the Triassic area maturely dissected hills with a maximum altitude of 900 to 1,080 feet form the gentle topography of the Piedmont Upland. The country is well dissected by streams that flow in broad valleys having composite slopes. The interstream areas are flat-topped hills, which are remnants of partial peneplains not entirely removed by erosion since their formation. The shape of the hills is largely dependent on the character of the underlying rocks.

The Piedmont Upland is underlain by crystalline schists and less altered slates, phyllites, and limestones. Crystalline schists and gneisses, including a few areas of marble, form the easternmost belt of the upland from Trenton, N. J., southward to Washington. For a large part of the westernmost belt the term "upland" is really a misnomer, inasmuch as this part is occupied by a rolling plain of low relief. This plain, which is underlain chiefly by limestones and dolomites, is known as the Lancaster, York, and Hanover Valleys in Pennsylvania and the Frederick Valley in Maryland. It has a wide range in width from a maximum of 20 miles in the vicinity of Lancaster to a minimum of 1 mile on Potomac River near Frederick.

Southeast of Lancaster, in the vicinity of Quarryville, the valley cuts across the northeastward trend of the schist hills and continues through the middle of the Piedmont Upland in a narrow depression known as Chester Valley, which 8 miles north of Philadelphia merges into the wide plain of the Triassic Lowland. The area of the Lancaster and York Valleys is interspersed by several anticlinal ridges underlain for the most part by phyllites and quartzites.

Post-Triassic sediments overlap the margin of the Piedmont province and occur in residual patches upon its surface. These sediments are unconsolidated deposits of sand, marl, and gravel that have been gently tilted southeastward. They range in age from Lower Cretaceous to Quaternary.

GEOLOGY OF THE PIEDMONT UPLAND IN MARYLAND

During a study of the crystalline rocks in Baltimore County, Md., and of their northeastward extension in Pennsylvania, begun in 1919, the writers have established the stratigraphic succession shown in Figure 3 for the pre-Cambrian crystalline schists of the eastern border of the Piedmont province in Maryland.

Baltimore gneiss.—The oldest rock of the region is the Baltimore gneiss, a light-colored, thoroughly recrystallized rock of sedimentary origin, which is in some places a thickly bedded gneiss of granitic aspect and in others a thinly banded ribbon gneiss. The deep-seated nature of the metamorphism that has produced the foliation is shown by the granoblastic texture and by the nature of the constituent minerals. The gneiss is composed dominantly of quartz, oligoclase, microcline, and biotite. Garnet and tourmaline are characteristic accessories. The sedimentary origin of the gneiss is shown in the field by the distinct and persistent even bedding, which can be traced in some quarries completely across the outcrop, for 100 feet or more. Study of thin sections indicates that recrystallization is complete and no trace of original sedimentary texture has been preserved. Many of the zircons occur in well-worn, rounded grains that are indicative of water-borne material. This criterion, however, must be used with considerable caution, because it is possible that zircon in igneous rocks may be rounded by magmatic corrosion. The Baltimore gneiss has been intruded by a granite that has produced a lit-par-lit injection in the upper part of the formation.

Glenarm series.—Unconformably overlying the Baltimore gneiss is a series of later pre-Cambrian sediments termed the Glenarm series, from its typical development near Glenarm, 13 miles northeast of Baltimore. The United States Geological Survey classifies the Baltimore gneiss as Archean and the Glenarm series as Algonkian, though in the writers' opinion there is lack of adequate evidence that

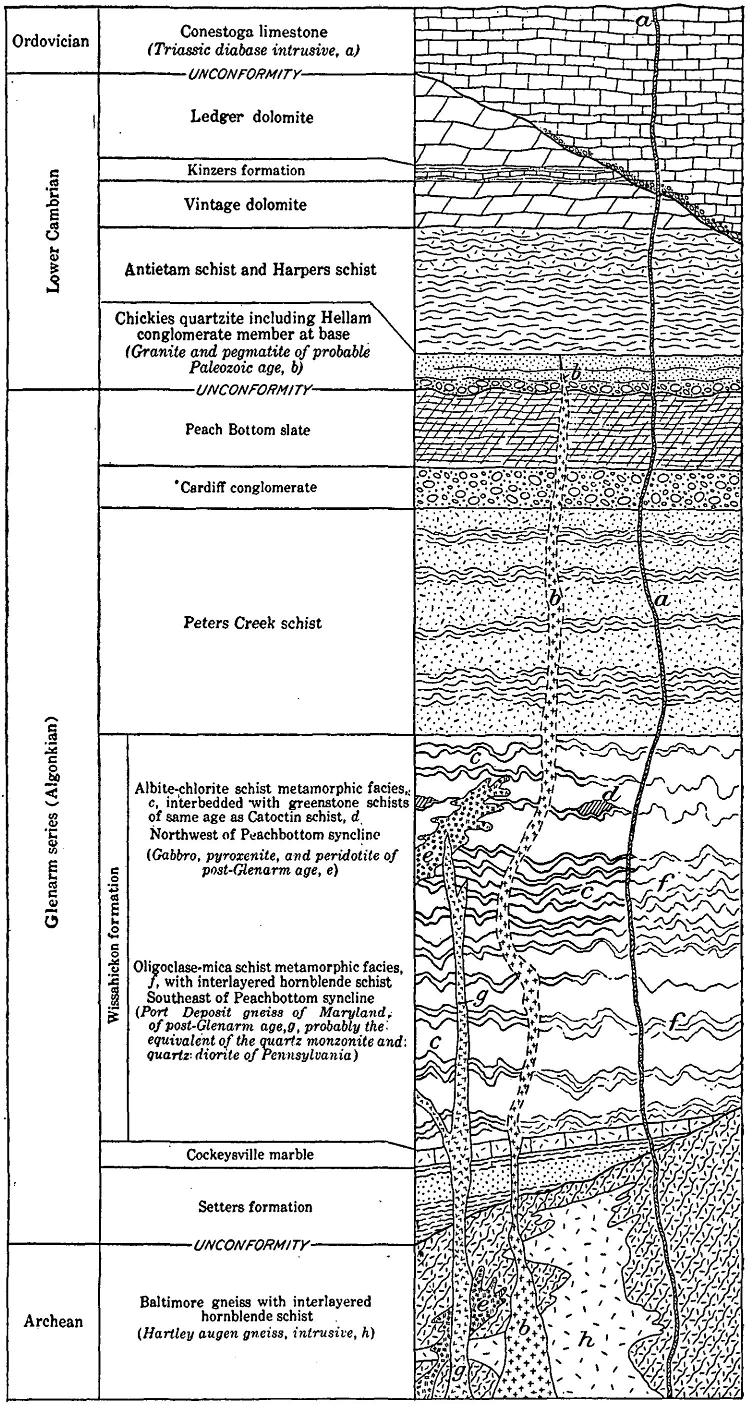

FIGURE 3.—Columnar section showing crystalline schists and sedimentary rocks of the McCalls Ferry-Quarryville district and adjoining region in Maryland and some of the igneous intrusions

they are the precise equivalents of rocks referred to these systems in other regions.

The Glenarm series comprises the Setters formation, the Cockeysville marble, the Wissahickon formation, the Peters Creek schist, the Cardiff conglomerate, and the Peach Bottom slate. The total thickness of the series probably amounts to 8,000 or 10,000 feet, although no accurate estimate can be made, for the middle formations have been repeated by close folding. In some places the Wissahickon formation lies directly upon the eroded surface of the Baltimore gneiss, owing to the fact that the earlier formations of the series are overlapping shore deposits. So far as now known the deposition of Glenarm sediments was continuous and not interrupted by erosion or by orogenic deformation.

Only the upper formations of the Glenarm series are exposed in the McCalls Ferry-Quarryville district, but a brief description of the lower Glenarm section as established in Baltimore County is introduced here in order to make clear the stratigraphic relations of the whole series.

The basal formation is the Setters, which overlies the Baltimore gneiss unconformably but with no basal conglomerate. Williams [14] considered that the basal schist of the Setters formation was really a transition bed between the Baltimore gneiss and the overlying formation, but recently it has proved possible to establish the unconformity by several facts. A structural discordance between the two formations can be recognized, the divergence in average strike amounting to 20°. There is more evidence of deformation in the Baltimore gneiss than in the Setters formation, even though the members of the gneiss are more resistant to deformation than the argillaceous layers of the Setters. Moreover, the Setters is irregular in thickness and in some places is entirely lacking, so that the Cockeysville marble overlaps the Setters deposits and lies directly upon the Baltimore gneiss. In the absence of positive evidence of erosion, it seems likely that the irregular thickness of the Setters formation and of the Cockeysville marble is due to transgression rather than to erosion. The Setters formation is composed of mica gneiss and mica schist with an intercalated quartzite member. The maximum thickness measured is 1,000 feet. The whole formation has been thoroughly recrystallized in the lowermost zone of metamorphism, with the production of deep-seated minerals such as garnet and biotite. Tourmaline is an abundant and characteristic accessory constituent, which is undoubtedly of pneumatolytic origin and associated with a granite known as the Gunpowder granite, which has invaded the Baltimore gneiss and the overlying Glenarm

[14] Williams, G. H., Geology of Baltimore and vicinity (extract from the Guidebook of Baltimore prepared for the American Institute of Mining Engineers), February, 1892.

series. The Setters formation is characterized by an unusually high content of potash which has crystallized as microcline and muscovite, indicating that the original sediment was a mud or sand high in potash feldspar and white mica.

The Cockeysville marble consists of alternating dolomitic and calcitic layers. It overlaps the Setters formation and Baltimore gneiss and grades through a calcareous mica schist into the overlying Wissahickon formation. Its probable thickness is 400 feet. The Wissahickon formation occurs in two metamorphic facies—an oligoclase-mica schist and an albite-chlorite schist. The oligoclase-mica schist facies is well developed in the vicinity of Baltimore. It is lithologically similar to the Setters, but the quartzite beds are less conspicuous than those in the Setters because they are thinner and more evenly distributed throughout the formation, whereas in the Setters the quartzite is chiefly confined to the middle member. The Wissahickon differs from the Setters also in its lower content of potash, which has caused oligoclase rather than microcline to be the dominant feldspar. The Wissahickon oligoclase-mica schist is a series of thoroughly recrystallized sediments that have been metamorphosed under deep-seated conditions with the production of garnetiferous muscovite-biotite schists, staurolite and kyanite schists, and biotite-plagioclase gneiss. In some places tourmaline is abundantly developed.

In northern Baltimore County, Md., and in the southern part of York and Lancaster Counties, Pa., metamorphism under different conditions has resulted in the formation of an albite schist that is considered to be the stratigraphic equivalent of the Wissahickon oligoclase-mica schist, the difference in mineral constituents being caused by metamorphism of moderate intensity in contrast to the intense metamorphism of the oligoclase-mica schist.

The stratigraphic relations of the Wissahickon albite schist and the upper formation of the Glenarm series to the underlying and overlying rocks have been established mainly by work in the McCalls Ferry-Quarryville district. Their lithologic character and stratigraphic relations are described in detail in the following section.

GEOLOGY OF THE McCALLS FERRY-QUARRYVILLE DISTRICT

PRE-CAMBRIAN ROCKS OF SEDIMENTARY ORIGIN

PRE-GLENARM ROCKS OF THE MINE RIDGE UPLAND

BALTIMORE GNEISS

North of Chester Valley an upland known as the Honeybrook upland extends from Valley Forge, on Schuylkill River, southwestward for a distance of 40 miles. The surface of this upland truncates a wide anticlinal arch that ends in several westward-pitching anticlinal folds. These anticlines underlie three elongated hills—Welsh Mountain, Barren Hill, and the Mine Ridge upland. The Mine Ridge upland has been studied in detail by the writers, but the results are unsatisfactory, owing to the unusual scarcity of outcrops, exceptional even in a region that has long been the bane of geologists on account of its deep, fertile soil and intensive cultivation. The only places where more than isolated outcrops are available for study are sections exposed in rock cuts along the railroads that cross the upland in three places. These three sections are (1) that between Christiana and Gap on the main line of the Pennsylvania System between Philadelphia and Harrisburg; (2) that between Coatesville and Brandywine Manor on the Wilmington division of the Philadelphia & Reading Railroad; (3) that north of Downingtown on the New Holland branch of the Pennsylvania Railroad.

A gneiss of sedimentary origin occurs in the core of the Mine Ridge upland below the Cambrian gneisses, quartzites, and schists that flank the uplift. Metamorphism has completely altered both Paleozoic and pre-Paleozoic sediments into highly crystalline schists and gneisses, in which the schistosity is so well marked that it has more or less obliterated original structural differences. But the unconformity between the Cambrian rocks and the underlying gneiss is well marked on the Philadelphia & Reading Railroad 1 mile north of Coatesville, where the Cambrian schists and quartzites dip southward over the northward-dipping pre-Cambrian schists and gneisses. Additional evidence of unconformity consists in the constant repetition of layers of hornblende schist or amphibolite in the underlying gneiss, combined with the absence of all such interlayered amphibolites in the Cambrian sediments.

The upper beds of the pre-Cambrian sedimentary series are biotite-albite schists. The writers have so far never succeeded in finding in this district an exposure showing the actual contact of pre-Cambrian and Cambrian rocks. Four miles northeast of Christiana and about 700 feet south of the Steprock quarry, in Cambrian quartzite, the rock that lies within a few feet beneath Cambrian schists is a gray finely banded biotite schist in which numerous narrow layers of light-colored granitic rock suggest an injection of a highly biotitic schist by a granitic magma.

A succession of quartzose and biotitic gneisses, in many places graphitic, underlies the biotite-albite schist. These rocks are interlaminated with fine-grained biotitic hornblende schists and amphibolites that appear to be dynamically altered basic sills or flows, although the formation of uralite is so complete that the original character of the rock can not be determined. The amphibolites and interlayered sediments are intruded by a light-colored rock of granitic texture that contains much saussuritized feldspar. The light-colored rock in many places near the upper part of the pre-Cambrian section is injected into the sedimentary series and the amphibolites, forming a banded lit-par-lit injection zone.

The biotite-albite schists are distinctly foliated rocks, composed of quartz, albite, biotite, and muscovite, with strong parallel dimensionalism. These constituents surround numerous lenticular areas of albite or albite and quartz. It is not easy to decide whether the rock was produced by the recrystallization of an original arkosic sandstone or by the injection of a monzonitic magma into a shaly sediment followed by strong metamorphism of the mixed rock. The albite, which is nearly pure (about $Ab_{95}An_5$) is crowded with fairly coarse inclusions of muscovite, biotite, and quartz. The whole rock shows a strong dynamic deformation superimposed upon an original crystalloblastic texture.

The quartzose and biotitic gneisses that underlie the biotite-albite schists are strongly foliated rocks in which layers of sutured quartz alternate with bands of biotite and muscovite. The crystalloblastic texture and the development of garnet, epidote, and biotite indicate a recrystallization under deep-seated conditions. Undulatory extinction and granulation of quartz show a cataclastic deformation subsequent to the recrystallization.

Arkosic beds of the pre-Cambrian gneissic series contain a considerable amount of plagioclase filled with muscovite and zoisite. The species of the plagioclase could not be accurately determined in the three sections studied, but the index and birefringence indicate the composition of a sodic oligoclase or albite.

As yet it is impossible to make a conclusive correlation between the Baltimore gneiss of the type locality in Baltimore, Md., and the pre-Cambrian gneiss of the Mine Ridge uplift, but the existing evidence seems to justify the belief that the core of Mine Ridge anticline is of pre-Glenarm age. About 12 miles southeast of Coatesville, between Pocopson and Chadds Ford, Brandywine Creek cuts through the Avondale-Kennett anticline, which exposes a quartzose biotite gneiss in the core and injection gneisses on the flank. The anticline is in the same line of strike as the anticlinal uplift of Baltimore gneiss north of Baltimore, Md., and the rocks of the anticline in Pennsylvania recall by their lithology the Baltimore gneiss of Maryland with its accompanying injection zone. Moreover, the sedimentary series on the flanks of the Avondale-Kennett uplift resembles in lithology and thickness the members of the Glenarm series in Maryland. The lowest formation overlying the biotite gneiss is a series of mica schists, mica gneisses, and quartzites of variable thickness, ranging from a few feet to at least 800 feet. It is overlain by crystalline marbles, also of variable thickness, and these in turn are overlain by garnetiferous mica schists and more or less quartzose mica gneisses.

Hawkins[15] is of the opinion that the sedimentary series which flanks the Avondale-Kennett anticline around Avondale is of the same age as the sedimentary series of Chester Valley. He bases his opinion on the belief that "the sedimentary succession and thickness of the various members in the Avondale series[16] are closely similar to those of the sedimentary series of accepted Cambro-Ordovician age in the adjacent Chester Valley," which he describes as a "basal quartzite followed by blue crystalline limestone, overlain by phyllite, all apparently conformable," and also on the belief that the apparent dissimilarity between the two series is the result of a regional contact metamorphism exerted by "a large buried intrusive mass below, whose roof cover has not yet been removed by erosion and which has but barely reached in its metamorphic action the series of the Chester Valley." In the first statement Hawkins makes the error of comparing the section of the sedimentary series at Avondale, not with the complete Paleozoic section around the Mine Ridge upland and northward, but with the incomplete section west of Coatesville, where a post-Beekmantown unconformity[17] has removed a considerable thickness of Cambrian and Lower Ordovician beds. As a result he ignores the fact that the full sequence of the Paleozoic sedimentary formations in the Chester Valley-Mine Ridge upland section is entirely different from that of the sedimentary series at Avondale, even if we accept his own estimate as to the thickness and structure of the latter series. For purposes of comparison two stratigraphic sections are given below; the first shows the section for the Glenarm series near Avondale, and the second shows the section for the sedimentary series in the vicinity of the Mine Ridge upland, as determined by the work of Stose, Jonas, and Knopf.

[15] Hawkins, A. C., Alternative interpretations of some crystalline schists in southeastern Pennsylvania: Am. Jour. Sci., 5th ser., vol. 7, pp. 355–364, 1924.

[16] "Avondale series" is used to denote "the Glenarm sedimentary series in the vicinity of Avondale, Chester County, Pa."

[17] Stose, G. W., and Jonas, A. I., Ordovician overlap in the Piedmont province of Pennsylvania and Maryland: Geol. Soc. America Bull., vol. 34, pp. 507–524, 1923.

Sections near Avondale and Mine Ridge anticline

Pre-Cambrian Glenarm series of Avondale		Paleozoic succession of Mine Ridge upland and northward[a]		
Formation	Thickness (feet) [b]	Thickness (feet)	Formation	Age
Wissahickon schist (correlated by Hawkins [c] with Ordovician phyllite).	(?)	1,000+	Cocalico shale. Unconformity.	Ordovician.
		500+	Conestoga limestone. Unconformity.	
		2,000±	Beekmantown limestone.	
		900±	Conococheague limestone.	Upper Cambrian.
Marble	600±	500±	Elbrook dolomite. Unconformity.	Middle Cambrian.
		1,000±	Ledger dolomite.	Lower Cambrian.
		150	Kinzers formation.	
		600-0	Vintage dolomite.	
		1,500	Antietam schist. Harpers schist.	
Quartzite	1,200-0	500±	Chickies quartzite, including Hellam conglomerate member (150 feet) at base.	

[a] Knopf, E. B., and Jonas, A. I., Stratigraphy of the crystalline schists of Pennsylvania and Maryland: Am. Jour. Sci., 5th ser., vol. 5, pp. 40-62, 1923.
[b] Hawkins, A. C., Alternative interpretations of some crystalline schists in southeastern Pennsylvania: Am. Jour. Sci., 5th ser., vol. 7, p. 357, 1924.
[c] Idem, p. 363.

The series around Avondale is a continuous stratigraphic sequence with no evidence of unconformity between the individual formations, although the younger deposits overlap the older in some places owing to the irregularity in deposition of the quartzite and the limestone. Thus in some places the marble overlies the Baltimore gneiss directly, and in others the Wissahickon formation overlies the Baltimore gneiss. The highly garnetiferous schist beds of the Wissahickon that overlie the Baltimore gneiss on the north flank of the Avondale-Kennett anticline are apparently the same beds that overlie the marble at all places where the contact between the marble and the Wissahickon formation is exposed. This suggests that the relation between the Wissahickon and the Baltimore gneiss where the marble and quartzite are absent is an overlap rather than a fault. The same irregularity in deposition of Cockeysville marble and Setters formation is seen in Maryland on the flanks of the Phoenix anticline in Baltimore County, where the Wissahickon overlies in some places the Cockeysville marble, in some the Setters formation, and in some the Baltimore gneiss.

Around the Mine Ridge upland, on the other hand, there is a marked unconformity,[18] as a result of which the Conestoga limestone, probably of Chazy age, overlies Lower Cambrian dolomites and quartzites. North and west of Coatesville the Conestoga limestone overlies the Lower Cambrian Harpers schist, but east of

[18] Stose, G. W., and Jonas, A. I., Ordovician overlap in the Piedmont province of Pennsylvania and Maryland: Geol. Soc. America Bull., vol. 34, pp. 507-524, 1923.

Coatesville the Conestoga limestone overlies the Upper Cambrian Conococheague limestone.[19]

If the marble at Avondale is the equivalent of the fossiliferous Vintage dolomite, which normally overlies the Lower Cambrian arenaceous and argillaceous rocks, the Wissahickon formation might be considered the stratigraphic equivalent of the Cocalico shale, which in the area north of Lancaster Valley overlies the Beekmantown limestone.[20] But the Vintage dolomite everywhere rests upon the Lower Cambrian arenaceous and argillaceous rocks and nowhere lies directly upon the Baltimore gneiss. Moreover, the aggregate thickness of the Wissahickon formation and the Peters Creek schist as exposed along Susquehanna River must be 8,000 to 10,000 feet, whereas the total thickness of all the Ordovician argillaceous sediments exposed south of Blue Mountain is probably of an order of magnitude of only 3,000 feet.

Nor could the marble at Avondale be explained as Conestoga limestone overlying Harpers schist, because although the absence of about 5,000 feet of calcareous sediments under the Conestoga and above the Harpers might be explained, as it is north and west of Coatesville, by erosion before the Conestoga was deposited, it is not so easy to explain the absence both of the Conestoga itself and of the Harpers schist in those places where the Wissahickon directly overlies the Baltimore. Nor is it easy to explain why the Cocalico shale, which nowhere overlies the Conestoga limestone north of Chester Valley, should overlie it around Avondale. Therefore "the sedimentary succession and thickness of the various members in the Avondale series" are not, as asserted by Hawkins, "closely similar to those of the sedimentary series of accepted Cambro-Ordovician age in the adjacent Chester Valley."

In the second statement concerning the metamorphic effects of a large buried intrusive mass, which has altered the series at Avondale but which has barely affected the series in Chester Valley, Hawkins ignores the fact that the Paleozoic rocks in the vicinity of the Mine Ridge upland are themselves highly metamorphosed and in that respect are strictly comparable to the rocks of the Glenarm series. The degree of metamorphism attained by the micaceous marbles of the Quarryville and Chester Valleys is analogous to that of the Cockeysville marble near Avondale. Therefore the dissimilarity between the two series can not be explained by a weakening of metamorphic action in the rocks of Chester Valley. The Paleozoic schists, quartzites, and marbles are cut by granitic intrusions, indicated by extensive pegmatitization, thus showing that there was igneous

[19] Stose, G. W., and Jonas, A. I., op. cit., p. 518.
[20] Idem, p. 521.

activity in the Paleozoic era, but the interlayered amphibolites that are characteristic of the pre-Cambrian core of the Mine Ridge upland occur throughout both the Glenarm series and the underlying Baltimore gneiss near Avondale and are absent in the Paleozoic sediments that flank the Mine Ridge upland. It seems evident, therefore, that the post-Glenarm basic intrusions must have been intruded prior to the deposition of the sediments flanking the Mine Ridge upland, because the amphibolitic layers occur with striking regularity and persistence interlaminated in the pre-Cambrian core of the Mine Ridge upland within a few feet of the base of the Cambrian. There is strong evidence for the pre-Cambrian age of the recrystallized sedimentary series near Avondale, which is stratigraphically continuous with the Glenarm series that underlies the basal Cambrian sediments in the western Piedmont belt of Maryland,[21] and it is reasonable to correlate the pre-Glenarm core of the Avondale-Kennett anticline with the Baltimore gneiss of Maryland.

The stratigraphic equivalence of the core of the Mine Ridge uplift and the Baltimore gneiss in the Avondale-Kennett anticline is difficult to prove because of the absence of the Glenarm series underlying the Paleozoic section on the flanks of the Mine Ridge upland. However, the pre-Cambrian sedimentary formations in the Mine Ridge anticline are strikingly similar in lithology to the Baltimore gneiss in the Avondale-Kennett anticline. About a mile north of Chadds Ford there is interwelded within the Baltimore gneiss a lentil of graphitic marble that is identical in structure and lithology with the lenticular areas of Franklin limestone which are in many places interbedded with the graphitic mica schists in the northeastward extension of the Mine Ridge upland known as the Honeybrook upland and in the graphitic schists of the Reading and Durham Hills and New Jersey Highlands. Therefore, it seems reasonable to assume that the metamorphosed sediments in the Mine Ridge upland are the equivalent of the Baltimore gneiss in the Avondale-Kennett anticline and are thus of pre-Glenarm age.

GLENARM SERIES IN PENNSYLVANIA

The oligoclase-mica schist facies of the Wissahickon formation, which overlies the Setters formation in Maryland, dips northwest on the northern flank of the Phoenix anticline north of Baltimore and passes under the Peters Creek schist. The Peters Creek schist underlies in apparent conformity Cardiff conglomerate and Peach Bottom slate in a tightly compressed syncline known as the Peach Bottom syncline. The Cardiff conglomerate and Peach Bottom slate crop

[21] Jonas, A. I., Pre-Cambrian rocks of the western Piedmont of Maryland: Geol. Soc. America Bull., vol. 35, pp. 355–364, 1924.

out in an area in the center of the syncline 18 miles long and half a mile to a mile in width, which extends from northern Maryland into southern Pennsylvania. The cleavage of the slate in the trough of the syncline is nearly vertical and is so strongly developed that the bedding is obscure. But it is possible to show that the bedding is parallel to the cleavage on the southeast limb of the fold, where the slate, for a distance of 100 feet, is clearly seen to be interbedded with the conglomerate. Farther northwest along Susquehanna River the Peters Creek schist forms the northwest limb of this compressed syncline and conformably overlies a wide anticline of albite-chlorite schist which appears from its position to be the stratigraphic equivalent of the Wissahickon oligoclase-mica schist that adjoins the Peters Creek schist on the southeast in Maryland.

COCKEYSVILLE MARBLE

About 1 mile south of Chanceford, York County, there is a small valley underlain by marble, which crops out only in a few isolated boulders. It is a coarsely crystalline micaceous marble with a considerable admixture of epidote that imparts a greenish color to the rock. No structure is discernible in the marble outcrops, but the rock shows a lithologic resemblance to the Cockeysville marble, which underlies the Wissahickon albite-chlorite schist farther southwest, in Carroll and Frederick Counties, Md. The Chanceford occurrence lies along the northern fold of the Mine Ridge-Tucquan anticline and is probably an anticline of Cockeysville marble exposed by the erosion of overlying Wissahickon albite-chlorite schist.

WISSAHICKON FORMATION

Oligoclase-mica schist facies.—The southern facies of the Wissahickon formation, which is characterized by strong intensity of metamorphism, occurs only in a small area in the extreme southeast corner of the McCalls Ferry-Quarryville district. It is well developed to the east and northeast of the district, along the eastern boundary of the Piedmont province northeastward to Trenton, N. J., and was named by Bascom from its fine exposure along the banks of Wissahickon Creek in Philadelphia. In the small area of oligoclase-mica schist in the McCalls Ferry-Quarryville district outcrops of fresh rock are almost lacking. A few isolated exposures show the typical fine-grained "pepper and salt" biotite gneiss and the white quartzite of the strongly metamorphosed facies. The mica schist beds, dominantly composed of biotite and muscovite, are almost entirely lacking in fresh outcrop, but their presence is indicated by the characteristic soil derived from the mica schist, a deep, unctuous clay, sparkling brilliantly with numerous minute flakes of mica.

The gneissic layers are intensely metamorphosed, consisting chiefly of quartz, biotite, and plagioclase, with some orthoclase. The plagioclase is oligoclase ($Ab_{70}An_{30}$). Coarse-grained layers in which quartz predominates alternate with fine-grained layers made up of biotite, plagioclase, and small interlocking grains of interstitial quartz. Strain shadows in the quartz indicate a certain amount of deformation subsequent to the recrystallization, but the rock in this area does not show the strong cataclastic action that has affected the Wissahickon formation south of Chadds Ford, 20 miles farther east. Characteristic accessories are garnet, apatite, zircon, and magnetite. Many of the garnets are poikilitic porphyroblasts containing quartz inclusions; where the garnets have been broken the fragments are healed by secondary quartz.

The quartzitic layers crop out about half a mile south of Tweedale, on the road that follows the left bank of Tweed Creek, where the rock is composed chiefly of recrystallized quartz, a little fine-grained biotite, and a small amount of feldspar.

The mica schist beds of the Wissahickon are finely plicated and coarse to medium grained. Biotite, muscovite, and quartz are dominant, associated with a variable content of oligoclase. The mica occurs in large blades that spangle the cleavage surfaces and wrap around lenticular areas of quartz or large pink garnets, some of which attain half an inch in diameter. Under the microscope the blades of mica are seen to be much twisted and frayed. The garnet is usually very abundant, standing out in knobs upon the cleavage surface of the weathered rock. In Maryland a notable feature of the mica schist is the abundant development of staurolite and kyanite. The rapid alternation of beds in the deposits that make up the Wissahickon formation makes it impossible to establish any bed as a key horizon. Although in numerous localities the garnetiferous schist overlies the Cockeysville marble, the lithologic variation is so great that the absence of garnetiferous schist over the marble can not be considered an evidence of faulting. In the gneissic and schistose members of the Wissahickon there is a well-marked foliation that is parallel to the bedding, accompanied by a secondary strain-slip cleavage that is parallel to the drag folds.

The oligoclase-mica schist shows, by its complete recrystallization and by its content of such minerals as garnet, staurolite, kyanite, and sillimanite, that the present condition has been developed by metamorphic agencies characteristic of a deep-seated zone. That the original material was a sediment may be inferred from the heterogeneous character of the formation and the presence of abundant rounded grains of zircon, suggestive of waterworn grains; from the occurrence of mineral high in alumina, such as staurolite, kyanite, and sillimanite; and from the lack of correspondence between the chemical composition of the rock and any known igneous type. The development of tourmaline along a certain well-defined zone indicates that the rock has been subjected to pneumatolytic action subsequent to its deep-seated metamorphism, and this conclusion is supported by the large amount of secondary quartz that penetrates the Wissahickon formation in lenses and stringers. The undulatory extinction in the quartz and feldspar is the effect of a stress to which the rock has been subjected subsequent to the formation of its metamorphic minerals.

Diaphthoritic schists.—In the vicinity of Octoraro Creek below the confluence of the East and West Branches there are several outcrops of a peculiar schist that indicates diaphthoritic metamorphism.[22] On Black Run, 1½ miles east of its confluence with Octoraro Creek, there is a fine example of such a diaphthorite. The rock is a garnetiferous schist, in many respects identical with the typical garnetiferous mica schist of the Wissahickon. However, in this rock the biotite of the typical deep-seated schist has been partly replaced by chlorite. The chlorite fibers are intimately intertwined and show in places remnants of the original brown color and polarization of biotite. Associated with the chlorite is a large amount of iron oxide liberated from the original biotite during the formation of chlorite. Another instance of the breaking down of a mineral formed under deep-seated conditions is seen in the garnet porphyroblasts, which are surrounded by rims of chlorite, showing that the chemical constituents of the garnet have adapted themselves to new conditions of metamorphism by the formation of chlorite, which is stable under the new environments. Such diaphthorites have hitherto been described only in Europe, where they occur in the lower part of the overriding block in areas of overthrusting. They are called by Suess[23] "deep-seated diaphthorites," because the retrogression in metamorphism induced by friction along the fault plane has taken place under deep-seated conditions. In this connection it is significant that three specimens of a cataclastic rock collected near Glenroy suggest a mylonitized quartz conglomerate by the presence of large lenticular areas of granulated quartz in a matrix of chlorite, muscovite, quartz, a little biotite, and abundant iron oxide. The southwestward extension of the diaphthoritic zone is interrupted by igneous masses that are of wide extent in the vicinity

[22] Diaphthorite is a term that was introduced by Becke to describe a schist that has been recrystallized under deep-seated conditions and later carried into an upper tectonic zone, where it has undergone a reversal of metamorphism, indicated by certain new minerals that are characteristic of the upper zone of metamorphism and that have been formed at the expense of the heavier minerals of the deep-seated zone. Becke, F., Über Diaphthorite: Min. pet. Mitt., Band 28, pp. 369–375, 1909.

[23] Suess, F. E., Moravische Fenster: K. Akad. Wiss. Wien, Math.-naturwiss. Klasse, Anzeiger, Jahrgang 47, pp. 428–432, 1910.

of the Maryland-Pennsylvania State line. The diaphthorite can not be traced in these igneous rocks, but still farther south, in Cecil County, Md., along Basin Run 1 mile south of Liberty Grove, Grimsley [24] and Bascom [25] have found what is apparently the reappearance of the Pennsylvania diaphthorites. The rock here is a garnetiferous-staurolite mica schist in which the garnets have partly altered to chlorite and the staurolite has almost entirely altered to muscovite. One mile southeast of Liberty Grove E. B. Knopf has found a brecciated rock in which numerous fragments of biotite schist and biotitic quartzite 1 to 2 inches in length stand out upon the weathered surface in long prismoids that are arranged roughly parallel to the plications of the schistose matrix. The rock is thickly studded with garnets and is penetrated by quartz veinlets.

East of the Quarryville district, in the West Chester quadrangle, Knopf has found a narrow belt of mylonitized schists, which are dense, schistose dark-colored rocks interleaved with paper-thin layers and elongated lenses of feldspathic material. The rock occurs near the contact between Baltimore gneiss and Wissahickon formation on the southern flank of the Avondale-Kennett uplift. The Wissahickon formation for at least a mile southward from the belt of mylonitization is a highly cataclastic, sillimanite-bearing oligoclase-biotite schist filled with lenses and stringers of pink granite material. The features described suggest strong mechanical deformation by which a biotite schist injected with granitic material was first mashed and finally rolled out so that the original constituents have been pulverized. Although the inference is strong that the presence of these diaphthoritic schists indicates a zone of overthrusting, it has so far been impossible to establish the overthrust by any other evidence.

Albite-chlorite schist facies.—The albite-chlorite schist facies of the Wissahickon formation is widespread on the northwest side of the Peach Bottom syncline, where it underlies almost half of the McCalls Ferry quadrangle. It is particularly well exposed along Susquehanna River, where the steep river gorge is cut into rugged, heavily wooded cliffs of albite schist for almost 17 miles between Turkey Hill and Fites Eddy. As in the southern oligoclase-mica schist facies of the Wissahickon, individual beds vary widely in lithologic character. The formation as a whole comprises a porphyroblastic feldspathic type and a micaceous schistose type, with a subordinate amount of quartzose gneiss in both types.

The first type, which is in some places massive and almost igneous in appearance, is characterized by poikilitic metacrysts of albite,

[24] Grimsley, G. P., The granites of Cecil County, in northeastern Maryland: Cincinnati Soc. Nat. Hist. Jour., vol. 17, pp. 66–67, 1894.

[25] Bascom, F., The crystalline rocks of Cecil County: Maryland Geol. Survey, Cecil County, p. 115, 1902.

embedded in a schistose matrix that is made up of chlorite and muscovite in varying proportion. The lower beds exposed in the core of the anticline between Safe Harbor and McCalls Ferry contain much biotite in deep-brown, strongly pleochroic individuals, some of which are intergrown with chlorite. The chlorite is a grass-green, strongly pleochroic clinochlore, and clinochlore, biotite, and muscovite alike contain numerous fine pleochroic haloes surrounding inclusions of zircon. The albite metacrysts are dull white to pinkish, some are as large as 10 millimeters in diameter, and they are most conspicuous on rough surfaces normal to the plane of schistosity. They occur in subhedral or anhedral individuals with ragged boundaries and commonly show Carlsbad twins, rarely polysynthetic twinning. They inclose poikilitically numerous rather large inclusions of garnet, epidote, magnetite, biotite, muscovite, calcite, and rarely tourmaline. The inclusions generally show a certain alinement, which is parallel to the schistosity of the groundmass in some specimens and inclined to it in others. (See pl. 2, *A*.) Out of eight thin sections studied only one showed a random arrangement of the inclusions. The inclined relation of schistosity of the groundmass to the alinement of the inclusions in the metacrysts has been considered by some writers to indicate that the metacrysts have been formed after the crystallization of the groundmass has ceased and that they have rotated during the later stages of their formation. In support of this view Bailey [26] has figured albites from the albite schists of the Scottish Highlands, in which the ends of the linear inclusions are curved in opposite directions, thus presumably indicating the direction of rotation. Such a curvature in the alinement of the inclusions has been interpreted by Clough [27] to show that the albite has been formed later than the main schistosity of the groundmass, which is fossilized, as it were, in the winding stream of inclusions that represents the original puckered foliation planes of the rock. The inclusions in the albite metacrysts of the Wissahickon schist show no evidence of crumpling or distortion in the linearity. It is hard to escape the conclusion that the alinement is caused by a rotation of the metacrysts by which the albites, in a final stage of growth, adjusted themselves in a certain way to their physical environment. However, it is difficult to understand why this adjustment to environment did not keep pace with the development of the foliation, manifesting itself in an elongation of the metacryst parallel to the foliation plane rather than by a bodily movement of the individual. The fact that some of the albite metacrysts are so elongated in response to the stress under which they

[26] Bailey, E. B., The metamorphism of the southwest Highlands: Geol. Mag., vol. 60, p. 329, 1923.

[27] Gunn, W., Clough, C. T., and Hill, J. B., The geology of Cowal (Scotland Geol. Survey Mem.), p. 40, 1897.

were formed suggests that the rotation of the metacrysts may be connected with a second period of metamorphism subsequent to that which produced foliation parallel to the bedding. The existence of such a later deformation is clearly indicated by the presence of a fine foliation that crosses the first foliation at a steep angle. This secondary foliation is best developed in the soft phyllitic beds, where it corresponds to the direction of the minute drag folds. It is excellently exposed on the north flank of Turkey Hill in the cuts of the Pennsylvania Railroad along Susquehanna River. The rock is filled with twisted and broken veinlets of quartz, which are in some places cut across by quartz veins of a later age than the deformed veins.

The schistose type of the albite-chlorite schist is characterized by an abundance of chlorite, which is deep-green clinochlore, and a variable quantity of sericite. Albite is almost universally present, but is subordinate in amount to the micaceous minerals, so that the rock is highly foliated.

Quartz is fairly abundant in both types. It fills the interstices between the albite, chlorite, and mica, and in some of the schists it occurs in layers, alternating with micaceous bands. Characteristic accessories are magnetite, zircon, carbonates (calcite or dolomite), and tourmaline. Other common constituents are pyrite, epidote, garnet, apatite, titanite, and hematite. The carbonate is in many specimens abundant and appears to have crystallized at the same time as the other constituents. Magnetite is of universal occurrence and in many places forms large porphyroblasts that show under the microscope a distinctly ragged outline.

The quartzose members of this formation occur in subordinate amount. Quartzite beds that probably belong in the Wissahickon albite-chlorite schist are well exposed in a railroad cut along Brandywine Creek in Coatesville, where they occur as layers from 4 to 8 feet in thickness, containing numerous closely cemented grains of bluish quartz. Quartzite layers in the albite-chlorite schist occur also at Martic Forge and at several places in the section exposed along Susquehanna River.

Both facies of the Wissahickon formation weather readily into a micaceous clay soil, which has the consistency of putty when wet and holds water for a long time. This clay forms deep ruts, thereby making roads that are almost impassable in rainy weather. The unctuous clay soil derived from the oligoclase-mica schist sparkles brilliantly, owing to the presence of numerous mica flakes. The soil derived from the albite-chlorite schist shows less mica and is full of silvery schist fragments.

Composition.—The chemical composition of the two facies is illustrated by the following analyses:

Chemical composition of schists and gneisses from the Wissahickon formation

	1	2	3	4	5		1	2	3	4	5
SiO_2	70.26	66.13	79.60	77.00	54.99	H_2O-	} 1.10	1.55	} 1.66	0.01	} 5.10
Al_2O_3	12.87	15.11	9.48	10.17	18.74	H_2O+		.24		1.18	
Fe_2O_3	2.61	2.52	1.77	2.41	2.90	TiO_2	1.85	.82	.71	.96	1.66
FeO	2.70	3.19	1.49	2.35	5.49	CO_2	----	----	----	.08	.18
MgO	1.49	2.42	.76	1.26	2.81	P_2O_5	.11	.22	.19	.10	.62
CaO	2.19	1.87	.72	2.02	.85	MnO	.05	.20	.67	.09	.04
Na_2O	2.55	2.71	1.83	1.14	1.68						
K_2O	2.15	2.86	1.54	1.94	5.36		99.93	99.84	100.42	100.71	100.42

1. Wissahickon formation (oligoclase-mica schist, gneissic variety, with fine schist beds and quartz veinlet), 1 mile northwest of Monkton, Md. Analysis made for Maryland Geological Survey by Penniman & Browne. Represents an arkosic sediment high in lime. Rock is composed of quartz, oligoclase, and biotite, with small amount of orthoclase and muscovite.
2. Wissahickon gneiss, composite sample from four different localities. Bascom, F., U. S. Geol. Survey Geol. Atlas, Philadelphia folio (No. 162), p. 4, 1909. Analyst, W. F. Hillebrand. Represents an arkosic sediment.
3. Wissahickon gneiss (quartzose variety), Elm (Narberth) station, Pennsylvania Railroad, Montgomery County, Pa. Idem, p. 135. Analyst, F. A. Genth, jr. Represents a sandstone high in soda. Rock is described as containing quartz, feldspar, and a little mica.
4. Wissahickon formation (fine-grained biotite gneiss). Rock is composed of quartz, oligoclase, and biotite.
5. Wissahickon formation (albite-chlorite schist, schistose variety), 1 mile south of Shenks Ferry station, Columbia & Port Deposit branch, Pennsylvania Railroad, Lancaster County, Pa. Analyst, R. K. Bailey, U. S. Geological Survey, 1920. Represents a potassic argillite. Rock is composed of muscovite, clinochlore, quartz, and albite.

The mineral composition of these rocks has been calculated from the chemical analyses in such a way as to correspond as nearly as possible with the mineral constituents of the rocks determined under the microscope.

Mineral composition of schists and gneisses from the Wissahickon formation

	1	2	3	4	5
Quartz	40	32	62	54	21
Plagioclase	$Ab_{70}An_{30}$ 31.5	$Ab_{76}An_{24}$ 31	$Ab_{85}An_{15}$ 18	$Ab_{72}An_{28}$ 15	Ab_{100} 14
Orthoclase	7.5	1.5	----	----	----
Muscovite	----	----	10	12	45
Biotite	13	25	4	15	----
Chlorite	----	----	----	----	13
Garnet	----	4	----	----	----
Titanite	----	.5	----	2	----
Magnetite	----	3	3	3	4
Ilmenite	2.5	----	1	----	3
Hematite	2.5	----	----	----	----
Apatite	----	----	----	----	1
	97.00	97.00	98.00	101	101

The variable content of Na_2O and K_2O in these analyses is such as would be expected from the heterogeneous nature of the original sediment. The 5.36 per cent K_2O of the albite-chlorite schist is high for a normal shale. In the computation of mineral composition this potash is recalculated as 45 per cent muscovite, which agrees very well with the content of muscovite estimated in the microscopic section. The computation of 45 per cent muscovite that is demanded by the high potash content and by the absence of biotite and orthoclase requires more alumina than is indicated by the analysis, and it seems probable that the alumina content of the analysis is a little low. Normal shale is calculated to contain an average of 3.60 per cent K_2O, as determined by a composite analysis of 51 samples of Paleozoic shale weighted in reference to the mass of the formation represented.[28] The high potash content of the albite-chlorite schist is not, however, incompatible with a sedimentary origin, because some sediments and slates of known sedimentary origin contain an unusually large amount of potash. The slate in the Cartersville formation of Georgia, for example, shows as much as 8.26 per cent K_2O in an average of 59 analyses.[29] Therefore although the potash content of the muscovitic variety of the Wissahickon albite-chlorite schist is high, it does not necessarily denote an igneous origin for the schist but rather indicates that the original sediment was a feldspathic mud rich in potash feldspar and white mica and similar in origin to the highly potassic sediments of the Setters formation in Maryland and of the Harpers schist in Pennsylvania.

