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THE OCCURRENCE AND ORIGIN OF ANALCITE AND MEERSCHAUM BEDS IN THE GREEN RIVER FORMATION OF UTAH, COLORADO, AND WYOMING

By Wilmot H. Bradley

ABSTRACT

Thin beds that consist almost wholly of euhedral analcite crystals occur at five or more horizons in the upper half of the Green River formation of Utah and Colorado, a series of Eocene lake beds that contain large deposits of oil shale. Most of these beds contain also minute crystals of apophyllite. The analcite crystals in some beds are cemented only by chalcedony, but in other beds that have less analcite the matrix is tuffaceous and consists of chalcedony in which are embedded splinters of feldspar, hornblende, and quartz, laths of biotite, and euhedral crystals of sanidine, apatite, and zircon. Analcite occurs plentifully also, disseminated in many of the oil-shale beds, and locally makes up as much as 16 per cent of weight of the rock. In these beds the analcite is accompanied by apophyllite, euhedral crystals of sanidine and orthoclase, angular fragments of quartz, and a little volcanic glass. The rocks associated with the beds of analcite and analcite-bearing oil shale contain salt-crystal cavities, that strongly suggest the former presence of glauberite and anhydrite.

Both clastic and hydrothermal hypotheses for the origin of the zeolites are considered and dismissed in the paper. The field and microscopic evidence presented leads to the conclusion that all the analcite and apophyllite formed in place on the lake bottom as a result of interactions between salts dissolved in the lake water and the dissolution products of volcanic ash that fell into the lake.

Meerschaum, or fibrous sepiolite, occurs in several thin beds near the top of the Green River formation in Duchesne County, Utah. It is mixed with structureless organic matter like that in the oil shale and apparently formed in place on the lake bottom.

Chemical analyses of the analcite rock and of the sepiolite are given.

INTRODUCTION

SCOPE AND PURPOSE OF THE REPORT

Analcite is rare in sedimentary rocks, but its occurrence as extensive beds is even more remarkable. In an earlier paper [1] the occurrence of thin beds of euhedral analcite crystals was noted briefly, and the hypothesis was advanced that they were derived from volcanic ash. These beds are so unusual as to merit more than incidental record. They represent a new phase in the alteration of volcanic ash, and as end products of that alteration they may perhaps be regarded as genetic relatives of bentonite, though their mineralogic composition and physical properties are utterly different. They indicate that analcite and apophyllite, which also occurs in these beds, can form at normal temperatures of the earth's surface and without the agency of gaseous or liquid emanations from lavas or volcanic sources; consequently, they provide a basis for supposing that definitely crystallized zeolites may be expected in either fossil or recent alkali soils. The analcite beds may have economic value as a source of zeolitic material for water softening. Finally, because of their remarkable areal extent they serve as useful correlative units within the part of the Green River formation that contains the richest beds of oil shale.

Therefore, the writer's purpose in this report is to describe the deposits of analcite in the Green River formation, to compare them with other similar deposits, and to present the observations and inferences that led him to explain them as alteration products of volcanic ash that fell into an ancient saline lake. The report also records the occurrence of several thin beds of sepiolite, or meerschaum, in the Green River formation and presents new data on the molds of saline minerals of the Green River formation whose determination affects directly the interpretation of the analcite and sepiolite deposits.

FIELD WORK AND ACKNOWLEDGMENTS

The observations and results recorded here are, in a way, by-products of the writer's study of the Green River formation, which began in 1922 and is still in progress. The specimens of analcite rock described were collected during the field season of 1925, when the writer was assisted by R. D. Ohrenschall. Some of the thin sections, however, which showed most clearly the tuffaceous matrix of the analcite beds were cut from specimens collected by D. E. Winchester in 1914. These sections were recorded with the obviously tentative field name sandstone.

OCCURRENCE OF ANALCITE IN OTHER SEDIMENTARY DEPOSITS

In clays of middle Cretaceous age near Duingen, Germany, Von Seebach [2] found clay ironstone concretions incrusted with dolomite and trapezohedrons of analcite. The crystals were colorless or whitish. Although apparently not genetically related to volcanic ash the deposit seems to have originated under conditions of low temperature, for Von Seebach remarks

[1] Bradley, W. H., Zeolite beds in the Green River formation: Science, new ser., vol. 67, pp. 73–74, 1928.

[2] Von Seebach, Karl, Notiz über ein neues Vorkommen des Analcim: Georg Augusts Univ. K. Gesell. Wiss. Göttingen Nachr., 1862, pp. 334–335.

that it is similar to the deposits of carbonates and quartz known to occur only in nonvolcanic regions. Guthe [3] confirmed Von Seebach's identification of the analcite by an analysis, which showed some ferric iron and an excess of water. These apparent impurities may, however, have been due to the traces of a reddish incrustation on the crystals noted by Guthe.

Another occurrence of analcite in sedimentary beds near Lehre, Germany, is recorded by Kloos.[4] This deposit also appears to have formed at a low temperature, despite the intergrowth of the analcite with sphalerite.

The formation of the analcite was clearly subsequent to the diagenesis of the beds in both the deposit described by Von Seebach and that described by Kloos. Neither is therefore strictly comparable to the syngenetic deposits of analcite and apophyllite in the Green River formation.

More nearly analogous to the zeolites of the Green River formation in its mode of occurrence is the zeolite phillipsite which Murray and Renard [5] found to be forming in large quantities in the red deep-sea oozes. In those oozes they found an exceptional abundance of fragmental lapilli of vesicular basalt, much of which had a glassy groundmass, also palagonite, a hydration product of basaltic glass. They regarded the phillipsite as produced from volcanic material by a process of hydration and alteration that went on very slowly because of the low temperature, 2°–3° C., and the almost total quiescence of the water.

A palagonite tuff from a small island 3 miles east-northeast of Reykjavik, Iceland, which has been partly altered to analcite, faujasite, and an impure potash analogue of analcite, has been described by Tyrrell and Peacock.[6] This tuff consists largely of clear yellow perfectly isotropic palagonite, which they characterize as gel-like. Of the significance of this variety of palagonite they say:

> It was previously indicated that gel-palagonite appears to have a somewhat higher water content than fibro-palagonite, and on considering the notes on the occurrence of the various specimens we find that the tuffs which are characterized by gel-palagonite are those which have been submerged and have, therefore, been acted on at raised pressure (hydrostatic) and low temperature, while those which contained dominant fibro-palagonite were always associated with hot springs.[7]

The analcite of the deposit is evidently very closely similar in its mode of occurrence to the analcite in the water-laid tuffs of the Green River formation.

Beds of analcite that formed in playas in western Yavapai County, Ariz., have been described recently by C. S. Ross,[8] of the United States Geological Survey. These beds resemble fine-grained light-green sandstone and consist almost exclusively of analcite trapezohedrons that average about 0.06 millimeter in diameter. The individual crystals are surrounded by thin films of glauconitic material. Ross suggests two possible explanations for the origin of the analcite: It may have formed as a result of reaction between the sodium salts concentrated in the playa lake and the hydrous aluminum silicates in the playa muds; or, on the other hand, as glassy volcanic ash is common in those playas and as the analcite beds contain a small quantity of minerals generally associated with ash, the concentrated sodium salts may have reacted with the decomposed ash to form the analcite. Ross's second interpretation of the origin of the analcite in the playas in Arizona agrees exactly with the writer's interpretation of the origin of the analcite beds in the Green River formation.

ANALCITE BEDS OF THE GREEN RIVER FORMATION

OCCURRENCE OF THE ANALCITE AND APOPHYLLITE

Lenses or beds of analcite were found at five different levels in the section of the Green River formation in the canyon of White River, in sec. 27, T. 9 S., R. 25 E., Uintah County, Utah. The largest of these lenses is about 40 feet long and 3 feet 3 inches thick at the thickest place and lies about 700 feet above the base of the formation. At about 13, 85, and 130 feet above it there are other thin beds and lenses. About 50 feet below the largest lens there is a thin bed of oil shale that contains short stubby lenses of analcite. Lenses of analcite were also found about 700 feet above the base of the formation in Hells Hole Canyon in sec. 22, T. 10 S., R. 25 E., Uintah County, Utah. Thin sections cut from material collected in Colorado by D. E. Winchester in 1914 showed that several of the beds consist chiefly of analcite. The samples came from localities (1) north of the iron bridge on the old Watson-Rangely wagon road about 2 miles east of the Utah line (probably in sec. 24, T. 1 N., R. 104 W.), Rio Blanco County; (2) in the cliffs on the north side of White River, along the same road but about 5 miles east of the State line (approximately in sec. 9, T. 1 N., R. 103 W.); (3) in the Cathedral Bluffs, in sec. 26, T. 1 N., R. 100 W., Rio Blanco County. The horizons in the Green River formation at which those beds occur are unknown. The writer found lenticular beds of analcite near the top of the formation farther south along the Cathedral Bluffs, in sec. 26, T. 3 S., R. 99 W., Garfield County, Colo.

Recently Mr. Winchester sent to the writer from De Beque, Colo., a sample of the "marker bed" that lies just above the group of rich oil-shale beds known as the Mahogany Ledge and from Grand

[3] Guthe, H., Mineralogische Notiz: Naturhist. Gesell. Hannover 12. Jahresb., p. 41, 1863.

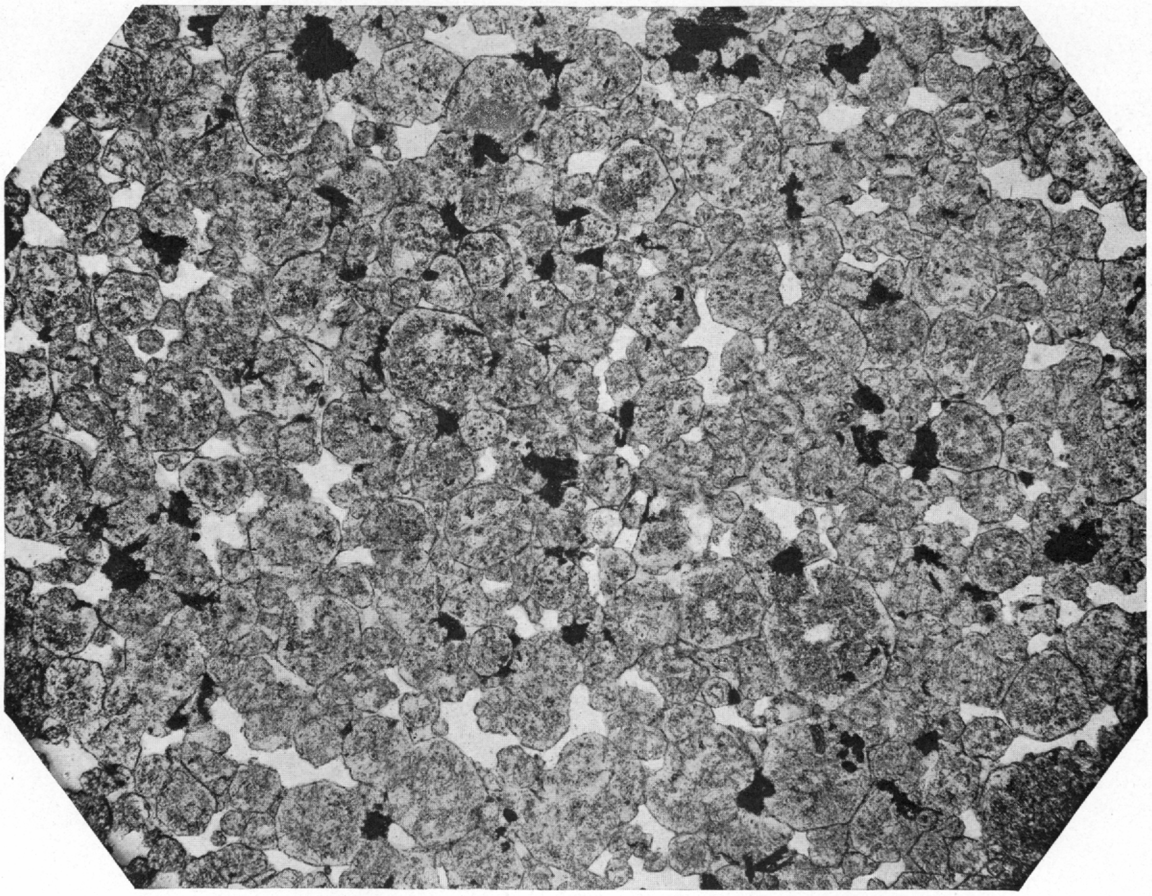
[4] Kloos, J. H., Über Analcim auf Steinkernen von Ammoniten und auf Sphärosiderit von Lehre: Ver. Naturwiss. Braunschweig 11. Jahresb., pp. 153, 167–168, 1899.

[5] Murray, John, and Renard, A. F., *Challenger* Rept., Deep-sea deposits, pp. 400–411, 1891.

[6] Tyrrell, G. W., and Peacock, M. A., The petrology of Iceland: Roy. Soc. Edinburgh Trans., vol. 55, pt. 1, No. 3, pp. 62–63, 1926.

[7] Idem, p. 71.

[8] Ross, C. S., Sedimentary analcite: Am. Mineralogist, vol. 13, pp. 195–197, 1928.

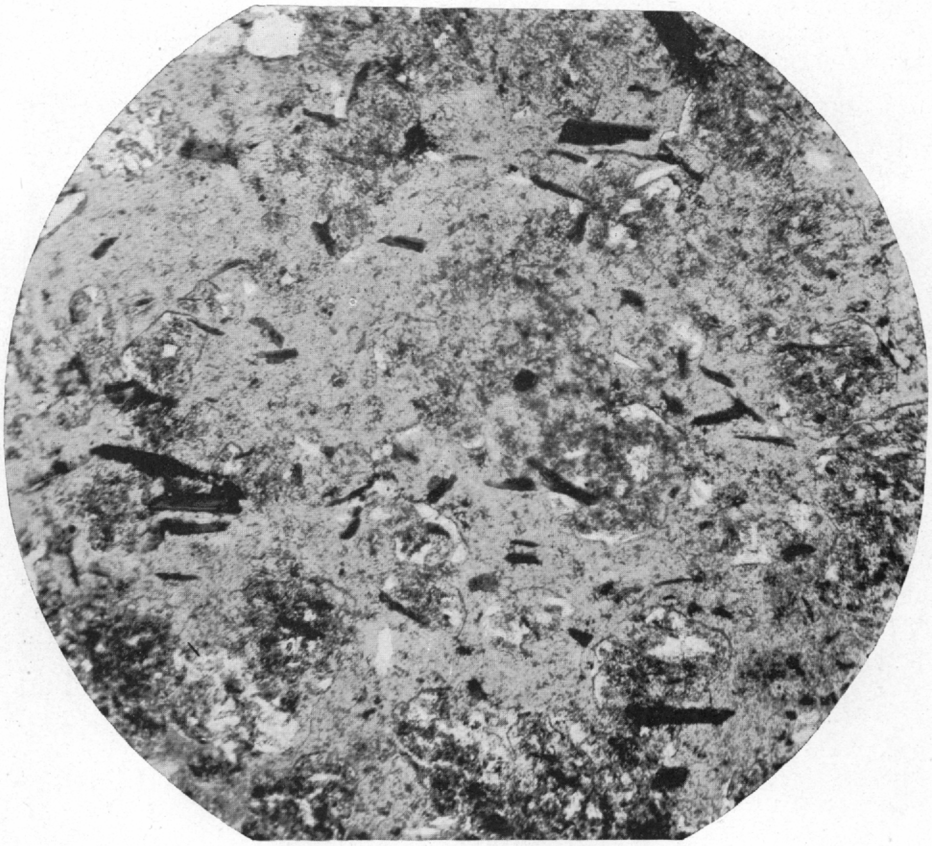


A. PHOTOMICROGRAPH OF A THIN SECTION OF A SPECIMEN FROM AN ANALCITE CRYSTAL BED

Showing the idiomorphism of the crystals and the abundance of dustlike inclusions. The jet-black is pyrite, and the clear gray areas between the crystals are holes in the section. From a bed about 750 feet above the base of the Green River formation in the canyon of White River, sec. 27, T. 9 S., R. 25 E., Uintah County, Utah. Enlarged 36 diameters

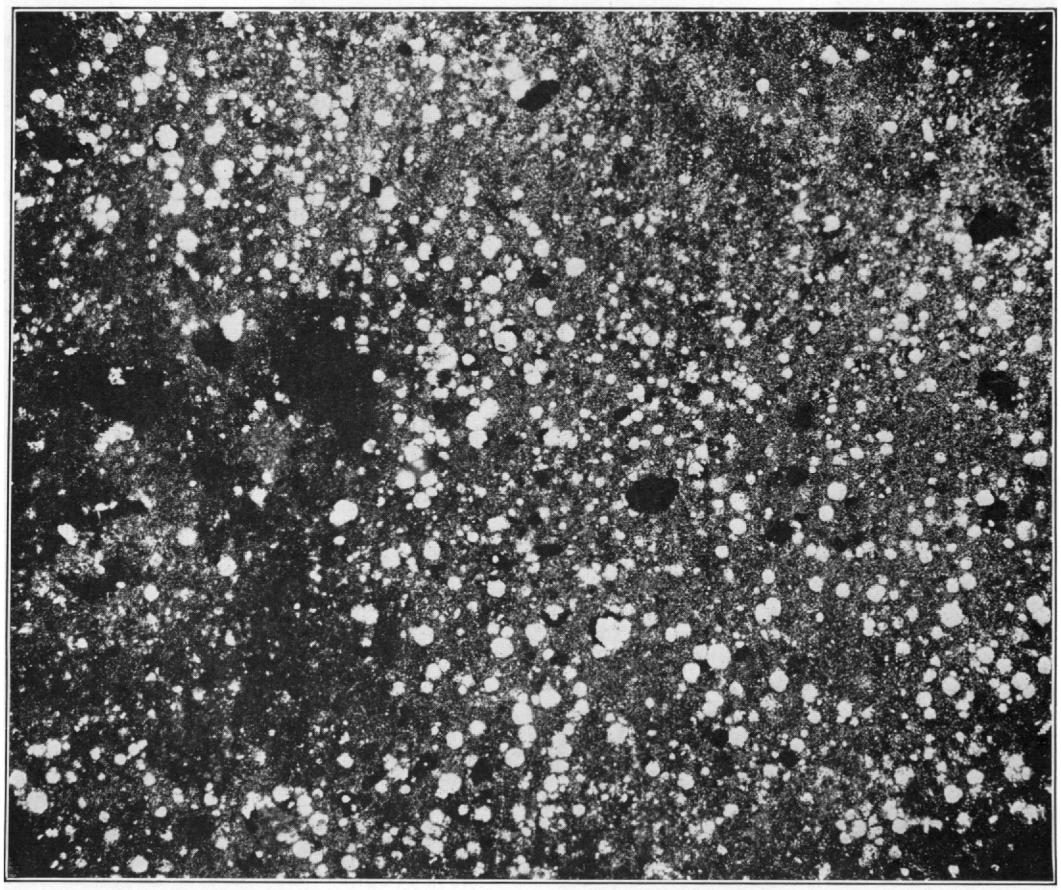
B. PHOTOMICROGRAPH OF A PART OF THE THIN SECTION SHOWN IN *A*

Showing at " a " inclusions of minute lozenge-shaped apophyllite crystals in analcite. Enlarged 500 diameters



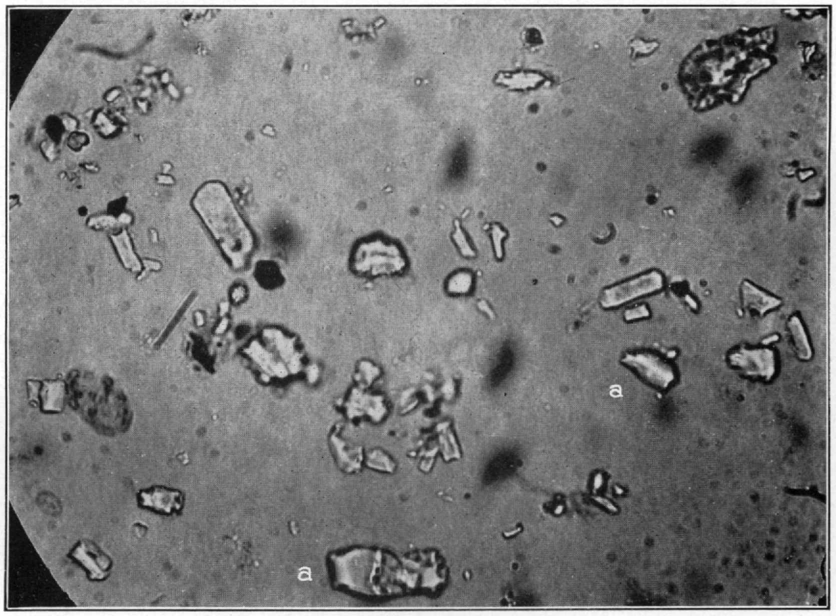
A. PHOTOMICROGRAPH OF A SPECIMEN FROM AN ANALCITE CRYSTAL BED WITH A DISTINCTLY TUFFACEOUS MATRIX

Showing the subparallel orientation of the large biotite flakes, some of which are deeply embedded in the analcite crystals. The medium-toned gray matrix is chalcedony. From a bed in the north wall of White River Canyon approximately in sec. 9, T. 1 N., R. 103 W., Rio Blanco County, Colo. Enlarged 80 diameters



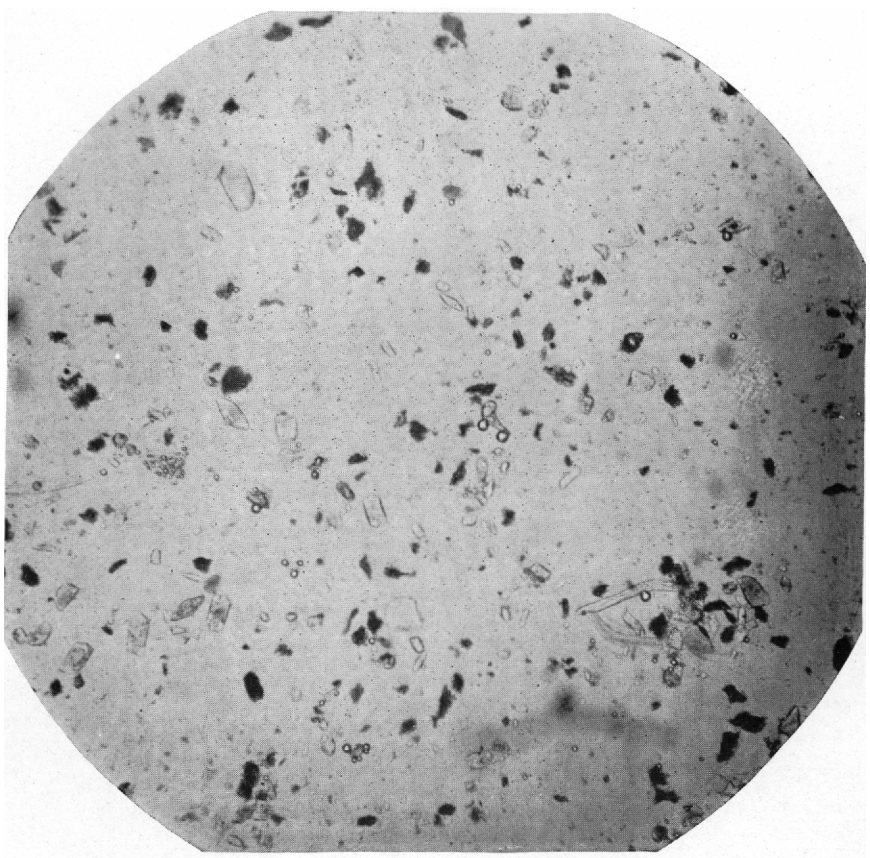
B. PHOTOMICROGRAPH OF A THIN SECTION CUT PARALLEL TO THE BEDDING OF A RICH STRATUM OF OIL SHALE STUDDED WITH ANALCITE CRYSTALS

The finest white specks are grains of dolomite and calcite; the jet-black is pyrite, which in places encircles analcite crystals; and the gray groundmass (almost obscured by the abundance of minute carbonate grains) is structureless organic matter. Enlarged 36 diameters



A. PHOTOMICROGRAPH OF A GROUP OF MINUTE APOPHYLLITE CRYSTALS ISOLATED FROM OIL SHALE

Showing nearly perfect prismatic crystals and also irregular forms as at "a." From a thin rich layer of oil shale in the " Mahogany ledge " at the U. S. Bureau of Mines experimental mine in sec. 23, T. 6 S., R. 96 W., Garfield County, Colo. The grains are immersed in an oil which has a refractive index of 1.58. Enlarged 500 diameters



B. PHOTOMICROGRAPH OF A GROUP OF EUHEDRAL SANIDINE AND ORTHOCLASE CRYSTALS ISOLATED FROM OIL SHALE

From a bed in the western part of T. 6 S., R. 95 W., Garfield County, Colo. The grains are immersed in an oil which has a refractive index of 1.58. Enlarged 80 diameters

Valley, Colo., a sample of another similar bed that lies about 70 feet below the "marker bed." Thin sections show that both these beds consist of deeply altered volcanic ash containing an abundance of analcite. Heretofore these "markers" have been regarded either as oolite or as thin beds of sandstone.

The analcite crystals in these beds differ greatly in size, but some are nearly 2 millimeters in diameter. Nearly all are euhedral, though the faces of many are rough and uneven. All are opaque and dull gray. Some are darkly stained with iron oxide.

Thin sections show that the analcite crystals are crowded with minute inclusions, most of which appear to be isotropic, though they may be anisotropic and yet too small to show birefringence. (See pl. 1, *A*.) In plain light these dustlike inclusions are gray or brownish gray. Fragments of quartz, sanidine, orthoclase, plagioclase, and biotite, together with euhedral crystals of apophyllite, apatite, pyrite, and zircon, are also included in the analcite. Few of the apophyllite crystals so included are more than 0.005 millimeter long. Without exception these tiny crystals are a combination of very short prism and basal plane, which are distorted so that they appear distinctly lozenge shaped in cross section. (See pl. 1, *B*.) In many analcite crystals the inclusions are grouped in a subparallel manner, and in several of the crystal beds the direction of parallelism is the same not only for the inclusions within the various analcite crystals but also for the oriented grains of the cementing material. The apophyllite crystals do not share in this general orientation, but are oriented at random. In a few analcite crystals there are relatively clear areas whose size and shape are wholly comparable to those of the angular grains of feldspar found as inclusions in analcite crystals of other beds. These clear areas are more or less definitely outlined by the dustlike inclusions and suggest complete replacement of foreign grains, presumably feldspars, by analcite. In fact, some of the very deeply altered orthoclase found as inclusions in analcite crystals may represent a transition step in this replacement.

There appear to have been two stages of analcite formation in at least one of the beds, for the fully grown trapezohedrons of analcite were cracked and the cracks were filled with perfectly clear analcite wholly free from inclusions. These cracks appear to have resulted from the same compression that caused a certain amount of flattening of the crystals during compacting and diagenesis of the beds.

Nearly all the analcite crystals in the crystal beds and also in the oil shale are completely isotropic. Only the borders of the largest crystals and the clear analcite that fills cracks in the older crystals of certain beds show feeble anomalous birefringence. The refractive index is 1.488 ± 2.

The matrix of the zeolite beds consists of microcrystalline chalcedony or opal clouded with dustlike inclusions. Embedded in the silica are various quantities of elongate splinters and sharply angular fragments of fresh sanidine and quartz and more or less altered orthoclase and plagioclase, together with a considerable number of laths and plates of biotite. In one bed the biotite is altered to chlorite. Minute euhedral crystals of apatite, apophyllite, and zircon and a few lath-shaped crystals of feldspar terminated by definite faces were also found. Pyrite occurs as aggregates of minute grains and also as tiny cubes and octahedra between the analcite crystals and intergrown with them. Clay minerals and minutely granular carbonates are virtually absent. This lack is noteworthy, as fine carbonate grains and particles of clay minerals predominate in most other beds of the Green River formation.

In one bed of oil shale a little below the Mahogany Ledge at the Monarch Shale Oil Co.'s mine, in sec. 32, T. 6 S., R. 97 W., Garfield County, Colo., analcite makes up a little more than 16 per cent by weight of the rock. (See pl. 2, *B*.) Another bed of oil shale is worthy of mention because it contains fragments of partly devitrified greenish-brown glass in all stages of replacement by small crystals of analcite. This bed, which occurs in sec. 36, T. 1 N., R. 96 W., Rio Blanco County, Colo., contains also a rather large number of zircon crystals and a little mica.

The trapezohedrons of analcite in the oil-shale beds are considerably smaller than those in the crystal beds. They average about 0.065 millimeter in diameter but range from about 0.004 to 0.15 millimeter. Also, they contain fewer and different inclusions. They are clear and are not clouded with dustlike specks but contain instead minute irregular grains and rhombs of carbonates and tiny flakes of clay minerals. In general, the crystals more than 0.04 millimeter in diameter have been distinctly flattened parallel to the bedding laminae. Vertical sections of some of these show only vague or badly distorted crystal faces, but considered together with their horizontal sections they are clearly trapezohedrons that have been compressed during the compaction of the shale to about half their normal diameter. These distorted crystals contain many irregular cracks, which are not confined to a peripheral zone like the shrinkage cracks of larger analcite crystals.

Apophyllite crystals in the oil shale are neither so large nor so plentiful as the analcite. Most of the crystals are between 0.02 and 0.03 millimeter in length, but many are less than 0.005 millimeter, and none of those measured exceed 0.075 millimeter. The most common form is a rather long first-order prism with second-order pyramid, though many crystals were found which had only one pyramidal termination, the other end being irregularly rounded off. (See pl. 3, *A*.) Smooth-surfaced fusiform grains that showed no distinct faces were also found; a few of these were more nearly pyriform and resembled Rupert's drops. The larger crystals generally contain a few minute inclusions. The writer estimated that apophyllite makes up as much as 5 per cent of a thin streak of exceedingly rich oil shale from the Mahogany Ledge in sec. 23, T. 6 S., R. 96 W., Garfield County, Colo. In most other oil-shale beds, however, the percentage is considerably smaller.

The apophyllite included in the analcite crystals is either older than the analcite or contemporaneous with it, but the apophyllite not so included may have formed at any time during the history of the Green River formation. There appears to be no way to correlate its growth with that of any other mineral in these beds. On the other hand, there is nothing to suggest two stages of apophyllite formation, and this lack of evidence might perhaps foster the inference that all the apophyllite formed along with the older analcite, very early in the history of the oil shale.

ORIGIN OF THE ZEOLITES

The hypothesis of a clastic origin for the zeolites does not seem to accord with their observed relations in the Green River formation. Despite the great range in size between the smallest crystals of analcite and apophyllite and the largest crystals of analcite, all occur together without semblance of sorting, and they are accompanied by only insignificant amounts of other clastic grains, such as quartz and feldspar, of comparable size and weight. Moreover, according to the writer's observations, the zeolites in the oil shale seem to be most plentiful in the beds whose clastic constituents, quartz, feldspar, and clay minerals, are markedly subordinate to their chemically precipitated constituents, calcite, dolomite, opal, and pyrite. Then, too, both zeolites appear to be consistently absent from the closely associated beds of fine-grained sandstone and limy siltstone, where they might confidently be expected if they were of clastic origin. Furthermore, many of even the largest analcite crystals have perfectly sharp interfacial angles and show no effects of abrasion. For these reasons the idea that these zeolites are a peculiar concentrate derived from disintegration of an analcite basalt or similar basic rock can be dismissed.

The hypothesis of a hydrothermal origin seems no better suited to explain the occurrence of these zeolites for they are very plentiful in beds of rich oil shale whose organic matter, if given sufficient time, will distill at temperatures as low as 100° to 200° C.[9] Plainly no such liquefaction has occurred. Furthermore, the oil-shale beds containing the zeolites and also some of the crystal beds extend over hundreds of square miles and occur in two unconnected intermontane basins. The very extent and uniformity of these deposits, coupled with the total absence of any indications of hydrothermal alteration, argue strongly against the probability of metasomatic action subsequent to the diagenesis of the beds. In fact, the formation of the zeolites within the obviously unaltered oil shale through the agency of solutions of any kind that entered after lithification of the beds seems inconceivable, because the oil shale is so extremely fine grained and compact that it appears to be utterly impervious to circulating waters. Further than that, during diagenesis both the oil shale and the crystal beds were being progressively compacted, and liquids were being expressed from them. Therefore, solutions from external sources would have had very little opportunity to enter until loading of the beds by younger rocks and the consequent compacting had ceased, but by that time the beds would have reached essentially their present state of imperviousness.

To postulate the action of hot springs as an explanation of such extensive and uniform deposits of zeolites would be to make the patently absurd assumption that not one but both of these ancient lakes, covering in all more than 25,000 square miles, were kept hot for long periods of time. The biologic evidence[10] alone is sufficient to make this hypothesis untenable.

Hence, it appears that the zeolites in the Green River formation are syngenetic and that they formed at a temperature probably below 30° C. This interpretation of the origin of analcite and apophyllite departs rather widely from the commonly accepted theory. The writer proposes the hypothesis that these minerals resulted from interaction between the dissolution products of volcanic ash that fell into the ancient Green River lakes and salts dissolved in the lake water. That the zeolites of the crystal beds grew in place seems to be clearly shown by the common orientation of the mineral grains and groups of dust-like inclusions in both the analcite crystals and in the matrix and by the random orientation of the apophyllite crystals, which, being prismatic, might be expected to coincide with the subparallel arrangement of the elongate splinters of quartz, feldspar, and biotite.

In some crystal beds the biotite laths and the elongate splinters of quartz and feldspar in the matrix are oriented roughly parallel to the bedding planes, and this indication of bedding, together with the presence of organic matter diffused through these beds, shows clearly that they accumulated under water. (See pl. 2, *A*.) Further reason for believing that these zeolites formed on the lake bottom is found in the occurrence of large quantities of both analcite and apophyllite disseminated through many of the oil-shale beds, not only in Colorado and Utah but also in Wyoming.

The general distortion of the analcite crystals in oil-shale seems to indicate that they formed when not deeply buried, for the greater part of the compaction of an argillaceous sediment occurs before it has been

[9] Karrick, L. C., Manual of testing methods for oil shale and shale oil: U. S. Bur. Mines Bull. 249, p. 4, 1926.

[10] Bradley, W. H., A contribution to the origin of the Green River formation and its oil shale: Am. Assoc. Petroleum Geologists Bull., vol. 9, pp. 247-262, 1925.

buried as much as 100 feet. This inference renders untenable a hypothesis that most of the analcite was formed at the high pressures and high temperatures incident to deep burial, although the clear, obviously secondary analcite that fills cracks may have formed at depth. Probably both the analcite and the apophyllite formed before the organic ooze had been greatly compacted and had consequently become inimical to the intermingling and reacting of the zeolite constituents.

The intimate relation between the pyrite and the analcite suggests that the analcite formed contemporaneously with or soon after the accumulation of the organic ooze. The pyrite seems plainly to have formed in the organic ooze from an interaction between the iron salts dissolved in the lake water and hydrogen sulphide derived either from the decay of organic matter or from the reduction of inorganic sulphates as a result of that decay. On this pyrite, much of which has a distinctive, more or less compact microgranular structure, the analcite seems to have exerted a marked localizing influence. Many of the analcite crystals are ringed with aggregates of pyrite granules or with minute cubes and octahedra (see pl. 2, *B*), and some appear to be entirely encased in a pyrite shell. Pyrite is also intergrown with analcite. This distribution of pyrite is quite different from that in oil-shale beds that contain little analcite. In them the minute individual pyrite grains or small groups of them are rather uniformly disseminated through the organic matter.

The genetic relation between the zeolites of several of the crystal beds and their tuffaceous matrices is evident. The occurrence of biotite flakes, zircon, and apatite as inclusions in the analcite crystals suggests a volcanic source for the material, but the general aspect, the composition, and the structure of the matrix inclosing the analcite crystals indicate more plainly that these crystal beds resulted from a rather thorough alteration of volcanic ash. Furthermore, the apparent absence from the analcite beds of clay minerals and minutely granular carbonates, which are the predominant constituents of the other beds in the Green River formation, strongly suggests that the original material of the zeolite beds accumulated so rapidly that it completely obscured them. Only a fall of ash seems to be consistent with this apparent rapidity of deposition and with the fine grain and mineral composition of the matrix in the analcite beds.

That the zeolites of the oil shale may also be genetically related to the volcanic ash is less apparent but yet seems reasonable. First, there is the analogy between the occurrence of analcite and apophyllite in the crystal beds and their occurrence in the oil shale, beds of the same two minerals bearing the same relations one with the other. Then, the oil shale that contains the zeolites, in common with the tuffaceous crystal beds, contains elongate splinters and sharply angular fragments of sanidine, orthoclase, plagioclase, biotite, and quartz and idiomorphic crystals of sanidine, orthoclase, apatite, zircon, and more rarely also of plagioclase. (See pl. 3, *B*.) A few zeolite-bearing oil-shale beds contain also fragments of hornblende and altered volcanic glass. The dissemination of the zeolites within an oil-shale bed and the admixture of normal sedimentary minerals, such as the carbonates and clay minerals, seem to be natural consequences if the volcanic ash fell into and became mixed with a nearly fluid organic ooze.

In connection with the alteration of the ash and the formation of the zeolites it is germane to consider the probable chemical character of the lake water into which the ash fell. At many horizons in and above the oil-shale zones of the Green River formation there are abundant salt-crystal molds. W. F. Foshag,[11] of the United States National Museum, examined some of these molds in a sample collected by the writer from the cliffs just east of the town of Green River, Wyo. He regarded them as probably the molds of glauberite and said:

> The simplest form noted is a rhomboid form, corresponding to the combination of prism and the base of glauberite. The interfacial angles are dull and distorted, but an approximate measurement of the prism angle gave 90°. Glauberite has a prism angle of 96°. The agreement is as close as can be expected from this type of material. Another type of crystal mold gives acute and lenticular cross sections corresponding to the type of glauberite crystal in which the prism is prominently developed.

Crystal cavities in rocks of the Green River formation collected later from several other localities were examined for the writer by W. T. Schaller, of the United States Geological Survey. Some of these strongly suggested glauberite and others anhydrite. In a few specimens cavities suggesting both minerals are closely associated. In one specimen from a quarry a short distance northwest of Green River, Wyo., in sec. 15, T. 18 N., R. 107 W., the cavities suggest not only glauberite and anhydrite but a third mineral whose angles do not agree with any known salt. These crystals had a flattened rhomboid form with the following interfacial angles:

b:c	a:b=a:c
64°	87°
65	84
60	87
62	86
65	88
66	89
Av. 64	Av. 87

Probably this mineral is neither new nor unusual but rather a distorted form of a well-known salt.

As the lake water into which the volcanic ash fell contained solutions of these salts it seems reasonable to conclude that the water was more or less alkaline.

Volcanic ash, especially the glassy portions, which are in a metastable condition, must have been quickly attacked by the alkaline lake waters, and thus rela-

[11] Personal communication, Jan. 14, 1924.

tively large quantities of alumina, silica, potash, and other constituents would have been made available for recombination. Many feldspar grains were deeply altered and seem also to have contributed material for recombination. New minerals, stable under the new conditions, could in this manner form contemporaneously over very large areas—indeed, areas coincident with the extent of the lake at the time of each ash fall. Thus the analcite and apophyllite in the Green River formation appear to have formed as reaction products of the silica and alumina liberated by dissolving the volcanic ash with the solutions of sodium, potassium, and calcium salts that were either normal constituents of the lake water or that were also derived from alteration of ash.

CHEMICAL COMPOSITION OF AN ANALCITE BED

A chemical analysis of a sample from a thin asphalt-saturated ash bed containing much analcite was made in the hope that some information as to the character of the original ash might be obtained. This sample was taken from a bed about 750 feet above the base of the Green River formation in the canyon of White River, in sec. 27, T. 9 S., R. 25 E., Uintah County, Utah. As shown by the analysis given below and the mineral composition computed from it, the ash is too much changed to afford any notion as to its original composition. Apparently much sodium, silica, iron, and water have been added.

Analysis and calculated mineral composition of analcite rock from the Green River formation of Utah

[J. G. Fairchild, analyst]

Analysis		Calculated mineral composition		
			By weight	By volume
SiO_2	52.54	Analcite	65.84	70.34
Al_2O_3	15.78	Chalcedony, quartz, and opal	15.66	14.49
Fe_2O_3, FeO	4.54	Sanidine and orthoclase	3.71	5.67
MgO	.63	Limonite(?)	11.27	7.58
CaO	.36	Pyrite	2.73	1.31
Na_2O	8.63	Calcite	.69	.61
K_2O	1.16			
H_2O	.05			
H_2O+	8.96			
CO_2	.29			
FeS_2	2.74			
Bituminous material	3.93			
	99.61			

Calculated in this way there is a small excess of Al_2O, and MgO and several per cent of H_2O. The large amount of limonite is evidently a weathering product of the pyrite, which is very plentiful in the freshest parts of the rocks. Presumably, however, the MgO and some of the SiO_2, Al_2O_3, and Fe_2O_3 (or FeO) should be combined to make up a part of the dustlike inclusions and tiny fragments of ferromagnesian minerals. The excess of water is puzzling. A part of it may be contained in opal, and if some of the dust-like inclusions in the analcite are really volcanic glass perhaps they contain a part also.

MEERSCHAUM BEDS OF THE GREEN RIVER FORMATION

OCCURRENCE AND COMPOSITION

Beds of meerschaum, or sepiolite, so far as the writer has been able to learn, have not heretofore been reported from any locality in this country, and it is therefore of interest to record the occurrence of several thin beds, which the writer found in 1925 in the Green River formation. They lie high in the west wall of Indian Canyon, about 4,400 feet above the base of the formation, in sec. 11 (?), T. 10 S., R. 7 W., Duchesne County, Utah. None of the beds of this small group exceed 1 centimeter in thickness. On its weathered surface the material is light bluish gray and has a distinctly silky luster. Below the weathered surface it is dark grayish brown, owing to the large admixture of structureless organic matter. In fact, it is a fairly rich oil shale whose principal inorganic constituent is sepiolite, and like some other oil shales of this formation it has a pronounced fissility and resembles matted coarse brown paper pulp. Unlike other oil shales, however, it is light enough when dry to float on water. The sepiolite layers are interbedded with chocolate-brown oil shale that contains an abundance of glauberite (?) molds that are now partly filled with calcite.

Thin sections show that the sepiolite is all crystalline and that it consists of a felt of excessively fine fibers that have positive elongation. The refractive indices, $\alpha = 1.528 \pm 2$; γ, $\beta = 1.538 \pm 2$, are somewhat higher than those given by Larsen [12] for β sepiolite, namely, $\alpha = 1.519$, $\gamma = 1.529$, $\beta = 1.52$, but the birefringence is approximately the same. The mineral is biaxial and negative, with 2V fairly large. It does not gelatinize when heated with strong HCl, but, instead, granular silica separates out. Hence, according to Vernadsky's distinction[13] between two varieties of sepiolite, this appears to be the β variety.

The material is mixed with minute crystals and spherulites of calcite and a considerable amount of organic matter, but some of the cleanest fibrous mineral was picked out and analyzed. The results of this analysis and the recalculated form are given below, together with the theoretical composition of sepiolite for comparison.

[12] Larsen, E. S., The microscopic determination of nonopaque minerals: U. S. Geol. Survey Bull. 679, p. 245, 1921.

[13] Vernadsky, W., Zur Theorie der Silicate: Zeitschr. Kryst. Min., vol. 34, pp. 46-47, 1901.

Analysis of fibrous sepiolite from the Green River formation of Utah

[E. T. Erickson, analyst]

	As reported	Recalculated	Theoretical [a]
SiO_2	35.92	58.40	60.80
Al_2O_3	[b] 1.16	1.89	
Fe_2O_3	[c] .26	.42	
MgO	16.34	26.58	27.10
CaO	1.88	.69	
TiO_2	.10	.10	
CO_2	1.14		
Loss on ignition	43.20		
H_2O		11.86	12.10
	100.00	100.00	100.00

[a] Dana, J. D., System of mineralogy, 6th ed., p. 680, 1909.
[b] Uncorrected for P_2O_5, if present.
[c] Calculated from total Fe.

In recalculating this analysis it was assumed that all the CO_2 was combined with CaO, for according to the refractive index ($\omega = 1.66 \pm 2$), the calcite is practically pure. This left an excess of 0.69 per cent of CaO as an impurity. Then, as the water could not be determined because of the organic matter present, an amount of water chemically equivalent to the amount of MgO was assumed. It should be noted, however, that, wholly apart from these assumptions, the silica-magnesia ratio of the Utah mineral, 2.20 agrees very closely with that of the theoretical sepiolite, 2.24. The other constituents shown by the analysis are probably to be regarded either as impurities of the mineral itself or as admixtures.

COMPARISON WITH EUROPEAN MEERSCHAUM BEDS

This Utah deposit of sepiolite is comparable to those in the Paris Basin and at Vallecas, about 5 miles south of Madrid, in both of which, according to Brongniart,[14] the sepiolite occurs in Tertiary lake beds, associated with gypsum and rock salt and interbedded with marly limestone and limy marl. It is also much like the sepiolite in the fresh-water Tertiary lake beds in southern France near Sommieres, described by De Serres.[15] It differs from all these, however, in several respects. Silica, usually as some variety of opal, is consistently associated with the European sepiolite beds, but it is apparently absent from the Utah deposit. On the other hand, the Utah sepiolite is intimately associated with a large quantity of organic matter and is accompanied by glauberite (?) and magnesite rather than gypsum or rock salt.

ORIGIN OF THE MEERSCHAUM

At the time the sepiolite formed, that part of the ancient Green River lake in the vicinity of Indian Canyon, Utah, seems to have been very shallow and saline. Most of the shale beds just below as well as just above the sepiolite horizon contain an abundance of salt molds (glauberite ?) and pseudomorphs. Sun-cracked bedding planes, too, are plentiful, and the polygonal pattern on several of these is made more conspicuous by the arrangement of the salt pseudomorphs along the cracks. One sepiolite layer rests upon a sun-cracked and salt-studded surface of this kind. Apparently during this phase of the lake the amount of magnesium in solution exceeded the amount of calcium, for several of the bedding planes of the rocks closely associated with the sepiolite layers are studded with small, crudely fusiform crystals of magnesite. A part of this readily available magnesium apparently combined with silica, perhaps as a hydrated gel, to form sepiolite.

[14] Brongniart, Alex., La magnésite du bassin de Paris: Annales des mines, vol. 7, pp. 303-304, 1822.

[15] De Serres, Marcel, Mémoire sur les terraines d'eau douce: Jour. physique, vol. 87, pp. 134-135, 1818.

THE CONTACT OF THE FOX HILLS AND LANCE FORMATIONS

By C. E. Dobbin and John B. Reeside, Jr.

ABSTRACT

The presence or absence of an extensive unconformity at the contact between the Fox Hills and Lance formations is an important element in the problem of the Cretaceous-Tertiary boundary in the Western Interior of the United States. Proponents of the theory that an unconformity exists have rested their case primarily on a correlation of the Arapahoe and Denver formations with the Lance and Fort Union formations, the supposed absence in the northern Great Plains of the Laramie formation, the relative thinness of the Fox Hills sandstone in the northern Great Plains as compared with that in the Denver Basin, and reported angular discordances between the Fox Hills and the Lance at several localities. The present writers describe, chiefly from personal observation, the contact of the Fox Hills and Lance formations over much of the area of their occurrence and of units equivalent to these formations in areas where other names are used, paying especial attention to localities in which there are reported discordances. They advocate the view that the Laramie, Arapahoe, and Denver formations of the Denver Basin are represented in the Lance formation; that the differing thicknesses of the Fox Hills are original and not due to varying amounts of pre-Lance erosion; that each supposed angular discordance is due to some other phenomenon, such as local small faulting or cross-bedding; and finally, as a general conclusion, that there is no unconformity between the Fox Hills and Lance other than such as is found along local minor erosion planes at many horizons within both formations. The suspicion that the peculiar lithology of the Colgate member of the Fox Hills sandstone might be due to a content of volcanic ash is disproved by petrographic examination.

INTRODUCTION

Many papers have been published on the problem of the Cretaceous-Tertiary boundary in the Western Interior of the United States, but many points nevertheless remain in dispute. The relation between the Fox Hills and Lance formations, a cause of particularly wide divergence of opinion, has been of especial interest to the writers because of intimate concern with it in the field, and they have ventured here to review the matter and to present certain observations and opinions regarding it.

The field work upon which the paper is based, carried out at different times during the years 1921 to 1926, covered most of the region where the Lance occurs. It included studies of the contact of the Fox Hills and Lance formations in areas that were mapped in detail; visits to localities cited by previous investigators as showing angular discordance or other marked evidence of unconformity between the formations; and examinations of the strata in areas where beds of Fox Hills age are present in formations known by other names and are overlain by nonmarine beds that are variously named.

For historical details and references to the literature of the subject the reader may consult recent papers by Knowlton,[1] Thom and Dobbin,[2] and Ward.[3]

QUESTIONS INVOLVED AND GENERAL CONCLUSIONS

In the northern Great Plains region the standard section of late Cretaceous and early Tertiary deposits contains the following formations, in ascending order: Pierre shale, Fox Hills sandstone, Lance formation, Fort Union formation, and Wasatch formation. The Pierre shale contains a large marine fauna, including such characteristic ammonites as *Placenticeras*, scaphites of the group of *Acanthoscaphites? nodosus*, and baculites of the large species *B. ovatus*, *B. compressus*, and *B. grandis*. The uppermost Pierre includes a group of scaphites of the genus *Discoscaphites* and certain other fossils that are more abundant in and more characteristic of the Fox Hills. The Fox Hills has a large marine fauna, of which the most distinctive fossils belong to the ammonite genus *Sphenodiscus*. At some places the uppermost Fox Hills contains brackish-water species which are found also in the lower part of the overlying Lance. The Lance also contains a great variety of fresh-water invertebrates, some reptiles, including especially the dinosaurian genus *Triceratops*, a few species of mammals, and a large flora. In a small area the upper part of the Lance contains a marine fauna similar to that of the Fox Hills but somewhat modified and without cephalopods. The Fort Union formation has yielded a small but characteristic fauna of fresh-water invertebrates, a fairly large fauna of primitive mammals, and a large flora. The Wasatch formation has yielded a small fauna of fresh-water invertebrates, a large fauna of mammals of more modern type, and a flora like that of the Fort Union formation.

It has been generally agreed that the Pierre shale and the Fox Hills sandstone are Cretaceous and that

[1] Knowlton, F. H., The Laramie flora of the Denver Basin: U. S. Geol. Survey Prof. Paper 130, pp. 3-82, 1922.

[2] Thom, W. T., jr., and Dobbin, C. E., Stratigraphy of the Cretaceous-Eocene transition beds in eastern Montana and the Dakotas: Geol. Soc. America Bull., vol. 35, pp. 500-506, 1924.

[3] Ward, Freeman, The Lance problem in South Dakota: Am. Jour. Sci., 5th ser., vol. 7, pp. 65-68, 1924.

the Wasatch formation is Eocene. Concerning the intermediate formations—the Lance and the Fort Union—three views have been expressed. One view is that the base of the Wasatch formation is the proper plane of division between the Cretaceous and the Eocene because the Fort Union is apparently inseparable from the Lance, which contains dinosaurs, whereas the Wasatch contains modern types of mammals, and because a pronounced discordance at many places indicates pre-Wasatch orogeny. Another view is that the contact of the Fort Union and Lance formations indicates the boundary between the Cretaceous and the Eocene because the Lance formation contains the latest known dinosaurs, the Lance invertebrates are more like those of the Cretaceous, and the marine member of the Lance, which is the final marine deposit in the whole Western Interior region, is Cretaceous in aspect, though at most places the contact with the Fort Union appears transitional. A third view is that the Lance and Fox Hills are separated by a hiatus of great magnitude, which is the first after a great unbroken sequence of Cretaceous deposits and which therefore constitutes the proper plane of division between the Cretaceous and the Eocene.

The question whether the Lance and Fox Hills formations are or are not separated by a great unconformity therefore becomes an important element in the general problem, and the history of the view that such an unconformity exists will be of interest here. This view arose from a comparison of the stratigraphy in northeastern Wyoming, eastern Montana, and adjoining portions of North and South Dakota with the stratigraphy of the Denver Basin in Colorado, where the detailed studies of Emmons, Cross, and Eldridge [4] were believed to have established the following propositions: The Cretaceous sea of the Western Interior invaded a region of low relief; there were no upward movements and no eroded areas of great size within the borders of this region until the end of Cretaceous time; there was, near the end of Cretaceous time, a passage into conditions of nonmarine sedimentation expressed by the Laramie formation; then followed an inception of mountain building over the present site of the Rocky Mountains and a period of active erosion, during which the sediments at places were stripped down to the pre-Cambrian; and a new set of deposits was then formed, resting in the Denver Basin itself on the eroded Laramie and beginning with the conglomerate of the Arapahoe formation.

The nonmarine Laramie formation was reported to contain a distinctive flora and a brackish and fresh water invertebrate fauna and to rest conformably upon the highest marine Upper Cretaceous beds in the basin, the Fox Hills sandstone. The conglomerate of the Arapahoe formation was reported to contain pebbles derived from all the older rocks in the adjacent mountains, indicating the long period of mountain making and erosion that preceded. The succeeding Denver formation was reported to consist largely of andesitic débris not seen in the Arapahoe, a fact that implies the elapse of the time necessary for the extrusion and erosion of the volcanic materials after the deposition of the Arapahoe. The Denver and Arapahoe formations were reported to contain a distinctive flora and a dinosaurian fauna that included the genus *Triceratops*.

In early discussions [5] the Arapahoe and Denver were referred to the Tertiary, and the Laramie was left in the Cretaceous. In the monograph on the Denver Basin, in consideration of the dinosaurian fauna and despite the great hiatus believed to exist beneath the Arapahoe, both Arapahoe and Denver were placed in the Cretaceous. In later papers, chiefly because of the evidence of diastrophism between Laramie and Arapahoe time, the Laramie was considered by Cross [6] to be the uppermost Cretaceous and the Arapahoe and Denver formations to be Tertiary. The *Triceratops* fauna in the post-Laramie beds was considered a relic of the Mesozoic fauna that lived on into Tertiary time.

In the 20 years that followed the publication of the Denver Basin monograph the Fox Hills sandstone and overlying strata at many places in northeastern Wyoming, eastern Montana, and western North and South Dakota were examined in detail by geologists of the United States Geological Survey, who did not find a thick Fox Hills formation, a typical Laramie flora, nor a conglomerate similar to the Arapahoe of the Denver Basin. On the contrary, a *Triceratops* fauna and a flora identified as like that of the Fort Union formation, which is widely accepted as Tertiary, were found in the Lance formation, directly above the relatively thin Fox Hills sandstone, and in the sandstone was found the same distinctive Cretaceous fauna as in the Denver Basin. For some geologists the only answer to the query thus raised seemed to be that the "*Triceratops* beds" (the Lance formation) are separated from the underlying marine Fox Hills sandstone by the same great unconformity as that supposed to be marked by the conglomerate of the Arapahoe formation in the Denver Basin; that erosion was greater in the north than in the Denver Basin; and that in the north there are no strata equivalent in time to the Laramie formation and part of the Fox Hills formation. In the north, however, no suitable plane of unconformity was at first observed, owing to the absence of a conglomerate similar to that of the Arapahoe. Further search revealed local details that were interpreted by some geologists as proof of a widespread and at some

[4] Emmons, S. F., Cross, Whitman, and Eldridge, G. H., Geology of the Denver Basin in Colorado: U. S. Geol. Survey Mon. 27, 1896.

[5] See, for example, Cross, Whitman, The Denver Tertiary formation: Am. Jour. Sci., 3d ser., vol. 37, pp. 261-282, 1889.

[6] Cross, Whitman, The Laramie formation and the Shoshone group: Washington Acad. Sci. Proc., vol. 11, pp. 27-45, 1909; Are the Lance and Fort Union formations of Mesozoic time?: Science, new ser., vol. 53, pp. 304-308, 1921.

places angular unconformity between the Fox Hills and Lance formations.

The interpretation of geologic history given above, the assignment of a great time value to the unconformity in the Denver Basin, and the evidence of the presence of a corresponding unconformity farther north have been vigorously disputed, and a list of papers contributed to the discussion is long. The strongest support for the theory of a large hiatus beneath the Lance formation lies in the supposed existence of an angular unconformity due to deformation and erosion, and no other specific explanation of some of this important physical evidence has yet been published.

In brief, then, the specific question to be considered here is whether the Lance formation rests conformably on the Fox Hills sandstone, or whether there is immediately beneath the Lance a great hiatus that represents the time required for the deposition of a considerable thickness of beds and for the occurrence of notable events in the geologic history of the Western Interior region. The writers' answer to this question is that there is nowhere in the areas considered a great unconformity between the Fox Hills and Lance formations and at most localities not even a minor one.

CRITERIA FOR SEPARATION OF THE FOX HILLS AND LANCE FORMATIONS

In a consideration of the nature of the boundary between the Lance and Fox Hills formations one must first determine what criteria shall be used for identifying the two formations and thus settle upon a means of locating the boundary between them. In general, minor individual lithologic units are not persistent in either the Fox Hills or the Lance. On the other hand, larger lithologic units, aggregations of these smaller, variable units, persist over large areas and are trustworthy features for identification. In some places, however, there is little lithologic distinction between beds that contain characteristic Lance species and others that contain characteristic Fox Hills species, and in these places the fossils must be made the chief reliance for separating the formations.

In eastern Montana, western North and South Dakota, and northeastern Wyoming the Fox Hills formation usually consists of an upper white sandstone (in eastern Montana called the Colgate sandstone member), beneath which occur alternating beds of gray, yellow, and buff sandstone and sandy shale. As noted on page 9, this formation contains a marine fauna, in part characteristic but in part similar to that found in the upper part of the underlying Pierre shale. Locally a brackish-water fauna that contains an admixture of marine forms occurs at the top of the formation. At and west of Glendive, Mont., the Colgate member contains only fossil plants, though the lower member of the formation contains a few marine invertebrates. Here both the lithology and the marine fauna are of assistance in identification.

In central and southern Wyoming and northwestern Colorado marine forms characteristic of the Fox Hills occur in sandstones that have been interpreted at some places as part of the Lewis shale and at others as part of the overlying beds, variously named, but there is no definite, separable Fox Hills formation. The beds immediately below the horizon of these Fox Hills fossils are in part nonmarine, and the beds above are all nonmarine. Here the marine fossils alone are characteristic.

With the exception of the Cannonball marine member in a small area in North and South Dakota, the Lance formation of Montana, the Dakotas, and eastern Wyoming is of nonmarine origin. The lower part is made up of somber-colored sandy shale alternating with lenticular beds of buff or brown sandstone and thin beds of coal. The contained fossils include plants, reptiles, especially dinosaurs of the genus *Triceratops*, and a few mammals. The fact that vertebrate remains are especially abundant near the base of the Lance provides an excellent criterion for separating it from the underlying Fox Hills formation in areas where the lithology is similar, but at many places the lithology also is of great service. The upper 75 to 350 feet of the Lance usually consists of soft buff and cream-colored calcareous sandstone and shale with numerous beds of coal.

In central Wyoming the Lance formation consists of interbedded fine-grained tan sandstone, light-gray shale, and thin layers of carbonaceous shale and coal and is not conveniently divisible into members. It contains, however, the same nonmarine fossils as the Lance of the areas to the east and north.

In southern Wyoming and northwestern Colorado beds believed by the writers to be the equivalent of the Lance formation, though known by several local names, have the same lithology as the Lance of central Wyoming.

FOX HILLS AND LANCE FORMATIONS IN DIFFERENT REGIONS

In setting forth the local features on which is founded the general conclusion that there is nowhere an important unconformity between the Lance and the Fox Hills the data are presented by convenient geographic units, recognizable on Figure 1.

MOUTH OF CANNONBALL RIVER, N. DAK.

In the area near the mouth of the Cannonball River the gray-white sandstone at the top of the Fox Hills forms a variable upper part of an upper member of the formation. The remainder of the member consists of light-gray clayey sandstone and olive-colored sandy shale; the two parts together attain a thickness of 100 feet. Above the white sandstone and to all appearances conformable with it are thick beds of impure lignite and carbonaceous shale of the Lance formation. (See pl. 4, *A*.) The lower member of the Fox Hills in this area is a thick-bedded yellowish-brown sandstone about 100 feet thick. Further details regarding the character of the Fox Hills in this area are given by Leonard.[7] A fairly representative section of the beds exposed on the Cannonball River is as follows:

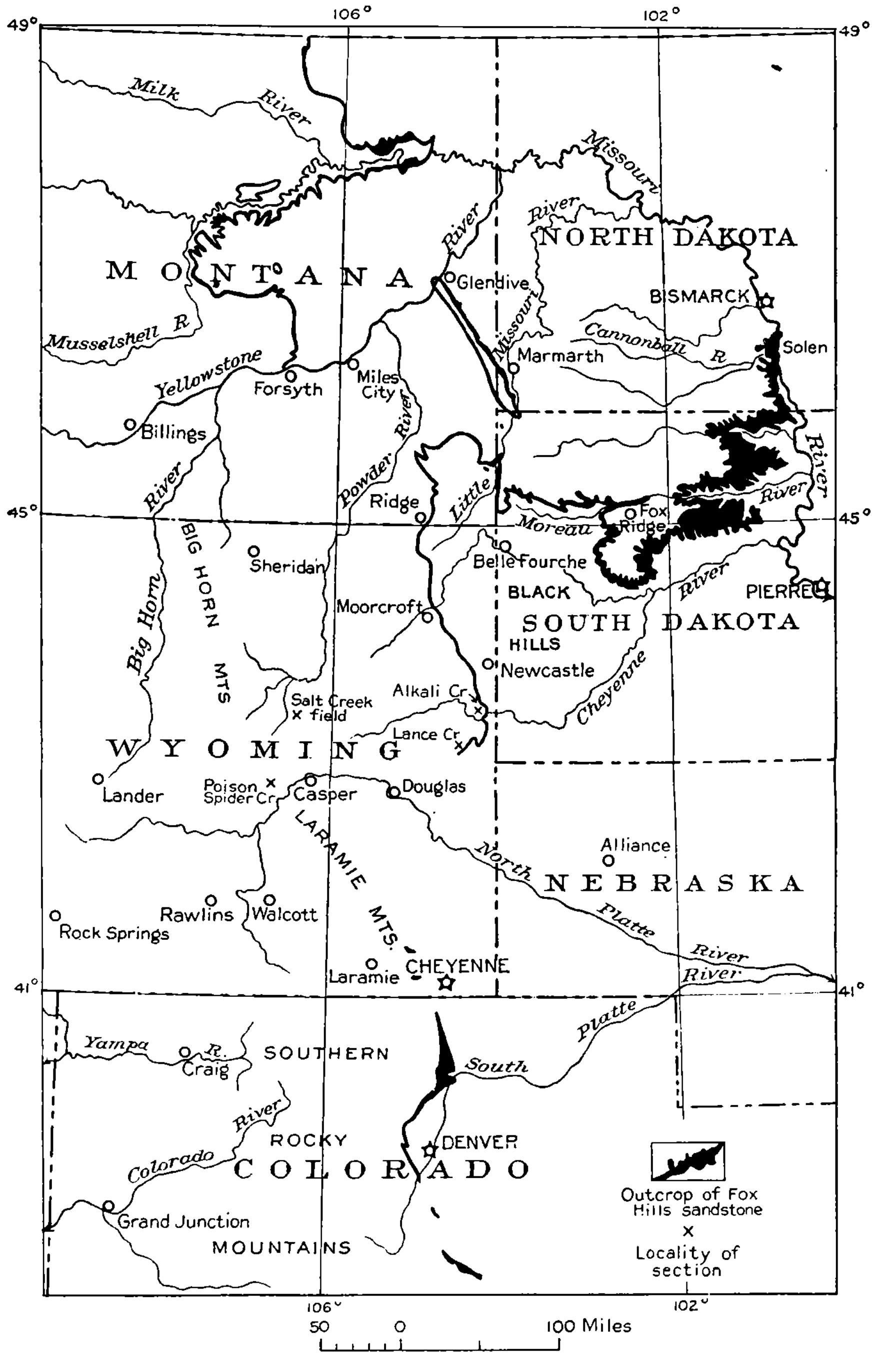


FIGURE 1.—Index map showing distribution of the Fox Hills sandstone. Certain outcrops in central South Dakota and eastern Colorado and beds of Fox Hills age included in other formations in central Wyoming and western Colorado are not shown

Section of parts of Fox Hills and Lance formations on south bank of Cannonball River 3½ miles west of Solen, N. Dak.