Metamorphism.—The nature of the mineral constituents in the two facies of the Wissahickon formation shows that the metamorphism of the oligoclase-mica schist took place under conditions of strong intensity and that the albite-chlorite schist was recrystallized in an upper zone of metamorphism. The lithologic characteristics of the two facies are essentially similar, in spite of the heterogeneity that is to be expected in widespread sedimentary deposits. In both facies mica schist alternates with more feldspathic gneisses, and quartzites or quartzose gneisses occur in subordinate amounts. The difference between the two facies appears, therefore, to be conditioned by a difference in intensity of metamorphism rather than by difference in stratigraphic position or in chemical composition. The constituents that have formed oligoclase under conditions of intense metamorphism have crystallized under more moderate metamorphic intensity as albite and calcite.

The gradation from rocks that were formed under moderate metamorphic intensity to rocks that have been formed by an intense metamorphism is well shown about 12 miles east of the McCalls Ferry-Quarryville district, where the Wilmington division of the Philadelphia & Reading Railroad follows the West Branch of Brandywine Creek southward from Coatesville to Embreeville. Within the borough limits of Coatesville, on the west bank of the West Branch of Brandywine Creek and along the north face of the South Valley Hills, there is a zone of much shattered rock that shows an alternation of quartzose and slaty beds. This rock overlies the Conestoga limestone in Chester Valley. From the lithology it is hard to determine whether the rock belongs in the upper part of the Wissahickon albite-chlorite schist or in the Peters Creek

[28] Clarke, F. W., Analyses of rocks and minerals, 1880 to 1914: U. S. Geol. Survey Bull. 591, p. 23, 1915.

[29] Knopf, E. B., and Jonas, A. I., Geology of the crystalline rocks of Baltimore County: Maryland Geol. Survey, Baltimore County (in press).

formation. Metamorphism, which has been of low intensity, has not progressed far enough to destroy the original clastic grains of blue quartz in the quartzose beds, and the argillaceous layers have scarcely passed the stage of slates. About half a mile south of Coatesville this strongly shattered zone is succeeded by a belt about 1,200 feet wide of steeply dipping slates and slaty schists, impregnated with quartz veinlets. A mile farther south the rock is a chloritic quartz schist, and about 5 miles southeast of Coatesville it has become a quartzose muscovite schist or an albite-chlorite schist. Metamorphism has advanced but is still of moderately low intensity. Still farther south the rock becomes coarser grained, and biotite begins to appear in association with the muscovite. From this point southward the rock gradually assumes the lithologic characteristics of the Wissahickon oligoclase-biotite schist, and the metamorphic intensity continues to increase until, half a mile north of Embreeville and 7 miles south of Coatesville, the rock shows every evidence of metamorphism under conditions of strong intensity. It is a fairly coarse-grained, highly garnetiferous muscovite-biotite schist and contains short crumpled augen of quartz that show no connection with channels through which siliceous solutions might penetrate the rock. Such detached lenses of quartz have been explained by Holmquist as a result of segregation from the surrounding rock during the ultra-metamorphism of schistose and folded gneisses.[30] In the Wissahickon mica schists, however, there is no indication of ultra-metamorphism, which implies the vanishing of the schistose structure. As the whole Wissahickon formation is traversed by innumerable veinlets of tourmaliniferous quartz and feldspathic quartz, it is evident that the rock has been permeated by pneumatolytic vapors at some time during its metamorphism, and the quartz augen described above have probably been introduced in connection with pneumatolysis. It is clear from the section between Coatesville and Embreeville that there is a progressive increase of metamorphism from the rock overlying the Conestoga limestone in Coatesville, which is probably upper Wissahickon albite-chlorite schist, through an apparently synclinal area where the rock has the lithologic characteristics of the Peters Creek formation to an underlying mica schist, which becomes, north of Embreeville, the intensely metamorphosed facies of the Wissahickon.

Age.—In the earlier work in this region the two mineralogic facies of the Wissahickon were considered parts of the same formation, but at that time the albite-chlorite schist that underlies the South Valley Hills was thought to be Ordovician because it appeared to overlie

[30] Holmquist, P. J., Swedish Archean structures and their meaning: Geol. Inst. Upsala Bull. 15, pp. 141, 142, 1916.

conformably Beekmantown limestone in Chester Valley, the limestones of Chester Valley being at that time all mapped under the name Shenandoah.[31] Later the southern facies of the Wissahickon, the oligoclase-mica schist, was called Wissahickon mica gneiss and shown to be pre-Cambrian.[32] The rock underlying the South Valley Hills was still assigned to the Ordovician and called "Octoraro schist," and the name "Octoraro" was used for all the rock in the area lying between the formation then mapped as Wissahickon gneiss and the limestone of Chester Valley. The "Octoraro schist" thus comprised both what is now mapped as the albite-chlorite schist facies of the Wissahickon formation and part of the overlying Peters Creek formation. Part of what is now mapped as Peters Creek was then called Wissahickon gneiss.

The interpretation of the "Octoraro schist" as Ordovician led to grave difficulties when it was discovered that the Wissahickon gneiss is pre-Cambrian, because the rock mapped as "Octoraro schist" grades imperceptibly into the Wissahickon gneiss. It was necessary to assume an overthrust fault along the contact between the Wissahickon gneiss and the "Octoraro schist" in order to explain the anomalous position of Wissahickon gneiss on top of the so-called Ordovician strata south of Chester Valley. The presence of the fault was apparently demanded by the stratigraphy, but no structural evidence of such a fault could be found.

Later work by the writers both in Maryland and in Pennsylvania, along the Susquehanna River section, shows that the pre-Cambrian Wissahickon gneiss north of the Phoenix anticline in Maryland, where the formation is an oligoclase-biotite schist, dips under a synclinal fold of Peters Creek schist and reappears from under the northern limb of the syncline in a different mineralogic facies, the albite-chlorite schist.[33] It was thus shown that the "Octoraro schist" is not a stratigraphic unit and that the formations which made up the old "Octoraro schist" are part of a conformable sequence of pre-Cambrian rocks known as the Glenarm series. Confirmation of the pre-Cambrian age of this series is the discovery by Jonas that basal Cambrian rocks rest unconformably upon the volcanic rocks and the Cockeysville marble of the Glenarm series in Frederick County, Md.,[34] and upon Wissahickon albite-chlorite schist in Carroll County, Md., and southern York County, Pa.

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[31] Bascom, F., Piedmont district of Pennsylvania: Geol. Soc. America Bull., vol. 16, p. 305, 1905.

[32] Bascom, F., U. S. Geol. Survey Geol. Atlas, Philadelphia folio (No. 162), p. 3, 1909.

[33] Knopf, E. B., and Jonas, A. I., Stratigraphy of the crystalline schists of Pennsylvania and Maryland: Am. Jour. Sci., 5th ser., vol. 5, pp. 40–62, 1923.

[34] Jonas, A. I., Pre-Cambrian rocks of the western Piedmont of Maryland: Geol. Soc. America Bull., vol. 35, p. 363, 1924.

It thus became evident that the contact between the pre-Cambrian Wissahickon albite-chlorite schist (formerly mapped as "Octoraro schist") and the so-called "Cambro-Ordovician" Shenandoah limestone of Chester Valley could not be conformable. The establishment of the lower Paleozoic stratigraphic column in the Lancaster, Quarryville, and Chester Valleys [35] showed that the Shenandoah limestone comprises several different formations ranging in age from Lower Cambrian dolomite to Ordovician limestone of probable Chazy age. Moreover, it was discovered by Stose and Jonas that the Ordovician Conestoga limestone unconformably overlies the Beekmantown.[36] Therefore the albite-chlorite schist underlying the South Valley Hills must be faulted over the Paleozoic rocks in Chester Valley. The existence of the overthrust is difficult to demonstrate on account of the fact that the fault plane itself has been seen in only a few localities, and there the structural relations are obscure. No mylonite has been observed in the few localities where the overthrust plane is visible. The absence of brecciated and mylonitized rock along such a plane of movement is not as surprising as it might at first seem, because a strong metamorphism has succeeded or accompanied the overthrusting. Mylonites if ever present may have been obscured by this later metamorphism. It is more probable, however, as suggested by Bailey [37] in his paper on the Scottish Highlands, where there is a similar absence of mylonites along a plane of overthrust, that the overthrusting took place in a deep zone where the temperature was sufficiently high to allow recrystallization to take place at the same time as the overthrusting, hence mylonites would not be formed. But the absence of structural evidence of faulting along the overthrust plane is offset by the fact that the writers have discovered, in the hills west of Quarryville and north of Shenks Ferry, Wissahickon albite-chlorite schist and Peters Creek schist overlying previously folded rocks, the Vintage dolomite, of Lower Cambrian age, and the Conestoga limestone, of Ordovician age. This discordant contact of Glenarm rocks against folded strata of different stratigraphic horizons is structural evidence of the overthrust relation of the pre-Cambrian rocks to the underlying Paleozoic formations.

PETERS CREEK SCHIST

The Peters Creek schist underlies a belt about 6 miles wide that covers most of the southeast corner of the McCalls Ferry-Quarry-

[35] Stose, G. W., and Jonas, A. I., The lower Paleozoic section of southeastern Pennsylvania: Washington Acad. Sci. Jour., vol. 12, pp. 358–366, 1922.

[36] Stose, G. W., and Jonas, A. I., Ordovician overlap in the Piedmont province of Pennsylvania and Maryland: Geol. Soc. America Bull., vol. 34, pp. 507–524, 1923.

[37] Bailey, E. B., The metamorphism of the southwest Highlands: Geol. Mag., vol. 60, p. 317, 1923.

ville district. It overlies the southeast flank of the wide double-crested anticline of Wissahickon albite-chlorite schist that crosses Susquehanna River at Tucquan, is eroded from the crest of the anticline, and reappears on the northwest flank in a narrow belt on the north side of Turkey Hill and Long Level. The southeastern belt of Peters Creek schist contains the narrow synclinal area that is underlain by Cardiff conglomerate and Peach Bottom slate.

The Peters Creek schist was named by the writers on account of its typical exposures along Peters Creek, which flows into Susquehanna River at Peach Bottom, Lancaster County. The formation has been traced, with unchanged characteristics, along the strike from Pennsylvania southwestward into Maryland across Baltimore and Carroll Counties. It comprises part of what was originally mapped as Wissahickon gneiss and part of the "Octoraro schist."

The formation consists of a series of chloritic and sericitic quartz schists interbedded with chlorite-sericite schist and grading toward the top into mildly metamorphosed quartzose and conglomeratic sediments. The sericitic and chloritic schists are most abundant in the lower beds that overlie the Wissahickon albite-chlorite schist. These lower beds of the Peters Creek schist are well exposed along Susquehanna River between Fites Eddy and Fishing Creek Station, where the rock is a light greenish-gray straight-bedded fine-grained quartz-chlorite schist. The banding is produced by an alternation of quartzose layers with thin layers of chlorite and sericite, which give a silvery luster on planes parallel to the bedding. The lowermost beds in which chlorite-sericite schist predominates are difficult to separate from the schistose beds of the Wissahickon albite-chlorite schist, which immediately underlies the Peters Creek. In fact, there is so perfect a lithologic gradation between the upper beds of the Wissahickon and the lower beds of the Peters Creek that the boundary between the two formations can not be accurately traced in the well-cultivated upland country above the valley of the Susquehanna. It has been represented on the map by a dotted line. Under the microscope the sericitic and chloritic quartz schists of the Peters Creek formation are seen to be almost completely recrystallized. The quartz, which predominates, is fine grained and clear, with local inclusions of rutile in minute rosettes. A fine foliation is imparted by the sericite and chlorite (a deep-green clinochlore). Albite, which is usually untwinned, is fairly abundant, amounting to about 20 per cent, and is very difficult to distinguish from quartz. Measurements on sections perpendicular to an optic axis indicate a species about $Ab_{95}An_5$. It differs from the albite of the Wissahickon albite schist in that it is finer grained, not noticeably porphyroblastic, and not spongiform. Magnetite and titanite are abundant accessories, and apatite, pyrite, tourmaline, and a carbonate (presumably dolomite) are of common occurrence.

In the upper members of the Peters Creek formation recrystallization is not complete, and there are numerous clastic grains of clear blue quartz scattered throughout the rock. In some places toward the top of the formation these quartz pebbles become so abundant that the rock might almost be called a sandstone. Under the microscope these quartz pebbles show up plainly, as deformed lenticular areas that are in some places clear, with slight strain shadowing, and in other places completely granulated. The argillaceous matrix surrounding the pebbles is completely recrystallized to chlorite or sericite. Some of the clastic fragments are feldspar, both a sodic albite and a microcline. The scarcity of microcline in the lower beds of the Peters Creek is probably due to the completeness of the recrystallization, which has taken place under a low metamorphic intensity, thereby using the potash of the microcline to form sericite.

The chemical and mineral composition of the lower schistose beds of the Peters Creek formation is given below.

Analysis of Peters Creek schist, Benton, Lancaster County, Pa.

Chemical composition [Glenn V. Brown, analyst]		Mineral composition	
SiO_2	73.00	Quartz	42.84
Al_2O_3	10.55	Albite ($Ab_{95}An_5$)	29.44
Fe_2O_3	2.42	Orthoclase	8.34
FeO	2.75	Chlorite	6.72
MgO	1.55	Muscovite	5.57
CaO	1.30	Magnetite	3.02
Na_2O	3.32	Titanite	2.16
K_2O	2.10	Dolomite and apatite	.89
H_2O+	2.01		98.98
TiO_2	.90		
CO_2	.25		
P_2O_5	.07		
MnO	.05		
	100.27		

CARDIFF CONGLOMERATE

The Cardiff conglomerate directly overlies the Peters Creek schist. It occurs in a belt of variable thickness about 18 miles long, extending southwestward from Fairmount, in Lancaster County, Pa., to Pylesville, in Harford County, Md., and surrounds an elliptical area of black roofing slates about 6 miles long and half a mile wide. The name Cardiff has been given to this conglomerate by the Maryland Geological Survey from the town of Cardiff, on the Maryland State line, which is the direct continuation of the town of Delta, in Pennsylvania, well known as a slate locality. In the lower beds of the conglomerate there is an alternation of schistose and conglomeratic layers, marking the transition from the Peters Creek pebbly sandstones to the Cardiff sediments. The Cardiff conglomerate, like the Peters Creek sandstones, is characterized by the predominance of quartz in the pebbles. Most of the pebbles are completely granulated and in many places are deformed and considerably elongated. Some of the pebbles, however, show a core of clear glassy quartz, commonly of a bluish tinge, and some of the smaller pebbles are clear all the way through. In one place a granulated quartz contained within it an albite crystal, which suggests that the detrital material was derived from an albitic igneous rock. The matrix of the conglomerate is schistose, composed of fine quartz and fibers of sericite and chlorite. Chloritoid, which occurs rather abundantly in several specimens, is in coarser fibers than the muscovite and is packed closely around the quartz pebbles. Magnetite is very abundant in some specimens, and titanite commonly occurs. In some places the magnetite is associated with an opaque whitish mineral, probably leucoxene.

The Cardiff conglomerate was for some time tentatively considered to be Ordovician and to rest upon so-called Ordovician phyllite and to underlie the Peach Bottom slate, then considered of doubtful Ordovician age.[38] The grounds for belief in its Ordovician age were made questionable when the phyllite was shown to be pre-Cambrian. The Cardiff conglomerate resembles the Lower Cambrian Hellam conglomerate member of the Chickies quartzite on the Mine Ridge upland, and on lithologic grounds it might be interpreted as basal Cambrian. However, the Cardiff conglomerate is overlain by more than 900 feet of black slates, and the Hellam conglomerate on the Mine Ridge upland and vicinity is overlain by more than 350 feet of white quartzite.

The Lower Cambrian series exposed in the area south of Lancaster Valley resembles the Cardiff conglomerate and Peach Bottom slate sequence more closely than it does the succession of Cambrian strata on the flanks of the Mine Ridge upland. Jonas has studied the area south of the Lancaster-York Valley in Pennsylvania and in its southwestward extension in Maryland and has found that the basal Cambrian conglomerate is overlain by black slate containing thin quartzite layers. This slate and quartzite series represents the quartzite member of the Chickies quartzite of the area that lies north of Chester Valley, where the siliceous beds are thicker than in the area south of Chester Valley.

[38] Mathews, E. B., Correlation of Maryland and Pennsylvania Piedmont formations: Geol. Soc. America Bull., vol. 16, p. 338, 1905.

The only places known in Pennsylvania or Maryland where recognized basal Cambrian is preserved on the overthrust mass are in the Sugarloaf syncline of Frederick County, Md., and in the Wentz syncline of Carroll County, Md., and York County, Pa.[39] There Lower Cambrian rocks rest unconformably upon Cockeysville marble and volcanic flows and Wissahickon albite-chlorite schist, of lower Glenarm age. The fact that the upper part of the Glenarm series had been removed in Maryland before the deposition of the basal Cambrian indicates deeper pre-Cambrian erosion in the western part of the Piedmont than in the southeastern area.

In Pennsylvania there is no difference in trend between the Cardiff conglomerate and Peach Bottom slate and the underlying Peters Creek formation, and there is an apparent gradation from fine pebbly sandstones in the upper beds of the Peters Creek formation through coarser beds into the Cardiff conglomerate. Therefore, if a depositional break occurred between the two formations the field evidence indicates that it was not an angular unconformity. In the absence of any definite proof that it is Cambrian, the Cardiff conglomerate has been included in the pre-Cambrian Glenarm series.

PEACH BOTTOM SLATE

The Cardiff conglomerate grades upward through beds of alternating conglomerate and slate into roofing slates, which form an area about 6 miles long and half a mile wide. These slates, which are well known to the trade as the Peach Bottom "black roofing" slate, have a probable thickness of 1,000 feet and show a uniform steep south dip, which coincides with the cleavage. On its southeast boundary the slate is distinctly interbedded with the conglomerate, and here it is possible to show that bedding and cleavage are parallel. The section measured from north to south on the road north of Peach Bottom station is as follows:

Section north of Peach Bottom station

	Feet
Black slate	45
Slate with interbedded conglomerate, 3 inches to 1 foot thick	15
Slate	20
Conglomerate	60

The Peach Bottom slate is dark bluish gray, nonfading, and of a strong luster on cleavage surfaces. It has a high crushing strength, 11,260 pounds to the square inch, and is exceedingly tough. The microscope shows a distinctly crystalline texture approaching that of the phyllitic stage of metamorphism. The essential constituents

[39] Jonas, A. I., Pre-Cambrian rocks of the western Piedmont of Maryland: Geol. Soc America Bull., vol. 35, p. 361, 1924.

are muscovite, quartz, andalusite, and graphite with variable amounts of magnetite and pyrite. The following table shows the chemical analyses of four specimens of Peach Bottom slate. The sedimentary origin of the slate is shown by the strong dominance of alumina which exceeds the molecular proportion of the combined lime and alkalies by over 100 per cent.

Analyses of Peach Bottom slate

	1	2	3	4
SiO_2	55.88	58.37	60.32	44.15
Al_2O_3	21.85	21.99	23.10	30.84
Fe_2O_3	--------	10.66 (Fe_2O_3 + FeO)	7.05	14.87
FeO	9.03			
MgO	1.50	1.20	.87	.27
CaO	.16	.30	--------	.48
Na_2O	.46	1.93 (Na_2O + K_2O)	.49	.51
K_2O	3.64			4.36
H_2O	3.39	4.03	3.83	4.49 (H_2O + C)
C	1.79	.93	4.08	
TiO_2	1.27	Trace.	--------	--------
CO_2	--------	.39	--------	--------
SO_3	.02	--------	--------	--------
S	--------	.11	--------	--------
MnO	.59	--------	--------	--------
FeS_2	.05	--------	--------	--------
CoO	Trace.	--------	--------	--------
	99.63	99.91	99.74	99.97

Specific gravity 2.89.

1. Peach Bottom "roofing" slate. J. Humphrey & Co.'s quarry, half a mile east of Delta, York County, Pa. Analysis by A. S. McCreath, Pennsylvania Second Geol. Survey, Rept. CCC, pp. 269–270, 1880.
2. Peach Bottom slate (perfectly fresh). J. Humphrey & Co.'s quarry. Analysis by Booth, Garrett, and Blair, Philadelphia, Pa., 1895. Mathews, E. B., Maryland Geol. Survey Rept., vol. 2, p. 226, 1898.
3. Slate, Lancaster County, Pa. Merrill, G. P., A treatise on rocks, rock weathering, and soils, p. 119, 1906.
4. Slate, Harford County, Md. Merrill, G. P., idem, p. 214.

In 1876 I. N. Rendell, president of Lincoln University, near Oxford, Pa., obtained from the dump pile of the Peach Bottom slate quarries on Susquehanna River six or seven slabs of roofing slate bearing markings that suggested fossils. These slabs were sent to Leo Lesquereux, who reported in 1879 that the markings were *Buthotrephis flexuosa.*[40] Later the supposed fossils were sent to James Hall, who reported in 1883 that "the conspicuous fossils have the character of marine vegetation and the aspect of algae."[41] He acknowledged that these markings may well be dendritic markings of inorganic origin. He then mentioned other forms that are "distinctly organic in their character and mode of occurrence and belong to some algoid form." He also described faint irregular marking and said, "In certain light I think I can see by the naked eye faint markings as if of branches, radiating in an irregular manner from the center and bifurcating in their course. This reminds me strongly of some graptolitic forms."

On the basis of the supposed *Buthotrephis* and supposed graptolites the Peach Bottom slates were assigned to the "Hudson River shale" horizon of the Upper Ordovician. However, the most careful and persistent search through the Peach Bottom slates by numerous

[40] Frazer, Persifor, jr., Age and position of the Peach Bottom slates: Am. Philos. Soc. Proc., vol. 18, pp. 366–369, 1879.

[41] Frazer, Persifor, jr., The Peach Bottom slates of southeastern York and southern Lancaster Counties: Am. Inst. Min. Eng. Trans., vol. 12, pp. 353–359, 1884.

geologists has failed to reveal any further fossil markings that would confirm the reported find of graptolites. The deep refuse piles that surround the quarries, both in Pennsylvania and in Maryland, contain many specimens with distinct dendritic markings that might easily suggest fossil forms, but are all of inorganic origin. It may be that the dubious graptolites mentioned by Hall were in reality nonorganic markings and that if any of the so-called fossils discovered in 1879 were really of organic origin they were algae, not necessarily Ordovician or even Paleozoic. If the Cardiff conglomerate underlying the slates is a basal conglomerate, then the slates are in all probability Paleozoic. The Cardiff conglomerate occurs in a region where conditions subsequent to deposition have been favorable for molecular rearrangement which would have obliterated evidence of unconformity. Nevertheless it has been possible to prove an unconformity at the base of the Glenarm series where basal conglomerates are lacking, and it is somewhat surprising that an unconformity at the base of the Cardiff conglomerate can not be demonstrated by field evidence. Therefore it seems best to include both the Cardiff and Peach Bottom in the pre-Cambrian, because it is impossible to prove that there is an unconformity below the Cardiff.

PALEOZOIC SEDIMENTARY ROCKS

PALEOZOIC STRATIGRAPHY OF WELSH MOUNTAIN AND VICINITY

The Honeybrook upland lies north of the Chester Valley and extends southwestward from Valley Forge, on Schuylkill River, to Quarryville on the south and from Phoenixville to a point north of Honeybrook on the north. It is cut off on the northeast and north sides by the sediments of the Triassic basin, which trends northwestward from Valley Forge to Phoenixville and Elverson. Its greatest width from north to south is about 12 miles. The upland terminates on the west in three anticlines, named Welsh Mountain, Barren Hill, and Mine Ridge, from the ridges of Lower Cambrian quartzite which partly inclose the pre-Cambrian crystalline rocks that form the core of the upland. These anticlines plunge southwestward and carry the Lower Cambrian arenaceous series under the Cambrian and Ordovician limestones of the Quarryville, Lancaster, and New Holland Valleys.

The Lower Cambrian series, which unconformably overlies the pre-Cambrian core of the Welsh Mountain and Barren Hill anticlines, is composed of the Chickies quartzite, including its basal Hellam conglomerate member, the Harpers phyllite, and the Antietam quartzite. This series is 2,200 feet thick in Welsh Mountain and is little metamorphosed. The calcareous Antietam quartzite grades upward into and is overlain by the Vintage dolomite, which together with the Kinzers formation and Ledger dolomite compose the equivalent of the Lower Cambrian Tomstown dolomite of Cumberland Valley. The Elbrook, Conococheague, and Beekmantown limestones overlie the Lower Cambrian limestone in the synclinal valley areas both north and south of the Welsh Mountain-Chestnut Hill axis, which crosses the Lancaster and New Holland Valleys. The Cocalico shale lies only north of the above-mentioned axis of uplift.

PALEOZOIC STRATIGRAPHY OF MINE RIDGE UPLAND AND VICINITY

The Mine Ridge upland is a double anticline lying partly in the northeastern part of the Quarryville quadrangle and is the southernmost of the three anticlinal folds that underlie the Honeybrook upland. The anticline is 30 miles long and 3 miles wide and extends from a point east of Downingtown southwestward to Quarryville. At both ends it plunges under the limestones of the Chester-Quarryville Valley, and it is separated from the anticlinal area of Barren Hill, to the north, by normal faults. Like the Welsh Mountain and Barren Hill anticlines, the Mine Ridge fold is formed of a core of pre-Cambrian rocks bordered by Lower Cambrian arenaceous rocks overlain by Paleozoic limestones. The Paleozoic rocks of the Mine Ridge upland and vicinity have been greatly metamorphosed and closely folded and do not closely resemble the less altered Paleozoic rocks north and west of them. Their relations and age have been worked out, however, in connection with a study of sequence of the slightly altered fossiliferous Lower Cambrian arenaceous rocks of Welsh Mountain, Chestnut Hill, and the Hellam Hills and the fossiliferous limestone series of the adjoining New Holland, Lancaster, and York Valleys.[42]

LOWER CAMBRIAN ROCKS OF THE McCALLS FERRY-QUARRYVILLE DISTRICT

CHICKIES QUARTZITE

The basal Cambrian formation of the McCalls Ferry-Quarryville district is the Chickies quartzite. It consists of the Hellam conglomerate member at the base, overlain by a considerable thickness of white vitreous quartzite, which is succeeded by a greater thickness of sericitic quartzite interbedded with quartz-mica schist and quartz schist. The Chickies quartzite was named in 1878 by Lesley and Frazer from outcrops of quartzite at Chickies Rock, in the gorge of Susquehanna River north of Columbia, Pa. At the type locality the quartzite is exposed in an anticline which does not bring to the surface the underlying Hellam conglomerate member. The quartzite

[42] Stose, G. W., and Jonas, A. I., The lower Paleozoic section of southeastern Pennsylvania: Washington Acad. Sci. Jour., vol. 12, pp. 358–366, 1922.

here is a heavy-bedded white vitreous quartzite at the base and a pebbly quartzite with black slate partings near the top.

Hellam conglomerate member.—The Hellam conglomerate member of the Chickies quartzite forms the base of the Cambrian and is exposed on the flanks of Mine Ridge. It was named [48] from the Hellam Hills, York County, Pa., where it is well developed and gives rise to high ridges.

In the Quarryville area the Hellam conglomerate is exposed as a narrow band around the pre-Cambrian rocks of the Mine Ridge anticline. The conglomerate together with the overlying quartzite forms a ridge which surrounds the central area underlain by pre-Cambrian rocks, above which the bordering quartzites rise 100 feet on the north and southwest sides. The conglomerate is well exposed in several places on both the north and south sides of the ridge. The freshest outcrops are in the valley of Nickel Mines Run north of Greentree and along Houston Run southwest of Gap. At the latter place the rock stands up as ledges 25 feet high exposing 40 feet of thick beds of pebbly quartzite with sericitic schist partings.

The Hellam conglomerate on the Mine Ridge upland is a coarse to fine grained siliceous conglomerate interbedded with mica schist. The conglomerate is a pebbly quartzite and quartz conglomerate composed of thoroughly recemented and recrystallized quartz grains with green muscovite developed on bedding planes. It contains small amounts of microcline and albite, with accessory zircon, magnetite, hematite, and titanite. The pebbles are milky and blue quartz, and have been so much granulated by pressure that weathering reduces them to an aggregate of sand grains, thus obscuring the conglomeratic character of the rock in weathered exposures. These pebbles have been flattened and elongated, and their edges interlocked with the quartz of the groundmass.

The quartz conglomerate occurs in beds a foot or two thick intercalated with thinner beds of crumpled mica schist. The schist beds represent the metamorphosed equivalent of the fine sericitic quartzite layers that occur between conglomerate beds of the Hellam conglomerate of Welsh Mountain. The effect of metamorphism on the quartz conglomerate beds has been elongation and granulation of the pebbles and the development of biotite and muscovite, producing a schistose texture in the rock.

The maximum thickness of the Hellam conglomerate of the Mine Ridge upland is 90 feet, measured north of Greentree. In the type locality of the Hellam Hills a thickness of more than 600 feet has been estimated. The Hellam conglomerate is thinner around the Mine Ridge upland than it is a few miles to the north, on Welsh Moun-

[48] Stose, G. W., and Jonas, A. I., op. cit, p. 96.

tain. This difference is due partly to difference in original thickness and partly to a considerable reduction in thickness of the formation around the Mine Ridge upland as a result of the intense compression which has flattened the pebbles in the harder beds and rendered the softer beds schistose. In the writers' opinion the Hellam member corresponds in general with the Loudoun formation of South Mountain and Catoctin Mountain and the basal Cambrian conglomerate of the Reading Hills.

The Hellam conglomerate is the basal member of the Chickies quartzite. It contains no fossils but unconformably overlies pre-Cambrian rocks throughout its extent. It contains pebbles of blue quartz, probably derived from quartz veins in the underlying pre-Cambrian gabbro exposed northeast of Gap and from the pre-Cambrian granite of the region, which also contains blue quartz.

Quartzite member.—In the McCalls Ferry-Quarryville district the quartzite member of the Chickies quartzite is exposed in the Mine Ridge upland, 15 miles east of the type locality. It encircles the upland and together with the underlying basal conglomerate forms a prominent ridge on the flanks of the uplift. Being less resistant to erosion, the underlying pre-Cambrian rocks have been reduced to a level below that of the quartzite on the northern and western flank of the anticline.

The quartzite has been quarried south of Gap in a quarry cut across the strike, where 70 feet of hard vitreous quartzite with a dip of N. 40° W. is exposed. On the southwest nose of the ridge, a mile north of Quarryville, the quartzite has disintegrated into fine white sand streaked with sericitic clay beds. This sand is dug in small pits and is used for concrete and plaster.

The quartzite member of the Chickies quartzite is made up of white vitreous quartzite, some of which is massive and some thin bedded, with sericite developed on bedding planes. The upper two-thirds of the quartzite member consists of thin-bedded crumbly white sericite quartzite interbedded with quartz-mica schist and quartz schist. The quartzite is seen in thin section to be made up chiefly of interlocking areas of quartz, which show strain shadows. The other constituents are straw-colored blades of muscovite, a little microcline, irregular albites with fine inclusions, well-rounded zircons, and magnetite. Black tourmaline, which is a characteristic constituent of the rocks on the flanks of the Mine Ridge anticline, is especially abundant in the quartzite. The original sand, which was rather pure, has been thoroughly recrystallized, but the stretching of the tourmaline would indicate that the rock has been subjected to dynamic disturbance subsequent to recrystallization.

The mica schist is a fine-grained schistose aggregate of biotite, quartz, and subordinate feldspar.

Of the following analyses No. 1 represents the quartzite at Chickies Station, Pa., and No. 2 the Montalto quartzite of South Mountain. Both analyses are those of a pure quartzite. The rock from South Mountain contained some sericite and epidote.

Analyses of quartzite

	1	2		1	2
SiO_2	97.71	[a] 92.00	K_2O		1.16
Al_2O_3	1.39	4.21	H_2O (loss on ignition)		.96
Fe_2O_3	1.25	1.80	TiO_2		.14
FeO			P_2O_5		.21
MgO	.13				
CaO	.18	.04		100.66	100.68
Na_2O		.16			

[a] 7.87 per cent is combined silica.

1. Pirsson, L. V., Rocks and rock minerals, 1st ed., p. 367, 1910.
2. Quartzite specimen No. 1157, 1½ miles south of burned sawmill on the Conococheague. Analysis by F. A. Genth. Pennsylvania Second Geol. Survey Rept. CCC, p. 363, 1880.

The thickness of the quartzite member of the Chickies overlying the Hellam member on the flanks of the Mine Ridge anticline has been estimated as 350 feet. The full section is not exposed at any one place in that area, hence the thickness can not be determined accurately. It is somewhat thinner on the Mine Ridge anticline than in Welsh Mountain or the Hellam and Chickies Hills.

The basal member of the quartzite, the Hellam conglomerate, unconformably overlies pre-Cambrian rocks. The quartzite is part of an arenaceous series that contains Lower Cambrian trilobite and brachiopod remains in its upper part, although the only fossil remains that have been collected from the quartzite member are scolithus tubes. Therefore both stratigraphic relations and fossil content establish the age of the quartzite member of the Chickies as Lower Cambrian. It is similar to the Montalto quartzite member of the Harpers schist of South Mountain, Pa., but as it lies at the base of the Harpers and not in the midst it is probably not the exact equivalent of the Montalto.

HARPERS AND ANTIETAM SCHISTS

The Harpers schist[44] received its name from Harpers Ferry, W. Va., where it is well exposed in the gorge of Potomac and Shenandoah Rivers. It lies on the flanks of the Blue Ridge, and an eastern belt of the formation extends northward along South Mountain into Pennsylvania.

The Harpers formation is widely exposed in the Piedmont of Maryland and Pennsylvania south of the Hanover-York Valley and in the Pigeon, Hellam, and Chickies Hills and Welsh Mountain.

[44] Keith, Arthur, Geology of the Catoctin belt: U. S. Geol. Survey Fourteenth Ann. Rept., pt. 2, p. 333, 1894.

The Harpers of this area is a greenish-gray phyllite composed of albite, muscovite, and quartz with some biotite. It has not, however, advanced beyond a phyllitic stage of metamorphism and should be called a phyllite. In its type locality near Harpers Ferry the rock is a shale. In South Mountain it is a phyllite like much of the Harpers of the Piedmont north and northwest of the Mine Ridge upland. From Welsh Mountain southward to the Mine Ridge upland there is a gradual increase in metamorphism of the Paleozoic sediments, and the Harpers phyllite has been altered into an albite-mica schist and the overlying Antietam quartzite into a quartzose albite schist.

The Antietam sandstone (quartzite), which overlies the Harpers formation in Virginia, West Virginia, and Maryland, was named by Keith [45] from Antietam Creek, on the tributaries of which it is well shown. The Harpers phyllite of Pennsylvania is overlain by a quartzite which is the equivalent of the Antietam sandstone of the Blue Ridge and South Mountain. The upper beds of the formation are a light-gray, somewhat calcareous quartzite containing angular grains of glassy quartz. These upper quartzite beds weather to a laminated, porous, highly ferruginous rock whose bedding surfaces show numerous rusty molds of *Obolella* and trilobite fragments. These beds are similar to fossiliferous beds characteristic of the Antietam sandstone of South Mountain and are excellent horizon markers. The fossiliferous quartzite is interbedded with gray quartz phyllite in which weathering produces a rusty banding. The Antietam of Welsh Mountain is about 120 feet thick and is an easily identified formation between the Harpers phyllite and Vintage dolomite. Metamorphism of this formation in the McCalls Ferry-Quarryville district has produced a quartzose albite schist which is difficult to separate from the underlying Harpers albite schist. As it has been possible to recognize the ferruginous quartzite beds only in a few places near the Mine Ridge upland and in the hills to the west, the Harpers and Antietam have been in part mapped together in this report.

Areal distribution.—The Harpers and Antietam schists of the McCalls Ferry-Quarryville district overlie the Chickies quartzite on the flanks of the Mine Ridge anticline, where they are exposed in a border around the Chickies. These rocks are brought to the surface also in a series of anticlinal hills extending from the Mine Ridge anticline west for 15 miles, nearly to Susquehanna River at Safe

[45] Keith, Arthur, as reported by Williams, G. H., and Clark, W. B., Maryland, its resources, industries, and institutions, p. 68, 1893. The formation was described, but not named, by Keith in Am. Geologist, vol. 10, p. 365, 1892.

Harbor. These hills, to many of which local names have been given, such as Bunker Hill and Oak Hill, trend N. 75° E., in a direction roughly parallel to the strike of the formation. They rise about 150 feet above the valley level but show few fresh outcrops except in stream gorges cut, for the most part, north and south across the hills. Quarries have been opened in a few places and the stone is used locally for building. The largest quarry, a mile north of Safe Harbor, yields fresh and well-jointed blocks, but is not operated at present.

Lithologic character.—Two facies of micaceous schist can be distinguished on the Mine Ridge upland. The lower overlies the Chickies quartzite and is certainly part of the Harpers formation. It will be called the blue garnetiferous chlorite schist. It is overlain by an albite-mica schist that includes the upper part of the Harpers and the overlying Antietam formation.

The blue chlorite schist has been mapped separately on the flanks of the Mine Ridge anticline. It forms a narrow band on the south face and widens considerably on the northwestern nose of the ridge, and extends northeastward toward Gap, where it is cut off by a fault. Erosion has not gone deep enough to expose this basal part of the Harpers in the outlying hills west of the Mine Ridge upland.

The blue chlorite schist is a fine-grained closely crumpled schist containing chlorite and some biotite with subordinate quartz and feldspar. The chlorite and mica are in large blades that give a spangled appearance to the rock and wrap around large red garnets. In weathered rock the garnets stand out upon the foliation planes, producing a knobby surface. The bluish color is due to the chlorite content of the schist. Thin sections of the rock show that the feldspar is sericitized. Chlorite occurs in large green blades without orientation. Quartz, some secondary albite, garnet, and magnetite make up the rest of the rock.

The overlying albite schist is a gray coarse-grained sparkling muscovite-biotite schist spotted with dull-white porphyroblasts of albite. The albite crystals are most conspicuous on rough surfaces normal to the plane of schistosity. Calcite, tourmaline, zircon, and apatite are constituents of common occurrence. Albite occurs in spongiform individuals filled with inclusions of the other constituents. These inclusions usually show a parallel arrangement, which may be either inclined or parallel to the foliation. (See pl. 2, *A*.) Quartz is fairly abundant. Calcite appears to have crystallized at the same time as the other constituents.

The albite mica schist includes the rest of the Harpers above the blue chlorite schist and the Antietam. The upper part of the Antietam is an albite schist but is more calcitic and dolomitic and contains less albite than the underlying beds. It is a fine-grained light-gray schist speckled with brown biotite containing many small needles of black tourmaline. Microscopically the rock shows abundant calcite, dolomite, and biotite; some of the biotite occurs as porphyroblastic blades with inclusions of the groundmass. Quartz, albite, titanite, and pyrite are present in small amounts. Tourmaline crystals which contain inclusions of the rock minerals show that the mineral was formed as a result of pneumatolytic action subsequent to recrystallization of the rock. (See pl. 2, *B*.)

At first sight such sparkling crystalline rocks as the albite-mica schists just described show absolutely no resemblance to the smooth phyllite or impure quartzite that occurs in the Hellam-Chickies anticline and in Welsh Mountain, but it has been possible to trace a graduation in metamorphism from phyllite to schist. The gradational belt is only 5 miles wide but has been shortened considerably by normal faulting. In the gradational belt the Harpers and Antietam are brought to the surface by several anticlines in which the phyllite progressively shows an increase of metamorphism from the flanks of Welsh Mountain southward. Metamorphism has for the most part obscured the fossil remains usually found in the ferruginous quartzose beds of the Antietam, but in many places, even in the area of most intense anamorphism, the ferruginous beds can be detected. These beds weather to a biotitic rusty-banded calcareous schist that contains albite and is streaked with limonite and that probably represents ferruginous fossiliferous layers of the less metamorphosed Antietam quartzite.

The Harpers and Antietam schists weather readily because of their content of calcite and form a deep, loose soil. The steep slopes of the hills are scarred by deep gullies that render the land difficult of cultivation. Because of the character of the soil these hills are locally called "the sand hills."

Chemical composition.—The following table gives analyses of the coarse albite Harpers schist and of Antietam calcareous albite schist, both from the Quarryville quadrangle; also a partial analysis of the Harpers phyllite from a point a short distance to the north.

Analyses of Harpers and Antietam schists

Chemical composition	1	2	3	Mineral composition (2)	
SiO_2	64.40	57.36	67.38	Quartz	34
Al_2O_3	12.62	11.84		Biotite	18
Fe_2O_3	1.80	1.26		Orthoclase	12
FeO	2.56	3.15		Muscovite	10
MgO	1.74	4.52		Dolomite	10
CaO	2.20	5.92		Calcite	7
Na_2O	2.34	1.06	2.18	Albite	8
K_2O	5.12	5.10	5.79	Ilmenite and magnetite	2
H_2O-	1.08	.01			
H_2O+		.95			101
TiO_2	1.13	.88			
CO_2	1.34	7.28			
P_2O_5	.50	.02			
MnO	.06	.12			
	100.89	99.47			

1. Harpers albite schist, State road, three-fourths of a mile west of Camargo, Lancaster County, Pa. Analyst, R. K. Bailey, U. S. Geol. Survey, 1920.
2. Antietam calcareous albite schist, 1 mile south of Lime Valley, Lancaster County, Pa. Analyst Glenn V. Brown, U. S. Geol. Survey, 1920.
3. Harpers phyllite, 1 mile south of Mountville, Lancaster County, Pa. Analyst, J. G. Fairchild, U. S. Geol. Survey, 1922.

Character of the metamorphism in the Harpers and Antietam albite schist.—In the garnetiferous chlorite schist metamorphism has apparently taken place in the upper part of the deep-seated zone, as is shown by the formation of such heavy minerals as garnet and plagioclase, which are characteristic of zones of high pressure and temperature, associated with chlorite, which forms in the upper zones of metamorphism. In the albite-mica schist also metamorphism has taken place in a transition zone, albite being characteristic of the upper zone and biotite of the lower zone. The lime content of the albite-mica schist, instead of uniting with soda to form plagioclase, has crystallized as calcite, and the soda as albite. All the minerals in the path of each growing albite crystal that were not needed in forming calcite were inclosed by the metacryst, which has irregular spongiform boundaries.

Thickness.—The thickness of the Harpers phyllite has been estimated to be 1,000 feet in the Hellam and Chickies Hills and about 1,500 feet in Welsh Mountain. The overlying Antietam quartzite is about 120 feet thick in Welsh Mountain. The albite-mica schist representative of the Harpers and Antietam formations on the flanks of the Mine Ridge upland probably attains a thickness comparable to that of the formation on Welsh Mountain, a few miles to the north, but the thickness of the schist can not be measured. Cleavage has developed to such an extent that the bedding planes are rarely discernible. It is not only difficult to trace the original bedding, but repeated minor folding makes it impossible to estimate the thickness of the formation with any exactitude.

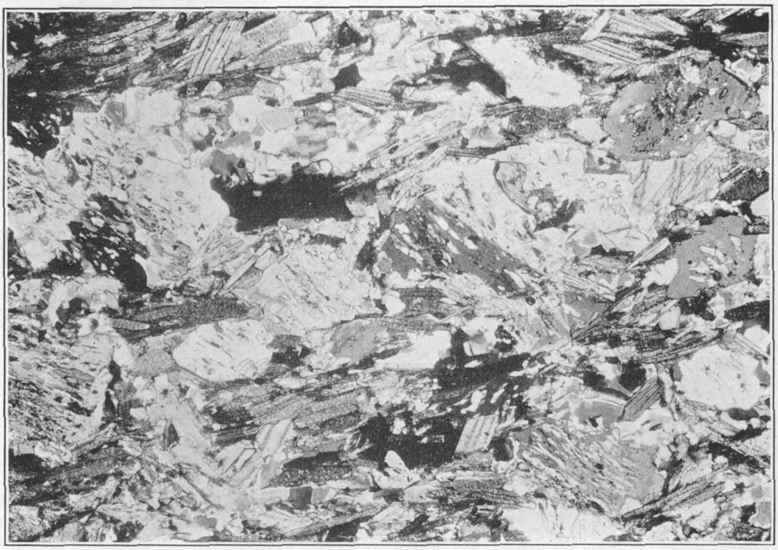
Age.—The resemblance of the Harpers albite schist to the albite schist facies of the Wissahickon is very striking, so that in earlier work the writers considered the rocks to be identical. This resemblance is especially marked between the Harpers albite schist and the biotitic albite facies of the Wissahickon that occurs along Susquehanna River near Safe Harbor and Shenks Ferry. However, the albite schist on the flanks of the Mine Ridge anticline and farther west has been proved on stratigraphic grounds to be Lower Cambrian Harpers schist and Antietam quartzose schist. The chlorite schist and albite schist overlie Chickies quartzite on the flanks of the Mine Ridge anticline. The albite and quartzose calcareous schist in the valley have been established as anticlinal and have been shown to be conformable under Vintage dolomite, into which they also grade through a calcareous schist. The gradation in fresh rock cuts has been seen at several localities. An excellent exposure three-fourths of a mile south of Conestoga shows in a thickness of 6 feet passage from a biotite dolomite schist through rusty-banded dolomite speckled with biotite into a cream to dove-colored dolomite free of mica and belonging to the Vintage formation. Metamorphism has almost entirely obliterated fossils, but fossils have been found in the Antietam half a mile north of Gap. Therefore, on structural and stratigraphic grounds, as well as by fossil content in adjacent areas, the albite-mica schist has been found to be identical with the Harpers phyllite and the calcareous biotite schist with the Antietam quartzite and hence Lower Cambrian.

VINTAGE DOLOMITE

The arenaceous series of Welsh Mountain and of the McCalls Ferry-Quarryville district is overlain by limestones, the oldest of which is the Vintage dolomite,[46] named from a small village 15 miles east of Lancaster and just north of the Quarryville quadrangle, where most of the section is excellently exposed in a cut of the Pennsylvania Railroad. (See pl. 3, *A*.)

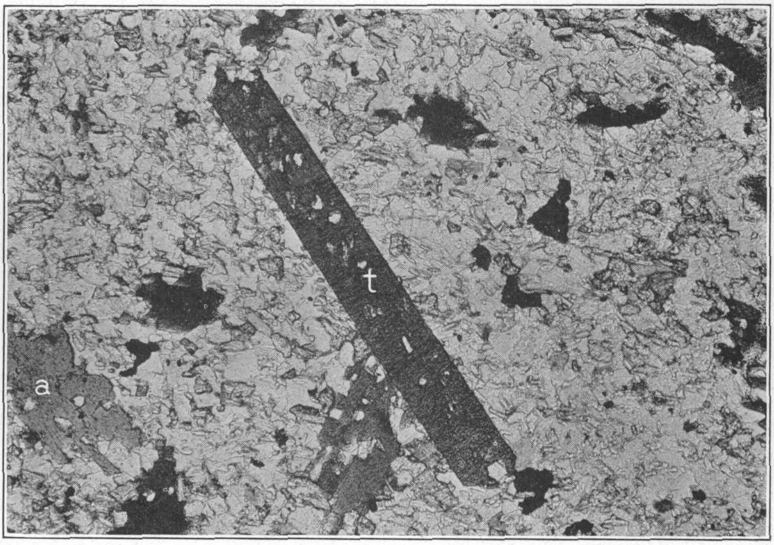
Distribution.—The Vintage dolomite is exposed north and west of Mine Ridge upland on the slopes of the "sand hills," made up of Antietam albite schist, which it surrounds. It occupies valleys also between the Cambrian schists and the blue slaty Conestoga limestone. North of Gap it underlies the valley west to Bellemont and Leaman Place and forms a narrow valley that extends 8 miles west of Bellemont. The best exposures lie in the cuts of the Pennsylvania Railroad from Leaman Place to Kinzers. There is a large quarry in Vintage dolomite (see pl. 3, *B*) now operating at Kinzers, north

[46] Stose, G. W., and Jonas, A. I., Lower Paleozoic section of southeastern Pennsylvania: Washington Acad. Sci. Jour., vol. 12, No. 15, p. 362, 1922.



A. HARPERS SCHIST

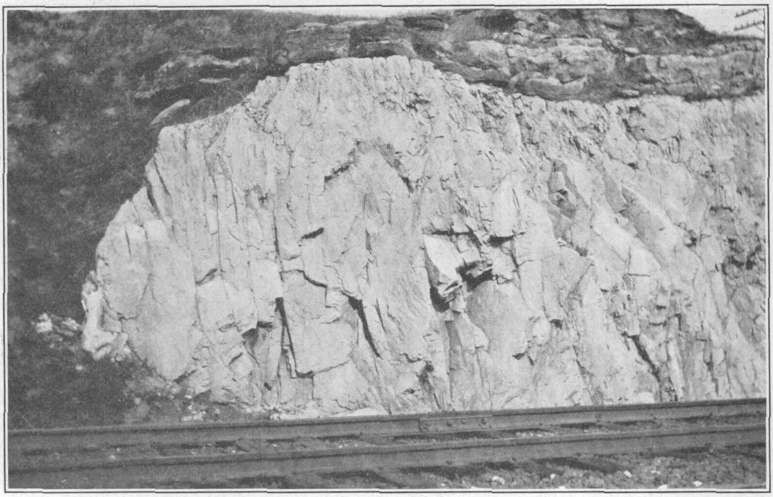
The albite is crowded with inclusions in a roughly parallel orientation that is inclined to the foliation of the rock



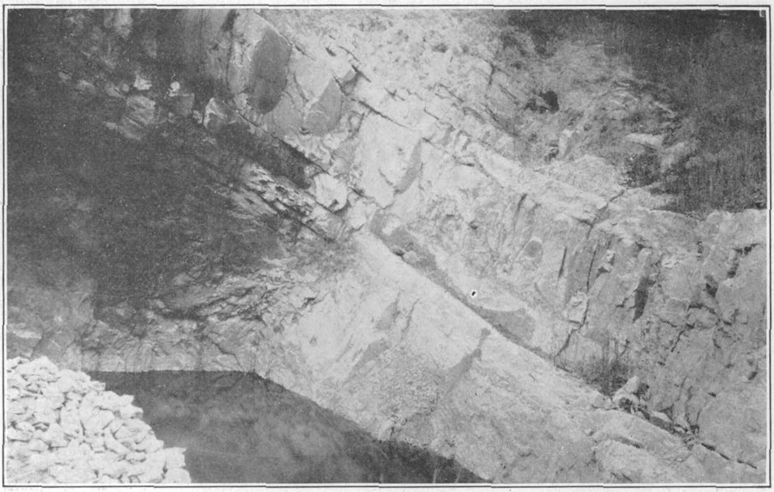
B. ANTIETAM SCHIST

Porphyroblastic biotite, a, in a dolomitic matrix. The tourmaline, t, is epigenetic

PHOTOMICROGRAPHS OF SCHIST

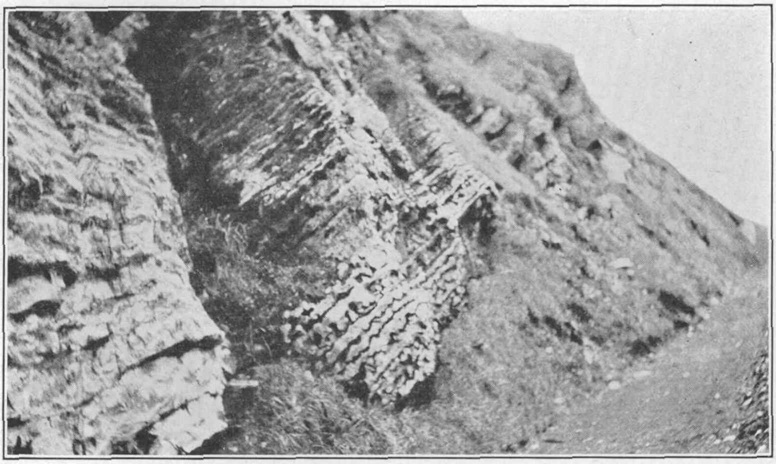


A. UPPER PART OF THE VINTAGE DOLOMITE OVERLAIN BY KINZERS SHALE EXPOSED IN CUT ON THE PENNSYLVANIA RAILROAD NEAR VINTAGE



B. VINTAGE DOLOMITE IN QUARRY AT KINZERS

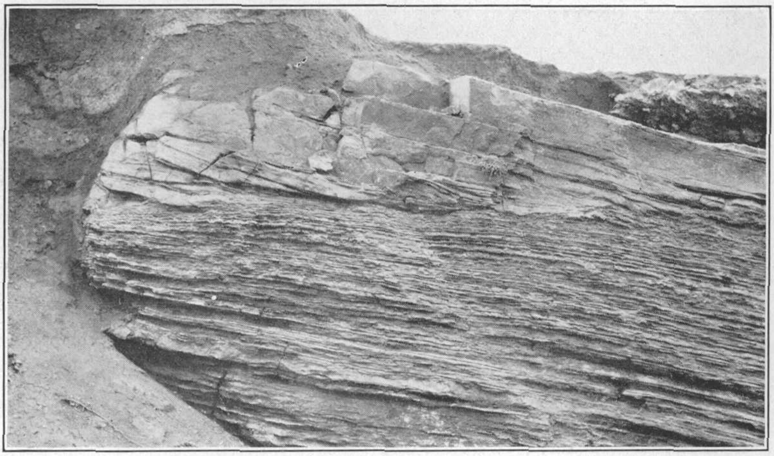
Shows well-bedded character of lower part of Vintage dolomite



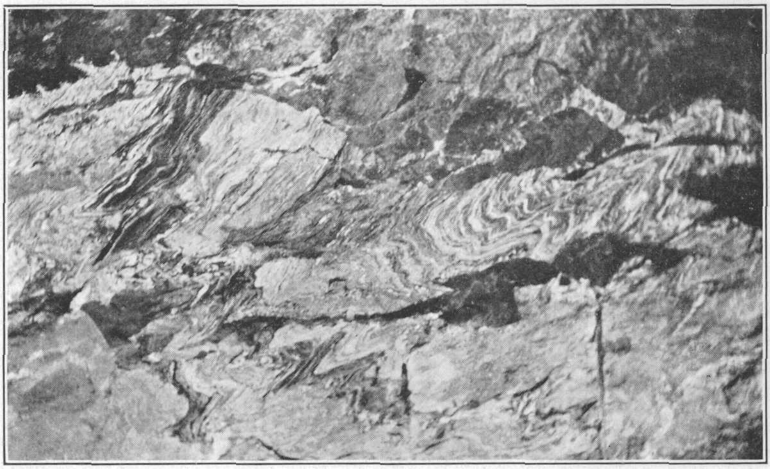
A. KINZERS PSEUDOCONGLOMERATE ON PENNSYLVANIA RAILROAD WEST OF KINZERS

Produced by differential weathering of white marble and dark impure dolomite

B. BASAL CONESTOGA SLATE AND LIMESTONE CONGLOMERATE, UNCONFORMABLY OVERLYING VINTAGE DOLOMITE, BELLEMONT QUARRY



A. THICK-BEDDED CRYSTALLINE LIMESTONE PASSING INTO FINELY LAMINATED SLATY LIMESTONE TYPICAL OF CONESTOGA LIMESTONE IN QUARRY 1 MILE NORTHWEST OF BELLEMONT



B. CLOSELY FOLDED MICACEOUS MARBLE OF CONESTOGA LIMESTONE IN QUARRY HALF A MILE NORTHWEST OF QUARRYVILLE of the Lincoln Highway, where the rock quarried is crushed for road metal. In the valley of Pequea Creek north and east of Lime Valley there are many good outcrops of dolomite in stream cuts and old quarries. It is well exposed farther southwest along Pequea Creek near Burnt Mills. No quarries are now operated in the Vintage dolomite in the McCalls Ferry-Quarryville district, except the one at Kinzers.

Lithology.—In general the Vintage consists of massive glistening dark-gray dolomite, weathering whitish, with scattered coarse crystalline blebs, and dark-blue dolomite with argillaceous partings, weathering knotty. At the base is a cream-colored schistose thin-bedded dolomite containing muscovite flakes, with overlying knotty blue and white dolomite weathering to a rough surface. Higher up there are some dark-blue beds with argillaceous partings, in places micaceous and wavy. In the upper part the rock is a massive sparkling gray to blue mottled dolomite in thick beds. It weathers white with siliceous and calcareous blebs. The uppermost beds are very massive, as is well shown in the Pennsylvania Railroad cut, where the Vintage underlies shales of the succeeding formation. The rock weathers to a deep-red clay soil and occupies the lower slopes of the sand hills. The red color of the soil derived from the dolomite contrasts sharply with that of the gray sandy soil derived from the Harpers and Antietam schists.

Thickness.—At the type locality, just north of the Quarryville quadrangle, the thickness of the Vintage dolomite has been estimated as 654 feet, although the base of the formation is not exposed at this place. The Vintage dolomite thins southward toward the Mine Ridge upland. At Lime Valley it is 150 feet thick, 2 miles farther south it is 50 feet thick, and south of Oak Hill it is absent. This irregularity of thickness and southward thinning is due to erosion subsequent to deposition and is discussed on page 57. The section at Vintage is given below.

Section of Vintage dolomite in cut on main line of Pennsylvania Railroad west of Vintage, Pa.

	Feet
Argillaceous dolomite; weathers rusty and earthy	7
Massive, sparkling, gray-blue mottled dolomite; weathers chalky, with siliceous and calcareous blebs in relief	30
Blue dolomite, wavy bedding, with mica partings	25
Thin shale, micaceous	5
Blue argillaceous dolomite; weathers to thin slivers, with some massive blue dolomite bands	15
Granular dolomite, light gray-blue mottled, in 5-foot beds	35
Dense darker blue massive dolomite, with some blebs, argillaceous partings; weathers chalky	100

	Feet
Light and dark mottled dolomite	35
Gray, wavy, thin-bedded dolomite; weathers chalky	148
Covered, probably dolomite	234
Dark, glistening dolomite; weathers white (base not exposed)	20
	±654

Age and correlation.—The Vintage dolomite has not yielded any fossils in the McCalls Ferry-Quarryville district. Stose and Jonas have collected remains of *Salterella*, trilobites, and brachiopods and plates of cystids from the formation at several localities in the York Valley. The section that has yielded the most fossils lies south of Emigsville and has been described by Walcott.[47]

The fossils and stratigraphic relations of the Vintage dolomite show its Lower Cambrian age. It overlies the Antietam schist, which grades up into the dolomite, and it is overlain by a shale that contains Lower Cambrian fossils. Lithologically it resembles the Tomstown dolomite of Cumberland County, Pa., and it is considered to represent the basal part of the Tomstown.

KINZERS FORMATION

Name.—The Kinzers formation takes its name[48] from an excellent exposure in the Pennsylvania Railroad cut at Kinzers, just east of Vintage. Here it is exposed in an almost complete section overlying Vintage dolomite.

Distribution.—The Kinzers formation in the McCalls Ferry-Quarryville district occurs in two synclinal folds in the Vintage dolomite 2 miles north of Gap. It is more widely exposed to the north, in the valley west of Barren Hill and Welsh Mountain, and north and northwest of Lancaster. Being a thin, shaly formation between two dolomites, it forms narrow hills by differential erosion. No quarries have been operated in it in the vicinity of the Mine Ridge upland and Welsh Mountain, but north of Lancaster, where the formation is largely a shale, it has been quarried extensively.

Lithologic character.—At the base of the formation there are a few thin beds of impure dolomite that weather to an earthy tripoli containing at many places remains of *Salterella*, brachiopods, and trilobites. These beds are followed by a variable thickness of blue hackly shale as much as 50 feet thick in places. The earthy tripoli and shale beds are shown in Plate 3, *A*, overlying Vintage dolomite. Above the shale is a variable series of dark banded argillaceous dolomites that weather to tough buff ribbed argillaceous rocks,

[47] Walcott, C. D., Cambrian rocks of Pennsylvania: U. S. Geol. Survey Bull. 134, pp. 15–16, 1896.

[48] Stose, G. W., and Jonas, A. I., The Lower Paleozoic section of southeastern Pennsylvania: Washington Acad. Sci. Jour., vol. 12, p. 362, 1922.

sparingly fossiliferous. Some beds are an intimate mixture of nodular white granular marble and dark impure dolomite that weathers to a knotty pseudoconglomerate of white marble. (See pl. 4, *A*.)

Thickness.—The thickness at the type locality is about 132 feet. South of Welsh Mountain the Kinzers formation is much thinner, in places not more than 25 feet, and the shale is not prominent.

The section from the type locality is as follows:

Partial section of Kinzers formation in railroad cut at Kinzers, Pa.

	Feet
Dark-blue limestone with wavy impure partings	10
Thick-bedded light-gray dolomite	12
Dark-blue limestone with wavy impure partings	6
White spotted marble with wavy buff dolomite partings	8
Blue limestone banded with slightly wavy siliceous layers	10
Highly siliceous banded dark limestone, weathering to skeleton of buff siliceous network	8
Impure thick-bedded dolomite, weathering to dense buff tripoli	3
White spotted marble with even buff dolomite banding	8
Wavy banded blue limestone, numerous argillaceous partings	10
Crumbly fissile gray shale	50±
Impure dolomite, weathering to buff tripoli and containing few trilobite fragments and *Salterella*	7
Massive light-blue dolomite (Vintage).	
	132±

Age.—The Kinzers formation contains Lower Cambrian fossils *Olenellus*, *Salterella*, and brachiopods. The trilobites, chiefly *Olenellus thompsoni* (Hall), occur abundantly in the shale and were described by Walcott and extensively collected in the Lancaster-York Valley by Professor Roddy, of Millersville, Pa., and other local collectors. In the McCalls Ferry-Quarryville district fragments of trilobites and *Salterella* occur in the basal Kinzers beds of the Kinzers cut at a horizon which is abundantly fossiliferous elsewhere. The Kinzers is an argillaceous fossil-bearing formation, which is represented by dolomite or limestone in the Lower Cambrian (Tomstown) of the Cumberland and Lebanon Valleys.

LEDGER DOLOMITE

Name and distribution.—The upper part of the Tomstown dolomite is represented in this region by the Ledger dolomite. It received its name from Ledger,[49] 3 miles northeast of Kinzers, and although it is not exposed in the McCalls Ferry-Quarryville district, it underlies most of the wide valley just to the north, from Barren Hill to Lancaster.

[49] Stose, G. W., and Jonas, A. I., op. cit., p. 363.

Lithology.—The Ledger is a granular gray to white dolomite, generally thick bedded, with few bedding planes. Outcrops are few because of the readiness with which the dolomite weathers to a granular red clay soil. It is quarried considerably 2 miles east of Gap, at Limeville, and northward. It has been used as building stone to a slight degree, but it weathers too readily to be satisfactory for that purpose.

Thickness and age.—The thickness of the Ledger has been estimated at 1,000 feet. Fossils have not been found in it in the Lancaster Valley, and its Lower Cambrian age was there determined on stratigraphic grounds. In 1922 Stose and Jonas found a Lower Cambrian brachiopod, *Nasusia festinata* Billings, in this formation west of Hanover. The presence of this brachiopod confirms the determination of Lower Cambrian age.

ORDOVICIAN ROCKS

CONESTOGA LIMESTONE

In the northern part of the McCalls Ferry-Quarryville district there is a great thickness of dark-blue slaty crystalline limestone and limestone conglomerate, which is called Conestoga limestone by the writers because of excellent outcrops along Conestoga Creek, south of Lancaster.

Distribution.—The Conestoga limestone lies south of the Welsh Mountain-Hellam Hills uplift and Lancaster and occupies most of the limestone valley area of the northern part of the McCalls Ferry-Quarryville district. South of Lancaster, which is its northwesternmost locality in the valley, it extends to the Martic Hills and to Quarryville. East of Quarryville it covers the entire narrow floor of Chester Valley nearly to Coatesville and borders the south side of the valley to Schuylkill River, 50 miles east of this area. It extends westward from a point near Gap, its most easterly known occurrence north of Mine Ridge upland, to Susquehanna River. It occurs on the west side of the river at Shenks Ferry in outcrops at the water's edge, and north of Long Level it underlies a valley area that extends to the northern edge of the McCalls Ferry quadrangle and is part of the wider valley included in the York quadrangle west of Long Level. The Conestoga limestone is bordered on the south by the albite-chlorite schist facies of the Wissahickon formation all the way from the eastern edge of the Quarryville quadrangle westward to Susquehanna River at Shenks Ferry. From Shenks Ferry north to Safe Harbor it adjoins on the south in turn the Vintage dolomite, Wissahickon schist, and Harpers schist.

Its northern limit lies north of this district. It underlies the northern part of the valley north of the Mine Ridge upland from a point south of Kinzers through Strasburg to the west side of Susquehanna River. South of London Grove and Willow Street it occupies narrow valleys between areas of Cambrian schist and Vintage dolomite. From the northwest side of the Mine Ridge upland southward and eastward the Conestoga limestone is in contact with the Antietam and Harpers schists of the Mine Ridge upland and contains, in the valley north of Quarryville, hills of the same schists. The apparently anomalous relations it sustains to other formations, both pre-Cambrian and Cambrian, are explained below in connection with the age of the Conestoga.

Lithology.—The Conestoga limestone is made up of thin-bedded dark slaty limestone, coarse conglomerate or breccia of limestone and marble pebbles and fragments, thin-bedded blue crystalline limestone, and thin dark graphitic slate. The limestone conglomerates occur at or near the base of the formation but are irregular in occurrence and thickness along the strike. They may occur in single beds or repeated many times through the lower part of the formation. The conglomerate is composed of both coarse and fine gray to white marble pebbles and angular fragments of dolomitic and siliceous limestone in a blue slaty matrix. Some of the fragments measure as much as 5 feet across, but for the most part the conglomerate is made up of fragments a few inches to a foot in length. Although most of the marble fragments are light in a dark-blue matrix, in some of the conglomerate this condition is reversed—dark pebbles occurring in a lighter-colored matrix.

The conglomerate may form the base of the formation but usually has beneath it a thin layer of black slate. In some places this underlying bed is a thick, hackly, blue-black slate which replaces the conglomerate and other basal beds. The slate is usually banded and micaceous, with abundant pyrite, which weathers to a rusty stain on the blue slate.

The greater part of the Conestoga limestone is blue crystalline limestone with slaty partings. (See pl. 5, *A*.) At some places there are dolomitic beds and beds of light-gray limestone containing quartz in round glassy grains. This rock weathers to a buff porous siliceous skeleton by removal of the lime.

The conglomerate is well exposed at many places in the northern part of the Quarryville quadrangle—in the Bellemont quarry, south of Vintage, at Lime Valley and north of it on Pequea Creek, where its outcrops are repeated many times by folding, also near Pequea Valley and Burnt Mills.

Southward from the northern edge of the district to the Mine Ridge anticline there is a gradual increase in the metamorphism of Paleozoic sediments and consequently of the Conestoga limestone.

South of Bunker Hill, Refton, and Marticville the Conestoga limestone is a coarsely crystalline marble with considerable brown mica developed on the partings and throughout the rock. It is closely folded and crumpled and converted to a crystalline schist, like the underlying Harpers and Antietam schists. West and south of the Mine Ridge upland the metamorphism and folding have been most intense, so that no individual beds of the Conestoga can be traced any distance and no sequence can be determined. Recrystallization has obliterated all trace of limestone conglomerate beds and produced a series of gray and blue marbles in which the slate beds have been altered to micaceous banded marble. (See pl. 5, *B*.) The mica in this marble is phlogopite. There are many good exposures of micaceous marble, highly folded or crumpled, in cuts of the Atglen and Susquehanna branch of the Pennsylvania Railroad north and east of Quarryville and for 2 miles east of Shenks Ferry. Here beds of marble and siliceous marble alternate with beds of micaceous calcareous schist. Cleavage diagonal to the bedding, which is marked in the micaceous layers, is absent in purer marble beds made up of calcite and without the platy mineral, mica.

Thickness.—The thickness of the Conestoga limestone is not known but is probably several hundred feet. The beds are so variable from place to place that no consecutive section has been recognized for any distance. There is no place where a full section can be obtained, and the upper beds have been removed by erosion.

Utilization.—Many quarries have been opened in the Conestoga limestone, both in the limestone conglomerate and in the crystalline limestone. The largest quarries in the limestone are those west of Strasburg, just east of Strasburg, near Letort, east of Rockhill, and near New Providence. There are quarries in the conglomerate at Bellemont, south of London Grove, south of Wheatland Mills, and both north and south of Lime Valley along Pequea Creek. Quarryville, as its name indicates, was the center of a considerable quarrying industry, and the marble taken out near by has been used for building stone in Quarryville, Lancaster, and vicinity, and burned for lime.

Surface form.—The Conestoga limestone weathers readily because of its easy solubility and forms a valley country in contrast to that produced by the harder Cambrian and pre-Cambrian schists. Its slaty beds furnish a certain amount of resistance to solution, and the country underlain by the Conestoga is rolling valley land with low hills that rise higher than the valleys underlain by the less resistant Vintage dolomite. On weathering the Conestoga limestone of the northern part of the district produces a slaty soil to which the presence of numerous slate fragments gives a blue to black color. The crystalline limestone and marble in the belt south of Bunker

Hill weather to a red micaceous soil more nearly resembling that derived from the Vintage dolomite.

Age and correlation.—As yet the Conestoga limestone has yielded no fossils in the McCalls Ferry-Quarryville district. Small crinoid stem segments and a brachiopod, determined by Bassler as *Strophomena stosei*, were collected by Stose and Jonas from slaty limestones interbedded with conglomerate near the base of the formation at the York Lime & Stone Co.'s quarry, 6 miles northeast of York, Pa. Bassler [50] described this form from the Frederick limestone as of probable Chazy age. Hence the Conestoga limestone is of probable Chazy age and in part, at least, of the same age as the Frederick limestone of Maryland.

Walcott in his study of the limestones of the York-Lancaster Valley called these limestone conglomerates and associated beds probably Lower Cambrian. Before Ordovician fossils were found near York detailed study of the York-Lancaster Valley by Stose and Jonas brought out the facts that these conglomerates are part of a younger formation that unconformably overlies the older limestones and that the Conestoga limestone laps progressively southeastward from the Ledger dolomite, north of Lancaster, down to the Harpers schist, on the south flank of the Mine Ridge anticline.

Southeast of Lancaster the Conestoga overlaps lower and lower beds of the Ledger until near Leaman Place, just north of Vintage, it is nearly down on the Kinzers formation. At the Bellemont quarries, south of Vintage (see pl. 4, *B*), limestone conglomerate and associated slaty limestone of the Conestoga rest on Vintage dolomite, and the unconformity can be seen between the basal slate of the Conestoga and the beveled edges of the underlying beds of Vintage dolomite. The Conestoga overlies the Vintage dolomite throughout the northern part of the McCalls Ferry-Quarryville district. Farther south, on the north flank of the Mine Ridge anticline and south of Refton and Marticville, the Conestoga almost completely overlaps the Vintage dolomite. On the south flank of the Mine Ridge anticline and in areas north of the Martic Hills west to Prospect Hill School the Conestoga rests on Antietam and Harpers schists. The Conestoga limestone in the Norristown quadrangle is said to overlie Beekmantown limestone. The Frederick limestone unconformably overlies Beekmantown limestone, hence the age of the unconformity is post-Beekmantown.

It has been suggested [51] that the Cocalico shale is the northern representative of the Conestoga limestone. When graptolites were first

[50] Bassler, R. S., Maryland Geol. Survey, Cambrian and Ordovician, p. 117, 1919.

[51] Stose, G. W., and Jonas, A. I., Ordovician overlap in the Piedmont province of Pennsylvania and Maryland: Geol. Soc. America Bull., vol. 34, p. 520, 1923.

found in the shale, in 1921–1923, they were considered by Ulrich to be of Normanskill type and the shale probably of Chazyan age and older than the Martinsburg shale, of Trenton and Utica age. Further study by Ulrich of the graptolites found in the Cocalico shale has thrown doubt on the earlier identification of these poorly preserved fossils. He now believes that they are probably lower Trenton and that the Cocalico shale is probably equivalent to the Martinsburg shale of the main belt and outlying areas in region north of the Triassic rocks.[52] The Cocalico shale occurs only north of an axis of uplift represented by the Honeybrook-Welsh Mountain upland and the Chickies and Hellam Hills, and the Conestoga limestone occurs only south of this axis. North of Lancaster the Conestoga limestone and the Cocalico shale are only 5 miles apart across the strike, and the uplift which separates them is structurally low at this point. The suggestion of the probable equivalence of the shale and argillaceous limestone was a natural one in view of the fact that both were considered to be of Chazy age.

The Cocalico shale is now considered to be younger than Chazy on fossil evidence. It unconformably overlies limestones of lower Trenton and upper Beekmantown age, and therefore was deposited after a period of uplift and erosion which was later than the limestone of lower Trenton age. If the fossils of the Conestoga limestone are correctly interpreted and that limestone is Chazyan, its deposition followed an older period of uplift and erosion, the post-Beekmantown. It is possible, therefore, that the uplift represented by the Honeybrook upland and Hellam Hills acted as a partial or complete barrier between the two basins whose erosional and depositional history after Beekmantown time was so different.

IGNEOUS ROCKS

The pre-Cambrian sedimentary core of the Mine Ridge upland is intimately penetrated by igneous rocks that range in composition from amphibolite to diorite and granite. Owing to the paucity and weathered condition of the outcrops it is difficult to establish the structural relations of these rocks, and, as in studying the stratigraphy of the Baltimore gneiss, the geologist, in interpreting the sequence of igneous intrusion, is forced to rely almost entirely upon a few railroad cuts. Mapping of the underlying rock in the highly cultivated farm country that intervenes between stream valleys becomes almost entirely a matter of the successful delimitation of areas characterized by float of varying composition. The probable sequence of igneous intrusion may be outlined as follows:

[52] Stose, G. W., and Jonas, A. I., Ordovician shale and associated lava in southeastern Pennsylvania: Geol. Soc. America Bull., vol. 38, pp. 505–536, 1927.

1. The intrusion of basic sills or the extrusion of flows in the early pre-Cambrian sedimentary series. These sills or flows, which were later metamorphosed to amphibolite schist, probably belong to an earlier period of igneous activity than the post-Glenarm gabbro, which covers so wide an area in southern Pennsylvania, Delaware, and Maryland.

2. The intrusion of gabbro, probably the equivalent of the post-Glenarm gabbroic invasions.

3. The intrusion of diorite, probably associated with granodiorite or quartz monzonite, which forms a widespread invasion of the sedimentary and igneous complex of the Mine Ridge upland, accompanied by pegmatitization, tourmalinization, and an intimate injection along bedding or cleavage planes in the interlayered sedimentary and hornblende schists. The pre-Cambrian or post-Cambrian age of this granodiorite, which has been highly deformed subsequent to consolidation, is a matter of doubt.

4. The intrusion of granite cut by dikelets of pegmatite, quartz, and ore-bearing solutions, all strongly tourmalinized.

5. Diabase intrusion of Triassic age.

PRE-CAMBRIAN IGNEOUS ROCKS

BIOTITIC AMPHIBOLITE SCHISTS

In the Mine Ridge anticline the biotitic amphibolite schists are interlaminated with the pre-Cambrian sediments and are not separated on the map from the Baltimore gneiss. They are completely recrystallized and strongly foliated, the planes of foliation being parallel to the bedding of the inclosing mica schists. The dominant constituents are hornblende, biotite, quartz, and plagioclase. Epidote (in some places surrounding allanite), clinozoisite, and garnet occur in considerable quantities, and leucoxene is abundant. Some biotite plates show distinct pleochroic haloes. Interlayered with the sedimentary beds are a few bands of a light-colored rock that is characterized by actinolite instead of hornblende, in association with biotite.

Complete recrystallization of the original constituents makes it difficult to determine the origin of these amphibolite schists, but the mineral composition and the prevalence of leucoxene are strongly suggestive of derivation from a basic igneous rock rich in ilmenite.

METAGABBRO

No unmetamorphosed gabbro is exposed in Mine Ridge, because the igneous rocks together with the including sedimentary series have been profoundly altered. Biotite-hornblende gabbro or metagabbro occurs in a massive rock whose outcrops are few. This rock is best seen at the Gap nickel mine, where the metagabbro is a coarsely crystalline, dark-gray, and brilliantly lustrous rock, composed chiefly of green hornblende, with subordinate plagioclase and a considerable amount of brown biotite. Under the microscope the original igneous texture is apparent, although there remains no trace of the mineral from which the hornblende was derived, and it is possible that a large part of the hornblende is pyrogenetic. Kemp[53] has found remains of orthorhombic pyroxene in his study of the ore rock, which he calls amphibolite. Biotite, of a deep reddish-brown color, appears to be universally present, though showing a considerable range in amount. Both hornblende and biotite contain numerous large and distinct pleochroic haloes. The radioactive minerals that form the cores of these haloes are zircon and a deeply colored prismatic mineral, possibly allanite. Titaniferous magnetite or ilmenite is of common occurrence in places surrounded by rims of titanite. The abundance of large apatite crystals is noteworthy. Chalcopyrite is a common accessory.

AGE OF BIOTITIC AMPHIBOLITE SCHIST AND METAGABBRO

A strongly foliated variety of fine-grained micaceous amphibolite schist found upon the dump at the Gap nickel mine is composed of hornblende, biotite, plagioclase (a saussuritic albite), some quartz, and both iron oxides and iron sulphides. This amphibolite schist may represent a sheared border of the gabbro that has yielded to dynamic action more completely than the central portion. On the other hand, it may be a layer in the inclosing sedimentary series and thus older than the cross-cutting gabbro rock. The similarity between the biotitic amphibolite interlaminated with the ancient pre-Cambrian sediments and the amphibolite at the Gap nickel mine would suggest that the two rocks are of the same age, and there is no evidence from field relations to show whether the amphibolite specimen from the abandoned nickel mine is really a part of the intrusive stock or a part of the intruded series. But along the Philadelphia & Reading Railroad north of Coatesville, where there are good outcrops of amphibolite schists, interlaminated with the ancient pre-Cambrian sediments, the amphibolites appear to be an integral part of the intensely metamorphosed complex and older than the intrusions of massive metagabbro on the Siousca Road about a quarter of a mile farther north. The massive metagabbro is in every respect identical with the gabbro at the nickel mine and is undoubtedly of the same age.

In the Reading and Durham Hills there is a similar suggestion of two ages of gabbroic intrusion, the older represented by thin layers

[53] Kemp, J. F., The nickel mine at Lancaster Gap, Pa.: Am. Inst. Min. Eng. Trans., vol. 24, p. 626, 1895.

of highly foliated biotitic amphibolites or hornblende schists, intimately interleaved with the sedimentary schists, and the younger represented by the slightly deformed hornblende norites and hornblende gabbros. But even in a comparatively detailed study of the areal geology of these hills that was made by the writers in 1912 to 1914 it was not possible to delimit the two types or to find a place where the massive rock cuts the amphibolite schists. Therefore until extremely detailed field work is done it is impossible to do more than state that a basic rock occurs in the pre-Cambrian sediments in cross-cutting bodies and in thin layers, which have intimately interleaved or interpenetrated the sedimentary layers. In these thin layers, which may have originated as sills or as basaltic flows, fine-grained amphibolite schists have resulted. The intense shearing of these schists is not in itself indicative of older or more intense metamorphism, for the massive metagabbro occupies the center of larger igneous masses where the bulk of the intrusion has somewhat protected the rock from shearing action. Fine-grained amphibolite schists occur also in association with gabbro intrusions, where they probably represent peripheral zones in which the rock has yielded readily to dynamic metamorphism. The sills or flows may be of the same age as the schists associated with the gabbro or they may belong to different periods of igneous activity. The gabbro in Mine Ridge is in every respect lithologically identical with the widespread intrusive bodies of post-Glenarm gabbro in the southeastern Piedmont of Pennsylvania and Maryland and is probably part of the same igneous intrusion.

SERPENTINE

A few pieces of float comprising serpentine and talcose schists were found in the Mine Ridge anticline at isolated localities. As no contacts between serpentine and the inclosing rock could be found it was impossible to determine the structural relations of this serpentine, although its association with pre-Cambrian schists at a locality 1 mile north of Christiana, on the Lancaster-Christiana trolley line, suggests that it is probably intrusive in the pre-Cambrian. In York County, about 2 miles northwest of Peach Bottom, fragments of serpentine and talc schist denote the existence of a pyroxenite or peridotite surrounded by schist. The areal distribution of the serpentines of Maryland indicates that they are intrusive in the Wissahickon and Peters Creek formations and in the gabbro. It is probable, therefore, that the serpentines in the McCalls Ferry-Quarryville district are younger than the metagabbro in which the nickel deposit occurs.

AMPHIBOLITE AND CHLORITE SCHIST—ALTERED VOLCANIC FLOWS

A mile and a quarter southwest of Woodbine, in York County, on Bald Eagle Creek, there is an isolated outcrop of a dense and massive dull-green rock speckled with lighter-colored areas. Under the microscope this rock is seen to be a garnetiferous amphibolite, in which the large garnets are surrounded by rims of chlorite. The rock is doubtless derived from the deep-seated metamorphism of a diabase or basalt. Subsequent dynamic metamorphism at higher tectonic horizons has crushed the garnet and produced a reversal of metamorphism indicated by the alteration of the garnet to chlorite. There are no visible relations between the amphibolite and the surrounding rock.

A similar occurrence is found in Chester County half a mile west of Tweedale, on the north bank of Tweed Creek near its junction with the East Branch of Octoraro Creek, where a dense green schist with light-colored lenses and streaks is in contact with the quartzose beds of the Peters Creek schist. The contact, which is obscure, suggests a crosscutting dike in the Peters Creek schist, but the greenstone schist under the microscope suggests an altered basic lava by the presence of chlorite, epidote, quartz, and plagioclase in a fine groundmass together with lenticular areas of calcite and epidote and quartz that may be amygdules. The rock contains a large amount of ilmenite altering to leucoxene and ilmenite intergrown with pyrite.

At a third locality, in Lancaster County near the mouth of Fishing Creek, green schist crops out in ledges of sharply inclined rock infolded in Peters Creek schist. The mineral composition of the green schist, calcite, epidote, and chlorite, with some albite, magnetite, and actinolite, is suggestive of tuffaceous layers interbedded with the Peters Creek sediments. It may represent a calcareous bed in the original sediment, but the absence of similar beds elsewhere in the Peters Creek schist makes the sedimentary origin of the green schist somewhat improbable.

The three outcrops described above show no field relations to the surrounding rock but suggest in their composition and manner of occurrence that they may belong to the altered lava flows that are more extensively developed in the Glenarm series farther southwest.[54]

IGNEOUS ROCKS OF UNDETERMINED PRE-TRIASSIC AGE

QUARTZ MONZONITE

Throughout the Mine Ridge anticline there are fragments of a light-colored rock of granitic texture, composed of quartz, feldspar, and a variable quantity of biotite that imparts an irregular gneissic

[54] Jonas, A. I., Pre-Cambrian rocks of western Piedmont of Maryland: Geol. Soc. America Bull., vol. 35, pp. 355–364, 1924.

banding to the rock where present in considerable amount. The rock usually appears in isolated occurrences as stray boulders, in the meadows or along roadsides, but in a few places, such as in the Pennsylvania Railroad cut north of Christiana, it forms layers of variable thickness in the amphibolites. In the cuts of the Philadelphia & Reading Railroad about 1 mile north of Coatesville it cuts across the schistosity of the pre-Cambrian schist and penetrates that rock in stringers, forming an undoubted igneous contact with the schist of sedimentary origin.

In many hand specimens it is possible to recognize the repeated twinning striation of a plagioclase feldspar, and under the microscope the constituents are seen to be albite, or albite and microcline, quartz, and more or less biotite. Some specimens are practically lacking in dark-colored constituents and might be termed alaskite. In other places there is as much as 25 per cent biotite and considerable muscovite. The rock everywhere shows evidence of strong mechanical deformation in the granulation of the quartz and in the numerous inclusions of zoisite, epidote, and quartz with which the albite individuals are crowded. The association of calcic minerals with albite points to the breaking down under dynamic action of a plagioclase that was originally more calcic than albite and the secondary formation of the albite. The variable percentage of microcline suggests that the original rock ranged in composition from a quartz monzonite to a quartz diorite.

A peculiar light-colored gneiss found upon the dump of the Gap nickel mine showed a pronounced development of "schachbrett" or chessboard albite, together with a high percentage of normal albite and some microcline. The occurrence in one or two other specimens of albite that shows indication of a "schachbrett" texture suggests that the microcline in some of the quartz monzonite has been replaced by albite.

QUARTZ MONZONITE PEGMATITE AND APLITE

In several localities, notably in the Pennsylvania Railroad cut north of Christiana, the quartz monzonite is cut by satellitic pegmatites that have been strongly deformed. The albite crystals in these pegmatites are filled with inclusions that indicate saussuritization of a calcic plagioclase. Like the magma with which they are genetically associated, the pegmatites contain varying amounts of microcline together with albite. The relations of these monzonitic rocks to the sedimentary and igneous complex and to the gabbro is shown in the Pennsylvania Railroad cuts north of Christiana and along the Philadelphia & Reading Railroad north of Coatesville, where they intrude both the gabbro and the sedimentary and igneous complex. The massive gabbro shows lenses, stringers, and dikelets of pegmatitic and aplitic material. In the foliated rocks the highly fluid magma has spread out in stringers that follow the foliation planes and form typical banded injection gneisses in the biotite schists and in the biotitic amphibolites. Large outcrops of these banded gneisses show the characteristics of lit-par-lit intrusion—in the lenticular areas of granitic material that fray out on the ends of the lenses into the including schist, in the pinching and swelling of the granitic bands, and in the ptygmatic folding[55] of the invading rock. In some places stringers, parallel to the foliation, can be traced into crosscutting dikelets. The evidence indicates, therefore, that after the invasion of the old complex by massive gabbro there was a batholithic intrusion of dioritic magma, whose highly fluid upper portion invaded both the massive rock and the foliates where well-established planes of schistosity offered easy pathways for intrusion.

QUARTZ DIORITE

A couple of isolated occurrences on the Mine Ridge upland show boulders of medium-grained light-gray rock of granitic texture, with a poorly defined gneissic banding. When examined under the microscope the rock shows considerable alteration. The original feldspar appears to have been chiefly oligoclase-andesine with so small an amount of potash feldspar that the rock may be termed a quartz diorite. Biotite is abundant, in some flakes accompanied by syngenetic chlorite. The biotite shows pleochroic haloes. A variable amount of quartz is present together with epidote, zoisite, and titanite.

The relation of the quartz diorite to the other igneous rocks and to the sediments is nowhere clearly shown, but at one place the diorite is intruded by a strongly deformed albitic pegmatite. It is highly probable that the quartz diorite is consanguineous with the quartz monzonite described above. It is impossible to separate these two rocks upon the areal map.

GRANITIC PEGMATITE AND GRANITE

There are no outcrops of true granite in the McCalls Ferry-Quarryville district, but in the northeastward extension of the Mine Ridge uplift about 1½ miles north of Downingtown on the New Holland branch of the Pennsylvania Railroad, there is exposed a pink microcline granite with a notable content of deep-blue quartz. This granite cuts a green hornblende-quartz diorite. Some of the quartz in the diorite shows a striking blue color, and the diorite is filled with

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[55] "Ptygmatic folding" (from πτύγμα, fold) is a term introduced by J. J. Sederholm, who defined it as folding "due to undulatory movements in a semimolten medium" (Comm. géol. Finlande Bull. 58, p. 85, 1923).

blue quartz veins. North of Coatesville, on the branch of the Philadelphia & Reading Railroad that parallels the West Branch of Brandywine Creek, massive gabbro and hornblende schists, injected with monzonitic aplites, are cut by veins of blue quartz that are folded together with the including rocks. It is possible that the blue quartz in the diorites and the blue quartz in the granites are integral parts of two different igneous intrusions, or it is possible that the blue quartz, which is better developed in the granite than in the diorite, is genetically related to the granite and is of secondary origin in the diorite.

Although granite outcrops are lacking in the west end of the Mine Ridge anticline, the abundant tourmalinization and pegmatitization throughout the flanking Paleozoic sediments is an evidence of underlying granite. In two localities, one about half a mile north of Atglen and the other half a mile north of Greentree, a pink microcline pegmatite heavily laden with magnetite and tourmaline cuts through the basal pebbly schistose beds of the Hellam conglomerate member of the Chickies quartzite. This pegmatite is accompanied by a large amount of pure quartz, and the quartz is loaded with tourmaline crystals. About 140 feet higher in the section tourmaline is abundantly developed along the bedding planes of white Chickies quartzite, interbedded with sericite schist carrying hematite crystals. One mile northeast of Simmonstown an albite-microcline pegmatite cuts across the bedding of basal Cambrian pebbly sericite schist.

AGE OF GRANITE AND QUARTZ MONZONITE

There are in the Mine Ridge anticline occurrences of igneous rocks that range in composition from quartz diorite to granite. The relations of the granite and quartz monzonite have not been determined, as the writers have nowhere seen them in contact, unless the hornblende-quartz diorite north of Downingtown is a part of the same intrusion as the quartz monzonite, in which case the granite is younger than the quartz monzonite. So far as known, with one possible exception, the outcrops of granite and quartz monzonite are confined to the pre-Cambrian sedimentary series, although pegmatites have been found in several places cutting Cambrian schists. The exception mentioned is in a cut where the Pennsylvania Railroad crosses the flanking sediments of the Mine Ridge anticline between Christiana and Atglen. Here a fine-grained rock of granitic texture occurs in the greenish sericite schist, with pebbly beds, that forms the base of the Paleozoic section. However, the outcrop is so much disintegrated that it is not possible to prove an intrusive contact between the Cambrian sediments and the granitic rock.