	Feet
Lance formation (basal part):	
Irregular assemblage of purple and gray shale and gray sandstone	60
Lignite	1
Purple carbonaceous shale	2
Fox Hills formation:	
Sandstone, white, more resistant than inclosing beds; contains many casts of roots	3
Sandstone, grayish white, soft; weathers into fluted channels; contains many layers and concretions of ferruginous brown sandstone and *Halymenites major* throughout (see pl. 4)	60
Large concretions of cinnamon-brown sandstone	2
Shale, olive-colored, sandy	50
Sandstone, brown, thick bedded, several feet; top of lower member.	
Level of the Cannonball River.	

Both members of the Fox Hills contain throughout *Halymenites major*, a fossil believed to be a usable criterion of marine origin. The lower member at the west end of the railroad cut one-third of a mile west of Solen, N. Dak., yielded, at a horizon 30 feet below the top, *Pteria nebrascana* Evans and Shumard, *P. linguaeformis* Evans and Shumard, *Tellina scitula* Meek and Hayden, and *Mactra formosa* Meek and Hayden, all of which are long-ranging species that serve only to confirm the indication of *Halymenites* that the beds are marine.

The Fox Hills near the mouth of the Cannonball River, therefore, shows two members and is apparently conformable beneath the Lance beds. Some hundreds of feet higher in the section the Cannonball marine member of the Lance appears with a large fauna so similar to that of the Fox Hills as to suggest that no long period could have intervened in addition to that during which the lower Lance sediments were being laid down.

CANNONBALL RIVER TO FOX RIDGE, S. DAK.

Observations made by the writers between the Cannonball River and Fox Ridge, the type locality of the Fox Hills formation, are in complete accord with the

[7] Leonard, A. G., The Cretaceous and Tertiary formations of western North Dakota and eastern Montana: Jour. Geology, vol. 19, pp. 511-514, 1911.

findings of Stanton,[8] who reports that the Fox Hills exhibits much evidence of irregular deposition, channeling between beds, and variability in thickness, but that it nowhere shows evidence of a major break at its top. The occurrence at isolated localities in this area at or near the top of the Fox Hills of beds that have an irregular base and contain a fauna of chiefly brackish-water species has led to the assumption of a widespread unconformity beneath these beds and the selection of this unconformity as the "pre-Tertiary" break.[9] The beds immediately above the "hiatus," however, contain a number of significant species of the Fox Hills fauna and preclude an interruption greater than that represented by any one of the admittedly minor erosion planes present in both Fox Hills and Lance formations. The fossils are delicate shells that would show any reworking very distinctly. Although the white sandstone of the upper Fox Hills was largely concealed along the route followed by the writers through this area, its presence locally is attested by Calvert, who says:[10] "In some areas the lignite-bearing beds of the Lance abruptly change to the white marine sandstone of the Fox Hills, but in others the rocks of the two formations are very similar and no hard and fast line can be drawn between them."

Particular attention was paid to the locality on Worthless Creek (locally known as Irish Creek), in the SE. ¼ sec. 23, T. 16 N., R. 20 E., where Calvert[11] observed what he regarded as an angular unconformity between the Fox Hills and the Lance that is cited a number of times in the literature.[12] Calvert's description is as follows:

> On Worthless Creek, in T. 16 N., R. 20 E., where exposures are especially good, the most striking example of unconformity between the Fox Hills and Lance formations was observed. On the west side of the Worthless Creek Valley, near the line between secs. 25 and 26, the "somber beds" of the Lance formation transgress across the Fox Hills sandstone, and the upper part of the Fox Hills down to the banded shale is absent. The unconformity at this locality is angular as well as erosional, for the banded shale dips 4° N., whereas the Lance is horizontal. A section of the strata 500 feet long reveals the Lance filling a channel eroded in the banded shale of the Fox Hills to a depth of 40 feet, so that the vertical amount of combined transgression and erosion is at least that amount. On the opposite side of the valley the undoubted Fox Hills is believed to be absent and the lignitic zone of the Lance rests on a brown banded shale, which may represent the top of the Pierre or may be part of the Fox Hills. In any event, there is surely less than 25 feet of Fox Hills present at this place. In view of the fact that the Fox Hills sandstone is normally at least 150 feet thick, it seems that the time during which erosion took place was of considerable duration.

The writers visited this locality early in May, when the outcrops were not hidden by vegetation and it was possible to study the structural relations in detail. A most casual inspection under such conditions shows that the "unconformity" is due to a small fault which has displaced a carbonaceous shale bed about 20 feet, the downthrown bed dipping into the fault at an angle of about 4°. (See fig. 2.) Dinosaur bones at the base of the bluffs indicate that the beds in the exposure are Lance rather than Fox Hills, and this fact explains why Calvert failed to find Fox Hills beds on the opposite side of the valley. There is probably a second small fault a few hundred feet away, though its presence is not very clearly shown.

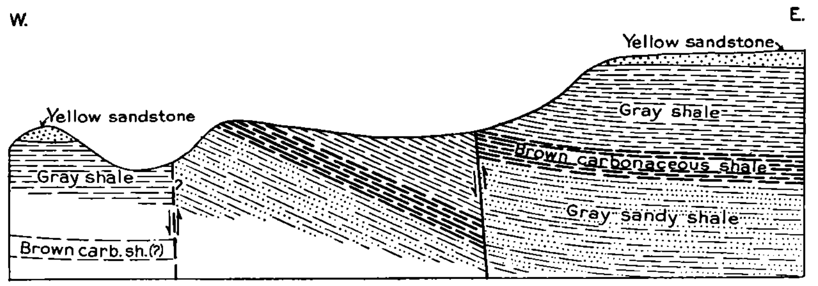


FIGURE 2.—Diagram of faulting in Lance formation on Worthless (Irish) Creek, in the SE. ¼ sec. 23, T. 16 N., R. 20 E., S. Dak. The throw is about 20 feet

Wilson and Ward[13] report that the contact of the Fox Hills and Lance is transitional in the Worthless Creek district and that the lowest bed of coal or carbonaceous shale in the Lance occurs just above an oyster bed that marks the top of the Fox Hills over a rather broad area. That small faulting is common in the whole area is also evident from their work and from that of Russell.[14] A small fault of the same type as the one at Worthless Creek is clearly shown on the north bank of the Moreau River about 11 miles east of Bixby, in sec. 15, T. 14 N., R. 14 E., where a throw of 35 feet introduces a local dip of 4° in beds that probably constitute the upper part of the Fox Hills. An example of irregularities of deposition is well shown on the bank of the Moreau River at Bixby. The upper part of the exposure may be Lance, but even so the relations between the irregular beds are so intricate that it is unreasonable to pick out any plane in the sequence as more significant than the rest.

[8] Stanton, T. W., Fox Hills sandstone and Lance formation ("*Ceratops* beds") in South Dakota, North Dakota, and eastern Wyoming: Am. Jour. Sci., 4th ser., vol. 30, pp. 173-182, 1910.

[9] Knowlton, F. H., Further data on the stratigraphic position of the Lance formation ("*Ceratops* beds"): Jour. Geology, vol. 19, pp. 362-369, 1911.

[10] Calvert, W. R., and others, Geology of the Standing Rock and Cheyenne River Indian Reservations, N. and S. Dak.: U. S. Geol. Survey Bull. 575, p. 18, 1914.

[11] Idem, p. 18.

[12] Knowlton, F. H., Further data on the stratigraphic position of the Lance formation ("*Ceratops* beds"): Jour. Geology, vol. 19, p. 346, 1911; Cretaceous-Tertiary boundary in the Rocky Mountain region: Geol. Soc. America Bull., vol. 25, p. 327, 1914.

[13] Wilson, R. A., and Ward, Freeman, The possibilities of oil in northern Ziebach County: South Dakota Geol. and Nat. Hist. Survey Circ. 13, pp. 6-7, 1923.

[14] Russell, W. L., The possibilities of oil in western Ziebach County: South Dakota Geol. and Nat. Hist. Survey Circ. 20, pp. 12-23, 1925.

FOX RIDGE TO LITTLE MISSOURI RIVER, S. DAK.

From Fox Ridge westward to the Little Missouri River the contact of the Fox Hills and Lance is, as a rule, exposed only in the more or less isolated bluffs along stream valleys. Practically all outcrops of both formations near their contact show local channeling, cross-bedding, faulting, and slumping on a small scale. The Fox Hills is about 125 feet thick and consists of light-colored marine sandstone which is overlain by the basal coal-bearing beds of the Lance. No distinct or persistent division into a lower yellow sandstone and an upper white sandstone could be recognized, though it must be admitted that the outcrops are scarcely adequate.

One locality that received particular attention in the area west of Fox Ridge was that in the SE. ¼ sec. 15, T. 15 N., R. 8 E., South Dakota, 1 mile south of Govert post office, which has been reported by several writers as showing an angular unconformity between the Fox Hills and the Lance.[15] The exposure at this locality is shown in Plate 4, *B*, and exhibits the following section:

Section on bank of Moreau River near Govert, S. Dak.

	Feet
Dark-colored wash and soil	10
Recent sand	6
Clay, gray	5
Sandstone and sandy shale with topset and foreset bedding	18
Talus.	

A detailed examination of this exposure shows that the inclined bedding planes represent merely foreset bedding and are not a result of folding and erosion. The sketch in Figure 3, made to scale, gives the details of bedding not clearly visible in the photograph. No evidence could be obtained to indicate the age of the individual beds in this outcrop, and it is not possible to assign them definitely to either the Lance formation or the Fox Hills sandstone.

The sandstone and conglomerate capping some of the high buttes on the divide between the Belle Fourche and Owl Rivers, northeast of Belle Fourche, S. Dak., which have been doubtfully called the Fox Hills sandstone,[16] are of White River (Oligocene) age, according to Rubey,[17] who has examined them in detail.

As a brief summary, it may be said that all data obtained by the writers regarding the contact of the Fox Hills and Lance in South Dakota are in accord with the findings of Ward,[18] who says: "It seems plain that the evidence points consistently and convincingly to the fact that there is no sharp or well-defined break between the Fox Hills and Lance—certainly no break which would warrant a division between the Mesozoic and Cenozoic."

MARMARTH, N. DAK.

At the south end of the Cedar Creek anticline, in the vicinity of Marmarth, N. Dak., the contact of the Fox Hills and Lance exhibits all the characteristics observed farther east. On Little Beaver Creek in the SE. ¼ sec. 7, T. 132 N., R. 106 W., the transition zone between the typical Pierre shale and the basal Fox Hills sandstone is about 40 feet thick. Through this zone the rock grades upward into a soft yellowish massive sandstone about 50 feet thick, which is overlain by the grayish-white sandstone that is seen farther east and is now known in eastern Montana as the Colgate sandstone member of the Fox Hills formation. The top of the Colgate sandstone shows local channels (see pl. 5, *A*), which have been interpreted to indicate an unconformity.[19] The section at this locality is as follows:

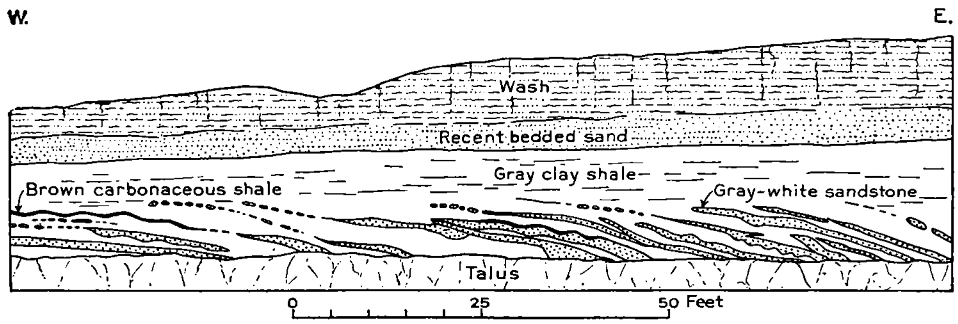


FIGURE 3.—Details of bedding of outcrop of Fox Hills sandstone (?) in the SE. ¼ sec. 15, T. 15 N., R. 8 E., near Govert, S. Dak. Length of section shown, 125 feet, drawn to scale. Part of this outcrop is shown in Plate 4, *B*

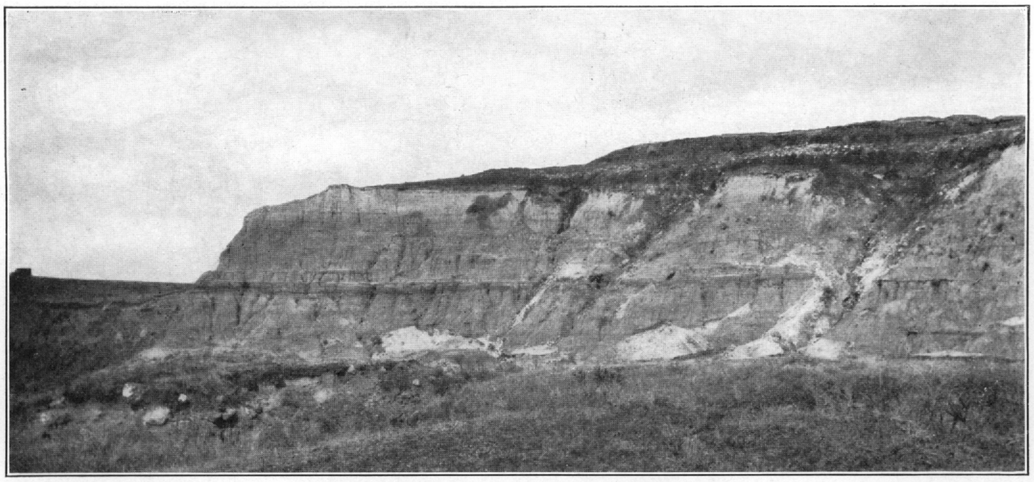
[15] Knowlton, F. H., Further data on the stratigraphic position of the Lance formation ("*Ceratops* beds"): Jour. Geology, vol. 19, p. 367, 1911; Cretaceous-Tertiary boundary in the Rocky Mountain region: Geol. Soc. America Bull., vol. 25, p. 327, 1914. Calvert, W. R., and others, op. cit., p. 19.

[16] Darton, N. H., U. S. Geol. Survey Geol. Atlas, Belle Fourche folio (No. 164), p. 5, 1909.

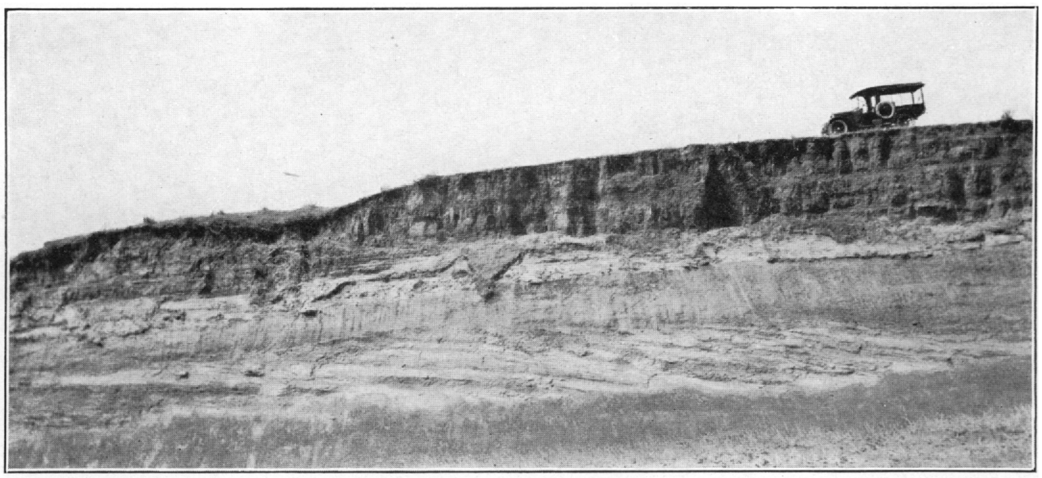
[17] Rubey, W. W., personal communication.

[18] Ward, Freeman, The Lance problem in South Dakota: Am. Jour. Sci., 5th ser., vol. 7, p. 67, 1924.

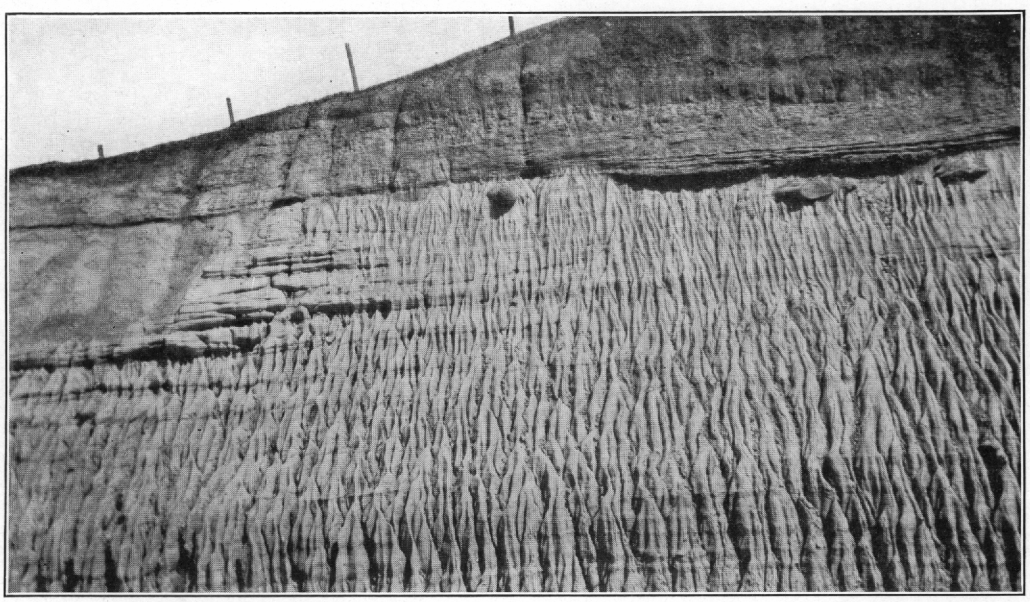
[19] Leonard, A. G., The geology of southwestern North Dakota with special reference to coal: North Dakota Geol. Survey Fifth Bienn. Rept., pp. 43-44, 1908. Knowlton, F. H., The stratigraphic relations and paleontology of the "Hell Creek beds," "*Ceratops* beds," and equivalents and their reference to the Fort Union formation: Washington Acad. Sci. Proc., vol. 11, pp. 201-202, 1909.



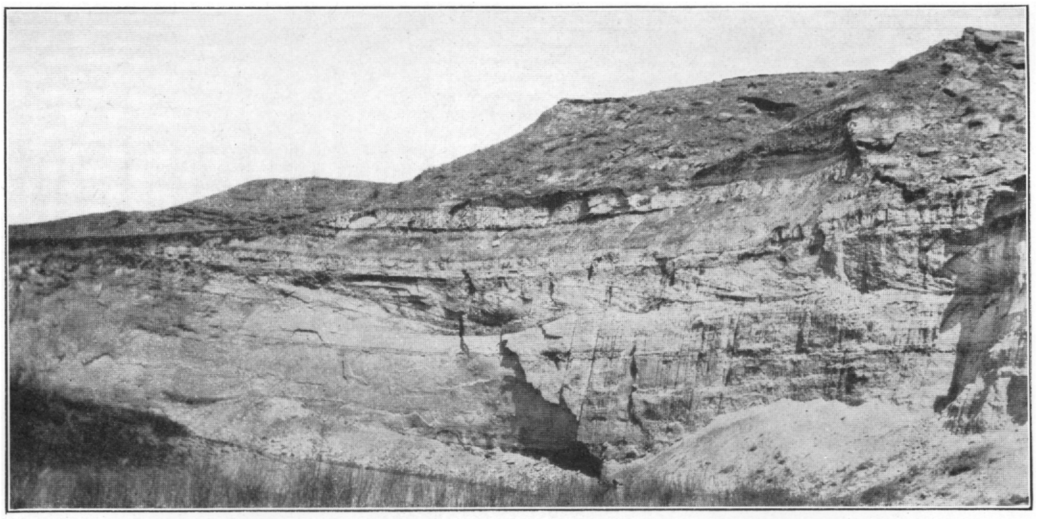
A. GRAYISH-WHITE SANDSTONE AT THE TOP OF THE FOX HILLS SANDSTONE IN THE SOUTH BANK OF THE CANNONBALL RIVER 3½ MILES WEST OF SOLEN, N. DAK.



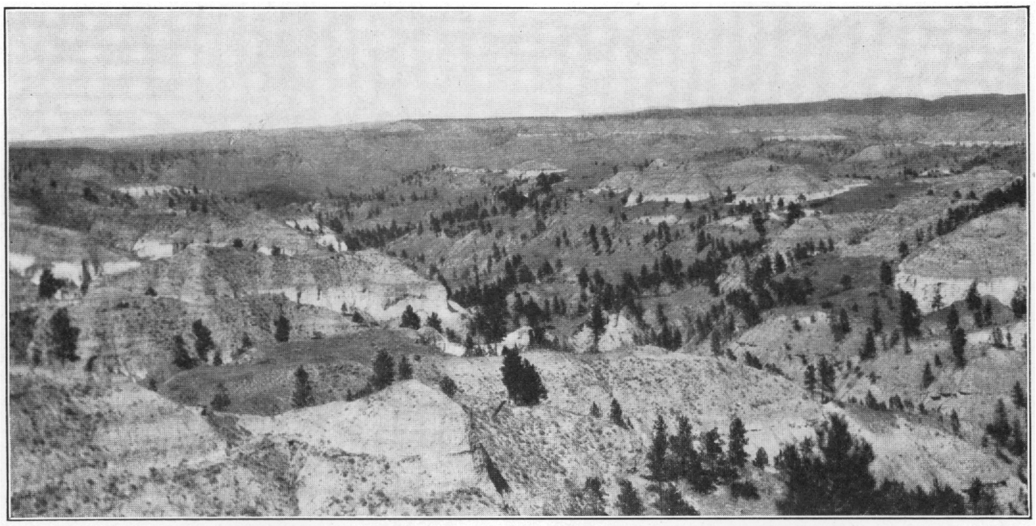
B. CROSS-BEDDING IN THE FOX HILLS SANDSTONE (?) IN THE SE. ¼ SEC. 15, T. 15 N., R. 8 E., NEAR GOVERT, S. DAK.



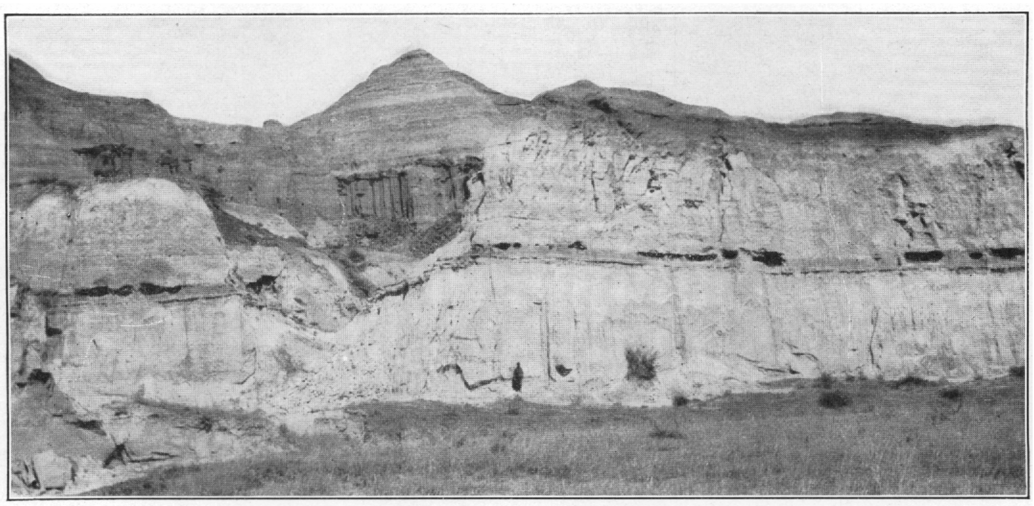
C. FLUTINGS IN THE WEATHERED SURFACE OF THE COLGATE SANDSTONE MEMBER OF THE FOX HILLS SANDSTONE IN THE RAILROAD CUT 3½ MILES WEST OF MARMARTH, N. DAK.



A. LOCAL CHANNELING IN THE FOX HILLS SANDSTONE ON LITTLE BEAVER CREEK IN THE SE. ¼ SEC. 7, T. 102 N., R. 106 W., NORTH DAKOTA



B. DEVIL CREEK, IN T. 21 N., R. 32 E., MONTANA, SHOWING THE WHITE COLGATE SANDSTONE AND ASSOCIATED BEDS



C. COLGATE SANDSTONE 4 MILES SOUTHWEST OF GLENDIVE, MONT.

Section in the Fox Hills and Lance formations on Little Beaver Creek in the SE. ¼ sec. 7, T. 132 N., R. 106 W., N. Dak.

[See pl. 5, *A*]

	Feet
Lance formation (lower part):	
Sandstone, light yellowish brown, coarse grained; contains scattered fragments of carbonized wood	1
Sandstone, yellowish brown, soft, cross-bedded, medium grained; contains large copper-colored concretionary masses that are hard and weather out unbroken	15
Sandstone, shaly, and sandy shale, light gray to drab, soft; contains thin crusts of limonitic material and small limonitic concretions; cross-bedded	22
Sandstone, rust-colored, coarse grained, much cross-bedded; contains many siderite concretions; at some places hard, at others soft	11
Sandstone, soft, light gray; contains rust-colored, iron-stained masses that are hard and contain many plant impressions	3
Shale, sandy, and shaly sandstone, brown, carbonaceous	5
Shale, black, carbonaceous	4
Shale, brown, carbonaceous	6
Sandstone, gray white, shaly, coarse grained, cross-bedded locally but otherwise massive	7
Shale, brown, carbonaceous, irregular	1
Shale, drab to light gray	7
Fox Hills formation:	
Colgate sandstone member (part)—	
Sandstone, light gray, soft; contains many fossil roots of plants	5
Shale, drab to light gray, with very irregular base	2
Sandstone, gray-white, and gray shale, very much cross-bedded and irregular	8
Sandstone, gray-white, massive to rather indistinctly bedded, shaly in places; lower 10 feet streaked with iron stains; locally contains irregular thin lenses of carbonaceous shale. *Halymenites major* throughout. Forms the smooth wall above creek level shown in Plate 5, *A*	40
Shale, carbonaceous	6
Creek level.	

In the railroad cut 3½ miles west of Marmarth the Colgate sandstone is about 30 feet thick, massive, locally cross-bedded, and overlain by a carbonaceous shale bed 1 foot thick. (See pl. 4, *C*.) About 5 miles south of Marmarth, in the NE. ¼ sec. 32, T. 132 N., R. 106 W., the Colgate is again well exposed and contains abundant remains of *Halymenites major*. Other details regarding the Fox Hills and Lance formations in the Marmarth area are given by Stanton,[20] with whose conclusions the writers concur—that "it is most probable that the abrupt change from marine to fresh-water and land conditions seen near Marmarth is purely local and that the eroded surface at the top of the Fox Hills does not represent a time interval of any geologic importance."

[20] Stanton, T. W., Fox Hills sandstone and Lance formation ("*Ceratops* beds") in South Dakota, North Dakota, and eastern Wyoming: Am. Jour. Sci., 4th ser., vol. 30, pp. 182-184, 1910.

MARMARTH, N. DAK., TO GLENDIVE, MONT.

In the SW. ¼ sec. 34, T. 7 N., R. 61 W., Mont., about 1 mile west of the North Dakota line, the Colgate sandstone displays a clean-cut contact with the overlying horizontal Lance beds. The sandstone is cross-bedded and channeled on a small scale, but the significance of the unconformity is minimized by the fact that at short distances away the contact is not sharp nor discordant.

Other unconformities between the Fox Hills and Lance formations, reported by Knowlton [21] on the authority of Calvert to occur in sec. 22, T. 6 N., R. 60 E., and sec. 32, T. 7 N., R. 61 E., Montana, were not found in these sections. The writers examined every outcrop along the contact in T. 7 N. but could find no evidence of unconformity other than cross-bedding and purely local channeling of a kind present at many levels in both Fox Hills and Lance. As Knowlton was quoting Calvert, who supervised the mapping of the area by Bowen,[22] it is possible that a mistake was made somewhere in the process of handing on the references and that it was intended to cite the localities as in secs. 27 and 32, T. 6 N., R. 60 E., as said elsewhere by Calvert,[23] though Bowen's map shows no Colgate in sec. 32, T. 6 N., R. 60 E. Sec. 32, T. 7 N., R. 61 E., is on the east side of the Cedar Creek anticline and not on the west side, as said by Knowlton. Published photographs [24] of one of these unconformities show local channeling and cross-bedding, such as characterize the upper part of the Fox Hills throughout its extent.

It has been stated that the Fox Hills and Lance are unconformable along the Cedar Creek anticline because in the Baker field the Colgate sandstone, which at the time the statement was made was considered a member of the Lance, "appears to rest directly on Pierre shale without the intervention of the lower strata so noticeable at Iron Bluff" (near Glendive).[25] The writers' examinations along the whole length of the anticline have convinced them that the basal brown sandstone of the Fox Hills is present throughout the greater part of the anticline, though it is replaced by gray sandy shale in a few localities, and that the upper white sandstone contains *Halymenites* at many places and is marine.

[21] Knowlton, F. H., Further data on the stratigraphic position of the Lance formation ("*Ceratops* beds"): Jour. Geology, vol. 19, p. 366, figs. 2, 3, 1911.

[22] Bowen, C. F., The Baker lignite field, Custer County, Mont.: U. S. Geol. Survey Bull. 471, pp. 202-226, 1912.

[23] Calvert, W. R., Geology of certain lignite fields in eastern Montana: U. S. Geol. Survey Bull. 471, p. 195, 1912.

[24] Knowlton, F. H., op. cit., p. 368.

[25] Calvert, W. R., Geology of certain lignite fields in eastern Montana: U. S. Geol. Survey Bull. 471, p. 195, 1912.

The rocks near Glendive were first studied by Leonard,[26] who says:

In the Glendive region there are no beds above the marine Pierre that correspond to the supposed Fox Hills clay and sandstone of Hell Creek. In fact, the line of contact between the Pierre and the overlying dinosaur-bearing beds, while not discordant so far as structure is concerned, may possibly represent a time break in which most of the upper fresh and brackish water beds of the Cretaceous are wanting.

It may be seen from this statement that all the rocks above the Pierre shale were included by Leonard in the dinosaur-bearing beds, the Lance formation. Leonard's section is as follows:

35 feet thick is 150 feet stratigraphically above Pierre shale. In the Baker field, however, this white sandstone, which Leonard states constitutes the most prominent stratum in the region, appears to rest directly on Pierre shale without the intervention of the lower strata so noticeable at Iron Bluff.

Observations made by the writers at Iron Bluff and vicinity show that the section published by Leonard and quoted by Calvert is in error. It was assumed by Leonard that the sandstone (No. 3 of Leonard's section) capping Iron Bluff (see fig. 4) dips eastward beneath the Colgate sandstone in the bluffs east of Sand Creek. It is clear, however, that the Colgate is present in Iron Bluff beneath the top of Leonard's No.

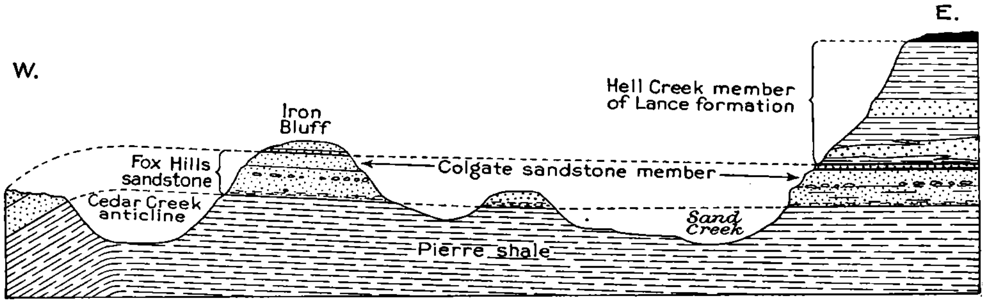


FIGURE 4.—Diagrammatic cross section across the Cedar Creek anticline at Iron Bluff, near Glendive, Mont.

Section of rocks at Iron Bluff and vicinity

	Feet
8. Coal bed, burned but probably 6 feet thick.	
7. Shale with a few thin beds of sandstone; abundant collection of fossil plants in sandstone bed 20 feet from base	150
6. Sandstone, massive, gray	40
5. Shale and sandstone; a few fossil plants at base	160
4. Sandstone, white, massive; most prominent stratum in the region	35
3. Sandstone, brown; fossil leaves in bottom part; forms summit of Iron Bluff	75
2. Shale and sandstone; fossil leaves in upper 20 feet	75
1. Shale, dark, Pierre, with limestone concretions containing abundant marine fossils; exposed to river level	100
	635

Calvert [27] later grouped Nos. 2, 3, and 4 of Leonard's section into the Colgate sandstone member of the Lance formation and wrote:

Although there is in the Iron Bluff section an appearance of transition from the Pierre shale into the overlying arenaceous strata, which suggests that the sandstone occupies the stratigraphic position of the Fox Hills, the evidence of fossil leaves indicates that much if not all of it is of later age.

Referring to the finding in units 2 and 3 of fossil plants that were identified by Knowlton as Tertiary species, he says:

It is certain that if the Fox Hills sandstone is present in this section it is restricted to the 70-foot interval between the location of the fossil plants and the top of the Pierre shale. * * * Southward from Iron Bluff the stratigraphy suggests even more strongly that the Fox Hills is not present. In the section near Iron Bluff, as given by Leonard, a white sandstone

3, as shown in the section on the east face of Iron Bluff given below. (See also fig. 4.) On the north face of Iron Bluff the Colgate is much discolored by brown ferruginous stain leached from the overlying rust-brown sandstone, a lens in the basal Lance which does not seem to extend eastward across Sand Creek. On Sand Creek the upper part of the lower member of the Fox Hills yielded marine fossils.

In summary, it may be said that near Glendive the Fox Hills sandstone maintains all the characteristics observed farther east, except that the Colgate member contains well-preserved fossil plants, which have been identified by Knowlton as Fort Union (Tertiary) species.[28]

As shown in Plate 5, *B*, beds of lignite also occur in the Colgate member and in the Hell Creek member, immediately above. Sections at Iron Bluff and Sand Creek are as follows:

Section of the Fox Hills and Lance formations at Iron Bluff, southwest of Glendive, Mont.

	Feet
Lance formation (part of Hell Creek member or "somber beds"):	
Sandstone, brown, platy; makes top of Iron Bluff	4
Concealed, probably like unit above	20
Sandstone, light gray with rusty stain locally, fairly cross-bedded	6
Concealed, probably sandstone	11
Fox Hills sandstone:	
Colgate sandstone member—Sandstone, white, fine grained, massive; deeply stained at places by material leached from the overlying ferruginous sandstone	44

[26] Leonard, A. G., The coal fields of parts of Dawson, Rosebud, and Custer Counties, Mont.: U. S. Geol. Survey Bull. 316, pp. 197-198, 1907.

[27] Calvert, W. R., Geology of certain lignite fields in eastern Montana: U. S. Geol. Survey Bull. 471, pp. 194-196, 1910.

[28] Calvert, W. R., Geology of certain lignite fields in eastern Montana: U. S. Geol. Survey Bull. 471, p. 195, 1912.

	Feet
Fox Hills sandstone—Continued.	
Lower member—	
Sandstone, brown, containing many ferruginous nodules	11
Sandstone, soft, yellow, containing discontinuous layers of dark-brown platy sandstone	25
Sandstone, soft, yellow	9
Sandstone, light gray, fine grained, concretionary	1
Sandstone, soft, yellow	10
Sandstone, dark brown, platy and concretionary	1
Sandstone, dark colored, soft; weathers yellow	10
Sandstone, soft, shaly, light gray; weathers yellow	16
Shale, sandy, gray, stained with limonite	5
Shale, brown and gray, sandy	13
Total lower member	101
Pierre shale: Shale, gray, sandy, with many large fossiliferous concretions in upper 20 feet.	

Section of the Fox Hills sandstone on east side of Sand Creek, 4 miles southwest of Glendive, Mont.

	Feet
Lance formation (Hell Creek member or "somber beds").	
Fox Hills sandstone:	
Colgate sandstone member—	
Sandstone, white, fine grained, massive, sugary; crops out in a prominent white bluff and contains many well-preserved fossil plants identified as Fort Union (Tertiary) species	7
Coal, impure	2
Sandstone, like that above	45
Total Colgate member	54
Lower member—	
Sandstone, concretionary, lenticular	5
Sandstone, brown, limonitic, slabby, grading upward into soft yellow shaly sandstone and sandy shale; contains *Vanikoro ambigua*, *Fusus dakotensis*, and *Anomia?* sp. 15 feet below top	70
Total lower member	75
Pierre shale.	

NORTHEASTERN MONTANA

In western and northwestern Garfield County, Mont., the lower member of the Fox Hills consists of a massive yellow-brown coarse-grained sandstone that averages about 75 feet in thickness and makes a prominent wall along the tributaries to the Musselshell and Missouri Rivers. The white Colgate member overlies the lower member and can be seen for miles as its outcrop meanders in and out of numerous coulees. (See pl. 5, *C*.) The lower member is marine in this region, but no fossils were found in the Colgate member. East of Hell Creek the white Colgate member is replaced by friable yellow to brownish sandstone and sandy shale.

Channeling and cross-bedding are common in the Colgate member in this region, but during detailed mapping by Mr. Dobbin, extending over two field seasons, no evidence was found to indicate a major break at the top of the Fox Hills. This conclusion is in accord with the findings of Brown,[29] who says:

On the east fork of Crooked Creek near the old Cook ranch, on the west fork of Crooked Creek near the Gus Colin claim, and on the east fork of Hell Creek near the EE cattle camp, these marine beds (Fox Hills) have been eroded in places, sometimes to a depth of 10 feet, before the succeeding massive sandstones of the fresh-water "Lance" were deposited. The strata are, however, in all cases parallel to the bedding plane of the succeeding sandstones, and the break is evidently of local erosional character. * * * No sign of an angular unconformity has been noted between the Fox Hills and the "Lance," and I have never yet seen any geologic evidence of the "great diastrophic break" which is alleged to occur here.

The following section is typical of the Fox Hills sandstone along the Musselshell and Missouri Rivers in Garfield County:

Section of Fox Hills sandstone on Devil Creek, in the SE. ¼ sec. 16, T. 21 N., R. 32 E., Mont.

	Feet
Lance formation (Hell Creek member or "somber beds)."	
Fox Hills sandstone:	
Colgate sandstone member: Sandstone, white, fine grained, massive	30
Lower member:	
Shale, yellow, sandy, with thin beds of sandstone	11
Sandstone, concretionary	1
Sandstone, brown, hard; makes a ledge; contains *Lingula nitida* Meek and Hayden in abundance	2
Shale, yellow, sandy	5
Sandstone, yellow, fairly hard; makes a wall	15
Sandstone, brownish yellow	15
Shale, gray, sandy	4
Sandstone, yellow, soft, shaly at base	20
Total Fox Hills sandstone	103
Bearpaw shale (=upper Pierre).	

In T. 16 N., R. 38 E., Montana, about 60 miles north of Forsyth, Bauer[30] found quartzite and porphyry pebbles at a horizon considered by him the base of the Lance formation. The pebbles are half an inch to 3 inches or more in diameter and are embedded in impure limonite, which ranges in thickness from 3 to 8 inches. A detailed examination of this locality by Mr. Dobbin has shown that the conglomerate is conformable with the underlying beds in most of the exposures, though locally there are indications of channeling such as characterize the upper part of the Fox Hills throughout its extent. As the conglomerate occurs only in this small area, its presence affords little proof of the existence of a major unconformity at the contact of the Fox Hills and the Lance.

The relations between the Fox Hills and Lance formations in the region between the Missouri River and the Canadian boundary can not be satisfactorily determined, owing to the covering of late Tertiary and Pleistocene deposits. Studies by Smith,[31]

[29] Brown, Barnum, Cretaceous-Eocene correlation in New Mexico, Wyoming, Montana, Alberta: Geol. Soc. America Bull., vol. 25, pp. 357-358, 1914.

[30] Bauer, C. M., Quartzite pebbles at the base of the Lance formation in Montana: Am. Assoc. Petroleum Geologists Bull., vol. 9, pp. 344-346, 1925.

[31] Smith, C. D., The Fort Peck Indian Reservation lignite field, Mont.: U. S. Geol. Survey Bull. 381, p. 42, 1910.

Beekly,[32] and Collier [33] indicate that the beds tentatively called Fox Hills in this region are probably of continental origin, as shown by the lack of marine fossils and by the finding of remains of a trachodont dinosaur in a sandstone "which surely lies only a short distance above the Bearpaw" (=upper Pierre).[34] These remains were identified as belonging to the species *Trachodon annectens* Marsh, a form which also occurs in the Lance of the Hell Creek area. This occurrence is noteworthy in view of the fact that the Fox Hills contains marine fossils on the south bank of the Missouri River and also in Canada, only a few miles from the locality where the dinosaur remains were found.[35] Studies by McConnell [36] indicate that the Fox Hills and "Laramie" (Lance) are conformable also in the Cypress Hills, which lie but a short distance northwest of the Wood Mountain Plateau.

CENTRAL MONTANA

West and southwest of Forsyth, Mont., the Bearpaw shale (=upper Pierre) is conformably overlain by nonmarine sandy shale and sandstone, locally indistinguishable from the rest of the Lance formation. These beds are indubitably of fluviatile origin, yield characteristic Lance fresh-water invertebrates, and probably represent the purely continental phase of the marine Fox Hills of the Dakotas.

SOUTHEASTERN MONTANA

The Fox Hills formation in the southeast corner of Montana is similar to the outcrops near Marmarth, N. Dak., except that locally the white Colgate member is replaced by brown sandstone. Near Ridge the Colgate member is more resistant than in areas to the north and east and makes bluffs 60 feet high. Excellent exposures at certain localities in this area fail to show any evidence of an unconformity between the Fox Hills and Lance.

RIDGE, MONT., TO LANCE CREEK, WYO.

Between the Montana-Wyoming State line and Moorcroft, Wyo., the contact of the Fox Hills and Lance could not be studied in detail, because of the lack of exposures in areas of low relief. The lower part of the Fox Hills in this area consists of brown slabby marine sandstone and sandy shale, which crop out in a pronounced escarpment that stands well above the Pierre lowlands to the east. Rubey [37] has mapped this area in detail and reports that the Colgate sandstone member of the Fox Hills is either absent or present as some other lithologic facies—for example, a brown unconsolidated sandstone that occurs locally between the marine Fox Hills and the *Triceratops*-bearing Lance beds may be equivalent to the Colgate sandstone member.

On Good Lad Creek, less than a quarter of a mile below Oshoto post office, in the SW.¼ sec. 7, T. 53 N., R. 67 W., Wyoming, sandy beds that are possibly of Fox Hills age stand vertical in a cut bank, and 400 feet upstream, directly opposite Oshoto post office, beds of brown and gray sandstone containing prominent concretions, such as characterize the Lance, dip about 5° SW. The great decrease in dip within so short a horizontal distance has been throught by some to indicate a structural unconformity between the Lance and Fox Hills. An examination of the locality by the writers showed that the exposures are too poor to reveal the structure with certainty. It is believed, however, that the abrupt change in dip is due either to sharp unsymmetrical folding or to folding and faulting on a prominent monocline, which has been mapped by Rubey [37] for a distance of 30 miles.

In the same area as that described in the preceding paragraph Darton and O'Harra [38] reported the Fox Hills to be apparently 400 to 450 feet thick and to contain vertebrate bones, fossil wood, and species of *Inoceramus*. That the record of *Inoceramus* was due to a misidentification of *Unio couesi* which caused these writers to include a portion of the Lance containing this fossil in the Fox Hills sandstone has been proved by a subsequent examination of the precise locality described. The true thickness is much less than that reported.

At 4 miles southeast of Moorcroft the Fox Hills sandstone is well exposed in a conspicuous ridge but affords no evidence of an unconformity with the Lance. The Fox Hills is about 200 feet thick at this locality.

The Fox Hills is also well exposed in the escarpment 7 miles northwest of Upton, Wyo. Its top is a white sugary sandstone overlain by a bed of purple carbonaceous shale. The white sandstone contains *Halymenites major* and is possibly a correlative of the Colgate sandstone of southeastern Montana, though this can not be proved. Dinosaur bones were collected from the Lance beds less than 50 feet above the purple carbonaceous shale, and marine fossils were obtained in the lower member of the Fox Hills.

About 13 miles west of Newcastle the Fox Hills and Lance exhibit the following section:

[32] Beekly, A. L., The Culbertson lignite field, Valley County, Mont.: U. S. Geol. Survey Bull. 471, pp. 329–330, 1912.

[33] Collier, A. J., Geology of northeastern Montana: U. S. Geol. Survey Prof. Paper 120, pp. 30–31, 1919.

[34] Collier, A. J., op. cit., p. 31.

[35] Rose, Bruce, Wood Mountain coal area, Saskatchewan: Canada Geol. Survey Summary Rept. for 1914, p. 65, 1915.

[36] McConnell, R. G., Report on the Cypress Hills, Wood Mountain, and adjacent country: Canada Geol. Survey Ann. Rept., new ser., vol. 1, pt. C, p. 26 c, 1885.

[37] Rubey, W. W., personal communication.

[38] Darton, N. H., and O'Harra, C. C., U. S. Geol. Survey Geol. Atlas, Devils Tower folio (No. 150), p. 5, 1907.

Section 13 miles west of Newcastle, Wyo.

	Feet
Lance formation (part):	
Sandstone, yellow-brown, containing flat hard brown sandstone concretions; unit weathers into "toadstools"	30
"Somber shale," composed of dark-gray sandy clay with small botryoidal masses of limonitic sand, brown concretions, and a little hard sandstone in thin layers; turtle and dinosaur bones at top	50
Slope probably underlain by gray sandy clay shale	70
Sandstone, soft, light gray, and shale, dark gray, alternating in thin layers; contains a little hard brown platy sandstone; no fossils noted, assignment arbitrary	30
Fox Hills sandstone:	
Sandstone, hard, platy, dark brown; forms cap rock of many small buttes	3
Sandstone, rather soft, yellow to light gray; weathers to a slope	62
Sandstone, argillaceous, very soft, slope forming; contains a few somewhat harder layers, some thin layers, and a few concretions of siderite. At 44 feet above base (U. S. G. S. locality 12833) contains *Crenella elegantula* Meek and Hayden, *Veniella humilis* Meek and Hayden, *Pyrifusus newberryi* Meek and Hayden, *Haminea* sp.; at 5 feet above base (U. S. G. S. locality 12832), *Serpula?* sp., *Nucula* sp., *Yoldia evansi* Meek and Hayden, *Pecten nebrascensis* Meek and Hayden, *Anomia gryphorhynchus* Meek and Hayden, *Modiola meeki* Meek and Hayden, *Crenella elegantula* Meek and Hayden, *Crassatellites subquadratus* Whitfield?, *Lunatia concinna* Hall and Meek, *Anchura americana* Evans and Shumard. *Pyrifusus newberryi* Meek and Hayden, *Fasciolaria culbertsoni* Hall and Meek, *Cinulia concinna* Hall and Meek, *Baculites ovatus* Say, and undetermined fish scales	60
Total Fox Hills sandstone	125
Pierre shale:	
Limestone concretions, gray, 3 to 4 feet in diameter, most of them septarian and without fossils; *Baculites ovatus* Say, *B. compressus* Say, *Yoldia evansi* Meek and Hayden seen	3
Shale, greenish gray, sandy, with a few small calcareous concretions	20
Concealed under a grassy flat, probably in part large sandy gray shale; extends eastward from Fox Hills escarpment	±150
Concretions brown, calcareous, with many fossils. About a mile east of the Fox Hills escarpment (U. S. G. S. locality 12757) contains *Nucula* sp., *Yoldia evansi* Meek and Hayden, *Cucullaea shumardi* Meek and Hayden, *Gervillia subtortuosa* Meek and Hayden, *Inoceramus fibrosus* (Meek), *Inoceramus sagensis* Owen, *Pteria nebrascana* Evans and Shumard, *Ostrea* sp., *Pecten nebrascensis* Meek and Hayden, *Syncyclonema halli* Gabb, *Anomia gryphorhynchus* Meek and Hayden, *Modiola meeki* Evans and Shumard, *Crenella elegantula* Meek and Hayden, *Cuspidaria* sp., *Lucina* sp., *Protocardia subquadrata* Evans and Shumard, *P. borealis* Whiteaves, *Corbula* cf. *C. crassimarginata* Meek and Hayden, *Dentalium gracile* Meek and Hayden, *Lunatia occidentalis* Meek and Hayden, *Anchura* sp., *Haminea subcylindrica* Meek and Hayden, *Baculites ovatus* Say. *B. grandis* Hall and Meek, *Discoscaphites nicolleti* Morton, *D. abyssinus* Morton, and *D. conradi* Morton var.	

The Fox Hills sandstone has been reported,[40] on the authority of D. E. Winchester and V. H. Barnett, to be 1,040 feet thick on Alkali Creek, on the same line of outcrop as the much thinner sections reported near Newcastle and Moorcroft. This greater thickness has been interpreted as normal for the formation and the smaller thicknesses found elsewhere as representing only a remnant left by pre-Lance erosion.[41] The great thickness reported is, however, incorrect, as shown by the detailed section measured by the writers and given below. The Fox Hills on Alkali Creek is thicker, it is true, than at the more northerly localities, but it reaches only the moderate thickness of 357 feet and is transitional into the overlying Lance, as described by Brown.[42] Stanton's section at the mouth of Lance Creek,[43] some miles south of Alkali Creek, agrees well with that measured by the writers, though it is somewhat thicker—about 400 feet.

Section measured in sec. 18, T. 40 N., R. 61 W., and secs. 11 and 13, T. 40 N., R. 62 W., Wyo., between Alkali Creek and Robbers Roost Creek

	Feet
Lance formation (part):	
Shale, somber colored; contains abundant remains of turtles and dinosaurs	50
Shale, carbonaceous, and coal	3
Fox Hills formation:	
Sandstone, brown, concretionary; contains abundant remains of dinosaurs, crocodiles, and turtles near top. V. H. Barnett and D. E. Winchester [44] reported marine fossils at a horizon 35 feet below top of this unit. The present writers did not succeed in verifying their report. This unit may belong in the Lance	50
Sandstone, gray, filled with shells of *Corbicula* cf. *C. subelliptica* Meek and Hayden; a brackish-water deposit that may belong to the Lance	23
Sandstone, gray, and soft sandy shale, inseparable in grass-covered slopes	74
Sandstone, yellowish, hard, capped by ripple-marked fissile concretions; contains *Yoldia evansi* Meek and Hayden, *Crenella?* sp., and *Haminea minor* Meek in basal part (U. S. G. S. locality 12586)	4
Shale, dark gray, and sandstone, soft, light gray, interlaminated; contains botryoidal masses of limonite and concretions of platy brown sandstone; mass weathers into slopes like the Lance "somber beds"; contains *Nucula* sp. indet., *Modiola attenuata* Meek and Hayden, *Cardium speciosum* Meek and Hayden, *Tellina* sp. indet., 25 feet below top (U. S. G. S. locality 12585)	70
Sandstone, platy, hard, brown	1
Clay, somber-colored	10
Concealed; probably interlaminated gray and brown sandy shale and sandstone much like third unit above	80

[40] Calvert, W. R., and others, Geology of the Standing Rock and Cheyenne River Indian Reservations, North and South Dakota: U. S. Geol. Survey Bull. 575, p. 12, 1914.

[41] Knowlton, F. H., Cretaceous-Tertiary boundary in the Rocky Mountain region: Geol. Soc. America Bull., vol. 25, pp. 326–327, 1914.

[42] Brown, Barnum, Cretaceous-Eocene correlation in New Mexico, Wyoming, Montana, Alberta: Geol. Soc. America Bull., vol. 25, p. 358, 1914.

[43] Stanton, T. W., Fox Hills sandstone and Lance formation ("*Ceratops* beds)" in South Dakota, North Dakota, and eastern Wyoming: Am. Jour. Sci., 4th ser., vol. 30, pp. 184–185, 1910.

[44] Unpublished data.

Fox Hills formation—Continued.	Feet
Sandstone, gray, platy; makes bench	10
Sandstone, brown, limonitic, hard and platy; makes the cap rock on the lowest Fox Hills escarpment	12
Sandstone, yellow, massive, soft; contains large hard brown calcareous concretions, which yielded (U. S. G. S. locality 12584) *Membranipora* n. sp., *Gervillia subtortuosa* Meek and Hayden, *Inoceramus* (*Actinoceramus*) *fibrosus* Meek and Hayden, *Veniella humilis* Meek and Hayden, *Protocardia subquadrata* Evans and Shumard, *Tellina equilateralis* Meek and Hayden, *Dentalium gracile* Meek and Hayden, *Fasciolaria cheyennensis* Meek and Hayden?, *Fasciolaria galpinianum* Meek and Hayden, *Cinulia concinna* Hall and Meek, *Discoscaphites nicolleti* (Morton), *Sphenodiscus lenticularis* (Owen), *Hypsodon* sp., and *Centrarchites?* sp	40
Sandstone, gray, shaly but hard enough to make a prominent bench	23
Total Fox Hills	357
Pierre shale.	

In the upper Lance Creek area the Fox Hills is about 500 feet thick and is transitional into the Lance. Winchester [45] reports a thickness of 748 feet, but examination by others has not substantiated so great a thickness as this. An increase in thickness toward the southwest has been established for all the marine formations of the Montana group and for the overlying Lance and Fort Union formations and is possibly due to the fact that the chief source of the sediments lay toward the southwest. Practically all sections of the Fox Hills in this particular area contain several beds of massive white sandstone, some of which are associated with beds of coal and impure shale. The following section, given by Stanton,[46] is typical of the Fox Hills in this area:

Section on Johnson Bros.' ranch near Buck Creek

	Feet
Lance formation: Sandy shale with thin beds of coal	25
Fox Hills sandstone:	
Massive white sandstone with *Halymenites major*	60
Yellowish massive sandstone with brown concretions	20
More thinly bedded brown sandstone with *Halymenites*	25
Massive white sandstone	75
Soft, somewhat sandy shales with thin sandstone bands containing marine Fox Hills shells	30
Brown shaly sandstone	5
Massive white sandstone	60
Thin-bedded brown and gray sandstone	130
Yellowish massive sandstone with concretions containing Fox Hills fauna	100
Total Fox Hills	505
Pierre shale.	

CENTRAL WYOMING

In the district near the Salt Creek oil field the fresh-water Lance formation is 3,200 feet thick and is underlain conformably by a series of beds designated the Lewis shale. The uppermost part of the Lewis is a pure-white sandstone 100 to 150 feet thick which bears *Halymenites* and is possibly a correlative of the Colgate sandstone of eastern Montana. Shale associated with the white sandstone contains Fox Hills fossils,[47] and the sandstone itself contains one or more coal beds, generally only a few inches thick but in some places as much as 5 feet thick. The beds through an interval of 600 feet below the white sandstone are predominantly sandy, though in at least half of this interval they contain distinctive Pierre fossils and are equivalent to the uppermost part of the Pierre shale of eastern Wyoming. Deposition of sandy beds, therefore, began earlier here than it did farther east. There are only arbitrary boundaries between representatives of the Pierre, Fox Hills, and Lance.

At Glenrock, east of Casper, the uppermost part of the Lewis shale, just under the Glenrock coal bed of the Lance, is a massive pure-white sandstone about 75 feet thick that contains *Halymenites* throughout. Under it lies perhaps 125 feet of yellow-brown massive sandstone containing many concretions of red-brown hard calcareous sandstone. Beneath this yellow-brown sandstone lies about 125 feet of shale, thin-bedded sandstone, and coal, beneath which in turn is a second yellow sandstone 70 feet thick. Barnett [48] reports the beds above the coal-bearing zone as 100 feet thick, but this figure is not quite large enough. Beneath the rocks above described, which are very likely of Fox Hills age, lie some 600 feet of interbedded sandstone and shale and, still lower, less sandy beds. These all contain a Pierre fauna and are therefore older than the Fox Hills.

In the Poison Spider district, 35 miles west of Casper, the Lewis shale is about 800 feet thick and consists of alternating beds of sandstone, shale, carbonaceous shale, and coal; the upper beds contain a distinctive Fox Hills fauna with *Sphenodiscus*, etc., and the lower beds an upper Pierre fauna. There are alternations of marine and nonmarine beds, and the entire section can be interpreted only as showing a gradual change from marine to continental conditions. A section illustrative of this district follows:

Section at the center of T. 33 N., R. 83 W., Wyo., 3 miles southwest of the Poison Spider oil field

	Feet
Lance formation (basal part): Shale, sandy, gray, with carbonaceous shale, thin coal beds, and concretionary sandstone; forms a prominent scarp. This unit is the lower part of the coal-bearing zone of the Lance formation	154

[45] Winchester, D. E., The Lost Spring coal field, Converse County, Wyo.: U. S. Geol. Survey Bull. 471, p. 478, 1912.

[46] Stanton, T. W., Fox Hills sandstone and Lance formation ("*Ceratops* beds") in South Dakota, North Dakota, and eastern Wyoming: Am. Jour. Sci., 4th ser., vol. 30, p. 186, 1910.

[47] Wegemann, C. H., The Sussex coal field, Johnson, Natrona, and Converse Counties, Wyo.: U. S. Geol. Survey Bull. 471, p. 445, 912; The Salt Creek oil field, Wyo.: U. S. Geol. Survey Bull. 670, p. 23, 1918.

[48] Barnett, V. H., Possibilities of oil in the Big Muddy dome, Converse and Natrona Counties, Wyo.: U. S. Geol. Survey Bull. 581, p. 114, 1915.

Lewis shale:	Feet
Sandstone, yellow, calcareous, concretionary; breaks into irregular pieces and contains *Halymenites*	1
Sandstone, gray-white, soft	80
Shale, gray to brown, sandy	65
Sandstone, brown, concretionary, limy; contains fossil plants in upper part (U. S. G. S. locality 7508), including *Taxodium occidentale* Newberry, *Ficus cockerelli* Knowlton?, *Rhamnus salicifolius* Lesquereux?, and *Ficus* sp.; also *Ostrea glabra* Meek and Hayden in lower 6 inches (U. S. G. S. locality 10708)	5
Shale, sandy, gray; contains thin lenses of papery sandstone	75
Shale, brown, sandy, grading upward into a prominent concretionary sandstone that contains a Fox Hills fauna, including *Nucula* sp., *Glycimeris wyomingensis* (Meek), *Pteria linguaeformis* Evans and Shumard, *Pteria nebrascana* Evens and Shumard, *Ostrea* cf. *O. glabra* Meek and Hayden, *Tancredia americana?* Meek and Hayden, *Cardium speciosum* Meek and Hayden, *Callista nebrascensis* Meek and Hayden, *Tellina scitula* Meek and Hayden, *Turritella* sp. undescribed, *Pyrifusus newberryi?* Meek and Hayden, *Sphenodiscus lenticulare* (Owen) (U. S. G. S. locality 10707)	25
Shale, gray, sandy	40
Lignite and brown carbonaceous shale	3
Sandstone, gray, fairly hard; contains much carbonaceous débris	5
Shale, gray, sandy	10
Coal	1
Shale, sandy, grayish white; contains thin lenses of yellow concretionary sandstone	90
Sandstone, gray, platy, concretionary	1
Shale, sandy, gray and brown, soft	80
Shale, gray, and carbonaceous shale, sandy shale, and concretionary sandstone; appears to be nonmarine	70
Sandstone, brown, hard; makes a small ridge	3
Shale, gray, carbonaceous	25
Sandstone, soft, carbonaceous; contains several hard layers	6
Shale, sandy, gray and brown	30
Shale, some sandy and some carbonaceous	1
Sandstone, gray, carbonaceous	8
Sandstone, white; contains many gray cannon-ball concretions as much as 3 feet in diameter	62
Shale, sandy, brown; occupies a low valley	66
Sandstone, gray and brown, with many large dark-brown concretions; contains in upper part a large Pierre fauna including *Baculites compressus* Say, *B. ovatus* var. *baculus* Meek, and *Scaphites nodosus* Owen (U. S. G. S. localities 10705, 10706)	100
Shale, gray, marine, sandy, grading upward into overlying brown sandstone; *Baculites compressus* Say (U. S. G. S. locality 10709)	200
Mesaverde formation: Teapot sandstone hogback.	

SOUTHERN WYOMING

In southern Wyoming the marine Lewis shale is succeeded by beds formerly called the Laramie formation but now designated "Laramie" or Medicine Bow formation. There is complete transition from marine to nonmarine deposits, and the transition beds are generally assigned to the succeeding formation because they are sandy and hence more like it in lithology than the underlying Lewis shale.

About 10 miles west of Rawlins, just north of Solon station on the Union Pacific Railroad, good exposures of Lewis and overlying beds show the transition from marine to nonmarine beds through strata that are undoubtedly of Fox Hills age but here, as in the Casper region, are not separable except arbitrarily from the beds above and below. The following section illustrates the sequence:

Section 10 miles west of Rawlins, Wyo.

"Laramie" formation (part):	Feet
Shale, sandy, brown, with thin carbonaceous shale beds and concretionary sandstone; contains *Unio priscus* Meek and Hayden, *Campeloma multilineata* Meek and Hayden, *Tulotoma thompsoni* White, and *Goniobasis tenuicarinata* Meek and Hayden 140 feet below top (U. S. G. S. locality 12507); *Unio priscus* Meek and Hayden 165 feet below top (U. S. G. S. locality 12507); and *Unio holmesianus* White 40 feet above base (U. S. G. S. locality 12505)	280
Sandstone, brown	6
Shale, sandy, brown, with a few concretionary sandstone lenses. Fragment of undeterminable dinosaur bone 85 feet below top; ceratopsian remains 300 feet below top	385
Sandstone, brown, concretionary; contains many *Halymenites;* highest marine horizon noted	9
Shale, brown, sandy, with thin platy sandstone	21
Coal	3
Sandstone, brown, concretionary; gray sandy shale; and several carbonaceous shale beds 2 to 4 feet thick	40
Sandstone, brown, concretionary; gray sandy shale; and carbonaceous brown shale. At about the middle of this unit and half a mile south of the line of the section (U. S. G. S. locality 10722) were found *Nucula* sp., *Yoldia evansi* Meek and Hayden, *Barbatia?* sp. undescribed, *Ostrea* sp., *Cardium* (*Ethmocardium*) sp. undescribed, *Mactra nitidula* Meek and Hayden, *Cuspidaria* aff. *C. moreauensis* Meek and Hayden, undetermined crab, and *Echidnocephalus?* sp	110
Shale, brown, sandy	38
Shale, carbonaceous	3
Sandstone and gray shale	33
Coal	2
Sandstone and gray shale	15
Shale, carbonaceous; contains at the middle (U. S. G. S. localities 10720, 12503) *Ostrea glabra* Meek	6
Shale, gray, sandy, with brown concretionary sandstone and beds of carbonaceous shale. From a soft sandstone 20 feet below the top *Nucula* sp., *Anomia gryphaeiformis* Stanton, *Unio subspatulatus?* Meek and Hayden, *Unio danae?* Meek and Hayden, *Unio* sp., *Campeloma multilineata?* Meek and Hayden, and *Viviparus* sp. (U. S. G. S. localities 102504, 10721); and *Rhamnus williardi* Knowlton, *Quercus* sp., *Viburnum* sp., and *Oreodaphne?* sp. (U. S. G. S. locality 7510)	140
Sandstone, brown, concretionary, marine; *Halymenites* abundant and fragments of shells	6
Shale, gray, sandy	28
Shale, carbonaceous	1
Shale, gray, and concretionary sandstone	10
Shale, carbonaceous	4

"Laramie" formation (part)—Continued.	Feet
Shale, gray, and concretionary sandstone	96
Coal	2
Shale, gray	40
Coal	3
Shale, gray	8
Coal	2
Shale, sandy, gray, and thin platy sandstone	70
Shale, carbonaceous; contains *Ostrea glabra* Meek and Hayden (U. S. G. S. locality 12502)	6
Shale, gray, sandy, with numerous discontinuous ferruginous concretionary beds	140
Sandstone, dark brown, massive to platy; contains *Ostrea* sp., *Cardium speciosum* Meek and Hayden, *Baculites* sp. (U. S. G. S. locality 12501)	6
Shale, sandy, dark brown	82
Shale, sandy, gray, with a few sandstone lenses	71
Sandstone, brown, concretionary, contains macerated plant remains	10
Shale, dark gray; weathers light colored	28
Sandstone, platy, calcareous	4
Shale, gray, sandy	112
Sandstone, brown, concretionary; contains *Ostrea* sp., *Cardium speciosum* Meek and Hayden, and *Corbula undifera* Meek (U. S. G. S. locality 12500)	4
Sandstone, gray, platy	53
Shale, sandy, dark gray; weathers light colored	55
Sandstone, grayish brown, concretionary	2
Lewis shale: Shale, gray, with minor thin beds of sandstone and calcareous concretions; contains Pierre fossils. Thickness not measured.	

E. E. Smith and M. W. Ball [49] collected at other localities in this area distinctive Fox Hills fossils from strata comparable to the uppermost marine beds of this section, and there is no doubt that the higher marine beds are of Fox Hills age. The lower beds are definitely of upper Pierre age. Plants collected from the nonmarine beds above the highest marine beds have been ascribed to the Laramie flora and given an age assignment older than that of the Lance beds, but an abundant invertebrate fauna is identical with that of the Lance of northeastern and central Wyoming. At no place in the sequence, either in the part described here or in considerable thicknesses above and below it, is there a plane suggesting an important interruption.

In the Hanna Basin [50] the change from marine to nonmarine beds—from Lewis shale to Medicine Bow formation—is exactly parallel to the change near Rawlins from Lewis to "Laramie." A typical example is the section 5 miles north of Walcott, Wyo. Here massive light-gray to brown sandstone 270 feet thick rests on several thousand feet of marine Lewis shale. Upon the sandstone lies about 300 feet more of shale, poorly exposed but probably marine and included in the Lewis by Bowen; then 550 feet of shale, soft gray to brown sandstone, and some hard dark-brown sandstone, which is not very fossiliferous but yields a few oysters and scattered plant remains and forms the basal part of the Medicine Bow formation. Next above is a thin zone of hard brown concretionary sandstone with Fox Hills fossils, including *Yoldia evansi* Meek and Hayden, *Yoldia?* sp., *Ostrea* sp., *Anomia* sp., *Modiola* sp. undescribed, *Anatina* aff. *A. doddsi* Henderson, *Pholadomya subventricosa* Meek and Hayden, *Thracia subgracilis?* Whitfield, *Cardium speciosum* Meek and Hayden, *Tellina scitula* Meek and Hayden, *Legumen planulatum* Conrad, *Mactra* cf. *M. warrenana* Meek and Hayden, *Dentalium* sp., *Lunatia concinna* Hall and Meek, *Haminea* sp., *Anisomyon?* sp., and *Sphenodiscus lenticulare* (Owen) (U. S. G. S. locality 10723). At a horizon 15 feet higher (U. S. G. S. locality 10724) occur *Ostrea glabra* Meek and Hayden, *Anomia gryphorhynchus* Meek, and *Corbicula* sp.—a brackish-water fauna—and at 15 feet higher still (U. S. G. S. locality 10725) *Ostrea glabra* Meek and Hayden and *Corbicula* sp. Above these only plant remains were found in a hasty search, though fresh-water invertebrates are abundant at many places in the formation in adjacent sections. The flora of the Medicine Bow formation has been identified with that of the Laramie of the Denver Basin and has been considered pre-Lance in age. The same zone of marine beds as that noted at Walcott, or at least a very similar one, well above the base of the Medicine Bow formation, was noted by Stanton [51] at several other places in the Hanna Basin, and there seems to be little doubt that a fairly thick transition zone of alternating marine and nonmarine deposits is widespread. A misinterpretation of the stratigraphy of this area [52] by some of the earlier students led to the concept of the presence here of a very great unconformity comparable to the supposed pre-Lance and pre-Arapahoe unconformities, but the careful restudy of the region by Bowen and others has shown that the only great unconformity in the section is at a much higher stratigraphic plane, the base of the Wasatch.

In this same general area the writers, with J. B. Eby, examined and measured the section on Muddy Creek near Dad, Wyo. Here is found essentially the same unbroken sequence from Lewis up into "Laramie."

For southern Wyoming as a whole the beds called "Laramie" or Medicine Bow are identical in lithology with those a short distance to the north, beyond the Sweetwater Mountains, which are accepted as Lance by everyone who has examined them; the same large nonmarine fauna occurs in both sets of beds and also the same *Triceratops* fauna; both succeed conformably

[49] Unpublished data.

[50] For a general description see Bowen, C. F., Stratigraphy of the Hanna Basin, Wyo.: U. S. Geol. Survey Prof. Paper 108, pp. 227-235, 1918.

[51] Stanton, T. W., personal communication.

[52] Veatch, A. C., Coal fields of east-central Carbon County, Wyo.: U. S. Geol. Survey Bull. 316, p. 250, 1907. Knowlton, F. H., Further data on the stratigraphic position of the Lance formation ("*Ceratops* beds"): Jour. Geology, vol. 19, pp. 359–362, 1911; Cretaceous-Tertiary boundary in the Rocky Mountain region: Geol. Soc. America Bull., vol. 25, p. 328, 1914.

a zone containing the *Sphenodiscus* fauna of Fox Hills age. The only discordant note is that the flora of the Lance beds is said to be a Fort Union flora, whereas the flora of the "Laramie" or Medicine Bow formation is said to be a Laramie flora and older than Lance. Hares [53] some years ago expressed the opinion that the Lance of the Casper region and much of the Medicine Bow ("Lower Laramie") of southern Wyoming are equivalent, an opinion which the writers believe to be substantially correct.

YAMPA VALLEY, COLO.

An area of especial interest as being the farthest west in Colorado at which a typical Fox Hills fauna is known is in the Yampa Valley in the vicinity of Craig. The section here includes a shale formation designated the Lewis shale, overlain by a formation still known, as are the similar beds in parts of southern Wyoming, by its old name "Laramie," originally applied because the upper part of it contains Laramie plants.[54] The "Laramie" in Yampa Valley begins with a massive sandstone, above which lie thin beds of shale and sandstone that include a coal bed and zones of marine and brackish-water fossils. These fossils, including *Sphenodiscus*,[55] belong to the Fox Hills fauna and range as high as 350 feet above the base of the "Laramie," as the formation was originally identified and as the name is applied now for convenience in mapping. Above the marine beds are shale, sandstone, and thin beds of coal, an assemblage much like that which in Wyoming is called Medicine Bow and Lance, though it is much thinner. The upper boundary of the "Laramie" has been drawn beneath a conglomerate that marks the beginning of the "post-Laramie" formation—a unit that has a flora of Fort Union aspect and a lithology somewhat different from that of the "Laramie." There is no evidence of unconformity in the section except the presence of the conglomerate—evidence of somewhat doubtful value.

NORTHEASTERN COLORADO AND THE DENVER BASIN

The Fox Hills sandstone has been interpreted in this general region as including a series of sandstones and sandy shales with a conspicuous sandstone at the top, the whole assigned a thickness stated by some geologists to be as much as 2,000 feet, though by others only 800 to 1,000 feet, the difference being due in part, perhaps, to difference in choice of basal boundary, in part to differences in calculation of thickness. The sandstone at the top was designated by Henderson [56] the Milliken sandstone, and the suggestion was made that its base would have been chosen as the logical plane of subdivision of the Montana group had the group been first studied in Colorado. The lower Fox Hills, taken in the usual sense as including beds well below the Milliken sandstone member, contains neither the fossils most characteristic of the Fox Hills nor those most characteristic of the Pierre, but it does contain a fauna of species that occur in both Fox Hills and upper Pierre; the Milliken sandstone itself contains the distinctive ammonite *Sphenodiscus*. It seems to the writers that only the Milliken and possibly any marine beds above it should be considered strictly as of Fox Hills age and comparable to the typical Fox Hills, though the grounds for this opinion are chiefly negative and therefore weak. Even if the Fox Hills is conceded to be very thick, the value of this thickness as proof of erosion in other areas is reduced by the fact that the underlying Pierre shale has in this area an astonishingly great thickness, in comparison with which even the great thickness of the Fox Hills would not be disproportionate.

According to Henderson [57] the Milliken sandstone consists of

> massive, rather soft, usually greenish-yellow sandstone, from 100 to 150 feet in thickness, almost entirely free from shales except a few 1-inch bands in the lower part. The sandstone contains many large brown concretions and bands, more or less ferruginous and calcareous and usually highly fossiliferous. The more gentle slopes above are occupied by alternating shales and soft sandstones, containing marine fossils and not sharply separated from the overlying Laramie shales and sandstones. The Laramie is not well exposed in this region, owing largely to the absence or weakness of the massive thick white sandstone, which is such a conspicuous and persistent feature of the lower Laramie in Boulder County.

In the Denver Basin the Milliken sandstone is overlain by two grayish-white sandstones, each about 60 feet thick, separated by about 4 feet of lignitic shale, which passes into coal in some places. A third and much whiter sandstone, about 10 feet thick, occurs 60 feet higher in the section and is a good marker in coal exploration because all workable beds in the lower part of the Laramie formation of the Denver field occur below it.[58]

The rock sequence as thus described might suggest that the Milliken sandstone is to be correlated with the basal brownish sandstone of the northern Fox Hills; that the coal-bearing white sandstones of the basal Laramie of the Denver Basin and northeastern Colorado are to be correlated with the Colgate sandstone member of the Fox Hills as developed in eastern Wyoming, eastern Montana, and the Dakotas; and that the Laramie clays are equivalent to the Hell Creek member of the Lance. Such long-range correlations on lithology alone are very uncertain, but it

[53] Hares, C. J., Correlation of some of the Cretaceous and Eocene formations of central Wyoming [abstract]: Washington Acad. Sci. Jour., vol. 5, pp. 328-330, 1915.

[54] Gale, H. S., Coal fields of northwestern Colorado and northeastern Utah: U. S. Geol. Survey Bull. 415, pp. 72-74, 1910.

[55] Hancock, E. T., Geology and coal resources of the Axial and Monument Butte quadrangles, Moffat County, Colo.: U. S. Geol. Survey Bull. 757, p. 22, 1925.

[56] Henderson, Junius, The Cretaceous formations of northeastern Colorado: Colorado Geol. Survey Bull. 19, p. 22, 1920.

[57] Henderson, Junius, op. cit., p. 23.

[58] Emmons, S. F., Cross, Whitman, and Eldridge, G. H., Geology of the Denver Basin in Colorado: U. S. Geol. Survey Mon. 27, pp. 73-74, 1896.

is on other grounds as nearly certain as such matters can be that the Milliken is equivalent to some part of the typical Fox Hills. The view that the Laramie clays are of the same age as the Hell Creek member of the Lance is supported, as Stanton long ago pointed out, by the invertebrate fauna and by the recent finding in the Laramie beds near Briggsdale, Colo.,[59] of several horn cores identified as probably *Triceratops*. The differences cited between the fossil floras of the Laramie and the Lance are no more impressive than those cited between the floras of the Denver and the Lance or Fort Union, both assigned to the Eocene by Knowlton.

No one has ever reported any evidence of unconformity between the Fox Hills and Laramie, and this fact, with the assignment of a great thickness to the Fox Hills, has been used as an argument that the Laramie and most of the Fox Hills were removed by pre-Lance erosion at places outside of the Denver Basin where no Laramie, as such, is recognized and the accepted Fox Hills is thin. If, however, the Laramie-Arapahoe unconformity represents a short interval, comparable to some of those represented by the admittedly minor unconformities within both Fox Hills and Lance formations, the Laramie, Arapahoe, and Denver with their *Triceratops* fauna become a logical equivalent for some part of the Lance, and the sections in the Denver Basin and in the northern plains match. Hay [60] and Stanton [61] long ago expressed the opinion that the unconformity noted between the Laramie and Arapahoe formations is not of stratigraphic importance—an opinion with which the writers fully agree.

The emphasis that has been laid by some writers on the importance of the unconformity between the Laramie and the Arapahoe rests, as stated on page 10, upon the propositions (1) that in the Rocky Mountain region the Cretaceous sea was unbroken by any islands and undisturbed by any regressions from its entrance into the region in early Upper Cretaceous time until its withdrawal in late Upper Cretaceous time and (2) that the conglomerate of the Arapahoe contains pebbles derived from near-by exposures of practically all the pre-Arapahoe rocks—a necessary corollary being that between the deposition of the last Cretaceous beds and the beginning of the Arapahoe there must have occurred the elevation and erosion of the whole preceding section of some 12,000 feet. The first proposition would certainly not be accepted by students of the Interior Cretaceous to-day. The second appears to the writers highly debatable. Even if the identification of the pebbles as pre-Cambrian, "Red Beds," Dakota, Niobrara, Laramie, etc., is granted—which appears not beyond some doubt, considering the types of rocks concerned—it is still a gratuitous assumption that they had to come from sources near enough to exclude any possible areas of Cretaceous erosion. The demand for a post-Laramie time interval long enough to permit elevation and erosion of the whole thickness of pre-Arapahoe sediments is too great a strain on the known facts, such as the occurrence of very similar if not identical ceratopsian dinosaurs in the Laramie and in the Arapahoe and Denver beds. The cited differences of the Laramie and post-Laramie in the floras are not impressive, in view of the differences in time and in physical conditions supposed to exist—in fact, they are no greater than those existing between the supposed Tertiary Arapahoe-Denver flora on the one hand and the supposed Tertiary Lance-Fort Union flora on the other, the latter said by the paleobotanists to be essentially one flora, though occurring through a great thickness of strata in which marked changes take place in the contained invertebrate and vertebrate faunas.

WESTERN AND SOUTHERN MARGINS OF THE INTERIOR REGION

In New Mexico, Arizona, Utah, and western Wyoming any beds of Fox Hills age that may be present are nonmarine and do not differ sufficiently in either lithologic constitution or faunal content from associated beds older or younger than the Fox Hills to permit any useful comparison to be made. In some regions, such as the San Juan Basin and the Raton coal field, New Mexico, unconformities are recognized within the series of nonmarine beds, but their relative importance can not now be determined. In the San Juan Basin dinosaur remains of very closely allied types occur both below and above the most conspicuous unconformity, suggesting that it is not very significant.

CONSTITUTION OF THE COLGATE SANDSTONE MEMBER OF THE FOX HILLS SANDSTONE

The record of volcanic ash in the Fox Hills sandstone east of the Missouri River near Linton, N. Dak., by Stanton [62] led to the suspicion that the white color of the Colgate sandstone might be due to a content of volcanic ash, and to the hope that the ash might afford a basis for definite correlation.

The deposit near Linton is described as consisting of chalk-white strata some 26 feet thick and 35 feet above the base of the Fox Hills sandstone. Specimens from this deposit were examined by G. F. Loughlin, who reported them to contain 80 per cent of volcanic glass, 15 per cent of quartz and feldspar, 2 or 3 per cent of biotite, and scattered grains of other minerals. About 16 feet higher in the section is a greenalite bed that also contains volcanic glass. The nearest possible source

[59] Toepelman, W. C., Notes on the Laramie formation in central Weld County, Colo.: Jour. Geology, vol. 34, p. 834, 1926.

[60] Hay, O. P., Where do the Lance Creek ("*Ceratops*") beds belong, in the Cretaceous or in the Tertiary?: Indiana Acad. Sci. Proc. for 1909, p. 284, 1910.

[61] Stanton, T. W., Boundary between Cretaceous and Tertiary in North America as indicated by stratigraphy and invertebrate faunas: Geol. Soc. America Bull., vol. 25, pp. 347-348, 1914.

[62] Stanton, T. W., A Cretaceous volcanic ash bed on the Great Plains in North Dakota: Washington Acad. Sci. Jour., vol. 7, pp. 80-81, 1917.

of the glass now known is far to the west, in the Livingston region of Montana.

Clarence S. Ross examined specimens of white Colgate sandstone obtained in T. 21 N., R. 32 E., Mont., but found no evidence of volcanic material, as the following memorandum submitted by him indicates:

The specimens of Colgate sandstone examined are composed of a great variety of detrital materials. Sand grains form about one-third, and claylike material the remainder of a typical specimen. Quartz is the most abundant single mineral element in the sand grains, but minor amounts of microcline, plagioclase, muscovite, biotite, chlorite, tourmaline, garnet, zircon, and apatite are present. The materials are not well sorted and range from 0.3 to 0.01 millimeter or less in diameter. Large grains are in direct contact with small ones, and there is no evidence of concentration of large or small grains along bedding planes. The sand grains are nearly all sharply angular.

Some specimens contain several per cent of calcite, and others contain almost none. Some of this is probably of detrital origin, but euhedral rhombs of calcite and sparse grains of glauconite were probably formed during the deposition of the sediments.

The material between the quartz grains is of the most heterogeneous nature. Most of it was not deposited as fine sediment but forms individual grains that have about the same diameter as the quartz. Much of it resembles the detrital material derived from more or less metamorphosed sediments, and the ferromagnesian minerals are those characteristic of metamorphic rocks. Some of the grains appear to be fine-grained quartzite, and others are chert or jasperlike quartz. Many grains are probably highly indurated shale; others resemble a sericitized rock, and some of the larger quartz grains appear to be partly sericitized. Qualitative chemical tests of the interstitial material show the presence of essential amounts of potash, and this indicates that this material is partly mica.

Clay minerals make up a considerable part of the interstitial material, and most of these appear to have formed in place by the alteration of detrital rock grains. At least two clay minerals are recognizable. The more abundant is a mineral of the kaolin group that is probably leverrierite or anauxite,[63] and the other is beidellite.[64]

The sandstone is nearly white, notwithstanding the fact that it contains grains of brown and other dark colors. Its light color is due to the white, chalklike appearance of the sericite and clay present as interstitial material between the grains of quartz—that is, the color of the sandstone is due to the tinting power of the large proportion of white interstitial material and not to freedom from impurities.

[63] Dittler, E., and Hibsch, J. E., Min. pet. Mitt., vol. 36, p. 85, 1923.

[64] Ross, C. S., and Shannon, E. V., Washington Acad. Sci. Jour., vol. 15, p. 467, 1925.

SUMMARY OF CONCLUSIONS

The writers believe that the foregoing description of the contact of the Fox Hills and Lance formations and of units equivalent to these formations in areas where other names are used, together covering a large part of the Western Interior of the United States, supports the following statements:

The contact of the Lance and Fox Hills is everywhere essentially transitional.

All angular unconformities reported to exist between these formations are misinterpretations of faulting, cross-bedding, or slumping.

Erosion planes in the Fox Hills and Lance formations are unimportant chronologically, being merely evidence of local channeling of near-shore deposits by tidal currents and wave scour in the Fox Hills or of flood-plain deposits by streams in the Lance—that is, the result of contemporaneous erosion and redeposition.

The localities of certain collections of plants identified as belonging to the Fort Union flora and hitherto considered to occur higher than the "unconformity" above the Fox Hills are really in the upper part of the Fox Hills sandstone.

The Fox Hills sandstone, from its type region at Fox Ridge, between the Cheyenne and Moreau Rivers, S. Dak., westward at least as far as the Musselshell River, Mont., and southward as far as Glenrock, Wyo., consists usually of a variable lower member of yellowish-brown marine sandstone and sandy shale and a variable upper member of white sandstone, which is locally of marine origin, though at some places it contains coal and plant remains; above the upper member at most localities occur beds of coal of variable purity, which belong to the Lance formation.

The Laramie and Arapahoe formations and at least part of the Denver formation of the Denver Basin, most of the Medicine Bow formation of southern Wyoming, and the "Laramie" formation of northwestern Colorado, conformable on beds of Fox Hills age, are equivalent to the Lance formation.

The white color of the Colgate member of the Fox Hills sandstone does not appear to be due to a content of volcanic ash, and the formations therefore can not be correlated on this basis.

THE HELDERBERG GROUP OF PARTS OF WEST VIRGINIA AND VIRGINIA

By Frank McKim Swartz[1]

INTRODUCTION

Although 90 years has elapsed since W. B. Rogers[2] first described the Paleozoic rocks of the Virginias, our knowledge of the more exact stratigraphic and paleontologic features of these beds is still relatively meager, at least when compared with the more extensive data gathered concerning the equivalent strata in the States to the north. The geologists concerned in the preparation of the folios of the Geologic Atlas covering this region[3] and of the State publications[4] have not, on the whole, been permitted time to make very detailed studies of these phases of their problems.

The present report is based on work done in the field during the summers of 1924 and 1925. Much of the laboratory work was completed and the report was originally written by the end of 1925. Several subsequent visits to Virginia have made possible some additions to the original manuscript; particularly with respect to the area south of the New River, which was visited with Charles K. Swartz during the summers of 1926, 1927, and 1928, in connection with a study of the Silurian deposits of that area.

The investigation was limited to a study of the development, in the Virginias, of the sediments that form the basal portion of the Devonian in this region and that constitute the Helderberg group, in the sense in which that term was used in the Maryland Geological Survey's report on the Lower Devonian. The strata concerned are here recognized as divisible into the Keyser, Coeymans, New Scotland, and Becraft limestones, named in ascending order. The relations of the Shriver chert have also been considered, as it seems to me probable that that formation is essentially equivalent to the Becraft limestone, at least as the Becraft is developed in Maryland and Virginia. The underlying Tonoloway limestone and the overlying Oriskany deposits were examined and roughly meas-

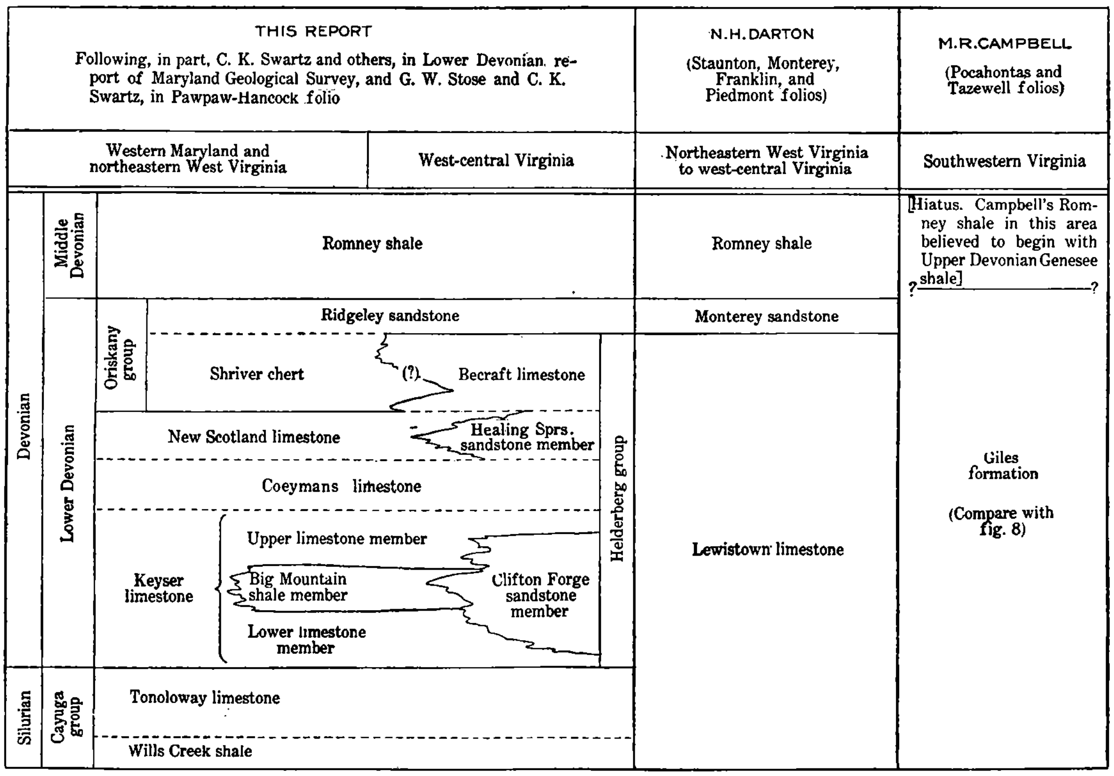


Figure 5.—Relations of Silurian and Devonian formations in western Maryland, northeastern West Virginia, and west-central and southwestern Virginia, as interpreted by F. M. Swartz

[1] Pennsylvania State College.

[2] Rogers, W. B., Virginia State Geologist Rept. for 1837, reprint, pp. 195–203, New York, D. Appleton & Co., 1884.

[3] Darton, N. H., U. S. Geol. Survey Geol. Atlas, Folios 14, 28, 32, 61, 1895–1899. Campbell, M. R., idem, Folios 12, 26, 44, 59, 1894–1899.

[4] Virginia Geol. Survey Bull. 11A, 1909; Bull. 23, 1922. West Virginia Geol. Survey, Report on Pendleton County.

ured, as they are involved in delimiting the Helderberg in the several sections.

Figure 5 indicates the relations of the formations as I would interpret them and compares the grouping used here with that of Darton and Campbell in the folios of the Geologic Atlas.

The area covered by the investigation was of necessity limited to that in which the Helderberg group crops out. This group is brought to the surface, in the Virginias, at many points along the belt of mountain ranges lying between the Shenandoah Valley and its continuations, on the southeast, and the Allegheny Front, on the northwest. Southeast of this belt the Helderberg and even the underlying Silurian have been removed by erosion, except where they are preserved in the syncline of Massanutten and the associated mountains east of Woodstock. To the northwest the Helderberg beds pass beneath the nearly horizontal higher Devonian and Carboniferous beds, except where brought up locally, as by the Browns Mountain anticline, near Frost, W. Va. The general extent of the belt is indicated by the distribution of the sections visited, as shown in Figure 6.

After an examination of the section at Keyser, W. Va., described in the Maryland report,[5] the Helderberg strata were followed, with frequent sectioning, southward to Clifton Forge, Va., a distance of about 130 miles. Beyond this point it was difficult to obtain exposures, and no satisfactory sections were seen between the Clifton Forge area and Hollybrook, Va. Farther south sections near Wytheville, Saltville, and Big Stone Gap, Va., and Sneedsville, Tenn., were visited. At these localities the Keyser is definitely

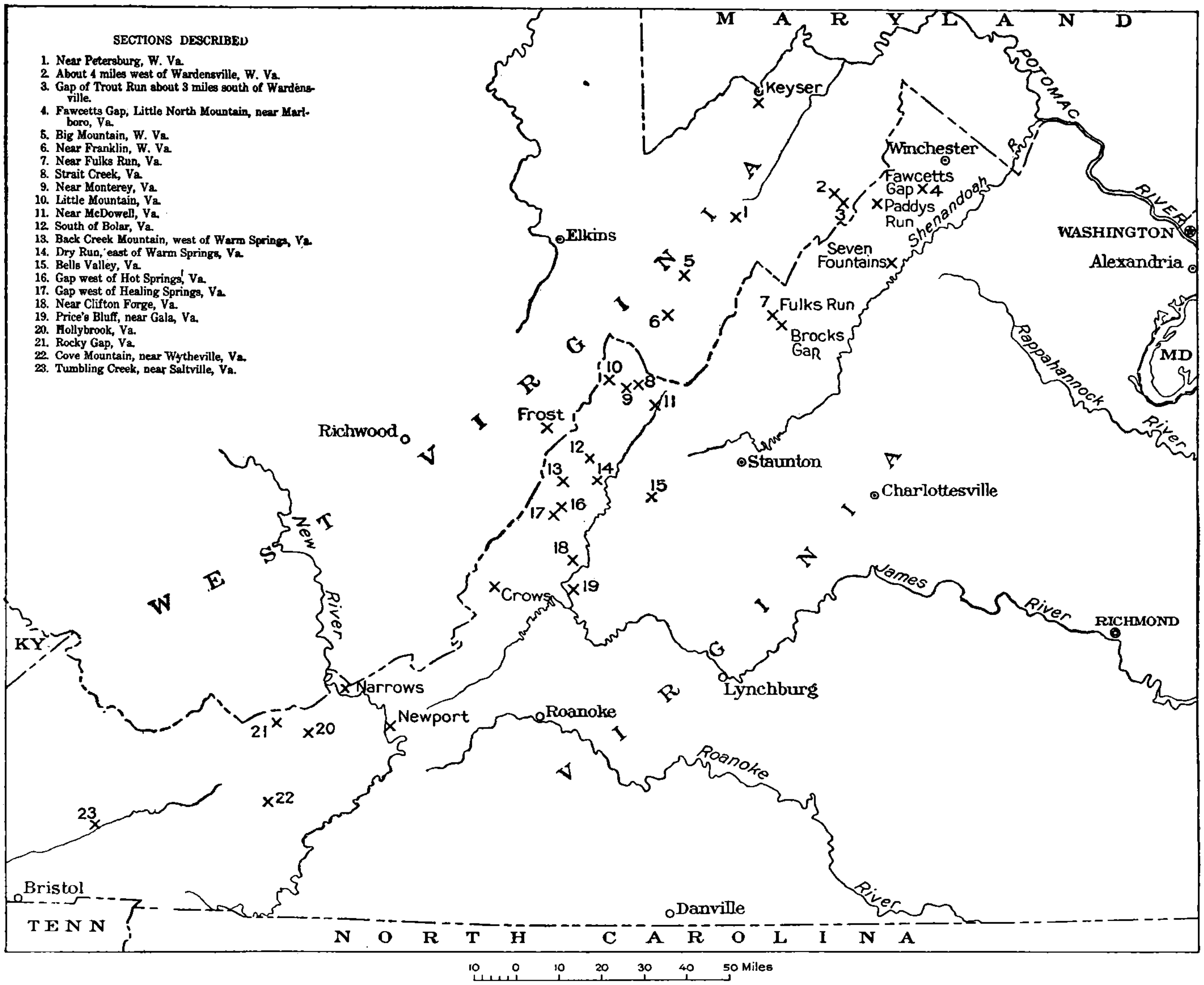


FIGURE 6.—Index map showing area in West Virginia and Virginia covered by investigation of Helderberg group

[5] Maryland Geol. Survey, Lower Devonian, p. 133, 1913.

absent, and there is some uncertainty as to the exact equivalency of the sandstones that form the basal portion of the Devonian.

KEYSER LIMESTONE

CHARACTER AND THICKNESS

At Keyser, W. Va., the type locality, the Keyser formation consists of a series of limestone beds 281 feet thick (see fig. 7), which are blue and very nodular below, more massive toward the middle, and rather shaly above. The upper portion resembles the Tonoloway somewhat in lithology and in the presence of *Tentaculites gyracanthus* and several ostracodes. The formation is underlain by the Tonoloway limestone, which is fine grained, thinly laminated, and sparingly fossiliferous—features which it was found to maintain, on the whole, at least as far south as Sneedsville, Tenn., where its lithology and fauna were identified. The Keyser is overlain by the Coeymans limestone, which is massive, crystalline, and highly crinoidal and carries characteristic fossils. This formation undergoes no essential lithologic changes as far south as Clifton Forge, Va., but its development farther south is questionable. The Keyser is thus limited, in much of the area investigated, by two well-defined formations.

At Petersburg, W. Va., the Keyser is 271 feet thick and consists of three well-defined members—an upper and a lower limestone and an intervening shale. The lower limestone is mostly heavy bedded and very nodular and aggregates 93 feet in thickness. The lowest beds, however, are somewhat shaly, and at 55 to 70 feet above the base there is a massive crinoidal and crystalline unit, overlain by a 13-foot concealed interval. The middle member is a calcareous shale, about 34 feet thick at Petersburg but thickening somewhat southward, replacing the upper beds of the underlying limestone, so that it finally rests upon the crystalline limestone noted above. This shale member is here named Big Mountain shale member, from exposures on Big Mountain, Pendleton County, W. Va. The upper limestone member carries massive and purer beds below, is more impure above, and is 144 feet thick. The three members described above are maintained southward to Warm Springs, Va.

As will be seen from Figure 7, there is little change in the thicknesses of these members, except as already noted for the shale, as far south as Bolar, Va. In the upper half of the upper limestone member there is, however, a well-marked variation from the impure phase seen at Keyser and Petersburg to a massive, purer, generally crystalline phase, well exhibited in the section west of Franklin, W. Va. The change in the character of the sediments is also reflected by changes in the fauna, including the disappearance of *Tentaculites gyracanthus* and of the ostracodes, seen at this horizon not only at Petersburg and Keyser, W. Va., but also northward through Maryland and Pennsylvania into the equivalent so-called Manlius limestone of New Jersey, which is considered by E. O. Ulrich and other geologists to be of Lower Devonian age and younger than the typical Manlius of New York; and the appearance of small varieties of *Camarotoechia altiplicata* and *Spirifer perlamellosus*, neither of which occurs below the Coeymans farther north.

Greater changes occur south of Bolar, Va. These include a marked thinning of the formation as a whole, the tonguing out of the Big Mountain shale member, the development of a heavy calcareous sandstone, seemingly equivalent to the lower half of the upper limestone member, the entire Big Mountain shale member, and the upper part of the lower limestone member. Thus in the section at Clifton Forge, Va., the lower limestone member is represented lithologically by only about 15 feet of somewhat sandy limestone, which retains the nodular character and the fauna of the lower Keyser as seen farther north; the Big Mountain shale member has disappeared as such; the bulk of the Keyser is formed by 66 feet of largely massive, unfossiliferous sandstone, here named the Clifton Forge sandstone member; and the overlying 34 feet of massive limestone seems to represent only the upper part of the upper limestone member as seen farther north.

The Keyser was not seen south of Gala, near Clifton Forge, Va. It is definitely absent in the sections near Saltville and Big Stone Gap, Va., where limestone of Tonoloway lithology and fauna was seen in contact with calcareous sandstones that are not older than the Coeymans. The Keyser is probably absent at Hollybrook and Rocky Gap also, although there are at those places concealed intervals of about 90 and 40 feet, respectively, between the exposures of the limestone of Tonoloway age and the Coeymans or younger beds. (See fig. 8.)

The Keyser also thins decidedly toward the eastern border of the belt of outcrop. Thus at Wardensville, W. Va., about 22 miles east of the Petersburg section, the entire interval between the Tonoloway and the Ridgeley sandstone is only 200 to 215 feet, as compared with 271 feet for the Keyser alone at Petersburg. The beds of the lower half of this interval are well exposed west of Wardensville, where the *Merista typa* subzone of the middle Keyser (No. 6, fig. 7) is 80 to 95 feet above the base. The higher beds are largely concealed in both of these sections, but the presence of considerable cherty material carrying some questionable New Scotland fossils, found loose about 160 feet above the base of the western section, indicates that the Keyser is possibly not over 150 feet thick here.

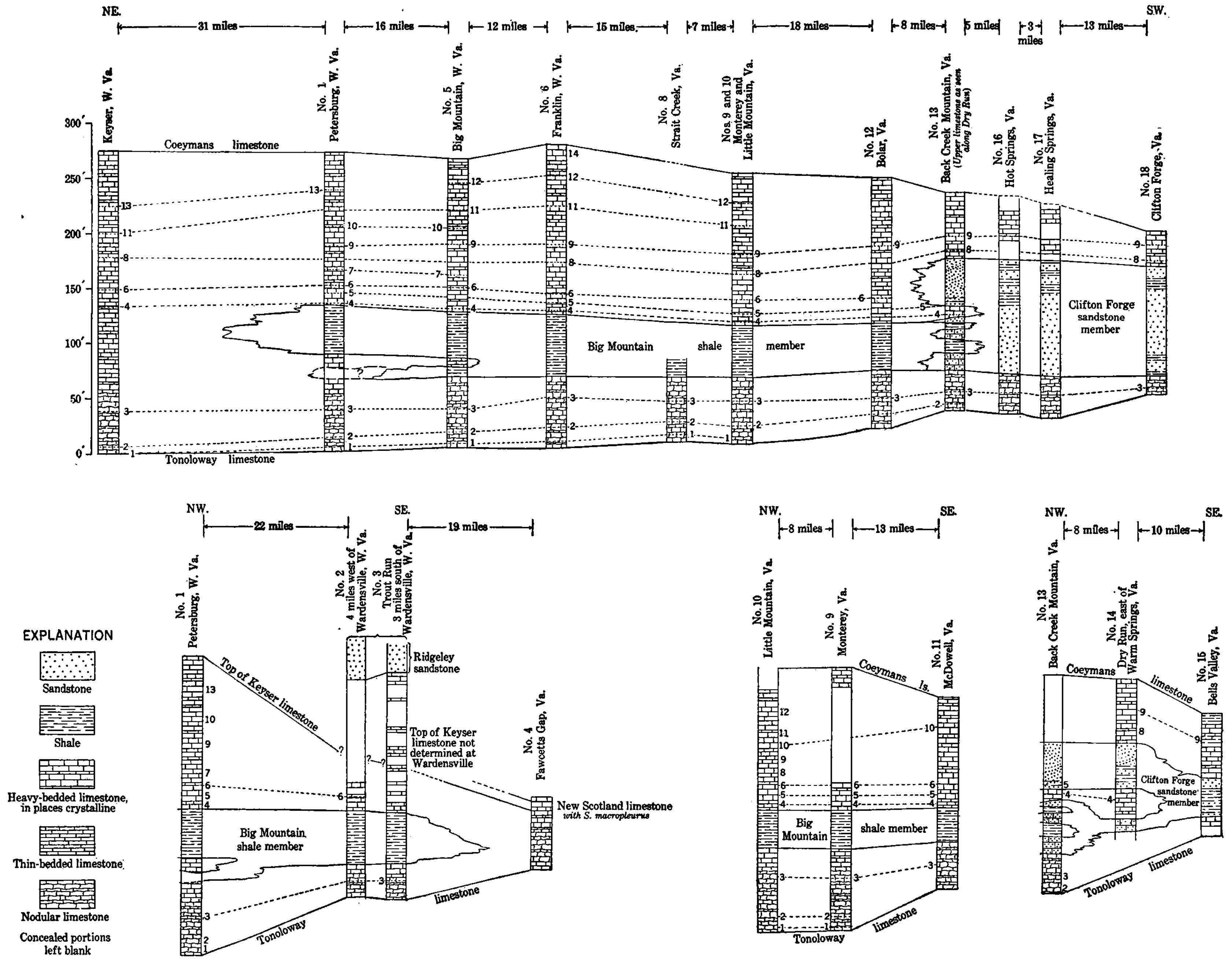


FIGURE 7.—Sections showing stratigraphy and faunal zones of the Keyser limestone from Keyser, W. Va., to Clifton Forge, Va., as interpreted by F. M. Swartz. Numbers at right of sections indicate the faunal zones and subzones. For forms characteristic of the faunal zones and subzones see table on page 32. (Section at Keyser, W. Va., after description in Maryland Geol. Survey, Lower Devonian, pp. 133–136, 1913)

The Keyser is comparatively thin throughout the mountains bordering the west side of the Shenandoah Valley, as far south as Bells Valley, Va. Thus at Fawcetts Gap, in Little North Mountain, about 8 miles southwest of Winchester, Va., the Keyser is only about 50 feet thick and is overlain by a massive limestone carrying *Spirifer macropleurus*, *Dalmanella perelegans*, and other New Scotland fossils. In this section the Keyser is composed entirely of limestones, which are chiefly nodular in the middle and lower portions. Fossils are rare. The Keyser can also be seen at Paddys Run, Va., where the lower beds carry *Chonetes jerseyensis* and *Uncinulus convexorus*. As the upper limits of the Keyser were not determined here, its thickness is uncertain. At Fulks Run, Va., the here also, and those seen are of little aid in correlation. The sandstone and shale of the lower half of the formation are roughly equivalent to the Big Mountain shale and Clifton Forge sandstone members and, like the ripple-marked shale developed at the top of the Keyser, are indicative of shore conditions.

The most easterly occurrence of the Lower Devonian in the Virginias is found in the Massanutten Mountain syncline, where the upper Silurian and Lower Devonian limestones are well exposed in the vicinity of Seven Fountains, Va. The lower portions of these limestones are definitely of Wills Creek and Tonoloway age, but the relations of the upper portion are uncertain. Fossils are scarce, and the distinctive forms of the Helderberg of the more westerly sections are absent,

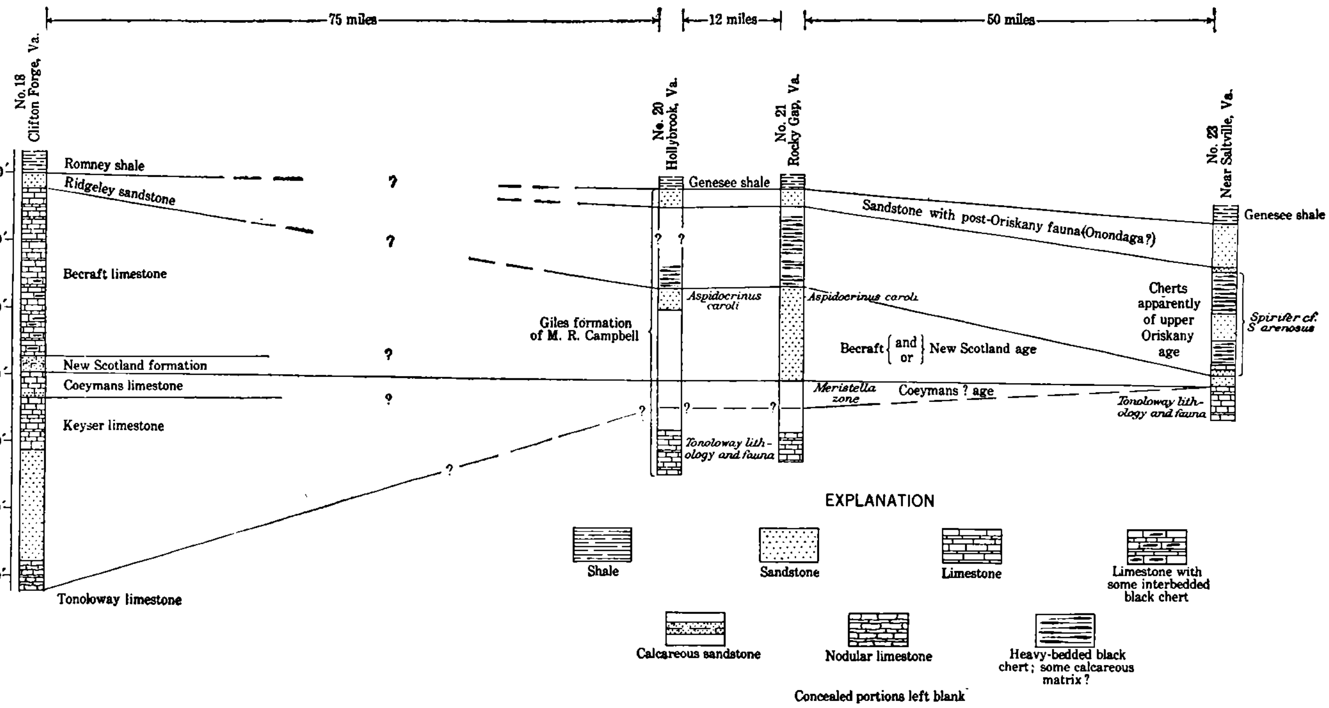


FIGURE 8.—Sections showing stratigraphy of the Helderberg group from Clifton Forge to Saltville, Va., as interpreted by F. M. Swartz. The limestone of Tonoloway age, the conglomerate of Helderberg age, the cherts of Oriskany (?) age, and the greenish post-Oriskany sandstone as seen at Hollybrook, Va., constitute what M. R. Campbell termed the Giles formation in folios of the Geologic Atlas

Keyser seems to be about 160 feet thick and is overlain by the New Scotland containing *Spirifer macropleurus*. *Merista typa* occurs in the middle of the Keyser at this locality, and *Chonetes jerseyensis* was observed near the base. The Big Mountain shale member is present at this locality, below the *Merista typa* zone. There is a concealed interval below the New Scotland, which may be occupied in part by the Coeymans, so that the exact thickness of the Keyser is uncertain, although it is in the neighborhood of 160 feet. The lower part of the Keyser was also seen at Brocks Gap, near by.

At Bells Valley, Va., east of the Warm Springs sections, the Keyser has a possible maximum thickness of 141 feet, of which the lower 34 feet is concealed and may belong with the Tonoloway. Fossils are rare although lithologically these upper beds suggest the Helderberg rather than the Tonoloway. The Oriskany is apparently absent in this area. Further work will be required in this district before exact correlation can be attempted.

FAUNA

On the whole, the Keyser is abundantly fossiliferous, in the Virginias as in Maryland. A list of the fauna occurring in the Virginias will be found in the table on pages 35–37.

One of the most interesting features of the Keyser is the development of a considerable number of marked faunal zones, which can be traced for many miles. In this region, as in Maryland and Pennsylvania, two main zones can be recognized. The lower one is characterized by an abundance of *Chonetes jerseyensis*; the upper one lacks that form and in the Maryland report was termed the *Favosites helderbergiae* zone, because of the abundance of the variety *praecedens* which it contains. Fourteen subzones, based on species of more restricted vertical range, can be recognized in the Virginias. Figure 7 and Table 1 show the occurrence of the subzones in the sections studied and their relations to the subzones established for Maryland and Pennsylvania.

TABLE 1.—*Faunal zones and subzones of the Keyser limestone*

Pennsylvania	Maryland	West Virginia-Virginia
Favosites helderbergiae var. praecedens zone		
17. *Leperditia* subzone.	14. *Leperditia* subzone. 13. *Stromatopora* reef.	14. *Spirifer vanuxemi* var. *prognosticus* and *Whitfieldella prosseri* subzone.
16. *Tentaculites* subzone. 15. *Stromatopora* reef. 14. *Rensselaeria mutabilis* subzone.	12. *Tentaculites* subzone. 11. Lower *Stromatopora* reef. 10. *Rensselaeria mutabilis* subzone.	13. *Tentaculites* subzone. 11. *Rensselaeria mutabilis* subzone. 12. *Camarotoechia* cf. *C. altiplicata* and *Nucleospira ventricosa* subzone.
13. *Pholidops ovata* subzone. 12. Lower *Leperditia* subzone.		10. *Meristella praenuntia* subzone. 9. Petersburg *Stromatopora* reef.
11. Coral subzone B. 10. *Spirifer vanuxemi* var. subzone.	9. Keyser coral reef.	8. Coral subzone, with *Cladopora rectilineata*.
Chonetes jerseyensis zone		
9. *Calymene camerata* subzone. 8. Bryozoan subzone.	8. Bryozoan subzone.	7. Bryozoan subzone. 6. *Merista typa* subzone. 5. *Camarotoechia gigantea* subzone.
7. *Gypidula coeymanensis* var. subzone.	7. *Gypidula coeymanensis* var. subzone.	4. *Gypidula coeymanensis* var. subzone.
6. *Dalmanella clarki* subzone. 5. *Spirifer modestus* subzone. 4. *Stromatopora* subzone. 3. Coral subzone A. 2. *Rhynchospira* subzone.	6. Cystid-*Spirifer* subzone. 5. Rawlings *Stromatopora* reef. 4. *Cladopora rectilineata* subzone. 2, 3. *Rhynchospira* subzones.	3. *Stenochisma deckerensis*, *Spirifer modestus*, *Uncinulus convexorus*, and *Nucleospira swartzi* subzone. 2. *Cyphotrypa corrugata* and *Stropheodonta bipartita* subzones (with *Rhynchospira globosa?*).
1. *Camarotoechia lamellata* subzone.	1. Warrior Mountain coral reef.	1. *Whitfieldella minuta* subzone.

1. *Whitfieldella minuta* subzone: In the somewhat shaly beds that occur at the base of the lower limestone member in a number of the sections, *Whitfieldella minuta* and *Camarotoechia litchfieldensis* occur in considerable abundance. *Whitfieldella minuta*, which occurs at the base of the Keyser at Keyser, W. Va., the type locality of the formation, seems to be restricted to this horizon. Chaetetoid Bryozoa are also prominent; and at this horizon at Petersburg were found some favositid corals, which suggest equivalency with the coral subzone at the base of the Keyser at Warrior Mountain, Md.

2. *Cyphotrypa corrugata*, etc., subzone: In the lower part of the Keyser *Cyphotrypa corrugata* and *Stropheodonta bipartita* have a nearly equivalent range, occurring not only in the *Whitfieldella minuta* subzone but continuing well up to the middle of the lower limestone member and overlapping the range of the species noted in the succeeding subzone. Both of these fossils are common but are rarely profuse. Some poorly preserved material found just above the *Whitfieldella minuta* subzone has been referred with some doubt to *Rhynchospira globosa* and suggests that this is essentially the horizon of the *Rhynchospira* zones noted farther north.

3. *Stenochisma deckerensis*, *Uncinulus convexorus*, etc., subzone: The nodular limestones of the lower limestone member of the Keyser contain a fauna which seems to be essentially a unit, although it is subject to facies development. Thus in the more shaly portions of the nodular limestone in the Petersburg, Big Mountain, Monterey, and Little Mountain sections *Stenochisma deckerensis* and *Uncinulus convexorus* occur in profusion, with an abundance of *Atrypa reticularis* and *Chonetes jerseyensis; Spirifer modestus*, *S. modestus* var. *plicatus*, and *Nucleospira swartzi* are rare in these beds at Petersburg and Monterey, although they are common to abundant at Big Mountain and Little Mountain. In the Strait Creek section this nodular limestone is highly argillaceous and contains an abundance of *Spirifer modestus*, *S. modestus* var. *plicatus*, *Atrypa reticularis*, *Chonetes jerseyensis*, *Nucleospira swartzi*, and *Dalmanella concinna*, but *Stenochisma deckerensis* and *Uncinulus convexorus* are comparatively rare, although the horizon seems to be exactly that at which these two species are abundant in the sections first mentioned. At Franklin, W. Va., and McDowell, Bolar, and Warm Springs, Va., the equivalent beds are more massive, the nodular character being less conspicuous

or even lacking, and although most of the above-named fossils were found, they occur much more sparingly. *Stenochisma deckerensis*, *Uncinulus convexorus*, *Chonetes jerseyensis*, and *Nucleospira swartzi* were also found in the thinning, still nodular remnant of the lower limestone member of the Keyser at Clifton Forge. *Uncinulus convexorus* is extraordinarily profuse in this limestone in the section south of Wardensville, where the other forms are rare or lacking. It should be noted that this is one of the two horizons in the Helderberg group at which *Atrypa reticularis* is abundant; the other one occurs at and above the *Gypidula* zone of the middle Keyser. Similar zones of abundance of *Atrypa reticularis* seem to be present in the Keyser of Maryland. It would be interesting to know what conditions permitted this species to flourish during the times represented by these two horizons, and what changes caused it to be practically excluded from the Maryland-Virginia area during the remainder of Keyser time. The *Stenochisma deckerensis*, etc., subzone is of considerable importance as a guide zone, because it is usually well developed in the sections from Maryland to Clifton Forge, Va., and can, indeed, be traced through Maryland into central Pennsylvania. *Stenochisma deckerensis*, *Uncinulus convexorus*, *Spirifer modestus*, and *Nucleospira swartzi* all seem to be restricted to these nodular limestones of the lower Keyser. There is, however, a very similar nodular limestone at about the middle of the underlying Tonoloway in the vicinity of Monterey, McDowell, and Bells Valley, Va., and this bed also contains an abundance of *Atrypa reticularis* and an *Uncinulus* that differs from *U. convexorus* only in its slightly finer ribbing. Owing to this lithologic and faunal resemblance these nodular beds might be confused, but the Tonoloway zone can readily be distinguished by an investigation of the stratigraphic succession, as well as by the absence of other fossils characteristic of the lower Keyser.

4. *Gypidula coeymanensis* var. subzone: One of the most striking faunal zones of the Keyser is that at about the middle of the formation, characterized by a *Gypidula* forerunning the *G. coeymanensis* of the over-lying Coeymans limestone. The zone can be traced from central Pennsylvania to Warm Springs, Va., and seems to occupy everywhere the same stratigraphic position; moreover, the fossil is both very abundant and greatly restricted in vertical range. The ventral valve tends to stand out on weathered surfaces and is readily recognized. The zone disappears south of Warm Springs, owing to the introduction of the shore conditions represented by the Clifton Forge sandstone member.

In Maryland and Pennsylvania the Gypidulas of this zone have been referred to the variety *prognosticus*, the separation resting largely upon the smaller size. In the Virginias, particularly in the vicinity of Monterey, the individuals of this horizon become larger, some reaching lengths of 35 millimeters, and can hardly be differentiated from the Coeymans material. However, in view of some differences in the plications, I have referred these large specimens to the new variety *similis*.

5. *Camarotoechia gigantea* subzone: The species *Camarotoechia gigantea*, which fails to the north, being relatively rare even in Maryland, has a profuse but stratigraphically restricted development in the Virginias, occurring only in and a few feet above the *Gypidula* zone, from Petersburg, W. Va., to Warm Springs, Va. This horizon is also that of the higher of the two zones in which *Atrypa reticularis* is abundant, as noted in the description of the *Stenochisma deckerensis* zone. *Atrypa reticularis*, however, ranges somewhat higher than *Camarotoechia gigantea*.

6. *Merista typa* subzone: *Merista typa* occurs 5 to 10 feet above the *Gypidula* zone in the Petersburg section, but ranges downward and mingles with both *C. gigantea* and *G. coeymanensis* farther south. It also occurs at this horizon in several of the sections of Maryland and Pennsylvania. The zone was not found south of Bolar, Va. *Chonetes jerseyensis* is of common occurrence at this horizon.

7. Bryozoan subzone: Above the *Merista typa* subzone in the Petersburg and Big Mountain sections was noted a bryozoan subzone, which would seem to correspond with that occurring near the top of the *Chonetes jerseyensis* zone in Maryland and Pennsylvania. No specific determinations were made, however, and the correlation is open to some question.

8. Coral subzone, with *Cladopora rectilineata:* In the section at Franklin, W. Va., *Cladopora rectilineata* occurs in profusion some distance above the *Merista typa* subzone, thus paralleling the occurrence noted by Reeside in Pennsylvania, at about the same horizon. This zone was also seen at Monterey and Little Mountain, Va. At Little Mountain *C. rectilineata* is accompanied by an abundance of *Aulopora schohariae*, *Striatopora bella*, and a number of other corals. *Cladopora rectilineata* is also abundant in the upper limestone member of the Keyser in the Clifton Forge and Gala sections. It is impossible to say whether this occurrence should be correlated exactly with that in the Franklin-Monterey area. In Maryland *C. rectilineata* seems to be restricted to the lower part of the Keyser, but it occurs in both the lower and upper parts in Pennsylvania.

9. Petersburg *Stromatopora* reef: The *Stromatopora* reef that is developed at about the middle of the upper limestone member of the Keyser in the Petersburg section seems to be paralleled by similar occurrences in most of the other sections southward to Bolar, Va. The reefs in the upper limestones of the Bells Valley and Clifton Forge sections may also be placed here tentatively, but the exact correlation can not be regarded as certain. The presence of *Rensselaeria mutabilis* above this reef in the Big Mountain, Franklin, and Little Mountain sections indicates that this reef

is below the lower of the two reefs noted at Corriganville, in the Maryland area, which lies above the *R. mutabilis* zone.

10. *Meristella praenuntia* subzone: The work done in Maryland has shown that *Meristella praenuntia* ranges through a considerable distance, stratigraphically, in the upper part of the Keyser of that State. In the Virginias, however, it was found only between the *Stromatopora* reef described above and the overlying *Rensselaeria mutabilis* subzone. As it occurs at this horizon in considerable abundance and in a number of the sections, it will be assigned to this zonal position, for the Virginias.

11. *Rensselaeria mutabilis* subzone: *Rensselaeria mutabilis* has been shown to have a very definite stratigraphic position in the upper part of the Keyser of Maryland and Pennsylvania, and it occurs at the same horizon in the Big Mountain, Franklin, and Little Mountain sections.

12. *Camarotoechia* cf. *C. altiplicata* and *Nucleospira ventricosa* subzone: With the change from the more shaly limestones of the upper part of the Keyser of Maryland to the purer and more massive beds that are developed at that horizon around Franklin and to the south, *Camarotoechia altiplicata* and *Nucleospira ventricosa*, both of which are present in the Coeymans and New Scotland, range downward into the upper part of the Keyser, where they occur in and above the *Rensselaeria mutabilis* subzone. Although common and rather persistent, the species are rarely abundant. The zone is also occupied commonly by *Schuchertella prolifica*, several species of bryozoans, and some corals. Other forms are of rare occurrence. The Camarotoechias here referred to *C. altiplicata* are somewhat smaller than the typical material but are otherwise similar.

13. *Tentaculites gyracanthus* subzone: The *Tentaculites* subzone that is so characteristic of the upper part of the Keyser in Pennsylvania and Maryland was observed in the Virginias only in the section at Petersburg, W. Va., although it was searched for carefully elsewhere. Its disappearance to the south is concomitant with the development of the more massive and purer limestones of the upper limestone member.

14. *Spirifer vanuxemi* var. *prognosticus* subzone: *Spirifer vanuxemi* var. *prognosticus* occurs in abundance in a zone at the very top of the Keyser in the section at Franklin, W. Va., together with *Whitfieldella prosseri* and *Schuchertella prolifica.* As its appearance at this horizon seems to correspond to its occurrence in Maryland [6] it is noted here as marking a faunal subzone. Although no similar occurrence was observed elsewhere in the Virginias, the subzone is of interest, because it perhaps represents a higher horizon in the Keyser than is found in most of the sections. This conclusion is also suggested by the exceptional thickness of the upper limestone member of the Keyser, which is greater in the Franklin section than in the other sections studied. The absence of the zone elsewhere might be due to the presence of the minor hiatus that has generally been thought to mark the Keyser-Coeymans contact.

Kloedenia smocki, the one ostracode found in the Keyser in the Virginias, was discovered about 30 feet beneath the top of the Keyser in the Big Mountain section. Its presence there is of interest in that the horizon seems to correspond to the so-called Manlius of the New Jersey section, from which *K. smocki* was originally described by Weller.

A complete list of the Keyser fauna is given in the following table:

[6] Maryland Geol. Survey, Lower Devonian, p. 404, 1913.