Nevertheless the universal distribution of epigenetic tourmaline throughout the Chickies quartzite, far too much to be merely recrystallized clastic tourmaline, and the presence in the Lower Cambrian schists of the magnetite-bearing pegmatites described above, together with the common occurrence of hematitic quartz, give conclusive evidence that the Cambrian sediments have been permeated by hot solutions and fumarolic vapors. It is, of course, possible, but it seems improbable that the pegmatites and ore-bearing solutions that penetrate the Paleozoic rocks are the only evidence at the surface of this cycle of igneous activity. If we search for evidence of the parent magma among the igneous rocks that have been discovered in the pre-Cambrian sediments there are three possibilities. The post-Cambrian magma may have been the source of the quartz monzonite or of the granite or of both. If it was the source of both, the granite would be consanguineous with the quartz diorite and quartz monzonite, although probably intruded during a welling up from the magmatic reservoir that succeeded the consolidation of the dioritic portion of the magma, as suggested by the lack of mechanical deformation in the granite contrasted with the quartz monzonite.

Support of the hypothesis that granite and quartz monzonite are derived from the same post-Cambrian magma is furnished by the composition of the pegmatites that cut the Lower Cambrian beds and the association of pegmatites of similar composition with both granite and quartz monzonite—facts that point to the derivation of granite, quartz monzonite, and pegmatite from a common magma. The fact that igneous rock of granitic texture has not been discovered in Paleozoic sediments except at one place, where the igneous contact could not be proved, may possibly be due to the lack of good exposures of the Lower Cambrian. It may also be accounted for by the hypothesis that the magma from which the quartz diorite, quartz monzonite, and granite have been differentiated has either invaded and largely swallowed up the pre-Cambrian sedimentary core of the Mine Ridge anticline or has been intruded as a concordant batholith whose upper limit is the plane of unconformity that marks the boundary between pre-Cambrian and Cambrian sediments. The highly argillaceous sediments of the upper beds of the pre-Cambrian would yield readily to injection by the more fluid portion of the invading batholith, but the more resistant quartzose beds of the Lower Cambrian would yield only to the permeation of the vapors and superheated solutions that emanated from the underlying batholith. That some highly active vapors derived from an underlying batholith penetrated upward for several thousand feet is shown by the presence of tourmaline nests in several places in the dolomites and limestones that overlie the Lower Cambrian arenaceous sediments. Such tourmaline nests have been found as far up as the Conestoga limestone, of probable Chazy age, thereby indicating that the igneous intrusions were post-Ordovician.

The hypothesis that the granite is post-Cambrian is somewhat shaken by the association of the granite with blue quartz, that seems to be the logical source for the blue quartz pebbles in the Cambrian conglomerate overlying the granite. Moreover, in the Reading and Durham Hills a granite that may be identical with the granite north of Downingtown furnishes material for the immediately overlying arkosic beds of the Lower Cambrian. However, the argument from the blue quartz is not conclusive, because the presence of blue quartz pebbles in both upper Glenarm and Paleozoic sediments shows that the blue quartz was a constituent of pre-Glenarm igneous rocks as well as of the post-Glenarm granites and thus was not confined to one period of igneous intrusion. If the quartz monzonite belongs to an earlier intrusion than the granite, the blue quartz associated with the quartz monzonite may be an original constituent of the diorite magma, and this may have furnished the quartz for the Paleozoic sediments.

In the absence of conclusive field evidence it is only possible to say that granite and quartz monzonite may well be differentiates of the same magma and that this parent magma may have been intruded in pre-Paleozoic time because of the absence of granite rocks in Paleozoic sediments and because of the presence in basal Cambrian sediments of blue quartz that may have been derived from the granite quartz. On the other hand granite was intruded in Paleozoic time, as shown by the effects of highly active vapors and solutions in Paleozoic sediments. Whether the quartz monzonite and granite in Mine Ridge are the source of these Paleozoic pegmatites or whether they are an older pre-Cambrian intrusion is still an open question.

TRIASSIC IGNEOUS ROCKS

DIABASE

Several diabase dikes cross the country with parallel northeasterly trends. These dikes are narrow and usually do not crop out. They weather into a trail of boulders, by means of which they may be traced across the country. In the few outcrops in which the diabase of the dikes can be measured it nowhere exceeds 100 feet in thickness.

In contrast to the other igneous rocks of the district the diabase is notably fresh, the only evidence of alteration being a slightly undulose extinction in some places and a little clouding of the feldspar and the formation of a small amount of chlorite and delessite. The essential constituents are labradorite, augite, and olivine.

Several of the diabase dikes of this district can be traced into the Triassic sediments farther to the north, in Lancaster and Berks Counties.

CORRELATION OF ROCKS OF M'CALLS FERRY-QUARRYVILLE DISTRICT WITH THOSE OF OTHER AREAS

The stratigraphic problems in the McCalls Ferry and Quarryville quadrangles were also recognized in Maryland during the areal mapping of Baltimore County, and similar problems exist in the region northeast of the Quarryville quadrangle as far as Trenton, N. J., where the pre-Cambrian and Paleozoic rocks disappear under Triassic strata and the sediments of the Coastal Plain. The pre-Cambrian rocks that lie east of the Triassic basin in the New York City area are lithologically identical with those of the Quarryville-Philadelphia belt, and the stratigraphic problems involved in their mapping appear to be the same as those in the Pennsylvania region. (See fig. 4.) In recent years Berkey[56] has published some conclusions concerning the age and correlation of the pre-Cambrian rocks of the West Point quadrangle that make it possible to suggest with reasonable certainty a correlation between the pre-Cambrian section along Hudson River south of Peekskill and that of southeastern Maryland and Pennsylvania. The probable stratigraphic equivalence is shown in the accompanying generalized correlation table.

The correlation of the pre-Cambrian rocks of the McCalls Ferry-Quarryville district with those of other areas of the Piedmont province involves (1) the relation of the metamorphosed pre-Cambrian sediments of the Honeybrook and Mine Ridge uplands to the Baltimore gneiss of Maryland and southeastern Pennsylvania and to the pre-Cambrian gneisses of the Reading and Durham Hills, and (2) the relation of the Glenarm series to the early pre-Cambrian Baltimore gneiss and to the Paleozoic rocks that occur in Chester Valley and on the flanks of the Mine Ridge uplift.

Correlation of the Baltimore gneiss in Maryland and Pennsylvania with the pre-Cambrian schists of the Highlands of New Jersey and New York.—The pre-Cambrian core of the Mine Ridge upland in the Quarryville quadrangle and in the Honeybrook upland is correlated with the Baltimore gneiss of Maryland because the schists of the Mine Ridge anticline are similar both in stratigraphy and history to the rocks that form the core of the anticlinal uplift north of Avondale and Kennett. The rock that occurs in the center of the anticline

[56] Berkey, C. P., and Rice, Marion, Geology of the West Point quadrangle: New York State Mus. Bull. 225–226, pp. 129–140, 1921. Berkey, C. P., and Sanborn, J. F., Engineering geology of the Catskill water supply: Am. Soc. Civil Eng. Trans., vol. 86, p. 7, 1923.

System	Series	Maryland: Baltimore County and Harford County. Knopf, E. B., and Jonas, A. I., Am. Jour. Sci., 5th ser., vol. 5, pp. 43–44, 1923; Maryland Geol. Survey, Baltimore County (in press).	Maryland: Carroll County. Jonas, A. I., Geol. Soc. America Bull., vol. 35, pp. 355–364, 1924.	Pennsylvania: McCalls Ferry-Quarryville district. This report.	Pennsylvania: Lancaster-York area. Stose, G. W., and Jonas, A. I., Washington Acad. Sci. Jour., vol. 12, pp. 358–369, 1922; Geol. Soc. America Bull., vol. 34, pp. 507–524, 1923; idem, vol. 38, pp. 505–536, 1927; Pennsylvania Top. and Geol. Survey Geol. Atlas, New Holland quadrangle, No. 179, pp. 1–15, 1926.	Pennsylvania: Trenton, Philadelphia, and Wilmington areas. Bascom, F., U. S. Geol. Survey Geol. Atlas, Folios 167, 1909; 162, 1909; 211, 1920. Bliss, E. F., and Jonas, A. I., U. S. Geol. Survey Prof. Paper 98, pp. 9–34, 1917.	New Jersey: Northern New Jersey. Bayley, W. S., Kümmel, H. B., and Salisbury, R. D., U. S. Geol. Survey Geol. Atlas, Folio 191, 1914.	New York: New York City and Westchester County. Berkey, C. P., and Rice, Marion, New York State Mus. Bull. 225–226, 1921.	New York: West Point quadrangle. Berkey, C. P., and Rice, Marion, New York State Mus. Bull. 225–226, 1921; Fettke, C. R., New York Acad. Sci. Annals, vol. 23, pp. 193–260, 1914.
		Woodstock granite and pegmatite (possibly Paleozoic).		Granite pegmatite.	[Granite pegmatite.]				
Ordovician.			Conestoga [Frederick] limestone. (Rests on Harpers phyllite.)	Conestoga limestone (lower part of Chazy? age). (Rests in places on Harpers schist.)	Conestoga limestone. (Rests in places on lower formations down to Harpers schist.) / Cocalico shale (lower part contains fossils of lower? Trenton age). (Rests in places on Beekmantown limestone.)	Octoraro schist [equivalent in part to Peters Creek schist; the rest to Wissahickon albite-chlorite schist].	Martinsburg shale ("equivalent of Trenton and Utica").	[Apparently not represented, although some authors place Manhattan schist, Inwood limestone, and Lowerre quartzite here.]	Hudson River formation (Trenton to Cincinnatian age).
					Unconformity.		Jacksonburg limestone ("equivalent of Lowville, Black River, and lower Trenton").		
			Unconformity.	Unconformity.	Unconformity.				
					Beekmantown limestone.				
Cambrian.	Upper Cambrian.				Conococheague limestone.				Wappinger limestone (Lower Cambrian to Trenton age).
	Middle Cambrian.				Elbrook limestone.	Shenandoah limestone.	Kittatinny limestone.		
	Lower Cambrian.			Equivalent to Tomstown dolomite: Ledger dolomite.	Ledger dolomite.				
				Kinzers formation.	Kinzers formation.				
				Vintage dolomite.	Vintage dolomite.				
			[Antietam quartz schist.]	Antietam schist.	Antietam quartzite.				
			Harpers phyllite.	Harpers schist.	Harpers phyllite.				
			[Chickies formation.]	Chickies quartzite, including Hellam conglomerate member at base.	Chickies quartzite, including Hellam conglomerate member at base.	Chickies quartzite.	Hardyston quartzite.		Poughquag quartzite.
			Unconformity.	Unconformity.	Unconformity.	Unconformity.	Unconformity.		Unconformity.
Algonkian.		Granite and granite pegmatite (possibly Paleozoic).[a] Port Deposit gneiss. Metagabbro and serpentine.	Metagabbro and serpentine.	Granite (Paleozoic or older). Quartz monzonite and quartz diorite (Paleozoic or older). Metagabbro and serpentine.	[Granite (Paleozoic or older).] [Quartz monzonite and quartz diorite.] Gabbro (and anorthosite) in Welsh Mountain and Honeybrook upland.	Gabbro gneiss.[b] Granodiorite and granite gneiss.	Pre-Cambrian: Byram gneiss (predominantly granite gneiss). Losee gneiss (predominantly diorite gneiss). Pochuck gneiss (probably comprising an older amphibolite and a younger metagabbro).	Basic dikes. Pegmatites. Yonkers gneissoid granite. Pegmatites. Ravenswood granodiorite. Injection granites and pegmatites. Fordham gneiss (in part injection gneiss). Occasional basic injections. Serpentines. Hornblende schists.	Basic dikes. Pegmatites. Storm King granite. Pegmatite. Reservoir granite. Canada Hill granite and associated injection gneiss. Pegmatites and magnetites. Occasional basic injections. Peekskill diorite injection gneiss. Pochuck gneiss (possibly equivalent of hornblende schist layers in Baltimore gneiss).
				Peach Bottom slate.	[Absent.]				
				Cardiff conglomerate.					
		Peters Creek schist.	Peters Creek schist.	Peters Creek schist.					
	Glenarm.	Wissahickon formation: Albite-chlorite schist. Oligoclase-mica schist. } Metamorphic facies.	Wissahickon formation: Albite-chlorite schist. Oligoclase-mica schist. } Metamorphic facies.	Wissahickon formation: Albite-chlorite schist. Oligoclase-mica schist. } Metamorphic facies.		Wissahickon mica gneiss. (Includes part of the Peters Creek schist.)	[Absent.]	Manhattan schist.	Manhattan schist.
		Cockeysville marble.	Metabasalt and tuffaceous slates interbedded with Cockeysville marble.	Amphibolite schist and chlorite schist (altered dikes and flows). Cockeysville marble.	Metabasalt and aporhyolite in Hellam Hills and Pigeon Hills. (Cockeysville marble absent.)	[Apparently represented in Elkton-Wilmington area but mapped as Chickies quartzite and Shenandoah limestone.]		Inwood limestone.	Inwood limestone.
		Setters formation.	Setters formation.	[Not exposed.]	[Absent.]			Lowerre quartzite.	Lowerre quartzite.
		Unconformity.	Unconformity.			Unconformity.		Not recognized.	
Archean.		Hartley augen gneiss (intrusive). Baltimore gneiss.	Baltimore gneiss.	Baltimore gneiss (graphite schists and limestone lenses) in Mine Ridge upland.	Baltimore gneiss (graphitic schists and limestone lenses) in Welsh Mountain and Honeybrook upland.	Baltimore gneiss. Franklin limestone.	Franklin limestone and graphitic schists.	Interbedded limestone and schist. Fordham gneiss (in part) (paragneiss).	Grenville metamorphics=mica schists, metaquartzites, metalimestones (Sprout Brook limestone). Paragneisses (Fordham gneiss).

[a] Cuts the Port Deposit gneiss in Harford County, Md.

[b] North of the Elkton and Wilmington quadrangles a granite is reported to cut the gabbro (Hawkins, A. C., Am. Jour. Sci., 5th ser., vol. 7, p. 360, 1924).

that extends from Avondale to Chadds Ford is known to be Baltimore gneiss, both because it is in line of strike with certain anticlines in Maryland that bring the Baltimore gneiss to the surface and because it underlies later pre-Cambrian strata.

The Baltimore gneiss of the Mine Ridge and Honeybrook uplands, together with the interbedded limestones, is correlated with the pre-

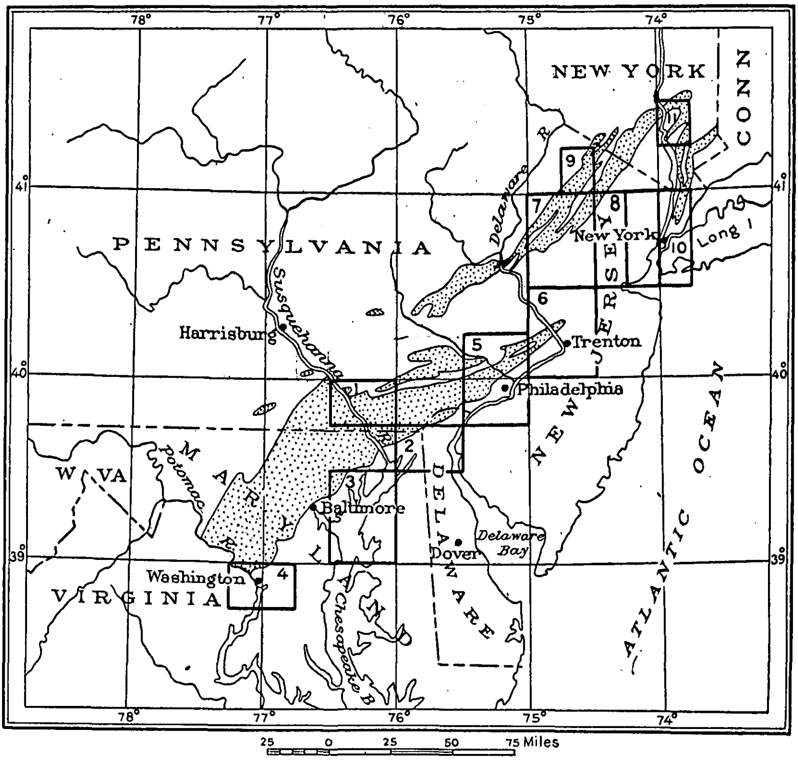

FIGURE 4.—Sketch map showing principal pre-Cambrian areas of the Piedmont province and Appalachian Mountains in part of Maryland, Pennsylvania, and New Jersey. 1, McCalls Ferry and Quarryville quadrangles; 2, Elkton and Wilmington quadrangles (U. S. Geol. Survey Geol. Atlas, Folio 211); 3, Tolchester quadrangle (Folio 204); 4, Washington quadrangle (Folio 70); 5, Philadelphia district (Folio 162); 6, Trenton quadrangle (Folio 167); 7, Raritan quadrangle (Folio 191); 8, Passaic quadrangle (Folio 157); 9, Franklin Furnace quadrangle (Folio 161); 10, New York City district (Folio 83); 11, West Point quadrangle (New York State Mus. Bull. 225–226)

Cambrian graphitic schists and limestones of the Reading and Durham Hills, which are lithologically similar to the rock in the center of the Mine Ridge anticline. Moreover, the pre-Cambrian rocks of the Reading and Durham Hills occupy the same stratigraphic position as the pre-Cambrian core of the Mine Ridge upland in that they unconformably underlie the Paleozoic rocks and are cut by a series of petrographically related igneous rocks that are presumably of late pre-Cambrian age. The old pre-Cambrian sedimentary gneisses and associated intrusive rocks of the Reading Hills continue northeastward through the Highlands of New Jersey into the Highlands of New York, which cross Hudson River north of Peekskill. These old pre-Cambrian rocks are the so-called Grenville metamorphics of Berkey's section, but the relation of this pre-Cambrian schist in the Highlands to the so-called Grenville schists of the Adirondacks and to the Grenville series of the type locality in Ontario is not yet definitely established.

Correlation of the Glenarm series and the Manhattan, Inwood, and Lowerre formations of New York.—The Manhattan, Inwood, and Lowerre formations are restricted in their occurrence to the New York City area south of the Highlands. They are considered by Berkey to be of Grenville age and to be part of a conformable section whose base is the Fordham gneiss, which he correlates with the banded Grenville gneisses of the Highlands. There is a strong similarity in lithology and thickness between the Glenarm series and the Manhattan, Inwood, and Lowerre formations, although the Cardiff conglomerate and Peach Bottom slate which have been provisionally placed in the Glenarm series, are apparently absent in the New York section. The rocks of the Glenarm series and underlying Baltimore gneiss in Pennsylvania have been driven northward from their original position by an overthrust fault, and they are absent from the stratigraphic succession in the Reading and Durham Hills, the Honeybrook and Mine Ridge uplands. In New York the crystalline schists of the Manhattan-Inwood-Lowerre succession come to an abrupt stop on the southern margin of the Highlands, just as the Glenarm schists cease on the south side of Chester Valley in Pennsylvania, nowhere reappearing to the north. There is thus a considerable ground for the belief that the Manhattan-Inwood-Lowerre succession may prove to be the northeastward continuation of the Glenarm series thrust northward over the southern rocks of the Highlands.

Apparently the unconformity that marks the base of the Glenarm series in Maryland has not been recognized in New York, for, according to Berkey, there is no unconformity between the Lowerre quartzite, where present, and the underlying Fordham gneiss, or between the Inwood limestone, where the Lowerre is absent, and the Fordham gneiss. However, Berkey [57] makes the point that a selective granitization has invaded and in many places completely transformed the strata underlying the Inwood limestone, while the limestone appears to have shown a remarkable resistance to the action of igneous magma. Such granitization may have obscured the relation

[57] Berkey, C. P., op. cit., p. 136.

between the Lowerre and the underlying Fordham gneiss, so that an unconformity between the Lowerre and the Fordham is unrecognizable, or, as suggested by Berkey, the Fordham gneiss may be really the downward continuation of the Manhattan-Inwood succession where the Lowerre is absent. However, in view of the fact that a comprehensive injection of both the upper beds of the Fordham and the lower beds of the Manhattan-Inwood-Lowerre succession would easily obliterate evidence of an unconformity, it seems best at present to correlate the Fordham gneiss with the Baltimore gneiss and the Manhattan, Inwood, and Lowerre formations with the Glenarm series.

STRUCTURE

The McCalls Ferry-Quarryville district is traversed by the southern part of a compound syncline, a major anticline, a major syncline, an overthrust fault of great magnitude, and normal faults that cross the northern part of the district.

The major anticline is the Mine Ridge anticline, which with its spurs forms a belt 4 to 7 miles wide, extending from the northeast corner of the district southwestward nearly to Susquehanna River. This anticline lies south of a more gently folded synclinal valley which is the southern part of the valley area of the New Holland and Lancaster quadrangles. The axis of the Mine Ridge anticline crosses the southern upland as the Tucquan anticline. The Mine Ridge anticline and its southwestern continuation, the Tucquan anticline, lie northwest of a wide syncline, the Peach Bottom syncline, so called from the youngest rock infolded in it, the Peach Bottom slate.

The district is traversed by a low-angle overthrust fault called the Martic overthrust, which has carried the pre-Cambrian rocks of the southeastern part over the Paleozoic rocks of the northern and northwestern part. Subsequent normal faulting in the northern part of the district has broken into blocks the Paleozoic rocks and the overlying overthrust mass. The relation of these structural features to the structure of a larger area may be seen in Figure 5.

FOLDS

The northern part of the district is occupied by a syncline which extends from a point just east of Gap through Strasburg, New Danville, and Washingtonboro and across the Susquehanna River north of Long Level. It is underlain by the Conestoga limestone and contains, in the eastern part, anticlines which bring to the surface older Paleozoic rocks. This syncline is the southern part of the synclinal valley area which lies south of the Welsh Mountain and

Chestnut Hill anticlines of the New Holland and Lancaster quadrangles and north of the Mine Ridge anticline and its outlying folds.

The Mine Ridge anticline is the southwestern spur of the broad Honeybrook upland, which is northeast of the Quarryville quadrangle. The anticline is a double fold which crosses the northern

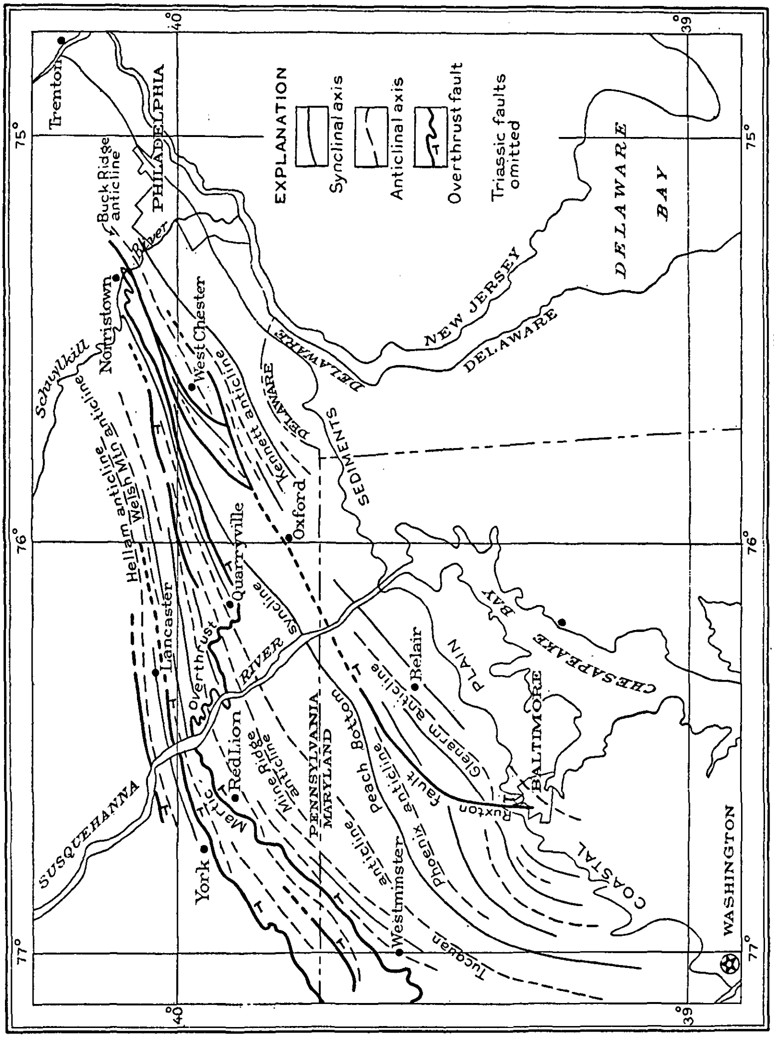

FIGURE 5.—Generalized structure of the region between Philadelphia and Washington

part of the Quarryville quadrangle in a direction S. 75° W. It exposes a core of pre-Cambrian rocks with overlying Lower Cambrian rocks on its flanks, dipping 75° S. on the southeast side and more gently northwest on the northwest side. North and northwest of the Mine Ridge anticline there are a series of anticlinal folds which expose Antietam and Harpers schists in the center and overlying limestones on the flanks. Bunker Hill and Oak Hill are formed by these anticlines, which extend S. 80° W. to Safe Harbor, where their west ends are beveled by overthrust Wissahickon albite-chlorite schist. The Shenks Ferry syncline lies south of these anticlinal hills from Susquehanna River to Oak Hill.

The Lower Cambrian quartzites and schists on the southwestern nose of the Mine Ridge anticline plunge steeply under Conestoga limestone of the Quarryville Valley, which in turn is overlain by overthrust Wissahickon schist. The folding of the Mine Ridge anticline is continued southwestward in the Wissahickon albite schist, where it is known as the Tucquan anticline. The southwestern continuation of the southern crest of the Mine Ridge anticline crosses the Susquehanna at Tucquan and extends southwestward through York County into Maryland. There it has been recognized as far as Woodbine, on the Carroll-Howard County line. The northern spur of the anticline crosses the river north of Pequea and south of Martic Forge, and it also extends southwestward into Maryland. At a point near Chanceford, York County, it brings to the surface a small area of Cockeysville marble, the formation which normally underlies the Wissahickon schist.

Railroad cuts along the left bank of Susquehanna River afford excellent opportunities for studying the structure of the overthrust rocks, continuously from Creswell Station for 20 miles to the southern edge of the district except for three-fourths of a mile near Shenks Ferry Station, where rocks of the autochthonous block cross the river. The major folds in the Wissahickon albite schist—the Tucquan and Pequea anticlines—are open folds, not overturned, with dips of 20° to 40°. The minor folds, however, are much compressed and the rocks are finely crumpled with quartz injections along the bedding. From Shenks Ferry to a point north of Safe Harbor the minor folds are recumbent toward the southeast. A cleavage more or less steeply inclined to the bedding occurs throughout the overthrust and autochthonous beds. The Paleozoic rocks of the district are closely folded, but the folds are for the most part not overturned. The less resistant of them are finely crumpled, and this fine crumpling is especially well developed in the Harpers and Antietam schists and the less massive beds of the Conestoga limestone.

The Peach Bottom syncline lies south of the Mine Ridge-Tucquan anticline and trends S. 50° W. across the southeastern part of the area. This syncline is occupied for the most part by the Peters Creek formation, which has an average width of outcrop of 5 to 8 miles. Cardiff conglomerate and Peach Bottom slate occupy the deepest part of the fold from Kings Bridge to Susquehanna River and extend southwestward as far as Pylesville, Md., 3½ miles southwest of Delta and Cardiff, at the Pennsylvania-Maryland line.

In the area northeast of Susquehanna River the syncline is overturned toward the northwest, and the beds on both limbs dip 50° to 60° SE. The dip of the cleavage also is southeast. The overturning was accompanied by a break on the northern limb of the fold, along which the Cardiff conglomerate is faulted out for 5 miles along the strike. The Peach Bottom syncline has been traced 50 miles southwest of this district to Sykesville, Md., where the intrusive granite that extends from Sykesville to Washington and southward interrupts it.

FAULTS

The principal fault of the region is an overthrust fault which carried pre-Cambrian schists of the southeastern Piedmont northwestward over the Paleozoic sediments of the northern part of the area. This fault has been named by the writers the Martic overthrust from the prominent line of hills south of the front of the fault in Marticville Township, east of Susquehanna River. The trace of the fault plane on the surface is the northern boundary of the Wissahickon albite schist and the northern area of the Peters Creek schist. This boundary follows the south side of the Chester-Quarryville Valley; just west of Quarryville it turns northwest for 5 miles, thence runs southwest again to Susquehanna River at Shenks Ferry. North of Shenks Ferry the trace of this fault swings northwestward in a wavy line nearly to Creswell Station and crosses the river at Long Level, southwest of which it passes out of the McCalls Ferry quadrangle.

All observed exposures of the fault contact are poor, but the best outcrop occurs in a cut of the Atglen & Susquehanna branch of the Pennsylvania Railroad 1 mile west of New Providence. This exposure shows Conestoga limestone in contact with folded Wissahickon albite schist, but the rocks are weathered and their relations are not clear. There is no evidence of a great amount of brecciation of rock along the plane of movement in any of the observed exposures. Near Quarryville the contact of albite schist and Conestoga limestone is a zone of limonite replacement. The limonite has cemented quartz fragments which may be part of a breccia produced along the fault contact. This limonite-quartz rock is best exposed in a cut of the Pennsylvania Railroad 2 miles west of Greentree. At many places along the fault contact in the vicinity of Quarryville limonite has been mined in the past, but the weathered exposures now to be seen throw no light on structural problems.

The existence of this extensive overthrust fault has therefore not been established by the discordant relations seen in exposures of fault contacts. It has rather been inferred from the abnormal contacts of the pre-Cambrian Wissahickon albite schist and other pre-Cambrian formations with rocks of Cambrian and Ordovician age on the north and northwest. From the irregular surface outline of the front edge produced by erosion and from the occurrence of "windows" or "fensters" of Conestoga limestone surrounded by Wissahickon albite schist near the northern edge of the overthrust mass west of Quarryville it is evident that the fault plane has a low dip.

This overthrust is part of a major fracture which has been traced from Schuylkill River north of Philadelphia across Pennsylvania and Maryland, where it is covered by Triassic sediments north of Union Bridge. The rocks of the overthrust mass are for the most part pre-Cambrian, comprising the Baltimore gneiss, the Glenarm series, and the intrusive and extrusive rocks; but in southern York County, Pa., and in Carroll, Frederick, and Montgomery Counties, Md., basal Cambrian rocks that overlie these pre-Cambrian crystalline rocks have participated in the movement.

It is not known how far northwestward the overriding block moved, but the distance must have been considerable. At present it covers the southern and greater part of the syncline south of the Mine Ridge anticline and the southwestern extension of the Mine Ridge anticline across Pennsylvania and Maryland. From the neighborhood of Susquehanna River into York County it overlies the anticlines north of Mine Ridge, but these anticlines emerge from the thrust in southern York County and Maryland. Southeast of these folds which emerge from its edge it covers a large area whose structure and rocks are a matter of conjecture.

It is believed that overthrust rocks extended over the Mine Ridge anticline and the region north of it, but at present its northwestern limit on Susquehanna River is 11 miles north of its front south of Quarryville. The Mine Ridge anticline, which has been a rising area since pre-Cambrian time, was probably higher at the time of thrusting than the region along Susquehanna River and southwestward, and the result of resistance to forward progress of the thrust offered by the Mine Ridge anticline was a thinning of the overthrust mass in that area. Folding subsequent to thrusting has further elevated the Mine Ridge area and made it easier for erosion to remove the overthrust in that area than to the southwest.

The Mine Ridge anticline has been thrust northwestward along a fault roughly parallel to the line of the Martic overthrust south of Chester Valley. This thrust has been traced northeast of Downingtown, but from structural relations northwest of the Mine Ridge anticline it seems to die out in the valley near Martinsville. The Mine Ridge anticline has moved across the syncline which lay south of Oak Hill and across the eastern ends of the anticlinal folds south of London Grove. A minor thrust extends for 5 miles north of Conestoga but is cut off by normal faults.

The northern part of the Quarryville-McCalls Ferry area is cut by normal faults trending about S. 70°–80° W. This area lies in the southern part of the northwestern Piedmont of Pennsylvania and Maryland, which is considerably broken by block faulting, and the normal faults which cut it are parts of faults of wider extent.

Welsh Mountain, Barren Hill, and Mine Ridge upland and its spurs are broken by a series of more or less parallel faults which produce marked effects in both pre-Cambrian and Paleozoic rocks. They are readily traceable except in the Conestoga limestone, in which it is difficult to recognize individual beds for any distance along the strike.

The most northerly fault of the Quarryville area is the fourth of the series of faults lying south of Welsh Mountain. It offsets the southern spurs of Barren Hill and passes south of Compass and westward into the Quarryville quadrangle near Kinzers. It separates the anticline of the Kinzers formation and the Conestoga limestone of the Bellemont area and extends westward into the Conestoga limestone, where for 15 miles it can not be certainly traced. It is in strike with the fault that runs west of Letort to Creswell Station and Long Level, where it cuts off the northern edge of the overthrust Wissahickon of the Turkey Hill area. The most extensive normal fault of the area is traceable 23 miles on the north side of Mine Ridge westward to a point north of Highville, where it offsets the Martic overthrust. In the vicinity of Burnt Mills a series of cross faults that strike southwest break up the folds of the Paleozoic rocks and offset the Wissahickon schist west of Safe Harbor and at Shenks Ferry station. These normal faults trend in the general direction of the post-Paleozoic folding and also cut the Paleozoic trends at angles of N. 30°–50° E. The latter direction is about that of the strike of the Triassic diabase dikes. Southwest of the McCalls Ferry-Quarryville district normal faults with a trend of about N. 30° W. are common in addition to those found in this district.

The movement seems to have been largely vertical and was accompanied by brecciation, which is not observed along trace of the Martic overthrust fault. A fault 1 mile north of Burnt Mills passes through a brecciated zone of Vintage dolomite and Conestoga limestone along which argentiferous galena occurs in small veins. Brecciation of Vintage dolomite has been noted at other places along these faults. Dolomite sheared and made slaty by faulting is well exposed on the north side of Pequea Creek at the trolley bridge west of Burnt Mills, and limonite replacing limestone occurs along the fault zones north of Shenks Ferry Station and just west of Safe Harbor, 1 mile southeast of Highville. Six of these normal faults cut the overthrust between Shenks Ferry and Creswell stations. There may be others which are not apparent; the most noteworthy extends from Letort, Creswell, and Long Level southwestward in York County, a distance of 15 miles. It has dropped down the overthrust rocks and uplifted Conestoga limestone which is in contact with them.

AGE OF THE STRUCTURE

The direction of the major folds of this district is in general southwest, parallel to the long folds of the Appalachian Mountains. Because of this parallelism, and because Paleozoic rocks are the latest rocks involved in the folding, the age of the dominant folds is considered to be post-Paleozoic, or the same as that of the Appalachian structure elsewhere.

Although the rocks of the region have major trends parallel to the Appalachian folding, they show also evidence of older folding, both pre-Cambrian and Paleozoic, as well as a younger structure acquired during Triassic deformation. All these structural disturbances are described below. The later Cretaceous, Tertiary, and Pleistocene deformation was in the nature of simple tilting and warping.

PRE-CAMBRIAN FOLDING

The pre-Cambrian rocks show evidence of at least one period of pre-Cambrian folding and probably two. In Maryland the rocks of the Baltimore gneiss basement have a trend discordant to that of the overlying Glenarm series, and hence had undergone folding before the younger series was deposited.

It is probable that the Glenarm series also was folded before Cambrian time, but subsequent folding has obscured the amount of late pre-Cambrian deformation. In the region of the Mine Ridge upland, however, elevation and complete removal of the Glenarm series took place before Cambrian deposition if Glenarm sediments were ever deposited in that region.

PALEOZOIC FOLDING AND FAULTING

The earliest Paleozoic movement of which there is record in this region is the uplift and arching along the Mine Ridge axis and to the northeast of it. This uplift occurred after the deposition of the Beekmantown limestone and before the Conestoga limestone of Chazy age was laid down. A later folding which involved the Conestoga limestone and which followed the same direction preceded the overthrust which bevels the folds.

The overthrust and underlying rocks alike were subsequently folded into longitudinal folds trending in general southwest, parallel to the folds of the Appalachian region. Although the youngest rock in this district that is involved in this folding is the Ordovician

Conestoga limestone, Carboniferous rocks in other parts of the Appalachian folded belt are affected but Triassic rocks are not. Hence this latest folding is assigned to post-Carboniferous time.

The age of the folding that involved the Conestoga limestone and preceded the overthrust and the age of the overthrust fault remain to be discussed. Although most of the folding once referred to the Taconic revolution [58] has been shown to be the result either of Middle Devonian or Appalachian mountain building, yet there is some evidence in the eastern United States of movement at the end of the Ordovician period. In southeastern New York, New Jersey, and northeastern Pennsylvania the Silurian Shawangunk conglomerate overlies unconformably Ordovician shales with a difference in dip of about 15°. Berkey [59] describes folding in Ordovician shales of the Hudson River area north of Kingston and pictures folded shales and an unconformity below the Shawangunk conglomerate in sections drawn along the line of the New York Aqueduct. This contact shows evidence of tilting, folding, and erosion at the end of the Ordovician in this part of the St. Lawrence trough. The absence of Silurian and Devonian beds west of the Green Mountains [60] is considered further evidence of folding at the same period in the northeastern part of this trough.

The mountain making which preceded the Upper Devonian epoch in New Brunswick and New England and in Appalachia northeast of the Piedmont of Pennsylvania seems to have been more vigorous than that of post-Ordovician time, but whether it was accompanied by folding is not known. Although it is not possible to fix the exact age of the prethrust folding it is known that the McCalls Ferry-Quarryville district and the Piedmont Plateau in general were subjected to diastrophism throughout the Paleozoic era. This area was part of Appalachia, that old land mass which arose in pre-Cambrian time and persisted as a borderland along the east side of the North American continent until after the Paleozoic era. The directions of folding that were established in pre-Cambrian time were followed by the later folding. Thus post-Beekmantown folding followed old trends of the pre-Cambrian rocks of the Mine Ridge-Honeybrook area, and so did later Paleozoic folding that involved the Conestoga limestone. The major folds that were developed in post-Carboniferous time were parallel to these old trends, and hence were posthumous folds along lines initiated by pre-Cambrian folding of the southeastern part of Appalachia.

[58] Clark, T. H., A review of the evidence for the Taconic revolution: Boston Soc. Nat. Hist. Proc., vol. 36, p. 163, 1921.

[59] Berkey, C. P., and Sanborn, J. F., Am. Soc. Civil Eng. Trans., vol. 86, pp. 1–9, 1923.

[60] Schuchert, Charles, Geol. Soc. America Bull., vol. 34, pp. 178–179, 1923.

AGE OF THE MARTIC OVERTHRUST

It is known that the overthrust occurred later than the folding of the Ordovician Conestoga limestone, which it bevels. It also preceded the latest Appalachian folding, because that folding involves both the overthrust and overlying rocks.

The folding in Appalachia was initiated in pre-Cambrian time. It is assumed as probable that the Baltimore gneiss has been practically inert since that time; also that the zones of folding in the early part of the Paleozoic era lay in the southeastern part of Appalachia and throughout that era moved northwestward into the belts of active sedimentation, while the land to the southeast kept rising and became immune to further folding. The McCalls Ferry-Quarryville district has been in a zone of active folding since the Beekmantown epoch, but it is probable that by post-Mississippian time the region to the southeast had become resistant to folding, and pressure found relief there by breaking and thrusting, which moved the rocks of the Martic overthrust northwestward, bringing them again into a zone where along with the underlying mass they were involved in the Appalachian folding.

Both southeast and northwest of the Martic overthrust are other thrust faults that are traceable for considerable distances. The most southeasterly known is the Ruxton overthrust of Baltimore County, Md. (See fig. 5.) This extends from Baltimore northeastward to the Phoenix anticline, and its probable connection with thrusts in Pennsylvania has been indicated on the generalized structure diagram. The Ruxton thrust and its probable northeastern extension in Pennsylvania are entirely in pre-Cambrian rocks; hence the age of the faulting is not known. It is possible, however, that the thrusting was later than the metamorphism and folding of these rocks, because the faulting has broken up the folds, especially in Pennsylvania, and has shortened the width of the belt of highly metamorphosed rocks in the vicinity of Philadelphia to no more than one-half of its width in Baltimore County. The Martic overthrust both in Pennsylvania and in Maryland is paralleled roughly on the north and northwest by other thrust faults. West of the Blue Ridge and 25 miles west of the Martic overthrust in Maryland there are thrust faults that extend southward through Virginia, Tennessee, and North Carolina and cut Carboniferous rocks. These faults mark the outer border of the overthrusts and may be somewhat later than the Ruxton and Martic faults. It seems probable that the zone of thrusting has moved progressively westward and that if the Martic overthrust belongs to the same period of deformation as the faults west of the Blue Ridge, the post-Carboniferous, it occurred in the early part of the period.

CLOSE FOLDING OF MINE RIDGE ANTICLINE

The Paleozoic formations on the flanks of the Mine Ridge anticline and in the folds lying to the north and west of this anticline are closely folded and greatly metamorphosed. This condition represents a locally intense metamorphism in the vicinity of the Mine Ridge anticline which dies out rapidly northward and northeastward. The cause of this locally strong metamorphism is discussed at length on pages 118 to 141, where the conclusion is reached that it may be explained as the result of regional contact influence of a post-Cambrian batholith.

The one pre-Triassic structure which all the rocks of the district possess in common is a cleavage, which was probably produced in the latest period of post-Carboniferous folding. This cleavage or schistosity is well developed in the softer rocks, the Conestoga limestone and the Antietam and Harpers schists.

The Antietam schist in an anticline north of London Grove (see pl. 6, *A*) shows a steep northwesterly schistosity which cuts across gently dipping beds of the anticline. Another outcrop in the same formation on the southwest limb of the Oak Hill anticline (see pl. 6, *B*) shows northwest schistosity brought out by weathering across the crumpled bedding outlined by quartz injection, which may be related in origin to the post-Cambrian batholith.

LOAD

The load under which the overthrusting and subsequent folding took place is not known. The nearest record of late Ordovician, Silurian, and Lower and Middle Devonian sedimentation occurs in the Appalachian Valley of Maryland and Pennsylvania, where a probable maximum of about 5,000 feet of sediments was deposited. There is no evidence of how far southeast these sediments extended, but it is believed that they covered a considerable area southeast of their present limits. In Upper Devonian[61] and lower Mississippian time a great thickness of sediments was deposited in central Pennsylvania, and although no remnant of them has been found in this area the mountain range that furnished them is considered to have been situated east of the present western margin of the Coastal Plain. The area between these mountains and the Upper Devonian delta of central Pennsylvania is thought to have been a piedmont gravel plain, while the shallow Upper Devonian sea extended west of the delta.

Late Ordovician, Silurian, and Devonian sediments therefore probably constituted the moderate load under which Appalachian thrusting and folding in this region took place.

[61] Barrell, Joseph, Upper Devonian delta of the Appalachian geosyncline: Am. Jour. Sci., 4th ser., vol. 37, p. 232, 1914.

TRIASSIC STRUCTURAL FEATURES

The McCalls Ferry-Quarryville district lies 15 miles south of the Triassic area of Pennsylvania and forms part of an area which has been considerably broken by normal faults, some of which are traceable into the Triassic rocks. These faults[62] both parallel the long Appalachian folds and cut across them. Near Ephrata, 14 miles north of the Quarryville quadrangle, blocks of Triassic rocks are faulted down in a syncline of Cocalico shale. Triassic faults break up the rocks of the northern part of the McCalls Ferry-Quarryville district into a series of blocks nearly parallel to the strike of the Appalachian folding.

Incorrect interpretation of Triassic deformation at some distance from Triassic basins has caused many errors in unraveling the structure in the Appalachian region. Recognition of normal faults in the areas of the Phoenixville, Honeybrook, New Holland, and Lancaster quadrangles, north of the McCalls Ferry-Quarryville district, and tracing of these faults across the York and Hanover quadrangles into Carroll and Frederick Counties, Md., has been the means of avoiding this error in this closely folded region, where thrust faulting rather than tension faulting would be expected and where fault outcrops are not preserved.

The normal faulting occurred during the crustal deformation which set in at the end of the deposition of the Newark group. This deformation produced the Palisade mountain system, in general parallel to that of the Appalachian Mountains. The stress was tensional, breaking into blocks the rising arch which lay east and southeast of the Pennsylvania-Maryland Triassic basin and west of the Connecticut Valley Triassic area. The record of this post-Triassic mountain making is left in the rocks of this region, although erosion has removed the Triassic sediments which probably were once there.

GEOLOGIC HISTORY

STRATIGRAPHIC RECORD

PRE-CAMBRIAN HISTORY

The oldest record of pre-Cambrian time in the Piedmont province, between Philadelphia and Baltimore, is the sedimentary Baltimore gneiss, derived from sands mixed with feldspathic débris and clay. In the northeastern part of the Piedmont province, in Pennsylvania, these sediments contained a considerable amount of carbonaceous matter and numerous calcareous lentils. The Baltimore gneiss in-

[62] Stose, G. W., New type of structure in the Appalachians: Geol. Soc. America Bull., vol. 35, pp. 465–480, 1924.

cludes bands of hornblendic composition, rich in iron oxides and sphene, from whose presence it may be inferred that the ancient sediments were either intruded by sills of basic igneous rock or interlaminated with basaltic flows.

The resultant sedimentary and igneous complex was deeply buried and profoundly altered, partly no doubt under the influence of high temperature and pressure generated by deep burial, and partly under the action of the strong tangential thrust of a period of mountain building, by which the sediments and included igneous material were folded and welded together into an intensely metamorphosed complex. Doubtless as an accompaniment to the folding, a granite batholith invaded the lower beds of the sedimentary series, furnishing to the upper beds alkalic vapors and a highly fluid magma that penetrated along the bedding planes of the batholithic cover, forming a banded gneiss by lit-par-lit injection. The folding, which either accompanied or followed the batholithic invasion, has left its mark in the intense cataclastic deformation of the granite, which has been stretched and crushed into a typical augen gneiss, known in Maryland as the Hartley augen gneiss. The trend of this ancient pre-Cambrian folding can be identified by a distinct divergence in strike between the early pre-Cambrian Baltimore gneiss and the overlying later pre-Cambrian Glenarm series.

The coarse texture of the granitic intrusions in the Baltimore gneiss shows that the folding and metamorphism took place under a thick cover. Prolonged erosion then removed this thick cover of sediments and exposed the coarsely crystalline core of the granite batholith. Upon the eroded surface the basal sediments of the Glenarm series were laid down. The irregular distribution and progressive overlap of Setters, Cockeysville, and Wissahickon deposits across the underlying Baltimore gneiss indicates that the Glenarm sediments, if marine, were laid down in a rapidly transgressing sea. The complete and in many places intense metamorphism of the Glenarm series makes it impossible to say with certainty whether the conditions of deposition were marine or terrestrial.

The deposition or nondeposition of the Glenarm series over the area of the Mine Ridge upland is a matter for speculation, because no Glenarm rocks occur on the flanks of the uplift between the ancient pre-Cambrian complex and the overlying Paleozoic beds. It is possible that the full thickness of Glenarm sediments may have once covered the area of pre-Cambrian rocks exposed in the Mine Ridge-Honeybrook upland. These beds may have been completely removed before the basal Cambrian strata were deposited. This assumption receives some justification from the fact that the pre-

Paleozoic plutonic rocks in the Mine Ridge upland, which have been tentatively correlated with the post-Glenarm igneous intrusions in Maryland, presuppose a sedimentary cover of considerable thickness. The necessary cover might have been furnished by Baltimore gneiss that was originally much thicker in the area of the Mine Ridge-Honeybrook upland than over the Avondale-Kennett uplift, farther south, where Glenarm sediments overlie the Baltimore gneiss. If Glenarm sediments were never present north of Chester Valley the stratigraphic horizon of the Baltimore gneiss that is overlain by basal Cambrian deposits would have been determined by erosion that continued throughout Glenarm time, whereas the Baltimore gneiss horizon that is overlain by Glenarm schists on the flanks of the Avondale-Kennett uplift, although approximately the same horizon as the floor of the Paleozoic strata in Mine Ridge, was obviously determined by pre-Glenarm erosion. It is an improbable coincidence that pre-Glenarm erosion and erosion operating over the whole period of Glenarm time should have arrived at the same place in the work of denudation, and it seems more probable that Glenarm sediments were laid down in both localities and later removed by pre-Paleozoic erosion.

The disappearance of Glenarm sediments north of Chester Valley is probably due to the thinness of sedimentation there as compared with the area of the Avondale-Kennett uplift, which originally lay at least 20 miles farther south of Chester Valley, its present position being due to northward overthrusting.

The middle beds of the Glenarm series in the western Piedmont province of Maryland include several volcanic flows, which have been interpreted by Jonas [63] as eruptions probably derived from volcanic vents farther west and northwest.

The deposition of the Glenarm series was succeeded by intrusion of basic rocks ranging from gabbro to pyroxenite and peridotite. Probably after the intrusion of the basic magma there was a second period of granitic invasion, as shown by rocks that range in composition from quartz monzonite and granodiorite to hornblende-quartz diorite. These light-colored plutonic rocks occur in large bodies in the Glenarm and pre-Glenarm strata of Maryland and southeastern Pennsylvania. Their relations to the basic intrusives are nowhere very clear, but at several localities west of Susquehanna River granodiorite correlated with the Port Deposit gneiss is intruded into hornblende schists and metagabbros that are presumably part of the larger gabbro masses with which they are associated. In several

[63] Jonas, A. I., Pre-Cambrian rocks of the western Piedmont of Maryland: Geol. Soc. America Bull., vol. 35, pp. 355–364, 1924.

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localities both granodiorite and gabbro are cut by a younger granite and pegmatite.

It is difficult also to interpret the history of igneous intrusion in Pennsylvania north of Chester Valley, because, as in Maryland, there is no definite evidence of the relative age of the gabbro and the rocks of monzonitic or dioritic composition; but the igneous sequence in the McCalls Ferry-Quarryville district appears to have been the same as in Maryland. The gabbro petrographically resembles gabbro of southeastern Maryland, and the monzonitic rocks are apparently intruded into the gabbro.

On the other hand, Bascom[64] describes gabbro dikes in Cecil County, Md., that cut the quartz monzonites represented by the Port Deposit gneiss. Likewise she describes the granodiorite in eastern Pennsylvania and in Delaware [65] as cut by the gabbro. It is probable that more detailed petrographic work will show the existence of either two periods of gabbroic intrusion, as has been suggested for the igneous complex of the Reading and Durham Hills, or else two periods of monzonite intrusion, one older and one younger than the gabbro. The granite of the Mine Ridge anticline, which may prove to be Paleozoic, is probably the same as the granite and pegmatite that cut the quartz monzonite in Maryland and the gabbro southwest of Wilmington, Del.[66]

The igneous intrusions may have produced regional contact metamorphism of the Glenarm sediments, which are profoundly altered throughout the eastern Piedmont area, in which igneous rocks are abundant. They were probably again accompanied by folding, as Jonas has found evidence in the western Piedmont belt of Maryland of a discordance in structure between the Glenarm rocks and the overlying basal Cambrian series.

To summarize, during early pre-Cambrian time there was deposition of sediments with possible intrusion of gabbroic sills or outpouring of basaltic flows. The basement complex was later invaded by a granite batholith, and the invasion was accompanied by folding and strong metamorphism. A long period of erosion followed, after which a series of later pre-Cambrian sediments was laid down and basic and acidic lava flows were extruded. A second period of igneous invasion and probable folding, followed by uplift and prolonged erosion, closed the pre-Cambrian history of the region.

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[64] Bascom, F., The crystalline rocks of Cecil County, Md.: Maryland Geol. Survey, Cecil County, p. 112, 1902.

[65] Bascom, F., U. S. Geol. Survey Geol. Atlas, Philadelphia folio (No. 162), 1909; Elkton-Wilmington folio (No. 211), 1920.

[66] Hawkins, A. C., Alternative interpretation of some crystalline schists in southeastern Pennsylvania: Am. Jour. Sci., 5th ser., vol. 7, p. 360, 1924.

PALEOZOIC HISTORY

GENERAL OUTLINE

The fact that the conglomerates occurring in the basal Cambrian beds on the flanks of the Mine Ridge upland contain only quartz pebbles in a matrix of silty origin suggests that the pebbles were derived from erosion of a low-lying land with an interior elevated tract in which the streams had a gradient sufficient to carry down fairly coarse detritus. During its long journey to the sea this detritus would lose most of the readily disintegrated material, retaining only the residual quartz pebbles, which were deposited at intervals from flooded streams, in silt derived from the low-lying shoreward plains. The fact that there are more than 2,200 feet of arenaceous Lower Cambrian deposits indicates prolonged erosion of a surface of moderate relief. The succeeding 1,800 feet of Lower Cambrian sediments are dominantly calcareous or dolomitic with subordinate layers of argillaceous material. As pointed out by Stose and Jonas[67] the axis of the Mine Ridge uplift was rising during Lower Cambrian time, and the Lower Cambrian section was probably thinner over the crest of the arch.

Calcitic and dolomitic sedimentation prevailed in the areas of Lancaster Valley and Chester Valley through Middle and Upper Cambrian time, although deposition was probably not continuous throughout this whole time. After the deposition of the Ordovician Beekmantown limestone slowly accumulated crustal stress found relief in a moderate flexure along the axis of the Mine Ridge anticline. This arching brought the strata above sea level long enough to allow erosion of the gently bowed sediments, so that when the Chazy sea transgressed over the eroded fold the Conestoga limestone was laid down unconformably over Upper, Middle, and Lower Cambrian strata down to the horizon of the Harpers schist.

After the deposition of the Conestoga limestone crustal stress, which had reaccumulated along the eastern border of the sinking Appalachian geosyncline, manifested itself in a moderately steep folding that followed the general trend line of the pre-Conestoga flexure. Subsequently a block of the earth's crust that originally lay some distance southeast of the Mine Ridge fold was driven northward across the folded edges of the Conestoga and pre-Conestoga strata. It is probable that this overthrust block extended at that time over the area occupied by the Mine Ridge and Honeybrook uplands.

The occurrence of folding subsequent to the overthrust is shown by the fact that the overriding fault block southwest of the Mine Ridge

[67] Stose, G. W., and Jonas, A. I., Ordovician overlap in the Piedmont province of Pennsylvania and Maryland: Geol. Soc. America Bull., vol. 33, p. 521, 1923.

anticline is folded in the double-crested arch of the Tucquan anticline. This anticline is a structural continuation of the double-crested Mine Ridge anticline, which must accordingly be later than the overthrust. The southwestward pitch of the Mine Ridge anticlinal axis was thus probably acquired or at least strongly intensified during the folding that succeeded the overthrust movement.

The only post-Conestoga and pre-Triassic events that are recorded in the McCalls Ferry-Quarryville district are, first, a post-Conestoga folding; later, extensive overthrusting; and finally, a folding of the overriding and underlying blocks. After the deposition of the Conestoga limestone there was a profound metamorphism of the Paleozoic rocks around the Mine Ridge anticline and westward to Susquehanna River at Shenks Ferry.

AGE OF THE POST-CONESTOGA DEFORMATION AND METAMORPHISM

There are no sedimentary rocks in the McCalls Ferry-Quarryville district that are younger than the Conestoga limestone, and the geologic evidence of the district shows only that the post-Conestoga deformation was pre-Triassic, because the Triassic strata north of the district are comparatively unaffected by folding. Therefore any discussion as to whether this deformation took place in one or more than one period of folding and as to the probable age of such deformation is based solely upon inferences drawn from the geologic history of adjoining regions and upon inferences as to the probable method of development of the observed structure.

Inferences drawn from the post-Chazy stratigraphy of adjoining regions.—The Martinsburg shale, on the northwestern border of the Lebanon and Cumberland Valleys in Pennsylvania, probably represents a depth of half a mile of sediments deposited during Trenton, Eden, and Maysville time. Such a thickness not more than 40 miles northwest of the McCalls Ferry-Quarryville district means that by the end of Martinsburg time there was probably a light load of sediments upon the Conestoga limestone in this district. At Lehigh Gap, in the Blue Mountain-Kittatinny range of the Appalachian Valley ridges, there is an angular unconformity between the Martinsburg and the overlying beds. Such an unconformity would represent the continuation into Pennsylvania of the crustal movement recorded at Otisville, N. Y., where Shawangunk conglomerate rests upon the tilted edges of shale of Martinsburg age with a difference in dip of 15°. The angular differences between the underlying and overlying beds at Lehigh Gap is also only about 13°, and therefore the crustal disturbance recorded in these two localities, although distinct, may be slight. A pronounced stratigraphic break at the base of the Richmond has long been recognized by Ulrich,[68] but the amount and extent of the folding involving pre-Richmond beds is still a matter of considerable uncertainty. The continental origin of the lower Silurian beds that occur in east-central Pennsylvania suggests land conditions in the McCalls Ferry-Quarryville district until the early part of the Niagaran transgression, which may have extended southwestward over a part of the Piedmont province. However, if Silurian sediments ever were deposited so far south they must have been removed before upper Silurian time, because in northern New Jersey there are no strata of Niagaran age beneath the Cayugan shales in Green Pond Mountain. The folding in the Green Pond syncline involves rocks laid down as part of the Upper Devonian gravel plain and is, therefore, post-Devonian. Barrell[69] has estimated a possible thickness of 2,500 to 3,500 feet of sediments embraced in upper Silurian and Lower and Middle Devonian time in the same area. He estimated the rate of southward and eastward thinning of Upper Devonian sediments at 100 feet to the mile, which would give not more than 1,200 feet of Upper Devonian sediments as a possible cover upon a piedmont gravel plain in the McCalls Férry-Quarryville district, which is 60 miles southeast of the Appalachian Valley ridges, where Upper Devonian sediments attain a thickness of 7,200 feet. If the full thickness of upper Silurian and Lower and Middle Devonian sediments in northern New Jersey continued southeastward as far as the McCalls Ferry-Quarryville district the probable maximum load upon the Conestoga limestone at the time of post-Devonian folding would have been of the order of magnitude of 5,000 feet. The actual thickness of sediments may, of course, have been much less than this figure, but it was probably not much greater.

There is no evidence in the eastern part of Pennsylvania that would show the nature of the uplift that occurred along the eastern border of the area of the Piedmont province in Devonian time. That such an uplift occurred is shown by the character of the Upper Devonian delta deposits. The location of the highland, from which the clastic deposits were derived, was somewhere in the area now covered by the Coastal Plain, as shown by the direction and rate of southeastward thinning of the delta deposits. In the Maritime Provinces of Canada and in the New England States, where Devonian rocks remain to record the history of the post-Middle Devonian movement, elevation was accompanied by folding and granitic intrusion. The amount of folding that accompanied the Devonian uplift

[68] Ulrich, E. O., The Ordovician-Silurian boundary: Cong. géol. internat., 12th sess., Compt. rend., pp. 614–615, 1913.

[69] Barrell, Joseph, Upper Devonian delta of Pennsylvania and New York: Am. Jour. Sci., 4th ser., vol. 36, pp. 234–239, 1914.

in eastern Pennsylvania, where Devonian sediments have been removed, can only be inferred.

During the Mississippian, Pennsylvanian, and early Permian time the area of the Piedmont province in Pennsylvania and Maryland was being eroded. In Virginia, along the western front of the Blue Ridge, Carboniferous rocks are folded and covered by an overthrust block of pre-Cambrian and Paleozoic rocks that has ridden westward along a nearly horizontal plane. This shows that folding at or near the end of the Paleozoic era was of the type that is accompanied by overthrusting. It does not prove that the Martic overthrust fault was formed during the same deformation as the overthrust west of the Blue Ridge, although it is a reasonable inference that they may be contemporaneous.

It was pointed out by Ulrich [70] that the folding of the Appalachian geosyncline took place in more than one period of orogeny and that the relative intensity of the folding in these different periods is hard to estimate. The Mine Ridge fold shows the existence of a periodically rising anticlinal axis that was initiated in early Paleozoic time or perhaps in pre-Cambrian time. Ulrich's suggestion of a westward migration of the belt of active folding leaves an open question how much earlier Paleozoic folds may have been intensified by later deformation. The evidence of two periods of metamorphism in post-Conestoga time suggests that the McCalls Ferry-Quarryville district and the immediately adjoining regions have undergone at least two compressive deformations, of which the second was the less intense, the inference being that the earlier folds remained more or less inert and resistant to the later folding or else that the later folding in this region was subordinate in intensity to the earlier.

Inferences drawn from the probable method of development of observed structure.—The absence of dislocation phenomena in the few localities where the thrust plane has been seen and the diminution in metamorphic intensity in the overriding block as compared with the metamorphic condition of the overridden strata in the McCalls Ferry-Quarryville district suggests that the thrusting took place under so-called "deep-seated" conditions. Such conditions are furnished not only by deep burial but also by batholithic intrusion, which raises the temperature gradient over wide areas. It has been shown that the evidence of adjoining regions does not indicate much more than a mile of sedimentary cover under which the rocks of the McCalls Ferry-Quarryville district have been folded. Such a load is comparable to the cover under which the great overthrusts of the western United States have taken place. In the Lewis overthrust, along the eastern flank of the Rocky Mountains, the older rocks of

[70] Ulrich, E. O., Revision of the Paleozoic systems: Geol. Soc. America Bull., vol. 22, pp. 436–442, 1911.

the mountain ranges have been shoved eastward over younger rocks, and in the eastward overthrust fault in the Philipsburg quadrangle, Montana, which is considered by Calkins to be the probable continuation of the Lewis overthrust, both underlying and overlying blocks were folded in rather close folds.[71] The attitude of the underlying beds is in general fairly simple, and the hypothesis is advanced that the rocks were not much folded before the overthrust, which followed a plane of slight resistance approximately parallel to the bedding. Such a movement as this overthrust in the Rocky Mountains is similar to the Martic overthrust in its wide extent, in the general parallelism between bedding and overthrust plane, and in the warping of the overthrust plane, which is, however, much more pronounced in the Martic overthrust. But a striking dissimilarity between the two faults is the present condition of the rocks involved. In the overthrust in the Rocky Mountains the rocks of the overriding block are ripple-marked sandstones, shales, and impure limestone that show only a very slight metamorphism, whereas in the Martic overthrust they are schists, phyllites, and slates. There is a well-marked brecciation along the base of the overthrust block in the Rocky Mountains, but so far as indicated in available exposures brecciation is conspicuously absent in the Martic overthrust. These differences indicate an essential difference in conditions of faulting, the overthrust in the Rocky Mountains having apparently taken place under relatively light load where mechanical deformation was dominant. The overthrusts of southern Nevada described by Longwell[72] have taken place under similar conditions, favorable to strong mechanical deformation, and yet the thickness of load in the Spring Mountain region at the time of overthrusting is estimated to be at least 10,000 feet. It is obvious that the Martic overthrust, at least in the McCalls Ferry-Quarryville district, took place under so high a temperature that mechanical deformation was at a minimum. Such a temperature could not be attained at depths of little more than a mile on a normal geothermal gradient, but it might have been furnished by the influence of a rising batholith that has never reached the surface but has produced the extensive pneumatolysis and pegmatitization of the Paleozoic rocks in the region.

SUMMARY OF PALEOZOIC GEOLOGIC HISTORY

During the early part of the Paleozoic era the McCalls Ferry-Quarryville district was accumulating first arenaceous, later calcitic

[71] Emmons, W. H., and Calkins, F. C., Geology and ore deposits of the Philipsburg quadrangle, Mont.: U. S. Geol. Survey Prof. Paper 78, pp. 146–149, 1913.

[72] Longwell, C. R., Geology of the Muddy Mountains, Nev.: Am. Jour. Sci., 4th ser., vol. 50, pp. 39–62, 1921; Thrust faults and flaws in southern Nevada (abstract): Geol. Soc. America Bull., vol. 35, pp. 64–65, 1924; Complex structure in the Spring Mountains, Nev. (abstract): idem, p. 150, 1924.

and dolomitic sediments. Even in Cambrian time, however, a slight unrest in the basin of deposition is manifested by the several probable emergences that correspond to breaks in sedimentation in adjoining regions. Such unrest apparently increased, so that before Chazy time the Paleozoic and underlying rocks of the district were folded into a gentle arch over the site of the Mine Ridge anticline. The Conestoga limestone was laid down over the eroded surface of this arch probably during Chazy time. The district was then folded in a period of deformation whose exact date can not be determined. Subsequently a block of the earth's crust that originally lay between 10 and 20 miles to the south was driven northward and westward, truncating the folded edges of Cambrian and Ordovician strata. A moderate folding was superimposed upon both overthrust and underlying rocks. The last record of compressive force in the district is the development of a schistosity that cuts across the older folding and foliation. The date of the folding subsequent to the overthrust can not be placed more definitely than pre-Triassic and post-Chazy.

POST-PALEOZOIC STRATIGRAPHIC RECORD

The arkosic conglomerates on the eastern edge of the Triassic basin in Pennsylvania and Maryland indicate that the McCalls Ferry-Quarryville district was a highland during Triassic time. Westward-flowing streams carried the débris from this highland into the basin of deposition. All during Triassic time dikes and sills of diabase forced their way upward into the sedimentary beds. These intrusions extended beyond the belt of sedimentation into the eastern highland and are recorded in the narrow dikes that cross the McCalls Ferry-Quarryville districts. The only other record of Triassic history in this area is the extensive normal faulting that is characteristic of Triassic deformation. This faulting broke the northern half of the district into several well-defined blocks and doubtless extended farther south than it is possible to trace the faults in the pre-Cambrian mica schists.

There is no record of Jurassic and early Cretaceous history in this district, but the presence of Upper Cretaceous marine deposits in the area southeast of the district that was a part of the old Appalachian highland shows that at some time during Jurassic or early Cretaceous time there was an upwarping of the Triassic basin together with a foundering of the eastern part of the old Appalachian land mass, which initiated the present eastward-flowing drainage of the Atlantic slope and the present relations of land and sea. The subsequent history of the region must be read in the physiographic record, which furnishes at least a partial epitome of erosion subsequent to the establishment of the eastward-flowing drainage.

PHYSIOGRAPHIC RECORD

Detailed topographic studies of the Piedmont province show that there exist in the highlands of the region a number of terracelike surfaces, and that these surfaces bevel the upturned underlying rocks. Moreover, in the lowlands the plane indicated by flat-topped hill summits of accordant altitude also bevels the underlying structure. Thus the hilltops are obviously remnants of old surfaces of low relief into which the present streams have incised their meandering course. The large number of residual erosion surfaces marking cycles and subcycles of erosion thus suggested can not be fitted into the idea that the topography of this region has been developed in three erosion cycles, as has been advocated by previous workers; nor can they be explained by a widespread warping of two or three uplifted peneplains, such as was invoked by Campbell in order to explain the occurrence of his uplifted Harrisburg peneplain at different altitudes in different places. The succession of numerous flat surfaces, one above another, cutting across the underlying rock structure of the highlands, and the appearance of the same level surfaces on both sides of the highland areas make the warping hypothesis inadequate to reduce the number of erosion cycles, indicated by the successive surfaces, to the three cycles that have been previously recognized. It is evident that the region between the Appalachian Valley ridges and the Coastal Plain furnishes evidence of numerous interrupted erosion cycles of relatively short duration, and level interfluves at various altitudes are recognized as remnants of old peneplains cut during these different cycles. But the proper correlation and the sequence of these remnants of peneplains or partial peneplains present a tantalizing problem, as the altitude of residual erosion surfaces depends so largely on the relative resistance of different rocks and on the distance of each part of the surface from base level as well as from the trunk stream of the drainage basin.

The fact that the master streams of the Atlantic slope show a lack of accordance with the structure of the underlying rock, combined with the extensive stream adjustment of the tributaries, suggests the possibility of tracing backward, step by step, the drainage history of the region, thereby determining the sequence of past cycles whose record is preserved in present and preexistent drainage channels and establishing the connection between such records of old drainage and the so-called peneplain remnants preserved in the level surfaces of the region.

EVIDENCE OF FORMER DRAINAGE COURSES

In a region where the drainage has become thoroughly adjusted the present streams will retain little evidence of their former courses.

However, in spite of extensive stream adjustment, successive rejuvenations may be recorded in the drainage of a region in one or more of four ways—(1) superimposition, (2) modification of normal cross profiles of stream valleys, (3) wind gaps, (4) "knickpoints," or points of abrupt change in the longitudinal profile of the stream valleys from a gentle concave curve to a curve that is convex upward.

Superimposition.[73]—Superimposition as the term was employed by Powell[74] in 1875 was applied to streams that acquired their courses on marine sediments and that have retained their courses after they have cut down to the underlying surface on which the sediments were deposited. In Powell's definition the idea of a marine cover through which streams cut their way into a discordant underlying terrane is an essential feature of superimposition, and this usage has been widely followed. However, even as far back as 1877 Gilbert[75] recognized and defined three classes of superimposition—namely, superimposition by sedimentation, or subaqueous deposition; superimposition by alluviation, or subaerial deposition; and superimposition by planation. His additions to the analysis of superimposition were twofold. (1) Subaerial erosion as well as marine erosion might produce conditions under which streams would be superimposed. He cites, as processes by which these conditions might be produced, alluviation and lateral planation, both associated with the mature or postmature degradation of a land surface. (2) Superimposition does not necessarily imply the previous existence of covering strata but may take place as a result of planation rather than of deposition.

McGee[76] carried the analysis a step farther when he defined superimposition "through base-level degradation so perfect that the old waterways were completely obliterated." It was thus gradually becoming apparent that the essential feature of a superimposed drainage pattern is the fact that the streams have been let down from above upon a rock surface, into which they are now incised without regard to the underlying rock structure. The attempt to explain many such superimposed drainage patterns as an inheritance from a cover of marine sediments has led to some unwarranted assumptions of the former existence of marine deposits in the east-

[73] The term "superimposition" was apparently first used by George Maw in 1866 (Notes on the comparative structure of surfaces produced by subaerial and marine denudation: Geol. Mag., vol. 3, pp. 441–445, 1866).

[74] Powell, J. W., Exploration of the Colorado River of the West and its tributaries, p. 166, Washington, 1875.

[75] Gilbert, G. K., Report on the geology of the Henry Mountains, 1st ed., p. 144, U. S. Geog. and Geol. Survey Rocky Mtn. Region, 1877.

[76] McGee, W J, Geology of the head of Chesapeake Bay: U. S. Geol. Survey Seventh Ann. Rept., pp. 562–563, 1888.

ern Appalachian Highlands in areas where there is no evidence of such deposition. Recognition of the fact that streams may be let down not only from a sedimentary cover of either marine or alluvial origin but also from an almost completely degraded surface (a peneplain) thus clears away one fruitful source of unjustifiable speculations in interpreting physiographic history.

In nonglaciated regions the essential feature of superimposition is that the unadjusted courses of major streams across ridges carved out by tributary streams indicate a reduction of the region to a low-lying level surface before the formation of the flat-topped ridges. If the original level surface was of marine origin and covered by sediments the drainage is said to be superimposed by sedimentation on the underlying hard rock. If the level surface on which the drainage has established its unadjusted course was formed by subaerial erosion the lack of accord between the major drainage channels and the underlying rock shown by the passage of the streams through resistant ridges is due to the fact that, consequent upon uplift of a peneplain, the streams that had been wandering upon the peneplain were intrenched and may be said to be superimposed by peneplanation. The term superimposed is thus used to denote the fact that the major streams which are flowing in courses that are out of accord with the underlying structure have originated not as initial consequent streams but as resuscitations of a previous drainage system whose courses were established on an approximately level surface produced by either marine or subaerial denudation. Whether the original level surface was of marine origin and once covered by sediments (superimposition by sedimentation), or whether it was of subaerial origin (superimposition by peneplanation), does not affect the value of the evidence of degradation that is afforded by the superimposition. Streams once firmly intrenched in a superimposed course will cut down across the hard ridges in valleys that will persist as long as the hard rock stands up above the level of the surrounding country. Thus are produced the water gaps that are so conspicuous a feature of the superimposed streams of the eastern Atlantic Slope, such as Delaware, Lehigh, and Susquehanna Rivers. A superimposed stream will record the surface of degradation on which it originally flowed as long as the flat top of the ridge remains unreduced, a period which may cover many subsequent partial cycles of erosion. In the relatively soft lowlands intervening between the hard ridges, such as the Lebanon, Lancaster, and Cumberland Valleys, these subsequent partial cycles of erosion have removed the rock so rapidly that the intrenched character of the master streams is not so apparent as in their courses athwart the resistant ridges. Therefore, in the stream gaps should be sought the evidence of superimposition that dates back to the time of formation of the highest flat-topped surfaces on both sides of the gap. The only clue to the erosion history of the region subsequent to the intrenchment of the old stream will be found in the shape of the cross section of the gap and in the longitudinal profiles of existing streams.

Cross profiles of stream valleys.—The cross profiles of rejuvenated stream valleys may show the outlines of wide older valleys in the bottom of which the present streams are intrenched. Composite valley slopes that indicate by their shape one or two rejuvenations in the course of normal erosion have been aptly termed by Matthes[77] "2-story valleys" or "3-story valleys." But, although the evidence of the oldest rejuvenations may be preserved in the composite upper slopes of the valley, the presence of a uniformly steep inner gorge does not indicate that the gorge has been formed during an uninterrupted erosion cycle. On the contrary, the downcutting of the lower gorge may continue through several partial cycles of erosion. For even though the progress of erosion in the region is quickened by an uplift or by successive uplifts, there can be no record of such quickenings in a gorge through hard rock, if lateral planation has not yet begun to take effect. In the vicinity of a hard ridge softer rocks may have been reduced to a local peneplain, not once but repeatedly, while a gorge in the hard rocks was simply being deepened.

Wind gaps.—The volume of water and the débris carried in the main stream of a superimposed drainage system is much greater than in the tributaries, therefore downcutting progresses most rapidly in the master valley, and the minor streams are one by one diverted from their superimposed courses by the subsequent tributaries of the master stream. The relic valleys of the original transverse minor streams are preserved as wind gaps in the hard ridges. Barrell[78] was the first to call particular attention to the value of wind gaps as indicators of former cycles of fluvial erosion. Briefly stated, the value of wind gaps in fixing points upon partly obliterated drainage systems lies in the fact that their present outline has been comparatively little modified since the time when water was flowing through them, because they are hung up, so to speak, out of reach of subsequent erosion.

The stage of erosion at which capture took place may be inferred from the form of the cross section of the wind gap and from the altitude of the base of the wind gap compared with the altitudes of contemporaneous stream channels recorded elsewhere in the region in the upper slopes of composite valleys. The subaerial erosion cycle

[77] Matthes, F. E., U. S. Geol. Survey Yosemite Valley map, 1922.

[78] Barrell, Joseph, The Piedmont terraces of the northern Appalachians: Am. Jour. Sci., 4th ser., vol. 49, p. 388, 1920.

during which any one group of wind gaps was cut can be fixed only by the correlation of the wind gaps with composite slopes in the adjoining stream valleys and with remnants of erosion surfaces whose present altitude shows an evident relation to the base of the wind gaps.

Longitudinal profiles of streams.—The value of the three evidences of past drainage that have been discussed above lies in the fact that in resistant rocks all three kinds of evidence will be preserved as long as the ridges stand above the surrounding lowland. They will thus persist long after all other indications of the original drainage course have been destroyed by extensive stream adjustment and by repeated planations of the surrounding country. In addition there remains, in the existent drainage courses of a region, still a fourth evidence of recent breaks in the continuity of erosion—namely, the knickpoints in the longitudinal stream profiles.

When a region is uplifted a stream whose channel had attained the normal concave stream profile [79] will be quickened and will begin to cut down its bed. In the lower reaches of the stream, where the longitudinal profile that has become nearly flat has been steepened by the uplift, there will be a development of a new concave curve of maturity, and inasmuch as erosion works progressively backward the reduction of the former stream profile to a new curve will progress backward from mouth to source. Obviously at the headward limit of this reduction to a new grade there will be an arch in the curve—a change in the stream profile from a concave to a convex curve. Therefore a stream that has become intrenched in a peneplain and has maintained its course through several partial erosion cycles can show, from mouth to source, evidence of successively older periods of uplift and consequent valley cutting by a succession of bends or knickpoints in the curve.

RELATIVE PRESERVATION OF SUCCESSIVE DRAINAGE COURSES

In a region that has been repeatedly uplifted we find near the seacoast widespread and clear records of the work of recent erosion. The dissection of successively emerged continental shelves by extended streams and the rejuvenation of drainage shown in the stream profiles of both recent and older stream courses, combined with the presence of residual deposits, make it comparatively easy to fix the duration and extent of each successive advance and withdrawal of the sea. But the amount of erosion accomplished near the seacoast, by either subaerial or marine agencies, has been a maximum in every cycle; therefore near the present shore line the clearness of the record of recent erosion is combined with an obliteration of the

[79] Penck, Albrecht, Morphologie der Erdoberfläche, vol. 1, pp. 319–341, Stuttgart, 1894.

record of earlier erosion cycles by the formation of new seaward peneplains and marine plains in the later cycles. The record of earlier erosion must be sought farther inland, near the present headwaters of the streams, where isolated residuals of earlier plane surfaces have been preserved from dissection and where hard ridges that have not preserved their original surfaces during the peneplanation of adjacent soft rocks retain, in a more or less ephemeral way, the records of earlier erosion in the form of wind and water gaps.

Therefore in order to trace the effects of successive rejuvenation it is necessary to study the course of a major drainage system from the mouth to a point so remote from present base-level that the record of the old drainage channels is still preserved in the cross profiles of the valleys. Such an investigation comprises a study of interstream areas that are remnants of former peneplains or of incompletely developed peneplains and a study of progressive stream adjustment and of the modifications of stream gradients and valley profiles during successive cycles of erosion. The method of investigation has been described at length in a previous paper,[80] and the results here presented are those which apply more particularly to the McCalls Ferry-Quarryville district.

STUDY OF INTERSTREAM AREAS

The flat-topped interstream areas were first outlined upon good topographic maps of a scale of 1:62,500. The individual levels thus estalished were then studied in their extended development across a wide stretch of country by the construction of a projected profile adjoining the northwest side of a profile projected by Barrell along a section plane running N. 35½° W. from the mouth of Susquehanna River at Havre de Grace. The foreground of the profile thus projected extends from the inland edge of the Coastal Plain northwestward as far as the eastern ridges of the Appalachian Valley. The projected profile embraces the plane surfaces that are upheld in the Blue Mountain-Kittatinny Ridge and Lebanon Valley divisions of the Great Appalachian Valley, in the South Mountain and Reading and Durham Hills of the Appalachian Mountains, in the Lancaster Valley and the Triassic Lowland divisions of the Piedmont Province, and in the Piedmont Upland to its junction with the Coastal Plain. (See pl. 7, *A*.)

The complexity of levels shown in such an extended profile is at first sight bewildering and is increased by the fact that in passing eastward from the Susquehanna Valley certain apparently level surfaces show a rise toward the divide. For example, in the Lebanon Valley the 400-foot level on limestone hills near the Susquehanna

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[80] Knopf, E. B., Correlation of residual erosion surfaces in the eastern Appalachian Highlands: Geol. Soc. America Bull., vol. 35, pp. 633–666, 1924.

rises to about 500 feet 6 miles northwest of Wernersville. The reason for the disappearance of the 400-foot surface near the divide is that on existent divides remnants of surfaces cut during previous cycles are more or less immune to the destructive agencies of the present cycle. The profile of the divide between the Susquehanna and the Schuylkill drainage systems (pl. 7, *B*) shows three conspicuous highlands occupied by the Blue Mountain, South Mountain, and Welsh Mountain-Barren Hill ridges. The southernmost highland terminates in Mine Ridge upland at an altitude of about 860 feet. These highlands are separated by intervening lowlands about 20 miles in width. Although Blue Mountain lies 24 miles farther inland than South Mountain, the general altitude of the intervening lowland between Blue Mountain and South Mountain is more than 100 feet lower than that of the lowland that separates South Mountain and Welsh Mountain.

The most striking feature of the divide between Susquehanna and Schuylkill Rivers is the difference between the portion that extends seaward from the Mine Ridge upland and the inland extension northward from the Mine Ridge upland. South of the Mine Ridge upland the profile of the divide is strikingly similar to Barrell's projected profiles of the Piedmont terraces in Maryland.[81] The descent from the Mine Ridge upland to sea level comprises a succession of seven stepped or terrace-like surfaces that are cut across crystalline schists of approximately uniform resistance to erosion. The outline of these steps, as shown upon the topographic maps, is not regular in its linear extent, as in a flight of stairs, but each successive level wraps around the dissected edges of the next higher level, thereby extending headward up the courses of the major tributaries.

In New England it was such a stairlike arrangement of levels, facing seaward, that suggested to Barrell the idea that the rises are old sea cliffs. In the Pennsylvania terrace-like surfaces the marine origin of the treads is suggested by their gentle seaward slope, the lack of monadnocks, and the scarplike descent of each to the next lower level. The three lowermost terraces, each carrying a well-defined mantle of unconsolidated deposits, are of Pleistocene age and are generally conceded to be of marine origin.

In the absence of marine deposits and of undoubted sea cliffs the origin of the four upper terraces must still remain hypothetical, but whatever was the erosional process that produced the level terrace-like surfaces, the uniformity of resistance in the underlying rock leads to the conclusion that they are distinct surfaces of erosion, partly dissected, the width of each step being dependent upon the duration of the erosion that produced the plane surface.

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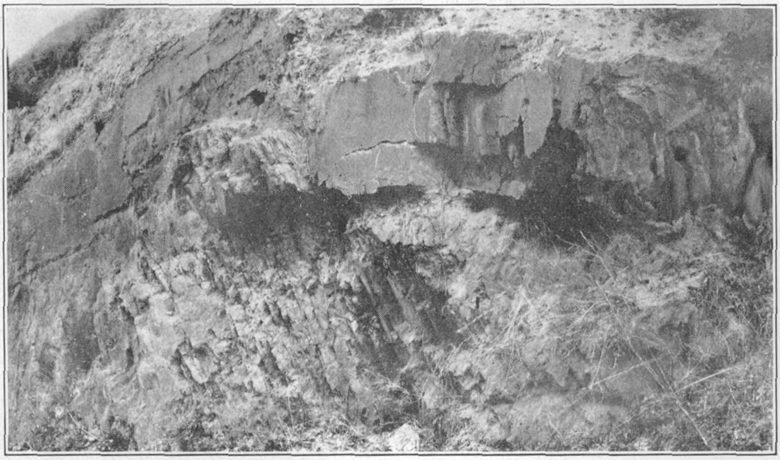
[81] Barrell, Joseph, The Piedmont terraces of the northern Appalachians: Am. Jour. Sci., 4th ser., vol. 49, pl. 6, 1920.

The dissected character of the divide north of the Mine Ridge upland presents a strong contrast to the terrace-like steps south of the ridge and is indicative of subaerial erosion. The divide is notched by depressions that fall into two classes. One class, to which belong only a few of the numerous depressions, comprises narrow, steep-sided transverse trenches, sunk into flat-topped uplands. Their shape clearly shows that they were carved by the erosion of former streams. The other class, which is represented by most of the depressions, embraces the wide, gently rounded col-like depressions that suggest at first a normal lowering of a divide crest by headward erosion of opposing streams. But these notches are not depressions in a sharp crest such as would be expected if the lowering were due to streams contesting a narrow divide. On the contrary, they extend with level floors for half a mile or more in a direction at right angles to the trend of the divide, and this trend suggests that they were originally water gaps, which have been abandoned by their streams in a late mature stage of erosion. If these depressions in the divide are really the result of stream erosion in previous erosion cycles the altitude of the base of the depressions should show a definite relation to that of remnants of erosion surfaces in other parts of the region, such as benches, shelves, or hill summits.

By a careful study of different localities it is possible to determine, for each, the relative position of the erosion surfaces. In the following table the base-levels and erosion surfaces recognized by Knopf in the region studied are compared with those previously determined by Barrell and Bascom.

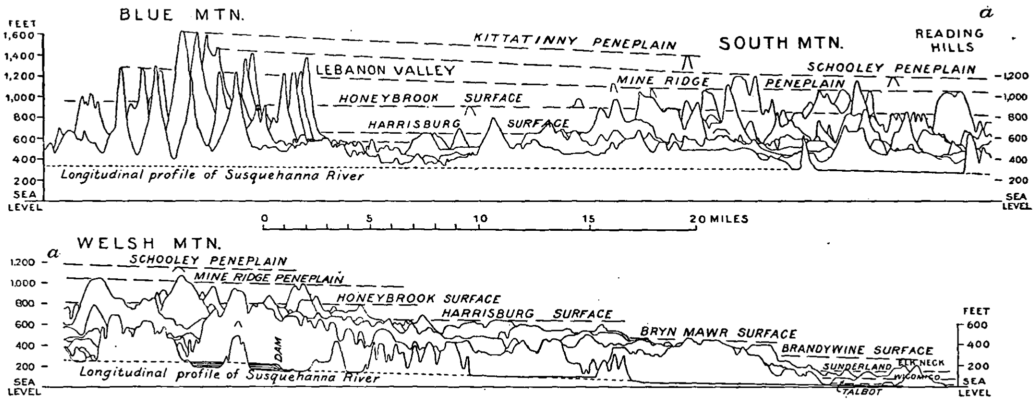


A. ANTICLINE IN ANTIETAM SCHIST, ROAD CUT 2 MILES WEST OF BELLEMONT

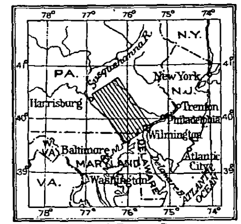
Shows gently dipping bedding and steeply dipping schistosity cutting it

B. CLOSELY FOLDED ANTIETAM SCHIST, ROAD CUT ON EAST SIDE OF OAK HILL

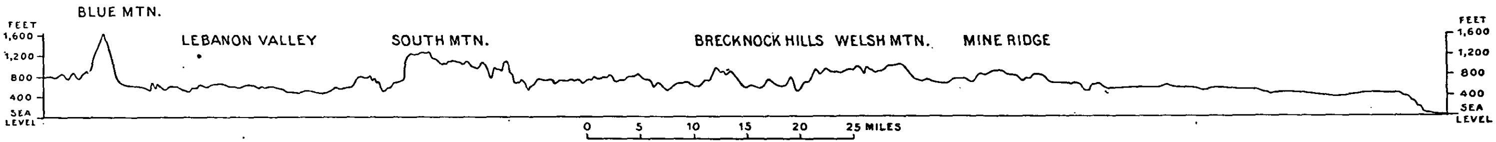
The folded bedding is outlined by quartz veins and cut at a steep angle by schistosity dipping northwest



A



INDEX MAP
Ruled area is region in which profiles were drawn



B

A. PROJECTED PROFILES OF AREA BETWEEN BLUE MOUNTAIN AND THE COASTAL PLAIN
B. PROFILE OF DIVIDE BETWEEN SUSQUEHANNA AND SCHUYLKILL RIVER DRAINAGE SLOPES

Altitude, in feet, of base-levels and erosion surfaces in Appalachian region

Fluvial base-levels in northern Appalachian Highlands (Joseph Barrell)		Piedmont province of Pennsylvania (F. Bascom)			Remnants of erosion surfaces established by present investigation (Eleanora B. Knopf)					
		Peneplains	West	East		Pocono Mountain	Blue Mountain and vicinity	South Mountain	Welsh Mountain and vicinity	Mine Ridge southward to sea level
Kittatinny.	1,600	Kittatinny.	1,800	1,600–1,100	Kittatinny.	1,800	1,660	1,340		
Schooley (?).	1,350	Schooley.	1,300	1,000–900	Schooley.		1,400–1,300	1,180–1,100	1,080	
					Mine Ridge.		1,100–1,000	1,000	940	880–820
(?).	950	Honeybrook.	860	800–700	Honeybrook.		900–840	840–800	800–720	720
Sunbury.	750	Harrisburg.	800	500	Harrisburg.		700–600		620	620
Harrisburg.	520	Early Brandywine.[a]	500	400–390	Bryn Mawr.		560–500			500–460
Somerville.	400	Terraces								
		Late Brandywine.[b]	400	300–200	Brandywine.		440–400		420	320
		Sunderland.	300	180–100	Sunderland.					200–100
		Wicomico.	90	45	Wicomico.					80–60
		Talbot.	45	40–0	Talbot.					40

[a] Now replaced by Bryn Mawr. Bascom, F., Resuscitation of the term Bryn Mawr gravel: U. S. Geol. Survey Prof. Paper 132, pp. 117-119, 1924.
[b] Now replaced by Brandywine. Idem.

Comparison shows that the results of the present study bear out to a large extent the conclusions of Barrell and Bascom and also suggest the existence of an additional erosion cycle, here called the Mine Ridge cycle because the surface then developed is well preserved on the summit of Mine Ridge.