Table 2.—*Distribution of the fauna of the Keyser limestone of West Virginia and Virginia*

[Note.—"r" represents a very closely related species]

	West Virginia-Virginia				Maryland-Pennsylvania				New Jersey				Eastern New York				Western New York		
	Keyser limestone					Keyser limestone													
	Chonetes jerseyensis zone																		
	Lower limestone and Big Mountain shale members	Base of upper limestone member	Favosites zone	Coeymans to Becraft limestones	Tonoloway limestone	Chonetes jerseyensis zone	Favosites zone	Coeymans to Becraft limestones	Decker limestone	Rondout limestone	So-called Manlius limestone	Coeymans to Becraft limestones	Cobleskill limestone	Rondout limestone	So-called Manlius limestone	Coeymans to Becraft limestones	Cobleskill limestone	Rondout limestone	Manlius limestone
COELENTERATA																			
1. Cyathophyllum radiculum Rominger	---	---	×	---	---	---	×	---	---	---	---	---	---	---	---	---	---	---	---
2. Striatopora bella C. K. Swartz	---	---	×	---	---	---	---	×	---	---	---	---	---	---	---	---	---	---	---
3. Favosites cf. F. helderbergiae var. praecedens Schuchert	---	---	×	---	---	---	×	---	---	---	---	---	---	---	---	---	---	---	---
4. Cladopora rectilineata Simpson	---	---	×	---	---	×	×	---	×	---	---	---	×	---	---	---	---	---	---
5. Aulopora schohariae Hall	---	---	×	---	---	---	×	×	---	---	---	---	---	---	---	×	---	---	---
6. Aulopora schucherti C. K. Swartz	---	---	×	---	---	×	?	×	---	---	---	---	---	---	---	---	---	---	---
7. Stromatoporoidea undetermined	---	---	×	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Favosites zone	---	---	6	---	---	2	5	3	1	---	---	---	1	---	---	1	---	---	---
ECHINODERMATA																			
8. Jaekelocystis hartleyi Schuchert	×	---	---	---	---	×	---	---	---	---	---	---	---	---	---	---	---	---	---
9. Mariacrinus sp	---	---	×	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Lower Keyser	1	---	---	---	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---
Favosites zone	---	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
BRYOZOA																			
10. Cyphotrypa corrugata (Weller)	×	---	---	---	---	×	---	---	×	---	---	---	---	---	---	---	---	---	---
11. Orthopora rhombifera (Hall)	×	---	---	---	---	×	×	×	---	---	---	---	---	---	---	×	---	---	---
12. Bryozoa undetermined	×	---	×	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Lower Keyser	2	---	---	---	---	2	1	1	1	---	---	---	---	---	---	1	---	---	---
BRACHIOPODA																			
13. Dalmanella concinna (Hall)	×	---	×	---	---	×	×	---	---	---	---	---	---	---	---	---	---	---	---
14. Rhipidomella emarginata (Hall)	×	---	---	---	---	×	---	---	---	---	---	---	---	---	---	---	---	---	---
15. Leptaena rhomboidalis (Wilckens)	×	×	×	×	---	×	×	×	×	---	---	×	×	---	---	×	×	---	---
16. Stropheodonta bipartita (Hall)	×	---	---	---	×	×	---	---	×	---	---	---	×	---	---	---	×	---	---
17. Stropheodonta cf. S. planulata (Hall)	---	×	---	×	---	---	---	×	---	---	---	---	---	---	---	×	---	---	---
18. Strophonella keyserensis C. K. Swartz	---	×	---	---	---	×	---	---	---	---	---	---	---	---	---	---	---	---	---
19. Schuchertella deckerensis (Weller)	×	---	---	---	---	×	---	---	×	---	---	---	---	---	---	---	---	---	---
20. Schuchertella sinuata (Hall and Clarke)	×	---	---	---	---	×	---	---	---	---	---	---	---	---	---	---	---	---	---

TABLE 2.—*Distribution of the fauna of the Keyser limestone of West Virginia and Virginia*—Continued

	West Virginia-Virginia				Maryland-Pennsylvania				New Jersey				Eastern New York				Western New York		
	Keyser limestone					Keyser limestone													
	Chonetes jerseyensis zone																		
	Lower limestone and Big Mountain shale members	Base of upper limestone member	Favosites zone	Coeymans to Becraft limestones	Tonoloway limestone	Chonetes jerseyensis zone	Favosites zone	Coeymans to Becraft limestones	Decker limestone	Rondout limestone	So-called Manlius limestone	Coeymans to Becraft limestones	Cobleskill limestone	Rondout limestone	So-called Manlius limestone	Coeymans to Becraft limestones	Cobleskill limestone	Rondout limestone	Manlius limestone
BRACHIOPODA—continued																			
21. Schuchertella prolifica Schuchert		×	×				×												
22. Chonetes jerseyensis Weller	×	×				×			×				×				×		
23. Gypidula coeymanensis var. prognostica Maynard		×				×													
24. Gypidula coeymanensis var. similis F. M. Swartz, n. var.		×		r		r		r				r				r			
25. Stenochisma deckerensis (Weller)	×					×			×										
26. Camarotoechia litchfieldensis (Schuchert)	×				×	×													
27. Camarotoechia gigantea Maynard		×				×													
28. Camarotoechia gigantea var. gigas F. M. Swartz, n. var		×																	
29. Camarotoechia cf. C. altiplicata (Hall)			×					×								×			
30. Uncinulus convexorus Maynard	×					×													
31. Uncinulus gordoni Maynard			×			×	×												
32. Uncinulus nucleolatus (Hall)			×			×	×	?				×				×			
33. Uncinulus angulatus Maynard			×			×													
34. Rensselaeria mutabilis (Hall)			×				×	×								×			
35. Atrypa reticularis (Linné)	×	×				×		×	×			×	×			×	×		
36. Spirifer modestus Hall	×					×													
37. Spirifer modestus var. plicatus Maynard	×					×													
38. Spirifer octocostatus (Hall)	×					×										×			
39. Spirifer vanuxemi var. prognosticus Schuchert			×			×	×												
40. Spirifer perlamellosus var. praenuntius F. M. Swartz, n. var.			×																
41. Cyrtina dalmani (Hall)			×													×			
42. Rhynchospira globosa (Hall)	×				×	×		×								×			
43. Rhynchospira formosa (Hall)	×					×			×			×				×			
44. Nucleospira swartzi Maynard	×					×													
45. Nucleospira ventricosa (Hall)			×			?	?	×				×				×			
46. Whitfieldella minuta Maynard	×					×													
47. Whitfieldella? prosseri Grabau			×				×												
48. Meristella praenuntia Schuchert and Maynard			×				×												
49. Meristella nasutaformis F. M. Swartz, n. sp.			×																
50. Merista typa (Hall)		×				×													
Lower limestone and Big Mountain shale [a]	16	1	1		3	16	1	1	5			1	2			1	2		
Base of upper limestone [a]	1	9	1	2		6	1	2	1			1	1			2			
Favosites zone [a]	1	1	14	1		5	8	6				2				5			

MOLLUSCA																			
51. Actinopteria cf. A. reticulata Weller			×			×			×										
52. Cypricardinia lamellosa Hall		×				×										×			
53. Straparollus welleri F. M. Swartz, n. name	×								×										
54. Tentaculites gyracanthus (Eaton)			×		×	×	×				×				×				×
55. Orthoceras cf. O. rigidum Hall	×						×									×			
Lower Keyser	2	1				1	1		1							2			
Favosites zone			2		1	2	2		1		1				1				1
ARTHROPODA																			
56. Proetus protuberans Hall	×					×		×				×				×			
57. Calymene camerata Conrad	×				×	×			×				×				×		
58. Kloedenia smocki (Weller)			×								×								
Lower Keyser	2				1	2		1	1			1	1			1	1		
Favosites zone			1								1								

[a] Atrypa reticularis and Leptaena rhomboidalis are excluded from this analysis.

CORRELATION

The essential continuity of the Keyser as a more or less definite lithologic unit from Maryland southward to Clifton Forge, Va., has already been demonstrated. The continuation of the series of faunal subzones, as shown in Figure 7 and Table 1, into the Keyser of Maryland and Pennsylvania is of even greater importance in the more exact correlation in time of the several horizons at which these subzones occur. The relationships of the fauna as a whole are given in Table 2 and the following analysis. *Atrypa reticularis* and *Leptaena rhomboidalis* are excluded from this and other analyses in this paper, because of their great stratigraphic range.

		Per cent
Species occurring in the Keyser limestone of West Virginia and Virginia	54	100
Previously described	49	91
Occurring in the Keyser of Maryland and Pennsylvania	43	82
Restricted to the Keyser in Maryland and Pennsylvania	27	54
Occurring in the Decker limestone to so-called Manlius limestone of New Jersey	14	27
Occurring in the Tonoloway limestone of Maryland	6	12
Occurring in the Coeymans to Becraft limestone of Maryland, etc	16	32
Occurring in the Coeymans to Becraft limestone of Maryland, etc., but not in the Keyser of Maryland	4	8
Forms very closely related to species occurring in the Coeymans to Becraft limestones of Maryland, etc., but without counterpart in the Keyser of Maryland	1	2

The large number of species common to the Keyser of the Virginias and that of Maryland and Pennsylvania and particularly the fact that the same species are generally the abundant ones in both areas indicate that the two areas were parts of one and the same depositional basin during Keyser time. The faunal zones recognized in these districts, as compared in Table 1, also show that the upper and lower limits of the Keyser are essentially of the same age from central Pennsylvania to west-central Virginia.

Ulrich,[7] C. K. Swartz,[8] and Reeside [9] have concluded that the interval from the Decker to the so-called Manlius of the New Jersey section, as described by Weller,[10] is the essential equivalent of the Keyser of the central Appalachian area. There is some uncertainty, at least in the minds of Swartz and Reeside, as to the relationships of the Keyser and also of the Decker, Rondout, and so-called Manlius of New Jersey to the Cobleskill, Rondout, and Manlius of New York. For further discussion the reader is referred to the Maryland Lower Devonian report.

An interesting point in the comparison of the Keyser and the New Jersey formations, seemingly unnoted before, is the similarity in position of the *Gypidula coeymanensis* var. *prognostica* subzone of the middle Keyser, and the *Gypidula* ("*Pentamerus*") *circularis* zone in the Decker. Both zones are near the top of the range of *Chonetes jerseyensis* in their respective sections. The general variability of the Gypidulas of the middle Keyser has been noted; in the Virginias they are large and comparable in size, at least, to the typical material; at the same horizon in Maryland and Pennsylvania the Gypidulas have been separated as the variety *prognostica*, chiefly because of their smaller size. With these facts in mind it seems possible, if not probable, that the *Gypidula circularis* of the Decker is the stunted counterpart of the Gypidulas of the Middle Keyser to the southwest. Evidence suggesting less favorable environment to the north is found in the changes going on in the fauna as a whole, which led Reeside [11] to conclude that the Keyser of Maryland, Pennsylvania, and New Jersey was "laid down in a basin which was connected with the open sea in the Maryland region and was progressively restricted northward and eastward to New Jersey."

Dunbar [12] has recently shown that the basal Devonian beds of western Tennessee, composed of the Rockhouse shale and the lower portion of the overlying Olive Hill formation, are perhaps of Keyser age. This view is suggested because the faunas of the Olive Hill, as a whole, and of the succeeding Birdsong shale are very definitely Coeymans and New Scotland, respectively, whereas there is no indication of a break of any magnitude between the Rockhouse and the Olive Hill which might make the former pre-Keyser. The fauna of the Rockhouse is too poor to give direct evidence of the age of the formation. It is an interesting hypothesis that the deposition of the Rockhouse shale resulted from the same crustal movements that caused the Virginia area to be flooded by sand and mud during the deposition of the Clifton Forge sandstone and Big Mountain shale members of the Keyser. The generally clearer water deposits of the Olive Hill might then be equivalent to both the upper Keyser and Coeymans limestones of the Virginias.

COEYMANS LIMESTONE

CHARACTER AND THICKNESS

The Coeymans of West Virginia and Virginia (see fig. 9) is in general a well-defined lithologic unit, consisting, as in Maryland, of very massive, crystalline, highly crinoidal limestone as far south as Clifton Forge, Va. It is separated from the underlying Keyser by a fairly sharp contact, the basal portion of the Coeymans

[7] Ulrich, E. O., U. S. Geol. Survey Geol. Atlas, Pawpaw-Hancock folio (No. 179), p. 8 (notes on lowest and second faunal zones of the Helderberg limestone), 1912.

[8] Swartz, C. K., Maryland Geol. Survey, Lower Devonian, pp. 105-117, 1913.

[9] Reeside, J. B., jr., The Helderberg limestone of central Pennsylvania: U. S. Geol. Survey Prof. Paper 108, pp. 193-197, 1917.

[10] Weller, Stuart, Paleontology of New Jersey, vol. 3, pp. 62-80, 1903.

[11] Reeside, J. B., jr., op. cit., p. 204.

[12] Dunbar, C. O., Stratigraphy and correlation of the Devonian of western Tennessee: Tennessee Geol. Survey Bull. 21, pp. 40-42, 54, 1919.

being somewhat sandy, particularly toward the south. The passage into the overlying New Scotland is transitional where that formation is a cherty limestone; but south of Monterey, where at least the lower portion of the New Scotland is highly arenaceous, the contact as here drawn becomes clear cut. At Bells Valley, Va., however, the New Scotland and Coeymans combine to form a massive limestone, which is highly crinoidal throughout, with diagnostic New Scotland fossils, including *Spirifer macropleurus*, restricted to the upper 20 feet. Only 4 miles to the north, along the same line of outcrop, at Craigsville, the *Spirifer macropleurus* zone is again cherty.

At Fawcetts Gap, in Little North Mountain, near Winchester, Va., the Coeymans is possibly absent. At least, faunal evidence of its presence is lacking, although analogy with the Bells Valley section might suggest Coeymans age for the lower few feet of the crinoidal limestone, which in its upper portion contains *Spirifer macropleurus*. Because of concealment of the interval, no data concerning the Coeymans were available at Wardensville, W. Va., and Fulks Run, Va.

The Coeymans seems to disappear in southwestern Virginia, as the calcareous sandstones in contact with the limestone of Tonoloway age at the sections near Saltville and Big Stone Gap are most probably of New Scotland age or even younger, although the faunal evidence observed is not altogether conclusive. At Rocky Gap, however, abundant casts of *Meristella* were observed in some weathered cherty fragments in the otherwise concealed interval just beneath the supposedly New Scotland sandstones; and this zone is possibly equivalent to the *Meristella arcuata* zone at the top of the Coeymans in the Clifton Forge area.

FAUNA AND CORRELATION

The Coeymans is in general abundantly fossiliferous. *Gypidula coeymanensis*, the most diagnostic form of this horizon, although profuse at Petersburg, W. Va., was found only after search in the sections from Petersburg to Hot Springs, Va., and seems to be entirely absent in the Coeymans at Bells Valley, Dry Run, and Clifton Forge. The specimens from the Coeymans in the Virginias are generally smaller than the material from New York and New Jersey and even from Maryland. Thus in the Virginias the Gypidulas of the middle Keyser are both larger and much more abundant than those of the Coeymans itself.

A very useful zone in the Monterey-Clifton Forge area is found at the top of the Coeymans, where *Meristella arcuata* occurs in considerable abundance, together with *Rhipidomella oblata* and, less commonly, *Uncinulus abruptus*. This zone is just beneath the sandy beds placed at the base of the New Scotland in the Bolar, Dry Run, Clifton Forge, and Gala sections and beneath the range of *Spirifer macropleurus* and *Dalmanella perelegans* in the Monterey and Bells Valley sections. The occurrence of this zone about 50 feet above the base of the crinoidal

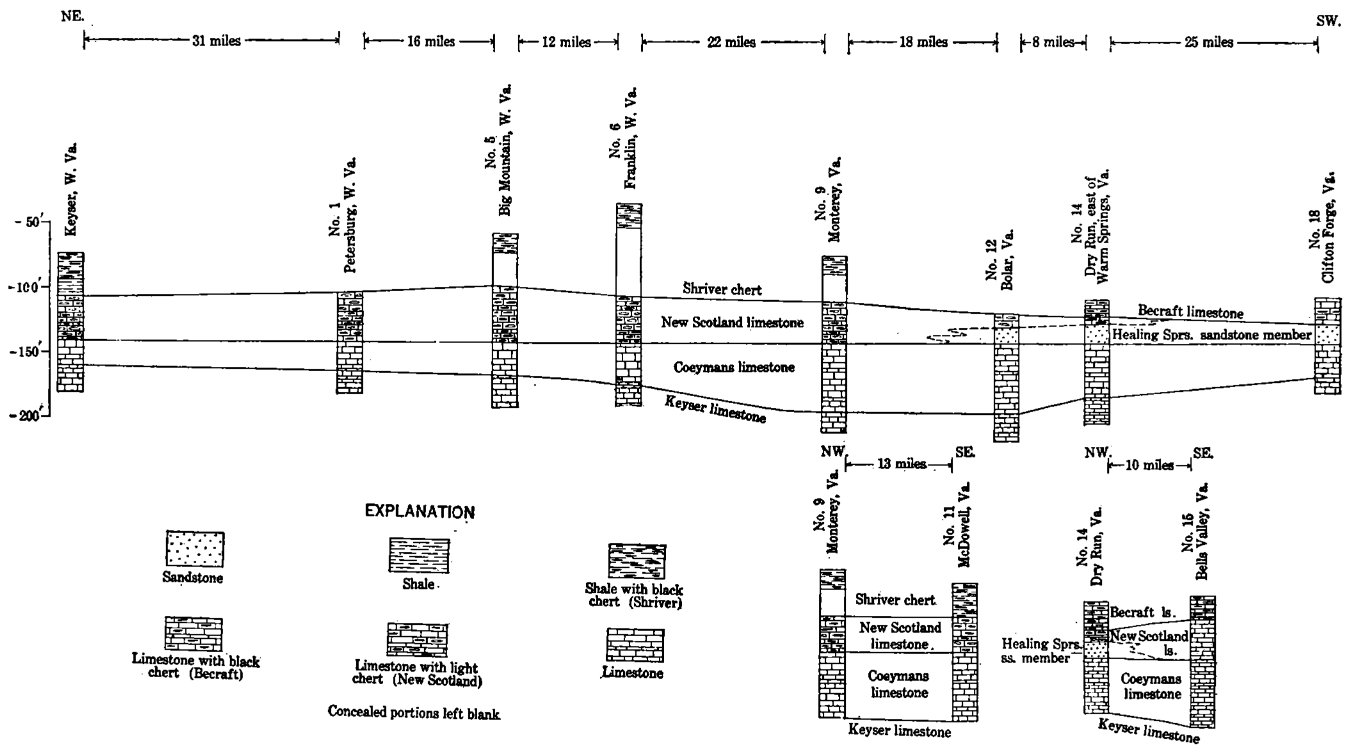


FIGURE 9.—Sections showing stratigraphy of the Coeymans and New Scotland limestones from Keyser, W. Va., to Clifton Forge, Va., as interpreted by F. M. Swartz. (Section at Keyser after description in Maryland Geol. Survey, Lower Devonian, p. 167, 1913)

limestone in the Monterey and Bells Valley sections is one of the chief reasons for considering the lower two-thirds of that unit Coeymans in age, rather than referring the whole of it to the New Scotland.

The following analysis shows the general relationships of the fauna; details of distribution are given in Table 3.

		Per cent
Species occurring in the Coeymans limestone of West Virginia and Virginia	18	100
Previously described	16	89
Occurring in the Coeymans of Maryland, New Jersey, and New York	11	68
Occurring in the New Scotland limestone of Maryland, etc	14	87
Occurring in the New Scotland of Maryland, etc., but not reported from the Coeymans or below	4	23
Occurring in the Keyser limestone (including two closely related forms)	4	23

These figures would seem to indicate a closer relationship to the fauna of the New Scotland of the Northern States than to that of the Coeymans. That the fauna and the beds containing it are Coeymans is shown by the presence of *Gypidula coeymanensis*, the most diagnostic species of the formation, and by the absence of *Spirifer macropleurus* and other characteristic fossils of the New Scotland, which appear in the overlying beds. Again, only one of the four species not noted from the Coeymans or lower horizons farther north—namely, *Meristella arcuata*—is at all abundant, and in some of the sections in the Virginias this form ranges well down into the Coeymans, occurring with *Gypidula coeymanensis* in the Big Mountain section. Of the other three, *Platyceras gibbosum* is represented only by two specimens and should not be considered decisive; and *Uncinulus abruptus* and *Camarotoechia campbellana* occur, in the Coeymans, only in the topmost beds, in the *Meristella arcuata* zone of the Monterey-Clifton Forge area, where they are not very common. This zone is placed in the Coeymans because it is below the range of *Spirifer macropleurus* in the sections where that form occurs, and particularly because it seems most logical to draw the New Scotland-Coeymans contact at the base of the Healing Springs sandstone member in the Bolar, Dry Run, and Clifton Forge sections. The zone seems, however, to be above the range of *Gypidula coeymanensis* and might well be considered equivalent to the beds included at the base of the New Scotland in Maryland, which are below the range of *Spirifer macropleurus*. The difference in usage is, of course, a reflection of the changes in lithology in the Clifton Forge area.

The close agreements between the faunas of the Olive Hill formation and the Birdsong shale of western Tennessee and those of the Coeymans and New Scotland of the Appalachian Basin, respectively, as recently reported by Dunbar,[13] indicate an open seaway between the two areas during the time of deposition of those formations. The middle of this passageway was probably in West Virginia and Kentucky, some distance west of Clifton Forge, Hollybrook, and Big Stone Gap, as shore and land conditions seem to be indicated in these localities.

NEW SCOTLAND LIMESTONE

CHARACTER AND THICKNESS

No essential changes in lithology, fauna, or thickness were noted in the cherty phase of the New Scotland from Keyser, W. Va., to Monterey, Va. (See fig. 9.) Throughout this area the formation is a cherty limestone, generally massive, and usually well exposed. Here, as in Maryland, the cherts are always white or at least light colored, even when quite fresh. The beds immediately overlying the cherty limestone are generally concealed and covered by débris; consequently there is some uncertainty as to the extent to which the shaly phase of the upper part of the New Scotland of western Maryland, so classed because of its contained fauna, is developed in the area studied. At Monterey and to the north the upper boundary of the New Scotland has everywhere been placed at the top of the exposed cherty limestone. Whether a portion of the overlying beds should be included in the formation must remain a problem for further investigation. At McDowell, however, and where the Becraft limestone is developed, the shaly phase is definitely absent.

To the east, at Fawcetts Gap, near Winchester, Va., the New Scotland is represented by a crinoidal limestone carrying *Spirifer macropleurus*. A somewhat similar occurrence is found at Bells Valley, Va., where *S. macropleurus* and other New Scotland fossils occur in the upper 20 feet of a very massive crinoidal limestone 78 feet thick. As noted in the discussion of the Coeymans limestone, the lower 50 feet or so of this limestone is apparently of Coeymans age. At Fulks Run, about halfway between Fawcetts Gap and Bells Valley and nearly along the strike line between them, the New Scotland is again a cherty limestone; and the *Spirifer macropleurus* zone is cherty at Craigsville, only about 4 miles north of Bells Valley, and along the same belt of outcrop.

As has been stated in the description of the Keyser, it has not been found feasible to correlate definitely the supposedly Lower Devonian limestones of the Seven Fountains section, in the Massanutten Mountain syncline, with the Helderberg. At least, no beds carrying the diagnostic New Scotland or other Helderberg faunas were observed in this section, and the lithologic sequence is not sufficiently similar to that of

[13] Dunbar, C. O., op. cit., pp. 54, 62–65.

the more western sections to permit confident correlation.

South of Monterey, Va., the New Scotland becomes highly arenaceous in its lower half, as at Bolar and Dry Run, and finally the calcareous sandstone (in part arenaceous limestone) entirely replaces the cherty limestone, forming what is here named the Healing Springs sandstone member. This sandstone contains some fossils in the sections west of Healing Springs and Warm Springs, including a few fragments of *Spirifer macropleurus*, but it seems to be quite unfossiliferous at Clifton Forge and Gala.

The New Scotland may be represented in the sandstones below the dark siliceous cherty beds of the Hollybrook, Rocky Gap, and Saltville sections. As stated in the descriptions of the Coeymans, a *Meristella* zone, possibly representing the *M. arcuata* of the upper Coeymans, was found just beneath these sandstones at Rocky Gap. The overlying dark cherts are unfossiliferous at Hollybrook and Cove Mountain, where they were first studied, and I then considered them definitely representative of the Shriver chert, although that formation had been last seen just west of Covington, about 70 miles to the north. Beds of upper Oriskany or Ridgeley age were thought to be entirely absent, as the Ridgeley is only a few feet thick near Clifton Forge, and the chert in the Hollybrook and Cove Mountains area is overlain by a thin sandstone carrying an Onondaga fauna. Study of the section on Tumbling Creek, 6 miles southwest of Saltville, cast a somewhat different light upon the subject, as the apparently equivalent cherts and some associated sandstones seen there carry *Diaphorastoma ventricosum*, *Ambocoelia umbonata* and a *Spirifer* which is essentially *S. arenosus*, although it differs slightly in its general aspect from the typical material of the Ridgeley sandstone farther north. Because of the pronounced flattening of its ribs it is here named *Spirifer arenosus* var. *planicostatus*. The presence of these fossils indicates late Oriskany age for the chert and associated beds containing them and suggest relationship to the Harriman chert of western Tennessee. Other upper Oriskany species, such as *Rensselaeria marylandica* and *Spirifer murchisoni* of the Ridgeley sandstone, were not observed. Charles Butts has referred a similar but unfossiliferous chert formation occurring near Gate City, Va., to the Onondaga, as there seemed to be no adequate reason for separating it from the overlying beds that carry an Onondaga fauna. The presence of *Ambocoelia umbonata* gives something of a Middle Devonian cast to the fauna of the chert at Tumbling Creek, but it seems best to regard the chert as being of late Oriskany age, though possibly with deposition continuous into the overlying Onondaga deposits.

At the Tumbling Creek section the lowest beds containing *S. arenosus* var. *planicostatus* are underlain by about 7 feet of calcareous sandstone in which was observed a *Rhipidomella* suggestive of *R. oblata* and a large *Dalmanella* that should probably be referred to *D. perelegans*. This sandstone should probably be correlated with the Becraft limestone, or possibly with the New Scotland. It is underlain in turn by the limestone of Tonoloway age.

Comparison of the sections at Hollybrook and Rocky Gap with that at Tumbling Creek strongly suggests that the chert of the Hollybrook-Rocky Gap area should be correlated with that seen at Tumbling Creek and should thus be considered younger than the Shriver chert of the sections farther north. At Hollybrook and Rocky Gap the chert is underlain by about 60 feet of calcareous sandstone, the upper part of which is somewhat conglomeratic, and contains at both localities a profusion of the empty molds of a cuplike crinoid fossil. Charles Butts, of the United States Geological Survey, examined a specimen of the crinoid and immediately referred it to *Aspidocrinus*, suggesting comparison with *A. scutelliformis* Hall, of the New York Becraft. Edwin Kirk agreed to the generic reference but thought it was not the species mentioned. It is here named *Aspidocrinus caroli*. The lower portion of the sandstone contains a few poorly preserved fossils, including a *Spirifer* close to *S. cyclopterus*. One fragmental cast is suggestive of *Trematospira equistriata*. The sandstone may be considered roughly equivalent to the Becraft and New Scotland limestones, although the fauna is a poor one for the purpose of correlation.

The similar calcareous sandstone that forms the basal portion of the Devonian in the vicinity of Big Stone Gap has been referred to the New Scotland by Ulrich, although no faunal lists accompanied Ulrich's published section.[14] The fauna, so far as I observed it during a short visit, is not distinctively New Scotland; that of the upper beds is in fact suggestive of the Onondaga. Further work must be done before the correlation of this sandstone can be considered assured

FAUNA AND CORRELATION

North of Healing Springs, Va., the beds assigned to the New Scotland carry *Spirifer macropleurus* and other characteristic New Scotland fossils. These are only sparingly present in the Healing Springs sandstone member but are generally profuse where the New Scotland consists of the cherty limestone. *Streptelasma strictum*, *Dalmanella perelegans*, and *Meristella arcuata* are generally very abundant.

[14] Ulrich, E. O., in Eby, J. B., The geology and mineral resources of Wise County, Va.: Virginia Geol. Survey Bull. 24, p. 40, 1923.

The New Scotland age of this fauna is shown by the following relationships:

		Per cent
Species occurring in the New Scotland limestone of West Virginia and Virginia	16	100
Previously described	16	100
Occurring in the New Scotland of Maryland, New Jersey, and New York	15	94
Occurring in the Coeymans limestone of Maryland, etc	6	38
Occurring in the Becraft limestone of Maryland, etc	11	68
Occurring in the Becraft of Maryland, etc., but not in the New Scotland	1	6
Restricted to the New Scotland in Maryland, etc	4	25

Further details of the distribution of the fauna are given in Table 3.

BECRAFT LIMESTONE

CHARACTER AND THICKNESS

At Clifton Forge, Gala, Healing Springs, Dry Run, and Bells Valley, Va., the interval between the New Scotland formation and the Ridgeley sandstone is occupied by a limestone 100 to 120 feet thick, with much interbedded black chert, chiefly in the lower and middle parts. (See fig. 10.) The upper chert-free beds are also present on Back Creek Mountain, along the road from Warm Springs to Driscol. The stratigraphic position of the limestone and more particularly the presence in it of *Spirifer concinnus*, *Rhipidomella assimilis*, and certain other fossils indicate general equivalency to the Becraft limestone of southeastern New York, New Jersey, and central Maryland. This limestone is also present farther north in Virginia, where it was seen at Fawcetts Gap, near Winchester; and it has been reported to occur in the vicinity of Cherry Run, W. Va.,[15] along the same general belt of outcrop.

FAUNA AND CORRELATION

The relationships of the fauna collected from this limestone in the Clifton Forge area are given in detail in Table 3 and are summed up in the following analysis:

		Per cent
Determined species occurring in the Becraft limestone of west-central Virginia	25	100
Previously described	24	96
Occurring in the Becraft limestone of Maryland, New Jersey, and New York	18	72
Occurring in the New Scotland limestone of Maryland, etc	14	56
Occurring in the New Scotland limestone of Maryland, etc., but not in the Becraft limestone of those States	1	4
Occurring in the Ridgeley sandstone of Maryland and the Oriskany sandstone of New Jersey and New York	9	36
Occurring in the above-mentioned sandstones but not previously reported from lower beds	3	12
Occurring in the Shriver chert of Maryland	2	8
Restricted to the Becraft limestone, at least in Maryland	3	12

[15] Maryland Geol. Survey, Lower Devonian, p. 179, 1913.

The large percentage of New Scotland survivors, together with the absence of such typical New Scotland forms as *Spirifer macropleurus* and *Spirifer perlamellosus*, is in itself very characteristic of the Becraft; particularly as these characteristic New Scotland species are found in the underlying beds. Of the New Scotland survivors, *Streptelasma strictum*, *Edriocrinus pocilliformis*, and *Uncinulus abruptus* are prominent in the lower part of the Virginian Becraft but do not appear to range higher. The two specimens of *Phacops logani* were also found in these lower beds, as were the fragmental Rensselaerias referred to *R. aequiradiata*, a species of the New York Becraft. On the other hand, *Schuchertella woolworthana* ranges throughout the Becraft, as do the new arrivals, *Rhipidomella assimilis*, *Eatonia peculiaris*, and *Spirifer concinnus*.

As has been noted, the upper part of the Becraft of the Clifton Forge area differs from the lower and middle portions in that as a rule it is quite free from interbedded black chert and is in many places massive and highly crinoidal. It also carries a fauna that differs somewhat from that of the lower portions, notwithstanding the presence of the four species last named. This is due not only to the absence of those New Scotland survivors which, as mentioned above, are restricted in the Becraft to its lower portion, but also to some progressive evolution in *Rhipidomella assimilis* and *Spirifer concinnus*, as well as to the introduction of a number of forms with definite Oriskany affinities.

Thus *Rhipidomella assimilis*, although mentioned above as ranging throughout the formation, is somewhat more abundant in these upper beds and tends to reach a somewhat greater size, with stronger muscle scars, so that whereas the average size of the species is 26 millimeters in length and 28 millimeters in width, several specimens from the upper part of the Becraft at Gala and Back Creek Mountain reach dimensions of 36 and 40 millimeters, respectively; and the ventral muscle scar is as much as 22 millimeters long, as compared with 17 millimeters for the more typical material. These specimens are approaching the *R. musculosa* of the Ridgeley sandstone, which measures 37 millimeters in length and 40 millimeters in width, with a ventral muscle scar as much as 27 millimeters long.

A similar development was noted in *Spirifer concinnus*. Associated with the more typical members of that species there are commonly in these upper Becraft beds specimens which have the same general aspect but differ in the presence of 16 or more plications on the lateral slope, as compared to the typical 12 to 14, and which also reach widths of 40 millimeters, as compared to the average 24 millimeters. This larger, more alate, and more numerously plicated form is here named *Spirifer concinnus* var. *progradius* and is thought to have developed out of earlier members of

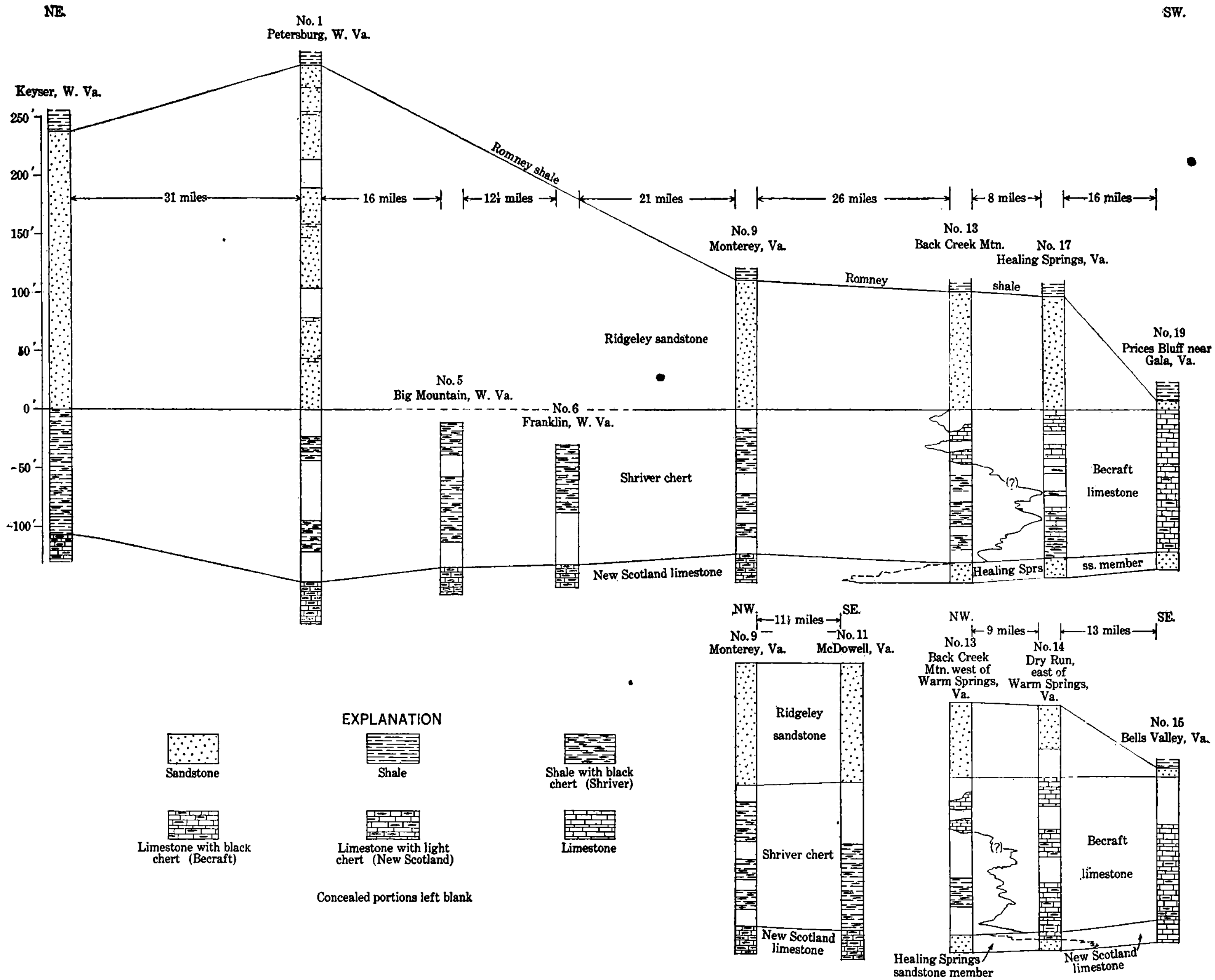


FIGURE 10.—Sections showing stratigraphy of the Shriver chert and Becraft limestone from Keyser, W. Va., to Clifton Forge, Va. See page 63 for correction of Shriver-Ridgeley interval at Monterey, Va.

S. concinnus proper. According to Schuchert,[16] some specimens of *S. concinnus* of the Maryland Becraft also reach widths of 40 millimeters or even more; he does not, however, mention the development of more numerous plications. The Maryland specimens are also more alate than those from New York—a condition that Schuchert ascribes to abrasion of the thin edges of the valves before deposition.

In the ventral valve used as the type for *Spirifer concinnus* var. *progradius* the sinus is broad and rounded; but in several of the associated and otherwise similar specimens the sinus is somewhat angulated and has a very feeble plication on each side toward the front. Similar but stronger plications were noted by Hall in some of the New York specimens which he included in *S. concinnus* but which Schuchert [17] later separated as *S. proavitus*. However, as the Virginian specimens agree more closely on the whole with the variety *progradius*, they are here retained in that form.

The presence at this horizon of a number of species with very definite Oriskany affinities is of particular interest. The most significant of these species are *Rhynchotrema cumberlandicum*, *Rensselaeria subglobosa*, *Spirifer angularis*, *Cyrtina varia*, *Anoplotheca flabellites*, and *Dalmanites* sp. near *D. dentatus*. The relationships of this portion of the fauna are as follows:

Relationships of Oriskany species in Becraft limestone

[r, closely related species]

	Becraft limestone, Maryland	Lower part of Oriskany, New Jersey	Shriver chert, Maryland	Ridgeley sandstone, Maryland
Rhynchotrema cumberlandicum				×
Rensselaeria subglobosa	×	×		
Spirifer angularis				×
Cyrtina varia		×		
Anoplotheca flabellites	×		×	×
Dalmanites sp		r		r

Of these species, *Rhynchotrema cumberlandicum* is rare. The one specimen from the upper part of the Becraft at Prices Bluff is, however, complete and quite typical. *Rensselaeria subglobosa* is common in the limestones of this horizon at Back Creek Mountain. The material so identified is quite typical in size and in the number and strength of the plications; but, as in that from the Maryland Becraft, the valves are somewhat less convex than in the New Jersey types. *Spirifer angularis* is common in several of the sections and seems to agree rather closely with the material from the Maryland Ridgeley. The presence of *Cyrtina varia* in the upper Becraft of Virginia parallels the occasional occurrence of *C. rostrata* in the Becraft limestone of Maryland. As was noted by Clarke in his original description, *C. varia* is very close to *C. rostrata*, from which it differs in its somewhat smaller size and smaller average number of plications. *Anoplotheca flabellites*, although typically an Oriskany fossil, has been reported from both the Becraft and the New Scotland of Maryland. Finally, the trilobite listed as *Dalmanites* sp. is of interest because the several fragments of pygidia are sufficient to show that the general shape of the pygidium, the width of its axis, the strength, grooving, and probably the number of the ribs, and the general character of the ornamentation are essentially those of *D. dentatus*. The ornamentation differs somewhat in detail from that of *D. dentatus* in that the tubercles lining the summits of the ribs are appreciably more numerous and not quite so prominent—a character which, with the granulation of the surface as a whole, suggests that this form was intermediate between *D. pleuroptyx* of the New Scotland and Becraft and *D. dentatus* of the Oriskany.

The Oriskany elements in this fauna indicate that the upper part of the Becraft limestone of the Clifton Forge area is younger than the top of the Becraft in the type area of southeastern New York. They also suggest that there is no essential time break between the Becraft and the Ridgeley such as there would be if the pre-Ridgeley Shriver chert of the neighboring sections to the north were entirely post-Becraft. This question is considered in further detail in the following discussion of the Shriver chert.

[16] Maryland Geol. Survey, Lower Devonian, p. 418, 1913.

[17] Idem, pp. 418, 419, pl. 71, fig. 17.

TABLE 3.—*Distribution of the fauna of the Coeymans, New Scotland, and Becraft limestones of West Virginia and Virginia*

Coeymans limestone

[r = very closely related form]

	West Virginia and Virginia				Maryland and Pennsylvania					New Jersey and New York		
	Keyser	Coeymans	New Scotland	Becraft	Keyser	Coeymans	New Scotland	Becraft	Ridgeley	Coeymans	New Scotland	Becraft
BRACHIOPODA												
1. Orthostrophia strophomenoides (Hall)		×			×		×				×	
2. Rhipidomella oblata (Hall)		×				×	×			×	×	×
3. Leptaena rhomboidalis (Wilckens)	×	×	×	×	×	×	×		×	×	×	×
4. Stropheodonta arata (Hall)		×				×	×			×	×	
5. Stropheodonta cf. S. planulata (Hall)	×	×				×	×	×		×		×
6. Strophonella punctulifera (Conrad)		×	×			×	×			×	×	×
7. Schuchertella woolworthana (Hall)		×	×	×		×	×	×			×	×
8. Gypidula coeymanensis Schuchert	r	×			×	×				×	×	
9. Camarotoechia campbellana (Hall)		×		×			×				×	×
10. Uncinulus abruptus (Hall)		×	×	×			×	×			×	
11. Spirifer perlamellosus Hall	r	×	×			?	×				×	
12. Spirifer cyclopterus Hall		×	×	×		×	×	×		×	×	×
13. Nucleospira ventricosa (Hall)	×	×			×	×	×			×	×	
14. Meristella arcuata (Hall)		×	×				×	×			×	
15. Meristella arcuata var. gigas F. M. Swartz, n. var		×										
16. Meristella symmetrica Schuchert		×				×						
MOLLUSCA												
17. Platyceras gibbosum Hall		×									×	
18. Platyceras multiplicatum F. M. Swartz, n. sp		×										
19. Platyceras trilobatum Hall		×					×				×	
TRILOBITA												
20. Dalmanites pleuroptyx (Green)		×					×		×	×	×	

New Scotland limestone

	Keyser	Coeymans	New Scotland	Becraft	Keyser	Coeymans	New Scotland	Becraft	Ridgeley	Coeymans	New Scotland	Becraft
PORIFERA												
1. Hindia sphaeroidalis Duncan			×					×		×	×	
COELENTERATA												
2. Streptelasma strictum Hall			×	×			×			×	×	
BRACHIOPODA												
3. Dalmanella eminens (Hall)			×				×				×	
4. Dalmanella perelegans (Hall)			×			×	×				×	
5. Leptaena rhomboidalis (Wilckens)	×	×	×	×	×	×	×	×	×	×	×	×
6. Strophonella punctulifera (Conrad)		×	×			×	×			×	×	×
7. Strophonella undaplicata C. K. Swartz			×				×					
8. Strophonella leavenworthana (Hall)			×	×		×	×				×	×
9. Schuchertella woolworthana (Hall)		×	×	×		×	×	×		×	×	×
10. Uncinulus abruptus (Hall)		×	×	×			×	×			×	
11. Spirifer perlamellosus Hall	r	×	×			?	×			×	×	
12. Spirifer cyclopterus Hall		×	×	×		×	×	×		×	×	×
13. Spirifer macropleurus (Conrad)			×				×				×	
14. Trematospira equistriata Hall and Clarke			×				×					
15. Meristella arcuata (Hall)		×	×				×	×			×	
16. Meristella lata (Hall)			×	×				×	×			
MOLLUSCA												
17. Platyceras gebhardi Conrad			×	×			×		×		×	×

TABLE 3.—*Distribution of the fauna of the Coeymans, New Scotland, and Becraft limestones of West Virginia and Virginia*—Continued

Becraft limestone

	West Virginia and Virginia				Maryland and Pennsylvania					New Jersey and New York		
	Keyser	Coeymans	New Scotland	Becraft	Keyser	Coeymans	New Scotland	Becraft	Ridgeley	Coeymans	New Scotland	Becraft
COELENTERATA												
1. Streptelasma strictum Hall			×	×			×			×	×	×
ECHINODERMATA												
2. Edriocrinus pocilliformis Hall				×			×	×			×	×
BRACHIOPODA												
3. Rhipidomella assimilis (Hall)				×				×			×	×
4. Leptaena rhomboidalis (Wilckens)	×	×	×	×	×	×	×	×	×	×	×	×
5. Strophonella leavenworthana (Hall)			×	×		×	×				×	×
6. Schuchertella woolworthana (Hall)		×	×	×		×	×	×		×	×	×
7. Schuchertella becraftensis (Clarke)				×					×			
8. Rhynchotrema cumberlandicum (Rowe)				×					×			
9. Camarotoechia campbellana Hall		×		×			×				×	×
10. Camarotoechia praespeciosa Schuchert				×				×				
11. Uncinulus abruptus (Hall)		×	×	×			×	×			×	
12. Uncinulus pyramidalis (Hall)				×						×	×	
13. Uncinulus vellicatus (Hall)				×			×	×			×	×
14. Eatonia peculiaris (Conrad)				×			×	×	×		×	
15. Rensselaeria cf. R. aequiradiata (Conrad)				×								×
16. Rensselaeria subglobosa (Weller)				×				×				
17. Spirifer concinnus Hall				×				×				×
18. Spirifer concinnus var. progradius F. M. Swartz, n. var.				×								
19. Spirifer cyclopterus Hall		×	×	×		×	×	×		×	×	×
20. Spirifer cf. S. angularis Schuchert				×					×			
21. Spirifer sp.				×					?			
22. Cyrtina varia Clarke				×				r	r			
23. Anoplotheca flabellites (Conrad)				×			×	×	×			
24. Meristella lata (Hall)			×	×				×	×			
MOLLUSCA												
25. Platyceras gebhardi Conrad			×	×			×		×		×	×
TRILOBITA												
26. Phacops logani Hall				×			×				×	×
27. Dalmanites sp.				×					r			

SHRIVER CHERT

From Cumberland, Md., to Monterey, Va., the New Scotland is overlain by a siliceous shale with much interbedded black chert, containing few fossils. In the Virginias the shale is generally subordinate to the bedded chert. The shaly chert is in turn overlain by the Ridgeley sandstone, carrying the characteristic *Spirifer arenosus* fauna, and is thus the southward continuation of the Shriver chert, so named from Shriver Ridge,[18] near Cumberland, Md.

The stratigraphic position of the Shriver chert is apparently the same as that occupied in the Clifton Forge area by the Becraft limestone. The conditions are similar in Maryland, where the Becraft lies between the New Scotland limestone and the Ridgeley sandstone in the exposures of the Lower Devonian east of Hancock; whereas at Hancock and westward to Cumberland the Becraft is absent, and the New Scotland-Ridgeley interval is occupied by the Shriver chert. In Virginia, at least, the displacement of the one lithologic unit by the other is relatively abrupt, the Shriver chert and the Becraft limestone appearing in full thickness and more or less typical development in sections less than 20 miles apart.

Although no section containing both formations was known, Schuchert,[19] Stose and Ulrich,[20] and C. K. Swartz [21] considered the Shriver chert to be of lower Oriskany age because the Shriver seems to thin somewhat in the vicinity of Hancock, and, more particularly, because of the following faunal relationships:

Species occurring in the Shriver chert of Maryland	28
Previously described	11
Occurring in the Becraft limestone or older formations of Maryland and New York	6
Occurring in the lower part of the Oriskany sandstone of New York	9
Occurring in the Ridgeley sandstone of Maryland	8
Occurring in the Ridgeley sandstone of Maryland and the Oriskany sandstone of New York	11
Occurring in the last-named sandstones but not in the Becraft or below	5

The outstanding features of the faunal relationships indicated above are the large proportion of species known from only the Shriver chert, the lack of agreement with the fauna of the Becraft, and the apparent affinity to the fauna of the Oriskany sandstone rather than to that of the Helderberg.

A more detailed analysis of the fauna places a somewhat different aspect on the problem from that given by the above figures. Taking the five species cited above as occurring in the Ridgeley and other Oriskany sandstones but not in the Becraft or below, we find that two of these—*Spirifer arenosus* and *Diaphorastoma desmatum*—are represented in the Shriver by but one known specimen each, and that there is some doubt as to the identification, at least of the *Diaphorastoma.* A third form, *Schuchertella becraftensis*, is also very rare; furthermore, one specimen from the Becraft of Virginia seems to be referable to this species. *Meristella lentiformis* is abundant in many places and is more definitely indicative of Oriskany relationships; even so, the Shriver chert material is consistently smaller than that of the Ridgeley sandstone. The citation of *Chonetes hudsonicus* from the New Scotland of New York is questionable; although not included in the above list, this seems to be definitely an Oriskany species and is abundant and typical in the upper part of the Shriver chert. The Shriver species *Beachia suessana* var. *immatura* also suggests the Oriskany. It should furthermore be noted that none of the species common to the Shriver and to the Becraft or lower formations is distinctively Helderbergian. The fauna is then suggestive of the Oriskany, though not distinctively so.

Of greater significance is the fact that the composition of the Shriver chert fauna is not what would be expected in a typical marine deposit of Lower Devonian time. Thus of the 28 described species, 11 are ostracodes, of which 10 are known only from the Shriver, 14 are brachiopods, and 3 are gastropods—including 2 species of *Tentaculites*. The nonostracode fauna is thus small, in number of species as well as in individuals. Furthermore, the corals, Pelmatozoa, Bryozoa, and trilobites are notably absent, and the brachiopod fauna itself is strikingly defective in the absence of the Lower Devonian genera *Dalmanella*, *Rhipidomella*, and *Camarotoechia* and the occurrence of but one rare species of each of the genera *Stropheodonta*, *Eatonia*, and *Beachia*. The impoverishment of this fauna, particularly when compared with the faunas of the Helderberg limestone and the Ridgeley and equivalent Oriskany sandstones, detracts greatly from its significance as a means of correlation.

These facts and the further fact that the Shriver chert and the Becraft limestone occupy analogous stratigraphic positions in near-by sections, without, at least in Virginia, a thinning of either formation toward the general boundary of the areas of development, greatly weaken the argument for the post-Becraft age of the Shriver chert. Indeed, these relationships, together with the fact that the Becraft of Maryland and Virginia contains pronounced Oriskany elements in the fauna of its upper part, suggest the possibility that the Shriver chert is a muddy bottom phase, equivalent in time to the Becraft limestone as developed in Maryland and Virginia, and with a fauna differing from that of the Becraft because of peculiar environmental conditions.

Since this question formulated itself in my mind, I have fortunately been able to revisit several key sections. The sections at Healing Springs and Back Creek Mountain, west of Warm Springs, deserve par-

[18] Swartz, C. K., Maryland Geol. Survey, Lower Devonian, p. 91, 1913.
[19] Schuchert, Charles, U. S. Nat. Mus. Proc., vol. 26, p. 422, 1903.
[20] Ulrich, E. O., in Stose, G. W., U. S. Geol. Survey Geol. Atlas, Pawpaw-Hancock folio (No. 179), p. 9, 1912.
[21] Swartz, C. K., Maryland Geol. Survey, Lower Devonian, pp. 121, 122, 1913.

ticular attention. At Healing Springs the New Scotland to Ridgeley interval is occupied by the Becraft limestone. As usual for the Clifton Forge area, the upper part of the Becraft is composed of very pure, rather massive gray limestone, containing an abundance of *Spirifer concinnus* and also *Spirifer cyclopterus* and *Eatonia peculiaris*. The middle and lower portions are composed of interbedded limestone and black chert, of typical Becraft type, but with the limestone much more subordinate than in the sections at Clifton Forge and Gala. The lower beds also lack the abundance of fossils which characterize them in the Clifton Forge and Gala sections.

At Back Creek Mountain, about 8 miles farther north, the Ridgeley sandstone is underlain by chert-free but not overly pure limestones carrying *Spirifer concinnus*, *S. cyclopterus*, *Eatonia peculiaris*, and *Rhipidomella assimilis*. These beds certainly represent the upper part of the Becraft, as seen at Healing Springs, Clifton Forge, and Gala. They are underlain, however, by bedded impure chert of Shriver rather than Becraft aspect; particularly as it weathers for the most part into soft buff spongy fragments. A minor portion of the chert is purer than is typical of the Shriver, and this weathers into harder whitish fragments. Farther north, at Monterey, the limestone of upper Becraft age is underlain by shaly chert that is even more suggestive of the Shriver.

The evidence offered by the above-mentioned sections favors the view that the Shriver chert and the Becraft limestone of Maryland and the Virginias are at least essentially equivalent in time. As has been shown in the discussion of the Becraft, however, the limestones referred to that formation in Maryland and Virginia seem to range higher than the top of the type Becraft of New York, the upper portion probably being at least as young as the Port Ewen limestone of southeastern New York and New Jersey.

Acceptance of the conclusions suggested above would still leave some questions as to terminology. If Helderberg group is used primarily with a time significance, I think that the Shriver would necessarily be included in it, although admitting that the Shriver and the Becraft of the Virginia-Maryland area range above the top of the Becraft, and thus above the Helderberg of the type area of New York. On the other hand, if the term is accepted as having a lithologic and faunal rather than a purely time significance, then the Shriver would be placed in the Oriskany group, being more closely related in those aspects to the lower Oriskany of southeastern New York than to the members of the type Helderberg. This usage seems somewhat the more acceptable and has been followed here.

POSITION AND NATURE OF THE SILURIAN-DEVONIAN BOUNDARY IN THE APPALACHIAN BASIN

One of the most unsettled questions of Paleozoic stratigraphy, both in Europe and in North America, has been that concerning the position in our sections of the Silurian-Devonian boundary. This doubt has resulted largely because the type Silurian gives way above to brackish and fresh water and not marine deposits, because the base of the Devonian is scarcely defined in Devonshire, and because the relationships of the Lower Devonian in Shropshire, Devonshire, the Rhine Valley, the Harz Mountains, and the Bohemian basins are obscured by the development of facies in the faunas. The controversies over the possible upper Silurian or Lower Devonian age of the Hercynian shales of the Harz Mountains,[22] the Koniepruss (F_2) limestone of Bohemia,[22] the Helderberg of New York,[23] and the Downtonian of Shropshire [24] have been concerned with the various local aspects of the same general problem.

Limiting ourselves to the Appalachian area and considering only the later phases of the problem even as it concerns that province, we may note that it was not until the end of the nineteenth century that the efforts of Clarke and Schuchert led the majority of American stratigraphers to concur in placing the Helderberg group (in which Clarke and Schuchert [25] in 1899 included the Coeymans, New Scotland, and Becraft limestones) at the base of the Devonian. The assignment was made because of what Clarke and Schuchert considered the definite Devonian affinities of the fauna as a whole, as well as because of its apparent equivalency to that of the Koniepruss (F_2) limestone of Bohemia, generally accepted as basal Devonian by European geologists, following Kayser and Frech.

Somewhat more recently it has been suggested, first by Ulrich [26] and then in further detail by C. K. Swartz [27] and Reeside,[28] that the Keyser limestone, so named from its exposure at Keyser, W. Va., should be considered the lowest division of the Helderberg and as such placed at the base of the Devonian of the Appalachian province.

[22] For review and bibliography see Clarke, J. M., The Hercynian question: New York State Mus. Forty-second Ann. Rept., pp. 408-437, 1889.

[23] Schuchert, Charles, Lower Devonic aspect of the lower Helderberg and Oriskany formations: Geol. Soc. America Bull., vol. 11, pp. 241-332, 1900.

[24] Stamp, L. D., The base of the Devonian: Geol. Mag., vol. 60, pp. 276-281, 331-336, 367-372, 385-410, London, 1923. Robertson, T., The Siluro-Devonian junction in England: Idem, vol. 65, pp. 385-400, 1928.

[25] Clarke, J. M., and Schuchert, Charles, The nomenclature of the New York series of geological formations: Science, new ser., vol. 10, pp. 874-878, 1899.

[26] Ulrich, E. O., Revision of the Paleozoic systems: Geol. Soc. America Bull., vol. 22, p. 590, pl. 28, 1911.

[27] Swartz, C. K., Maryland Geol. Survey, Lower Devonian, pp. 105-117, 1913.

[28] Reeside, J. B., jr., The Helderberg limestone of central Pennsylvania: U. S. Geol. Survey Prof. Paper 108, pp. 193-199, 1917.

Several aspects of the present investigation are of interest in this connection. In the Virginias, as in Maryland and Pennsylvania, a considerable percentage of the Keyser fauna continues into the higher Helderberg, there being 17 Keyser species that are identical with or at least very closely allied to forms occurring in the Coeymans, New Scotland, and Becraft limestones. This is a somewhat smaller number than the 35 Coeymans or younger species reported from the Keyser of Maryland and Pennsylvania but is of interest in that several of the strictly Helderberg species that occur in the Keyser of the Virginias have not been reported from the Keyser of Maryland and Pennsylvania.

Among the Helderberg species occurring in the Keyser limestone of the Virginias, but not known from that formation farther north, the specimens placed in *Striatopora bella* and *Cyrtina dalmani* may be considered strictly identical with the Coeymans and New Scotland material. *Striatopora bella* is locally common in the upper part of the Keyser at Little Mountain, Va.; *Cyrtina dalmani* is comparatively rare but is known from several localities. Mention might also be made here of the abundance of typical *Nucleospira ventricosa* in the upper part of the Keyser at Big Mountain, Monterey, and several other localities, as there is possibly some uncertainty about previous citations of this species from the lower part of the Keyser farther north.

The upper Keyser Camarotoechias referred in this report to *C. altiplicata* are somewhat smaller than the New Scotland material but seem to be similar in other respects. They should perhaps be separated under a varietal name. The new *Spirifer perlamellosus* var. *praenuntius* is somewhat more definitely removed from the type material of the New Scotland, having fewer plications as well as being smaller; nevertheless, its similarity in aspect shows it to be closely related to the species to which it is introductive. The new *Gypidula coeymanensis* var. *similis* is also very suggestive of the younger material of the Coeymans limestone—more so than the variety *prognosticus* of the Keyser of Maryland and Pennsylvania.

The two middle Keyser Stropheodontas cited as *Stropheodonta* cf. *S. planulata* are more questionable and less significant. The one *Meristella* which I am using as the type of *M. nasutaformis*, although suggestive of the younger Meristellas in its relatively large size, does not seem to be closely akin to any of the later forms described; it is of importance, however, as another Keyser representative of this Devonian genus.

The presence of these additional Helderberg elements in the Keyser fauna gives some added weight to the opinion that the Keyser limestone should be considered a part of the Helderberg group. It must nevertheless be admitted that the number of well-determined species continuing from the Keyser into the younger beds is not exceptionally great. Thus out of the total known Keyser fauna of about 180 species, 38 are known from the Coeymans and younger formations, 9 are derived from the Tonoloway limestone, and an additional 11 come from other pre-Keyser formations. The Helderberg aspect of the Keyser fauna gains somewhat by comparison if the species of little stratigraphic value are eliminated. Thus among the 20 Keyser species derived from the known older faunas, *Halysites catenularia*, *Stromatopora constellata*, *Leptaena rhomboidalis*, and *Atrypa reticularis* are devoid of any exact stratigraphic significance, while two other species are also long ranging forms, as they continue into the New Scotland. Of the remainder, two are Bryozoa and four are ostracodes and are species whose stratigraphic value is somewhat uncertain. On the other hand, a similar removal of the stratigraphically worthless or obscure species from the Helderberg element of the Keyser fauna results in the loss of a smaller number.

Too much weight should not be placed upon such statistics as those above given without some examination of the faunas with which comparison is being made. A study of the faunas of the Coeymans, New Scotland, and Becraft limestones shows them to consist of many and varied species and leads to the conclusion that these faunas are finely representative of the marine life of their day. This is not true of the faunas of the Cayugan deposits, including the Tonoloway limestone. Thus in New York the impoverishment of the Cayugan faunas has long been recognized and was suggested by Schuchert [29] to be an important factor contributing to the notable break in that State between the faunas of the Helderberg group and those of the subjacent Silurian deposits.

The case is similar in Maryland, where the Tonoloway limestone and the Wills Creek shale are comparatively unfossiliferous, except at a few horizons. Although twice as thick as the overlying Keyser, the Tonoloway contains less than half as many species. Furthermore, about 40 per cent of the known species are ostracodes, almost all new; the nonostracode fauna is composed of but 46 species, including 21 brachiopods, 6 Bryozoa, 5 gastropods, 4 pelecypods, 4 cephalopods, 2 corals, 1 stromatoporoid, 1 trilobite, 1 eurypterid, and 1 fish. Of these, the corals, Bryozoa, cephalopods, eurypterid, and fish are extremely rare, and only a few of the ostracodes, brachiopods, and gastropods are really abundant. The fauna of the underlying Wills Creek shale is still smaller in number of species and individuals.

The comparative impoverishment of the Cayugan faunas detracts somewhat from the force of the plea that the Keyser should be placed in the Helderberg group and thus at the base of the Devonian because of the greater specific agreement of its fauna with those

[29] Schuchert, Charles, Geol. Soc. America Bull., vol. 11, p. 275 and footnote, 1900

of the overlying than with those of the underlying deposits. If it were merely a question whether the Keyser should or should not be placed in the Helderberg group the apparent faunal relationships would be of greater significance. The injection of the question whether the Keyser should be considered Devonian or Silurian places greater emphasis upon the problem of age relationships, and the impoverished character of the Cayugan faunas leaves us with the doubt as to whether there might not be greater specific agreement between the fauna of the Keyser and those of the older deposits if only the older deposits carried more representative marine faunas.

The question, then, is not so much whether the Keyser should be placed in the Devonian because it belongs in the Helderberg group, as whether the Keyser should be placed in the Helderberg group because it is Devonian. We are thus not primarily concerned to know whether the fauna of the Keyser limestone contains a considerable number of Helderberg species; the question is, Does the Keyser fauna contain a marked number of those elements which have led stratigraphers to place the Helderberg of Clarke and Schuchert's definition in the Devonian? Great weight must therefore be accorded to the fact that members of the distinctive Devonian genera *Chonostrophia, Rensselaeria, Beachia, Meristella, Merista, Actinopteria,* and *Aviculopecten* make their first known appearance in the Keyser limestone, at least so far as the Appalachian province is concerned. The several species of *Uncinulus* of the type of *U. nucleolatus* probably also deserve a place in this category. The presence of these forms outweighs the evidence of the Silurian stragglers, of which *Calymene camerata* and *Whitfieldella prosseri* are the most abundant, and their testimony is upheld by the disappearance of the upper Silurian genera *Hindella* (*Greenfieldia*) and *Hormatoma,* both of which are well represented in the underlying Tonoloway limestone. I would thus concur in placing the Keyser limestone at the base of the Devonian and in including it in the Helderberg group, primarily because of the Devonian affinities of the Keyser fauna.

If the Keyser limestone is to be considered Devonian, as suggested, and the Silurian-Devonian boundary accordingly placed at the Keyser-Tonoloway contact, the nature of that contact becomes a matter of greater interest, particularly with respect to its time significance. So far as the physical aspect of the contact is concerned, the passage from the older formation into the younger seems to be transitional rather than abrupt, except south of Bolar, Va., where both the Tonoloway and the Keyser become much thinner. However, such evidence as is thus afforded is rather negative; and on the basis of the statement that the Keyser contains a large and varied fauna, of which only a comparatively few are derived from the Tonoloway or other known pre-Keyser faunas, it might be supposed that the Keyser is separated from the underlying Tonoloway by a hiatus of some magnitude.

We have already seen that the faunas of the Cayugan deposits of the Appalachian province seem to be considerably impoverished, and it has been suggested that the absence from the Cayugan deposits of a fauna with more direct affinities to that of the Keyser may be due more to the exclusion of such a fauna from the Appalachian Basin than to its actual nonexistence. It remains to be shown, of course, that such was the case.

A zone of fossils discovered in the middle of the Tonoloway at Monterey, McDowell, Fulks Run, and Bells Valley is of some interest in this connection. The most abundant species are *Atrypa reticularis,* which, although it has no precise bearing on the relationships of the fauna, does indicate marine connections; and an *Uncinulus,* which differs from the *U. convexorus* characteristic of the lower part of the Keyser only in its somewhat finer ribbing. One small specimen of *Stenochisma deckerensis* was found at the exposure of the zone at Monterey; although this specimen was loose, the supposition that it could have survived a trip down the hillside from the exposure of the lower Keyser *S. deckerensis* zone, some 250 feet higher up the slope, seemed very improbable. Another species occurring in this zone is a fairly large lamellose *Spirifer,* whose description will be reserved for another paper.

The above-described occurrence represents a short and restricted invasion of the area by a fauna with at least some definite affinities to that of the Keyser. It seems possible that, if the opportunities for invasion had been of longer duration and more widespread, the fauna might have been larger, and the affinities to the younger fauna more pronounced. There is thus no definite evidence of a marked time break between the Tonoloway and the Keyser limestones.

An aspect of the views above set forth that merits further comment is the diastrophic control of the apparent paleontologic relations of these formations in the sense that geographic and environmental conditions are ultimately controlled by diastrophic processes. Thus the Keyser of Maryland and the Virginias contains a large fauna, with a considerable number of species that continue into the overlying Coeymans, New Scotland, and Becraft limestones, and a lesser number derived from older deposits. The pronounced faunal change is found at the Tonoloway-Keyser contact in this area. As the Keyser is traced northward through central Pennsylvania the number of distinctly Helderberg species is diminished, and in the apparently equivalent Decker, Rondout, and so-called Manlius limestones of New Jersey and Cobleskill, Rondout, and so-called Manlius limestones of southeastern New York the fauna is much smaller, the species that continue into the overlying formations are only 6 in number, and the proportion of species derived from the underlying formations is greater in comparison to the total number of species in the fauna. Furthermore, the change in the fauna eliminates the members of the genera upon whose presence is here based the assignment of the Keyser to the Devonian. In the New Jersey and southeastern New York area the pronounced faunal change occurs at the base of the Coeymans limestone. The facts are, then, that a large and varied fauna, with a considerable number of Helderberg and Devonian species, was present in the Virginian and Maryland seas during Keyser time; the fauna present in the New Jersey and southeastern New York seas was smaller and practically devoid of Helderberg or Devonian species, although the seas of the two areas appear to have been connected, as shown by the fact that the Keyser sediments are continuous across Pennsylvania, with a more or less progressive change in lithology and fauna from the Maryland to the New Jersey development. The differences in fauna must then be accounted for by a lateral change in environmental conditions in a continuous sea, rather than by a barrier of some type, separating two seas. In other words, the oceanic connections of the Keyser sea must have been such that the marine fauna was able to penetrate in considerable numbers to Virginia and Maryland, while adverse environmental conditions, such as a change to more brackish water, excluded much of the fauna, and particularly its Devonian elements, from the New Jersey-New York area during this time; and it was not until Coeymans time that the progressive change in geographic relations permitted a representative marine fauna to invade the New York as well as the Maryland area. It is this difference in local development that has caused the difference in conclusions arrived at by the New York and New Jersey workers, on the one hand, and by those of Maryland and the adjoining States, on the other.

SUMMARY

In brief, then, as the Helderberg group is traced southward from Maryland, the formations recognized in that State are found to persist for a considerable distance with little change in thicknesses or faunas. Throughout the area covered by this study the Helderberg deposits are underlain by the Tonoloway limestone or limestones of Tonoloway lithology and fauna, while the younger Ridgeley sandstone persists as far south as Clifton Forge.

The most notable changes are as follows: In the Keyser limestone a shale unit, here named the Big Mountain shale member, is developed, separating the lower limestone member from the upper limestone member. The lower limestone member preserves, as far south as Clifton Forge, Va., the fauna and the essentially nodular character which are features of the lower part of the Keyser of Maryland. At Clifton Forge, however, this member has thinned considerably, its upper beds being replaced by sandstone, beginning in the sections west of Warm Springs and Hot Springs, Va. The Big Mountain shale member is maintained as far south as Bolar, Va., and then tongues out, giving place to sandstone. South of Petersburg, W. Va., the upper limestone member of the Keyser becomes purer and simulates the Coeymans limestone in lithology, and with this change in composition appear a number of forms that are not known beneath the Coeymans farther north. At Warm Springs a tongue of shaly sandstone appears near the base of the upper limestone member of the Keyser; as it is followed to Clifton Forge, this tongue of sandstone thickens, becomes more massive, and, with the sandstone beds replacing the Big Mountain shale and the upper part of the lower limestone members, forms what is here named the Clifton Forge sandstone member of the Keyser. This member is 66 feet thick at Clifton Forge, where the lower and upper limestone members of the Keyser are only 15 and 33 feet thick, respectively. The Keyser was not seen south of Clifton Forge. It is absent in the sections seen near Saltville and Big Stone Gap, in southwestern Virginia, and is probably missing in the intervening sections at Hollybrook and Rocky Gap.

The Coeymans remains a massive, highly crinoidal limestone, at least as far south as Clifton Forge. The "guide fossil" *Gypidula coeymanensis*, so abundant farther north, is very rare in the Clifton Forge area; one characteristic specimen was, however, seen in the Coeymans limestone at Hot Springs. Still farther south, at Rocky Gap and Hollybrook, the Coeymans may possibly be represented by a part of the sandstones that form the middle portion of the Giles formation of M. R. Campbell and lie between platy limestone of Tonoloway age and bedded chert of probable Oriskany age.

The New Scotland is a massive limestone with much interbedded white chert as far south as Monterey, Va. At Bolar, however, the cherty limestone is much thinner, and the lower part is replaced by a calcareous sandstone (in part an arenaceous limestone), here named the Healing Springs sandstone member, from the section west of Healing Springs, where the cherty limestone has disappeared. Part of the sandstone forming the middle of the Giles formation at Rocky Gap may be of New Scotland age. The New Scotland has been reported to occur as an arenaceous limestone in Wise County, supposedly with characteristic fossils. The exact relations of this supposed New Scotland are, however, debatable; its correlation presents a special problem.

The Shriver chert overlies the New Scotland in most of the sections studied but is replaced in the more easterly sections by the Becraft limestone with its characteristic fossils. The suggestion also arises that the Shriver chert is a muddy phase of the Becraft, as that formation is developed in Maryland and the Virginias.

The bedded chert forming the upper part of the Giles formation in the vicinity of Saltville, Va., simulates the Shriver chert in character and occupies an analogous stratigraphic position. Its post-Shriver age is shown by the presence in it of *Diaphorastoma ventricosum* and a variety of *Spirifer arenosus*. The similar bedded chert of the upper part of the Giles at Rocky Gap and Hollybrook is here correlated with that at Saltville, in spite of the absence of the fossils seen at the latter locality. This chert is possibly equivalent to the Harriman chert of western Tennessee.

The Silurian-Devonian boundary is placed at the base of the Keyser limestone chiefly because of the presence in the Keyser of members of important Devonian genera. That the species common to the faunas of the Keyser and the younger rocks are greater in number than those common to the faunas of the Keyser and the older rocks is thought to be due to the fact that the known upper Silurian deposits of the Appalachian area do not contain an entirely representative marine fauna, rather than to any considerable time break between the Tonoloway and the Keyser.

NOTES ON THE LOCAL FAUNAS OF THE HELDERBERG GROUP, WITH DESCRIPTIONS OF NEW FORMS

PORIFERA

Hindia sphaeroidalis Duncan.—Spherical; about 25 millimeters in diameter; skeleton of minute spicules uniting to form delicate radiating canals. Common in the New Scotland limestone.

ANTHOZOA

Streptelasma strictum Hall (pl. 8, figs. 13, 14).—Simple, regularly expanding, nearly straight, conical corallum; length 25 millimeters; diameter of calyx 16 millimeters; septa about 50; alternate septa short, frequently coalescing with the primary septa a short distance from the walls; some of the primary septa unite at the center to form a pseudocolumella; surface strongly ribbed. Abundant in the New Scotland limestone and basal part of the Becraft limestone.

Cyathophyllum radiculum Rominger.—Simple, conical, slightly curved corallite; length 14 millimeters; diameter of calyx as much as 8 millimeters; calyx fairly deep; septa low, denticulate, about 40. Common in the upper part of the Keyser limestone, Little Mountain, Va.

Favosites cf. *F. helderbergiae* Hall.—Corallum large, more or less hemispherical; base with wrinkled epitheca; corallites prismatic, intimately united, about 1.5 millimeters in diameter; 10 to 15 tabulae in space of 10 millimeters; mural pores in one or two ranges. The *Favosites* from the Coeymans and New Scotland limestones were generally referred to this species, but without critical study; those from the upper part of the Keyser limestone were mostly placed with the variety *praecedens* Schuchert.

Striatopora bella C. K. Swartz.—Corallum dendroid; stems 5 to 10 millimeters in diameter, consisting of closely united prismatic corallites, terminating obliquely at the surface, unequal in size and irregular in cross section; calyces funnel-shaped, as much as 2.5 millimeters in diameter; sides striated by 12 low ridges. Common in the upper part of the Keyser limestone, coral zone, Little Mountain, Va.

Cladopora rectilineata Simpson (pl. 7, figs. 9–11).—Corallum ramose, branching dichotomously; stems circular, about 2 millimeters in diameter; consisting of intimately united corallites, which ascend parallel to the axis and then turn outward and terminate obliquely at the surface; orifices small, arranged in 8 to 10 vertical rows, which are separated by low ridges. Abundant in the upper part of the Keyser limestone at many localities; rare in the lower part of the Keyser.

Aulopora schohariae Hall.—Corallum incrusting, composed of a loose network of elongate, thick-walled, nearly tubular corallites, which increase by basal or double laterobasal gemmation; corallites 5 to 7 millimeters in length; greatest diameter 1.5 millimeters. Abundant in upper part of Keyser limestone, coral zone, Little Mountain, Va.

Aulopora schucherti C. K. Swartz.—Corallum consisting of an intricately branched network of small nearly tubular corallites, irregularly fused so as to form a more compact corallum than in *A. schucherti*. Rare at the middle of the Keyser limestone.

ECHINODERMATA

Jaekelocystis hartleyi Schuchert.—Theca small, more or less pyriform; ambulacra narrow, prominent, extending nearly to the column; about eight folds in each pectinirhomb. Rare, lower part of Keyser limestone, Clifton Forge, Va.

Mariacrinus sp. (pl. 7, fig. 8).—Dorsal cup of medium size, subglobose; plates depressed convex, unornamented; basals unknown; radials rather large, hexagonal; primibrachs 2×5, the lower hexagonal, smaller than the radials; the upper pentagonal and axillary; secundibrachs 2×10, the upper axillary and bearing two tertibrachs each; tertibrachs 2×20, more or less pentagonal, each bearing an arm on the upper truncated edge; intersecundibrach one, heptagonal; interradials numerous, arranged in three irregular rows, with about four plates to the row above the level of the second primibrach; tegmen, stem, and arms unknown. Height of dorsal cup 26 millimeters; lateral diameter 23 millimeters. This description applies to one specimen from the upper part of the Keyser limestone at Petersburg, W. Va., which might well be considered a new species.

Edriocrinus pocilliformis Hall (pl. 8, fig. 15).—Cup very small, hemispheric or subturbinate, always detached; interior more or less deeply concave; when well preserved, the upper margin is scalloped with five large and one smaller depression; plates obscure; sutures generally obliterated. Very abundant at the base of the Becraft limestone, Clifton Forge area, Virginia.

Aspidocrinus caroli F. M. Swartz, n. sp. (pl. 9, figs. 10–12).—Body cup-shaped, fairly large, depressed subhemispherical, the thin edge flaring outward; interior broadly concave, with a sharper conical depression at the center; the apex of this depression reaching nearly to the attachment scar, which is about 6 millimeters in diameter and set in a depressed area; known only from the empty molds, which show no trace of plates and no points of attachment for arms. Like the other species referred to this genus, *A. caroli* probably represents the basal expansions of crinoid columns. Greatest diameter 38 millimeters; height 10 millimeters. The empty molds of *A. caroli* are very abundant in the upper portion of the sandstone forming the middle part of the Giles formation at Hollybrook and Rocky Gap, Va. Mr. Charles Butts, of the United States Geological Survey, who was so kind as to examine a specimen, identified it as an *Aspidocrinus* and suggested comparison with *A. scutelliformis* of the Becraft limestone of New York. Mr. Edwin Kirk agreed with the generic determination but did not believe it to be the species mentioned. As compared with *A. scutelliformis*, the Virginian material is distinguished by the shorter radius of curvature of the body, by the sharp conical depression at the center of the interior, and particularly by the peculiar outward flare of the margin. Also, the attachment scar is larger. *A. caroli* is named in honor of Dr. Charles K. Swartz, with whom I visited the sections at Hollybrook and to the south in connection with a study of the underlying Silurian.

BRYOZOA

Although Bryozoa are rather abundant at various horizons in the Helderberg group, particularly in the Keyser limestone, only the more obvious species have been identified.

Orthopora rhombifera (Hall).—Zoarium ramose; branches slender, elongate, 1 millimeter or less in diameter; zoecial apertures oval, set into rhomboidal or hexagonal, inward-sloping vestibular areas and arranged in diagonal intersecting lines. Abundant in Big Mountain shale member, 94 feet above base of Keyser limestone, Little Mountain, Va.

Cyphotrypa corrugata (Weller) (pl. 6, fig. 6).—Zoarium petasiform or hemispherical, as much as 50 millimeters in diameter; maculæ not elevated, consisting of large zoecia; zoecia tubular, thin-walled, polygonal; mesopores apparently wanting; acanthopores frequently developed at the angle of junction; diaphragms distant one or two times the diameter of the tubes. Although the lower Keyser material placed in this species has not been studied in thin section, the identification seems unquestionable. Common in the lower part of the Keyser limestone.

BRACHIOPODA

Orthostrophia strophomenoides (Hall).—Shell transverse, semioval; ventral valve with narrow mesial elevation from beak to front; corresponding narrow sinus confined to anterior half of dorsal valve; surface marked by coarse radiating striae, with concentric striae visible in depressions between the radiating striae. One specimen from the Coeymans limestone.

Dalmanella concinna (Hall).—Shell subcircular to longitudinally oval; hinge line straight; ventral valve the more convex and almost subcarinate, with rather prominent beak; surface with fine, even radiating striae. Length and width of Virginian material about 13 millimeters. Abundant at a few places in either the upper or the lower part of the Keyser limestone.

Dalmanella planiconvexa (Hall).—Shell of medium size, about 17 millimeters in length and 19 millimeters in width; transversely suboval; the dorsal valve almost flat, the ventral convex. Occurs sparingly in the New Scotland limestone.

Dalmanella perelegans (Hall) (pl. 8, figs. 10, 11).—Shell large, about 27 millimeters in length and 30 millimeters in width; transversely suboval; hinge line half the width of the shell or a little less; valves about equally convex; the dorsal valve with a shallow mesial sinus extending from the beak to the front margin; the ventral valve elevated along the middle, beak small, pointed and incurved. Surface with fine fasciculate radial striae. This species is a conspicuous element of the New Scotland fauna, to which it seems to be confined in the Virginias.

Dalmanella eminens (Hall).—Similar to *D. perelegans*, from which it differs in its somewhat more subquadrate outline and particularly in the much wider hinge line, which is about three-quarters as long as the width of the shell, and the greater height of the ventral area. Occurs sparingly in the New Scotland limestone.

Rhipidomella emarginata (Hall).—Shell small, subtrigonal; dorsal valve the more convex; the broad ventral sinus producing a marked emargination of the anterior margin. Length 15 millimeters; width 18 millimeters. Common in the lowest part of the Keyser limestone, to which it seems to be restricted.

Rhipidomella oblata (Hall).—Shell of medium size, transversely suboval to nearly circular; ventral valve depressed convex toward the beak, flattened to concave anteriorly, with a broad, shallow sinus; ventral beak extending only a little beyond the opposite; area very small; surface finely striated; ventral vascular impressions large and foliate. Length 20 millimeters; width 25 millimeters. This species is particularly characteristic of the top of the Coeymans limestone, where it is abundant at many places in the Monterey-Clifton Forge area.

Rhipidomella assimilis (Hall) (pl. 8, fig. 22).—Shell suborbicular, suggestive of *R. oblata* but differing in its larger size and relatively larger area and ventral beak. Length 27 millimeters; width 30 millimeters. Characteristic of the Becraft limestone, where it occurs more commonly in the upper portion. A few of the specimens from the upper part of the Becraft are larger, reaching 36 millimeters in length and 40 millimeters in width.

Leptaena rhomboidalis (Wilckens).—Shell semicircular, the hinge line slightly extended as a rule; dorsal valve nearly flat toward the umbo and sharply deflected near the margin; the flattened portion with strong concentric undulations; ventral valve deeply concave; radiating striae prominent. Common in all divisions of the Helderberg group.

Stropheodonta bipartita (Hall) (pl. 6, fig. 7).—Shell subsemielliptical; the hinge line produced into small mucronate extensions; dorsal valve nearly flat; ventral valve only slightly convex; surface with fine, angular striae, which are irregularly alternating in size, are not continuous to the beak, and curve outward on the sides of the shell in passing to the lateral margins. Length 28 millimeters; width 30 millimeters. This is a characteristic element in the fauna of the lower part of the Keyser limestone.

Stropheodonta cf. *S. planulata* (Hall).—Shell semielliptical, hinge line slightly extended; dorsal valve flat; ventral valve depressed convex; surface with extremely fine, rounded radiating striae, which are crossed by much finer, closely arranged concentric striae. Length 30 millimeters; width 40 millimeters. Several fragmental specimens from the middle part of the Keyser limestone seem to be most closely allied to this species. Also rare in the Coeymans limestone.

Stropheodonta arata (Hall).—Shell relatively small, semielliptical; dorsal valve more or less concave; ventral valve convex to gibbous. Surface marked by sharp angular plications and intermediate fine undulating striae. Length 15 millimeters; width 24 millimeters; five to six plications in 5 millimeters along anterior margin in the Virginian material. Rare in the Coeymans limestone.

Strophonella keyserensis C. K. Swartz.—Shell subsemicircular in outline; dorsal valve concave toward the umbo, convex toward the margins; ventral valve convex in the umbo, concave toward the margins; surface with numerous fine striae, about eight in 5 millimeters along the anterior margin; interior of valves pustulose. Length as much as 30 millimeters; width 37 millimeters. This forerunner of *S. punctulifera* is rare in the middle and upper parts of the Keyser limestone.

Strophonella punctulifera (Conrad).—Shell subsemicircular, somewhat contracted below the extremities of the hinge; ventral valve concave; dorsal valve concave near the umbo, very convex near the middle; hinge line straight; area narrow, linear; surface with strong sharp radiating striae, which are distinctly punctate on well-preserved shells. Length 35 millimeters; with 45 millimeters; about six plications in 5 millimeters. Occurs sparingly in the Coeymans limestone.

Strophonella leavenworthana (Hall).—Shell large, subsemicircular; both valves flattened except toward the margins, where the ventral valve is concave and the dorsal highly convex; surface with fine, obscure radiating striae, which are crossed on the flattened portions of the valve by small regular concentric wrinkles. Length 35 millimeters; width 45 millimeters. New Scotland and Becraft limestones.

Strophonella undaplicata C. K. Swartz.—Shell very large, subsemicircular, with the hinge line slightly extended; ventral valve convex anteriorly, umbo concave, and paralleled by dorsal valve; surface with coarse, peculiarly sinuous plications, which may unite, forming elliptical depressions between them. Length as much as 45 millimeters; width 52 millimeters. About four plications in 5 millimeters near anterior margin. Occurs sparingly in the New Scotland limestone.

Schuchertella deckerensis (Weller).—Shell transversely subelliptical; pedicle valve depressed, wrinkled and distorted by attachment in the umbonal region; brachial valve regularly convex; surface with rounded radiating striae, of which about 15 occupy 5 millimeters along the anterior margin. Length 20 millimeters; width 28 millimeters. Occurs sparingly in the lower part of the Keyser limestone, to which it is restricted.

Schuchertella sinuata (Hall and Clarke).—Shell transversely subelliptical; dorsal valve very convex, with a well-defined sinus extending from the strongly incurved beak to the anterior margin; surface with fine rounded radiating ribs, of which 10 occupy a space of 5 millimeters along the anterior margin. Length 27 millimeters; width 35 millimeters; these are the measurements of the Virginian specimens, which are larger than those from Maryland. Rare in the lower part of the Keyser limestone, to which it is restricted.

Schuchertella prolifica Schuchert.—Shell of medium size, semielliptical, flattened; hinge line straight; cardinal area narrow, linear; surface with low, rounded, fine radiating striae, of which there are 13 to 20 in 5 millimeters. Length 26 millimeters; width 35 millimeters. This species is closely allied to *S. woolworthana*, of the Coeymans limestone, to which it is introductory. It is common in the upper part of the Keyser limestone, and in the section at Petersburg, W. Va., it occurs very profusely in a zone just below the *Tentaculites* zone.

Schuchertella woolworthana (Hall).—Similar to *S. prolifica* but larger, slightly more convex, and with 7 to 12 radiating striae in 5 millimeters, instead of 13 to 20, as in that species. Length 35 millimeters; width 45 millimeters. This is an abundant element of the Coeymans, New Scotland, and Becraft faunas.

Schuchertella becraftensis (Clarke).—Shell small, suborbicular; pedicle valve erect at the beak; with triangular, broad, fairly high cardinal area; brachial valve depressed at the beak, becoming more convex toward the middle; surface with strong rounded radiating striae, increasing by implantation; also with exceedingly fine concentric striae. Length 13 millimeters; width 17 millimeters. One specimen from the Becraft limestone seems to belong to this species, which has been reported only from the Shriver chert and Ridgeley sandstone in Maryland.

Chonetes jerseyensis Weller (pl. 6, figs. 1, 2).—Shell subquadrate; brachial valve flat or slightly concave; pedicle valve depressed convex, beak small; cardinal area low, with about seven oblique marginal spines; surface with rather coarse radiating ribs; the lateral ribs commonly have a peculiar and characteristic anterior curvature. Length 14 millimeters; width 22 millimeters; about 8 to 10 ribs in 5 millimeters along anterior margin. Typical specimens of this species are common to abundant in the lower part of the Keyser limestone, to which they are restricted.

Gypidula coeymanensis Schuchert (pl. 6, figs. 25, 26; pl. 7, figs. 29, 30).—Shell ovoid to subglobose; ventral valve gibbous, becoming in old shells very ventricose about the umbonal region, beak swollen, arched, incurved; ventral valve convex, nearly circular. Surface generally with 15 to 20 well-developed plications, of which about five are raised on the anterior half of the ventral valve to form a distinct mesial elevation; the medial plications on both valves reach the beaks. Internally the ventral valve has a long spondylium and strong median septum, which stand out on weathered specimens. Length, 30 millimeters; width, 26 millimeters. This fossil is generally abundant in the Coeymans limestone, although it disappears from that formation in the Clifton Forge area. The Virginian material is considerably smaller than the dimensions given above and, as in the northern specimens, shows great variations in the plications and other features.

Gypidula coeymanensis var. *prognostica* Maynard (pl. 6, fig. 24).—Differing from *G. coeymanensis* proper in its smaller size, less inflated ventral beak, lack of distinct fold or sinus, and less distinct plications, which do not reach the beaks. Usual dimensions, length, 18 millimeters; width, 17 millimeters. Abundant in the middle of the Keyser limestone at Petersburg and Big Mountain, W. Va., also in Maryland and Pennsylvania.

Gypidula coeymanensis var. *similis* F. M. Swartz, n. var. (pl. 6, figs. 20–23).—I am proposing this name for the Gypidulas occurring in the middle of the Keyser limestone in the vicinity of Monterey and Warm Springs, Va. These specimens are comparable to the largest Coeymans material in size, some individuals measuring as much as 45 millimeters in length and 33 millimeters in width. The chief differences lie in the slightly less swollen ventral beak and the somewhat broader, less elevated plications, of which there are only about 12. However, about four of the medial plications of the ventral valve are raised into a low though fairly distinct elevation toward the front of the valve, and these plications reach the beak, as in *G. coeymanensis* proper. Internally, the ventral spondylium and median septum are quite comparable to those of the typical material in size and strength. As has been noted, the Coeymans material is itself rather variable, and the middle Keyser Gypidulas here noted are almost identical with the pauciplicate specimens found in that formation. The material placed in the variety *similis* should, however, be considered most closely akin to that included in the variety *prognostica* in the more northerly sections, which evidently occurs at exactly the same horizon, as is shown by the position of the overlying *Camarotoechia gigantea* and *Merista typa* zones.

Rhynchotrema cumberlandicum Rowe (pl. 8, figs. 23–25).—Shell trigonal to subpentagonal; somewhat wider than long; ventral valve depressed convex, beak nearly erect, rather prominent; sinus shallow, with three plications; dorsal valve moderately convex, with four plications on the depressed fold; surface with four or five sharply elevated, angulated plications on each side of the fold and sinus. Length, 18 millimeters; width, 19 millimeters. Rare at top of Becraft limestone, Gala, Va.

Stenochisma deckerensis (Weller) (pl. 6, figs. 10–12).—Shell transversely subovate; ventral valve less convex than the dorsal, its beak prominent, arched but not strongly incurved; sinus and elevation abrupt near the front and reaching the beaks; surface of each valve with 20 to 24 simple angular plications, of which three rather coarser than the rest lie in the ventral sinus and four on the dorsal fold. Length, 15 millimeters; width, 19 millimeters; thickness, 10 millimeters. Abundant in the lower part of the Keyser limestone, of which this species is diagnostic.

Camarotoechia litchfieldensis (Schuchert) (pl. 6, figs. 3–5).—Shell subtrigonal, very small; ventral sinus shallow; dorsal fold low; surface with 16 to 22 simple angulated plications, of which three occupy the sinus and four the fold. Length 9 millimeters; width 9.5 millimeters; thickness 5 millimeters. Common in the basal Keyser *Whitfieldella minuta* subzone; locally abundant at higher horizons of the Keyser.

Camarotoechia gigantea Maynard (pl. 6, figs. 27, 28).—Shell subtrigonal, rather large; ventral beak closely incurved over that of the dorsal valve; ventral sinus broad in front, not reaching the beak; dorsal valve much the more convex; surface with 18 to 20 simple, prominent rounded plications on each valve, with five or six in the ventral sinus and a corresponding number on the low dorsal fold. Length 25 millimeters; width 30 millimeters. Profuse at the middle of the Keyser limestone, just above the *Gypidula* zone.

Camarotoechia gigantea var. *gigas* F. M. Swartz, n. var. (pl. 6, figs. 29–31).—Similar in general character and proportions to *C. gigantea* proper but nearly twice as large and with 26 instead of 20 plications to the valve; seven of the plications, instead of five, are depressed toward the anterior margin of the ventral valve to form a broad sinus, which is rather sharply deflected. Length 40 millimeters; width 45 millimeters. Occurs in the *C. gigantea* zone, Monterey, Va.

Camarotoechia cf. *C. altiplicata* (Hall) (pl. 7, figs. 26–28).—Shell subtrigonal, rather small; ventral valve depressed convex, beak pointed, somewhat arched; dorsal valve the more convex, surface with 10 to 19, usually 14 simple angular plications, with three or less depressed on the middle of the ventral valve into a distinct sinus, with four or less on the dorsal elevation. Length 12 millimeters; width 14 millimeters. The material here referred to *C. altiplicata* is common in the upper part of the Keyser limestone south of Big Mountain, W. Va. It is somewhat smaller than the typical New Scotland material but otherwise similar.

Camarotoechia praespeciosa Schuchert.—Shell longitudinally ovoid, vertically flattened on the sides; ventral valve much the less convex; beak incurved; sinus shallow, broad, undefined; surface with 32 to 36 strongly elevated, subangular plications on each valve. Length 30 millimeters; width 26 millimeters. This species is introductory to *C. speciosa* of the Oriskany group. Occurs sparingly in the upper part of the Becraft limestone at Clifton Forge, Va.

Camarotoechia campbellana (Hall).—Shell longitudinally ovate; ventral valve depressed convex, beak small, incurved, sharply deflected at the lateral margins, depressed into a broad rounded sinus, which curves upward toward the other valve, forming a linguiform extension; dorsal valve much more convex, elevated into a broad fold near the front margin; surface with 22 to 24 plications to each valve, with about five or six on the sinus and fold. Length 25 millimeters; width 23 millimeters. Occurs sparingly at the top of the Coeymans limestone in the Clifton Forge area, Virginia.

Uncinulus convexorus Maynard (pl. 6, figs. 8, 9).—Shell trigonal to subpentagonal; ventral beak prominent, pointed, somewhat incurved; dorsal valve the more convex; no distinct fold or sinus; surface of each valve with 30 or more low simple rounded plications. Length 18 millimeters; width 16 millimeters. Abundant in the lower part of the Keyser limestone, of which it is characteristic.

Uncinulus gordoni Maynard.—Shell subpentagonal, small; ventral beak fairly prominent; dorsal valve much the more convex; surface of each valve with 19 to 26 subangular plications, three of which are depressed in the shallow ventral sinus and four are raised anteriorly to form a low dorsal prominence. Length 10 millimeters; width 9 millimeters. Several specimens were found in the upper part of the Keyser limestone.

Uncinulus nucleolatus (Hall).—Shell subpentagonal; ventral valve depressed convex; beak small, closely incurved; dorsal valve often gibbous. Surface with 15 to 20, usually 18 simple rounded plications; three to five depressed at the front of the ventral valve into a shallow sinus, which is continued into a short linguiform extension; a corresponding prominence on the dorsal valve. Length 14 millimeters; width 13 millimeters. Small specimens of this New Scotland form occur sparingly in the upper part of the Keyser limestone.

Uncinulus angulatus Maynard.—Shell subpentagonal; both valves depressed convex; surface with strong angular plications, of which there are said to be 23 in Maynard's description, although not more than 14 are indicated on the figured specimens; three plications depressed toward the front in a shallow ventral depression; a corresponding number on the dorsal prominence. Length 12 millimeters; width 9 millimeters. Several specimens in the upper part of the Keyser limestone, with 14 plications to the valve. Does not seem closely allied to *U. nucleolatus*, although it was originally described as a variety of that form.

Uncinulus abruptus (Hall).—Shell transversely oval to subpentagonal; ventral beak small, closely incurved; dorsal valve much the more convex. Surface with 25 to 33 simple subangular plications on each valve; six to eight depressed into a broad undefined sinus, and seven or eight elevated into a rather distinct dorsal prominence. Length 20 millimeters; width 24 millimeters. Common at the top of the Coeymans limestone and the base of the Becraft limestone in the Clifton Forge area, Virginia; also in New Scotland limestone.

Uncinulus pyramidatus (Hall).—Shell pyramidal, dorsal valve very gibbous, highest toward the front and declining toward the beak; cardinal slopes abruptly vertical, with a distinctly impressed suboval space beneath the beaks; surface with 13 to 22 simple strong subangular plications, four to six of which are somewhat elevated toward the front of the dorsal valve, forming a mesial prominence. Length 24 millimeters; width 23 millimeters; gibbosity of dorsal valve as much as 17 millimeters. One dorsal valve from the upper Becraft of Back Creek Mountain, Va., although suggestive in many ways of the *U. ventricosa* of the New York Becraft, agrees with *U. pyramidatus* in the presence of six plications on the mesial prominence and in that it is highest toward the front and flattened at the middle, rather than being nearly circular in vertical section.

Uncinulus vellicatus (Hall).—Shell subtrigonal to subovate; dorsal valve the more convex, gibbous; surface with 24 to 36 (usually 30) simple rounded plications, of which five to seven are depressed into a shallow indistinct sinus toward the front of the ventral valve and prolonged into an upward inflected lingual extension, while six to eight are slightly elevated at the front of the dorsal valve. Length 20 millimeters; width 22 millimeters. Occurs sparingly in the upper part of the Becraft limestone at Gala, Clifton Forge, and Healing Springs, Va. The specimens here included in *U. vellicatus* are somewhat narrower than is typical.

Eatonia medialis (Vanuxem) (pl. 8, fig. 12).—Shell transversely subquadrate, fairly large; hinge line very obtuse; ventral beak small, closely incurved; dorsal valve much the more convex; surface of each valve with 12 to 16 broad rounded plications, three of which are depressed to form the broad ventral sinus, with four on the mesial fold of the dorsal valve. Length 25 millimeters; width 28 millimeters. Abundant in the New Scotland limestone.

Eatonia peculiaris (Conrad) (pl. 8, figs. 16–18).—Shell longitudinally ovate, cardinal margins sloping rather abruptly from the beaks; ventral valve flattened, abruptly inflected along the cardinal slopes; toward the front, depressed into a broad rounded sinus, continued into a deflected linguiform extension; dorsal valve depressed convex toward the beak, rising into a high rounded mesial prominence toward the front; surface marked by fine radiating striae, with a stronger elevated one along the center of the mesial sinus. Length 20 millimeters; width 18 millimeters. Occurring throughout the Becraft limestone, but more abundant in the lower portion.

Rensselaeria mutabilis (Hall) (pl. 7, figs. 21–25).—Shell longitudinally subpentagonal, with a peculiar and characteristic outline; valves depressed convex, generally compressed toward the front border; ventral beak arched, pointed; surface with as many as 28 obscure radiating striae, which are usually obsolete in the upper part of the shell. Length 10 millimeters; width 9 millimeters. Locally abundant in the upper part of the Keyser limestone.

Rensselaeria cf. *R. aequiradiata* (Conrad) (pl. 8, figs. 19–21).—Shell subglobular, of medium size; dorsal valve somewhat the more convex; surface with 50 to 60 simple rounded fine plications, which become obsolete before reaching the beak. Length 18 millimeters; width 16 millimeters; thickness 12 millimeters. An abundance of fragmentary material, chiefly weathered posterior halves of brachial valves, found at the base of the Becraft limestone in the Clifton Forge area, Virginia, agrees with this New York species in size and shape. The lateral and anterior margins were not seen, so the number and strength of the plications are unknown, and the determination is uncertain. Not *R. subglobosa.*

Rensselaeria subglobosa (Weller).—Shell subglobular, length a little greater than the breadth; pedicle valve a little the more convex; beak small, closely incurved; surface marked by 50 to 60 simple rounded plications, becoming obsolete toward the beak. Length 18 millimeters; width 16 millimeters; thickness 11 millimeters. Occurs sparingly in beds representing the upper part of the Becraft limestone on Back Creek Mountain, west of Warm Springs, Va.

Atrypa reticularis (Linné).—Shell subrotund; dorsal valve much the more convex, gibbous; surface with fine dichotomous rounded radiating striae, of which the Helderberg material carries about six in 5 millimeters along the anterior border. Length 26 millimeters; width 24 millimeters; thickness of adult 20 millimeters. Abundant in the lower and middle parts of the Keyser limestone; otherwise rare in the Helderberg group.

Spirifer modestus Hall (pl. 6, figs. 16, 17).—Shell small, transversely subellipsoidal; extremities rounded; ventral beak prominent, acutely pointed, incurved; sinus and elevation faint, undefined; distinct plications absent on the lateral slopes; surface with very faint concentric striae. Length 10 millimeters; width 12 millimeters; thickness 8 millimeters. Very abundant in the lower part of the Keyser limestone in many of the sections.

Spirifer modestus var. *plicatus* Maynard (pl. 6, figs. 18, 19).—Very similar to *S. modestus* proper but differing in its slightly larger size and in the presence of three or four indistinct plications on the lateral slope. Length 15 millimeters; width 19 millimeters. As this form is strictly identical in stratigraphic and geographic range with *S. modestus* proper, it should be considered consanguineous with that form.

Spirifer octocostatus Hall.—Shell transversely suboval; central beak moderately elevated, incurved; sinus subangular; surface with four or five rounded plications to the lateral slope; these are crossed by very fine, closely arranged, sublamellose striae. Length 20 millimeters; width 25 millimeters; thickness 13 millimeters. Occurs sparingly in the lower part of the Keyser limestone.

Spirifer vanuxemi var. *prognosticus* Schuchert.—Shell small; subsemicircular; moderately gibbous; extremities rounded; surface with five or six plications on the lateral slope, instead of only three or four, as in *S. vanuxemi* proper. Length 8 millimeters; width 10 millimeters. Profuse at the top of the Keyser limestone at Franklin, W. Va.

Spirifer perlamellosus Hall. —Shell of medium size, more or less extended on the hinge line; ventral beak prominent, incurved at apex; sinus and fold strongly developed; surface with three to six elevated plications on each lateral slope; the plications are crossed by strong imbricating lamellae, which are abruptly arched in passing over the plications. Length 17 millimeters; width 30 millimeters. Common in the New Scotland limestone; rare in the Coeymans limestone.

Spirifer perlamellosus var. *praenuntius* F. M. Swartz, n. var. (pl. 7, figs. 18–20).—Seven specimens found in the upper part of the Keyser limestone agree very closely with *S. perlamellosus,* except in their somewhat smaller size. There are generally three or four plications on the lateral slope. Although the plications are not less numerous than in some of the New Scotland material, and although there is close agreement in the strength of the plications and of the imbricating lamellae, it seems better to separate these earlier and somewhat smaller forms as a distinct variety. Length 12 millimeters; width 16 millimeters. Occurs sparingly in the upper part of the Keyser, from Big Mountain, W. Va., to Monterey, Va.

Spirifer cyclopterus Hall.—Shell semicircular, extremities rounded; ventral valve the more convex, often gibbous; ventral beak moderately elevated, incurved; sinus of medium depth, flat-bottomed; dorsal fold strongly elevated; surface with five to seven strong rounded plications on each lateral slope, and with fine, close concentric striae. Length 17 millimeters; width 26 millimeters. Common in the Coeymans, New Scotland, and Becraft limestones.

Spirifer macropleurus (Conrad) (pl. 8, fig. 9).—Shell very large, transversely semielliptical; valves almost equally convex; extremities blunt; generally with four broad rounded plication on each side of the strong rounded sinus and elevation; entire surface with distinct closely arranged radiating striae. Length 40 millimeters; width 65 millimeters. Profuse in the New Scotland limestone as far south as Warm Springs, Va.

Spirifer concinnus Hall (pl. 9, fig. 1).—Shell semicircular, of medium size; extremities rounded; ventral beak somewhat elevated, abruptly incurved at apex; sinus wide, subangular; dorsal fold obtusely angular; surface with 12 to 14 rounded simple plications on each lateral slope; best specimens show radial lines of minute interrupted granules. Length 21 millimeters; width 24 millimeters. Common throughout the Becraft limestone in the Clifton Forge area, Virginia.

Spirifer concinnus var. *progradius* F. M. Swartz, n. var. (pl. 9, figs. 2–5).—Shell transversely subsemielliptical, cardinal angles more or less rounded; ventral valve most convex toward the beak; beak somewhat elevated, apex incurved; cardinal area of medium size, extending to the extremities of the hinge line; mesial sinus fairly broad, rounded, of moderate depth; surface with about 16 simple, rounded, little elevated plications on each lateral slope; the plications become obsolete in the region of the cardinal extremities, no trace of any plications being seen on these areas in the specimens at hand; dorsal valve not observed. Length 23 millimeters; width 40 millimeters. The general aspect of this species is that of *S. concinnus,* of which it is evidently a derivative. It differs from that form in its larger size and more transverse outline and particularly in its somewhat more numerous, relatively narrower plications. The finer surface detail is not preserved on the specimens seen, but is most likely that of *S. concinnus.* Occurs sparingly at the top of the Becraft limestone in the Clifton Forge area, Virginia.

A number of specimens occurring with the type material of *S. concinnus* var. *progradius* at the top of the Becraft limestone present the same expression but differ in the subangulation of the ventral mesial sinus, and also in the development of a faintly suggested plication on each side of the sinus, in the anterior half of the valve. The latter feature is seen in some of the Spirifers of the New York Becraft, which Hall included in *S. concinnus* but which Schuchert has separated as *S. proavitus.* However, as the Virginian specimens agree with *S. concinnus* var. *progradius* rather than with *S. proavitus* in size, in the width of the plications, and in the obsolescence of the plications toward the cardinal extremities, they are retained with the former species.

Spirifer angularis Schuchert (pl. 9, fig. 6.)—Shell transversely semielliptical; ventral beak moderately elevated, somewhat incurved; area of medium height; mesial sinus deep, angular; dorsal fold sharply elevated, angular. Surface with seven to nine elevated subangular plications to the lateral slope. Length 20 millimeters; width 35 millimeters. One specimen from the top of the Becraft limestone at Gala, Va., agrees closely with Schuchert's description of the type material from the Ridgeley sandstone of Maryland.

Spirifer arenosus var. *planicostatus* F. M. Swartz, n. var. (pl. 9, figs. 13–15).—Shell very large, semielliptical to subquadrate, extremities rounded; valves almost equally convex; ventral beak moderately elevated, incurved at the apex; area fairly high, extending to the extremities of the hinge line; mesial sinus of ventral valve gradually expanding, fairly wide, rather shallow, rounded; mesial elevation of the dorsal valve pronounced, broad, rounded, not sharply defined; surface with about 15 broad, low, very flat plications to the lateral slope; the intervening sulci are very narrow, angular, and of but little depth; the mesial fold and sinus are also plicate, carrying five or six plications similar to those of the lateral slopes; the plications are less numerous in the posterior portions of the fold and sinus, addition resulting from dichotomous division, as in *S. arenosus* proper. Maximum length 54 millimeters; maximum width 80 millimeters. This form differs from the *S. arenosus* of the Ridgeley sandstone farther north chiefly in the very flat plications; the plications of *S. arenosus* are rounded or even subangular. The cardinal angles also seem to be somewhat more definitely rounded and somewhat flatter. Abundant in the dark cherts and interbedded sandstones in the upper part of the Giles formation on Tumbling Creek, near Saltville, Va.

Cyrtina dalmani (Hall) (pl. 7, figs. 4–7).—Shell small, transverse; ventral valve pyramidal; beak simple, angular, not incurved; area high, flat or nearly so; foramen narrow. Dorsal valve depressed convex, beak scarcely elevated. Surface with five to seven plications to the lateral slope, crossed by strong imbricating concentric lamellae. Length 10 millimeters; width 14 millimeters; ventral area 5 to 7 millimeters in height. Rare in the middle of the Keyser limestone at Big Mountain, W. Va., and Monterey, Va.

Cyrtina varia Clarke.—Shell of moderate size; ventral valve pyramidal; beak angular, not incurved; area high, nearly flat; foramen narrow; dorsal valve depressed convex, beak scarcely elevated; surface with 7 to 10 plications on the lateral slope, crossed by concentric striae, which are not prominent on the specimens seen. Length 9 to 15 millimeters; width 18 to 23 millimeters; area 8 to 12 millimeters in height, forming an angle of 80° to 85° to the plane between the two valves. Occurs sparingly in beds representing the upper part of the Becraft on Back Creek Mountain, west of Warm Springs, Va.

Rhynchospira globosa (Hall).—Shell small, globose; ventral valve more convex than the dorsal; beak prominent, arched, perforated at the extremity by a round aperture; surface of each valve with 12 to 16 somewhat angular plications, of which two or three are slightly depressed along the middle of each valve. Length 11 millimeters; width 11 millimeters. Occurs sparingly in the lower part of the Keyser limestone.

Rhynchospira formosa (Hall).—Longitudinally subovate; ventral beak arched, tapering; dorsal valve the more gibbous; surface of each valve with 18 to 22 simple rounded plications, of which two or three are smaller and are slightly depressed along the median line of each valve. Rare in the lower part of the Keyser limestone.

Trematospira equistriata Hall and Clarke.—Shell transversely oval, of medium size; valves about equally convex; ventral beak only slightly extended beyond the dorsal valve, incurved; ventral sinus small; dorsal fold obsolete; surface with fine, simple radial plications, of which there are six plications on the sinus and 23 on each side. Occurs sparingly in the New Scotland limestone, to which it is restricted.

Nucleospira swartzi Maynard (pl. 6, figs. 13–15).—Shell subcircular, slightly transverse, hinge line shorter than greatest width of the shell; ventral valve somewhat the more convex, beak somewhat elevated, strongly incurved over that of the dorsal valve; valves without distinct mesial depressions, although these may be very faintly developed in some specimens; surface with very faint concentric growth lines and some stronger growth varices. Length 15 millimeters; width 16 millimeters. Abundant in the lower part of the Keyser limestone, *Stenochisma deckerensis* subzone.

Nucleospira ventricosa (Hall).—Shell globose, valves almost equally convex, very small; ventral valve with a narrow sinus from beak to front along medial line, while the median depressed line of the dorsal valve is nearly obsolete; ventral beak the more prominent, pointed, strongly incurved. Surface with some concentric growth lines. Length 8 millimeters; width 9 millimeters. Occurs in the Coeymans limestone and is abundant at the top of the Keyser limestone at Monterey and McDowell, Va.

Whitfieldella minuta Maynard (pl. 6, figs. 3, 4).—Shell minute, longitudinally subovate; both valves strongly convex; ventral beak prominent, pointed, incurved; no fold or sinus on either valve; surface smooth or marked by obscure concentric growth lines. Length 3 millimeters; width 2 millimeters. Abundant at the base of the Keyser limestone, to which it is restricted.

Whitfieldella? prosseri Grabau.—Shell elongate, oval to subpentagonal; ventral valve somewhat the more convex, gibbous; beak prominent, strongly incurved, with a faint longitudinal depression, enlarging somewhat toward the front and truncating the anterior margin; surface smooth except for some faint concentric striae and a few stronger growth lamellae, developed near the front. Length 20 millimeters; width 15 millimeters. Abundant at the top of the Keyser limestone at Franklin, Va.; rarer at the same horizon at McDowell, Va.

Meristella praenuntia Schuchert (pl. 7, figs. 12–15).—Shell ovoid, variable as to relative length and width; ventral valve much the more convex, gibbous in the unbonal region; beak strongly incurved, a mesial sinus more or less developed in the anterior part of the valve; dorsal valve often with a linear median depression toward the front; surface smooth. Length 19 millimeters; width 18 millimeters. Common to abundant in the upper part of the Keyser limestone. A forerunner of *M. arcuata* of the Coeymans and New Scotland limestones.

Meristella nasutaformis F. M. Swartz, n. sp. (pl. 7, figs. 16, 17).—Shell longitudinally ovoid; fairly large; greatest width a little front of the middle. Ventral valve convex to gibbous, arched toward the beak, which is closely incurved over that of the opposite valve; the anterior margin prolonged into a gently raised, very broad nasute extension, marked centrally by a narrow linear depression; this raised portion is continued to the middle of the valve, where it merges into the general convexity of the valve. The brachial valve in this specimen is crushed and broken, but the small incurved beak, fitting closely beneath the ventral beak, and interior median septum of the posterior half are shown. Surface with obscure lines of growth. Length 28 millimeters; width 25 millimeters. The one specimen came from about the middle of the Keyser limestone at Big Mountain, W. Va. It is sharply differentiated from the other Meristellas of the Keyser by its large size and from those of the Helderberg group in general by the nasute extension. *M. nasuta*, of the Schoharie grit of New York, also possesses a ventral nasute extension, but it is narrower and lacks the median depression seen on the Keyser specimen. *M. nasutaformis* should be regarded as a development from those members of *M. praenuntia* whose ventral valves have fairly strong plications on each side of a narrow median depression, such as is suggested by the larger of the two specimens figured in this report. It does not seem to have been ancestral to any of the later Meristellas described from the Appalachian areas.

Meristella arcuata (Hall) (pl. 8, figs. 1–4).—Shell large, broadly ovate; ventral valve gibbous in the central and umbonal regions; beak incurved, fairly prominent; a shallow rounded depression at the front of this valve; dorsal valve often abruptly elevated along the middle and sloping laterally but without a distinct fold; surface smooth. Length 20 millimeters; width 29 millimeters. Common to abundant in the Coeymans and New Scotland limestones; very abundant at the top of the Coeymans limestone in the Monterey and Clifton Forge area, Virginia; also seen just beneath the sandstones of the middle of the Giles formation at Rocky Gap, Va.

Meristella arcuata var. *gigas* F. M. Swartz, n. var. (pl. 8, figs. 5-7).—Associated with the normal *M. arcuata* of the top of the Coeymans limestone at Monterey and Bolar, Va., are some individuals reaching in maturity 40 millimeters in length. These are longitudinally oval, with the greatest width just in front of the middle. The ventral valve is highly convex and arcuate in the umbonal region, the beak is rather prominent and incurved, and there is a rather narrow, shallow sinus extending from the anterior border about halfway to the beak. The growth lines show that in the more youthful stages this form exhibited all the features of the typical *M. arcuata*, with the breadth about equal to the width. Surface with a few distinct growth lines. The interior features are those of *M. arcuata*. Length 40 millimeters; greatest width 35 millimeters.

Meristella cf. *M. symmetrica* Schuchert (pl. 8, fig. 8).—Shell almost circular; valves regularly and almost equally convex; no fold or sinus on either valve; surface with some concentric growth lines. Length 50 millimeters; width 50 millimeters. One specimen, with the dimensions stated, agrees with Schuchert's description of this Oriskany species. At the top of the Coeymans limestone at Monterey, Va.

Meristella lata (Hall) (pl. 9, figs. 8, 9).—Shell subquadrate to longitudinally ovate; ventral valve gibbous in the middle, with a shallow depression in front; beak elevated and closely incurved over the opposite; dorsal valve abruptly elevated in the middle, forming a rather prominent dorsal ridge. Length 30 millimeters; width 28 millimeters. Occurs sparingly in the Becraft limestone.

Merista typa (Hall) (pl. 7, figs. 1-3).—Shell longitudinally subovate; subglobose; ventral valve much the larger. The interior of the ventral valve has a more or less highly arched transverse septum, which, rising from beneath the rostral cavity, extends forward nearly to the middle of the valve. This "shoe-lifter process" is commonly seen on the weathered interiors and is very distinctive. Length 25 millimeters; width 21 millimeters. Profuse at about the middle of the Keyser limestone, just above the *Gypidula* zone, forming a distinct subzone, to which it is restricted.

PELECYPODA

Actinopteria cf. *A. reticulata* Weller.—Left valve large, subrhomboidal; body subovate, with an obliquity of 27° between the hinge line and the umbonal region; surface with conspicuous concentric lines of growth, and more or less dicontinuous radiating costae, giving the upper portion of the shell, where the lines are strongest, a nodose appearance. Height 35 millimeters; length of hinge line 29 millimeters; oblique diameter from beak to posterolateral extremity 48 millimeters. One fragmental specimen, found in the upper part of the Keyser limestone at Little Mountain, Va.

Cypricardinia cf. *C. lamellosa* Hall.—Shell subovoid; gibbous; umbones slightly elevated; posterior slope prominent, with a scarcely defined ridge; surface with strong elevated distant lamellae. Length 12 millimeters; height 8 millimeters. One specimen from the middle of the Keyser limestone at Monterey, Va.

GASTROPODA

Straparollus welleri F. M. Swartz, n. name (cf. *Straparollus* sp. Weller, Paleontology of New Jersey, vol. 3, p. 246, pl. 22, fig. 14, 1903).—Shell consisting of four or five volutions, very gradually expanding; shell plane above, with a broad umbilicus below, in which all the volutions are visible; surface features uncertain, although the shell seems to have been essentially smooth. Diameter of shell 25 millimeters; of body whorl 5 millimeters. One incomplete specimen from the lower part of the Keyser limestone on Big Mountain, W. Va.

Platyceras gibbosum Hall.—Spire small, of one or two closely coiled volutions; body volution very rapidly expanding, highly ventricose below, free; surface marked by fine, undulating, transverse striae. Diameter of shell 35 millimeters; of aperture 28 millimeters. One specimen from the Coeymans limestone at Monterey, Va.

Platyceras multiplicatum F. M. Swartz, n. sp. (pl. 7, figs. 31, 32).—Shell large, obliquely subovate, composed of three or four volutions, somewhat gradually expanding, with the last volution free near the aperture; flattened on the upper side, somewhat ventricose below; spire closely coiled, rising slightly above the plane of the body whorl; aperture unknown; surface with eight or more strong longitudinal plications, which are crossed by fine, obscure, closely arranged transverse striae. Diameter of shell 60 millimeters; of whorl near aperture 35 to 40 millimeters. This species is somewhat similar to *P. multisinuatum*, of the New Scotland limestone of Maryland and New York, but differs from that species in its larger size and in the more numerous longitudinal plications, which are continued farther toward the apex and occur on the lower as well as the upper and outer sides of the whorls. Common in the Coeymans limestone at Bells Valley and McDowell, Va. One fragment from the sandstones of the middle part of the Giles formation at Hollybrook, Va., is referred with some doubt to this species.

Platyceras gebhardi Conrad.—Shell of four gradually expanding volutions, which are contiguous, except the last one near the aperture; apex in the plane of the outer whorl; body whorl becoming ventricose; aperture expanded; surface with fine transverse undulating striae; shell occasionally carinate along the dorsolateral curvature. Diameter 40 millimeters. Common in the New Scotland limestone; rare in the Becraft limestone.

Tentaculites gyracanthus (Eaton).—Shell elongate conical, gradually tapering to the apex; surface with smooth rounded annulations, at irregular intervals, with about three in 1 millimeter; interspaces between annulations with fine annual striae. Length 5.5 millimeters; diameter 0.7 millimeter. Profuse near the top of the Keyser limestone at Petersburg, W. Va.; this subzone, which is prominent in Maryland and Pennsylvania, disappears south of Petersburg.

CEPHALOPODA

Orthoceras cf. *O. rigidum* Hall.—Shell of medium size, elongate, gradually tapering; cross section circular; siphuncle central; septa closely spaced, moderately convex, about six in a space equal to the diameter of the tube; surface with fine transverse striae; apical angle 10°. The specimens found are small, with diameter toward the aperture as great as 20 millimeters. Occurs sparingly near the base of the Keyser limestone.

TRILOBITA

Proetus protuberans Hall.—Pygidium semicircular; axis very prominent, nearly one-third the width of the pygidium, with eight annulations; lateral lobes with four or five segments, which are grooved and do not reach the margin, leaving a fairly broad border. Length 10 millimeters; width 18 millimeters. Several pygidia from the lower part of the Keyser limestone agree with Hall's description of this New Scotland trilobite. Reeside has also described *P. protuberans* from the lower Keyser in Pennsylvania.

Calymene camerata Conrad.—Pygidium with prominent axis, which is about one-fifth the width of the pygidium and carries eight annulations; the axis tapers gradually to the obtusely rounded extremity lying well within the posterior margin; lateral lobes with six broad ribs, which are strongly bent downward and are bifurcated toward the margins by deep longitudinal grooves. Length 14 millimeters; width 20 millimeters. One pygidium found in the basal part of the Keyser limestone at Clifton Forge, Va.

Phacops logani Hall.—Cephalon semicircular; genal angles obtuse; glabella subpentagonal, very prominent in front, projecting somewhat beyond the frontal limb; upper surface convex to gibbous, highly pustulose; lateral furrows undefined; occipital furrow strong; eyes fairly large. Occurs sparingly at the base of the Becraft limestone at Gala and Bells Valley, Va.

Dalmanites pleuroptyx (Green).—Pygidium triangular, convex, posterior extremity extended in a caudal spine; axis prominent, well defined, one-fifth the width of the pygidium in front, gradually tapering to the rounded extremity; annulations 17; lateral lobes with 11 to 13 ribs, which are bent backward. Length 45 millimeters; width 60 millimeters. One specimen from the Coeymans limestone at Petersburg, W. Va.; a second from the middle of the Becraft limestone at Dry Run, Va. Two specimens obtained above the middle of the Shriver (?) chert at Back Creek Mountain, Va., seem to belong to this species.

Dalmanites sp.—Fragments of several pygidia in beds representing the upper part of the Becraft limestone on Back Creek Mountain, west of Warm Springs, Va., show agreement with the pygidia of *Dalmanites dentatus* Barrett in the general shape of the shield, in the strong grooving of the pleurae, in the lack of a conspicuous marginal border, apparently in the width of the axis, and in the presence of a row of pronounced tubercles on either side of the groove of the pleurae. These tubercles are, however, smaller and more numerous, there being as many as 10 in a row, rather than 5, the average number seen on specimens of *D. dentatus* from the lower part of the Oriskany sandstone at Nearpass, N. J. The surface is granulose near the margins. This species should probably be considered intermediate and transitional between *D. pleuroptyx* and *D. dentatus*.

OSTRACODA

Kloedenia smocki (Weller).—Shell subelliptical; hinge line straight, with a narrow margin set off by a narrow furrow; valves divided into three lobes each; two vertical furrows bending toward each other at their lower extremities; posterior furrow somewhat the longer, reaching slightly below the middle of the valve; subcentral in position; surface smooth. Common near the top of the Keyser limestone at Big Mountain, W. Va.

LOCAL SECTIONS OF THE HELDERBERG GROUP

[The numbers of the following sections correspond to those showing locations on Figure 6. In the lists of fossils striking profusion or great rarity is indicated by the symbol "aa" or "r," respectively, after the name of the fossil concerned; otherwise the species will be found readily enough, at least after some search. A number after the name of a fossil indicates feet above the base of the unit and means that within the unit the species is very profuse at or is restricted to that horizon]

1. *Section near Petersburg, W. Va.*

[This section is to be seen about 2½ miles east of Petersburg, W. Va., along the road to Moorefield, in the east end of the gap of the South Branch of the Potomac, where the Helderberg beds are brought up in the eastern limb of a low anticline. The base of the exposure is about 105 feet west of the Hardy-Grant County boundary marker; the New Scotland is exposed along the road 800 feet east of the county line and 280 feet west of the bridge by which the highway crosses the South Branch; and the Shriver-Ridgeley contact is about opposite the bridge. The measurements of the Ridgeley and Shriver are only approximate]

	Feet
Oriskany group:	
Ridgeley sandstone:	
Medium to thick bedded hard gray sandstone	50
Concealed along road; some sandstone on hill	70
Thick-bedded to massive calcareous gray sandstone	60
Concealed	25
Thin to medium bedded calcareous gray sandstone; middle part concealed	60
Oriskany group—Continued.	
Shriver chert:	
Concealed	10
Hard black chert, bedded, and some interbedded sandstone; *Anoplotheca flabellites, Spirifer* sp.	50
Mostly concealed; some bedded black chert 80 and 110 feet above base	125
Helderberg group:	
New Scotland limestone:	
Thin-bedded shaly limestone. *Spirifer macropleurus, S. cyclopterus*	3
Medium to thick bedded hard blue limestone, with much interbedded white chert. *Streptelasma strictum, Dalmanella perelegans, Leptaena rhomboidalis, Strophonella undaplicata, Schuchertella woolworthana, Spirifer macropleurus, S. perlamellosus* (r)	30
Coeymans limestone: Thick-bedded to massive hard crystalline crinoidal gray limestone, grading into the overlying cherty limestone of the New Scotland but separated more abruptly from the underlying Keyser. *Leptaena rhomboidalis, Schuchertella woolworthana,* and *Meristella arcuata* (aa), all in upper 3 feet; *Gypidula coeymanensis* (aa), 9; *Dalmanites pleuroptyx* (r)	19
Keyser limestone:	
Upper limestone member:	
Hard dark-blue thick-bedded, somewhat nodular limestone. Chaetetoid Bryozoa (aa), 16; *Tentaculites gyracanthus* (aa), 12–14; *Schuchertella prolifica* (aa), 3	26½
Medium-bedded hard blue limestone	8½
Thick-bedded hard blue, partly crystalline limestone; weathered surfaces rutted. *Strophonella keyserensis* (r)	16½
Nodular impure limestone, partly concealed. *Mariacrinus* sp. (r), Bryozoa, *Schuchertella prolifica*	6½
Concealed	4
Band of hard gray limestone	½
Nodular impure shaly limestone. *Meristella praenuntia*, 5	6½
Band of hard gray crystalline limestone	½
Nodular impure limestone, partly concealed. *Camarotoechia* cf. *C. altiplicata*, Bryozoa	11
Heavy-bedded hard nodular blue limestone	6½
Concealed	9
Thick-bedded hard grayish, somewhat crystalline limestone; a few chert nodules toward center. *Stromatopora* reef near center	14
Nodular impure limestone. *Dalmanella concinna* (aa)	3½
Massive gray, somewhat crystalline limestone, weathering yellow. *Atrypa reticularis* (aa)	5
Thick-bedded hard dark-blue limestone. Chaetetoid Bryozoa, 2	9½
Nodular impure limestone	6
Hard blue limestone	2
Concealed	3
Hard blue fossiliferous limestone. *Merista typa* (aa), *Strophonella keyserensis* (r), *Leptaena rhomboidalis*	2
Somewhat nodular blue limestone; this unit is largely concealed along the roadside but may be seen on the slope above. *Gypidula coeymanensis* var. *prognostica* and *Camarotoechia gigantea* in lower 2 feet	9

	Feet
Helderberg group—Continued.	
Keyser limestone—Continued.	
Big Mountain shale member: Concealed; some shale fragments seen in upper part	28
Lower limestone member:	
Hard nodular fine-grained limestone. Unfossiliferous	10
Concealed; probably shale	13
Thick-bedded pearly-gray crystalline crinoidal limestone	15½
Blue, somewhat crinoidal limestone	3⅖
Massive impure, very nodular limestone, weathering into grayish lumpy masses, with a somewhat cross-bedded appearance. *Chonetes jerseyensis, Atrypa reticularis, Stenochisma deckerensis,* and *Uncinulus convexorus* are more or less abundant throughout	15
Massive nodular hard fine-grained blue limestone. *Stenochisma deckerensis, Spirifer octocostatus* (r)	23
Thin-bedded blue, somewhat crystalline limestone, with gray calcareous shale at top and bottom. *Favosites* sp., *Cyphotrypa corrugata, Rhipidomella emarginata, Camarotoechia litchfieldensis* (aa), *Whitfieldella minuta* (aa), *Rhynchospira globosa*(?)	11
Hard blue fine-grained limestone	2
Tonoloway limestone:	
Thin-bedded gray calcareous shale	7
Laminated blue limestone, weathering into plates	2
Thin-bedded soft calcareous shale	5
Concealed to western limb of anticline.	

2. *Section in field just north of road to Moorefield, about 4 miles west of Wardensville, W. Va.*

[The section can be located by the prominent exposure of the Ridgeley sandstone, which dips steeply to the west and then is followed, in succession, by the Helderberg, Tonoloway, and Wills Creek, the lower half of the Tonoloway and the entire Wills Creek are well exposed along the roadside]

	Feet
Oriskany group: Ridgeley sandstone: Massive calcareous sandstone, weathering yellowish	275±
Helderberg group:	
Concealed. Some white chert fragments found at the middle of this interval apparently represent the New Scotland, as they contain large strophonellids and a fragmental *Spirifer* which is perhaps *S. macropleurus.* Whether the New Scotland is overlain in this section by the Becraft or the Shriver chert (Oriskany) is uncertain	95
Keyser limestone (probably should include basal portion of overlying interval).	
Upper limestone member:	
Dark-gray fine-grained limestone	5
Nodular shaly limestone, with some interbedded shale. *Merista typa, Atrypa reticularis*	19
Dark-blue limestone, with a chaetetoid Bryozoa zone. *Merista typa?*	4
Big Mountain shale member:	
Calcareous greenish shale and some impure interbedded limestone	8
Shaly limestone. *Spirifer modestus*	4
Calcareous greenish shale	33
Lower limestone member:	
Massive dark-blue limestone	17
Shaly nodular limestone and some calcareous shale	14
Tonoloway limestone, excellently exposed in the field and along the highway.	

3. *Section in gap of Trout Run about 3 miles south of Wardensville, W. Va.*

[The Ridgeley sandstone is exposed at the west end of the gap, the beds dipping to the west. The Helderberg is exposed along the banks of the creek]

	Feet
Oriskany group: Ridgeley sandstone: Massive gray quartzitic sandstone; lower beds softer	275
Helderberg group:	
Upper part of Helderberg (Becraft, New Scotland, Coeymans):	
Massive dense, somewhat impure limestone, containing interbedded black chert. Probably Becraft	7½
Concealed	24
Massive dark, somewhat crinoidal limestone, containing fragmental fossils, possibly *Gypidula.* Coeymans?	1½
Concealed	28
Very massive dark-blue limestone	3
Massive gray, highly crinoidal crystalline limestone. All or part of this unit may belong in the Keyser	15
Keyser limestone (top doubtful):	
Upper limestone member:	
Lumpy drab-gray limestone	9
Concealed	8
Lumpy dark-blue limestone	8
Concealed	10
Massive gray crinoidal limestone	3
Concealed	9
Massive gray crinoidal limestone	2½
Impure lumpy limestone	9
Big Mountain shale member: Shaly limestone to calcareous shale	38
Lower limestone member:	
Massive gray crinoidal limestone	23
Lumpy impure grayish limestone. *Uncinulus convexorus* (aa) throughout, *Spirifer modestus*	16
Tonoloway limestone: Thin-bedded platy limestone.	