STUDY OF DRAINAGE MODIFICATIONS

Mine Ridge cycle of erosion.—The highest level surface in the summit of Mine Ridge at altitudes ranging from 820 to 880 feet is upheld by steeply inclined quartzite beds. The hill summits on the two sides of the wind gap at Gap are distinctly lower than the southwest end of Mine Ridge, being only 740 to 760 feet above sea level, but the northeastward continuation of Mine Ridge in the area of the Coatesville and Honeybrook quadrangles rises to altitudes of 800, 840, and 880 feet. That the altitude of this ridge is not determined by the relatively high resistance to erosion of the underlying rock is shown by the facts that the quartzite ridge has been reduced to the above-mentioned wide depression marked by the 760-foot hill summits at Gap and that in this depression there is an old valley preserved in a wind gap, showing that a transverse superimposed stream once flowed across a surface of low relief, the remnants of which remain in the highest 820 to 880 foot summits of the quartzite ridge on the two sides of the gap. Moreover, the same general altitude as that in the highest summits of Mine Ridge reappears at Hickory School, 10 miles nearer to the trunk stream of the district, where the flat-topped ridge at about 900 feet is underlain by micaceous and chlorite schists whose resistance to erosion is much less than that of quartzite.

The identification of the Mine Ridge cycle and its distinction from the Schooley cycle rests on the existence of a well-defined level surface on South Mountain below the remnants of the Schooley peneplain. This lower surface suggests a dissection of the Schooley surface with the resultant formation of the present 1,000-foot level. A study of the profile of the Blue Mountain front of the Appalachian Valley shows the record of this erosion cycle in a group of wind gaps, which lie between 1,100 and 1,200 feet. These gaps are approximately 200 feet lower than the gaps recording the Schooley channels.

Altitude of depressions that are probably wind gaps recording the Mine Ridge erosion cycle, during which the Schooley peneplain was dissected

Locality	Name	Altitude (feet)	
		Base	Adjoining hill summits
Blue Mountain	Myers Gap	1,180	1,340
	Millers Gap	1,180	1,340
	Little Gap	1,100	1,440-1,500
South Mountain	Transverse depressions in divide crests	1,000-1,080	1,200

Present altitude of valleys cut during the Mine Ridge erosion cycle as shown in the restored cross profiles of composite valleys

Locality	Name	Altitude (feet)	
		Base	Adjoining hill summits
Blue Mountain	Manada Gap	1,200	1,340
	Indiantown Gap	1,200	1,300
	Swatara Gap	1,180	1,440

These gaps indent the Schooley surface preserved upon Blue Mountain at altitudes between 1,300 and 1,400 feet and trench the mature Schooley valleys. This post-Schooley valley cutting therefore coincides with the development of a peneplain or partial peneplain south of Blue Mountain whose remnants are lower than the residuals of Schooley erosion on Schooley Mountain. A wind-gap record of the Mine Ridge erosion cycle is seen in Little Gap, 5 miles northeast of Lehigh Water Gap. Little Gap, whose base lies at 1,100 feet, records three erosion cycles in its relation to the highest summits of Blue Mountain. The oldest cycle is recorded by the Kittatinny surface, whose remnants lie at 1,620 feet 3 miles northeast of Little Gap and 3 miles southwest of Lehigh Water Gap. The record of the succeeding Schooley cycle appears in the somewhat irregular surface between 1,400 and 1,500 feet and in the upper slopes of the composite cross profiles of the valleys of Little Gap and Lehigh Gap shown in Figure 6. The lower steep gorge of Little Gap records the down-cutting produced by quickened erosion in the Mine Ridge cycle.

Therefore, during the erosion cycle that succeeded the Schooley cycle Blue Mountain was dissected to a depth now recorded at 1,100 feet. Local plane surfaces were also cut in the vicinity of South Mountain that are recorded in plateau remnants at about 1,000 feet, and the region south of South Mountain was reduced to a partial peneplain whose remnants are so well preserved at 800 to 900 feet in Mine Ridge and in the flat-topped hills southwest of it that this erosion cycle is called the Mine Ridge cycle. During the Mine Ridge

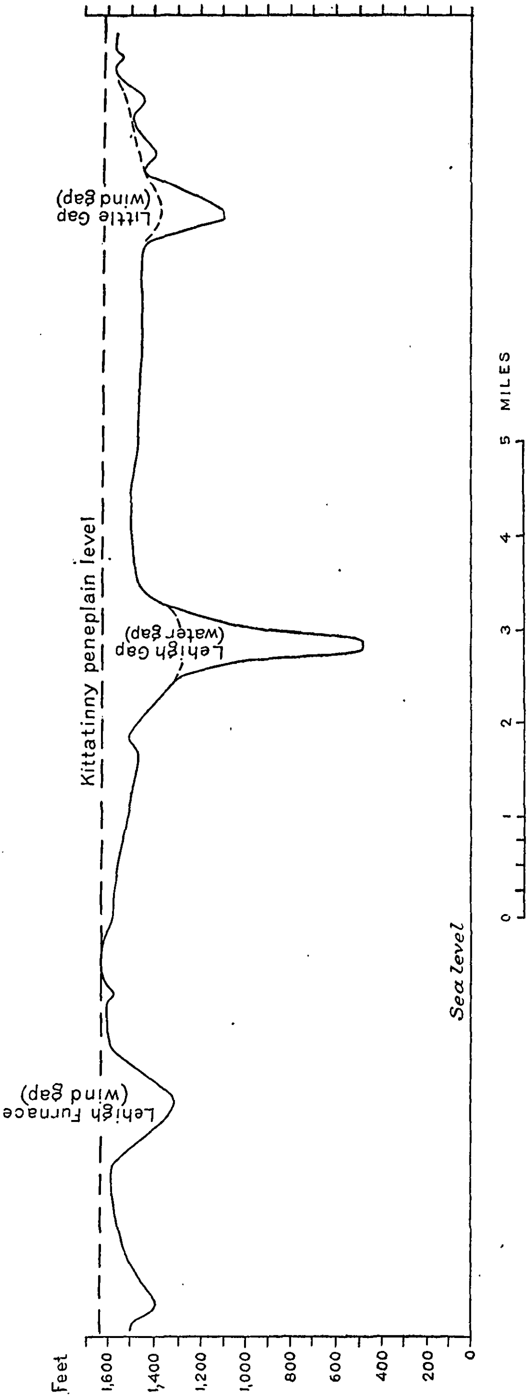


FIGURE 6.—Projected profile of Blue Mountain showing cross profiles of Lehigh Gap and Little Gap

cycle a smaller area was reduced than during the two previous cycles, and the highlands of the present topography began to appear. At the end of the Schooley cycle peneplanation had progressed so far inland that the whole area southeast of Blue Mountain was reduced to a peneplain, whereas at the end of the succeeding Mine Ridge cycle there remained unreduced remnants of the Schooley peneplain, which persist to the present day in the highlands of South Mountain, the Reading Hills, and Welsh Mountain.

With the exception of a few notches that are possible wind gaps in the South Mountain in Lebanon County, there is no evidence to prove the subaerial origin of the Mine Ridge partial peneplain south of Blue Mountain, or to indicate the extent to which it was developed eastward.

Honeybrook subcycle of erosion.—The rejuvenation that interrupted the continuity of erosion during the Mine Ridge cycle was recognized by Bascom[82] in the remnants of erosion surfaces now preserved at 720 feet near Honeybrook, Pa. As a result of this rejuvenation terraces were cut on the resistant slopes of Mine Ridge upland at 720 feet. South of Mine Ridge the seaward erosion of the Honeybrook cycle is preserved in the terrace-like surface at 720 feet. In Blue Mountain streams quickened by the Honeybrook uplift cut deep gorges, which were abandoned by their streams in consequence of the next rejuvenation and now stand as the well-known wind gaps represented by Pen Argyl Gap (980 feet), Culvers Gap (920 feet), and Sterrett Gap (920 feet).

Harrisburg subcycle of erosion.—The work of the next succeeding cycle has been fully outlined by Barrell[83] for the region between Sunbury and Harrisburg. He named the resultant surface the Sunbury peneplain because remnants are well preserved at about 850 feet in the flat hilltops near Sunbury, Pa.

The upland surface underlain by Martinsburg shale north and east of Harrisburg shows vestiges of two erosion surfaces. A local peneplanation is recorded in the general upland surface at about 540 feet, and remnants of an earlier planation rise above the 540-foot level in isolated monadnocks with an altitude of 660 feet. These upper residuals approximate the altitude of the Sunbury peneplain in the Harrisburg quadrangle described by Barrell. Campbell, who was the first to point out the operation in this region of more than two erosion cycles,[84] has since decided to restrict the name Harrisburg to the surface that is preserved at altitudes of 600 to 660 feet northeast of Harrisburg. Therefore the name Sunbury is abandoned in favor of the name Harrisburg, which has priority of usage. It is to be noted, however, that the name Harrisburg as here employed is the equivalent of the name Sunbury employed by Knopf in an earlier paper on the subject.[85]

The following table shows the altitude of several valleys, probably wind gaps, that were occupied by stream courses during Harrisburg erosion.

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[82] Bascom, F., U. S. Geol. Survey Geol. Atlas, Elkton-Wilmington folio (No. 211), p. 2, 1920.

[83] Barrell, Joseph, The Piedmont terraces of the northern Appalachians: Am. Jour. Sci., 4th ser., vol. 49, p. 336, 1920.

[84] Campbell, M. R., Geographic development of northern Pennsylvania and southern New York: Geol. Soc. America Bull., vol. 14, p. 284, 1903.

[85] Knopf, E. B., Correlation of residual erosion surfaces in the eastern Appalachian Highlands: Geol. Soc. America Bull., vol. 35, pp. 653–658, 1924.

Altitude of probable wind gaps recording the Harrisburg erosion during which the Mine Ridge peneplain was dissected

Location	Name	Altitude of base (feet)
New Jersey Highlands	Stewart Gap	740
	Gap north of Port Murray	740
North of Blue Mountain and west of Susquehanna Gap	Rattlesnake Ridge	750
Mine Ridge	Gap	580

Paleophysiographic maps of the region near the end of Honeybrook and Harrisburg erosion (figs. 7 and 8) show that the outlines of the present highland areas south of Lebanon Valley were clearly delimited and persistent, thus indicating that the normal progress of erosion was interrupted before either cycle had attained even approximate completion. The upland of Mine Ridge was initiated in the Honeybrook cycle. The shore line at this time probably lay southeast of the Mine Ridge upland, for old drainage courses recorded in the highlands north of the Mine Ridge upland suggest a persistence of fluvial erosion in that part of the region throughout Honeybrook and Harrisburg time. Gravel has been found by the writers at 620 feet on the Mine Ridge upland about 3 miles northwest of Coatesville.

Bryn Mawr subcycle of erosion.—In the region of Harrisburg below the remnants of the Harrisburg peneplain there are also remnants of a lower level surface preserved on the Martinsburg shales at altitudes of 500 to 560 feet. Barrell[86] ascribed to the remnants of this erosion surface traced northward from Harrisburg a gradient of 2 feet to the mile, and this slope projected southward would place residual surfaces of the same erosion cycle near Chesapeake Bay at an altitude of 440 feet, an estimate which accords remarkably well with the 460-foot surface at Woodlawn, northeast of Perryville, Md. North of Woodlawn and south of Chester Valley, in the McCalls Ferry-Quarryville district, the same surface is well developed at 500 to 460 feet. At Woodlawn this plane surface bevels the underlying Cretaceous formations and gravel formerly mapped as "Lafayette,"[87] now known as Bryn Mawr gravel.[88] As it is probable that the gravel associated with the 460-foot surface at Woodlawn is a marine deposit upon a wave-cut plane that was formed at the same time as the subaerial erosion surface now lying at 540 feet near Harrisburg, the

[86] Barrell, Joseph, op. cit., p. 336.

[87] Shattuck, G. B., Development of knowledge concerning the physical features of Cecil County: Maryland Geol. Survey, Cecil County, p. 166, 1902.

[88] Bascom, F., The resuscitation of the term Bryn Mawr gravel: U. S. Geol. Survey Prof. Paper 132, pp. 117–119, 1924.

name Bryn Mawr is used to denote this subcycle of erosion. In an earlier paper [89] this subcycle was called the Harrisburg.

Brandywine subcycle of erosion.—South of Harrisburg the 560-foot level of the shale hills descends abruptly to a wide lowland at about 400 feet underlain by calcareous rocks. This lowland, although known as Lebanon Valley, is not genetically related to a single stream, for Swatara Creek, which occupies the valley during a portion of its course, flows out of the limestone into the shale hills, utterly regardless of lithologic control. South of Lebanon Valley Lancaster Val-

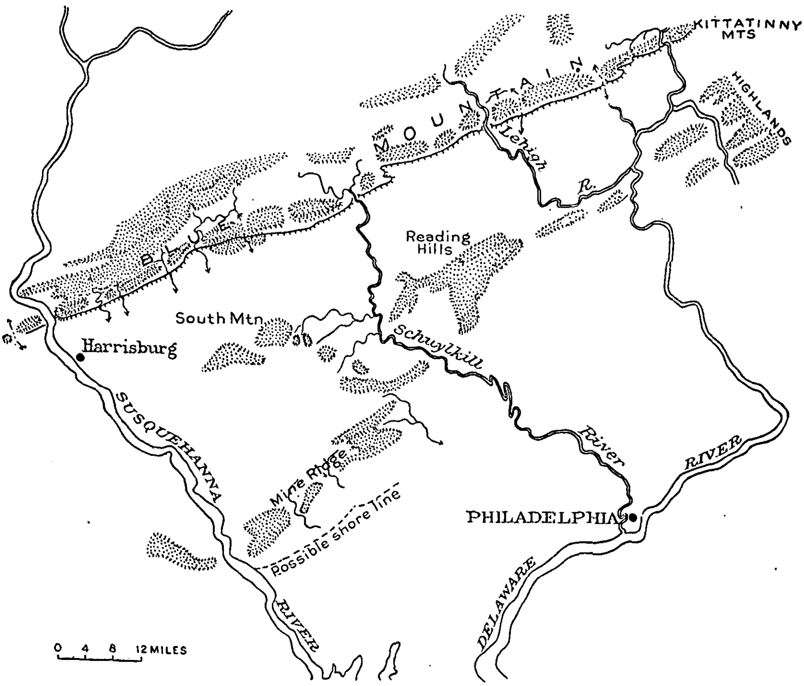

FIGURE 7.—Generalized topography of eastern Pennsylvania and Maryland at the end of the Honeybrook subcycle of erosion. Stippled areas are uplands; hachured lines, escarpments

ley stretches out in a wide, gently undulating surface of low relief, underlain by limestone and dolomite. About 130 square miles of country east of Susquehanna River maintains an approximate level of 380 to 420 feet. Above this level rise the low ridges whose summits are remnants of the dissected Bryn Mawr peneplain. That the 400-foot level has been produced by erosion and is not due to the character of the underlying calcareous rock is established by the fact that the surface bevels the folded schist, limestone, and dolomite in the McCalls Ferry quadrangle east of Pequea Creek and north of the Martic Hills. South of Conestoga the presence of a notch 100 feet

[89] Knopf, E. B., op. cit., p. 658.

deep whose base is at an altitude of 400 feet suggests a probable wind gap, which therefore fixes a point upon a stream that dissected the Bryn Mawr peneplain.

The remarkable incised meanders of Swatara and Conodoguinet Creeks were evidently initiated upon the surface of the local peneplain formed in the Bryn Mawr erosion cycle and cut down during the erosion that dissected the Bryn Mawr surface. The lowlands of the present topography are therefore due to post-Bryn Mawr erosion,

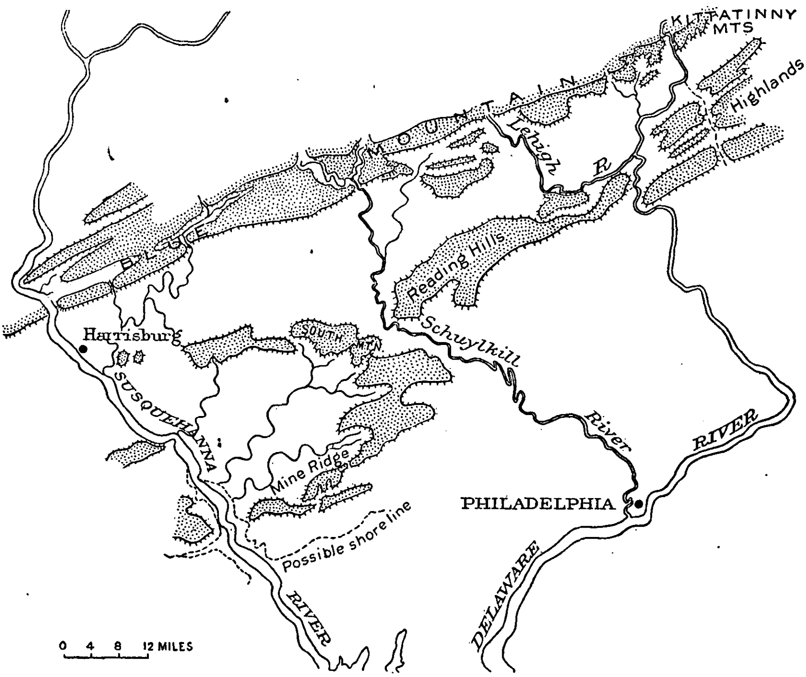

FIGURE 8.—Generalized topography of eastern Pennsylvania and Maryland at the end of the Harrisburg subcycle of erosion. Stippled areas are uplands; hachured lines, escarpments

whereas the rolling hills that bound the wide lowlands have been formed by erosion during the Bryn Mawr subcycle.

The duration of the subcycle of erosion that dissected the Bryn Mawr surface was relatively short, as pointed out by Campbell,[90] who called attention to the rise of the country underlain by limestone to more than 500 feet on the divide between the Susquehanna and Schuylkill drainage systems. Similarly on the divide between the Schuylkill and the Lehigh the 400-foot level is absent, showing that the Brandywine erosion, although relatively extensive where the underlying rock is soft, was confined to areas near the master streams

[90] Campbell, M. R., Geographic development of northern Pennsylvania and southern New York: Geol. Soc. America Bull., vol. 14, p. 283, 1903.

and nowhere succeeded in reducing the divide, although it is also on soft rock, to the level of Lancaster Valley.

The brevity of the time during which the limestone was being eroded is also shown by the fact that south of Mine Ridge the widespread surface at 380 to 400 feet is lacking but is replaced by a well-defined terrace-like surface at 320 to 340 feet. This shoulder or terrace can be followed up the valleys of the Susquehanna and its tributaries for a certain distance and then disappears, being replaced

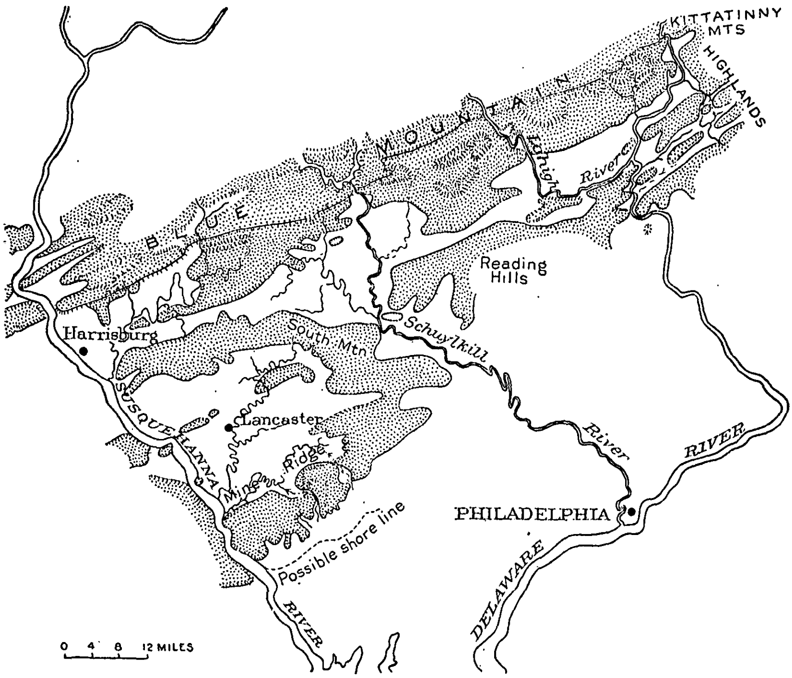

FIGURE 9.—Generalized topography of eastern Pennsylvania and Maryland at the end of the Bryn Mawr subcycle of erosion. Stippled areas are uplands; hachured lines, escarpments

farther upstream by another terrace at a higher level. The point on the stream valley where the 340-foot terrace disappears is marked by a knickpoint in the longitudinal stream profile, thus suggesting that this point may be the headward limit of rejuvenated erosion for the partial cycle that cut the stream terrace at 320 feet. Although this partial cycle lasted only long enough to permit stream terracing of valleys on resistant rocks in southern Pennsylvania, a local peneplain was developed farther north, across the easily eroded limestone of Lancaster Valley. (See figs. 9 and 10.)

The 380 to 420 foot surface in the region east of the Susquehanna appears to correspond to what has been called the Somerville plain by Campbell.[91] But the name Somerville has been used to describe a low-lying surface that occupies much of the Piedmont area in New Jersey and slopes southeastward within the Raritan quadrangle from an altitude of 200 to 300 feet near the base of the Highlands escarpment to about 100 feet in the southeast corner of the quadrangle.[92] This gives a slope of 11 feet to the mile, which is excessive as compared with the slopes of the older erosion surfaces and which

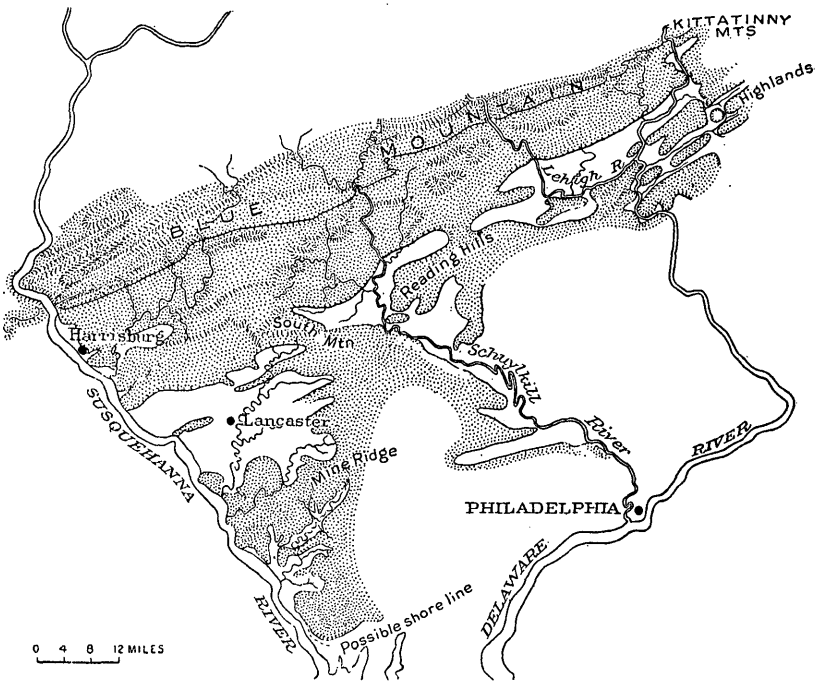

FIGURE 10.—Generalized topography of eastern Pennsylvania and Maryland at the end of the Brandywine subcycle of erosion. Stippled areas are uplands; hachured lines, escarpments

undoubtedly indicates the presence of more than one plane of erosion. In the neighborhood of Frankfort the local plane surface at 140 feet, carrying Pensauken gravel, is probably equivalent to the surface having the same altitude around Somerville.

The name Somerville, as originally used, clearly embraces several different erosion surfaces, and because the surface around Somerville was developed later than the local peneplain preserved in the floor of Lancaster Valley the latter can not now be considered Somerville. It is possible that the fluvial erosion surfaces preserved in Lancaster Valley can be correlated with the marine gravel of the Brandywine

[91] Campbell, M. R., Geographic development of northern Pennsylvania and southern New York: Geol. Soc. America Bull., vol. 14, p. 283, 1903.

[92] Bayley, W. S., Kümmel, H. B., and Salisbury, R. D., U. S. Geol. Survey Geol. Atlas, Raritan folio (No. 191), p. 3, 1914.

that lies at an altitude of 230 feet at the type locality in Maryland. This gravel is undoubtedly the equivalent of the low-level gravel (Brandywine formation) at an altitude of 200 feet in Elk Neck, on the eastern shore of Chesapeake Bay. The relation of the "late Brandywine"[93] surface to the Sunderland terrace on Elk Neck corresponds so closely to the relation between the 320-foot benches along the Susquehanna and the Sunderland benches at 200 to 220 feet that the contemporaneity of these 320-foot terraces and "late Brandywine" erosion is suggested, and the name Brandywine[94] is used to denote the subcycle of erosion in which the terraces were cut and the local peneplain was formed in Lancaster Valley.

Late Pleistocene erosion.—The Atlantic coast line is bordered by three terraces known as the Sunderland, Wicomico, and Talbot terraces. The oldest of these is the Sunderland, and the deposits of each of the two younger terraces wrap around the scarp of the next older one and extend up the valleys cut in the older terrace, thus showing that after the formation of each successive terrace the sea level was lowered and the terrace underwent a certain amount of erosion, after which the sea once more encroached upon the land and the deposits of the succeeding terrace were laid down in the lower portions of the stream valleys. This sequence of events was pointed out by Shattuck,[95] who also called attention to the fact that the surface was in many places stripped of one formation before the succeeding deposit was laid down.

SIGNIFICANCE OF KNICKPOINTS IN STREAM GRADIENT

On the principle that uplift will cause an alteration in the normal profile of a stream course and that maturity is first established in the lower part of the stream course, the new mature profile after rejuvenation will appear near the mouth and will extend backward as far as the stream has succeeded in cutting back its channel before interruption by another uplift. Therefore, the uplifts during which the Pleistocene terraces were dissected and partly removed should have left some record of their quickening effects upon the stream erosion in the cross profiles and longitudinal profiles of the larger streams of the region.

However, it must be remembered that abrupt changes in the longitudinal profile may be produced in other ways than by rejuvenation: (1) A steepening of the curve at the point where the river channel passes from limestone into quartzite or vice versa might produce a sudden steepening of gradient but would obviously have no effect

[93] Bascom, F., U. S. Geol. Survey Geol. Atlas, Elkton-Wilmington folio (No. 211), p. 3, 1920.

[94] Knopf, E. B., op. cit., p. 659.

[95] Shattuck, G. B., The Pliocene and Pleistocene deposits of Maryland: Maryland Geol. Survey, Pliocene and Pleistocene, pp. 135–136, 1906.

upon the upstream limit of accelerated stream cutting. The effect of harder beds athwart a stream would not, however, be seen in a stream that has already cut its channel to maturity. A knickpoint that indicates the headward limit of rejuvenated erosion may coincide with the interposition of hard layers across the river channel and thus may not be caused by the hard layer alone. The stream profile will then not show similar irregularities at all such points where hard layers occur athwart the stream course. Knickpoints are particularly significant as indicating rejuvenation where they occur as gentle arches upon an approximately horizontal curve in the profile of a stream that has attained maturity, rather than where they occur as steps in a steep curve that denotes a stream that is cutting downward in a youthful valley. (2) Deviations from the normal stream profile may also be produced by climatic factors, such as climatic changes during Pleistocene time. But such secular changes in climate would influence the eroding power of streams in this region as a whole and would not therefore produce knickpoints on the stream gradient. (3) Sudden steepening in a prevailing gentle curve may be caused by local uplift across the bed of a stream, as pointed out by Campbell [96] in the explanation of Muscle Shoals on Tennessee River. But the value of knickpoints as an aid in recognizing regional uplift lies in the fact that the headward limit of stream rejuvenation caused by such an uplift will be recognizable not only in the main stream but also in the tributaries at points which clearly are not conditioned by local factors, such as bedrock, faulting, or local upwarping. Knickpoints can be used as records of physiographic history only after many data have been assembled and critically evaluated for the whole region. The irregularities in longitudinal stream profile thus determined to be significant must be corroborated in the corresponding cross profiles of the stream valleys. The present study of the drainage of that part of the eastern Atlantic slope in which the McCalls Ferry-Quarryville district lies is obviously only a small fraction of the whole study necessary to establish conclusively the physiographic history of the region as a whole. It is presented here as suggestive rather than as complete, in the hope that other investigators may be tempted to apply it to adjoining regions and that thus the results suggested by this investigation may be tested.

LONGITUDINAL PROFILES OF SUSQUEHANNA RIVER AND MINOR STREAMS

As far back as 1888 McGee [97] noted the fact that Susquehanna River, in common with other rivers of the eastern Atlantic slope that

[96] Campbell, M. R., Drainage modifications and their interpretation: Jour. Geology, vol. 4, p. 667, 1896.

[97] McGee, W J, The geology of the head of Chesapeake Bay: U. S. Geol. Survey Seventh Ann. Rept., pp. 548–558, 1888.

flow across both the Piedmont Upland and the Coastal Plain, is characterized by a mature concave profile at the mouth, which changes abruptly, a short distance from the mouth or from the entrance of the river into the Coastal Plain, into a convex profile of high declivity characteristic of an actively corrading stream. The available map data for Susquehanna River are shown in the following table:

Approximate altitude and distance from the mouth of several points on the longitudinal profile of Susquehanna River

Locality	Altitude (feet)	Distance from mouth (miles)	Distance between points (miles)	Fall between points	
				Feet	Feet per mile
Havre de Grace	0	0			
Amos Falls	40	9.5	9.5	40	0.4
Holtwood Dam	120	25	17	80	4.7
Turkey Hill	200	39	12	80	6.6
York Haven	260	56	18.5	60	3.24
Half Falls	360	93	38	100	2.6
Liverpool	380	100	7	20	2.9
Selinsgrove	420	118	20	40	2

A longitudinal profile of Susquehanna River made from the available map data is shown in Figure 11. In this profile a slightly concave curve extends almost horizontally for a distance of 3 miles from the mouth; thence upstream a steeper curve extends to a point about 9 miles from the mouth, where there is a low fall or rapids, shown on the topographic map as Amos Falls. From this point northwestward to the Holtwood Dam, a distance of 25 miles from the mouth, the river flows along a slightly convex immature profile. The Holtwood Dam has utilized a natural falls or rapids known as Cullys Falls. From the dam northwestward the river gradient maintains a convex curve as far as Turkey Hill, 39 miles from the mouth.

It is a notable fact that these changes in gradient take place in a rock bottom that is of approximately homogeneous resistance to erosion, and from a point about 3 miles north of Havre de Grace to Turkey Hill the river flows in a rocky gorge, whose local widening and constrictions show no relation to the steepenings of the longitudinal profile. Thus there are three stretches of the river channel between the mouth and Turkey Hill delimited by knickpoints at altitudes of about 40 feet and 120 feet. These three stretches suggest the influence of three different periods of stream cutting interrupted by successive uplifts.

North of Turkey Hill the river channel passes into limestone and dolomite. There is a suggestion that the flattening of the profile north of Turkey Hill is not caused by change in the underlying rock, because for a distance of 18 miles from Turkey Hill to the vicinity of Yorkhaven the river maintains an even and relatively low gradient over limestone, quartzite, shale, and sandstone. Although not

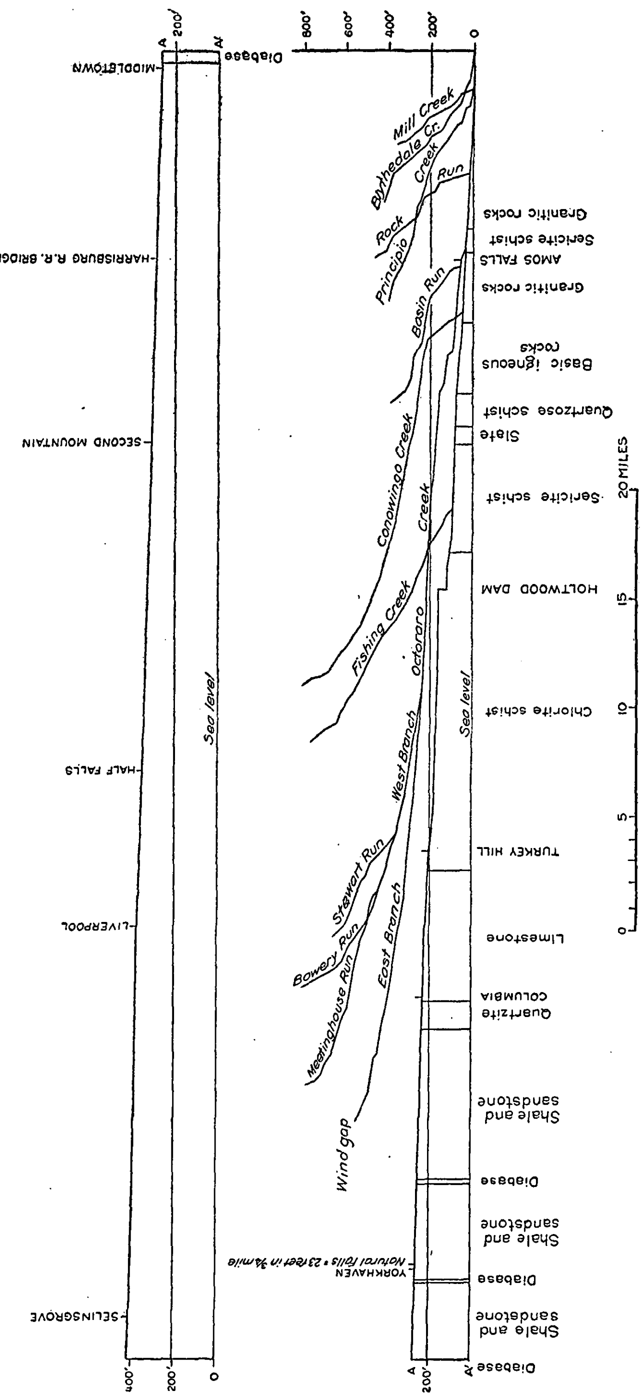

FIGURE 11.—Longitudinal profiles of Susquehanna River from its mouth to a point 3 miles southwest of Sunbury, Pa., and of eight minor streams

shown in the profile constructed from map data (see fig. 11) the natural fall of the river near Yorkhaven is steep, amounting to 23 feet in three-quarters of a mile at Conewago Falls.[98] This steepening is conspicuously shown in the Susquehanna River profile published in the report by Hoyt and Anderson and might be ascribed to the presence of diabase in the river bed at this point, but between Yorkhaven and Half Falls, a distance of about 38 miles, the river flows across Triassic strata, diabase, and Paleozoic limestone and shale along a uniform gradient which is once more gentle, with a total fall of 60 feet. Therefore the fact that Conewago Falls occurs at a point where the channel is in diabase does not appear to explain the knickpoint. Between Half Falls, at 340 feet, and a point 2 miles north of Selinsgrove, where the altitude is 420 feet, the river falls 60 feet in 20 miles, and above Selinsgrove the gradient is almost horizontal for a long distance up the main branch of the Susquehanna.

There are therefore knickpoints on the profile of the Susquehanna River at 40, 120, 200, 260, 340, and 420 feet. To test the significance of these knickpoints a number of longitudinal profiles were constructed for tributaries of the Susquehanna and also for two streams draining directly into the head of Cheasapeake Bay. In order to remove the subjective element as far as possible the profiles of each stream were constructed on more than one scale using different vertical exaggeration. The lithologic character of the bedrock was platted on each profile, and the relation between stream cutting and the resistance of the eroded rock was thus taken into account in estimating the significance of knickpoints. Some of the profiles are shown in Figure 11.

The following table shows the knickpoints that were recognized in nine profiles and that were regarded as significant.

Altitude, in feet, of significant knickpoints on longitudinal profiles of Susquehanna River and its tributaries

<table>
<tr><th>Point</th><th>Susquehanna River</th><th>Principio Creek</th><th>Blythedale Creek [a]</th><th>Mill Creek</th><th>Rock Run</th><th colspan="2">Octoraro Creek</th><th>Basin Run</th><th>Conowingo Creek</th><th>Fishing Creek</th></tr>
<tr><td>Mouth</td><td>0</td><td>0</td><td>0</td><td>20</td><td>20</td><td colspan="2">40</td><td>60</td><td>50</td><td>90</td></tr>
<tr><td>1</td><td>40</td><td>40</td><td>40</td><td>60</td><td>-----</td><td colspan="2">60</td><td>-----</td><td>-----</td><td>-----</td></tr>
<tr><td>2</td><td>120</td><td>120</td><td>120</td><td>-----</td><td>-----</td><td colspan="2">120</td><td>-----</td><td>120</td><td>120</td></tr>
<tr><td>3</td><td>200</td><td>180</td><td>200</td><td>220</td><td>160</td><td colspan="2">200</td><td>200</td><td>220</td><td>200</td></tr>
<tr><td></td><td></td><td></td><td></td><td></td><td></td><td>East branch</td><td>West branch</td><td></td><td></td><td></td></tr>
<tr><td>4</td><td>260</td><td>260</td><td>-----</td><td>-----</td><td>240</td><td>-----</td><td>-----</td><td>280</td><td>300</td><td>260</td></tr>
<tr><td>5</td><td>360</td><td>380</td><td>380</td><td>-----</td><td>380</td><td>420</td><td>380</td><td>-----</td><td>420</td><td>420</td></tr>
<tr><td>Source</td><td>-----</td><td>400</td><td>420</td><td>360</td><td>440</td><td>560</td><td>780</td><td>380</td><td>800</td><td>750</td></tr>
</table>

[a] This stream, which is not named on the map of the Havre de Grace quadrangle, flows through the village of Blythedale.

[98] Hoyt, J. C., and Anderson, R. H., Hydrography of the Susquehanna River drainage basin: U. S. Geol. Survey Water-Supply Paper 109, p. 199, 1905.

From the table it is apparent that three out of the six streams whose mouths lie at altitudes of 40 feet or less show knickpoints at 40 feet and two show knickpoints at 60 feet. From 40 feet to the mouth Principio Creek and Blythedale Creek have cut through the Talbot terrace, showing that this stretch of the stream profiles was formed during the post-Talbot erosion and suggesting that the headward limit of post-Talbot erosion in the streams whose mouths lie above sea level is 60 feet. Above the knickpoint at 40 feet Blythedale Creek and Principio Creek are cutting into the Wicomico terrace. The Wicomico deposits extend up to 80 feet, thereby suggesting that the knickpoints at 120 feet indicate the headward limit of post-Wicomico erosion. It is probable that the knickpoint at 160 feet on Rock Run has the same significance, although the profile of Rock Run shows no change in gradient at 40 feet or at 120 feet. There is a knickpoint at 160 feet in the profile of Octoraro Creek which coincides with the presence in the stream bed of a pegmatite ledge shown in the geologic map of Cecil County. This knickpoint is therefore disregarded, as being due to rock control. A knickpoint at 80 feet on the profile of Principio Creek is disregarded, because it does not appear on the other eight profiles. Every one of the nine profiles shows a knickpoint at elevations ranging from 160 to 220 feet. On the profile of Blythedale Creek the knickpoint occurs at 200 feet and on Principio Creek at 180 feet. These streams have both cut through a Sunderland cover that attains an altitude of 180 feet, thus suggesting that 200 feet is the headward limit of post-Sunderland erosion.

The knickpoint at 260 feet that coincides with Conewago Falls on the Susquehanna is duplicated in five out of the other eight profiles. It does not show in the bed of Octoraro Creek, a fact which may possibly be explained by confluence of the two main branches of Octoraro Creek at 240 feet. Figure 12 shows cross profiles of the Octoraro Valley at various places and of the wind gap against which the Octoraro heads. The cross section near Porter Bridge (fig. 12, C) shows the 3-story valley formed by the entrenchment of a valley superimposed upon the surface that carries Bryn Mawr gravel at Woodlawn. *abcd* shows the outline of the post-mature valley cut in the Brandywine erosion cycle, during which a rock bench at 300 to 320 feet was developed by lateral planation. Within this valley of Brandywine age a shallow valley at *cdfg* is associated with a similar bench at 220 feet. These 220-foot benches can be traced down Susquehanna River to the mouth, where they appear to merge into the level occupied by remnants of Sunderland deposition. This valley at *cdfg* is probably a Sunderland valley, and the streams of the region that formed the 220-foot benches were

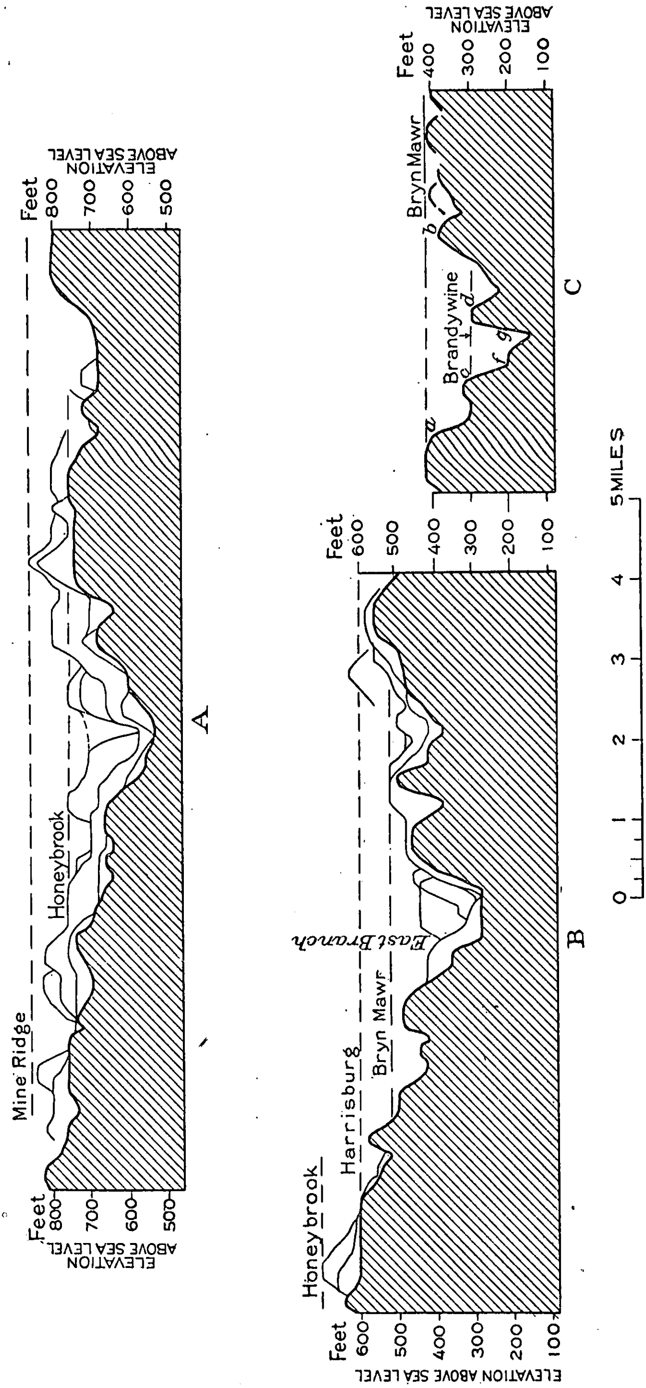

FIGURE 12.—Cross profiles of Octoraro Valley and of the wind gap above the source of Octoraro Creek. A, Projected profiles along Mine Ridge across Gap wind gap N. 75° E.; B, projected profiles across valley of East Branch of Octoraro Creek northeast of confluence of East and West Branches N. 50° W.; C, cross profile of the valley of Octoraro Creek between Porter Bridge and the mouth of Stone Run

probably cutting along the gradient above the 200-foot knickpoint during the time that a marine plain was formed and covered with Sunderland deposits, which now lie at altitudes from 100 to 180 feet. The 260-foot knickpoint thus probably represents the headward limit of stream cutting during Sunderland marine deposition. In each stream whose source lies above 360 feet there is a knickpoint at an altitude between 360 and 420 feet, in four at 380 feet, and in three at 420 feet. It is probable, therefore, that during the erosion cycle that cut the 300 to 340 foot benches which are conspicuous along Susquehanna River and Octoraro Creek for a certain distance the headward limit of stream erosion had attained approximately 400 feet, which thus marks the limit of Brandywine rejuvenation.

Above 420 feet on the longitudinal profile of the East Branch of Octoraro Creek the cross profiles just below the wind gap show shallow valleys in the wide Harrisburg erosion surfaces at 600 to 620 feet. Therefore from 420 feet to the source, at 560 feet, a stream was cutting into the Harrisburg surface. Directly above the source of the East Branch the wind gap at Gap records the course during Harrisburg time, when the ancestral Octoraro Creek, which flowed through the gap, still had its source north of Mine Ridge. During the Bryn Mawr subcycle the stream was diverted and the Harrisburg valley was hung up as a wind gap.

Mine Ridge is a local divide, and above the level of the wind gap the record of present stream courses fails, because the streams that flow northward from the divide are of more recent age than the summit divide. The evidence of previous cycles lies in the superimposition of the ancestral Octoraro Creek across the Mine Ridge surface and in the shape of the composite valley of the wind gap (fig. 12, A), which records in its gentle upper slope the intrenchment of the superimposed stream accompanied by extensive lateral planation during the Honeybrook subcycle and in its steep inner gorge the downcutting of the Harrisburg valley from which the Octoraro Creek was diverted, probably in the early stages of Bryn Mawr rejuvenation.