4. *Section at Fawcetts Gap, Little North Mountain, near Marlboro, Va.*

[The section is exposed at west end of gap, on slope above the county road running north from the road from Marlboro to Star Tannery]

	Feet
Romney shale: Somewhat fissile brownish shale.	
Concealed (Romney shale?)	30
Helderberg group:	
Becraft limestone:	
Massive arenaceous gray limestone with much interbedded black chert. Probably Becraft	9
Concealed	14½
Gray limestone, some chert	3½
New Scotland limestone: Massive gray crystalline limestone, with a few lenses of chert. *Dalmanella perelegans, Spirifer macropleurus, Meristella arcuata* (aa)	10
Keyser limestone (the upper crystalline beds may belong to the Coeymans limestone; division of the Keyser into the members recognized elsewhere is not feasible):	
Gray crystalline limestone, weathering thin-bedded. Possibly Coeymans rather than upper Keyser	14
Massive gray crinoidal limestone	8

	Feet
Helderberg group—Continued.	
Keyser limestone—Continued.	
Thick-bedded dark-blue limestone, weathering somewhat nodular	25
Medium-bedded gray crinoidal limestone; coral zone at top. *Stenochisma deckerensis*	5
Tonoloway limestone: Platy limestone, shaly above.	

5. *Section on Big Mountain, W. Va.*

[The Helderberg group is well exposed on Big Mountain along the county road running through the gap of the South Fork of the Potomac. The section is in the eastern limb of the Big Mountain anticline and is about 1½ miles west of Upper Tract, a small settlement on the highway to Franklin, about 17 miles south of Petersburg, W. Va.]

	Feet
Oriskany group:	
Shriver chert:	
Concealed. The measurement to the first exposures of the Ridgeley sandstone is unsatisfactory but could be obtained in the western limb of the anticline.	
Bedded black chert, weathering buff; layers 4 to 6 inches thick	45
Concealed	6
Bedded black chert, weathering buff; the lower half is largely concealed	70
Helderberg group:	
New Scotland limestone:	
Thin-bedded, somewhat shaly limestone, with some interbedded white chert. *Spirifer macropleurus*	16½
Thick-bedded limestone with much interbedded white chert. *Hindia sphaeroidalis, Streptelasma strictum* (aa), and *Spirifer macropleurus* (aa) in all but the lower 2 feet or so. *Dalmanella perelegans, Leptaena rhomboidalis, Schuchertella woolworthana,* and *Meristella arcuata* (aa), throughout	20½
Coeymans limestone: Thick-bedded to massive hard somewhat crystalline gray crinoidal limestone. *Orthostrophia strophomenoides* (r), *Rhipidomella oblata, Schuchertella woolworthana,* and *Uncinulus abruptus* (r) near top; *Spirifer cyclopterus* and *Meristella arcuata* (aa) throughout; *Gypidula coeymanensis* (r) near middle	24
Keyser limestone:	
Upper limestone member:	
Thin to medium bedded hard bluish-gray, somewhat crinoidal limestone. *Camarotoechia* cf. *C. altiplicata; Kloedinia smocki,* 13	44½
Thick-bedded hard blue crinoidal limestone. *Camarotoechia* cf. *C. altiplicata, Schuchertella prolifica, Rensselaeria mutabilis* (r), 4	13½
Shaly impure limestone, partly concealed. *Schuchertella prolifica*	10
Thick-bedded hard bluish-gray, somewhat crinoidal limestone. *Meristella praenuntia* (aa), 14; *Stromatopora,* 10; chaetetoid Bryozoa	19
Thin-bedded shaly limestone, with several beds of gray crinoidal limestone. Many Bryozoa near top	21
Helderberg group—Continued.	
Keyser limestone—Continued.	
Upper limestone member—Continued.	
Thin to medium-bedded gray, somewhat crinoidal limestone; the weathered surfaces have a peculiar rutted appearance. *Cyrtina dalmani* (r), 20; *Stromatopora* (aa), 18; Bryozoa; *Merista typa* (aa), 4–6; *Gypidula coeymanensis,* 4; *Meristella nasutaformis* (r), 1	21
Big Mountain shale member:	
Thin-bedded greenish shale, with some beds of very impure lumpy limestone. *Dalmanella concinna, Leptaena rhomboidalis* (r), *Chonetes jerseyensis, Atrypa reticularis, Orthoceras rigidum* (r)	25
Impure lumpy limestone	7
Thin-bedded greenish shales	9
Lower limestone member:	
Gray, somewhat crystalline limestone. *Chonetes jerseyensis, Schuchertella deckerensis, Uncinulus convexorus*	14
Blue, very nodular limestone. *Cyphotrypa corrugata, Dalmanella concinnus, Stropheodonta bipartita, Chonetes jerseyensis, Stenochisma deckerensis, Uncinulus convexorus, Atrypa reticularis, Spirifer modestus, S. modestus* var. *plicatus, Nucleospira swartzi, Straparollus welleri* (r), *Orthoceras rigidum* (?) (r), *Proetus protuberans*	15
Hard blue nodular limestone, partly concealed. *Rhipidomella emarginata, Chonetes jerseyensis, Stenochisma deckerensis,* and *Uncinulus convexorus,* all in the upper half of the unit. *Camarotoechia litchfieldensis* and *Whitfieldella minuta* in lower 5 feet	29
Thin-bedded blue limestone, weathering yellow	7½
Hard blue somewhat crinoidal limestone. *Cyphotrypa corrugata*	4½
Hard blue fine-grained somewhat nodular limestone	3
Tonoloway limestone:	
Thin-bedded laminated blue limestone	1½
Concealed. The Tonoloway is well exposed in the western limb of the anticline, where it can be seen along the roadside.	

6. *Section near Franklin, W. Va.*

This section is exposed on the south end of the Pond Range, along the State highway to Circleville, W. Va., about 4 miles west of Franklin. It is in the eastern limb of a large anticline that brings up the Tuscarora sandstone farther west. The Helderberg group is for the most part concealed along the immediate roadside but is almost continuously exposed on the slope above the road]

	Feet
Oriskany group:	
Ridgeley sandstone: Thick-bedded gray sandstone.	
Shriver chert:	
Bedded black chert, partly concealed	105
Concealed	45

	Feet
Helderberg group:	
New Scotland limestone: Thick-bedded to massive limestone, with much interbedded white chert. *Hindia sphaeroidalis, Streptelasma strictum* (aa), *Dalmanella perelegans, D. eminens* (r), *Leptaena rhomboidalis, Schuchertella woolworthana, Eatonia medialis, Spirifer perlamellosus* (r), *S. cyclopterus, S. macropleurus* (aa), *Trematospira equistriata, Meristella arcuata*	22
Coeymans limestone: Massive gray crystalline crinoidal limestone, more or less transitional with the overlying New Scotland. *Meristella arcuata* (aa), near top; *Schuchertella woolworthana; Gypidula coeymanensis* at middle	20
Keyser limestone:	
Upper limestone member:	
Very arenaceous limestone, weathering thin-bedded. This bed might be placed at the base of the overlying Coeymans. *Spirifer vanuxemi* var. *prognosticus* (aa) and *Whitfieldella prosseri* (aa), 7	10
Thin-bedded argillaceous limestone, somewhat concretionary. *Schuchertella prolifica, Spirifer vanuxemi* var. *prognosticus, Whitfieldella? prosseri*	4
Thick-bedded to massive gray, coarsely crystalline crinoidal limestone. *Cladopora rectilineata*, 45 to 50; *Mariacrinus?* sp. (aa), 50; *Camarotoechia* cf. *C. altiplicata*, 16 and above; *Rensselaeria mutabilis* (aa), 16; *Uncinulus keyserensis* (r); *Uncinulus nucleolatus* (r)	70
Thick-bedded gray crystalline limestone, forming a series of small cliffs on the hillside. *Stromatopora*—coral reef near top; *Cladopora rectilineata*, 2 and 32; *Camarotoechia* cf. *C. altiplicata*, 22; *Uncinulus nucleolatus* (r), 22; *Rensselaeria mutabilis*, 22	36
Medium-bedded gray, somewhat crystalline limestone, mostly concealed	18
Massive pearly-gray crystalline limestone, forming a series of small cliffs. *Stromatopora* bed, 28	32
Medium to thick-bedded, somewhat crystalline gray limestone, weathering more readily than the overlying beds. *Gypidula coeymanensis* var. *prognosticus* (aa), *Camarotoechia gigantea* (aa), *Atrypa reticularis* (aa), and *Merista typa* (aa), all in the lower 6 feet. *Merista typa*, however, ranges a little higher than the three other species, and the Gypidulas are practically confined to the basal foot or so	12
Big Mountain shale member: Soft yellowish shale, concealed for the most part	45
Lower limestone member:	
Crystalline crinoidal limestone, weathering thin-bedded	6
Massive dark-blue, very nodular limestone. *Cyphotrypa corrugata, Dalmanella concinna* (r), *Stropheodonta bipartita, Chonetes jerseyensis, Stenochisma deckerensis, Uncinulus convexorus, Spirifer modestus, S. modestus* var. *plicatus, S. octocostatus.* The above fossils are not as abundant here as they are in the equivalent beds in most of the other sections	28
Helderberg group—Continued.	
Keyser limestone—Continued.	
Lower limestone member—Continued.	
Thick-bedded blue limestone. This bed should probably be considered the base of the Keyser, although it was not seen in contact with the underlying Tonoloway	6

7. *Section near Fulks Run, Va.*

[A fine exposure of the Lower Devonian can be seen along the highway from Broadway, Va., about half a mile west of the Fulks Run store and post office]

	Feet
Oriskany group: Ridgeley sandstone: Massive calcareous sandstone	100+
Shriver chert:	
Bedded black chert, matrix sandy	7
Bedded black chert, largely concealed	43
Sandy, impure limestone, with some black chert. The character of this unit suggests that it should be considered transitional with the Becraft limestone	6
Helderberg group:	
New Scotland limestone:	
Drab shale, partly concealed	3½
Limestone with some interbedded dark-colored chert. *Spirifer macropleurus, Meristella arcuata*	13½
Keyser limestone (including beds of age of the Coeymans limestone?):	
Upper limestone member:	
Concealed. Any beds of Coeymans age would lie in the upper part of this interval	85
Impure, somewhat nodular limestone	4
Concealed	8
Calcareous sandstone, weathered buff	1
Impure shaly limestone above to calcareous shale below	14
Massive crinoidal limestone. *Merista typa*	3
Big Mountain shale member:	
Concealed; probably shale	21
Nodular blue limestone	1½
Greenish shaly limestone and calcareous shale	16½
Lower limestone member:	
Massive crinoidal limestone	3
Greenish shaly limestone. *Chonetes jerseyensis*	15
Massive crinoidal limestone	17
Tonoloway limestone: Blue platy limestone.	

8. *Section at Strait Creek, Va.*

[The lower part of the Keyser is exposed along the road from Franklin, W. Va., to Monterey, Va., about ½ mile east of the store at Strait Creek, Va., in the eastern limb of the anticline through which the road runs at this place. The higher part of the Helderberg is concealed, although some New Scotland float was seen]

	Feet
Helderberg group:	
Keyser limestone:	
Big Mountain shale member:	
Concealed.	
Fissile olive-green shale	10
Lower limestone member:	
Thick-bedded hard gray, somewhat crystalline limestone	11

	Feet
Helderberg group—Continued.	
Keyser limestone—Continued.	
Lower limestone member—Continued.	
Shaly, very lumpy limestone, weathering yellowish. *Dalmanella concinna* (aa), *Stropheodonta bipartita*, *Chonetes jerseyensis* (aa), *Atrypa reticularis*, *Spirifer modestus* (aa), *S. modestus* var. *plicatus*, *Nucleospira swartzi* (aa), *Orthoceras* cf. *O. clavatum* (r)	35
Medium-bedded dark gray limestone. *Chonetes jerseyensis* (r)	3
Tonoloway limestone: Laminated blue limestone, upper part concealed.	

9. *Section near Monterey, Va.*

[The Helderberg group is well exposed half a mile east of Monterey, Va., just north of the State highway to Staunton, on the first foothill of Jacks Mountain. The Becraft limestone and the cherty New Scotland limestone are conspicuous on the hillside]

	Feet
Oriskany group:	
Ridgeley sandstone: Mostly concealed, with boulders of friable sandstone, weathered whitish on the surface but iron stained within	50±
Concealed, with a few exposures of very fine grained calcareous sandstone. *Anoplotheca flabellites*	50±
Helderberg group:	
Becraft limestone:	
Heavy-bedded white calcareous sandstone	10
Massive grayish sandy limestone. *Rensselaeria subglobosa*, *Spirifer cyclopterus*	4
Massive gray sandy limestone to limy sandstone	5
Hard massive bluish-gray limestone, very fossiliferous; a few lenses of dark chert in lower part. *Eatonia medialis*, *E. peculiaris*, *Spirifer cyclopterus*, *S. concinnus*, etc	71
Partly concealed; some beds of limestone with much interbedded black chert	20
Shriver (?) chert:	
Mostly concealed, with much impure dark bedded chert of Shriver type on the surface	85
New Scotland limestone: Fine-grained limestone with much interbedded white chert, conspicuous on the hillside. The upper 6 feet is less resistant than the remainder. *Hindia sphaeroidalis*, *Streptelasma strictum* (aa), *Favosites helderbergiae?*, *Dalmanella perelegans* (aa), *Leptaena rhomboidalis*, *Strophonella undaplicata*, *S. leavenworthana*, *Schuchertella woolworthana*, *S. punctulifera*, *Eatonia medialis*, *Spirifer cyclopterus*, *S. macropleurus* (aa), *Trematospira equistriata*, *Meristella arcuata*	25
Coeymans limestone:	
Massive hard gray crystalline crinoidal limestone. *Rhipidomella oblata*, *Leptaena rhomboidalis*, *Stropheodonta arata* (r), *Schuchertella woolworthana*, *Uncinulus abruptus*, *Meristella arcuata*, and *M. arcuata* var. *gigas*, all in the upper 5 feet. *Schuchertella woolworthana*, *Gypidula coeymanensis* (r), and *Meristella arcuata* near middle and about 10 feet above base. *Nucleospira ventricosa* (aa) near base	44
Thin-bedded, somewhat arenaceous limestone	6
Keyser limestone:	
Upper limestone member:	
Thick-bedded blue limestone, with narrow brown sandy lines on the weathered surface	13

	Feet
Helderberg group—Continued.	
Keyser limestone—Continued.	
Upper limestone member—Continued.	
Medium-bedded blue limestone, with a few lenses of light-colored chert	4
Concealed	81
Thick-bedded gray, somewhat crinoidal limestone. *Aulopora schucherti*, 2; *Strophonella keyserensis* (r); *Stropheodonta* cf. *S. planulata*, 4; *Schuchertella prolifica; Gypidula coeymanensis* var. *similis*, 5; *Camarotoechia gigantea* and *C. gigantea* var. *gigas*, 2; *Cyrtina dalmani* (r), 19; *Nucleospira ventricosa* (aa), 19; *Merista typa* (aa), 4 to top	20
Gray, somewhat crinoidal crystalline limestone; upper part concealed. *Striatopora bella*, *Leptaena rhomboidalis*, *Chonetes jerseyensis*, *Atrypa reticularis*	7
Big Mountain shale member: Yellowish shale, with some thin arenaceous shale layers; largely concealed	28
Lower limestone member:	
Thick-bedded gray, somewhat crinoidal limestone	7
Shaly, much weathered nodular limestone	21
Massive, very nodular blue limestone. *Favosites* sp. (r), *Stropheodonta bipartita*, *Chonetes jerseyensis*, *Uncinulus convexorus*, *Stenochisma deckerensis*, *Atrypa reticularis* (aa)	39
Thin-bedded grayish limestone. *Cyphotrypa corrugata*, *Rhipidomella emarginata*, *Strophonella keyserensis* (r), *Stropheodonta bipartita*, *Chonetes jerseyensis*, *Stenochisma deckerensis* (r), *Camarotoechia litchfieldensis*, *Reticularia* cf. *R. bicostata* (r), *Whitfieldella minuta*	8
Tonoloway limestone:	
Laminated blue limestone, largely concealed	219
Thick-bedded nodular blue limestone. *Atrypa reticularis* (aa), *Stenochisma deckerensis* (one specimen, loose)	75
Laminated blue limestone, largely concealed	262

10. *Section on Little Mountain, Va.*

[All but the topmost part of the Keyser is exposed on the western slope of Little Mountain, along the State highway to Bartow, W. Va., about 5 miles west of Monterey, Va. The section is in the western limb of the anticline that centers in Crabbottom Valley, east of Little Mountain; but as the beds are overturned, they dip to the east. The Coeymans, New Scotland, Shriver, and Ridgeley are not exposed here and have possibly been cut out by faulting. The section here given is a composite of two exposures along the roadside. In going down the road from the mountain top, the Tonoloway is the first limestone seen. It crops out for some distance, as the road makes only a small angle with the strike; the road then turns sharply to the left and cuts through the upper Tonoloway and through the lower Keyser to a level about 15 feet above the *Gypidula* zone. The road then cuts back into the underlying shale member, until it bears to the left again and cuts through the upper part of the Keyser about 500 feet farther on. The two exposures may be matched by means of the *Gypidula* zone, seen on the bank above the second exposure]

	Feet
Helderberg group:	
Keyser limestone:	
Upper limestone member:	
Concealed. (Compare with the Monterey section.)	
Thin to medium-bedded drab-gray limestone, with a few minor lenses of white chert. *Stromatopora*, 13; *Camarotoechia* cf. *C. altiplicata*, *Nucleospira ventricosa*	16

	Feet
Helderberg group—Continued.	
Keyser limestone—Continued.	
Upper limestone member—Continued.	
Thin-bedded, somewhat argillaceous limestone. Many Bryozoa throughout. *Camarotoechia* cf. *C. altiplicata*, *Uncinulus gordoni* (r), *Rensselaeria mutabilis*, 27; *Spirifer perlamellosus* var. *praenuntia*, *Actinopteria* cf. *A. reticulata* (r)	28
Massive gray crystalline limestone. *Stromatopora*, 26; *Cladopora rectilineata* (aa), *Striatopora bella*, 16–28; *Ceratopora marylandica*, 16–28; *Aulopora schohariae*, 16–28; *Cyathophyllum radiculum*, 16–28; *Schuchertella prolifica*, *Spirifer perlamellosus* var. *praenuntia*, *Nucleospira ventricosa*	28
Massive, somewhat crystalline crinoidal limestone, with a fine-grained greenish matrix	6
Thick-bedded gray crinoidal limestone. Chaetetoid Bryozoa, 15. The following fauna occurs in the lower 8 feet: *Leptaena rhomboidalis*, *Strophonella keyserensis*, *Schuchertella prolifica*, *Chonetes jerseyensis*, *Gypidula coeymanensis* var. *similis* (aa), 5; *Camarotoechia gigantea*, 5–7; *Atrypa reticularis* (aa), 7; *Merista typa*, 7	27
Big Mountain shale member: Soft olive-green shale, with a 1-foot bed of shaly sandstone 5 feet above base. *Orthopora rhombifera* (aa), 12; *Strophonella keyserensis*, *Schuchertella sinuata*, *Chonetes jerseyensis* (r)	28
Lower limestone member:	
Massive hard gray crystalline crinoidal limestone	4½
Thin-bedded shaly nodular limestone. *Dalmanella concinna*, *Chonetes jerseyensis*, *Stenochisma deckerensis* (aa), *Uncinulus convexorus* (aa), *Atrypa reticularis*, *Spirifer modestus*, *Camarotoechia litchfieldensis*, 15	18
Thick-bedded nodular blue limestone. *Chonetes jerseyensis*, *Atrypa reticularis* (aa)	29½
Thin-bedded, somewhat nodular grayish limestone. *Atrypa reticularis*, *Spirifer octocostatus* (r)	3
Hard, somewhat crystalline limestone. *Atrypa reticularis*	3
Thin-bedded grayish, somewhat crinoidal limestone. *Cyphotrypa corrugata*, *Dalmanella concinna*, *Rhipidomella emarginata*, *Stropheodonta bipartita*, *Chonetes jerseyensis*	7½
Hard grayish limestone	4
Thin-bedded grayish shaly limestone. *Camarotoechia litchfieldensis*	12½
Tonoloway limestone:	
Concealed	49
Laminated blue limestone	28
Concealed.	

11. *Section near McDowell, Va.*

[The Helderberg group is finely exposed on the eastern slope of Bullpasture Mountain, about 4 miles east of McDowell, Va., on the highway to Staunton]

	Feet
Oriskany group:	
Ridgeley sandstone: Thick-bedded gray sandstone, weathered buff	115
Shriver chert:	
Concealed. Comparison with the section on Black Creek Mountain, west of Warm Springs, suggests that this interval may contain tongues of the upper part of the Becraft. So far as seen, the soil does not indicate chert of the Shriver type	75
Bedded black chert, unfossiliferous	57
Helderberg group:	
New Scotland limestone: Massive gray limestone, with much interbedded white chert. The chert is not, however, as predominant here as at Monterey and farther north. *Hindia sphaeroidalis*, *Streptelasma strictum*, *Dalmanella perelegans*, *Schuchertella woolworthana*, *Uncinulus abruptus* (r), *Spirifer perlamellosus* (r), *S. cyclopterus*, *S. macropleurus* (aa), *Meristella arcuata* (aa)	23
Coeymans limestone:	
Massive gray crystalline limestone. *Rhipidomella oblata*, *Schuchertella woolworthana*, and *Meristella arcuata* (aa) in upper 2 feet. *Spirifer cyclopterus* (aa), 20–45; *Whitfieldella prosseri*, 8; *Platyceras multiplicatus* (r)	53
Arenaceous gray limestone	4
Keyser limestone:	
Upper limestone member:	
Medium-bedded gray, somewhat crystalline crinoidal limestone. *Nucleospira ventricosa* (aa)	11
Medium-bedded, somewhat nodular blue limestone. *Nucleospira ventricosa* and *Meristella praenuntia*, 16–20	28
Thick-bedded arenaceous blue limestone	8
Medium-bedded gray crystalline limestone	5
Blue, somewhat nodular limestone	4
Thick-bedded to massive gray crystalline crinoidal limestone. *Gypidula coeymanensis* var. *similis* (aa), *Camarotoechia gigantea*, and *Atrypa reticularis* (aa), 0–5; *Merista typa* (aa), 5	42
Big Mountain shale member: Soft fissile yellowish shale	15
Lower limestone member: Massive gray crystalline crinoidal limestone; the nodular character that is generally so conspicuous is lacking here. *Cladopora rectilineata* (r), 40; *Chonetes jerseyensis*, *Stenochisma deckerensis*, *Uncinulus convexorus*, *Spirifer modestus*, *S. modestus* var. *plicatus*, *Whitfieldella minuta* (r), 3. These fossils are by no means so abundant here as they are in the equivalent beds at Monterey and farther north	42
Tonoloway limestone:	
Laminated blue limestone; upper 30 feet shaly	185
Lumpy shaly blue limestone. *Atrypa reticularis*, *Uncinulus convexorus* var	16
Laminated blue limestone	

12. *Section south of Bolar, Va.*.

[The Helderberg group is exposed about 3½ miles south of Bolar, Va., on the southern slope of a foothill west of Jacks Mountain, just east of the bridge over which the road from Monterey to Warm Springs crosses the Jackson River. A large eastern loop of the river cuts into the north side of this hill, and the Helderberg is exposed there also. The formations above the New Scotland are concealed]

	Feet
Helderberg group:	
New Scotland limestone:	
Hard gray arenaceous limestone with some interbedded white chert. This member forms the southward-thinning tongue of the cherty limestone phase of the New Scotland. *Streptelasma strictum* (aa), *Dalmanella perelegans, Leptaena rhomboidalis, Schuchertella woolworthana, Spirifer cyclopterus, S. macropleurus*	6
Medium to thick-bedded gray arenaceous limestone to calcareous sandstone. This member (here named the Healing Springs sandstone member) forms the entire New Scotland farther south	13
Coeymans limestone: Massive gray crystalline, somewhat crinoidal limestone. *Rhipidomella oblata, Schuchertella woolworthana, Meristella arcuata* (aa), and *M. arcuata* var. *gigas* in upper 5 feet; *Stropheodonta arata* (r), *Schuchertella woolworthana, Gypidula coeymanensis* (r), and *Meristella arcuata* (aa) at about middle	53
Keyser limestone:	
Upper limestone member:	
Medium to thick-bedded crystalline limestone, largely concealed	29
Massive grayish, somewhat nodular limestone. *Stromatopora*	9
Concealed	30
Massive limestone, abounding in reef corals and Bryozoa embedded in a greenish limestone matrix	15
Thick-bedded gray, rather coarsely crystalline limestone. *Stromatopora* and corals at top; *Cladopora rectilineata* (r), *Aulopora schucherti*, and *Merista typa*, 25	38
Concealed	6
Gray arenaceous limestone, weathering buff and porous. This is a northern tongue of the Clifton Forge sandstone member of the Keyser	3
Concealed. It is possible that part of this interval belongs with the Big Mountain shale member	22
Big Mountain shale member: Thin-bedded greenish, somewhat arenaceous shale	18
Lower limestone member:	
Hard gray crystalline, somewhat arenaceous limestone	2
Massive bluish-gray nodular limestone	6
Medium-bedded hard gray, somewhat crystalline limestone	25
Medium to thick bedded hard gray arenaceous limestone	8
Tonoloway limestone:	
Concealed	180
Thick-bedded blue limestone, partly concealed	170
Concealed.	

13. *Section on Back Creek Mountain, west of Warm Springs, Va.*

[The Helderberg group is exposed on the eastern slope of Back Creek Mountain about 4 miles west of Warm Springs along the road to Frost, W. Va. The road ascends for a considerable distance along the slope of the Oriskany sandstone and then, turning sharply to the left, cuts through the Helderberg, Tonoloway, and Wills Creek formations, finally exposing the Bloomsburg sandstone member of the Wills Creek shale near the summit of the mountain. The Helderberg group is mostly concealed above the *Gypidula* zone of the Keyser; consequently the thicknesses given for that part of the section are only approximate]

	Feet
Oriskany group: Ridgeley sandstone: Massive sandstone, weathering buff. Upper contact not seen	100
Helderberg group:	
Becraft limestone:	
Concealed	30
Dark-gray, somewhat crystalline limestone. *Schuchertella woolworthana, Rhipidomella assimilis, Eatonia peculiaris, Rensselaeria subglobosa, Spirifer cyclopterus, S. concinnus, Cyrtina varia, Dalmanites* sp. (suggests *D. dentatus*)	4
Concealed	12
Hard gray, somewhat crystalline limestone. *Rhipidomella assimilis, Schuchertella woolworthana, Eatonia peculiaris, Uncinulus pyramidalis, Rensselaeria subglobosa, Spirifer cyclopterus, S.* cf. *S. angularis, S. concinnus, Anoplotheca flabellites* (r), *Meristella lata* (r), *Dalmanites* sp. (suggests *D. dentatus*). This and the higher bed of limestone represent part of the upper beds of limestone assigned to the Becraft in the Bells Valley and Clifton Forge sections	3
Concealed	20
Shriver? chert: Impure bedded dark chert, weathering buff, partly concealed. Some of the chert is purer than is typical of the Shriver and weathers whiter. In my opinion, this chert is more or less transitional between what I regard as the equivalent lower part of the cherty Becraft limestone of the Healing Springs section and the lower part of the more typical Shriver chert of Monterey and farther north. Affinity to the Becraft is shown by the occurrence of *Eatonia medialis* and *Dalmanites pleuroptyx?* 70 feet above the base	100
New Scotland limestone (Healing Springs sandstone member): Calcareous sandstone, weathered buff and porous	10
Coeymans limestone: Concealed. The thickness given is taken from the Dry Run section	34±
Keyser limestone:	
Upper limestone member:	
Concealed	60
Calcareous sandstone, weathered buff. This, with the underlying unit, forms a tongue of the Clifton Forge sandstone member	12
Thin-bedded, somewhat concretionary shaly sandstone. *Atrypa reticularis* and *Camarotoechia gigantea* abundant in the lower 5 feet	11
Gray crystalline limestone	2
Blue fine-grained limestone, abounding in *Gypidula coeymanensis* var. *similis*	2
Soft, weathered arenaceous limestone. This may be considered another tongue of the Clifton Forge sandstone member	3

	Feet
Helderberg group—Continued.	
Keyser limestone—Continued.	
Upper limestone member—Continued.	
Hard blue arenaceous limestone, with a somewhat lumpy appearance	5
Big Mountain shale member:	
Yellowish fissile shale, partly concealed	7
Hard blue arenaceous limestone, weathering buff and sandy; another tongue of the Clifton Forge sandstone	18
Olive-green shale, with thin sandy layers	11
Lower limestone member:	
Massive, somewhat lumpy blue limestone, very arenaceous in the upper half. *Cyphotrypa corrugata*, abundant in the lower half; *Stenochisma deckerensis, Stropheodonta bipartita, Chonetes jerseyensis, Uncinulus convexorus*	46
Concealed	4
Sandy limestone	6
Tonoloway limestone:	
Laminated blue limestone	21
Blue limestone and some calcareous shale	160
Thin-bedded calcareous sandstone	60
Dark-blue limestone.	

14. *Section along Dry Run east of Warm Springs, Va.*

[An instructive section of the Helderberg group can be seen on the north bank of Dry Run, just west of the point where it cuts across the road to Williamsville, about 2¾ miles northeast of the Alleghendoah Club (shown as Bath Alum on the United States Geological Survey's topographic map of the Monterey quadrangle). The section is in the western limb of the Tower Hill syncline. The Ridgeley sandstone crops out immediately west of the Williamsville road, and the underlying rocks are exposed more or less continuously to the base of the sandstone overlying the *Gypidula* zone of the Keyser. The beds from the sandstone of lower New Scotland age (Healing Springs sandstone member) to a horizon 30 feet below the *Gypidula* zone are much better exposed about 200 feet farther west along the bank of the run, where they are repeated, probably by faulting. The two sections are here combined. Neither section reaches to the base of the Keyser]

	Feet
Romney shale: Black shale seen along run.	
Oriskany group: Ridgeley sandstone: Massive gray sandstone, weathering buff. At the time of my visit the contact with the Romney could be seen along the creek	32
Helderberg group:	
Becraft limestone:	
Concealed; probably Becraft	23
Medium to thick bedded limestone, with minor lenses of black chert. *Leptaena rhomboidalis, Spirifer cyclopterus, S. concinnus, Meristella lata, Dalmanites pleuroptyx*	57
Blue limestone, with much interbedded black chert; largely concealed	98
New Scotland limestone:	
Massive gray limestone, with a few lenses of white chert. *Streptelasma strictum, Dalmanella perelegans, Leptaena rhomboidalis, Schuchertella woolworthana, Eatonia medialis, Spirifer cyclopterus, S. macropleurus*	6
Hard gray arenaceous limestone to calcareous sandstone (Healing Springs sandstone member)	10
Coeymans limestone: Massive crystalline, somewhat crinoidal gray limestone. *Leptaena rhomboidalis, Schuchertella woolworthana, Rhipidomella oblata,* and *Meristella arcuata* (aa) at top	34
Keyser limestone:	
Nodular blue limestone	5
Thick-bedded gray, somewhat crystalline limestone	10

	Feet
Helderberg group—Continued.	
Keyser limestone—Continued.	
Concealed	12
Thick-bedded gray, somewhat crinoidal limestone. *Cladopora rectilineata*	13
Thin-bedded calcareous sandstone, with pronounced cross-bedding. This and the underlying sandstone form a small cliff at both exposures. These sandstones, like the corresponding ones in the section west of Warm Springs, form a northward tongue of the Clifton Forge sandstone member	20
Sandy shale to shaly sandstone, somewhat concretionary	14
Thick-bedded gray limestone, abounding in *Gypidula coeymanensis* var. *similis*	7
Thick-bedded gray crystalline limestone	12
Arenaceous limestone, showing some cross-bedding on weathered surfaces; the lower half is colored red by hematite. *Camarotoechia litchfieldensis* (aa)	15
Concealed. (See Back Creek Mountain section.)	

15. *Section at Bells Valley, Va.*

[The section as given is a composite of the exposures along the county road leading southeast from Bells Valley, and of those on the southern side of the hollow, in and near the old quarry in the New Scotland-Coeymans limestone. The sections were matched chiefly by means of the 4-foot bed of hard sandstone which forms the backbone of the knolls both north and south of the hollow]

	Feet
Oriskany group: Ridgeley sandstone: Hard gray sandstone, weathered buff	5
Helderberg group:	
Becraft limestone:	
Concealed, most probably Becraft	40
Thick-bedded dark-blue limestone, with minor layers of black chert. *Spirifer concinnus* (r)	84
Medium-bedded limestone, with much interbedded black chert. *Edriocrinus pocilliformis, Schuchertella woolworthana, S. becraftensis?* (r), *Phacops logani*	8
New Scotland and Coeymans limestones: Massive gray crystalline, highly crinoidal limestone, forming the quarry bed of the area. The presence of *Spirifer macropleurus* and other characteristic New Scotland forms in the upper 20 feet shows that that part is of New Scotland age. That the lower 55 feet should be assigned to the Coeymans is indicated by the absence of the New Scotland fossils and by the presence, about 55 feet above the base, of the *Meristella arcuata* zone which is so prominent at the top of the Coeymans at Monterey, Bolar, and Dry Run. *Leptaena rhomboidalis, Schuchertella woolworthana, Uncinulus abruptus, Eatonia medialis,* and *Spirifer macropleurus* (aa) in the upper 20 feet; *Favosites helderbergiae* and *Meristella arcuata* (aa) 52 to 56 feet above base; *Platyceras multiplicatum* in lower half	78
Keyser limestone:	
Thin-bedded shale and limestone; the upper surface is extensively ripple marked	6
Thick-bedded crystalline limestone	7
Thin-bedded shaly limestone. *Dalmanella concinna, Schuchertella prolifica* (r)	3
Massive fine-grained limestone, weathering greenish; many Bryozoa and *Stromatopora* 10 feet beneath top	42

	Feet
Helderberg group—Continued.	
Keyser limestone—Continued.	
Thick-bedded hard gray sandstone, weathered buff to white, forming the backbone of the knoll. Like the underlying sandstones, this is a tongue of the Clifton Forge sandstone member of the Keyser	4
Thin-bedded yellowish-green, somewhat arenaceous shale (tongue of the Big Mountain (?) shale member)	9½
Medium-bedded gray sandstone (tongue of the Clifton Forge sandstone member)	4½
Dark-grayish soft shale, mostly concealed	8
Calcareous sandstone, weathered buff (tongue of the Clifton Forge sandstone member)	6
Thick-bedded gray, somewhat crystalline limestone	10
Concealed. Some arenaceous shale toward top. Keyser (?)	34
Tonoloway limestone:	
Laminated blue limestone, concealed in part	135
Nodular blue limestone, much weathered. *Uncinulus convexorus* var., *Atrypa reticularis*	10

16. *Section in gap just west of Hot Springs, Va.*

[Section along railroad at west end of gap]

	Feet
Romney shale: Brownish shale exposed along railroad cut.	
Oriskany group: Ridgeley sandstone: Massive buff, iron-stained sandstone, forming a prominent ledge on each side of the railroad	80
Helderberg group:	
Concealed. Includes Becraft limestone (possibly with some beds of Shriver chert facies) and New Scotland limestone	200
Coeymans limestone: Highly crinoidal crystalline, limestone. *Gypidula coeymanensis* (r), *Platyceras trilobatum*	23
Keyser limestone:	
Upper limestone member:	
Concealed	2
Arenaceous limestone	3
Concealed	20
Clifton Forge sandstone member:	
Calcareous sandstone, weathered buff	12
Arenaceous shale to shaly sandstone, somewhat concretionary. This unit is probably equivalent to the concretionary shaly sandstone above the *Gypidula* zone in the sections east and west of Warm Springs	20
Thick-bedded hard calcareous sandstone, weathered buff, making prominent ledges along the railroad, except the upper 7 feet, which is more weathered	70
Lower limestone member: Blue to gray limestone, somewhat nodular above; crinoidal and somewhat crystalline below	25

17. *Section in gap west of Healing Springs, Va.*

[Seen along road and creek near west end of gap about 3 miles south of section 16. This locality is of interest because it affords a fine exposure of the sandstone phase of the New Scotland (the Healing Springs sandstone member), and because the Becraft limestone presents characters that are more or less intermediate between those of the Becraft of Clifton Forge and those of the Becraft and Shriver (?) chert as seen on Back Creek Mountain, west of Warm Springs]

	Feet
Oriskany group: Ridgeley sandstone: Massive gray sandstone. A bridge crosses the stream at the basal contact	101
Helderberg group:	
Becraft limestone:	
Massive gray crystalline, somewhat crinoidal limestone, lower part more or less concealed. *Eatonia peculiaris*, *Uncinulus vellicatus*, *Spirifer concinnus* (aa), *S. cyclopterus*	35
Limestone, with much interbedded black chert; the lower part in particular is almost solid chert; the upper part is somewhat concealed. Unfossiliferous	70
New Scotland limestone (Healing Springs sandstone member): Massive calcareous sandstone to arenaceous limestone. *Leptaena rhomboidalis*, *Schuchertella woolworthana*, and *Spirifer macropleurus* (r) were found in the exposure north of the creek. South of the creek, above the road, the sandstone makes a prominent ledge on the hillside	20
Coeymans limestone and upper limestone member of Keyser limestone: Massive limestone, largely concealed	105
Clifton Forge sandstone member of Keyser limestone: Massive calcareous sandstone	92
Lower limestone member of Keyser limestone.	
Lumpy blue limestone	14
Somewhat crinoidal arenaceous limestone	19
Tonoloway limestone: Laminated blue limestone.	

18. *Section near Clifton Forge, Va.*

[The most complete exposure of the Helderberg group in the immediate vicinity of Clifton Forge is that at the east end of the gap of the James River, along the road to Iron Gate. Other partial sections can be seen along the railroad just north of Clifton Forge and at Island Ford Bridge, about 8 miles south of Clifton Forge on the road to Covington. The Becraft can also be seen in quarries in the vicinity of Low Moor, 3 or 4 miles south of Clifton Forge. The exposure along the railroad north of Clifton Forge of the sandstone overlying the Becraft limestone is of particular interest in that the sandstone there carries *Spirifer arenosus*, *S. murchisonia*, and other fossils that permit definite correlation with the Ridgeley sandstone (of Oriskany group) of Maryland. The section given below is the one first mentioned]

	Feet
Helderberg group:	
Becraft limestone:	
Concealed. The upper part of the Becraft as seen north of Clifton Forge along the railroad is gray, crinoidal, and massive and contains *Uncinulus vellicatus*, *Eatonia peculiaris*, *Spirifer concinnus*, and other fossils.	
Thick-bedded to massive gray limestone, with some interbedded black chert	43
Massive gray, somewhat nodular limestone, with a little interbedded black chert. *Streptelasma strictum*, *Edriocrinus pocilliformis* (aa), *Uncinulus abruptus*, *Eatonia peculiaris*, *Rensselaeria* cf. *R. aequiradiata*, *Spirifer concinnus*. (r)	8
New Scotland formation (Healing Springs sandstone member): Thick-bedded gray calcareous sandstone to arenaceous limestone	16
Coeymans limestone: Massive gray crystalline crinoidal limestone, arenaceous at base. *Rhipidomella oblata* (aa), *Stropheodonta arata* (r), *Schuchertella woolworthana*, *Camarotoechia campbellana*, *Uncinulus abruptus* (r), and *Meristella arcuata* (aa), all in upper 5 feet	18
Keyser limestone:	
Upper limestone member:	
Massive gray limestone, abounding in reef corals and *Stromatopora* best seen below road, along railroad	15

Helderberg group—Continued.
Keyser limestone—Continued.

	Feet
Upper limestone member—Continued.	
Massive hard gray, somewhat crystalline limestone, seen along railroad. *Cladopora rectilineata* is very abundant at this horizon along the railroad north of Clifton Forge and at Island Ford Bridge	19
Cilfton Forge sandstone member:	
Thin to medium-bedded shaly sandstone	12
Massive hard white sandstone	8½
Thin-bedded arenaceous shale	2½
Massive hard white to buff calcareous sandstone	25
Thin-bedded shaly sandstone, thicker bedded toward the middle	18
Lower limestone member: Somewhat lumpy arenaceous limestone. *Dalmanella concinna, Stropheodonta bipartita, Stenochisma deckerensis, Uncinulus convexorus, Nucleospira swartzi.* The following additional fossils were found at this horizon in exposures along the railroad north of Clifton Forge: *Atrypa reticularis, Calymene camerata* (r), *Jaekelocystites hartleyi* (r)	15

19. *Section at Prices Bluff, near Gala, Va.*

[An excellent section of the Helderberg group can be seen at Prices Bluff, along the Norfolk branch of the Chesapeake & Ohio Railway, 1 mile north of Gala, Va. R. J. Holden, of the Virginia Polytechnic Institute, at Blacksburg, Va., kindly told me of this exposure]

	Feet
Romney shale: Brown shale with some interbedded limestone; of Onondaga age (?).	
Oriskany group: Ridgeley sandstone: Gray quartz sandstone, weathering brown; exposed at southeast end of bluff	1½
Helderberg group:	
Becraft limestone:	
Massive gray crystalline, highly crinoidal limestone, with some brown sandstone lines on weathered surfaces. *Rhipidomella assimilis, Leptaena rhomboidalis, Schuchertella woolworthana, Spirifer concinnus, S. concinnus* var. *progradius,* and *S.* cf. *S. angularis* in upper part	59
Massive gray crystalline limestone, with some lenses of black chert	18
Massive, very arenaceous limestone	11
Very massive gray limestone, somewhat nodular toward the base. *Streptelasma strictum, Edriocrinus pocilliformis, Rhipidomella assimilis*(r), *Rensselaeria* cf. *R. aequiradiata, Eatonia peculiaris, Platyceras* cf. *P. gebhardi,* and *Phacops logani* in the lower 15 feet	31
New Scotland formation (Healing Springs sandstone member): Massive gray arenaceous limestone to calcareous sandstone	15½
Coeymans limestone:	
Very massive crinoidal limestone, abounding in large crinoid rings. *Meristella arcuata* (aa) at top	19½
Massive gray arenaceous limestone	5½
Keyser limestone:	
Upper limestone member:	
Massive limestone, with many *Stromatopora* and *Favosites*	15
Medium-bedded, somewhat nodular limestone, with some interbedded chert	10

Helderberg group—Continued.
Keyser limestone—Continued.

	Feet
Upper limestone member—Continued.	
Arenaceous limestone. *Cladopora rectilineata* (aa)	3
Massive dark-blue limestone, with a few chert lenses	4½
Clifton Forge sandstone member:	
Massive calcareous sandstone. Some interbedded arenaceous shale in lower 14 feet	37
Sandy limestone, with chaetetoid bryozoan zone at middle	1
Very massive blue limestone, weathering gray. This and the underlying unit might be placed in the lower limestone member of the Keyser	10
Massive calcareous sandstone to arenaceous limestone	21
Tonoloway limestone: Laminated blue limestone.	

20. *Section at Hollybrook, Va.*

[This section is fairly well exposed in a washway on the hillside north of the road along No Business Creek, about 720 feet northeast of the intersection with the Kimberling road, at Hollybrook, Va. The exposures around Hollybrook were cited by M. R. Campbell in his description of the Giles formation, in the Pocahontas folio. The sandstone at the top of the Giles is not seen in this particular section, as it has been eroded from the hilltop. An exposure showing this sandstone, practically in contact with the overlying Genesee shale, can be seen along the road to Bland about a quarter of a mile east of the crossroads. The sandstone is greenish, carries a post-Oriskany (apparently Onondaga) fauna, and is evidently equivalent to the greenish sandstone at the top of the Giles at Rocky Gap, Cove Mountain, and Tumbling Creek, near Saltville. The Hollybrook, Rocky Gap, Cove Mountain, and Tumbling Creek section were measured by C. K. and F. M. Swartz in 1927, in connection with a study of the underlying Silurian.]

	Feet
Giles formation:	
Mostly concealed at top of hill; fragments of sandy chert and some weathered sandstone strew the surface. (See Rocky Gap section.)	
Thick-bedded to massive quartz sandstone conglomerate, forming a prominent ledge about 35 feet below top of hill. The many cavities formed by the weathering out of bulbs of *Aspidocrinus caroli* are a very striking feature. There are also poor molds of a *Spirifer* that suggests *S. cyclopterus* of the Helderberg group. The stratigraphic evidence afforded by this section and those at Rocky Gap and near Saltville suggests that this conglomeratic sandstone is of New Scotland or Becraft age	22
Concealed. Probably calcareous sandstone for the most part. If any portion of the Keyser limestone is present here it would lie in the lower part of this interval	86
Laminated blue limestone (Tonoloway lithology and fauna). This limestone was included in the Giles formation by Campbell. In exposures on Ding Run, in and near Burton's quarry, about 3 miles northeast of Hollybrook, this limestone carries *Leperditia* sp. (probably *L. alta*) and *Hindella congregata*	68

21. *Section at Rocky Gap, Va.*

[Section at east end of gap 2 miles east of Rocky Gap, about 15 miles south of Narrows, Va.]

	Feet
Genesee shale: Black shale seen along road to Bland.	
Giles formation:	
Greenish sandstone, same as the sandstone at top of the Giles at Hollybrook and Cove Mountain	5+

Giles formation—Continued.	Feet
Bedded black chert, forming riffles in the creek. The contact with the overlying sandstone was not seen, but the sandstone was projected along the strike from the road to the creek for a rather unsatisfactory measurement of the interval. This black chert, although somewhat suggestive of the Shriver chert of the sections farther north, is most likely equivalent to the chert that carries *Spirifer arenosus* var. *planicostatus* in the Saltville region. Upper part concealed	65+
Calcareous sandstone, exposed in part in bed of creek and seen also along the State road to Bland. *Aspidocrinus caroli* (aa) in upper 20 feet; *Spirifer cyclopterus?* at about middle; a large *Schuchertella* near base	62
Concealed. A few weathered chert masses were seen at the top of this interval, along the road to Bland; these contained an abundance of casts of a *Meristella* which is apparently *M. arcuata*. The zone is suggestive of that at the top of the Coeymans limestone in the Clifton Forge area, although correlation with that zone is not assured. It seems probable that much of this concealed interval is occupied by the Tonoloway limestone	40
Blue limestone (Tonoloway lithology and fauna) mostly concealed; the most definite exposures were seen in diggings for the abutments of the highway bridge.	

22. *Section at Cove Mountain, near Wytheville, Va.*

[The top of the Giles formation is exposed along the old road from Wytheville to Bland at the gap through Cove Mountain, about 4 miles northwest of Wytheville. The mountain is made by the Clinch sandstone, which dips to the east and is overlain by the black chert and the upper greenish sandstone of the Giles formation. The intervening beds are absent, probably owing to faulting rather than to an unconformity. The best exposures are on the hillside south of the creek, just opposite an old mill]

	Feet
Genesee shale: Black shale, seen along creek bank.	
Giles formation:	
Greenish, very fossiliferous sandstone. *Amphigenia?* sp., *Athyris spiriferoides*, *Spirifer* cf. *S. manni*, *Reticularia fimbriata*, etc. Probably Onondaga	10±
Black bedded chert; a little interbedded shale. No fossils seen. Probably equivalent to the chert of the Saltville section	25
Fault (?). Lower part of Giles (Tonoloway, etc.), and Clinton absent.	
Clinch sandstone: Massive white to gray sandstone.	

23. *Section along Tumbling Creek, near Saltville, Va.*

[The strata above and below the Silurian-Devonian boundary are exposed along Tumbling Creek, about 6 miles southwest of Saltville, Va. (The United States Geological Survey map shows a second Tumbling Creek north of Saltville.) The fossils collected from this section have not yet been thoroughly studied]

	Feet
Genesee shale: Black shale exposed east of abandoned mill.	
Giles formation:	
Weathered sandstone, largely concealed	14
Greenish calcareous sandstone, with a few lenses of chert. *Chonetes coronatus?*, *Spirifer* cf. *S. manni*, *Anoplotheca acutiplicata*, etc. Probably of Onondaga age	22
Red sandstone with some limonite mottling	2
Bedded black chert. Some calcareous matrix? *Spirifer arenosus* var. *planicostatus* and *Diaphorastoma ventricosum* at middle; *Ambocoelia umbonata* a few feet lower. Of late Oriskany age	21
Sandstone, with red staining at top	9
Massive blue limestone	4
Greenish calcareous sandstone	12
Red sandstone with limonite mottling	2½
Greenish-brown sandstone, with some chert	21
Blue limestone; a few minor chert lenses. *Meristella* sp.; corals	4
Greenish-brown sandstone, with some lenses of black chert; upper 8 inches red, with some mottling. *Spirifer arenosus*, rather abundant at top. Of Oriskany age	4½
Thin-bedded yellowish-brown sandstone	3
Greenish-brown sandstone, with 3 inches of cherty limestone, in somewhat irregular lenses, at base. *Rhipidomella* cf. *R. oblata*, *Dalmanella perelegans?*, etc. This and the overlying unit appear to be of New Scotland or Becraft age	4½
Hard blue limestone (Tonoloway lithology and fauna) weathering platy in part.	

PLATES 6–9

PLATE 6

KEYSER FAUNA

CHONETES JERSEYENSIS ZONE

1, 2. *Chonetes jerseyensis* Weller (p. 54).

1. Nearly complete ventral valve, showing spines of cardinal margin. A very characteristic specimen. Petersburg, W. Va., about 50 feet above base of Keyser limestone.
2. Incomplete ventral valve, showing somewhat stouter ribs. Little Mountain section, west of Monterey, Va., *Gypidula* subzone, 113 feet above base of Keyser limestone.

3–5. *Camarotoechia litchfieldensis* (Schuchert) (p. 54) and *Whitfieldella minuta* Maynard (p. 57).

3. Slab showing ventral valves of these two associated species. Petersburg, W. Va., *Whitfieldella minuta* subzone, base of Keyser limestone.
4. Ventral valve of *Whitfieldella minuta.*
5. Ventral valve of *Camarotoechia litchfieldensis.*

6. *Cyphotrypa corrugata* (Weller) (p. 53). Vertical section of a large zoarium. West of Warm Springs, Va., *Cyphotrypa corrugata* subzone, 20 feet above base of Keyser limestone.

7. *Stropheodonta bipartita* (Hall) (p. 53). Exterior of ventral valve. Warm Springs, Va., *Cyphotrypa corrugata* subzone, 20 feet above base of Keyser limestone. Specimen partly exfoliated.

8, 9. *Uncinulus convexorus* Maynard (p. 55).

8. Dorsal view. Franklin, W. Va., *Stenochisma deckerensis*, etc., subzone, 20 feet above base of Keyser limestone.
9. Side view of somewhat imperfect specimen, showing detail of interlocking of serrated margins of the two valves. Near Warm Springs, Va., *Stenochisma deckerensis*, etc., subzone, 30 feet above base of Keyser limestone.

10–12. *Stenochisma deckerensis* (Weller) (p. 54).

10. Dorsal view showing the relatively great width of this species. Petersburg, W. Va., *Stenochisma deckerensis*, etc., subzone, about 45 feet above base of Keyser limestone.
11. Side view of same specimen showing the high dorsal elevation.
12. Dorsal view of somewhat smaller specimen. Clifton Forge, Va., *Stenochisma deckerensis*, etc., subzone, 10 feet above base of Keyser limestone.

13–15. *Nucleospira swartzi* Maynard (p. 57).

13. Ventral view. Big Mountain section, south of Petersburg, W. Va., *Stenochisma deckerensis*, etc., subzone, 40 feet above base of Keyser limestone.
14. Side view of same specimen.
15. Dorsal view. Clifton Forge, Va., *Stenochisma deckerensis*, etc., subzone, 10 feet above base of Keyser limestone.

16, 17. *Spirifer modestus* Hall (p. 56).

16. Ventral view. Strait Creek, north of Monterey, Va., *Stenochisma deckerensis*, etc., subzone, 30 feet above base of Keyser limestone.
17. Dorsal view of same specimen.

18, 19. *Spirifer modestus* var. *plicatus* Maynard (p. 56).

18. Dorsal view, showing the low plications from which this variety derives its name. Strait Creek north of Monterey, Va., *Stenochisma deckerensis*, etc., subzone, 15 feet above base of Keyser limestone.
19. Side view of same specimen.

20–23. *Gypidula coeymanensis* var. *similis* F. M. Swartz, n. var. (p. 54).

20, 21. Exterior and side views of type specimen of ventral valve, showing large size, somewhat inflated beak, and broad, rather indistinct plications. Little Mountain, near Monterey, Va., *Gypidula* subzone, 113 feet above base of Keyser limestone.

22, 23. Weathered interiors of ventral valves; same place and horizon.

24. *Gypidula coeymanensis* var. *prognostica* Maynard (p. 54). Ventral valve. Petersburg, W. Va., *Gypidula* subzone, 122 feet above base of Keyser limestone. The material which I am including under the variety *similis* is undoubtedly consanguineous with the variety *prognostica* of the more northern sections.

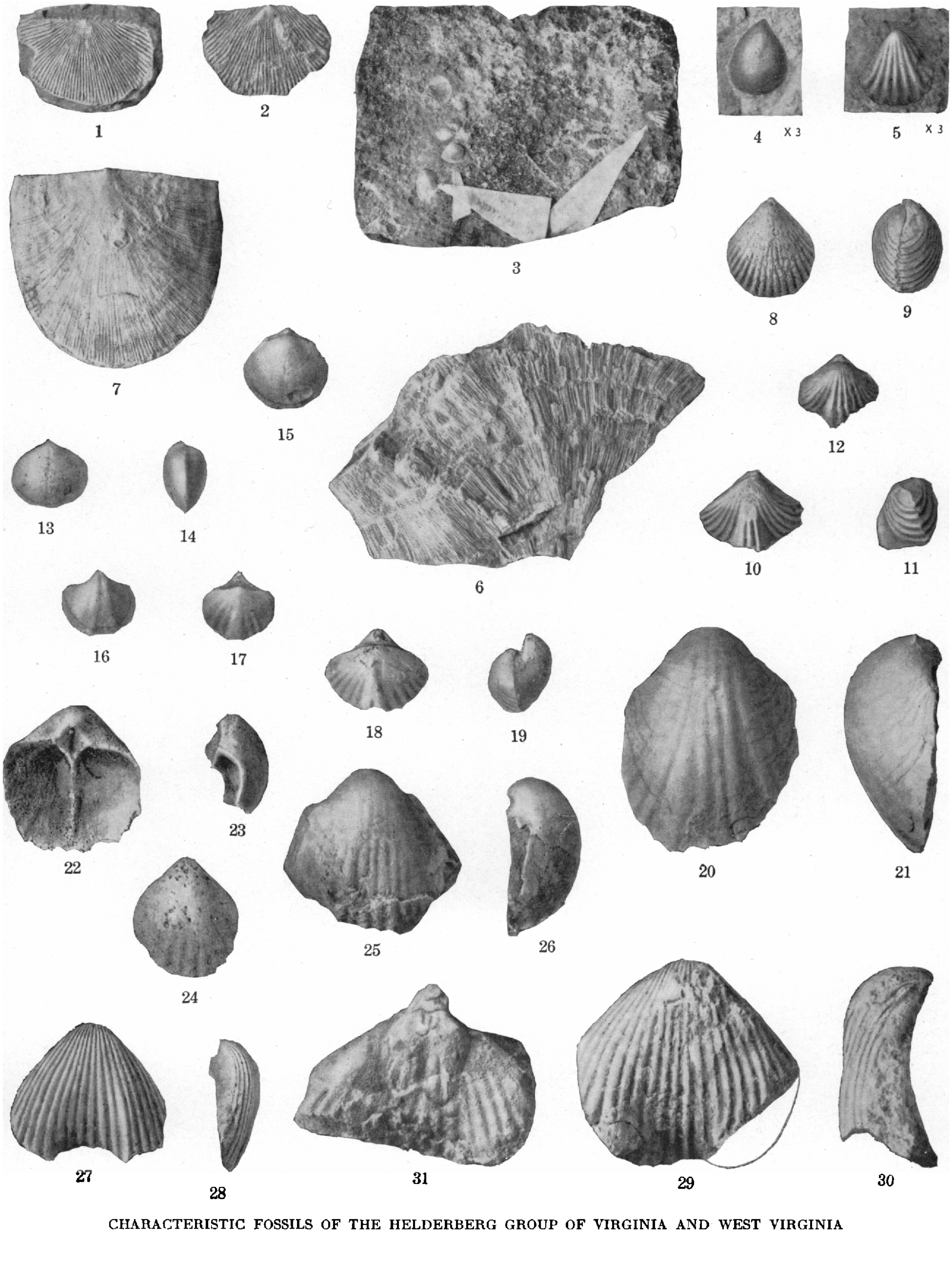
25, 26. *Gypidula coeymanensis* Schuchert (p. 54). Exterior and ventral views. Coeymans limestone, Nearpass, N. J. Show the close similarity of the Little Mountain middle Keyser material to typical specimens of the species.

27, 28. *Camarotoechia gigantea* Maynard (p. 54).

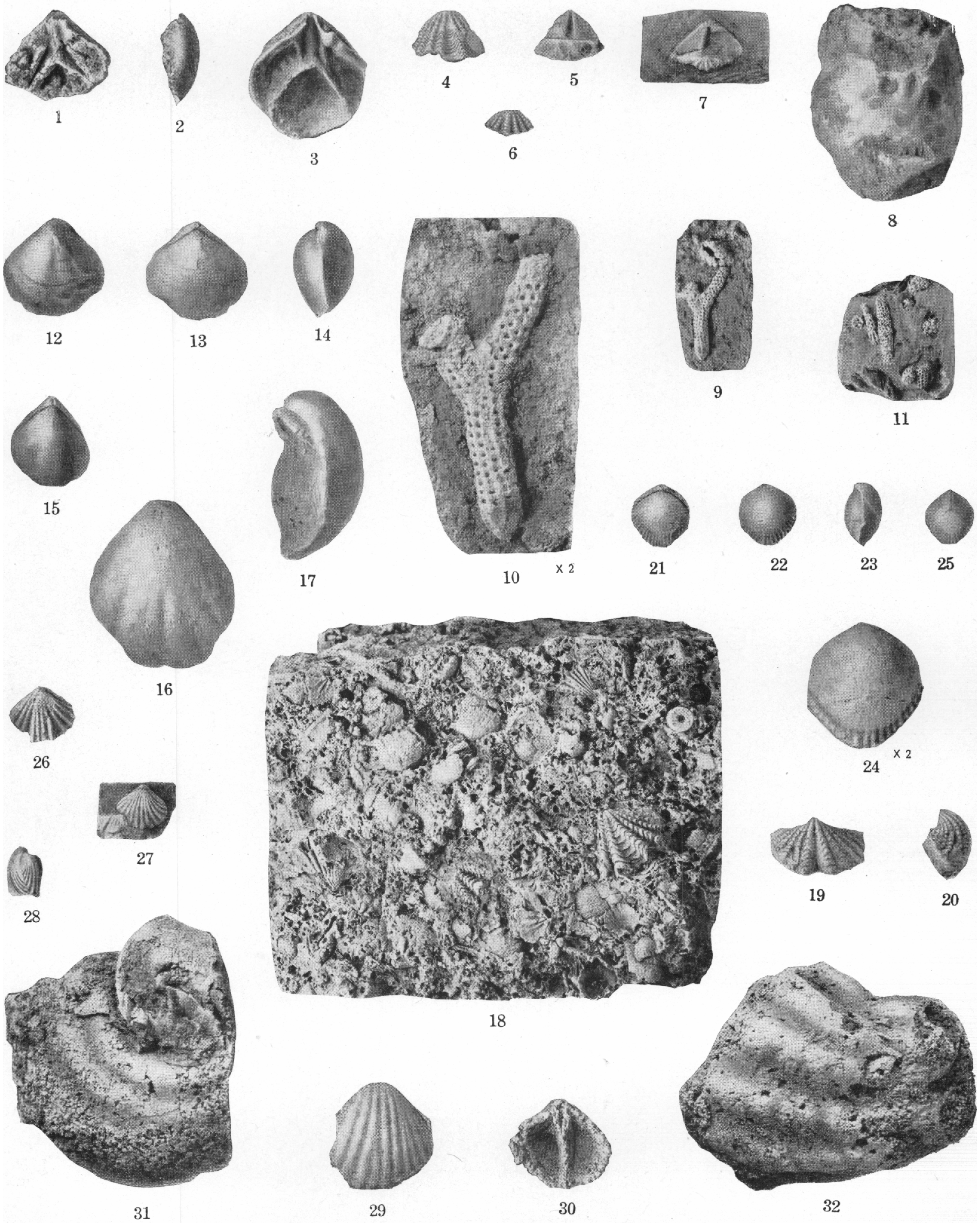
27. Characteristic ventral valve of this abundant species. Warm Springs, Va., *Camarotoechia gigantea* subzone, 105 feet above base of Keyser limestone.
28. Side view of same specimen.

29–31. *Camarotoechia gigantea* var. *gigas* F. M. Swartz, n. var. (p. 55).

29. Ventral valve, showing large size and broad, deep sinus. Monterey, Va., *Camarotoechia gigantea* subzone, 112 feet above base of Keyser limestone.
30. Side view of same specimen, showing closely incurved beak.
31. Internal cast of ventral valve showing casts of muscle scars. Same locality and horizon.



CHARACTERISTIC FOSSILS OF THE HELDERBERG GROUP OF VIRGINIA AND WEST VIRGINIA



CHARACTERISTIC FOSSILS OF THE HELDERBERG GROUP OF VIRGINIA AND WEST VIRGINIA

PLATE 7

KEYSER FAUNA

CHONETES JERSEYENSIS ZONE

1–3. *Merista typa* (Hall) (p. 58).
- 1. Interior of ventral valve, showing characteristic "shoe lifter" arched plate. East of McDowell, Va., *Merista typa* subzone, 64 feet above base of Keyser limestone.
- 2. Side view of same specimen.
- 3. Weathered interior of ventral valve. West of Wardensville, W. Va., *Merista typa* subzone, 95 feet above base of Keyser limestone.

FAVOSITES HELDERBERGIAE VAR. PRAECEDENS ZONE

4–7. *Cyrtina dalmani* (Hall) (p. 57).
- 4. Imperfect ventral valve, showing elevated plications and strong lamellae. Monterey, Va., 130 feet above base of Keyser limestone.
- 5. Cardinal view of same specimen.
- 6, 7. Dorsal and cardinal views. Big Mountain, W. Va., 139 feet above base of Keyser limestone.

8. *Mariacrinus* sp. (p. 52). Side view of imperfect calyx. Petersburg, W. Va., 218 feet above base of Keyser limestone.

9–11. *Cladopora rectilineata* Simpson (p. 52).
- 9. A typical corallum. Bolar, south of Monterey, Va., 114 feet above base of Keyser limestone.
- 10. The same specimen enlarged.
- 11. Several specimens from upper part of Keyser limestone, Island Ford Bridge, south of Clifton Forge, Va.

12–15. *Meristella praenuntia* Schuchert (p. 57).
- 12–14. Ventral, dorsal, and side views of a transverse specimen. McDowell, W. Va., *Meristella praenuntia* subzone, 136 feet above base of Keyser limestone.
- 15. Dorsal view of a more elongate specimen. Same locality and horizon.

16, 17. *Meristella nasutaformis* F. M. Swartz, n. sp. (p. 57). Ventral and side views of ventral valve. Big Mountain, W. Va., 113 feet above base of Keyser limestone.

18–20. *Spirifer perlamellosus* var. *praenuntius* F. M. Swartz, n. var. (p. 56).
- 18. Slab showing ventral valve and small portion of dorsal valve, associated with *Dalmanella concinna* and *Camarotoechia altiplicata*. Little Mountain, west of Monterey, Va., upper part of Keyser limestone.
- 19, 20. Ventral and side views. Big Mountain, W. Va., 181 feet above base of Keyser limestone.

21–25. *Rensselaeria mutabilis* (Hall) (p. 55).
- 21–23. Dorsal, ventral, and side views of a large specimen. Little Mountain, west of Monterey, Va., *Rensselaeria mutabilis* subzone, 195 feet above base of Keyser limestone.
- 24. Enlarged ventral view of same specimen, showing ornamentation.
- 25. Ventral view of a smaller specimen. Franklin, W. Va., *Rensselaeria mutabilis* subzone, 200 feet above base of Keyser limestone.

26–28. *Camarotoechia* cf. *C. altiplicata* (Hall) (p. 55).
- 26. Dorsal view. Little Mountain, west of Monterey, Va., *Camarotoechia* cf. *C. altiplicata* subzone, 195 feet above base of Keyser limestone. The Keyser material here referred to *C. altiplicata* is smaller and somewhat less gibbous than the New York types, but more similar to the specimens from the Maryland New Scotland.
- 27. Dorsal view, showing ventral beak. Franklin, W. Va., *Camarotoechia* cf. *C. altiplicata* subzone, 200 feet above base of Keyser limestone.
- 28. Side view of imperfect specimen, Franklin, W. Va., *Camarotoechia* cf. *C. altiplicata* subzone, 195 feet above base of Keyser limestone.

COEYMANS FAUNA

29, 30. *Gypidula coeymanensis* Schuchert (p. 54).
- 29. Exterior of ventral valve. Monterey, Va., Coeymans limestone.
- 30. Weathered interior ventral valve. Petersburg, W. Va., Coeymans limestone. Note small size and stronger plications of the Coeymans material as compared to the Gypidulas from the Keyser limestone.

31, 32. *Platyceras multiplicatum* F. M. Swartz, n. sp. (p. 58).
- 31. Top view of imperfect specimen, showing general character of the shell, McDowell, W. Va., Coeymans limestone.
- 32. Fragment of outer whorl. Hollybrook, Va., conglomeratic sandstone of Giles formation.

PLATE 8

COEYMANS FAUNA

1–4. *Meristella arcuata* (Hall) (p. 57).
 1, 2. Ventral and side views of a gibbous specimen. Monterey, Va., New Scotland limestone.
 3. Ventral view. Near Clifton Forge, Va., *Meristella arcuata* zone, top of Coeymans limestone.
 4. Interior of ventral valve, showing characteristic muscle scars. Bells Valley, Va., *Meristella arcuata* zone, top of Coeymans limestone.

5–7. *Meristella arcuata* var. *gigas* F. M. Swartz, n. var. (p. 58).
 5, 6. Ventral and side view of ventral valve. Monterey, Va., *Meristella arcuata* zone, top of Coeymans limestone.
 7. Ventral valve, Bolar, Va., *Meristella arcuata* zone, top of Coeymans limestone.

8. *Meristella* cf. *M. symmetrica* Schuchert (p. 58). Ventral valve, Monterey, Va., *Meristella arcuata* zone, top of Coeymans limestone.

NEW SCOTLAND FAUNA

9. *Spirifer macropleurus* (Conrad) (p. 56). Dorsal view of weathered specimen. Monterey, Va., New Scotland limestone.

10, 11. *Dalmanella perelegans* (Hall) (p. 53). Ventral and side views of a specimen whose dorsal sinus is that of *D. perelegans*, but whose ventral cardinal area approaches that of *D. eminens* in height. Monterey, Va., New Scotland limestone.

12. *Eatonia medialis* (Vanuxem) (p. 55). Dorsal view of characteristic specimen. Monterey, Va., New Scotland limestone.

13, 14. *Streptelasma strictum* Hall (p. 52).
 13. Side view of specimen showing the strong exterior septal ridges and the growth constrictions. The apex has been lost. Monterey, Va., New Scotland limestone.
 14. Weathered transverse section, showing the septa. Same locality and horizon.

BECRAFT FAUNA

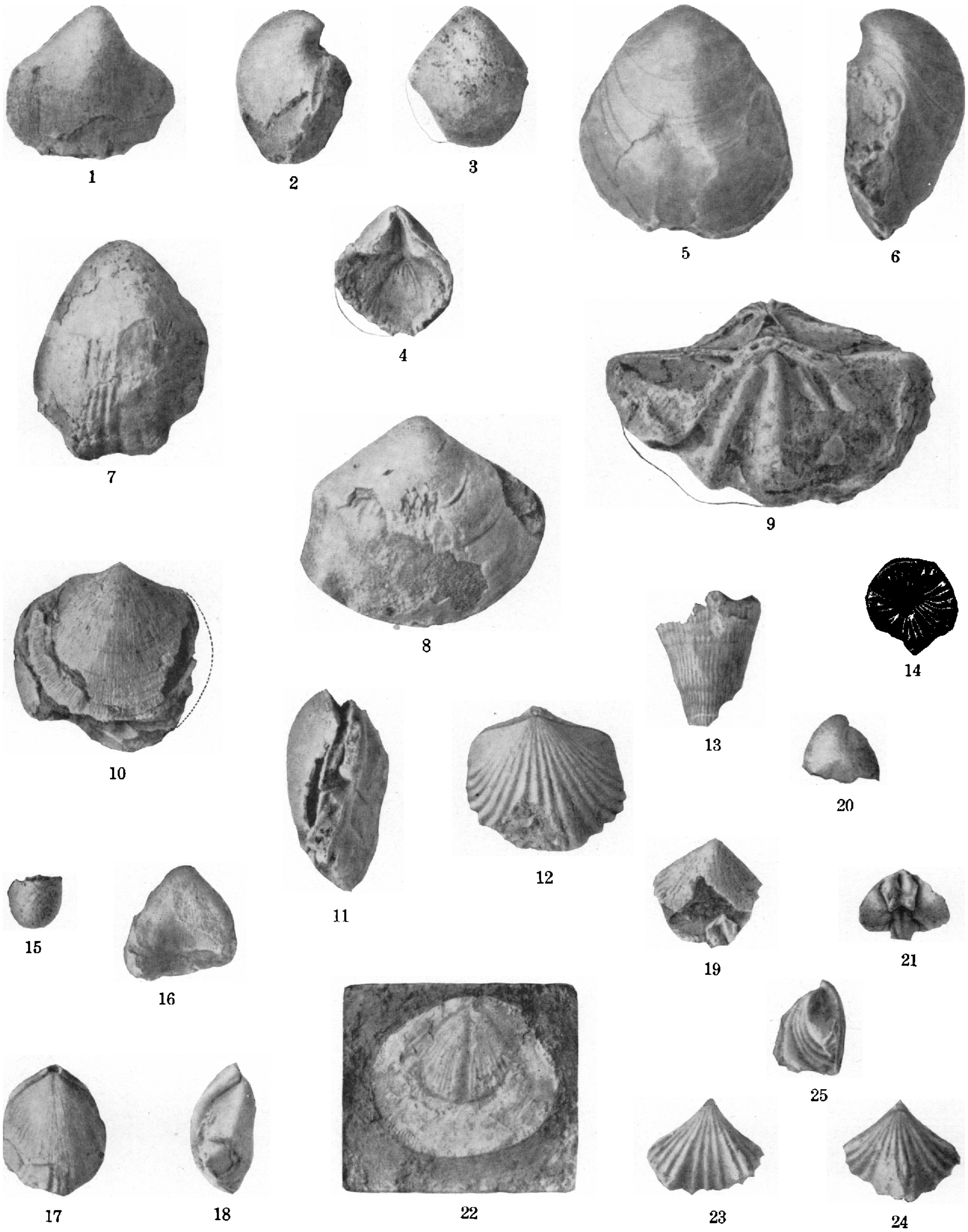
15. *Edriocrinus pocilliformis* Hall (p. 52). Side view of weathered calyx. Gala, Va., base of Becraft limestone.

16–18. *Eatonia peculiaris* (Conrad) (p. 55).
 16. Ventral view. Gala, Va., 10 feet below top of Becraft limestone.
 17, 18. Dorsal and side views. Gala, Va., lower part of Becraft limestone.

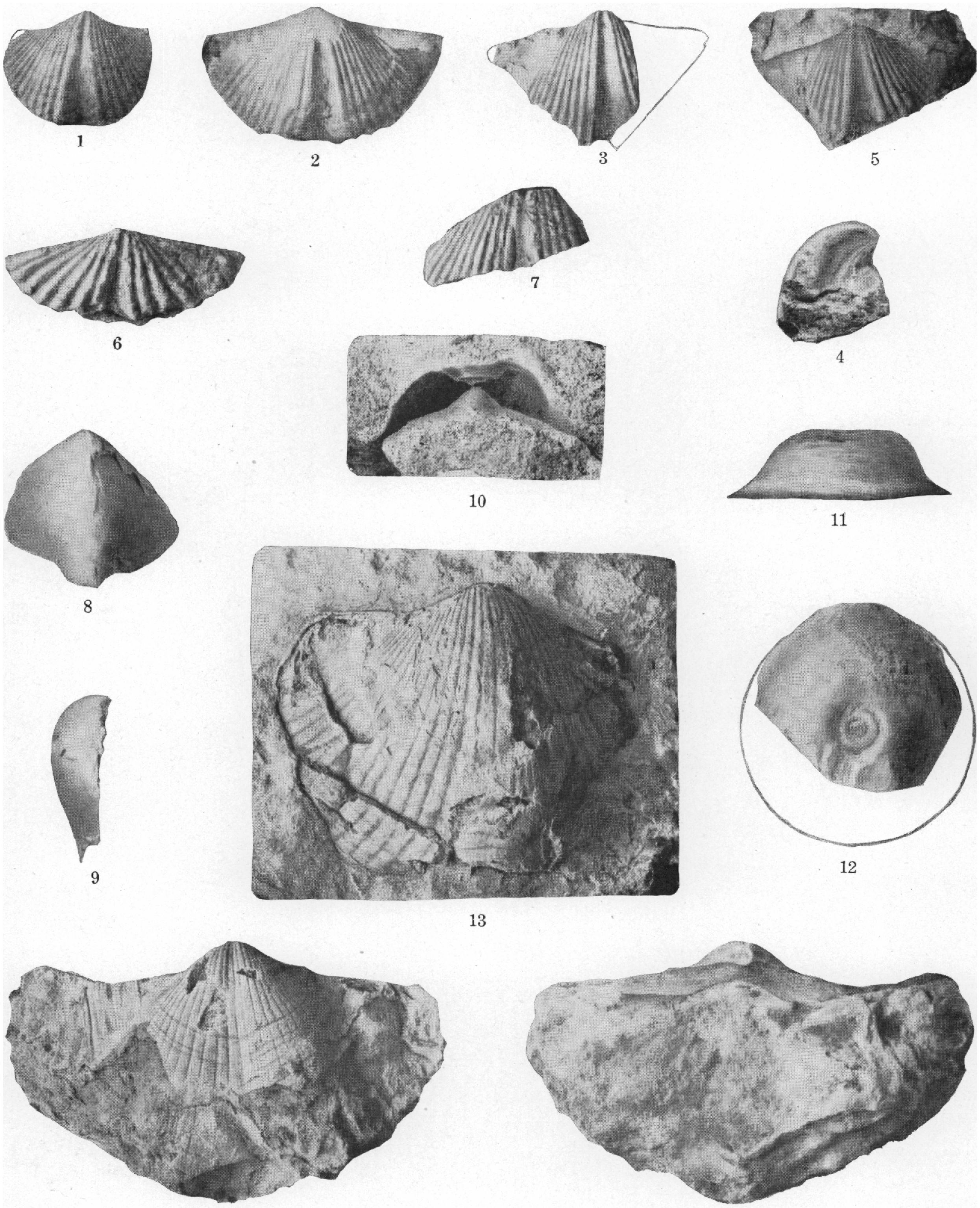
19–21. *Rensselaeria* cf. *R. aequiradiata* (Conrad) (p. 55).
 19. Posterior portion of ventral valve. Gala, Va., basal part of Becraft limestone.
 20. Side view of posterior part of shell; ventral beak somewhat imperfect. Bells Valley, Va., base of Becraft limestone.
 21. Weathered interior of posterior part of brachial valve, showing crural plates. Gala, Va., basal part of Becraft limestone.

22. *Rhipidomella assimilis* (Hall) (p. 53). Exfoliated ventral valve, showing casts of muscle scars. Gala, Va., upper part of Becraft limestone.

23–25. *Rhynchotrema cumberlandicum* Rowe (p. 54). Ventral, dorsal, and side views. Gala, Va., upper part of Becraft limestone.



CHARACTERISTIC FOSSILS OF THE HELDERBERG GROUP OF VIRGINIA AND WEST VIRGINIA



CHARACTERISTIC FOSSILS OF THE HELDERBERG GROUP OF VIRGINIA AND WEST VIRGINIA

PLATE 9

BECRAFT FAUNA

1. *Spirifer concinnus* Hall (p. 56). Ventral valve. Gala, Va., upper part of Becraft limestone.

2–5. *Spirifer concinnus* var. *progradius* F. M. Swartz, n. var. (p. 56).

2. Ventral valve, showing large size, rather narrow ribs, and apparent alation of this progressive variety. Gala, Va., 10 feet below top of Becraft limestone.
3. Exterior of fragmentary ventral valve. Same locality and horizon.
4. Longitudinal section of specimen shown in Figure 3, showing the rather strong dental plate and indicating the moderately high cardinal area.
5. Smaller specimen with a feeble plication on each side of the ventral sinus, near the front margin. This feature, which is also found in the specimen illustrated in Figures 3 and 4, suggests *Spirifer proavitus* Schuchert.

6. *Spirifer angularis* Schuchert (p. 56). Ventral valve, Gala, Va., 6 feet below top of Becraft limestone.

7. *Spirifer* sp. Fragmental ventral valve. Clifton Forge, Va., upper part of Becraft limestone. The narrow sinus and the fairly prominent but narrow plications suggest *S. cumberlandiae* Hall, of the Oriskany group.

8, 9. *Meristella lata* (Hall) (p. 58). Dorsal and side views of dorsal valve. Dry Run, east of Warm Springs, Va., middle part of Becraft limestone.

GILES FAUNA

[See text for relations of Giles formation to Helderberg group]

10–12. *Aspidocrinus caroli* F. M. Swartz, n. sp. (p. 52). (Named for Dr. Charles K. Swartz, with whom the sections in southwestern Virginia were visited in connection with work on the Silurian.)

10. Vertical section of mold of the cup, as ordinarily seen in the rock. Hollybrook, Va., at top of sandstones of middle part of Giles formation.

11, 12. Side and top views of wax casting from same specimen.

13–15. *Spirifer arenosus* var. *planicostatus* F. M. Swartz, n. var. (p. 56).

13. Dorsal view of large specimen. Tumbling Creek, near Saltville, Va., dark chert beds of upper part of Giles formation.

14, 15. Exterior and cardinal views of a smaller ventral valve; margins imperfect. Same locality and horizon.

PETROGRAPHY OF THE PIOCHE DISTRICT, LINCOLN COUNTY, NEVADA

By Joseph L. Gillson

LOCATION AND GENERAL GEOLOGY

The geology of a section of the Great Basin in the vicinity of Pioche, Nev., has been studied by members of the United States Geological Survey, and a brief summary of the results, by Westgate and Knopf,[1] has already been published. The writer had the pleasure of serving in the district with Professor Westgate during two field seasons.

The part of the area here described lies in the southern part of the Bristol Range quadrangle and the northern part of the Highland quadrangle. (See fig. 11.) Extending through these quadrangles from north to south is a high range of faulted but gently dipping Cambrian limestone, called in the north the Bristol Range and south of a pass known as Stampede Gap the Highland Range. At the north end of the Bristol Range occurs an overthrust block of Devonian sediments, which is known also in a small range 4 miles to the west, where it overlies Miocene (?) lavas. In the vicinity of Blind Mountain, in the southern part of the Bristol Range quadrangle, and on the western slopes of the Bristol Range, a down-faulted section of this Devonian overthrust mass occurs, there involved with the Miocene (?) lavas and intruded by a quartz monzonite mass, which has caused extensive contact metamorphism not only of the overthrust mass but of the Cambrian rocks adjacent to it.

This quartz monzonite is the only large intrusive mass exposed in the Pioche district, and its outcrops are confined within an area of about 2 square miles, although the desert wash filling the valley west of the Bristol Range probably buries a considerable mass of it. The exposures are considered to represent an upper cupola of a large batholith which underlies the whole Pioche area and from which the ore solutions emanated. The large size of the batholith is indicated by widespread marmarization of certain beds of the limestone and the wide distribution of the ore deposits, of zones of intense contact metamorphism, and of lamprophyre dikes.

The exposed igneous rock has two principal facies, but this study has shown that they are related. Both illustrate the type of differentiation described by Fenner,[2] in which the character of the magma during crystallization was changed because of the upward passage of gases. The already solidified border phase of the rock was intensely endomorphosed by these emanations, and throughout the body of the rock many mineral changes took place after consolidation.

THE QUARTZ MONZONITE

DISTRIBUTION AND GENERAL CHARACTER

As shown on Figure 11, a granitoid quartz monzonite crops out in two belts which are separated by a deep wash-filled valley, here called Kiln Valley. A porphyritic facies, in which the groundmass ranges from fine granitoid to aphanitic, forms two masses on Blind Mountain. Many narrow apophyses also there cut the volcanic rocks and sediments, and a large apophysis extends southeastward for over a mile. The smaller dikes are not shown on Figure 11.

The granitoid rock is light gray to pink, medium grained, and slightly porphyritic, and contains quartz, feldspar, and biotite, with subordinate amounts of other ferromagnesian minerals. The largest crystals in the rock are about 3 millimeters in diameter.

The coarser masses of the porphyritic rock contain white plagioclase phenocrysts as much as 2 millimeters in diameter and less abundant black phenocrysts of like size, set in a groundmass made dark by the abundance of minute grains of the ferromagnesian minerals. Quartz is confined to the groundmass and can be recognized only with difficulty. The finer-grained apophyses differ only in that the phenocrysts are smaller and the groundmass, being more dense, is darker.

At three places the granitoid rock is so intensely endomorphosed and differs so much in appearance from the main body of the rock that it was not correctly identified during the field study. It is a dark-colored medium to fine grained rock containing glistening phenocrysts of plagioclase and in some places of quartz, in a groundmass in which biotite is the only mineral present in crystals large enough to recognize. The identification of the rock in the field was made difficult by its close similarity to metamorphosed lavas and to the cordierite hornfels formed from the metamorphism of the Lower Cambrian shale.

The granitoid facies is cut by aplite dikes and veins and by narrow stringers of pegmatite. The aplite dikes are light-colored fine-grained rocks in which dark silicates are very subordinate. Large masses of

[1] Westgate, L. G., and Knopf, Adolph, Geology of Pioche, Nev., and vicinity: Am. Inst. Min. and Met. Eng., Bull. 1647-I, 21 pp., 1927.

[2] Fenner, C. N., The Katmai magmatic province: Jour. Geology, vol. 34, pp. 743-744, 1926.

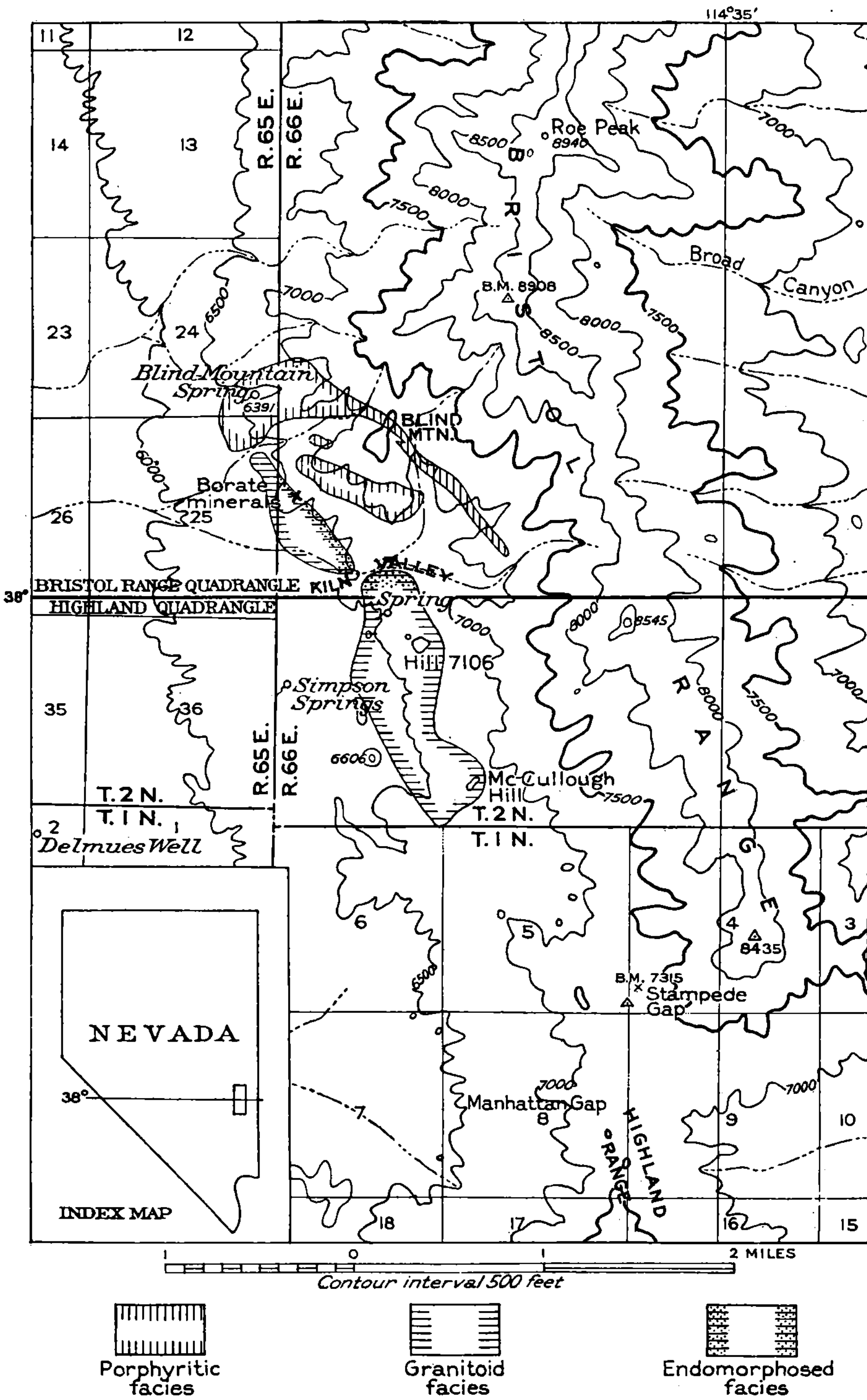


FIGURE 11.—Outline map of the exposures of the quartz monzonite of Blind Mountain, at the west base of the Bristol Range, Pioche district, Nev., showing the distribution of the two facies of the quartz monzonite and the endomorphosed zones

true pegmatite do not occur, but several small miarolitic cavities were found containing attractive crystals of tourmaline, feldspar, and magnetite. In the endomorphosed quartz monzonite narrow pegmatite stringers are abundant, and in several of these large druses occur in which are euhedral crystals of quartz and microcline as large as 2 centimeters in diameter.

PETROGRAPHY

GRANITOID FACIES

The rock away from its margins is subporphyritic and hypidiomorphic in texture and contains the following minerals in about the proportion shown in the table:

Average mineral composition, by weight, of the main mass of the granitoid quartz monzonite

	1	2	3
Quartz	15	16	25
Microcline	25	22	32
Plagioclase	46. 5	49	35
Biotite	9	6	4
Hornblende	2	3	2
Augite	Rare.	2	0
Accessories	2. 5	2	2

1. South slope of McCullough Hill.
2. North slope of hill 7106.
3. 0.2 mile north of latitude 38° and 2.07 miles west of longitude 114° 35′.

The augite occurs as cores of the hornblende grains. The accessories are apatite, magnetite, zircon, allanite, and titanite.

The rock near its border, where almost free from endomorphic effects, is similar to the normal type of the main mass but is richer in dark minerals, especially in pyroxene. The mineral composition of the border rock is as follows:

Average mineral composition, by weight, of the border facies of the granitoid quartz monzonite

	1	2	3
Quartz	15	12	12
Microcline	24	18	24
Plagioclase	40	52	39
Biotite	7	7	9
Hornblende and pyroxene	11	7	13
Olivine	0	Rare.	0
Accessories	3	3	3

1. 1.1 miles south of latitude 38° and 1.32 miles west of longitude 114° 35′.
2. 0.09 mile south of latitude 38° and 1.66 miles west of longitude 114° 35′.
3. 0.42 mile north of latitude 38° and 2.17 miles west of longitude 114° 35′.

The border facies is interpreted as being a chilled margin, representing the composition of the rock magma earlier in the process of differentiation than that shown by the normal type. Hypersthene is almost as abundant as augite in some specimens. As some of the potash feldspar is deuteric, the proportion of it present is not an indication of the composition of the magma during consolidation.

PORPHYRITIC FACIES

The porphyritic facies contains zoned plagioclase phenocrysts, the cores of which are labradorite, set in a groundmass of acidic plagioclase and very interstitial quartz and microcline. Green hornblende and brown biotite occur both as phenocrysts and in the groundmass. Pyroxene is present as cores in a small proportion of the hornblende grains. Biotite is less abundant in the narrow dikes than in the larger masses. The feldspar phenocrysts are all more or less replaced by potash feldspar, by a process here called orthoclasization and described in detail on a later page. Other alteration by emanations after consolidation is conspicuous in most specimens studied.

Two apophyses of the porphyritic facies were found that differ from the more common type. They cut the Devonian sediments on the west spur of Blind Mountain and contain phenocrysts of quartz and potash feldspar. Biotite was the primary ferromagnesian mineral, although deuteric epidote, chlorite, and another finely divided brown but not pleochroic micaceous mineral occur.

CHEMICAL COMPOSITION

Only the granitoid facies of the quartz monzonite has been analyzed. The results are as follows:

Analysis and norm of the granitoid facies of the quartz monzonite of Blind Mountain

[J. G. Fairchild, analyst]

Analysis		Norm	
SiO_2	65. 84	Quartz	20. 64
Al_2O_3	15. 29	Orthoclase	26. 69
Fe_2O_3	2. 48	Albite	29. 34
FeO	2. 26	Anorthite	12. 79
MgO	1. 06	Diopside	1. 33
CaO	3. 05	Hypersthene	2. 60
Na_2O	3. 47	Magnetite	3. 71
K_2O	4. 53	Ilmenite	1. 98
H_2O-	. 01	Apatite	. 34
H_2O+	. 67		
TiO_2	1. 00		
P_2O_5	. 19		
MnO	. 09		
	99. 94		

The norm places the rock in Class II, order 4, rang 2, subrang 3 of the quantitative classification.

DIFFERENTIATION OF THE MAGMA

The border facies of the quartz monzonite, the mineral composition of which has already been given, is inferred to be the quenched product of the magma in a stage of differentiation prior to that from which the main mass finally solidified. Augite and hypersthene are prominent constituents of this facies, and in one specimen crystals of olivine were found. Much of the augite is surrounded by hornblende, indicating that with the progress of crystallization an increase in volatile material took place, making the hydrous mineral the stable form. The amount of potash feldspar present in the border facies is lower than in the normal rock, but in both much of it was formed after consolidation, and hence the quantity is not an indication of the character of the magma.

Both in the border facies and in the main mass an earlier chapter in the differentiation than that revealed by the general composition of the border facies is indicated by corroded cores of labradorite in some of the plagioclase crystals. In many grains scarcely more than a ghost of the corroded core remains. The labradorite is made conspicuous by the presence of tiny black rod-shaped microlites, of general definite arrangement, of the kind that are characteristic of the labradorite of such rocks, for example, as the Duluth gabbro, the Marcy anorthosite of the Adirondacks, the gabbros of the Adirondacks, and the anorthosite of Labrador. These microlites are significant for two reasons. First, the identity of the intensely endomorphosed quartz monzonite was made certain partly by the presence in it of plagioclase crystals containing just such corroded cores of labradorite, with microlites. Second, these microlites are characteristic of the feldspar of basic igneous rocks. An idea of the shape of the corroded cores of labradorite is given by the sketch in Figure 12. These corroded grains are not the normal central cores of zoned plagioclase crystals. The labradorite crystals after a considerable period of growth were so severely attacked by the melt in which they were no longer in equilibrium that many of them entirely and the rest of them largely went back into solution. Then normal zoned plagioclase, the cores of which are andesine, began to crystallize.

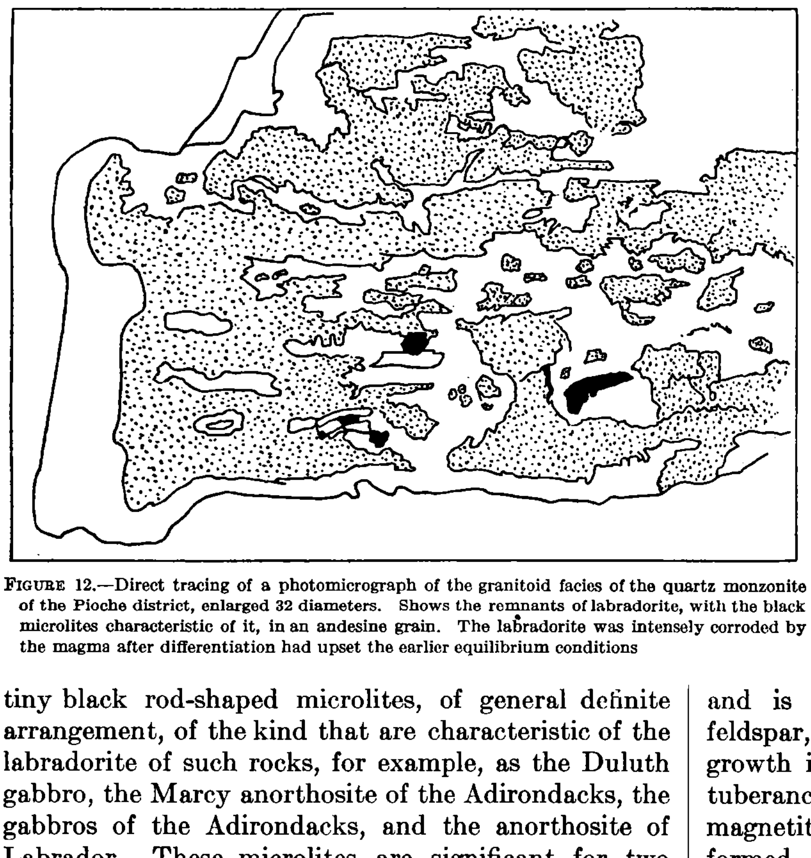


FIGURE 12.—Direct tracing of a photomicrograph of the granitoid facies of the quartz monzonite of the Pioche district, enlarged 32 diameters. Shows the remnants of labradorite, with the black microlites characteristic of it, in an andesine grain. The labradorite was intensely corroded by the magma after differentiation had upset the earlier equilibrium conditions

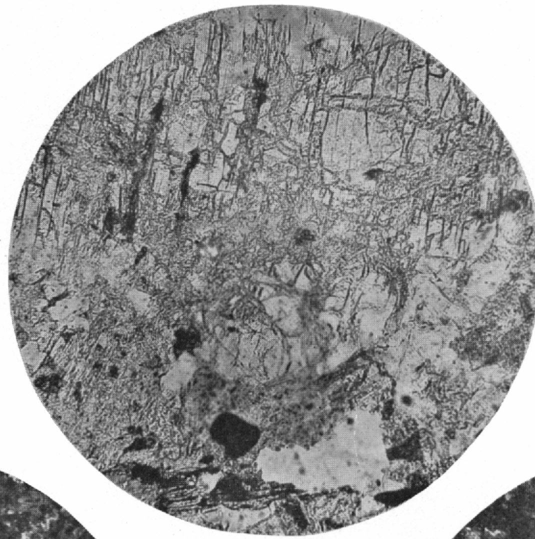
These labradorite cores and the pyroxene and olivine of the border zone indicate that the rock began to crystallize as a gabbro. Differentiation then took place, these minerals were no longer in equilibrium, and others of more acidic character began to crystallize. The change from augite to hornblende indicates that this differentiation was accompanied or caused by an increase in the volatile materials, and the endomorphism and deuteric mineral formation testify to the abundance of such materials.

EVIDENCE OF MINERAL FORMATION AFTER CONSOLIDATION

The potash feldspar was not only a late pyrogenetic mineral, but it continued to grow after consolidation and formed by the replacement of quartz and feldspar. Evidence of this growth is seen in every thin section. (See pl. 10, *A*, and fig. 13.) This replacement, which was carried on most intensely in the endomorphosed rock (pl. 10, *C*), is here called orthoclasization and indicates a tremendous introduction of potash from the reservoir below. The introduction must have begun during the crystallization of the magma and must have been one of the factors that upset the equilibrium at the time when labradorite, augite, and hypersthene were the stable minerals. The only way in which potash could have been introduced into the magma, as it was in the solid rock after consolidation, was by means of volatile emanations. The potash feldspar so formed has a small to moderate optic angle, ranging from 20° to 50°, and rarely shows twinning. Much of it is therefore sanidine, but that in other specimens is perthitic and has a larger optic angle and is therefore microcline. Besides the potash feldspar, hornblende and biotite continued their growth into the deuteric stage, as is shown by protuberances on many of the grains, and also deuteric magnetite, apatite, titanite, and probably zircon formed. Their age is shown by the shape of the magnetite and titanite, and also by the apatite, many of the grains of which are long needles, some of which crosscut the boundaries of more than two adjacent grains. The age of zircon is never satisfactorily proved.

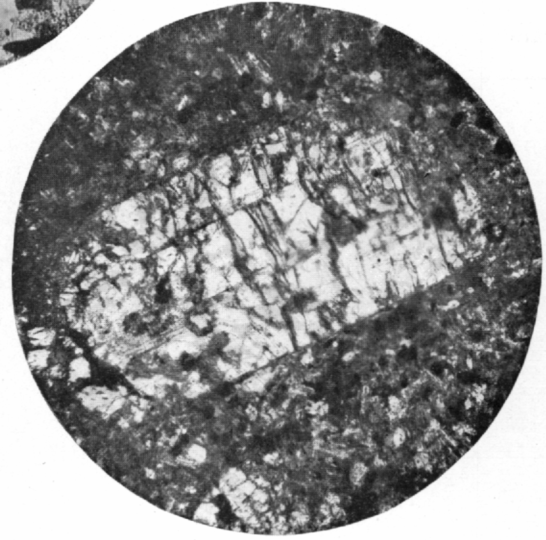
The type of alteration which is loosely described as "hydrothermal" but which is considered by the writer to be a late deuteric stage of mineral formation [3] is much more intensely developed in the porphyritic facies than in the granitoid facies. The primary horn-

[3] Gillson, J. L., Granodiorites of the Pend Oreille district of northern Idaho: Jour. Geology, vol. 35, pp. 18-20, 1927.



A

B



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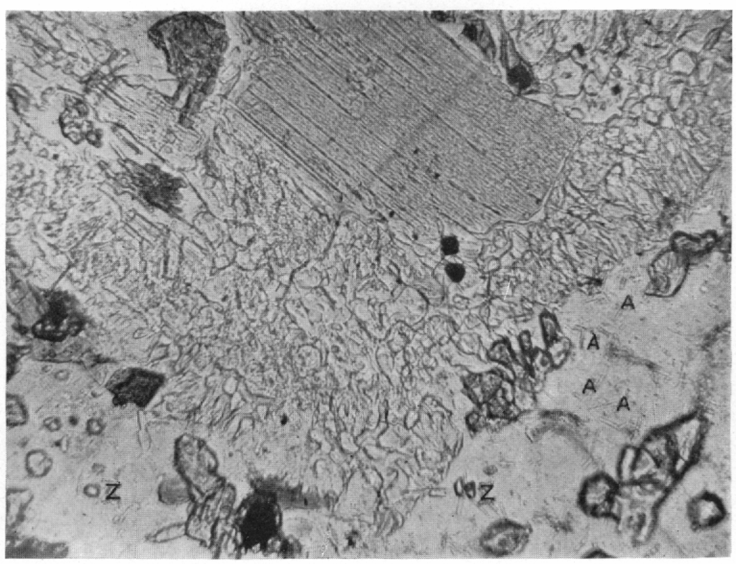


C

PHOTOMICROGRAPHS OF QUARTZ MONZONITE AND METAMORPHOSED LAVA OF THE PIOCHE DISTRICT, NEV.

A, Porphyritic phase of the quartz monzonite, plane polorized light; enlarged about 40 diameters. Shows intense orthoclasization of the plagioclase.
B, Endomorphosed quartz monzonite, enlarged about 40 diameters. An original phenocryst of plagioclase is cut by tongues of andesine and orthoclase, in which there formed also small pyroxene crystals.
C, Endomorphosed quartz monzonite, enlarged about 300 diameters. Shows the graphic intergrowths formed by the replacement of an original plagioclase grain by microcline. Several of the anhedral grains of pyroxene characteristic of the endomorphism are seen, as are also many small needles of apatite (A) and zircon (Z).
D, Metamorphosed Miocene (?) lava, under crossed nicols; enlarged about 40 diameters. Shows the riddling of an old plagioclase phenocryst by veins of potash feldspar.

blende has been replaced by an actinolitic variety, and the chlorite and epidote are so abundant that the rock has a greenish tint. A carbonate was formed in some of the more intensely altered zones, and a zeolite of very low refractive index and strong birefringence was found in many of the altered plagioclase grains.

Alteration of a different type occurred a short distance east of Blind Mountain Spring, where topaz, epidote, and actinolite were abundantly developed along with the intense orthoclasization. The replacement of quartz by orthoclase had proceeded in such a manner that the residuals of quartz have the structure of a graphic intergrowth. Proof that this structure is the result of replacement was afforded by a quartz grain lying between two microcline grains. A graphic intergrowth occurs in the quartz grain on each side, and the quartz component of both intergrowths is in optical continuity with the main part of the grain free from microcline. The microcline component of one is in optical continuity with the nearest microcline grain; that of the other is in optical continuity with the microcline grain adjacent to it.

ENDOMORPHOSED QUARTZ MONZONITE

The endomorphosed quartz monzonite was found at three places. The largest area of its exposures is on the north slope of the spur that extends north from hill 7106 (fig. 11), and the same rock is found on the north side of Kiln Valley and can be traced northwestward along the base of Blind Mountain. A small exposure was also found on the south slope of McCullough Hill, where it seems almost to grade into the hornfels formed by the metamorphism of the Pioche shale. At two points coarsely crystalline aggregates of minerals of unusual variety are interpreted as metamorphosed limestone inclusions in the endomorphosed igneous rock and are described on pages 82–83, under the heading "Metamorphism of mixed rocks."

The identity of the black rock as endomorphosed quartz monzonite was established by microscopic study. It contains plagioclase crystals of the same type, attacked in the same manner (although more intensely) by potash feldspar, and has the corroded labradorite cores in which are the same black microlites as are seen in the quartz monzonite. Furthermore, the border facies of the quartz monzonite contains the first stages of the endomorphism, of which the most characteristic feature is the wide distribution of minute anhedral grains of augite which followed the orthoclasization in a manner so nearly unique that it would probably not be found in different rock types. (See pl. 10, *B*.) The presence of hypersthene in both rocks corroborates the other evidence, as that mineral was not found in any of the other metamorphosed rocks that resemble the endomorphosed quartz monzonite macroscopically.

The character of the rock varies with the degree of metamorphism. The rock has a porphyritic texture and contains plagioclase crystals averaging a millimeter in diameter, set in a groundmass of clear, anhedral, equidimensional grains of potash feldspar, and replacement of the plagioclase by the orthoclase is seen in all stages. Intense orthoclasization of plagioclase produced structures resembling graphic intergrowths, an example of which is illustrated in Plate 10, *C*. Widely scattered through the rock are anhedral augite crystals, many of them subspherical, colorless in section in the less metamorphosed facies,

FIGURE 13.—Direct tracing of a photomicrograph of the granitoid facies of the quartz monzonite of the Pioche district, enlarged 32 diameters. Shows the remnants of two plagioclase grains in orthoclase, the replacement having been a deuteric effect. All outliers marked 1 extinguish together and were originally parts of a single large grain; the same for the outliers marked 2

green in the more metamorphosed. Their size ranges from 0.01 to 0.03 millimeter. Rambling biotite crystals that send out arms or branches inclosing or replacing adjacent minerals are abundant and, being much larger than the other ferromagnesian minerals, can be identified with the naked eye. Biotite is absent, however, from the specimens showing the most intense metamorphism. Widely distributed also are minute grains of magnetite and ilmenite, and large crystals of these minerals of irregular form are found inside of many of the biotite grains. Minute prisms of apatite, seen only with moderately high magnification (pl. 10, *C*) occur by the million in most of the specimens studied. Associated with them are numerous small, generally anhedral crystals of zircon.

When examined closely the plagioclase phenocrysts are seen to have been replaced by potash feldspar only where a previous replacement by glassy andesine had occurred. This replacement was not magmatic corrosion, for where the replacement by orthoclase has gone so far that the pyrogenetic plagioclase crystals are reduced to small residuals of bizarre form, the margin of andesine is invariably found around the edges of the residuals and is also found in the interiors of sections, as spots where the solutions have drilled from above or below.

The small augite crystals never entered the plagioclase phenocrysts except where they had been made over into andesine or potash feldspar. Their formation did not require, however, the presence of the potash feldspar, as they occur in the secondary feldspar of both types.

Quartz is absent from the most intensely endomorphosed varieties, having been replaced by the potash feldspar. In specimens illustrating partial elimination of the quartz, graphic structures showing replacement of the quartz by potash feldspar are very striking.

A small amount of green hornblende of a late generation occurs in some places. Finely divided titanite is abundant locally, and where it is present many of the ilmenite grains are surrounded by a halo of titanite. Very minute prisms, practically opaque although slightly brownish on thin edges, are very abundant, but their identity could not be established.

The most intense endomorphism produced a rock consisting dominantly of potash feldspar in anhedral grains, 0.05 millimeter in diameter, in which are some residuals of plagioclase and abundant anhedral grains of greenish augite and magnetite. Quartz, biotite, and hypersthene are absent, and apatite is not so abundant as in less metamorphosed phases.

At one point, possibly owing partly to the influence on the solutions from passage through limestone blocks near by, the endomorphism went further. Garnet veins cut through the pegmatite stringers, and locally the old igneous rock has been altered to a nearly solid mass of grossularite garnet and two pyroxenes, one having a β refractive index of 1.705 and the other of 1.680. Topaz and calcite also occur in smaller amounts. Thin sections show this stage of metamorphism superimposed on the orthoclasization stage. Residuals of plagioclase remain to prove the genesis of the rock.

In some places nests of epidote occur within the garnet rock.

The paragenesis of mineral formation seems to have been approximately andesine, orthoclase, pyroxene, apatite, zircon, biotite, magnetite, ilmenite, titanite, followed locally by garnet, augite, topaz, and epidote. It is not clear just where in the sequence the pegmatite stringers so commonly found through the rock had formed. They contain principally quartz, potash feldspar, and biotite, with a little andesine.

In some druses found in wider pegmatite veins a few less common minerals occur in small amount. These include small honey-yellow crystals of titanite and a deep-green nonpleochroic pyroxene, the β index of which is 1.725 and the extinction angle $C \wedge Z$ 52°. An unidentified mineral of which a few small grains were found has the following properties: Color black, brittle, β index 1.815, birefringence strong, optically negative, $2V = 15° \pm$, X light brown, Z black.

At one place on the east side of McCullough Hill evidence of endomorphism of a different type was found. In a zone a few feet wide, exactly at the contact with the limestone, the rock is soft and black and consists dominantly of tourmaline, which is pleochroic in shades of pale pink and very deep blue. Quartz, calcite, and small anhedral grains of zircon are disseminated through the tourmaline.

METAMORPHISM OF MIXED ROCKS

On the north slope of the spur extending north from hill 7106 an inquisitive prospector had opened a trench in a 5-foot veinlike mass of green diopside, spinel, and calcite. About 20 feet south of this trench is the normal facies of the granitoid quartz monzonite, and the diopside rock grades sharply on both sides of the vein into recognizable quartz monzonite. Between the vein and the main mass of the intrusive rock no outcrops are found, but the surface is littered with float containing minerals characteristic of the metamorphism of limestone, including vesuvianite and wollastonite. Accordingly, this block of coarsely crystalline minerals grading into the quartz monzonite is considered to be the metamorphosed product of a limestone inclusion in the border phase of the quartz monzonite.

The diopside of the diopside-spinel rock has a β index of 1.692, and the subhedral crystals reach 1 centimeter in diameter and have a grass-green color. The spinel is black, though deep green in color by transmitted light, has a refractive index of 1.740, and gives qualitative tests for magnesium and aluminum and weakly for iron. Most of the grains have curved surfaces but do not appear corroded. A few are octahedrons.

The quartz monzonite on the south side of the "vein" grades sharply within the width of a hand into a nearly white dense rock, consisting of a white pyroxene, the β index of which is 1.690, and an isotropic grossularite garnet. This rock grades into another greenish rock, tinted with red, containing vesuvianite and locally a considerable amount of wollastonite. In several specimens of float there is a second augite, green in section, with a β index of 1.705. Calcite is present in subordinate amount in all specimens.

Another variation, not found in place, is an aggregate of pyroxene, vesuvianite, calcite, and a greenish mica, colorless in section, occurring as pseudohexagonal tabular crystals reaching 1 centimeter in diameter. The indices of refraction of the mica are α 1.555,

β and γ 1.588. The mica is intergrown with or altered to another micaceous mineral, also colorless in section, with indices of 1.530 and 1.555.

Another unusual aggregate of minerals was found at the place marked on Figure 11 "Borate minerals." In a zone 20 feet across, bounded by white marble on one side and by quartz monzonite on the other, occur successively a peculiarly subspherically banded rock, black and white; then a green diopside-spinel aggregate similar to that at the other locality; and then some of the typical endomorphosed quartz monzonite, which gives way to the more normal intrusive rock. The peculiarly banded black and white rock contains the borate minerals ludwigite, szaibelyite, and fluoborite, associated with a great deal of magnetite and serpentine and cut by veins of dolomite and hydromagnesite. This occurrence was described in 1925,[4] but at that time the mineral now considered to be fluoborite was not identified, as the grains were too small and too intimately intergrown for analysis. The optical properties, however, were published, and recently Geijer[5] in studying a similar mineral association identified a new mineral which he called fluoborite, with the composition $3MgO.B_2O_3.3Mg(F,OH)_2$, and which had the same optical properties. Doctor Geijer has suggested that the unknown mineral in the Pioche occurrence is fluoborite.

With the borate minerals occur small amounts of bornite and chalcopyrite, now partly oxidized, and the opening of a prospect on their account has well exposed this interesting occurrence, which otherwise might have passed unnoticed.

At the south contact of the quartz monzonite on McCullough Hill the endomorphosed igneous rock grades sharply but almost imperceptibly into the hornfels that is the product of metamorphism of the Pioche shale. The two rocks appear so similar that the distinction was not recognized in the field.

This grading of endomorphosed igneous rock into exomorphosed sedimentary inclusions or walls indicates that the metamorphism of both followed the intrusion and solidification of the border facies of the quartz monzonite.

CHEMICAL CHANGES DURING ENDOMORPHISM

A composite sample of the endomorphosed quartz monzonite from three places was analyzed. A comparison of this analysis with that of the main mass of the granitoid quartz monzonite suggests the changes in composition caused by the metamorphism, but the comparison is not an exact one because the endomorphosed rock was from the border facies and not the main mass, and border rock free from endomorphism and deuteric potash feldspar could not be found. The endomorphosed rock is low in quartz and rich in feldspar and pyroxene. The analysis suggests an increase in lime, iron, and alumina and a loss in silica and soda. It is probable that if the amount of potash present in the border zone at the time of consolidation could be determined, a notable increase of that oxide would be recorded by the analyses.

Analysis of composite sample of the endomorphosed quartz monzonite of the Pioche district

[Naima Sahlbom, analyst]

SiO_2	57. 87
Al_2O_3	18. 68
Fe_2O_3	3. 13
FeO	3. 37
MgO	1. 26
CaO	7. 38
Na_2O	1. 83
K_2O	4. 14
H_2O-	. 32
H_2O+	. 97
TiO_2	. 80
P_2O_5	. 28
MnO	. 07
	99. 78

CONCLUSIONS

When the magma stoped its way upward and came to rest, it had the composition of a norite, and augite, hypersthene, labradorite, and olivine were crystallizing from it. Differentiation took place, and the composition of the magma was changed so that the labradorite became unstable and was vigorously attacked by the liquid. The evidence for such vigorous attack indicates that though the change in stability was rather sudden it was not the normal progression of equilibrium following a lowering of temperature. If it had been, the labradorite crystals would exist as uncorroded cores in the plagioclase. A slow lowering of temperature would have completely eliminated the labradorite. Corroded labradorite grains as shown in Figure 12 indicate a more special type of differentiation.

Evidence that this differentiation was the result of the introduction into the magma of new material by gases passing upward from below, as a result of which an original norite magma crystallized into a quartz monzonite rock, is as follows: The change of augite to amphibole, the oscillatory zoning of the plagioclases, and the intense endomorphism and exomorphism testify to the presence of gases in large quantity. The large-scale replacement of plagioclase and quartz by deuteric orthoclase and the orthoclasization in the contact metamorphism indicate that these gases were rich in potash. If these gases were abundantly present during the final period of consolidation, it is reasonable to think that they began to appear earlier and accounted for this rather sudden change in equilibrium conditions above noted.

[4] Gillson, J. L., and Shannon, E. V., Szaibelyite from the Pioche, Nevada, district: Am. Mineralogist, vol. 10, pp. 137-139, 1925.

[5] Geijer, Per, Some mineral associations from the Norberg district: Sveriges geol. Undersökning, ser. C, No. 343, pp. 26-27, 1927.

That crystal settling did not play much part in this differentiation is indicated by the presence of the corroded labradorite crystals in quantity at the roof of the batholith and by the presence of oscillatory zoning in the plagioclase rims around them.

The conclusion is thus drawn that this quartz monzonite illustrates the type of differentiation suggested by Fenner.[6] It shows, also, how different a consolidated rock may be from the magma that made the intrusion.

CONTACT-METAMORPHOSED MIOCENE (?) VOLCANIC ROCKS

The rocks around Blind Mountain are in a downfaulted part of an overthrust block containing Devonian sediments and some of the volcanic rocks. Because of the faulting and the intrusion of many apophyses of a porphyritic facies of the igneous rock, the bedding of the sedimentary and volcanic rocks can not in general be distinguished. Moreover, the close resemblance between the metamorphosed lavas and the fine-grained dikes made mapping difficult in the field.

The volcanic rocks consist dominantly of latite, dacite, and andesite tuffs, breccias, and flows and are in general crumbly and have suffered severely from erosion. Only a few of the more massive flows make bold hills. Blind Mountain, however, one of the sharpest and most conspicuous of the minor summits, consists of tuff and breccia, indurated by the metamorphism into rocks as hard and resistant as quartzite.

A description of the character of the metamorphism which has so profoundly changed these rocks is given below.

GENERAL CHARACTER

The unmetamorphosed flows are black, gray, or reddish fine-grained porphyritic rocks, in many of which flow lines are conspicuous. The breccias are light in color and contain fragments of other volcanic rocks of various size and character. Open cavities are numerous. The tuffs differ from the breccias in the absence of the rock fragments. Most of them contain biotite and feldspar phenocrysts and the microscopic glass fragments described by Pirsson[7] as characteristic of such rocks.

Many of the metamorphosed volcanic rocks have a cream or a brownish-brick color on weathered surfaces and a light greenish color on the fresh fracture. Others are very dark brown or nearly black. Most of them break with a very hackly fracture, and the greater number contain minute feldspar phenocrysts. Many have open cavities containing garnet, epidote, magnetite, and amphibole. Nearly all contain disseminated pyrite and pyrrhotite.

The volcanic breccias can be identified the most easily. They contain angular to rounded fragments of dense dark chocolate-brown rocks embedded in a mottled green matrix. Small subspherical or irregularly shaped green or black masses, with white borders, formed by aggregates of secondary minerals, are widespread and very characteristic of the breccias.

In general, the metamorphism of the volcanic rocks involved the elimination of all the pyrogenetic ferromagnesian minerals and the formation of much potash feldspar and a new generation of ferromagnesian silicates, the latter generally in very finely divided form. To illustrate the process of metamorphism a description will be given of a few flows, tuffs, and breccias.

ILLUSTRATIVE EXAMPLES OF THE METAMORPHISM

East of hill 7106, at the west base of the spur projecting toward the north slope of that hill (see fig. 11), a considerable area is covered with fragments of a nearly black rock, which does not, however, crop out conspicuously. It weathers a light brown and contains scattered plagioclase crystals reaching 1½ millimeters in length. The original character of the rock is shown under the microscope by the old plagioclase phenocrysts, by the quartz and apatite, and by pseudomorphs of the ferromagnesian minerals. An intense orthoclasization of the groundmass and a similar attack on the phenocrysts had nearly eliminated the original minerals. Minute grains of biotite, pyroxene, and magnetite of metamorphic origin give the dark color to the rock. In the groundmass these are so generally distributed as to appear with moderate magnefication as a cloud of minute specks. The plagioclase phenocrysts are zoned and have labradorite cores tinted dark because of submicroscopic inclusions. All the phenocrysts show to a greater or less extent an attack by the orthoclase. This had begun by the formation of narrow irregular veins, in general following cleavage cracks, and from these veins the replacement had extended through the phenocrysts. The minute grains of biotite, pyroxene, and magnetite that are so abundant in the orthoclase of the groundmass are less common in the orthoclase of the phenocrysts, except near the edges. Accumulations of nearly solid masses of the dark silicates and magnetite represent imperfect pseudomorphs of pyrogenetic ferromagnesian minerals, many of which have the form of augite crystals. In addition to the widespread, finely divided magnetite, larger grains of it are fairly abundant. Many have the form of deuteric magnetite found in igneous rocks and are presumably of a late period of formation. A subordinate amount of rutile is also present. In specimens of this rock showing the most intense alteration the plagioclase phenocrysts have been so completely replaced by orthoclase that their pseudomorphs are conspicuous only between crossed nicols, where the outline of orthoclase grains larger than those in the groundmass pre-

[6] Fenner, C. N., The Katmai magmatic province: Jour. Geology, vol. 34, pp. 743-744, 1926.

[7] Pirsson, L. V., The microscopical characters of volcanic tuffs—a study for students: Am. Jour. Sci., 4th ser., vol. 40, pp. 195-197, 1915.

serves the shape of the plagioclase phenocrysts. Through the groundmass irregularly shaped aggregates of pyroxene grains are widespread. The individual grains are about 0.003 millimeter in diameter; the aggregates are as large as 0.25 millimeter. Many aggregates contain cores of magnetite. Aggregates of biotite also occur, and in many of these are single grains of magnetite, larger than those in the pyroxene aggregates. Many of these magnetite grains have projecting fingers penetrating into the biotite, indicating a later formation. They are similar to the late magnetite crystals in the peculiar aggregate structures or "reaction rims" of the Adirondack gabbros.[8] Some grains of ilmenite occur in the groundmass, and these are surrounded by a halo of minute titanite crystals.

Another metamorphosed lava from the summit of Blind Mountain is similar to that just described. Orthoclasization and the formation of very minute silicates and magnetite had been caused by the metamorphism. Besides the finely divided augite, characteristic of most of these rocks, an amphibole of actinolitic habit, epidote, titanite, and a little allanite occur. The rock contains numerous ellipsoids 1 to 5 millimeters in diameter, black but with white rims. The black core in some is augite, in others an amphibole with a β index of refraction of 1.690, weak birefringence, and distinct dispersion, ρ greater than v. With the amphibole is a brown isotropic garnet, index 1.87. The white rim is andesine, partly replaced by orthoclase. The apatite is so abundant in this and in many of the other metamorphosed lavas that it must represent an introduction of phosphate. Long needles of apatite, crosscutting the boundaries of several grains, are certainly of a late period of formation.

A lava flow illustrating the most intense alteration was found on the summit of the spur east of the north end of hill 7106. Along joint seams garnet and rhodonite crystals were found. The phenocrysts in the rock are so completely replaced by orthoclase that the crystal boundaries are distinguishable only between crossed nicols, where the difference in birefringence due to grain size is emphasized. In the pseudomorphs, however, a few remnants of plagioclase occur. The groundmass is made up largely of small equidimensional orthoclase grains in which very minute crystals of high refractive index are abundant. These include apatite, garnet, titanite, and a micaceous mineral similar to biotite in color but nonpleochroic.

In some of the lavas in which considerable quartz was originally present, replacement graphic structures made by the introduction of potash feldspar into the quartz are numerous.

The metamorphosed volcanic breccias represent a similar process of metamorphism but offer wider variations, inasmuch as the original rock was not uniform. The tuffaceous nature of the groundmass is generally readily apparent under the microscope. The irregular streaks and angular fragments of original glass are preserved as pseudomorphs of orthoclase in which there are countless minute grains of dark silicates, the average individual size of which is about 0.001 millimeter. Pseudomorphs of twisted and bent biotite crystals and angular fragments of foreign rocks testify to the original character of the rock. Spots made of spherical aggregates of minerals and partly filled druses are widespread. The plagioclase phenocrysts of the breccias are all more or less replaced by orthoclase. Although several grains of pyrogenetic ferromagnesian minerals were found, their rarity indicates that a complete elimination of such minerals, so common to these rocks, had occurred. One old hornblende crystal was found in one thin section, partly replaced by a carbonate and surrounded by a rim of finely divided pyroxene. In another section a large augite had begun to break down into small crystals of the same mineral. Aggregates of the tiny augite and hornblende crystals of metamorphic origin made crude pseudomorphs of pyrogenetic ferromagnesian minerals.

The rock fragments found in the volcanic breccias show a great variety of mineralogic features. Most of the metamorphosed fragments resemble the metamorphosed lavas and consist of a groundmass of finely divided potash feldspar and larger plagioclase grains, also replaced by orthoclase. Disseminated ferromagnesian grains and magnetite occur in all. Epidote and titanite seem to be more abundant in the rock fragments in the breccias than in the lavas, and biotite, common in the lavas, is absent from most of the metamorphosed volcanic breccias.

Green or black spots or ellipsoids with white rims are characteristic of the metamorphosed volcanic breccias. In a specimen from the summit of Blind Mountain the green cores consist of augite aggregates and of an actinolitic hornblende, the β refractive index of which is 1.650. The white rim is made up of calcite, many of the grains of which are optically biaxial.

Garnet occurs in most of the breccias. Many grains half a centimeter across are readily visible, and under the microscope scattered garnet crystals are found in many thin sections. Their irregular form indicates a late time of crystallization. Where tested the garnet has an index of refraction near 1.690 and a reddish-brown color.

Druses, partly filled with pretty crystals of pyroxene and lined with chlorite and magnetite, were found. In one of them slender needles of a green hornblende are perched on octahedrons of magnetite.

A breccia metamorphosed to a different product than the others was found on the spur east of hill 7106. Although the common process of orthoclasiza-

[8] Gillson, J. L., Callahan, W. H., and Millar, W. B., The age of certain of the Adirondack gabbros and the origin of the reaction rims and peculiar border phases found in them: Jour. Geology, vol. 36, pp. 153-157, 1928.

tion had occurred, most of the rock consists of augite and scapolite, with radiating fibers of chlorite and abundant minute grains of titanite. Large grains of garnet occur haphazardly.

A striking feature in these metamorphosed volcanic rocks is the occurrence on joint surfaces of drusy coatings of crystals. An occurrence of green garnet and red rhodonite, already mentioned, is an example of the beauty of these occurrences. Along another joint seam the zeolite steelerite is associated with an unidentified mineral having the following properties: β 1.615, positive sign, moderate birefringence, $2V=75°$, hardness 5–6, color white, cleavage none.

CONCLUSIONS

The geologic importance of the contact metamorphism of the Miocene (?) lavas lay principally in the evidence it afforded on the date of the intrusion of the quartz monzonite. Descriptions of contact-metamorphosed volcanic rocks are not numerous, however, and thus the problem deserved study.

When the quartz monzonite was intruded it had the composition of a gabbro (see p. 80), but during its crystallization a rapid differentiation took place because of the upward passage of potash-bearing emanations from below. Because of the chemical changes caused by the reaction between these emanations and the magma the final product of crystallization was a rock of quartz monzonite composition. The border facies, solidified earlier, was intensely endomorphosed by these emanations. Orthoclasization had occurred, and finely divided biotite, augite, apatite, and magnetite formed in the feldspar. In places of intense endomorphism garnet also is found.

The metamorphism of the lavas is in harmony with that of the quartz monzonite. The solutions were rich in potash and iron and carried phosphate and titanium. Whether the magnesium, aluminum, lime, etc., of the metamorphosed rocks are due simply to a recrystallization of material already present or represent some introduction, it is impossible to state.

In the endomorphosed quartz monzonite and in the metamorphosed lavas sericite, chlorite, serpentine, and zeolites, characteristic of late stages of metamorphism such as that in the Pend Oreille district of northern Idaho,[9] are not abundant. The metamorphism was intense but rather short lived and was completed while the temperatures were high, for garnet was one of the late minerals. The heated waters given off in long progression from the cooling solid intrusive and causing "hydrothermal" alteration in other districts were not abundant here or were impotent to make many mineralogic changes.

[9] Gillson, J. L., Contact metamorphism of the rocks in the Pend Oreille district of northern Idaho: U. S. Geol. Survey Prof. Paper 158, pp. 111-120, 1929.

THE VARVES AND CLIMATE OF THE GREEN RIVER EPOCH

By Wilmot H. Bradley

ABSTRACT

The Green River formation is a series of lake beds of middle Eocene age which occupies two large intermontane basins, one in Colorado and Utah, the other in Wyoming. The formation averages about 2,000 feet in thickness and covers an area of more than 25,000 square miles. Many of its beds of marlstone, oil shale, and fine-grained sandstone contain varves. As the origin of these varves is closely linked with the climate the writer has attempted rough quantitative estimates of several elements of the climate of the Green River epoch. These estimates are based largely upon the relative area of the lake and its drainage basin. A climate is postulated which was characterized by cool, moist winters and relatively long, warm summers. Presumably the temperature fluctuated rather widely from a mean annual temperature that was of the order of 65° F. The rainfall varied with the seasons and probably also fluctuated rather widely from a mean annual precipitation between 30 and 43 inches.

One type of varve predominates. This consists of a pair of laminae, one of which is distinctly richer in organic matter than the other. The contacts between the two parts of the varve and between successive varves are generally sharp. The varves differ considerably in thickness according to the type of rock in which they occur and range from a minimum of 0.014 millimeter in the beds of richest oil shale to about 9.8 millimeters in the beds of fine-grained sandstone. The average thickness of the varves, weighted according to the quantity of each type of rock in the formation, is about 0.18 millimeter.

The assumption that the pairs of laminae are varves is tested, first, by analogy with the varves in the deposits of modern lakes and, second, by calculation of the thickness of annual laminae to be expected in the ancient Green River Lake based upon data of present stream loads.