The profiles of Octoraro Creek and of the gap above its source, therefore, suggest the physiographic history of the region from the period of valley cutting that dissected the Wicomico marine terrace as far back as the Mine Ridge cycle.

SEQUENCE AND CORRELATION OF EROSION RECORDS

It has been possible, therefore, by an analysis of present and past drainage to suggest that eight successive uncompleted erosion cycles are recorded in the region between Chesapeake Bay and Mine Ridge.

North of Mine Ridge vestiges of two older cycles are preserved in remnants of former erosion surfaces occupying the summits of South Mountain.

AMOUNT AND DURATION OF EROSION

The relative amount and duration of erosion comprised in each of the 10 complete or partial cycles of erosion, whose existence has been indicated, are impossible to estimate, because there is no means of determining the maximum area undergoing erosion during the older cycles, the agents by which the work of these cycles was accomplished, or the rate at which they acted. During the Kittatinny cycle the region between South Mountain and Blue Mountain was reduced to a plane surface of low relief that was cut across the most resistant formation, giving for the region lying between Delaware and Susquehanna Rivers a minimum area undergoing erosion amounting to 2,000 square miles, which is undoubtedly only a fraction of the original extent. In the succeeding three cycles the corresponding eroded area amounted to at least 5,000 square miles, but the reduction was much less complete than in the previous smaller area. It is evident that a large part of the 5,000 square miles was undergoing fluvial erosion, which probably progresses more slowly than marine planation; and the failure to reduce the region completely during all post-Schooley erosion may be due to the fact that the Kittatinny and Schooley peneplains were formed by marine erosion, which reduces alternating belts of hard and soft rock with variable rapidity, but with equal completeness. If, on the other hand, the 2,000 square miles that was eroded during the Kittatinny and Schooley erosion cycles was degraded under subaerial conditions the length of these two cycles must have been appreciably longer than that of any succeeding partial cycle.

PERIODICITY OF UPLIFT

Whatever the duration of the earliest cycles it must be emphasized that the present study of the evolution of drainage in the region between the Blue Mountain range of the Appalachian Ridge and Valley province and the Coastal Plain clearly disproves the idea of numerous successive peneplanations of the area and at the same time equally clearly establishes the recurrence of numerous slight uplifts that caused repeated interruptions to the continuity of base-leveling. The three youngest uplifts were oscillatory in character along the inland margin of the present Coastal Plain. Whether the older uplifts were accompanied by similar oscillations in level is not clear, but the net result of coastal movement has been an intermittent but progressive elevation that has raised the Blue Mountain-

Kittatinny range at least 1,600 feet in a saltatory series of slight uplifts. Such a succession of slight uplifts operating throughout a long time suggests a periodic relief of a steadily accumulating vertical strain. It is an interesting speculation whether such periodic movements record the periodic establishment of isostatic equilibrium along the border of an area of deposition, rather than the continuous maintenance of isostatic balance demanded by the extreme isostasists.

AGE OF EROSION CYCLES

The effect of such repeated uplift is to postpone indefinitely the total degradation of a region by constant addition to the area undergoing reduction and by setting back the progress of the work, so to speak, after each interruption. It is therefore idle to speculate as to the age of the oldest surface in the region on the basis of the probable time required to reduce such a surface to sea level.

The evidence on the age of erosion cycles that is based upon the correlation of the present slope of the old fluvial gradient with the present slope of buried peneplains is perhaps more reliable, and according to this criterion the gentle seaward slope of the old fluvial gradients suggests that no erosion cycle recorded in the region between Blue Mountain and the Coastal Plain is older than late Tertiary. The gaps in the geologic record over late Cretaceous and early Tertiary time may eventually be filled, at least partly, by a physiographic study of the Atlantic drainage slope in areas more remote from present base-level than those covered by the present study.

ORIGIN OF METAMORPHISM IN THE CRYSTALLINE SCHISTS OF THE PIEDMONT PROVINCE

VARIABLE METAMORPHIC INTENSITY IN THE McCALLS FERRY-QUARRYVILLE DISTRICT

The most striking feature of the crystalline schists of the McCalls Ferry-Quarryville district is the relatively intense metamorphism of the autochthonous Paleozoic strata in comparison with that of the pre-Cambrian rocks in the overlying thrust block. The argillaceous Cambrian sediments on the flanks of the Mine Ridge anticline have been metamorphosed to garnetiferous biotite schists, and impure Ordovician limestone in the Chester and Quarryville Valleys has been altered to micaceous marble that contains numerous beds of mica schist. The metamorphism of the pre-Cambrian rocks that have been thrust northward is most intense in the cores of the double-crested Tucquan anticline, which is the southwestward continuation of the downward plunging anticlinal fold underlying the Mine Ridge upland. But in the Mine Ridge anticline, where the pitch of the anticline brings up the Paleozoic rocks that underlie the thrust plane, the overriden Paleozoic schists show an intensity of metamorphism that exceeds even the maximum alteration of the older overthrust rocks. As the metamorphic intensity diminishes on both sides of the Mine Ridge-Tucquan anticlinal axis the maximum metamorphism may be said to coincide with the core of the double-crested arch.

PREVIOUS IDEAS ON METAMORPHISM OF THE PIEDMONT PROVINCE

A pronounced variation in intensity of metamorphism throughout the Piedmont province has long been recognized. Williams[99] in 1891 first called attention to the fact that the rocks of the Piedmont Plateau consist of an eastern area of strongly crystalline rocks and a western area of less metamorphosed rocks, which he called semicrystalline. His generalizations from the facts then available were comprehensive and striking, but he drew the erroneous conclusion that the highly crystalline eastern rocks are older than the less metamorphosed series on the west. In 1904 Mathews[1] pointed out that the rocks of the eastern and western areas are probably contemporaneous and that the difference in metamorphism is caused by the presence of large bodies of plutonic rocks in the eastern area, whereas the rocks of the western area inclose only smaller bodies of surface volcanic rocks.

The results of the present more extended and detailed field study of the Piedmont province has shown that the variation in intensity of metamorphism does not correspond to any simple generalization of a western area of semicrystalline rocks and an eastern area of thoroughly recrystallized rocks. This field study combined with a detailed petrographic examination of the rocks both in hand specimen and in thin section has furnished data for a fairly detailed and accurate generalization of relative metamorphic intensity. This generalization is shown in Figure 13, which is a diagrammatic representation of the region between meridians 75° and 77° and parallels 40° 15′ and 39°. Three stages of metamorphic intensity are represented by three different patterns. These three stages have been determined by applying certain definite criteria for measuring the degree of metamorphism.

[99] Williams, G. H., The petrography and structure of the Piedmont Plateau in Maryland: Geol. Soc. America Bull., vol. 2, pp. 301–317, 1891.

[1] Mathews, E. B., The structure of the Piedmont Plateau as shown in Maryland: Am. Jour. Sci., 4th ser., vol. 17, p. 158, 1904.

CRITERIA FOR DETERMINATION OF METAMORPHIC INTENSITY

As the chemical reactions of metamorphism depend mainly upon temperature and upon unequal pressure (shearing stress),[2] it follows that the higher the temperature and the greater the shearing stress the more intense will be the metamorphism. Two indices of the conditions under which metamorphism takes place are the texture and the mineral constituents of the resultant schist or gneiss. The

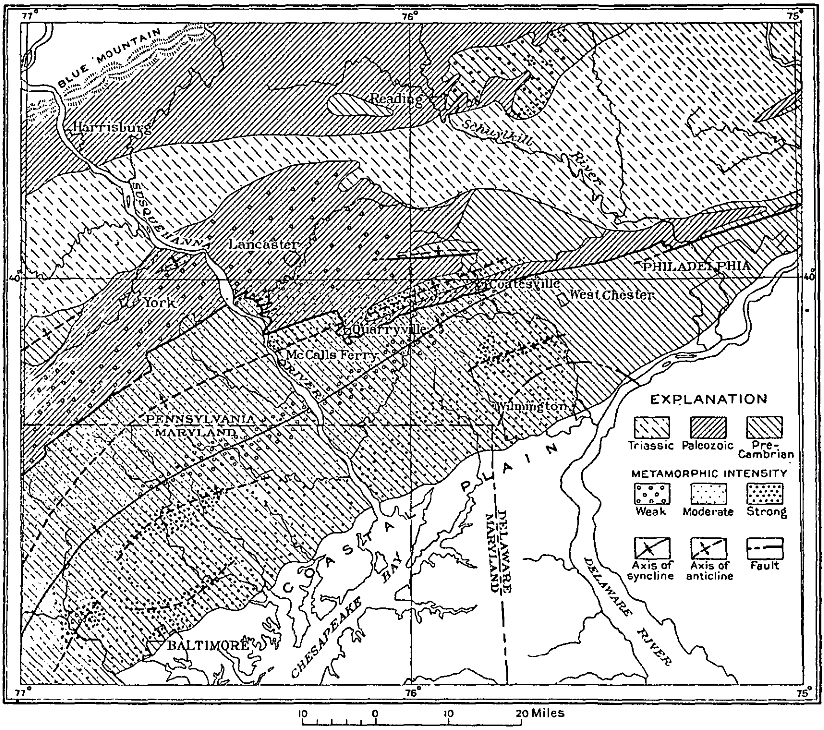

FIGURE 13.—Diagrammatic map showing distribution of metamorphic intensity in the crystalline schists of the Piedmont province of portions of Pennsylvania and Maryland

significance of these indices, of course, depends upon the character of the rock that has been altered. The same intensity of metamorphism will produce from an argillite a mica schist, in which the original constituents are completely obliterated, and from a conglomerate a schistose conglomerate, in which most of the original constituents are recognizable in a more or less mechanically deformed condition.

For purposes of comparison, therefore, the original rocks in different localities must have been of approximately the same suscepti-

[2] Johnston, John, and Niggli, Paul, The general principles underlying metamorphic processes: Jour. Geology, vol. 21, pp. 482–516, 588–624, 1913. Johnston, John, Pressure as a factor in the formation of rocks and minerals: Idem, vol. 23, pp. 730–747, 1915.

bility to metamorphosing influences, or else the application of criteria for metamorphic intensity must be accompanied by a certain correction for relative degree of susceptibility. Thus the character of the original rock from which the schist has been derived will be considered in the discussion of criteria of metamorphic intensity.

METAMORPHIC TEXTURES

Metamorphic textures fall into two main groups—cataclastic texture and crystalloblastic texture.

Cataclastic texture.—The cataclastic texture is associated with mechanical deformation in the upper zone of metamorphism and is regarded as a metamorphism of low-grade intensity. It is characterized by all stages of mechanical breaking down, from strain shadows, granulation, and mortar structure in the brittle constituents, such as quartz and feldspar, to slicing and twisting of the fibrous and platy minerals, such as hornblende and mica. The end result of cataclastic deformation is a mylonite in which the original constituents have been rolled out and sheared to such an extent that they are recognizable only with difficulty.

Such mechanical deformation is usually combined with the development of new minerals, owing to the fact that some rock constituents readily undergo a rearrangement into new minerals whose shape conforms to the metamorphic requirements, while other more resistant constituents—as, for example, quartz—show only the mechanical effects of stress. Thus low-grade metamorphism will produce new minerals and parallel arrangement in a fine-grained rock such as a shale, whose constituents yield readily to rearrangement. In such a rock the low intensity of the metamorphism is indicated not by the cataclastic texture but by the fine grain of the constituents in the resultant schist. Under the same intensity of deformation a rock that is coarser grained than a shale will show only mechanical strain in the original constituents. Therefore while cataclastic texture is always indicative of metamorphism of low-grade intensity the reverse relation is not necessarily true—that is, low-grade metamorphism does not always produce cataclastic texture.

Crystalloblastic texture.—A complete recrystallization of the original constituents of the rock produces the so-called crystalloblastic texture, which differs from igneous texture in the order or rather in the lack of definite order of crystallization. Each mineral in a crystalloblastic rock has been formed at nearly the same time, idiomorphism of form being conditioned not by the laws that govern crystallization from a rock melt but by the relative ease with which certain minerals break down and form new constituents and by the power of crystallization, by virtue of which the newly formed mineral can maintain its boundaries against its associates.

Crystalloblastic texture is an index of complete recrystallization and is therefore the result of high-grade metamorphic intensity, which is characteristic of the deep-seated zone of metamorphism. But, once more, the converse is not true, for crystalloblastic texture may also result from a complete recrystallization of the original constituents under conditions of temperature and pressure that are characteristic of the upper zones of metamorphism and therefore indicative of fairly mild metamorphic intensity.

METAMORPHIC MINERALS

Fortunately, although texture is in many respects of negative value in estimating metamorphic intensity, we have an additional criterion in the nature of the metamorphic minerals that are formed. The formation of new minerals by recrystallization is dependent upon three factors—namely, cubical pressure, temperature, and shearing stress. The most influential factor is the temperature. Under normal conditions the temperature gradient increases with depth, but under such conditions as batholithic intrusion into the upper levels of the earth's crust the temperature even at comparatively shallow depths may be abnormally high, and the regional contact metamorphism produced by such an intrusion may be of a strong order of intensity. Therefore minerals that are stable at high temperatures, such as sillimanite, biotite, and oligoclase, are indices of high-grade metamorphic intensity but do not necessarily indicate the depth at which alteration took place. On the other hand, such low-temperature minerals as albite, chlorite, and sericite are indices of a relatively low metamorphic intensity, which is never associated with great depth.

The character of the new minerals formed during recrystallization will often furnish a clue to the conditions under which metamorphism took place and thus evaluate relative degrees of metamorphic intensity in rocks whose original character was such that cataclastic alteration predominates over the formation of new minerals. For example, a quartz conglomerate whose matrix is argillaceous will alter to a schistose conglomerate in which the stretched or mashed pebbles are embedded in a mica schist. Such a micaceous matrix in which biotite is the dominant mica indicates the operation of more intense metamorphism than is shown by a matrix composed of muscovite, chlorite, or chloritoid.

PROGRESSIVE METAMORPHISM

The application of the criteria that have just been outlined shows the wide range in the effects of metamorphism, from simple undulatory extinction in quartz caused by a slight mechanical deformation to the complete transformation of the original constituents of a rock into minerals that are stable under conditions of high temperature and pressure. The complete process of metamorphism may be illustrated by the passage of a shale into a coarse-grained mica schist under conditions of steadily increasing metamorphic stress. When the normal process is arrested before completion we find a slate, phyllite, or fine-grained mica schist, according to the stage reached.

The first attempt to classify metamorphic rocks according to their stage of development at the moment of arrest was made by Sederholm, who grouped them into rocks altered in an upper zone of metamorphism and rocks altered in a lower zone. Later Becke and Grubenmann amplified the scheme of classification so as to embrace rocks formed in three metamorphic zones that correspond to upper, intermediate, and lower depth.

In the present regional study the relative intensity of metamorphism is represented upon the map by three degrees, as follows:

First degree, low metamorphism. The original texture is easily recognizable. Cataclastic deformation is marked. Certain new minerals have been formed, chiefly sericite. There is incipient dimensional parallelism.

Second degree, moderate metamorphism. Crystalloblastic texture is complete or nearly so. Relict structures are absent or rare. There is strong dimensional parallelism. New minerals have been formed that are characteristic of low temperature, such as albite, chlorite, and muscovite.

Third degree, high metamorphism. Crystalloblastic texture is complete. There is strong dimensional parallelism. New minerals have been formed that are characteristic of high temperature, such as biotite, oligoclase, and sillimanite.

These classes are called degrees rather than stages, because they do not necessarily represent completeness of metamorphism, although the first degree is obviously associated with partly advanced alteration. Both the last two degrees are associated with the end products of the metamorphic process under different conditions. Second-degree intensity is often produced at the end of alteration in the upper or intermediate zones, under conditions of moderate metamorphic intensity, and third-degree intensity is the final result of alteration under conditions of very strong metamorphism, characteristic of the deep zone. Thus a rock that has been completely reconstituted in the upper metamorphic zone may be subjected to increased metamorphic intensity by deeper burial or as a consequence of batholithic intrusion, and it will therefore renew the process of chemical interaction and crystallization in an attempt to fit itself to the new environment. The sum total of all such changes may be considered progressive metamorphism, because the ultimate product of the most intense metamorphism is a rock that has undergone great condensation of volume and increase in density. Such a rock may be considered the end product of anamorphism. Certain changes in its physical environment may result in its further alteration by refusion into a liquid state, but as a rock it can undergo no further change except a reversal of metamorphism.

RETROGRESSIVE METAMORPHISM

In order to appreciate the complex metamorphic history of rocks like the pre-Cambrian schists that have undergone repeated alterations it must be remembered that the effects of repeated metamorphism are not necessarily cumulative and that later deformation often results in the breaking down of textures and constituents that have been built up during an earlier metamorphic process. Such a breaking down may be recognized by the superimposition of cataclastic effects and low-grade metamorphism upon a high-grade metamorphism, the result being the formation of diaphthoritic schists such as those of the Wissahickon formation. (See pp. 27–28.) The original intense metamorphism by which the schist was formed is disclosed by certain relict minerals that are characteristic of deep-zone metamorphism. For example, garnets formed in an earlier intense metamorphism have later been converted, at least partly, to chlorite. Other minerals, such as biotite, which are stable under deep-seated conditions, have recrystallized in the upper zone as chlorite. In many chlorites the titanium content of the original biotite has been freed as rutile. The question how far earlier metamorphism may be masked or even obliterated during a later deformation thus adds an increased complexity to the already obscure history of rocks that have passed through more than one orogenic period.

REGIONAL VARIATION IN METAMORPHIC INTENSITY

In the preparation of the map shown in Figure 13 field study over a large part of the region was supplemented by a careful study of the published petrographic descriptions of the rocks of Cecil County, Md.,[3] and the southeastern portion of Pennsylvania.[4] A summary of the data for the construction of the map is given below.

EARLY PRE-CAMBRIAN PARAGNEISSES (BALTIMORE GNEISS)

The number of thin sections studied was 25. All show crystalloblastic texture and minerals characteristic of the deep zone of metamorphism. Two show superimposed mechanical deformation in a zone of overthrust faulting on the east side of the Woodstock anticline, 3 miles north of Ellicott City, Md.

[3] Bascom, F., The crystalline rocks of Cecil County: Maryland Geol. Survey, Cecil County, pp. 83–143, 1902.

[4] Bascom, F., U. S. Geol. Survey Geol. Atlas, Philadelphia folio (No. 162), 1909; Elkton-Wilmington folio (No. 211), 1920.

LATER PRE-CAMBRIAN PARAGNEISSES (GLENARM SERIES)

Setters formation.—Seven thin sections were studied. All show crystalloblastic texture developed by an intense metamorphism, as shown by the nature of the mineral constituents. Four show superimposed mechanical deformation in areas where there has been igneous invasion of the Setters by the post-Glenarm Gunpowder granite.

Wissahickon formation (oligoclase-mica schist facies).—The number of thin sections of Maryland rocks that were studied was 25. All show crystalloblastic texture and minerals characteristic of the intermediate or of the deep-seated zone of metamorphism. The metamorphic intensity represented by these rocks ranges from the second to the third degree. Three show superimposed mechanical deformation. The thin sections of Pennsylvania rocks that were studied were 11 in number. The rocks are sillimanite-bearing biotite gneisses. Five show intense crystalloblastic metamorphism superimposed by intense cataclastic deformation. One specimen is completely mylonitized. Six sections show crystalloblastic schists whose metamorphic intensity lies between the second and third degree.

Wissahickon formation (albite-chlorite schist facies).—Three thin sections of Maryland rocks were studied. All show the crystalloblastic texture of the upper zone of metamorphism. The number of thin sections of Pennsylvania rocks that were examined was 10. They show a metamorphism similar to the albite schists of Maryland. Two show superimposed mechanical deformation. On Susquehanna River between Conestoga Creek and Pequea Creek there is a strong local intensification of metamorphism shown by the presence of biotite instead of chlorite or in combination with chlorite.

Peters Creek schist.—Of 18 thin sections studied 15 are crystalloblastic schists of the second degree of metamorphic intensity. They show textures that are characteristic of the upper zone of metamorphism. Three are cataclastic schists in the lowest degree of metamorphism containing relicts indicative of an original graywacke. Three of the 15 crystalloblastic schists contain some biotite, thus indicating a transition from moderate to strong metamorphic intensity.

Cardiff conglomerate.—Two thin sections were studied. They indicate cataclastic deformation of an arenaceous sediment composed of variably coarse constituents. Field study shows cataclastically deformed pebbles in a crystalloblastic matrix composed of muscovite, chlorite, and chloritoid, all minerals that indicate moderate metamorphic intensity.

Peach Bottom slate.—One thin section was studied. The data were obtained chiefly from field study and from microscopic descriptions of the slate by Dale.[5] The rock shows abundant fine-grained platy minerals with strong dimensional parallelism producing slaty cleavage, which indicates low-grade metamorphism of a shale.

PALEOZOIC ROCKS OF SEDIMENTARY ORIGIN

LOWER CAMBRIAN ROCKS

Hellam conglomerate member of the Chickies quartzite as exposed on the Mine Ridge upland.—No thin sections were studied. In the field the rocks, which are mica schists and conglomerates, clearly indicate strong differential metamorphism. The matrix of the schistose conglomerate is highly foliated. The quartz pebbles are cataclastically deformed (stretched) and show varying

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[5] Dale, T. N., Slate in the United States: U. S. Geol. Survey Bull. 589, pp. 111–113, 1914.

degrees of granulation. Such rocks as the conglomerates in the Hellam member of the Chickies are poor indices of metamorphism, on account of their uneven grain. The intensity may be fairly inferred from the degree of metamorphism of the mica schists that underlie and are interbedded with the conglomerate, from the overlying beds, which are crystalloblastic quartzites, and from the prevalence of biotite in the matrix of the quartz pebbles. The Hellam conglomerate is lithologically very similar to the Cardiff conglomerate, being composed of quartz pebbles in a micaceous matrix; but the relatively low-grade metamorphism of the Cardiff is shown by the fine-grained matrix of minute muscovite, chlorite, and chloritoid flakes contrasted with the fairly coarse biotite schist matrix of the Hellam. Another clue to the relatively mild metamorphism of the Cardiff is the underlying rock, which is a schistose graywacke, and the overlying rock, which is a slate—both in a low stage of metamorphism.

Quartzite member of Chickies quartzite.—The thin sections studied were eight in number. They indicate a crystalloblastic metamorphism of a relatively pure sandstone. The purity of the original sand makes the rock an unsatisfactory index of metamorphic intensity.

Harpers schist or phyllite.—Seven thin sections were studied. Six of these sections show fine-grained phyllites of first-degree metamorphic intensity. One thin section cut from a specimen collected on the west flank of the Mine Ridge anticline shows complete recrystallization and a strong dimensional parallelism owing to the dominance of micaceous minerals, chiefly biotite. The metamorphism in this locality is strong, between second and third degree intensity.

Antietam schist or phyllite.—The thin sections studied were 18 in number. They show extreme variability of metamorphic intensity, dependent on geographic location. Increase in intensity from north to south is shown by gradation from fine-grained biotitic phyllites on the eastward continuation of the Hellam Hills, north of Columbia, to fairly coarse-grained mica schist on the flanks of the Mine Ridge anticline.

Vintage dolomite.—No thin sections were studied. The massive beds of this formation are not readily susceptible to metamorphism, but the schistose beds show the development of a considerable amount of muscovite. The lower beds that grade into the underlying Antietam are dolomitic schists that indicate a strong metamorphism in the Paleozoic area west and northwest of the Mine Ridge upland, where the schists contain a considerable amount of biotite.

ORDOVICIAN ROCKS

Conestoga limestone.—Ten thin sections were studied. This formation is a valuable index of metamorphic intensity because the calcareous beds in the original sediment contained much argillaceous matter. As in the Antietam, the intensity of metamorphism in the Conestoga increases from north to south. Near Lancaster, where the formation is an impure banded limestone with graphitic laminae, the metamorphism of the Conestoga, like that of the adjoining arenaceous sediments, is of low intensity. This is shown by the fine grain of the crystallization in the impure beds and by the formation of graphite, a mineral that may indicate a low metamorphic intensity where present in calcareous rocks. Under conditions of high temperature[6] the partial pressure of carbon dioxide will be increased, and it is possible that a certain amount of carbon dioxide may combine with the carbon, thereby eliminating the graphite

[6] Johnston, John, The general principles underlying metamorphic processes: Jour. Geology, vol. 21, p. 511, 1913.

from the rock. Caution must be exercised in the use of graphite as an index of metamorphism, because it occurs in some highly metamorphosed marbles, and in pelitic sediments it will persist in the most intensely altered rocks, such as garnetiferous mica schists. The strong metamorphism of the Conestoga formation in the Quarryville and Chester Valleys is shown by the coarse grain of the marble and by the abundant development of mica in the impure argillaceous layers.

ORTHOGNEISSES

One of the chief factors in determining the alteration of igneous rocks is chemical composition. Basic igneous rocks are ready victims to metamorphic changes and alter with surprising readiness to chlorite schists, amphibolite schists, and numerous other schists collectively designated by the widely embracing term "greenstones." Siliceous plutonic rocks, on the other hand, show a truly marvelous resistance to metamorphic action. The condition of the granite gneiss at Hartley, in Baltimore County, Md., is an excellent example of the stability of granitic rocks as contrasted with the susceptibility of sediments under the same intensity of metamorphism. This granite, which was intruded into the Baltimore gneiss before the deposition of the Glenarm series, has been altered into a typical cataclastic gneiss (augen gneiss), representing a low degree of metamorphism. It is unconformably overlain by highly metamorphosed sedimentary gneisses of lower Glenarm age. Here the younger sediments have been completely reconstructed by the action of metamorphism while the older igneous rocks have only been mechanically deformed.

Moreover, the granodiorite gneiss at Port Deposit, although younger than the Hartley gneiss, shows by the inverse zoning of its plagioclase that its constituents have been recrystallized. The only explanation for the greater susceptibility of the Port Deposit gneiss is that the calcic feldspars of the granodiorite have yielded to molecular rearrangement while the potash feldspars of the granite have been merely granulated or crushed. Igneous rocks, therefore, are not valuable as specific indices of metamorphic intensity, except where rocks of approximately the same composition show marked difference in condition of metamorphism, and they are unreliable interpreters of metamorphic history except when used in combination with other facts of broader significance.

For example, the Woodstock granite in Maryland, which is a quartz monzonite containing about 40 per cent of plagioclase, is practically unaltered, but rock near by, which was originally a granodiorite containing the same percentage of slightly more calcic feldspar, has become the Port Deposit gneiss. The similarity in composition between the Woodstock and the Port Deposit is sufficient to justify the inference that the lack of metamorphism in the Woodstock is due to the fact that it was intruded during a later period of igneous action than the Port Deposit, thereby escaping most of the deformation that has left its mark upon the Port Deposit gneiss. The occurrence of the Woodstock, surrounded as it is on all sides by strongly deformed rocks into which it is intrusive, precludes the idea that it owes its comparative immunity from metamorphism to a fortunate position in which it has been protected from deforming influences.

On the other hand, the relatively slight metamorphism of the igneous complex of the Reading and Durham Hills is due to the fact that these rocks were not exposed to the stress that affected rocks of the same age in other localities. In the Reading Hills the igneous rocks as a whole are notably free from any evidence of stress except a moderate amount of mechanical deformation that is in many places concentrated along recognizable zones of faulting and brecciation. The igneous rocks intrude pre-Cambrian crystalline schists that have been formed under conditions of intense metamorphism. The fact that they are unconformably overlain by Cambrian strata makes it impossible to explain their lack of alteration by the assumption that they are younger than the lithologically similar pre-Cambrian igneous rocks of southern Pennsylvania and Maryland. The alternative hypothesis seems to be that since the intrusion of the igneous rocks the Reading and Durham Hills have been but slightly affected by the influence that was so effective in producing the Paleozoic metamorphism farther south. On the whole, the irregular distribution and widespread variation in intensity of the regional metamorphism is the most striking fact brought out by the study of the Piedmont metamorphism.

CAUSES OF REGIONAL VARIATION IN METAMORPHIC INTENSITY

A glance at the map that represents the distribution of metamorphic intensity in the area studied (fig. 13) shows three parallel belts of strong, weak, and again strong metamorphism in the Piedmont province north and northeast of Baltimore. The southern belt of strongly metamorphosed rocks extends northeastward from the neighborhood of Baltimore beyond the limits of the map (fig. 13) as far as Trenton. It attains its maximum width of about 18 miles north of Baltimore and becomes somewhat narrower toward Philadelphia, where it is scarcely more than 10 miles wide. In this belt of strong metamorphism there are certain isolated areas where the intensity reaches a maximum—namely, in the anticlinal uplifts that bring to the surface the oldest pre-Cambrian rocks. Northwest of this belt of great intensity there lies another narrower belt in which the metamorphism, though distinct, is of low intensity and even in some localities incomplete—for example, in the area underlain by Peach Bottom slate, which occupies the middle of the belt. Still farther northwest metamorphic intensity once more increases, attaining a maximum along a line that corresponds to the axis of the Mine Ridge-Tucquan anticline.

These three belts of differing intensity may be designated for convenience the Baltimore-Philadelphia belt, the Peach Bottom belt, and the Mine Ridge belt. The metamorphism varies considerably in the Mine Ridge belt along the direction of the northeast strike. Metamorphism along the strike of the Tucquan anticline in Maryland and in Pennsylvania as far northeast as the Quarryville Valley is complete but of moderate intensity. But from Quarryville northeastward along the axial plane of the Mine Ridge anticline the metamorphism, even in the youngest Paleozoic rocks, is once more intense and of a grade comparable to the metamorphism in the oldest rocks in the Baltimore-Philadelphia belt. An exhaustive and detailed metamorphic study will be required, both in the region comprised in the compound anticline that underlies the Honeybrook upland and in the region that lies west of the Tucquan anticline in southern

Pennsylvania and in Maryland, before the local variations in metamorphic intensity can be platted. But the data now available justify the generalization that north and west of the Mine Ridge upland the metamorphism wanes at first rapidly and then gradually until the metamorphosed Paleozoic strata disappear under practically unaltered Triassic rocks.

Where Paleozoic and pre-Cambrian rocks reappear in the area of the Reading and Durham Hills, north of the Triassic belt, intense metamorphism is confined to the oldest pre-Cambrian rocks. The moderately metamorphosed Paleozoic strata of this area nowhere attain the degree of alteration shown by the Paleozoic schists of the Mine Ridge upland, nor do the pre-Cambrian intrusive rocks of this area show as much metamorphism as the pre-Cambrian intrusive rocks of the Mine Ridge upland.

CAUSES OF REGIONAL METAMORPHISM

Regional metamorphism may be grouped under three heads according to the dominant causal factor—dynamic metamorphism, in which the dominant influence is nonuniform shearing stress; load metamorphism, in which the causal factor is static pressure under conditions of high temperature; and regional contact metamorphism, produced under the influence of increased temperature caused by batholithic intrusion.

Apparently the regional metamorphism of the Piedmont province does not show a consistent causal relation to either the structure or the age of the rocks involved. In some places rocks on the limbs of anticlines are highly metamorphosed, as might be expected from their structural position; in other places rocks occupying the limbs of folds are only moderately or incompletely metamorphosed, and rocks in the center of an appressed syncline have not advanced beyond the condition of slates, whereas rocks in the arches of open anticlines are completely metamorphosed with a moderate or strong degree of intensity. Likewise in some places pre-Cambrian rocks are phyllites or slates, and in others Ordovician rocks are calcareous mica schists. Rocks that are equally susceptible to metamorphism and that have been exposed to equal intensity of dynamic deformation show more metamorphism in one place than in another. Rocks of similar composition, supposedly deformed under the same load, are more intensely altered in one place than in another.

These peculiar localizations of metamorphic intensity are at first sight extremely puzzling. Regional contact metamorphism is suggested in certain areas because rocks that have been intruded by igneous material are intensely metamorphosed, whereas in areas where igneous rocks are not exposed there is a decrease in intensity.

However, in certain other areas where no igneous rocks are exposed the intensity of metamorphism is fully as great as in areas exhibiting large igneous intrusions. Therefore, if regional contact metamorphism is to be invoked to explain the intense metamorphism in an area that is lacking in igneous outcrops, it is necessary to fall back upon the hypothesis that the area is underlain by an unexposed batholith of sufficient size to produce the observed result, an idea that has been prevalent in geologic literature since the days of Michel Levy and Termier.

Evidently the puzzling variations in metamorphism in the Piedmont province can not be explained by any one alone of the three dominant causes of metamorphism. However, regional metamorphism has taken place not once but repeatedly and therefore under the influence of different factors or combinations of factors at different times. The solution to the problem lies in the correct interpretation of the whole sequence of metamorphic changes during the pre-Cambrian and Paleozoic history of the region.

DYNAMIC AND CONTACT METAMORPHISM IN THE OLDEST PRE-CAMBRIAN ROCKS

One generalization may be made in respect to stratigraphic horizon. Wherever the oldest pre-Cambrian basement gneiss is exposed it shows the maximum intensity of metamorphism. It is probable that the alteration of the Baltimore gneiss has been produced by a combination of both contact and dynamic action, because the presence of injection gneisses and of granitic intrusions in the Baltimore gneiss shows the influence of an active magma and because the structural discordance between the Baltimore gneiss and the overlying Setters formation shows that folding had taken place before the deposition of the Setters. It is impossible to tell how far the pre-Glenarm alteration of the Baltimore may have been intensified by later anamorphism. The strong metamorphic intensity in these oldest basement rocks suggests a possible cumulative effect of later metamorphism, but in view of the fact that repeated metamorphism is often retrogressive rather than cumulative in its effects, and also in view of the fact that the microscope shows the Baltimore gneiss to have yielded in some degree to a cataclastic action which seems to have been connected with the pre-Setters igneous activity and whose effects have not been obliterated by subsequent regional metamorphisms, it is possible that the Baltimore gneiss remained an inert mass so far as later metamorphosing influences were concerned.

The intrusion of an active magma in the Baltimore gneiss would insure a reconstruction of the rock in connection with the folding, although the absence of intense metamorphism in certain strongly folded regions like the Appalachian shows that folding alone is not necessarily sufficient to cause crystalloblastic metamorphism.

LOAD METAMORPHISM AND REGIONAL CONTACT METAMORPHISM IN THE GLENARM SERIES

It has been pointed out that the metamorphism of the Glenarm series shows a relation to the folding that is exactly the reverse of what would be expected if the alteration had been caused by the folding alone. The cores of open anticlines show a metamorphic intensity that is not attained on the limbs and that far surpasses the intensity in the trough of a tightly compressed syncline.

On the other hand, the relation of the metamorphism to the stratigraphic horizon in the Glenarm series is such as might be expected if the load had been the causal factor, for the lowermost formation of the series is a biotite schist and the uppermost beds are slates. Thus in so far as the pre-Cambrian rocks of the region are concerned the stratigraphic horizon might be construed as the dominant factor in their transformation into schists. On the hypothesis of load metamorphism, however, it is not obvious why the Wissahickon formation should occur in two metamorphic facies that differ so widely in intensity as the albite-chlorite schist and the oligoclase-mica schist. A possible explanation of this difference in mineralogic facies may be the presence of large bodies of plutonic rocks along the southeastern border of the Piedmont province, where the oligoclase-mica schist occurs, and the absence of similar rocks in the area underlain by albite-chlorite schist. The contact effects of these intrusions on the oligoclase-mica schist facies of the Wissahickon must have been appreciable and may have amounted to regional contact metamorphism.

If, as seems probable, the post-Glenarm igneous intrusions of Maryland are pre-Paleozoic, such regional contact metamorphism must have antedated the overthrust of the Glenarm series upon the Paleozoic rocks. Therefore, if the combined influence of load and batholithic intrusion was sufficient to produce metamorphism of the Glenarm series as we now see it, that metamorphism must have been pre-Paleozoic and earlier than the overthrusting. However, the area underlain by Glenarm schists was affected by a Paleozoic metamorphism, because the igneous intrusions invoked as a possible cause of regional contact metamorphism in the Glenarm series are themselves metamorphosed in areas where the Paleozoic strata are schists and only slightly altered in areas where the Paleozoic rocks are less metamorphosed, thus suggesting that the metamorphism of the Paleozoic beds was contemporaneous with the alteration of post-Glenarm igneous rocks.

DOUBLE METAMORPHISM IN THE GLENARM SERIES

The Glenarm schists in the overthrust block show two foliations—an older, more strongly developed foliation that is parallel to the bedding and a younger schistosity that makes an angle with the bedding. Likewise, the Glenarm schists are cut by two systems of quartz veins, and the older veins, which are parallel to the bedding, are folded. The relations of these two metamorphisms to the metamorphism in the Paleozoic rocks should throw some light upon the age of the metamorphism in the Glenarm series.

VARIATION IN METAMORPHIC INTENSITY IN THE PALEOZOIC ROCKS

The intense local metamorphism of the Lower Cambrian sediments that flank the Mine Ridge anticline is the most surprising feature of the metamorphic history of the region. Metamorphism due to the normal load of the overlying stratigraphic series can hardly be invoked as a cause, because on the slopes of Welsh Mountain, north of the Mine Ridge upland, there are Lower Cambrian rocks at the same stratigraphic horizon as the strata in the Mine Ridge upland, yet the Paleozoic beds of Welsh Mountain are much less metamorphosed than the Paleozoic beds of the Mine Ridge upland, although the load of the normal stratigraphic series in both places must have been approximately the same. Dynamic metamorphism operative over a large area is at first sight equally inadequate to account for this peculiar localization of intensity, for any regional deformation under strong tangential stress that would have reconstructed Paleozoic rocks in the Mine Ridge metamorphic belt could hardly have left closely folded rocks farther southeast, in the Peach Bottom belt, in the condition of slates.

A possible explanation for the puzzling localization of metamorphism is suggested by the idea that intense metamorphism is produced by frustrated movement rather than by free movement of the rock masses involved, an idea that was employed by Greenly to explain the regional metamorphism on the island of Anglesey,[7] where there are three superimposed recumbent folds of great amplitude. The metamorphism in this tectonic series shows two maxima which are not related to the recumbent folding, because the conditions necessary for molecular readjustment do not exist in freedom of movement. They are ascribed rather to a frustration of move-

[7] Greenly, Edward, Geology of Anglesey: Great Britain Geol. Survey Mem., p. 908, 1919.

ment, which takes place when the limit of free horizontal movement is reached in any one recumbent fold and which is expressed in a buckling of the recumbent folds into close secondary folds. If the core of the Mine Ridge anticline represents a buttress of pre-Cambrian igneous rocks and crystalline schists it is possible that there was a buckling of the Paleozoic sediments that was locally intensified and concentrated around the resistant buttress, waning gradually as the distance from the unyielding core became greater. A serious objection to this hypothesis is the fact that the pre-Cambrian buttress underlies the whole Honeybrook upland and is not confined to the Mine Ridge upland. The metamorphism of the Paleozoic strata, on the other hand, is confined to the Mine Ridge upland and is not uniform around the entire resistant core.

DOUBLE METAMORPHISM IN THE PALEOZOIC ROCKS

The Paleozoic rocks, like the Glenarm schists, show the influence of two metamorphisms. On the east flank of Bunker Hill the dominant metamorphism of the rock corresponds to the older schistosity, which is parallel to the folded bedding. A younger metamorphism whose effects are distinctly subordinate to the first is shown in a cleavage that cuts across the older schistosity. (See pl. 8, *A*.) The older schistosity and folding is cut across by a system of younger quartz veinlets that are presumably contemporaneous with the younger schistosity. In addition to these younger quartz veinlets there is an older vein system that may be genetically associated with the pegmatites which are so widespread throughout the Paleozoic rocks in the region. A vein of blue quartz, cutting the injected hornblende schist north of Coatesville, is folded along with the including rock, thereby showing the influence of an orogenic period that was later than the igneous epoch during which the vein was formed. The Paleozoic pegmatites themselves show evidence of strong dynamic deformation. The plagioclases are saussuritized, the potash feldspar is microcline, and the quartz is granulated. The effects of this second deformation are particularly conspicuous in the anticlines south of Columbia that bring Lower Cambrian argillaceous beds to the surface. Jonas has found that the banded slates in this locality show a well-developed cleavage that crosses the bedding. There is also a distinct but weak metamorphism that has produced an earlier foliation parallel to the gently folded bedding. (See pl. 8, *B*.)

In the older of these two metamorphisms the intensity has waned perceptibly in a distance of 3 to 4 miles across the strike of the formation. The intensity of the younger metamorphism is practically constant in the same distance and has only produced a slaty cleavage. But the younger foliation is less conspicuous in the strongly metamorphosed schist from Bunker Hill than in the less altered rock near Columbia, thereby illustrating the fact that a second metamorphism is comparatively ineffectual when superimposed upon an already highly metamorphosed rock.

DEVELOPMENT OF OLDER METAMORPHISM IN THE GLENARM SERIES AND IN PALEOZOIC ROCKS

As noted above the marked characteristic of the later metamorphism in both Glenarm schists and Paleozoic strata is its uniform and low intensity over a wide area. In the same area the earlier metamorphism shows a wide range in intensity. The intensity of this older metamorphism is at a maximum in the Paleozoic schists on the flanks of Mine Ridge anticline and in the immediately surrounding neighborhood. It wanes northward very rapidly in the Paleozoic rocks and wanes southward less rapidly but perceptibly in the pre-Cambrian schists of the overthrust block. The puzzling feature of the metamorphic history of the area is the age of the earlier metamorphism and the cause of its variation in intensity.

The variation in metamorphic intensity in the Glenarm series has already been discussed, and it is obvious that there are logical objections to ascribing the metamorphism of the series as a whole to dynamic metamorphism accomplished under the influence of non-uniform pressure or to load metamorphism under the influence of uniform pressure. The influence of large igneous intrusions furnishes a possible explanation of the concentration of metamorphic intensity in some areas, and it is possible that the part of the Glenarm section that was beyond the thermal influence of these intrusions may have acquired a certain amount of load metamorphism.

The metamorphic gradient in the overriden and overriding blocks strongly suggests that the metamorphism now seen in the Glenarm section along Susquehanna River was not altogether developed before the thrusting. A waning of intensity between the overridden Paleozoic rocks and the overthrust pre-Cambrian beds is a marked contrast to the metamorphic gradient in many overthrusts involving pre-Cambrian rocks where highly altered older rocks have been shoved over slightly metamorphosed younger formations.

An a priori reason for assuming that a metamorphism of the faulted block accompanied or succeeded the thrusting is the notable absence, so far as observed, of strong brecciation along the thrust plane. The inference is that the thrusting took place under conditions favorable for molecular reconstitution during and subsequent to the movement.

The question then arises, under what conditions could such molecular rearrangement take place? The work of Adams and Bancroft[8] suggests that "decken" are probably formed only in the upper part of the zone of flow, beneath the zone of fracture, where adjustment takes place by means of faults and overthrusts. Great movements of adjustment, therefore, must take place at or comparatively near the surface. The experimental work of Adams[9] would suggest that granite rocks do not flow at depths of less than 11 miles below the surface of the earth's crust and that even limestones do not yield to deformation at depths of less than 9 miles, provided the temperatures at these depths are such as would be expected from the normal temperature gradient. Nevertheless, plastic deformation has frequently taken place under a load of sediments that is much less than the load that would furnish the pressure required by the experiments of Adams.

The intense Paleozoic metamorphism in the Quarryville district is apparently another illustration of the difficulty in reconciling the results of experimental work with the available geologic evidence. The geologic evidence indicates a load upon the Paleozoic rocks in the McCalls Ferry-Quarryville district that did not exceed 1 or 2 miles at the time that the sediments were folded. This load was certainly less than the weight of sediments under which the Cambrian and Ordovician rocks of the Lebanon Valley must have been deformed, because the strata in the Lebanon Valley were deposited in a deeper part of the geosyncline than the rocks of the Quarryville Valley. And yet the metamorphism in the Lebanon Valley is very slight in contrast to the Paleozoic metamorphism around the Mine Ridge upland. Moreover, the overthrusts of southern Nevada, which have taken place under a load that is of the order of magnitude of 2 miles, show all the effects of shattering and brecciation that would be associated with a light load, whereas the Martic overthrust, which is estimated to have taken place under a maximum load of less than 2 miles, shows effects that would be attributed to formation at great depth. Of course, if the Paleozoic metamorphism took place in the course of the earth movement which accompanied or followed the overthrusts and during which the thrust plane was folded, the load under which the alteration was accomplished would have been locally increased by an unknown thickness in the overriding block.

According to the experimental indications that the zone of flow lies at a depth of 11 miles, the overriding block must have been at least

[8] Adams, F. D., and Bancroft, J. A., On the amount of internal friction developed in rocks during deformation and on the relative plasticity of different types of rocks: Jour. Geology, vol. 25, p. 635, 1917.

[9] Adams, F. D., An experimental contribution to the question of the depth of the zone of flow in the earth's crust: Jour. Geology, vol. 20, pp. 97–118, 1912.

9 miles thick in order to furnish conditions of molecular rearrangement along the plane of the overthrust. At such a depth rock transfer would be expected to take place by the rolling out of great recumbent folds or "decken" rather than by a translation of a large block of the earth's crust along a fracture plane.

Moreover, Lawson[10] has calculated the limiting value for the length of the overthrust prism along different inclinations of the fault plane under different assumed coefficients of friction. Figure 14 is his diagram of a cross section through a wedge-shaped thrust block. *ab* represents the length of the overthrust block, *ac* the maximum depth of the overthrust block, and *bc* the horizontal displacement of the thrust. The angle *abc* is the angle made by the plane of the overthrust with the surface. Lawson's table is given below showing the values of *ab* that he calculated for several assumed values of *abc* and of the coefficient of friction (*f*).

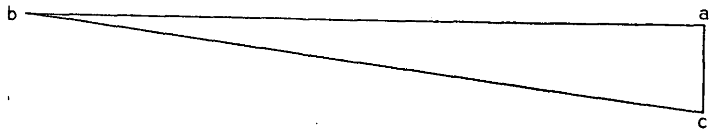


FIGURE 14.—Cross section through a wedge-shaped thrust block. (After A. C. Lawson)

Limiting values for length of overthrust block, ab, in miles

abc	*f*=0.5	*f*=0.4	*f*=0.3
2°	20	25	32
5°	18	22	26
10°	16	19	23
20°	12	14	16

The corresponding values for the depth of the overthrust block have been calculated and are given in the table below.

Limiting values for the depth of the overthrust block, ac, in miles

abc	*f*=0.5	*f*=0.4	*f*=0.3
2°	0.7	0.9	1.1
5°	1.5	1.9	2.3
10°	2.8	3.4	4.5
20°	4.4	5.5	5.6

This table would indicate that the maximum depth of the overthrust block along such a thrust as the Martic overthrust, the angle of which was doubtless less than 10°, would not amount to 6 miles and

[10] Lawson, A. C., Isostatic compensation considered as a cause of thrusting: Geol. Soc. America Bull., vol. 32, p. 341, 1922.

probably not even to 3 miles. Therefore it is hardly likely that the added load of the overriding block in itself was sufficient to cause deformation under conditions of flowage without the introduction of some other factor of molecular reconstruction. It would seem more probable that the deformation originated as a fracture, along which the displaced block was driven toward the northwest, and that, either during the movement or immediately subsequent to it, some factor other than depth changed the environment of the fault from the zone of fracture to that of flowage.

There remain two factors that would affect the depth at which the observed deformation and folding may have taken place—namely, time and temperature. It is undoubtedly possible that a comparatively low pressure exerted continuously over a long period of time may furnish conditions of deformation strictly analogous to that reproduced in Adams's laboratory under enormously higher pressure. A long-continued and relatively slight pressure might have produced the high metamorphism of the Mine Ridge belt, but it would not account for the weak metamorphism of the near-by rocks.

The temperature factor thus appears to be the only variable by which differences in intensity can be explained, and the raising of the geothermal gradient by batholithic invasion seems to be the only possible explanation for so peculiar a localization of high-temperature effects. If it is assumed that the metamorphism of the Paleozoic strata and possibly of the overthrust pre-Cambrian beds took place at or near the end of the Paleozoic era, the load under which a batholith might have been intruded would be represented by about 8,000 feet of Paleozoic strata in the Lancaster Valley plus a possible additional overthrust cover of 1 to 2 miles in thickness. At this depth of about 3 miles the normal temperature gradient would give a temperature of about 170° C., which is presumably inadequate to account for a coarse-grained crystalloblastic metamorphism of a high order of intensity.

But the presence of highly heated igneous bodies at the comparatively shallow depth of 2 miles below the surface would easily raise the temperature gradient to an amount that would be sufficient to account for all the puzzling features of the peculiarly localized variations in metamorphic intensity.

Such a hypothesis is theoretically possible and therefore attractive, but it is peculiarly dangerous in that it is inherently insusceptible of proof, the igneous rock that is supposed to cause the effect being unexposed at the surface and therefore a matter of speculative origin. The theory was suggested by Barrell[11] to account for the metamorphism in the highlands of western Connecticut, where segre-

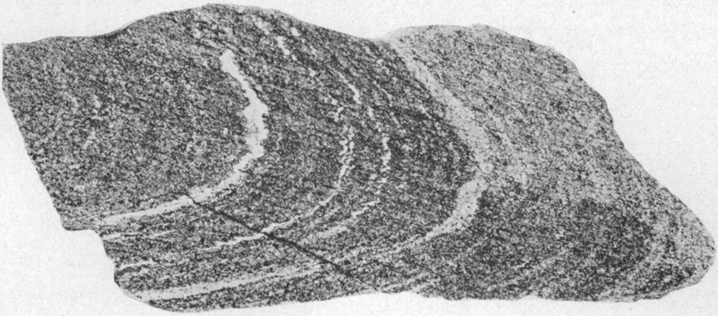
[11] Barrell, Joseph, Relation of subjacent igneous invasion to regional metamorphism: Am. Jour. Sci., 5th ser., vol. 1, pp. 1–19, 174–264, 1921.

gation seams and lenses of tourmaliniferous quartz and feldspar are abundant. He also accounted for the lack of metamorphism in the Lebanon Valley of Pennsylvania, as compared with high metamorphism in the Connecticut highlands, by the absence of evidences of igneous intrusion in the Pennsylvania locality. There is some evidence for the presence of underlying batholithic intrusion along the Mine Ridge metamorphic belt, because the Paleozoic rocks are penetrated by pegmatites, many of which were accompanied by ore-bearing solutions. Tourmaline is of abundant and widespread occurrence even as high in the section as the Conestoga limestone. Many of the tourmaline crystals have been broken and pulled apart along the foliation planes. (See pl. 8, *C*.) The spaces between the stretched fragments are filled with quartz that has apparently accompanied the boron vapors from which the tourmaline was deposited. As pointed out by Barrell, the presence of tourmaline is not an infallible evidence of igneous activity, yet its widespread occurrence, even in the calcareous rocks, in rosettes and nests that are clearly of epigenetic origin is a reasonable indication that the Paleozoic rocks in the Chester, Lancaster, and York Valleys have been penetrated by emanations from an underlying batholithic intrusion. Whether the presence of such a batholith is sufficient to account for the strong metamorphism of the Mine Ridge belt is a pure speculation unless it proves possible to work out in the field some relationship between the intensity and extent of the visible effects of igneous activity, such as pegmatitization and tourmalinization, and the amount and intensity of metamorphism.

RELATIVE AGE OF DOUBLE METAMORPHISM IN THE GLENARM SERIES AND IN THE PALEOZOIC ROCKS

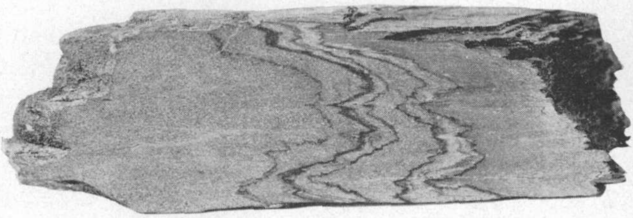
There is a striking parallelism between the double metamorphism that is found in the Glenarm and the Paleozoic schists, but a most exhaustive study of the metamorphism in the two areas will be required in order to prove that they are contemporaneous. The earlier metamorphism of the Glenarm series may be pre-Cambrian, and the younger cleavage may be the only record in the Glenarm schists of the first Paleozoic metamorphism. Metamorphism would doubtless register its effects with great intensity upon the unaltered Paleozoic rocks but would not necessarily be so effective in altering the Glenarm rocks if they had been already transformed to schists in pre-Cambrian time.

On the other hand, if the pre-Cambrian metamorphism in the Glenarm series was only slight or local and confined to the neighborhood of the epi-Glenarm intrusions, the metamorphism in the Wissahickon albite-chlorite schist may be largely Paleozoic and of the same



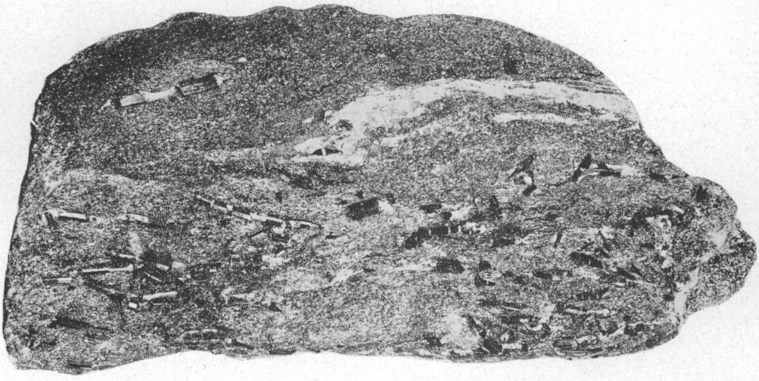
A. SPECIMEN OF HARPERS SCHIST

Shows older folded schistosity parallel to bedding cut by younger cleavage inclined to bedding



B. SPECIMEN OF CHICKIES BANDED SLATE

Shows older folded schistosity parallel to bedding cut by younger cleavage inclined to bedding



C. SPECIMEN OF MICA SCHIST FROM UPPER BEDS OF CHICKIES QUARTZITE

Shows stretched epigenetic tourmaline

age as the earlier metamorphism that is so strongly developed in the Harpers schist around the Mine Ridge uplift. If the earlier metamorphism in the Paleozoic and pre-Cambrian section is of the same age and was associated with or subsequent to the overthrusting, the abnormal metamorphic gradient in the area of the overthrust becomes intelligible, for whether the cause of the metamorphism is uniform pressure or the thermal influence of an underlying batholith, the rocks that lie the deepest in the earth's crust would be the most intensely altered.

SUMMARY OF CONCLUSIONS

The variable metamorphic intensity in the rocks of the Piedmont province is the result of a variety of causes that have operated at different times. The relative intensity of metamorphism of different formations is no clue to their stratigraphic horizon, and there is no simple relation between the degree of metamorphism and geographic position.

The pre-Cambrian basement complex was doubtless formed during the pre-Glenarm folding as a result of dynamic metamorphism intensified by the thermal influence of batholithic invasion. The metamorphism of the early pre-Cambrian sediments is everywhere intense but shows no evidence that it is a cumulative effect produced during superposed metamorphisms. On the contrary, it suggests that the resultant crystalline schists have remained more or less inert since their formation, owing doubtless to their prolonged deep-seated position, which has preserved them from retrogressive metamorphism.

The later pre-Cambrian formations (Glenarm series) show a strong degree of metamorphic intensity in the schists that form the base of the section and a progressive waning of intensity toward the slates at the top. Load metamorphism is, however, inadequate to explain the occurrence of the same formation in two metamorphic facies of different intensity, because the facies were presumably at the same depth when they were formed. The rocks of the intensely metamorphosed facies have been invaded by numerous widespread plutonic intrusions, and these intrusions may have produced regional contact metamorphism.

The metamorphism of the Paleozoic sediments in general is of low to moderate intensity, but the Paleozoic strata that emerge from under the overriding fault block in the neighborhood of the Mine Ridge anticline show a puzzling development of abnormally strong intensity. The metamorphic gradient in the underlying and overlying blocks wanes from bottom to top, and this abnormal relation of strong metamorphic intensity in the Paleozoic rocks, contrasted with lower intensity in the pre-Cambrian thrust block, suggests that the dominant metamorphism in both Paleozoic and Glenarm strata was subsequent to the overthrust. This idea is confirmed by the character of the thrust plane so far as observation of it has been possible. The thrusting appears to have taken place under conditions that were favorable to molecular rearrangement and metamorphism.

The strong degree of metamorphic intensity localized in the autochthonous Paleozoic beds near the surface emergence of the fault plane suggests that the increased load produced by the weight of the thrust block may have had some influence in the metamorphism. But the conditions for crystalloblastic alteration could hardly have been produced by depth of burial alone. The normal load of sediments upon the Conestoga formation at the time of metamorphism could hardly have amounted to as much as 10,000 feet. Even if this normal load were considerably increased by the weight of the thrust block the additional load could not have been sufficient to cause anything but a cataclastic metamorphism in the autochthonous Paleozoic sediments or in the displaced thrust block.

The only feasible explanation for a folding and metamorphism of the thrust block that would inhibit or obliterate the effects of cataclastic metamorphism is a rise of geothermal gradient. A localized rise of geothermal gradient suggests the influence of highly heated underlying igneous rock, and such thermal influence combined with the locally increased load of the faulted rock would doubtless account for the peculiar concentration of intense metamorphism in the overridden Paleozoic rocks. The hypothesis is alluring, but has been insusceptible of proof except in so far as the extensive tourmalinization and pegmatitization of the Paleozoic rocks indicate the passage of magmatic emanations from an underlying igneous intrusion.

There has been a double metamorphism in both the Glenarm series and the Paleozoic strata, which is shown by foliation that is parallel to the bedding and cut across by a younger slaty cleavage. The older foliation has a considerable range in intensity in the Paleozoic rocks, being developed most strongly near the Mine Ridge anticline. A short distance north of the anticlinal axis of the Mine Ridge uplift the intensity of the older metamorphism diminishes and the younger cleavage becomes more noticeable. The younger metamorphism was subordinate in degree to the older, and is therefore conspicuous only when the older metamorphism was at a minimum. The older foliation was post-Conestoga and presumably later than the overthrust or possibly contemporaneous with the overthrusting, because of the absence of cataclastic effects along the thrust plane and in the overriding block. The younger foliation may have coincided with the last Paleozoic deformation of the region, but so far it is impossible to fix the dates of these two periods of metamorphism more closely than to say that they occurred in the Paleozoic era after Chazy time.

ECONOMIC GEOLOGY

NICKEL ORE

GAP MINE

The Gap nickel mine is on Mine Ridge, about 12 miles southeast of Lancaster and 3½ miles south of Kinzers on the Philadelphia division of the Pennsylvania Railroad. The nickel ore occurs in places along the border of a small gabbro mass that intrudes the early pre-Cambrian schists. Pyrrhotite and chalcopyrite with a small amount of pyrite make up the bulk of the metallic sulphides. The ore minerals are pentlandite, which is intergrown with pyrrhotite, and millerite. The millerite has been formed by downward enrichment of the primary sulphides.

The presence of ore was noted apparently as early as 1718 in the copper-bearing waters of a mineral spring that issued from the ground at the place where the mine was afterward established. For 80 to 90 years the Gap mine was worked at intervals for copper, but the operations proved unsuccessful, and the mine had stood idle for about 30 years, when it was reopened in 1849, still as a copper-mining enterprise. At that time the nickel ore was being discarded on the dump as refuse, called by the miners "mundic," the Cornish word for pyrite. In 1852 a miner named Doble, who afterward became superintendent of the mine, suggested that some other mineral than mundic was present in the discarded material. An analysis by Genth finally showed the presence of nickel in the so-called mundic, and the Gap Mining Co. then devoted its attention to the mining of nickel ore. The ore was smelted in a preliminary way about a mile north of the mine. The operations were not financially successful until 1862, when the mine and smelter were purchased by a Philadelphian named Joseph A. Wharton, who established a nickel refinery, known as the American Nickel Works, in Camden, N. J., where the matte produced in the Gap smelter was refined and manufactured. The ore as it left the mines contained from 1 to 3 per cent of nickel.[12]

In 1877 the average annual production of the Gap mine was estimated at 7,200 short tons of ore, carrying from 1 to 3 per cent of nickel.[13] At that time the largest producer of nickel in the world

[12] This account is taken from a mine report prepared by Charles Doble, superintendent of the Gap nickel mines, quoted by Frazer, Persifor, jr., The geology of Lancaster County: Pennsylvania Second Geol. Survey Rept. CCC, pp. 163–168, 1880.

[13] Idem, p. 174.

was Norway, whose maximum annual output in 1876 was 42,500 tons of ore, or 360 tons of nickel. The Gap mine at this time produced about one-sixth of the total annual nickel output of the world. But in 1877, when New Caledonia exported 8,000 tons of ore averaging from 8 to 10 per cent of nickel, the Pennsylvania nickel industry received a severe blow, and the advent of the Sudbury nickel ores in 1887 proved a death stroke to the Gap mine, which was closed in 1893.

COMPARISON WITH SUDBURY ORES

The Gap nickel ores are similar in character and occurrence to the Sudbury ores, and the problem of their genesis presents the same puzzling features that have caused a geologic controversy of many years' standing in the Sudbury district. Four theories have been advanced to account for the nickeliferous pyrrhotite deposits of Ontario—(1) that the ores have originated in a laccolithic chamber as magmatic segregations of heavy molten sulphides from a silicate melt;[14] (2) that the ores have been introduced by thermal waters that have no necessary genetic relationship to the norite mass in which the pyrrhotite occurs;[15] (3) that the ores were produced by magmatic differentiation into fluid sulphides but that the differentiation took place in the magma reservoir, not in the laccolithic chamber, and that the resultant liquid sulphides were injected upward into the laccolithic chamber;[16] (4) that the ores are of magmatic origin but that their consolidation has taken place at a late magmatic stage, when the temperature was so far reduced that the sulphides could not have existed in the fluid condition unless mineralizers were present in such large proportions that "the characteristics of the mixture are those of a high-temperature solution and not of a melt."[17]

Nevertheless, after almost 40 years of work on the origin of these ores, the question still remains to a large extent open. The answer is probably, as suggested by Bateman, that the ores as they exist to-day have been formed by a combination of magmatic segregation, injection of molten sulphides, and hydrothermal action.[18] Even so confirmed a believer in the segregated origin of the nickel ores as

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[14] Coleman, A. P., The Sudbury nickel field: Ontario Bur. Mines Rept., vol. 14, pt. 3, 1905; The Sudbury laccolithic sheet: Jour. Geology, vol. 15, pp. 759–782, 1907.

[15] Dickson, C. W., The ore deposits of Sudbury, Ontario: Am. Inst. Min. Eng. Trans., vol. 34, pp. 1–65, 1903. Campbell, William, and Knight, C. W., On the microstructure of nickeliferous pyrrhotites: Econ. Geology, vol. 2, pp. 350–366, 1907.

[16] Howe, Ernest, Petrographical notes on the Sudbury nickel deposits: Econ. Geology, vol. 9, pp. 505–522, 1914.

[17] Tolman, C. F., and Rogers, A. F., A study of the magmatic sulfid ores: Leland Stanford, Jr., Univ. Pub., Univ. series, p. 16, 1916.

[18] Bateman, A. M., Magmatic ore deposits, Sudbury, Ontario: Econ. Geology, vol. 12, pp. 391–426, 1917.

Coleman[19] states that these different processes have all operated in the Sudbury region. But on the assumption that all these processes were involved in the formation of the ores, the extent to which each one contributed to the formation of the valuable ore bodies is still a matter of doubt. Coleman takes exception to Bateman's idea that the injection of molten sulphides was responsible for the greater part of the ores and states that in the main they are due to magmatic segregation in place.[20]

OCCURRENCE OF THE GAP ORE

The discussions on the genesis of the Sudbury ores are, of course, greatly aided by the fact that the surface outcrops of the ore bodies and of their associated rocks have been carefully mapped in much detail. Moreover, the underground development of the numerous mines has given and still gives ample opportunity to study the structural relations of the ore body to the surrounding rock. On the other hand, students of the Gap ore are forced to rely entirely upon the old mine reports for the structure and field relations of the ore body, because the workings have been for years filled with water. It is not even possible to map with accuracy the intrusive mass with which the ore is associated, because the area in which it occurs is notably lacking in surface outcrops. Therefore, there is practically nothing new to be gained by a field study of the Gap ore body, and hypotheses for its genesis must be based upon field observations made in former years, which are scanty, and upon petrographic study of the ores and to a considerable extent upon a general knowledge of the stratigraphic and structural relations of the ores and the including rocks. The only geologic description of the ore body was given by Kemp,[21] who visited the mine in April, 1894, shortly after it had been closed. The map and sections of the ore body shown in Figure 15 have been taken from Kemp's paper, which, together with the old mine reports,[22] has furnished the data for the following description.

The ore occurs in places around the eastern, southeastern, and northeastern margins of a more or less lenticular mass of metamorphosed gabbro or norite. The gabbro mass is approximately 2,400 feet in length and between 300 and 500 feet in width, with a sharp

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[19] Coleman, A. P., Magmas and sulphide ores: Econ. Geology, vol. 12, pp. 427–434, 1917; Geology of the Sudbury nickel deposits: Idem, vol. 19, pp. 570, 574, 1924.

[20] Coleman, A. P., Geology of the Sudbury nickel deposits: Econ. Geology, vol. 19, p. 574, 1924.

[21] Kemp, J. F., The nickel mine at Lancaster Gap, Pa., and the pyrrhotite deposit at Anthony's Nose, on the Hudson: Am. Inst. Min. Eng. Trans., vol. 24, pp. 620–633, 883–888, 1895.

[22] Frazer, Persifor, jr., Geology of Lancascer County: Pennsylvania Second Geol. Survey Rept. CCC, pp. 163–176, 1880.

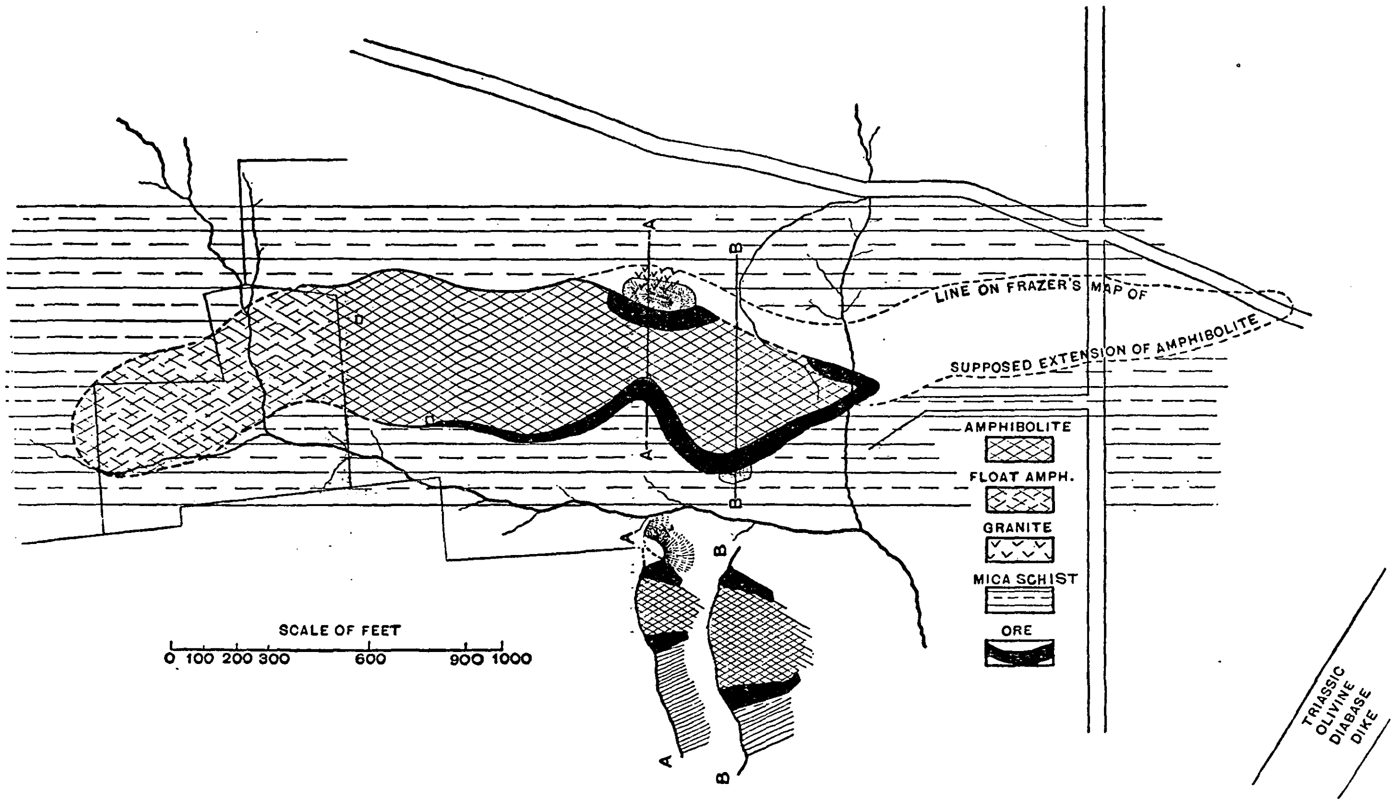

FIGURE 15.—Geologic map and cross sections of the Gap nickel mine. (From Kemp, J. F., Ore deposits of the United States, 2d ed., fig. 159, 1895)

constriction toward the east end. The ore body ranged in thickness from 4 to 35 feet. It was apparently confined to the east end of the gabbro intrusion and occurred along the contact of the gabbro and the inclosing mica schist. The least depth at which ore was found is 14 feet below the surface, and in many places the barren and much decomposed upper zone extended down to a depth of 60 feet. The ore body, which was nearly vertical, was worked to a depth of 250 feet before mining ceased. Kemp [23] states that the ore occurred to some extent in bunches within the lens, but most of the productive material was near the contact. Frazer notes the presence of detached masses of hornblende (the local name for the basic intrusion) within the ore, and Hess [24] mentions the occurrence of orbicules of hornblende in the pyrrhotite masses.

The available field evidence does not show whether the lenticular intrusion is a dike or a laccolith. The dip of the contact between the intrusion and the inclosing rock was apparently fairly steep on both the north and the south sides of the intrusion. Therefore, if the intrusion was laccolithic it must have been sharply tilted subsequent to its formation.

The igneous intrusive rock with which the ore is associated has been described by Kemp as an amphibolite in which the green hornblende is secondary, indicating that the rock is derived from a gabbro, norite, or pyroxenite. He describes the change to hornblende as so complete that a careful search was required to reveal any trace of the original pyroxene. Recently Phemister [25] described the intrusive rock as a gabbro containing large amounts of pyrogenetic hornblende and secondary biotite with subordinate amounts of quartz, calcite, and plagioclase.

Several specimens that were collected by the writers from the mine dump show both a massive and a foliated basic rock. The massive variety is an equigranular rock of igneous texture composed of hornblende and feldspar with a small amount of metallic sulphides scattered throughout. Under the microscope it is seen to consist roughly half of feldspar and half of hornblende and biotite. The feldspar, which was apparently a fairly calcic plagioclase, has been so thoroughly saussuritized that it is difficult to determine the original species. Float from a locality about 3 miles east of the nickel mine, which is apparently identical with the rock at the mine, shows a less altered feldspar, which was determined as $Ab_{64}An_{36}$. A large part of the green hornblende shows absolutely no trace of secondary

[23] Kemp, J. F., op. cit., p. 627.
[24] Hess, F. L., U. S. Geol. Survey Mineral Resources, 1915, pt. 1, p. 756, 1917.
[25] Phemister, T. C., A note on the Lancaster Gap mine, Pa.: Jour. Geology, vol. 32, pp. 500–502, 1924.

origin and may well be pyrogenetic or of magmatic origin. The biotite, which is almost equal to the hornblende in amount, is of a peculiar reddish-brown color and is in some places intergrown with a deep-green clinochlore. A most striking feature of both hornblende and biotite is the prevalence of large and particularly strong pleochroic haloes that surround zircons and prismatic crystals of a deep-brown mineral that appears to be allanite. The biotite shows none of the evidences of secondary origin that were mentioned by Phemister [26] as proofs of the hydrothermal origin of the biotite in the rocks that he studied. The individual crystals are fairly large, about 0.5 to 1 millimeter in length, and are of approximately uniform grain. The direction of the amphibole cleavage does not appear to have determined the orientation of the biotite flakes, which is distinctly random in respect to both the hornblende and the other biotite individuals. Therefore there is a strong indication that at least a large part of the biotite in this gabbro is of pyrogenetic origin. The rock may be called a biotite-hornblende gabbro. Titaniferous magnetite altering to leucoxene is of frequent occurrence, as are also apatite and to a lesser extent zircon. Tourmaline is a common accessory. Quartz is subordinate in amount.

The laminated variety is a dense, finely foliated schist, which shows under the microscope practically the same constituents as the massive rock. The tourmaline lies parallel to the foliation in the rocks, having been apparently directed by the planes of foliation. The metallic sulphides have a tendency to form stringers in the direction of the foliation, and there is a suggestion in some places that they may replace the biotite. In one place an irregular mass of sulphides is entirely surrounded by biotite. As pointed out earlier in this paper, there is no proof that this foliated amphibolite is really an integral part of the intrusive body associated with the nickel ore; but its close petrographic resemblance to the coarse-grained gabbro and the prevalence of metallic sulphides, which are more abundant in the foliated than in the massive variety, together with the fact that there is marked laminated structure at the west end of the gabbro intrusion,[27] suggest that this schist is really the border facies, which has been severely squeezed along the walls.

A light-colored rock of which several specimens were collected from the dump appears to be identical with the altered quartz monzonite or granodiorite that injects the pre-Cambrian sedimentary series. These specimens, however, show some interesting features. The rock is of granitic texture and is composed almost entirely of plagioclase with a variable amount of quartz. Under the micro-

[26] Phemister, T. C., op. cit., p. 502.
[27] Kemp, J. F., op. cit., p. 626.

scope it is seen that the plagioclase is of two kinds. One, a saussuritic albite, is clearly derived from a more calcic plagioclase. The other, which is less abundant, has the optical properties of albite and shows the peculiar "chessboard" texture that is formed when a crystal of microcline is replaced by homoaxial albite.[28] The percentage of albite in the rock is extremely variable, but in those sections that show the most "chessboard" albite the microcline is scanty. A little microperthite is visible. The quartz, where present, is granulated, and the rock shows a fairly strong cataclastic deformation. Dark minerals are conspicuously absent, but there are small amounts of biotite, muscovite, and chlorite, which rather strongly suggest by their method of formation the action of high-temperature solutions. The chlorite occurs in radiating fibrous aggregates, in fibrous veins, or in stringers that traverse the rock irregularly. One specimen in which the "chessboard" albite is so abundant that the rock is practically an albitite contains an irregular mass of metallic sulphides of considerable size. This rock is probably identical with the granite shown on Kemp's map (fig. 15).

The ore minerals are pentlandite, mixed with pyrrhotite, and millerite. The millerite was found as incrustations on cracks and is evidently, as pointed out by Kemp[29] and Hess,[30] a product of downward enrichment from the leached and barren upper zone. The pentlandite is intimately intermixed with pyrrhotite, from which it can be separated only by use of the reflecting microscope. Buddington[31] has called attention to the fact that the pentlandite of the Gap nickel ores is replaced by a weakly magnetic secondary nickel mineral, which he called "nickel mineral X" to emphasize the fact that it may prove to be bravoite, polydymite, or violarite, or even a mixture of more than one mineral. Pyrrhotite is associated with chalcopyrite, the order of formation being pyrrhotite, pentlandite, and chalcopyrite.

GENESIS OF THE GAP ORE

The first theory to explain the origin of the Gap nickel ores was proposed by Kemp. In 1893 he advanced the idea that these nickel sulphides were integral parts of the magma that formed the basic igneous intrusion. The ores were separated from the magma as it cooled and concentrated along the walls in the coolest part of the solution. In this explanation of the ores Kemp applied the theory

[28] Becke, Friederich, Zur Physiographie der Germengtheile der Krystallinen Schiefer: K. Akad. Wiss., Math.-naturwiss. Klasse, Denkschr., vol. 75, p. 125, 1913.
[29] Kemp, J. F., op. cit., p. 626.
[30] Hess, F. L., op. cit., p. 756.
[31] Buddington, A. F., Alaskan nickel minerals: Econ. Geology, vol. 19, pp. 529–531, 1924.

that had just been outlined by Vogt[32] in his discussion of ores formed by magmatic segregation. Both Kemp and Vogt invoke the Soret principle, that the concentration of the dissolved substances is greatest in the coolest part of the solution, as the probable explanation of the segregation of the sulphides along the walls of the intrusion. Vogt makes it very clear, however, that the consolidation of these sulphides does not take place until after the silicates have crystallized,[33] owing to the fact that the melting point of the sulphides is much lower than that of the silicate minerals. Petrographic study of nickeliferous pyrrhotite ores from different localities has shown that the sulphides have formed after the silicates, and the metallic sulphides of the Gap ores conform to this rule, having clearly been formed later than the silicates around which the sulphides have molded themselves. This easily recognized relation of sulphides to silicates has recently been emphasized by Phemister, who believes that the Gap ores are of hydrothermal origin. In fact, he goes so far as to state[34] that this relation is the insurmountable difficulty to the igneous view of the ore genesis. The late consolidation of the sulphides is, however, no obstacle to the theory that they have been differentiated in place in the fluid condition. The pyrrhotite, pentlandite, and chalcopyrite nowhere show relations to the silicates that could not be adequately explained by the hypothesis that the sulphides were present to a late stage of crystallization as molten patches in the midst of earlier consolidated silicates. Even their presence in the mineralized schist on the north wall of the intrusion, which is noted by Phemister,[35] might be explained by the theory that in certain places the molten sulphide material penetrated for a short distance into the adjoining heated wall rock. A much more serious difficulty in the way of explaining these ores as magmatic segregations is their position along both walls of a steep-sided intrusion. Since it has been recognized that the difference in concentration between different parts of a cooling magma is not sufficient to produce an important segregation of ore the localization of sulphide ores has been ascribed chiefly to the action of gravity. In order to explain the Gap ores as a gravitative differentiate it is necessary to assume that the igneous intrusion is a small and closely folded laccolith. In other words, if the ores have assumed their present position under the influence of gravity they must have been intensely deformed after their formation.

The relation of the ores to the deformation is not, however, easy to make out. The positive proof adduced by Phemister in support

[32] Vogt, J. H. L., Bildung von Erzlagerstätten durch Differentiations-processe in basischen Eruptivmagmata: Zeitschr. prakt. Geologie, vol. 1, pp. 4–11, 125–143, 1893.

[33] Idem, p. 139.

[34] Phemister, T. C., op. cit., p. 508.

[35] Idem, p. 501.

of the secondary character of the ores is the fact that the sulphides replace not only hornblende but biotite, which he considers to be of hydrothermal origin. If this biotite. which is surrounded by sulphides, is really secondary, there seems to be no escape from the fact that at least a part of the Gap ores have been introduced by highly heated solutions.

The somewhat cursory investigation of these ores made during the geologic study of the region would suggest, although it can not be said to prove, that the greater part of the biotite is pyrogenetic rather than hydrothermal. It is an interesting fact that the gabbros of Maryland, which are in many respects petrographically similar to the gabbros of Mine Ridge, are notably free from biotite. Biotite as a subordinate constituent is present in the gabbros that intrude the Glenarm schists in the West Chester quadrangle; and it is an accessory constituent in the gabbros of the Philadelphia belt.[36] It is of universal occurrence in the gabbros and hornblende schists throughout the Mine Ridge anticline. This leads to the conclusion that either the gabbros of the Mine Ridge anticline differ essentially from the Maryland gabbros in the presence of magmatic biotite or else if the biotite is largely due to hydrothermal replacement the mineralization is not confined to the igneous rocks of the Gap mine.

There has been undoubtedly a widespread mineralization in the region, as indicated by the prevalence of tourmaline and pyrite throughout rocks of widely different ages, by the presence of magnetite and hematite in the pegmatites, and by the occurrence of pyrrhotite here and there in the Conestoga limestone. But whether all this mineralization was confined to one metallogenetic epoch is not yet clearly established. The probable sequence of events after the intrusion of the gabbro mass of the Gap mine is as follows: (1) Intrusion of a gabbroic magma; (2) consolidation of pyroxene; (3) consolidation of magmatic hornblende; (4) consolidation of pyrogenetic biotite; (5) intrusion of granodiorite; (6) formation of chessboard albite in the granodiorite during the latest stages of consolidation; (7) intrusion of albitic pegmatites that are satellitic followers of the granodiorite intrusion; (8) a strong dynamic deformation that affected both the gabbro, as shown by the twisting and bending of the biotite and the uralitization of the pyroxene, and the granodiorite and pegmatites, as shown by the saussuritization of the feldspar; (9) intrusion of granite, perhaps contemporaneous with the dynamic action; (10) intrusion of microcline pegmatites accompanied by boron vapors and ore-bearing solutions carrying magnetite, hematite, pyrrhotite, pyrite, chlorite, and sericite.

[36] Bascom, F., U. S. Geol. Survey Geol. Atlas, Philadelphia folio (No. 162), p. 6, 1909.

The sulphides that occur in the Conestoga limestone are in many places sharply defined crystals, with no evidence of deformation, and therefore can not be ascribed to the earlier pegmatites that have been dynamically deformed. In some places the pyrite is flattened and slickensided on planes of movement. This movement was doubtless Triassic and local, thus affecting only the pre-Triassic sulphides in certain places near which Triassic faulting took place. If the Gap nickel pyrrhotites have been manifestly deformed they can hardly be the equivalent of the undeformed sulphides in the limestones. If they have been deformed there are two possibilities—that they may be magmatic segregations from the gabbro magma, or that they may have been introduced by emanations from the granodiorite magma. So far petrographic study of the ore has failed to reveal any marked evidence of cataclastic action.

The sulphides that occur in the massive variety of the dynamically deformed and saussuritized gabbros are notably free from any evidence of cracking and stretching and are molded as large blebs around the silicates. But sulphides, which are quasi plastic under stress that breaks down silicate minerals, would not register evidences of cataclastic deformation of the rock. It is true that the sulphides occurring in the laminated schist are arranged parallel to the foliation, but it is impossible to prove whether they have been dragged out along the foliation planes by the shearing action of the deforming force or whether they were introduced into the rock subsequent to the foliation and their position determined by foliation. As previously pointed out, it is impossible even to tell whether this schist belongs to the nickel-bearing intrusion or to the invaded schists. Therefore it seems absolutely impossible to come to any definite conclusion in regard to the magmatic or hydrothermal origin of the ores unless their relation to the metamorphism can be established. If they are later than the deformation they can not be magmatic. The country rock has been subjected to pneumatolytic action on a large scale, but whether this pneumatolysis produced the nickel ores is not known. If the nickel ores were not produced during the regional pneumatolysis and if they were formed before the metamorphism of the gabbro, it is highly probable that they are of magmatic origin.

IRON ORE

Limonitic ore that occurs in limestone has been mined to some extent at several localities in the district. The largest workings were the C. B. Grubb ore banks in Conestoga limestone, half a mile east of Shenks Ferry station. They were worked from 1840 to 1850. The ore was a concretionary, somewhat manganiferous limonite, which was mined by open cuts and then washed. An analysis of this ore showed 53 per cent of metallic iron.[37]

Limonite was also mined at many places in Providence, Eden, and Bart Townships. The ore was the same in character and geologic occurrence at all these localities, being a concretionary or geodiferous limonite deposit that has replaced limestone, in many places along the contact of the limestone and the overlying schist. The ore has been formed through the action of downward-moving water carrying iron in solution and depositing it as limonite most abundantly along contacts or fault surfaces. The high-grade ore appears to have been removed, and the remaining deposits are believed to be small; at least present conditions do not warrant any attempt to mine the known deposits.

In 1893 an unsuccessful attempt was made to extract magnetite from the Cambrian biotite schist on Pequea Creek 1½ miles northwest of Marticville.

CHROME ORE

About 3 miles east of Susquehanna River, in Fulton Township near Lyles (locally known as Texas), there is an area of chromiferous serpentine. This area lies 2 miles north of the so-called "State Line" serpentine belt of southern Pennsylvania and Maryland.[38] Chromite occurs at intervals along the northern boundary of this belt and is reported to occur also in the southern part of the serpentine area near Lyles. The Brown pit, east of Little Conowingo Creek, is the only mine known to have been opened in this area, and it has been long abandoned and filled up. The ore occurs in three ways—as massive, compact chromite in nodules and pockets of variable size in a serpentine gangue; as a disseminated ore, locally known as spotted ore or birdseye, in which the grains of chromite are rather evenly disseminated through the serpentine; and as placer deposits. The Pennsylvania ore has in general proved to be of good grade and superior in quality to the California ores. The chromite of Maryland and Pennsylvania was the world's only source of chromium from the beginning of the industry until the Civil War. At that time, on account of the lowered prices of chromite caused by the importation of Turkish and other foreign ores, the Pennsylvania deposits were relegated to the position of a reserve supply. The increased demand for chrome ore during the World War led to a revival of mining in the chromite deposits of the eastern United States, but the life of the industry, which was dependent upon the abnormal demand, ceased with the termination of the war.

[37] Frazer, Persifor, jr., Geology of Lancaster County: Pennsylvania Second Geol. Survey Rept. CCC, p. 221, 1880.

[38] Bliss, E. F., Chrome ores of southeastern Pennsylvania and Maryland: U. S. Geol. Survey Bull. 725, pp. 85–98, 1921.

SILVER ORE

There is an abandoned silver mine on Silver Mine Run in the McCalls Ferry quadrangle, about 1 mile east of Conestoga. It was worked about 125 years ago, when a shaft was sunk that connected with a tunnel running 400 feet into the hill. The ore occurred in small isolated lenses, and no large body was discovered. The ore mineral is argentiferous galena in a quartz vein. The country rock is Vintage dolomite overlain by the Conestoga formation. In the much weathered open cuts, which are all that remain of the old workings, the quartz veins appear to have had their maximum development in the Vintage dolomite.

ROOFING SLATE

A high-grade black roofing slate occurs in the belt of Peach Bottom slate, which extends northeastward from Susquehanna River for a distance of 3 miles and which also extends southwestward 6 miles into Maryland, where it is extensively quarried on the Maryland-Pennsylvania State line at the towns of Cardiff and Delta. The slate belt along Susquehanna River is half a mile wide, but it pinches out toward the northeast. The trend of the slate outcrops along the river is N. 50° E.

Most of the slate quarrying has been done along the trend of Slate Hill, which follows the Maryland and Pennsylvania State line for several miles on the west side of Susquehanna River. There are two quarries on the east side of the river a quarter of a mile northwest of Peach Bottom station. Work was begun here about 1777 and continued for 100 years. In 1910 the upper quarry was reopened by Gorsuch Bros., who operated it on a small scale for local trade. The depth of the quarry is more than 80 feet. The good slate, which stands nearly vertical, alternates with nonproductive layers, and in following the valuable layers downward along the steep dip the working became too deep to operate profitably. There is much waste in connection with the quarrying and the refuse was dumped over workable beds. When the Pennsylvania Water Power Co. constructed a dam at Holtwood in 1910 the Columbia & Port Deposit branch of the Pennsylvania Railroad was obliged to relocate its tracks north of Peach Bottom station. The dump from the lower quarry was then used as ballast upon the new line. In 1920 the Gorsuch Bros. Co. built a mill for crushing the waste slate into granules to be used for making slate-surfaced prepared roofing.

The Peach Bottom slates are lustrous, blue-black, unfading slates, very sonorous.[39] They surpass the slates of Northampton and Lehigh

[39] Dale, T. N., Slate in the United States: U. S. Geol. Survey Bull. 586, pp. 110–113, 1914.

Counties in strength, toughness, and impact tests. They are extensively used for roofing and to a lesser extent as gravestones, and the excellent condition of slate roofs and gravestones that have been in existence for over 100 years testifies to their durability and quality.

LIMESTONE

The limestone belt in the northern part of the district contains valuable stone that has a wide range in use for building stone, ballast, and road metal and for making lime and cement. The largest quarry now in operation is at Kinzers, on the north side of the Lincoln Highway. This is the E. L. Slaymaker quarry in Vintage dolomite that is worked for crushed stone. The rock is a heavy-bedded dolomite dipping gently north and quarried with the dip.

Zittle Brothers' quarry, west of Strasburg, near Pequea Creek, is operated intermittently for crushed stone, and the Aldus Zittle quarry, northeast of Strasburg, is worked for the same purpose. These quarries are in the Conestoga blue limestone. The quarry of the Quarryville Lime & Stone Co., north of Quarryville, is in the micaceous marble facies of the Conestoga limestone. It is known as the Orchard quarry and was opened in 1826. The stone is burned for lime, and considerable pure marble is available in the quarry. Many quarries throughout the Lancaster and Quarryville Valleys are now abandoned or worked only to a small extent for local use. Many of them are in the marble conglomerate of the Conestoga formation. The J. P. McIlvain quarries, at Bellemont, are among the larger ones. Lime was burned there for many years, but the kilns are no longer used and the quarries are for the most part idle. There are many old quarries around London Grove, Wheatland, and New Providence, from which much limestone has been taken in former years. At present the cheapness and availability of lime produced in large plants has caused a decline in local lime burning, but crushed limestone is in considerable demand for making roads, and most of the limestone now quarried is used for that purpose.

SAND

The Chickies quartzite on the southwest side of Mine Ridge has disintegrated into white sand, which is dug for building and furnace sand. A sand pit north of Quarryville was opened by Leander Kunkel in 1921 and was being operated in a small way in 1922.

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