The bipartite character of the varves is explained by postulating a more or less continuous sedimentation of mineral and organic constituents, with first a peak in the production of the carbonates and then a peak in the production of the plankton, both peaks apparently occurring during the summer, and by assuming that the primary difference in composition was accentuated by the differential settling rates of the two principal constituents. The preservation of the varves suggests that the lake water was thermally stratified and that the lake may not have been more than 75 or 100 feet deep where the varved deposits accumulated.

Three cycles of greater length than the varve cycle are suggested by fairly regular recurrent variations in the thickness of the varves and in the thickness and character of certain beds and by the fairly regular spacing of certain salt-mold layers. The first of these cycles averaged a little less than 12 years in length and appears to correspond to the cycle of sunspot numbers. The second cycle had an average length of about 21,600 years and suggests the average period of about 21,000 years which is the resultant of the cyclic changes of eccentricity of the earth's orbit and the cycle of the precession of the equinoxes. The third cycle, which was about 50 years long, agrees with no well-established rhythm.

From measurements of the varves the Green River epoch is estimated to have lasted between 5,000,000 and 8,000,000 years. The rate of accumulation of the fluviatile deposits above and below the Green River formation is estimated as about 1 foot in 3,000 years, which would indicate that the combined length of the Wasatch, Bridger, and Uinta epochs was between 8,000,000 and 25,000,000 years. From these figures the duration of the Eocene epoch is estimated as between 13,000,000 and about 33,000,000 years, the average of these estimates being a little less than 23,000,000 years. This estimate agrees rather closely with estimates of the duration of the Eocene epoch based on the age determinations of radioactive minerals, but it is entirely independent of them.

INTRODUCTION

Estimates of the age of the earth and of intervals of geologic time in terms of years have long fascinated geologists. Of late the accuracy of these estimates has been materially increased by using the rate of disintegration of certain radioactive minerals. As the methods involved in these determinations come to be more and more refined the results may be relied upon with increasing confidence. Even so, independent evidence which can be used as a check will always be desirable. Annual strata or varves [1] in sedimentary rocks seem most likely to yield data from which reliable check estimates can be made, but only rarely is such a record of the seasons preserved. Varves in the Pleistocene and recent glacial deposits furnish probably the best-known examples of this type of seasonal record, though varves in the glacial deposits of older geologic periods are becoming better known. Seasonal laminae in marine deposits, not related to glacial phenomena, are even more rare; nevertheless they are known, and one of the most interesting occurrences has recently been described and thoroughly discussed by Rubey.[2] The recognition of varves in the

[1] Ernst Antevs (Retreat of the last ice sheet in eastern Canada: Canada Dept. Mines Mem. 146, p. 1, 1925) gives the following definition and etymology of varve: "De Geer (1912, p. 253) proposed the use of the word *varve*, *n.*, pl. *-s*, English and French—*Warw*, pl. *-e*, German—as an international term for the distinctly marked *annual deposit* of a sediment regardless of its origin. A varve usually consists of two but might consist of more different layers, and in the varved late glacial clay mostly of one layer of silt and one layer of stiff clay. The Swedish word *varv* (old spellings *hwarf* and *hvarf*, Icelandic *hwerf*) means turn, round, revolution (of a body), lap (sport term), time (wind a band once, four times round), row, tier, course, and layer. (Cf. the English words warve, *n.*, same as wherve, *n.*, meaning a whirl of a spindle, and wherve, *n.*, same as whorl, *n.*, meaning the fly of a spinning-wheel, etc.)"

[2] Rubey, W. W., Lithologic studies of fine-grained Upper Cretaceous sedimentary rocks of the Black Hills region: U. S. Geol. Survey Prof. Paper 165-A (in press).

Green River formation is not entirely new. Sayles [3] says:

Blackwelder has recently made a study of oil shales from the Green River region. This shale has very regularly alternating brown and black bands. Blackwelder took the specific gravity of the material of these brown and black layers and found the former had an average specific gravity of about 1.9 and the latter about 1.3. This, in the opinion of Blackwelder and the writer, is a case which strongly suggests seasonal deposition.

White [4] wrote of certain beds in the Green River formation, "The lamination of the sapropelic deposits may be seasonal or due, in many cases at least, to the generative periods of the principal organic constituent of the raw material."

The records afforded by varved rocks are chiefly valuable for measuring short intervals of geologic time, but when they can be used for estimating intervals of time long enough to be compared with estimates based on the lead-uranium ratio of radioactive minerals an unusual interest attaches to them. The lacustrine varves in the rocks of the Green River formation provide the basis for an estimate of the length of the Eocene epoch which seems to agree rather closely with current estimates based upon radioactive determinations. Accordingly it is the primary purpose of this report to describe the thin rhythmic laminæ found in certain beds of the Green River formation, to test the hypothesis that they are varves, and to interpret their significance in terms of time and in terms of the conditions that prevailed during their deposition.

Quantitative estimates of the various factors of past climates are as desirable as estimates of geologic time. Usually, however, the results of such inquiries are disappointing because the available evidence is indefinite. Certain relations between the size of one of the ancient Green River lakes and the area of its hydrographic basin appear to offer that desired definiteness which is necessary in order to attempt even a partial reconstruction of an ancient climate. For this reason and because the origin of the varves is so intimately linked with climate, the secondary aim of this report is to attempt a quantitative estimate of several factors of the climate that prevailed in the vicinity of the ancient Green River lakes during the middle of the Eocene epoch.

FIELD WORK AND ACKNOWLEDGMENTS

The field work upon which this report is based was done in the years 1922 to 1925. In 1922 the writer assisted J. D. Sears, of the United States Geological Survey, in Moffat County, Colo., and in southern Sweetwater County, Wyo. In 1923 the writer was assisted by C. H. Dane, and in 1924 by C. E. Erdmann, both also of the Geological Survey; in 1925, by R. D. Ohrenschall. Most of the field data used in this report, however, were obtained in 1925 while the writer was studying the stratigraphy of the Green River formation in northwestern Colorado and northeastern Utah.

The writer wishes to express his thanks to Walter N. White, of the United States Geological Survey, who furnished the data on evaporation rates and criticized the section on the climate of the Green River epoch; to Prof. E. W. Berry, of Johns Hopkins University, who read the section on climate and offered critical suggestions; and to his colleagues of the Geological Survey, particularly W. W. Rubey and C. H. Dane, whose suggestions and criticism have been very helpful.

TOPOGRAPHY AND DRAINAGE OF THE ANCIENT GREEN RIVER BASIN AND ITS VICINITY

The Green River formation comprises a series of lake beds of middle Eocene age laid down in two large intermontane basins, one in Wyoming, the other in Utah and Colorado. The formation covers an area of more than 25,000 square miles and has an average thickness of about 2,000 feet. For the lake that occupied the Green River Basin of Wyoming during the Green River epoch the writer proposes to use the name Gosiute Lake, which King [5] suggested long ago. The limits of this ancient lake as determined by subsequent studies show that instead of extending from somewhere in the vicinity of Middle Park, Colo., westward as far as longitude 116°, as King supposed, the lake was restricted to the Green River Basin. (See fig. 14.) South of the Uinta Mountains, in Utah and Colorado, a single large body of water occupied the Uinta and Piceance Creek Basins during the greater part of the Green River epoch, and this lake the writer proposes to designate Uinta Lake. Although it seems probable that Uinta Lake was divided into two or possibly three lakes at certain stages of low-water level there appears to be no need to name these parts.

The drainage basin of Gosiute Lake apparently included more than the Green River Basin of Wyoming; as nearly as the writer can make out, it was bounded by the Uinta and Williams Fork Mountains on the south, by the Wasatch and Wyoming Ranges on the west, by the Gros Ventre and Wind River Ranges and a partly collapsed eastward extension of the Wind River Range on the north, and by the Laramie Mountains on the east. Into this basin the Medicine Bow Range, Sierra Madre, and Elkhead Mountains projected. All these mountains were formed at the end of the Cretaceous period, and the principal modification of them since has been effected by erosion.

[3] Sayles, R. W., The dilemma of the paleoclimatologists: Am. Jour. Sci., 5th ser., vol. 3, p. 458, 1922.

[4] White, David, The carbonaceous sediments, in Twenhofel, W. H., Treatise on sedimentation, p. 302, 1926.

[5] King, Clarence, U. S. Geol. Expl. 40th Par. Rept., vol. 1, p. 446, 1878.

Blackwelder [6] has shown that during the Tertiary period the mountains were first nearly buried under thick deposits; then they were modified somewhat by faulting and warping, which, in general, only accentuated the structure formed at the end of the Cretaceous period; later in Tertiary time they were exhumed by deep erosion of the early Tertiary rocks; and during in common with the general level of that part of the continent, was probably less than 1,000 feet above sea level. Presumably the great regional uplift did not occur until some time late in the Tertiary period. The assumptions which follow and from which are deduced the features of the Eocene climate depend more or less directly upon this conception of the Eocene topog-

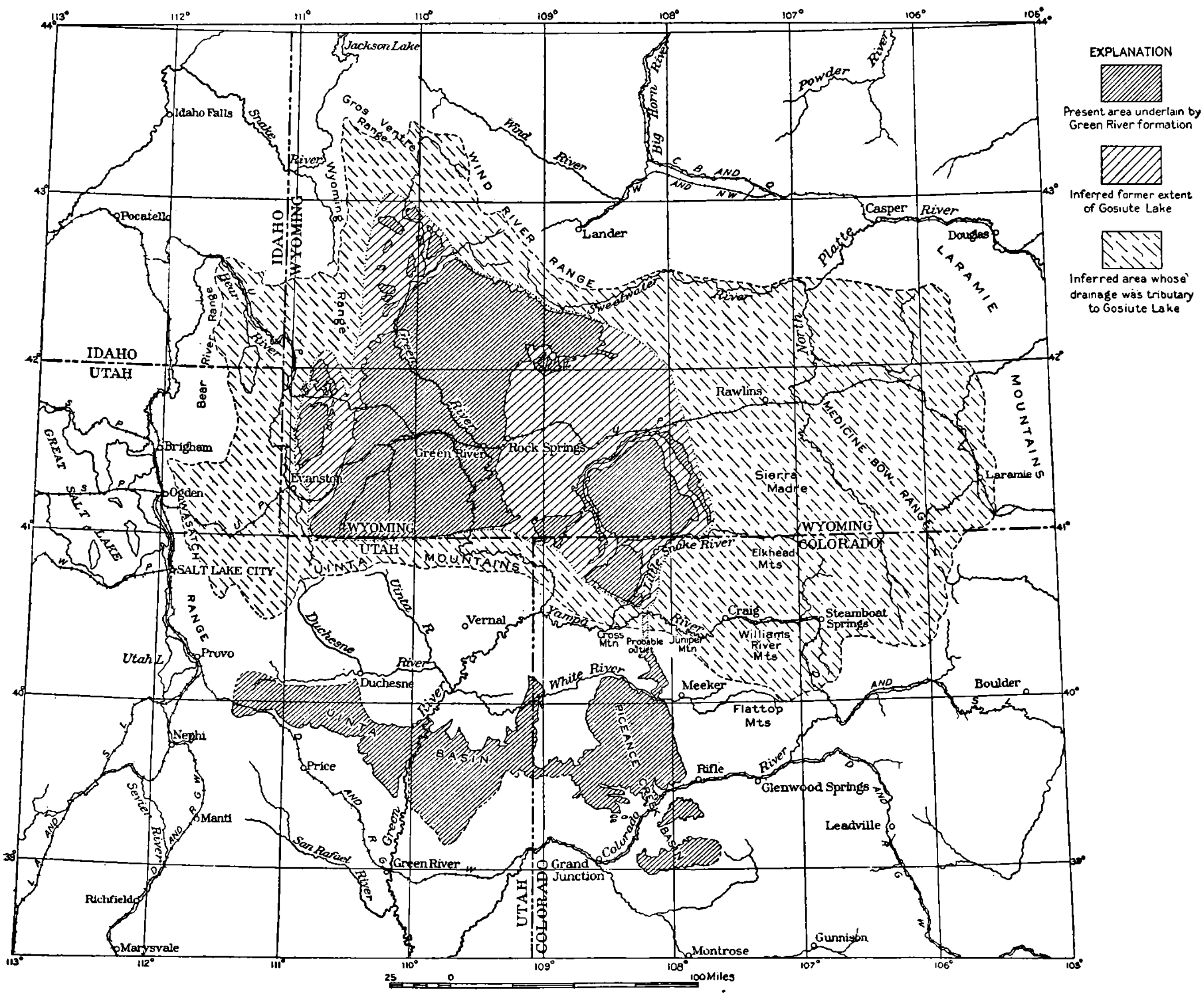


FIGURE 14.—Outline map of the Green River Basin and vicinity, Wyoming, Utah, and Colorado, showing the present area underlain by the Green River formation, the estimated mean position of the shore line of the ancient Gosiute Lake, and the estimated boundaries of its hydrographic basin

and after this exhumation the mountain ranges themselves were more or less deeply eroded, with the result that they were considerably narrowed and their crests were lowered.

From this history it appears that the mountain ranges and high divides that form the rim of the Gosiute drainage basin were probably somewhat higher with respect to the floor of the basin during Eocene time than at present. The floor of the basin, however, raphy of the Green River Basin and the surrounding country.

The boundaries of the hydrographic basin of Gosiute Lake were, in accordance with this conception, probably almost coincident with the positions of the present mountain crests, except along the eastward extension of the Wind River Range. (See fig. 14.) These crests are, in general, high above the level of the beds of the Green River formation. At a low place in the rim in northwestern Colorado, at the east end of the Uinta Mountains, between Cross and Juniper Mountains,

[6] Blackwelder, Eliot, Post-Cretaceous history of the mountains of central-western Wyoming: Jour. Geology, vol. 23, pp. 101-117, 193-217, 1915.

there apparently was an outlet to Gosiute Lake. This low place is only a few miles north of a thick sandy deltaic facies of the Green River formation, and as this sandy facies grades out southward into the normal very fine grained rocks that are characteristic of the formation in the Piceance Creek and Uinta Basins it suggests that a stream flowing across this low place in the rim connected Gosiute Lake with Uinta Lake, to the south, at least during certain intervals. A thick sandy facies perhaps should hardly be expected to form just below the outlet of a lake as large as Gosiute Lake, in which most of the detrital material must have settled. But the stream discharging from that lake apparently flowed for a considerable distance, perhaps 10 or 15 miles, across hogbacks of the more or less sandy Cretaceous and older rocks that form the eastward extension of the Uinta Mountain uplift. This hypothesis concerning the outlet of Gosiute Lake might be tested by a mineralogic comparison of the Cretaceous sandstone beds of that locality and of the sandstone in the deltaic facies of the Green River formation.

Most of the streams within the Gosiute hydrographic basin were apparently rather short and flowed directly into the lake, but those in the eastern part were longer and may have had considerable volume.

The mean position of the shore of Gosiute Lake is known fairly well along the west side of the Green River Basin and approximately along the south flank of the Wind River Range and the north flank of the Uinta Mountains. In other parts of the basin its position can be estimated from the position and character of the present outcrop of the Green River formation. The relations between this inferred outline of the lake and its hydrographic basin are shown in Figure 14, from which the writer calculated that the lake occupied about 36 per cent of the basin. This relation, though it varied from time to time with changes in lake level, remained approximately the same through the greater part of the Green River epoch.

CLIMATE OF THE GREEN RIVER EPOCH

Some features of the climate that prevailed in the vicinity of the ancient Green River lakes will be considered here, as they have a direct bearing on the interpretation of the lamination in the rocks of the Green River formation. The conclusions regarding the climate of Green River time have been drawn almost wholly from data on the Green River Basin of Wyoming. It is assumed that the climate in this basin, north of the Uinta Mountains, and that in the Uinta and Piceance Creek Basins, south of the Uinta Mountains, were essentially the same.

The fact that the area of Gosiute Lake remained approximately the same through the greater part of the Green River epoch in itself indicates that the climate was relatively humid or relatively cold rather than arid. Lake Bonneville[7] at its maximum extent, which was during a relatively cold epoch when the rate of evaporation was not great, flooded about 38 per cent of its hydrographic basin. Three of the Great Lakes—Superior, Michigan, and Huron—together occupy a little more than 36 per cent of their combined hydrographic basins.

If it is accepted that Gosiute Lake covered approximately 36 per cent of its hydrographic basin during the greater part of the Green River epoch, then the average annual rainfall of the region can be roughly approximated by using certain assumed rates of evaporation from the free water surface and from the land and plant surfaces. These assumed rates were taken from the Gulf Coast States, because of similarity between the Green River flora and the flora of those States, and from Lakes Superior, Michigan, and Huron, because of similarity in ratio of free water surface to drainage basin.

COMPARISON WITH THE GULF STATES

The flora of the Green River formation has lately been revised by Knowlton,[8] who also gives his interpretation of the ecology. Brown[9] has added more than 40 new species and has also given an excellent discussion of the environment. Brown reached essentially the same conclusion as Knowlton—that the flora had a dual aspect, many of its forms representing a warm, moist lowland type of flora and the others representing a cooler and perhaps somewhat drier upland type. As Brown's and Knowlton's conclusions are so nearly alike and as Berry[10] has recently analyzed Knowlton's interpretation of the Green River flora and virtually epitomized the conclusions of both Knowlton and Brown, it will suffice to consider here only Berry's summary, which comprises also conclusions drawn from his own studies. Berry points out a striking similarity between the flora of the Green River formation and a new flora found in the so-called Bridger beds, which overlie red-banded beds of the Wind River formation in the Wind River Basin. In analyzing Knowlton's interpretation[11] of the Green River flora Berry[12] says:

> Certainly the ensemble suggests a warm and genial climate, but that this borders on tropical in any precise use of that term or that the winter season was without frost is most doubtful. Nor is it necessary to assume that the fossils include the mechanically mixed representatives of lowland and upland associations. There is not a single well-authenticated genus

[7] Gilbert, G. K., Lake Bonneville: U. S. Geol. Survey Mon. 1, pp. 20, 105, 297, 1890.

[8] Knowlton, F. H., Revision of the flora of the Green River formation: U. S. Geol. Survey Prof. Paper 131, pp. 133-197, 1923.

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

[10] Berry, E. W., Flora and ecology of so-called Bridger beds of Wind River Basin, Wyo.: Pan-Am. Geologist, vol. 44, pp. 357-368, 1925.

[11] Knowlton, F. H., op. cit., p. 147.

[12] Berry, E. W., op. cit., pp. 366-367.

recorded from the Green River beds that is not found in the United States at the present time, and some of the genera range northward to New England and Canada.

I consider it utterly impossible to estimate the summer's heat or the winter's cold or to attempt to give mean annual temperatures, but a picture of temperature requirements can be given by citing a modern region where such a fossil flora would find optimum conditions for growth and reproduction. I have already suggested southern Louisiana as such a region, as regards temperature, rainfall, and humidity. This comparison might be somewhat extended by the statement that I know of no member of the Green River or so-called Bridger floras that would not be perfectly at home somewhere in the region between South Carolina and Louisiana in our South Atlantic and Gulf States at the present time. Hence, if general terms are imperative, these floras are warm temperate and not tropical.

The collections from the so-called Bridger beds of the Wind River Basin included some petrified wood with well-marked annual rings that clearly indicated seasonal changes, either hot and cold, or wet and dry. Although there was nothing to indicate a choice between these two alternatives Berry [13] concluded from the structural details of the wood that the seasonal contrasts throughout the year were not extreme.

The analogy between the fossil flora of the Green River formation and the present flora of the South Atlantic and Gulf Coast States serves as a basis in choosing evaporation rates for the purpose of estimating the probable average annual rainfall of the Gosiute Lake Basin during the Green River epoch. Accordingly, it is assumed that Gosiute Lake occupied a region whose climate was comparable to that of Alabama to-day.

The mean annual evaporation rate from a standard Weather Bureau pan at Silver Hill, Ala., is 58.5 inches. However, W. N. White,[14] of the United States Geological Survey, has found that in arid and semiarid regions these pans give too high a rate and that the observed rate has to be multiplied by a factor of 0.66. Although experimental data are lacking which would justify the application of this or any other corrective factor to readings of the standard Weather Bureau pan in humid regions, it seems likely that these pan readings are too high and that some correction should be made. In the lack of more pertinent information the factor 0.66 is used here. Probably this gives too low a rate, but if so the estimates based upon it will err in being conservative rather than extreme. The corrected mean annual rate of evaporation from the free water surface at Silver Hill, Ala., is therefore taken as 38.6 inches, or in round numbers 39.

The mean annual rate of transpiration from plants and evaporation from the land surface in Alabama is about 37 inches.[15] This figure is based upon the records of the Coosa River between Clanton and Montgomery, Ala. According to Berry's interpretation of the flora,[16] it might be legitimate to assume that this average rate of evaporation from the land and plant surfaces of Alabama is applicable to the entire land area of the ancient Green River Basin. Nevertheless the physiographic evidence seems to compel at least a rough division of the Gosiute Lake Basin into a lowland area and an upland area, and in the absence of more definite information it is assumed that the areas of the lowland and upland are equal. The mean annual evaporation rate of 37 inches will be used for the lowland. As a basis of estimate for the upland zone, which is supposed to represent the higher parts of the encircling mountain ranges, the basin of the Lenville River, in southwestern North Carolina, is chosen. The probable mean annual rate of evaporation from the land surface and transpiration from plants for the Lenville River Basin is about 30 inches. The average annual rate of evaporation and transpiration from the entire land and plant surfaces of the Green River Basin is therefore assumed to have been 33.5 inches, the arithmetical mean of the rates chosen here for the lowland and upland zones.

Now, as the land portion represented about 64 per cent of the total area of the hydrographic basin it would require 64 per cent of 33.5 inches, or 21.4 inches of precipitation over the whole basin, to balance the evaporation from the surface of the land and transpiration from plants. In the same way, as 39 inches is the assumed mean annual rate of evaporation from the free water surface, it would require 36 per cent of 39 inches, or 14 inches of precipitation over the whole basin, to balance the annual evaporation from the lake surface. A total mean annual rainfall of 35.4, or in round numbers 35 inches, is therefore indicated if it is assumed that the lake maintained nearly its maximum dimensions but did not overflow. Salt molds widely distributed at certain levels in the rocks of the Green River formation provide a good basis for this assumption, as fairly large quantities of salt could be deposited only from a body of water which had not overflowed for a considerable period of time. Mud cracks far out in the basin indicate further that at certain stages the lake contracted greatly from its maximum extent. But there is equally good reason to believe that at other stages the ancient lake overflowed for long periods. The particular stage at which evaporation exactly balances precipitation was chosen for these computations, first, because it obviates estimates either of discharge from the lake or of rate of contraction of the lake, and second, because it is a critical stage between a saline and a fresh condition of the lake and a stage through which Gosiute Lake evidently passed repeatedly.

[13] Berry, E. W. op. cit., pp. 366-367.

[14] Oral communication.

[15] Mr. Walter N. White, of the Geological Survey, has kindly supplied the writer with practically all the data available on evaporation rates in the United States.

[16] Berry, E. W., op. cit., pp. 366-367.

COMPARISON WITH THE GREAT LAKES

Another rough approximation, wholly independent of that based upon comparison with the Gulf States, can be reached by very indirect means through a comparison of the climatic factors in the vicinity of the Great Lakes with the same factors for the basin occupied by Gosiute Lake. Taken together, Lakes Superior, Huron, and Michigan cover a little more than 36 per cent of their combined hydrographic basins. In this respect they are comparable to the ancient Gosiute Lake, which covered about 36 per cent of its basin. But in all other respects these modern lakes differ decidedly from the ancient one. The mean annual temperature, the mean rates of evaporation from free water surfaces and from land and plants, and the mean annual precipitation are all probably much lower than those that prevailed in the region of Gosiute Lake during the Green River epoch. Moreover, although the three Great Lakes considered here occupy a large part of their drainage, they also have a large overflow. Apparently, Gosiute Lake had nothing in common with Lakes Superior, Michigan, and Huron except the ratio between free water surface and drainage area. But temperature and mean rates of evaporation are interdependent, and it might be assumed that if a reasonable figure for any one of these factors in the vicinity of Gosiute Lake during the Green River epoch can be estimated, then the others can be calculated. The assumption, however, that the temperature and evaporation rates bore the same relations to one another in the Gosiute Lake Basin during part of the Eocene epoch as they do to-day in the vicinity of the Great Lakes is plainly open to criticism. First, the two hydrographic basins which are compared are quite different topographically: one was more or less mountainous; the other is relatively flat. The topography would have a direct influence upon the air circulation over the lakes, and in a flat country the circulation and consequently the evaporation would be greater. Next, the Great Lakes lie in the path of frequent cyclonic storms, whereas Gosiute Lake presumably did not. Accordingly, the mean cloudiness is probably greater over the Great Lakes than it was over Gosiute Lake; this would tend to lower the evaporation rate but is probably offset to some extent by the greater storminess and therefore greater mean wind velocity over the Great Lakes. Finally, it seems likely from Brooks's analysis of geologic climates [17] that during the Eocene epoch the precipitation was seasonal rather than uniformly distributed through the year, as it now is within the north temperate cyclonic storm belt. This also would favor evaporation. Taken in the aggregate, therefore, the conditions postulated for the region of Gosiute Lake seem to indicate evaporation rates somewhat higher with respect to temperature than those that now prevail over the Great Lakes. Consequently an estimate of the mean annual precipitation necessary to balance the total evaporation from Gosiute Lake and its drainage basin in the Green River epoch based upon comparison with the present conditions in the region of the Great Lakes seems likely to err in being too low rather than too high.

Although temperature and evaporation are closely related, obviously other factors, particularly wind velocity and relative humidity, affect this relationship. Freeman[18] has constructed a curve based on data from several lakes in temperate regions to show the relation between air temperature and rate of evaporation from large free water surfaces. In order to make this a useful tie between the climate of the Great Lakes and that of the ancient Green River Lake an estimate of the mean annual temperature of the Green River epoch is necessary.

A reasonable approximation of this temperature can be deduced from an estimate made by Brooks.[19] In preparing this estimate Brooks made an elaborate statistical study of the effects upon climate produced by the relative distribution of land and sea, the altitude of the land, the direction and strength of ocean currents, and the relative amount of volcanic activity. From these studies he arrived at a mean annual temperature of 46.5° F. during Eocene time for that part of the earth north of latitude 40° north—that is, if a uniform temperature gradient is assumed, the mean annual temperature for the parallel of latitude that falls midway between 40° and 90°, namely 65° north latitude, was about 46.5° F. This is, of course, only an estimate, but he regarded it as of the proper order of magnitude. Brooks [20] also estimated the probable temperature gradient over open oceans in winter from 50° north latitude to the pole during the so-called nonglacial periods as about 0.9° F. for each degree of latitude. The temperature gradient over open oceans in the summer from 40° to 70° north latitude he estimated[21] as about 0.7° F. for each degree of latitude. Then as the mean annual temperature for a point in any given latitude is very nearly the same whether that point is located in the middle of an ocean or in the interior of a continent, it may be justifiable to assume that the mean of Brooks's winter and summer gradients (0.8° F.) calculated for conditions over open oceans during the nonglacial periods will serve as a reasonable approximation to the mean annual temperature gradient for the region of Gosiute Lake Basin during the Eocene epoch. Using this gradient and Brooks's cal-

[17] Brooks, C. E. P., Climate through the ages, p. 63, New York, R. V. Coleman, 1926.

[18] Freeman, J. R., Regulation of the Great Lakes—a report to the Chicago Sanitary District, p. 109, 1926.

[19] Brooks, C. E. P., op. cit., p. 233.

[20] Idem, p. 45.

[21] Idem, p. 42.

culated mean annual temperature of 46.5° F. for the Eocene epoch for that part of the earth between latitude 40° north and the pole[22] gives the mean annual Eocene temperature at 40° north latitude, the latitude of the Gosiute Lake Basin, as about 66.5° F. In view of the fact that the flora of the Green River formation indicates a climate comparable to that now found in the Gulf Coast States, it is interesting to note that the present mean annual temperature along the Gulf coast is about 70° F.

According to Freeman's curve [23] showing the relation between air temperature and rate of evaporation from a free water surface this temperature, 66.5° F., indicates an average annual rate of evaporation of about 54 inches from the free water surface of Gosiute Lake. However, actual evaporation rates according to Freeman's computations [24] are about 13 per cent lower than those indicated by this curve. Applying this correction, which seems to be justifiable, makes the mean annual rate of evaporation from the free water surface about 7 inches less, or about 47 inches. An assumption that the ratio between the mean annual rate of evaporation from a free water surface and the mean annual rate of evaporation and transpiration from land and plants was the same in the vicinity of Gosiute Lake as it is to-day in the vicinity of the Great Lakes probably introduces no great error, as the two processes are closely allied and, being contemporaneous, are controlled by the same conditions. Then, as calculated by a simple proportion, the mean annual rate of evaporation and transpiration from land and plants of the Gosiute Lake Basin is 32.4 inches, or approximately 32.5 inches. As the land portion represented about 64 per cent of the total area of the hydrographic basin it would require 64 per cent of 32.5 inches, or about 21 inches of precipitation over the whole basin, to balance the evaporation from the land and plants. In the same way it would require 36 per cent of 47 inches, the rate of evaporation from a free water surface, or about 17 inches of precipitation over the whole basin, to balance the evaporation from the lake surface. A total mean annual rainfall of about 38 inches is therefore indicated by this method of calculation.

A comparison of Lakes Superior, Michigan, and Huron with Gosiute Lake is shown in the table below. The data on the Great Lakes are taken from a report by Freeman,[25] except the figure for the mean annual rate of evaporation and transpiration from land and plants, which is a mean between Freeman's estimate [26] for the three lakes and that of Horton and Grunsky [27] for Lakes Michigan and Huron. Horton and Grunsky give no data on Lake Superior from which this factor can be derived. The other estimates taken from Freeman's report correspond with those of Horton and Grunsky.

[22] Brooks, C. E. P., op. cit., p. 233.
[23] Freeman, J. B., op. cit., p. 109.
[24] Idem, p. 142.

[25] Freeman, J. R., Regulation of the Great Lakes—a report to the Chicago Sanitary District, 1926.
[26] Idem, p. 145.
[27] Horton, R. E., and Grunsky, C. E., Hydrology of the Great Lakes: Report of the Engineering Board of Reviews of the Sanitary District of Chicago on the lake-lowering controversy and a program of remedial measures, pp. 15, 231, 1927.

Comparison of Lakes Superior, Michigan, and Huron with Gosiute Lake

	Drainage area	Area of water surface	Water surface in hydrographic basin	Mean annual precipitation	Mean annual temperature	Mean annual evaporation from free water surface	Mean annual evaporation from land and plants	Ratio of mean annual precipitation to total mean annual evaporation
	Square miles	*Square miles*	*Per cent*	*Inches*	*° F.*	*Inches*	*Inches*	
Lakes Superior, Michigan, and Huron	145, 120	77, 220	36. 4	29. 5	41. 7	25. 2	17. 2	1. 47
Gosiute Lake	21, 900	12, 300	36	[a]35 [e]38	[b]66. 5	[c]39 [f]47	[d]35. 5 [e]32. 5	1. 00 1. 00

[a] Calculated on the basis of comparison with Alabama.
[b] Calculated from data by Brooks on the climate of the Eocene.
[c] Estimated from data obtained at Silver Hill, Ala.
[d] Estimated from data on Coosa River, Ala., and Lenville River, N. C.
[e] Calculated on the basis of comparison with the Great Lakes.
[f] Estimated from Freeman's air temperature-evaporation rate curve.

SUMMARY OF CLIMATIC FEATURES

The annual rainfall over the basin of Gosiute Lake as indicated by a comparison with Lakes Superior, Huron, and Michigan is 3 inches greater than that indicated by comparison with the Gulf States. Nevertheless, as the two results are based on data that are wholly independent, save for the relation between water surface and land area in the two basins, they seem to indicate the probable order of magnitude of the precipitation necessary to balance evaporation—namely, 35 and 38 inches, or to be safe let us say between 30 and 43 inches. It is interesting to note that an average rainfall of about 45 inches would be necessary to balance the total evaporation to-day in the region included in Georgia, Alabama, Mississippi, and Louisiana.

Livingston and Shreve[28] have found that the most significant single expression of climatic conditions

[28] Livingston, B. E., and Shreve, F., The distribution of vegetation in the United States, as related to climatic conditions: Carnegie Inst. Washington Pub. 284, pp. 389-519, 1921.

governing the distribution of types of vegetation in the United States is the ratio between the precipitation and evaporation of the period extending from 30 days before the beginning of the frostless season to 30 days after the end of the frostless season. They call this ratio the moisture ratio. The calculated data on the climate of the Green River epoch provide no means of estimating this ratio, but the ratio between the total mean annual precipitation and the total mean annual evaporation, which is numerically nearly the same, is given in the last column of the table on page 93. These ratios must obviously be 1.0, as the mean annual precipitation of the Gosiute Lake Basin is calculated by assuming that the total evaporation balances the total precipitation. Livingston and Shreve [29] indicate an average ratio of about 1.0 for the general region of the Gulf Coast States. The ratio computed here for the region of Lakes Superior, Huron, and Michigan also agrees fairly well with the ratios given for that general region by Livingston and Shreve, who used data independent of those used by the writer. The ratios indicated by Livingston and Shreve range from a little less than 1.20 to a little more than 1.40; the ratio obtained by the writer is 1.47.

Brooks [30] gives some general conclusions regarding the probable climate and weather of the nonglacial periods that not only help to evaluate some of the assumptions used here but also add considerably to the general picture of the conditions that probably existed in southern Wyoming during the Eocene time. According to Brooks there is a critical polar temperature below which a polar ice cap forms and expands rapidly to a maximum size and above which the circumpolar oceans, if relatively free of land, remain continuously open. The great cooling effect of floating ice accounts for the rapid growth of ice caps and their stability. Once established they produce a system of cold winds that flow outward from the poles and meet opposing warm winds from equatorial regions in a zone called the polar front, which is marked by a belt of cyclonic storms. On the contrary, when the polar ocean is free from ice the warm equatorial winds, if not obstructed by land, flow far up into the polar regions and tend to effect a uniform distribution of temperature. Accordingly, there have been only two predominant types of climate during geologic time, one cool and with relatively large polar ice caps, the other warm and without ice caps. These types are stable and differ greatly in their relative distribution of temperature, storminess, wind systems, and ocean currents. The transitional type of climate in which a small ice cap forms in winter and breaks up in summer is unstable and requires temperature adjustments so delicate that it must have been rare and have existed only for brief periods.

[29] Livingston, B. E., and Shreve, F., op. cit., pl. 60, p. 336.

[30] Brooks, C. E. P., op. cit., pp. 29–251.

Brooks [31] also discusses in detail the distribution of the pressure belts and of the ocean currents and how these must change as various geographic and climatic factors change. It will suffice here merely to state his generalization that during the nonglacial periods the polar front with its zone of storminess retreats to high latitudes and the other climatic zones spread out into broader belts and so extend into higher latitudes than they now occupy. Thus the subtropical belt of high pressure moves somewhat northward, bringing generally mild weather with infrequent and erratic storms. According to this interpretation the Gosiute Lake Basin during the Eocene epoch should have had "a small rainfall during the mild winter and a long dry hot summer." [32] But as there were mountains in that region which rose several thousand feet above the general level of the continent it seems probable that the precipitation was somewhat greater there than in the adjacent regions of less relief and that in general the effect of the northward spreading of the climatic zones was less marked. Furthermore, as the region was remote from the ameliorating influences of the ocean, the seasonal extremes were probably accentuated.

Berry [33] has recently discussed rather fully the climate of the Wilcox epoch (lower Eocene) in the Mississippi embayment. For his use in this discussion Brooks calculated, for two localities, the probable differences between the present mean January and July temperatures and those that prevailed during the Wilcox epoch. According to these figures Berry computed the mean January temperature of the station on the east shore of the Wilcox embayment as about 53° F. and the mean July temperature as about 76° F. For the station on the west shore of the embayment he gives 45.5° to 49° F. for the January mean and 82° to 83° F. for the July mean. The average annual temperature indicated by the calculated temperatures at these two stations is about 65° F. Berry says that Brooks emphasized the fact that his formula was developed primarily for the study of glacial and postglacial climates and rests on the assumption that the general meteorologic system of the earth in Pleistocene time was similar to that of the present. And Berry remarks:

> That is, of course, a doubtful assumption for Wilcox time, for we have every reason to believe that there was a diminution of the polar ice caps in the Eocene. This diminution of the polar ice caps would greatly modify the general meteorologic system and would result in much higher temperatures along the Wilcox coasts than those given above, because of the fact that the subtropical anticyclonic belts would extend poleward into temperate latitudes. * * * Although the changes involved can not be stated quantitatively, their general tenor agrees admirably with the qualitative ideas regarding Wilcox climate derived from an analysis of the extensive Wilcox flora.

[31] Brooks, C. E. P., op. cit., pp. 47–95.

[32] Idem, p. 63.

[33] Berry, E. W., Revision of the lower Eocene Wilcox flora of the southeastern States: U. S. Geol. Survey Prof. Paper 156 (in preparation).

An estimate of the mean annual temperature in the vicinity of the Wilcox embayment is not directly comparable with an estimate of the mean annual temperature in the basin of Gosiute Lake, for aside from discrepancy in age, which, however, is not great, the Wilcox embayment is about 6° of latitude farther south than the basin of Gosiute Lake. Yet as the climatic zones of the Eocene were probably broader and less sharply defined than those of to-day the effect of this difference of latitude would be lessened.

The general correspondence of several factors indicated by these independent attempts at quantitative estimates of the climatic conditions during the Eocene epoch seems worth remarking. Hence it is perhaps reasonable to postulate a climate for the vicinity of the Gosiute Lake Basin during Eocene time which was characterized by cool, moist winters and relatively long, warm summers. Presumably the temperature fluctuated rather widely from a mean annual temperature that was of the order of 65° F. Likewise the rainfall varied with the seasons and probably departed rather widely from a mean annual precipitation that lay somewhere between 30 and 43 inches.

THE VARVED ROCKS

Regular rhythmic laminations have been found and studied in four different varieties of rock in the Green River formation. These four varieties—organic marlstone or low-grade oil shale, moderate-grade oil shale, rich oil shale, and fine-grained limy sandstone—differ from one another chiefly in the relative proportions of constituents that are common to all. Resistant and generally structureless organic matter characterizes each kind of rock and predominates in the beds of richest oil shale. Grains of calcium and magnesium carbonate are abundant in all the rocks and are mixed in various proportions with small angular grains of quartz, orthoclase, sanidine, plagioclase, and small micaceous flakes of clay minerals.

The usual type of varve consists of a pair of laminae, one of which is distinctly richer in organic matter than the other. The contacts between the two parts of the varve and between successive varves are generally sharp and regular. (See pl. 11.) In some beds, however, especially the richer oil shales, the boundaries are irregular and less plain.

The other type of varve, which is restricted to beds of very fine-grained limy sandstone, is really a modification of the predominant type, though in size and appearance the two are rather unlike. The varves in the sandstone differ from those in the other rocks in that the mineral layers and to a less extent also the organic layers are considerably thickened by the admixture of more or less silt or very fine sand. They also differ from nearly all the varves in the other rocks in that they show a gradation in grain size from coarse at the bottom to fine at the top. Near the base the grains average about 0.02 millimeter in diameter and at the top they average between 0.004 and 0.006 milimeter. The upper, fine-grained part of these varves contains the organic matter, but the boundary between this organic part and the lower, mineral part is in most varves transitional. On the other hand, the contact between successive varves—that is, between the fine-grained organic layer and the next mineral layer above—is invariably sharp and clearly defined. (See pl. 12, *A*.) The average thickness of 32 of these varves is 1.16 millimeters, but the range in thickness is great, from 0.6 to 9.8 millimeters. The structure of some of the thinnest of these varves is nearly identical with that of the predominant varve type, even to the relatively sharp separation of the organic layer from the mineral layer.

The varves in the beds of organic marlstone and oil shale, which are of the predominant type, are consistently thinner than those in the sandstone. (See pl. 12, *B*.) The average thickness of 268 varves measured in four different beds of organic marlstone from localities in Colorado and Wyoming is 0.167 millimeter. The extreme range in thickness is from 0.014 to 0.37 millimeter, but the thickest varves are unusually coarse grained and the thinnest are correspondingly fine grained. In two of the marlstones the mineral-rich parts of the varves are two or three times as thick as those rich in organic matter. Another marlstone, collected near the mouth of Piceance Creek, Colo., is unusual in that the organic parts of the varves average about one and one-half times as thick as the mineral parts. A fourth marlstone, which is virtually a limestone, is unusually coarse grained and shows a nearly uniform diminution of grain size upward from the base to the top of each mineral layer. At the base most of the grains are about 0.034 millimeter in diameter, but at the top the average grain size is between 0.003 and 0.005 millimeter. The parts of these varves rich in organic matter are excessively thin, and the mineral layers contain only the minutest quantities of organic matter. This distribution of organic matter is also abnormal, for most commonly the mineral layers of marlstone varves contain sufficient organic matter to give them a perceptible yellowish tint, and the organic layers contain enough mineral grains to have a distinctly granular aspect.

In both the rich and the moderately good grade of oil shale the varves are of one type and differ only in average thickness. (See pls. 13, *A* and *B*; 14, *A*.) In the richest beds of oil shale, those whose yield ranges from about 35 gallons to more than 60 gallons to the ton, the organic parts of the varves generally contain a relatively small quantity of mineral matter and the mineral-rich parts a relatively large quantity of organic matter.

In the mineral-rich laminae carbonate grains and flakes of clay minerals predominate, though clastic grains of quartz and feldspar are more plentiful than in the organic laminae. The minerals most common in the organic laminae are minute anhedral and euhedral carbonate grains and clay minerals. Sanidine grains and angular splinters of quartz, glass, and plagioclase, which are presumably of volcanic origin, may be found in either part of the varve without apparent relation to its composition. The average thickness of 143 varves measured in four different samples of very rich oil shale from a place near the head of Clear Creek, Colo., is 0.037 millimeter. The maximum range in thickness is from 0.014 to 0.153 millimeter. It is worthy of note that the average thickness is near the minimum. In these richest oil-shale beds, three of which are from the Mahogany Ledge, in sec. 9, T. 5 S., R. 98 W., Garfield County, Colo., the organic parts of the varves equal or exceed a little in thickness the mineral-rich parts. In the bed yielding nearly 60 gallons of oil to the ton the organic laminae are themselves more or less plainly laminated. (See pl. 14, *A*.) The possible significance of these most minute rhythms is considered on page 102, where the origin of the varves is discussed.

In moderately good oil shale—that is, shale yielding 15 to 35 gallons of oil to the ton—the varves range from 0.03 to 0.114 millimeter in thickness; the average thickness is 0.065 millimeter. This figure is based on the measurement of only 18 varves in a single specimen from the type locality of the Green River formation at Green River, Wyo. The varves of this specimen, however, are particularly interesting because they provide a means for measuring a larger cycle recorded in the same bed. (See pl. 14, *B*.) For some unknown reason easily measurable varves in oil shale of this grade are much more rare than in beds of very rich oil shale or in the much leaner beds of organic marlstone.

Laminations that may possibly be varves in rock of another type are represented by the successive regular layers of algal deposits in many of the algae reefs of the Green River formation. These average about 6 millimeters in thickness. As they have been discussed by the writer in an earlier report [34] and as they are quantitatively unimportant they will not be treated again here, although they will be used in making estimates of the duration of the Green River epoch.

The above descriptions show that the varves range in thickness from 0.014 millimeter in the beds of richest oil shale to 9.8 millimeters in the fine-grained sandstone. The average thickness, weighted according to the quantity of each type of rock in the Green River formation, is about 0.18 millimeter.

[34] Bradley, W. H., Algae reefs and oölites of the Green River formation: U. S. Geol. Survey Prof. Paper 154, pp. 203–223, 1929.

PROBABLE TIME VALUE OF THE RHYTHMIC LAMINAE

ANALOGY WITH MODERN LAKE DEPOSITS

In the preceding description of the regular, rhythmic laminations in the rocks of the Green River formation the time value of the rhythm has, for convenience of exposition, been assumed as one year, and accordingly the couplets of laminae have been called varves. The evidence upon which this assumption is based is drawn from several more or less independent sources. First of all is the analogy between the lamination found in these Eocene lake beds and that found in the deposits of modern lakes.

Perfiliev [35] has shown that the black organic mud now forming in Sakski Lake, in northern Crimea, is distinctly varved. Each varve consists of a thin black lamina of organic matter and a thicker gray lamina that consists either of fine-grained quartz sand of eolian origin or of gypsum. The black organic layers consist chiefly of minute aquatic organisms that reached their peak of production in the spring. The average thickness of these varves is 1.3 millimeters. If compacted to one-tenth of this thickness (0.13 millimeter) they would be of very nearly the same size as the varves in the marlstone beds of the Green River formation, which average 0.167 millimeter. Perfiliev, by a refined technique of sampling, was able to count with confidence 1,620 varves in a layer of ooze about 2 meters thick, and drilling by I. Mushketov in 1894 shows a thickness of lake deposits that suggests the possibility of a record as long as 16,000 years. Sakski Lake is highly saline, and the alternating laminae of gypsum and organic matter in its deposits are of unusual interest. The gypsum layers themselves are 0.1 to 0.2 millimeter thick. The writer recently found thin laminae of gypsum that alternate with laminae of organic matter in a sample of oil shale which he examined for C. H. Dane, of the United States Geological Survey. This sample came from a thin lens of oil shale in the gypsiferous series of southeastern Utah, which is of Carboniferous age. Its couplets of laminae average about 0.06 millimeter in thickness and very probably are varves.

Nipkow [36] found the ooze on the bottom of the Lake of Zurich below a depth of 295 feet to consist of alternating laminae of lime-poor organic ooze and of microgranular calcite which contained but little organic matter. The organic layers were derived largely from winter maxima in the production of a plankton alga, *Oscillatoria rubescens*. The calcite layers he regards as having been precipitated in the summer, in part

[35] Perfiliev, B. V., Ten years of Soviet science, pp. 402–403, Moscow, 1927. The writer is indebted to Miss Taisia Stadnichenko for calling his attention to this article and for translating it from the Russian.

[36] Nipkow, Fritz, Vorläufige Mitteilungen über Untersuchungen des Schlammabsatzes im Zürichsee: Zeitschr. Hydrologie, 1920 (cited by Collet, L. W., in Les lacs, pp. 267–269, 1925).

through the increased temperature of the water and in part through the activity of plants. The similarity between this annual lamination and the lamination in the beds of the Green River formation is at once evident. It is germane to point out an additional minor feature of the analogy—namely, that the calcite in the organic layers is in euhedral crystals, whereas that in the other layers is not. Euhedral carbonate crystals are also more or less distinctive of the organic laminae in the richer oil shale beds of the Green River formation. The varves in the deposits of the Lake of Zurich average about 3 millimeters in thickness, but a few are as much as 10 millimeters thick. After compaction under a load of 1,000 or 2,000 feet of rock these thicknesses would be considerably reduced. According to a rough calculation, W. W. Rubey, of the United States Geological Survey, estimated that layers of such fine-grained material would be compacted at least to one-third and perhaps to more nearly one-tenth of the thickness that they had when buried only 1 or 2 feet. Even if reduced to one-tenth these varves would indicate a rate of accumulation in the Lake of Zurich about three times as fast as that indicated by the lamination in the most nearly comparable rock of the Green River formation, the organic marlstone.

The bottom deposits of McKay Lake, Ottawa, were recently studied by Whittaker,[37] who found in them a clearly defined seasonal lamination. He found in his core samples, which were somewhat compacted, a bed of ooze 24 centimeters thick that consisted of 440 pairs of laminae. One layer of the pair was chocolate-brown and consisted of organic matter (algae of various types and a few sponge spicules); the other layer was gray and consisted largely of particles of marl washed in from the shore. The organic matter was deposited during the summer and fall, and the mineral layer was washed in from the shore during the spring rains. Thus the couplet represents an annual deposit. These paired laminae or varves are about 0.43 millimeter thick. In other parts of the core samples the varves, which could not be clearly diffentiated, were about 0.127 millimeter thick. These varved deposits formed in about 32 feet of water, the deepest part of the lake. They are plainly of the same character as the varves in the Lake of Zurich and the alternating laminae of certain beds in the Green River formation. If compacted to one-tenth of their original thickness the thicker varves of McKay Lake would be about comparable to the couplets of similar laminae in the oil-shale beds, which are 0.037 to 0.065 millimeter thick. The thinner varves of the deposits of McKay Lake, if similarly compacted, would equal almost precisely the thinnest couplet of laminae measured in the oil shale, which is 0.014 millimeter thick.

Coit and Collet [38] found that the deposits in certain parts of the lake of Geneva, Switzerland, are varved. Each varve consists of two layers, a light coarser-grained one that represents the summer deposit and a dark finer-grained one that represents the winter deposit. These couplets average about 2.5 millimeters in thickness.

Holmboe, according to Osvald,[39] was able to discern varves in an algal ooze associated with peat deposits in Norway. Lundqvist,[40] in commenting upon the varved algal oozes that Stalberg found in Lake Wetter, says he has found that varves are really not uncommon in the algal oozes of certain lakes in southern Sweden.

There is undoubtedly an analogy between the varves in these recent lake deposits and the rhythmic laminae in the rocks of the Green River formation.

PROBABLE THICKNESS OF ANNUAL LAMINAE AS COMPUTED FROM THE LOADS OF PRESENT STREAMS

Another kind of evidence bearing on the time value of the rhythmic laminae in the rocks of the Green River formation is different and is, moreover, independent of that just discussed. It consists of an estimate of the probable average thickness of the deposits supplied each year to Gosiute Lake.

In large inland lakes sedimentation must be nearly equivalent to denudation within the hydrographic basin. In a closed basin this balance is manifestly perfect. But in a lake that overflows either continuously or periodically the proportion of the suspended material that is deposited depends upon several factors. In general, the greater the volume of a lake with respect to the volume of water passing through it the more efficient it is as a settling basin. Other factors, however, affect this efficiency. For example, if the mouths of streams that feed the lake are close to the outlet suspended material is more likely to be carried beyond the lake. Again, the number and size of feeding streams would be significant, for, in general, numerous small streams would distribute the detritus more uniformly and thus facilitate deposition, whereas a single large stream, especially one emptying into a shallow lake, would be likely to scour out a channel and in this way carry much of its load through to the outlet. Upper Klamath Lake, Oreg., is a good illustration. This lake has a remarkably uniform depth

[37] Whittaker, E. J., Bottom deposits of McKay Lake, Ottawa: Roy. Soc. Canada Proc. and Trans., 3d ser., vol. 16, sec. 4, pp. 141-156, 1922.

[38] Coit, G. E., and Collet, L. W., Nouvelles recherches sur la sédimentation dans le Lac de Genève: Cong. internat. géographie Compt. rend., vol. 2, pp. 59-69, Cairo, 1925.

[39] Osvald, H., Till gyttjornas genetik: Sveriges geologiska undersökning Årsbok, vol. 15, no. 3, p. 36, 1921.

[40] Lundqvist, G., Geol. Fören. Förh., Band 46, p. 193, 1924 (review of Stalberg, Nils, Några undersökningar av Vättergyttjans beskaffenhet: Södra Sveriges Fiskeriförening, 1923).

of about 6 or 8 feet everywhere except along the west shore, where Klamath River has scoured out a narrow channel 20 to 40 feet deep. The current in this channel is strong enough to keep the bottom bare of any detritus except coarse sand, gravel, and boulders.

The efficiency of a lake in collecting the dissolved material in the water it receives depends in part upon the same factors that affect the deposition of the suspended material. More particularly, however, it depends upon the relation between the volume of water supplied to the lake and that lost by evaporation and upon the difference between the temperature of the lake water and that of the feeding streams. The greater the proportion of water lost by evaporation the more nearly the conditions approach those of a closed basin, wherein all the dissolved material is retained. The temperature differential between lake and stream water affects principally the carbonates, whose solubility, by reason of its dependence upon the solubility of carbon dioxide, is decreased as the temperature of the water rises; and as carbonates constitute the principal dissolved load of most streams this temperature factor is significant.

As Gosiute Lake covered a large part of its hydrographic basin and seemingly offered favorable conditions for the deposition of both suspended and dissolved material it might be reasonable to expect that sedimentation in the lake nearly equaled the erosion in its basin. Accordingly, if the thickness of the annual deposit can be calculated and expressed in terms of compacted rock it should be comparable to the observed thickness of the varves in the rocks of the Green River formation. However, as the rate of erosion during the Eocene is unknown and can only be inferred by analogy with the rate in certain modern regions this calculation involves considerable uncertainty. Also, the estimate of the relation between the size of the lake and its hydrographic basin is subject to revision, and use of that estimate may introduce still other errors. Consequently, a calculation of sedimentation based upon the present rate of erosion can not be expected to show more than the probable order of magnitude of the annual layers.

Dole and Stabler [41] have assembled many determinations of the suspended and dissolved loads of rivers in practically all parts of the United States. From these figures they have computed the mean loads of streams in the several drainage basins of the country. In order to apply these data to the present problem it is necessary first to choose the subdivision of the country whose climatic conditions and topography seem most nearly to approximate those that prevailed in the basin of Gosiute Lake.

Two of the primary drainage basins outlined by Dole and Stabler [42]—namely, the southern Atlantic and eastern Gulf of Mexico drainage basins—together seem to offer conditions of rainfall, vegetation, and topography that coincide most nearly with those postulated for the basin of the ancient Gosiute Lake. Although the average relief is probably less in these Southern States the rainfall is probably greater, and these two factors would oppose and tend to compensate each other. If, therefore, the average loads of the streams that fed Gosiute Lake are assumed to have been the same as those carried to-day by the streams in these Southern States, the thickness of the average annual deposit upon the lake bottom can be computed from the relative areas of the lake and its hydrographic basin. (See fig. 14.) The average annual load of dissolved solids for the streams of the South Atlantic and eastern Gulf Coast States as given by Dole and Stabler [43] is equivalent to 105 tons per square mile of the total drainage area, and the average annual load of suspended material is equivalent to 160 tons per square mile of drainage area. Accordingly, as the land area of the Gosiute Lake Basin was approximately 22,000 square miles, the streams working at these assumed rates would have delivered each year to the lake, partly in solution and partly in suspension, about 5,830,000 tons of material. Evenly distributed over the maximum area of the lake, about 12,300 square miles, this would be equivalent to about 474 tons per square mile. The average rock density [44] of 31 different beds of very low-grade oil shale (virtually marlstone with only a small quantity of organic matter) from the Green River formation is 2.24. The average density of the rock free from organic matter may be taken as 2.3. According to this, the annual deposit of 474 tons of material per square mile of lake bottom would be equivalent to a layer 0.074 millimeter thick of rock comparable in degree of compaction, mineralogy, and porosity to the rocks of the Green River formation. This calculated thickness approximates the average thickness of the varves in the marlstone and oil-shale of the Green River formation—marlstone about 0.167 millimeter, moderately good grade oil shale about 0.065 millimeter, and richest oil shale about 0.037 millimeter.

Certain corrections should be applied to the thickness of the annual deposit as calculated here from Dole and Stabler's data, and most of these corrections would make the thickness greater. Dole and Stabler did not consider the rolling load of the streams, and presumably part of the material so moved found its way out into the fine-grained laminated deposits. Again, in making the present computations it was assumed that the material was distributed uniformly over the entire area of the lake bottom; but this is improbable, as there are reasons for believing that the lake had rather broad marginal zones beyond which the very fine material, like that considered here, would

[41] Dole, R. B., and Stabler, Herman, Papers on the conservation of water resources—Denudation: U. S. Geol. Survey Water-Supply Paper 234, pp. 78-93, 1909.

[42] Idem, pp. 84-87.

[43] Dole, R. B., and Stabler, Herman, op. cit., p. 85.

[44] Rock density is distinguished from the specific gravity of the minerals making up the rock in that it takes into consideration the rock's porosity.

have been transported by wave and current action, thus making the thickness of the deposit in the middle of the lake greater than that on the margins. The varved rocks studied came from the middle of the basin. Considerable volcanic ash fell into the lake, and no allowance for this increment has been made. Finally, the quantity of organic matter, especially in the richer oil-shale beds, is considerable, but this was allowed for in part by assuming for the material a rock density higher than the observed rock density of many low-grade oil-shale beds.

Only one correction that should be applied would tend to reduce the thickness of the annual deposit as calculated here on the basis of present stream loads. This correction rests upon the probability that some fraction of both dissolved and suspended loads was lost from the lake by overflow. However, this loss was probably small for Gosiute Lake, because, presumably, most of the annual increment of water was removed from the enormous surface of the lake by evaporation rather than by overflow.

These corrections, though real enough, can hardly be evaluated quantitatively. They seem merely to indicate that the thickness of the annual deposit as calculated from Dole and Stabler's figures is probably too small. But it is very unlikely that, even in the aggregate, they would reach the same order of magnitude as the errors inherent in the assumptions which have been necessary to arrive at a figure for stream loads in the drainage basin of Gosiute Lake and upon which these calculations chiefly depend. Consequently such refinements as would be involved in applying these corrections may be neglected. But despite the crudity of the method the thickness of the annual deposit calculated in this way is of the same order of magnitude as the observed thicknesses of the varves in the Green River formation. This correspondence, therefore, also indicates that the time value of the rhythmic lamination in the Green River beds is one year.

ESTIMATED RATE OF ACCUMULATION OF THE VARVED ROCKS

From the measurements of the varves observed in the Green River formation estimates have been made of the average rate of accumulation of each kind of rock. These are given in the table below.

Estimated rate of accumulation of the varved rocks of the Green River formation

	Thickness of varves (millimeters)		Time necessary to accumulate 1 foot (years)
	Range	Average	
Sandstone, fine grained	0.6 –9.8	1.16	250
Marlstone and related rocks	.014– .370	.167	2,000
Oil shale, moderately good (yielding 15 to 35 gallons a ton)	.03 – .114	.065	4,700
Oil shale, rich (yielding more than 35 gallons a ton)	.014– .153	.037	8,200
Weighted average		.18	2,200

EFFECT OF NATURAL RHYTHMS ON THE CHARACTER OF SEDIMENTS

In any attempt to evaluate thin rhythmic lamination the power of various natural rhythms [45] to affect the character of sediments should be considered.

Periodic storms, as for example the cyclonic storms of to-day, might perhaps have produced a rhythmic lamination in the lake deposits. Storm waves in stirring up the bottom ooze would have thrown it into suspension, from which it would have settled gradually, and as the ooze consisted of particles of calcite and of organic matter, whose respective specific gravities are very different, a density stratification must have resulted. Moreover, this stratification would have been similar to that observed in the rocks of the Green River formation, for the lower layer would have consisted chiefly of the heavier carbonate particles and the upper layer chiefly of organic matter. But when the next storm occurred this same material, or at least part of it, plus whatever sediment had since come down, must have been again thrown into suspension. Upon settling it would resume a density stratification, but despite the fact that there had been two storms only one couplet of laminae would have been perfect, and it would have rested upon the remnants of the preceding couplet.

For storms to have produced a regular series of uniformly perfect pairs of laminae such as that illustrated in Plate 11 it is necessary to postulate one or the other of two conditions—(1) that each successive storm stirred up only the ooze that had accumulated since the completion of the preceding couplet of laminae, or that it stirred up the fresh supply and in addition one or more complete couplets of laminae, but never only part of a couplet; or (2) that between the formation of each couplet of laminae and the next storm the stratified material became so much indurated that it was unaffected by subsequent storm waves. Either of these necessary postulates seems to the writer adequate to rule out periodic storms as agents competent to produce the regular, uniform series of paired laminae observed in the rocks of the Green River formation. Moreover, there is a small but significant difference between pure density stratification and that observed in most beds in the Green River formation. In pure density stratification the grain size should diminish regularly upward from the base of each pair of laminae, but the grains in most laminae in the Green River beds are practically unsorted, even though the distinction between the layers rich in organic matter and those poor in organic matter is very sharp. If the material had been thrown into suspension in saline water coagulation would have prevented a clean separation of the various sizes of mineral grains, but at the same time it would just as surely have prevented a distinct separation of organic and mineral constituents.

[45] Barrell, Joseph, Rhythms and the measurement of geologic time: Geol. Soc. America Bull., vol. 28, pp. 824–834, 1917.

Barrell [46] considered rhythms of another kind, which he called orbital rhythms and which depend on motions in the solar system. Of this sort only two, the alternation of day and night and of summer and winter, are short enough to be considered here. Tides in the lake obviously must have been too slight to affect the sedimentary record. Diurnal temperature changes might conceivably have been large, and evaporation rates, which depend in part on the temperature, would have fluctuated in a corresponding manner; yet it is difficult to see how these changes could have affected perceptibly the rate of deposition, arrangement, or composition of the sediments.

The seasonal changes, however, have an evident effect upon the factors that influence sedimentation. Erosion rates and consequently rates of supply of material vary with the changing rainfall. The rate at which carbonates precipitate depends, in part at least, upon the temperature of the water, and the temperature of the water varies chiefly with the seasons. The rates of growth of planktonic organisms are controlled closely by temperature, and consequently the supply of organic matter in lakes also changes with the seasons. Furthermore, the thermal stratification in lakes (see p. 102), which is a distinctly seasonal phenomenon, may have a marked influence upon the differential settling rates of various kinds of sediment. Apparently, therefore, the march of the seasons is the only natural cycle of the right order of magnitude which combines pronounced effects upon the factors controlling sedimentation with the regularity necessary to explain adequately the long series of thin, uniform laminae found in the rocks of the Green River formation. The hypothesis that the couplets of laminae represent an annual rhythm and are therefore varves affords also a satisfactory basis for explanation of the origin of their bipartite character.

ORIGIN OF THE VARVES

BIOCHEMICAL REACTIONS

Of the three hypotheses that have been considered to explain the rhythmic alternation of carbonate and organic laminae of the varves in the Green River formation one depends upon biochemical reactions.

Abundant evidence has been assembled to prove conclusively that the organic matter in the oil shale passed through a stage of putrefaction. The ooze or sapropel which was later lithified into oil shale must have been wholly analogous to the black fetid organic oozes now forming in both fresh and salt lakes. The characteristics of such lacustrine ooze or "Faulschlamm" have been described with considerable detail by Naumann,[47] Potonié,[48] Nadson,[49] Wesenberg-Lund,[50] and others. In the microbian decomposition of such ooze the albumin is broken down, and among other compounds ammonia and hydrogen sulphide are formed. The ammonia immediately reacts with the HCO_3 ions in solution and precipitates the normal carbonates of calcium and magnesium. The products of this reaction assume particular significance for the problem in hand as they are interpreted by Nadson [51] in describing the bottom deposits of Weissowo-Salzee, which is near the city of Slawjansk, Kharkof, Russia. He isolated four bacteria and three fungi which play the principal part in the formation of the fetid black ooze in that lake.

These organisms are *Bacillus mycoides* Flügge, *Proteus vulgaris* Hauser, *Bacterium albo-luteum* Nadson, *Bacillus salinus* Nadson, *Actinomyces alba* Gasperini, *A. verrucosus* Nadson, and *A. roseolus* Nadson. They all bring about the decomposition of albumin with the release of abundant ammonia and hydrogen sulphide. Nadson finds that by the activity of these organisms thin laminae of calcium and magnesium carbonates are deposited on the surface of the ooze. These laminae of microcrystalline carbonates are apparently analogous to those in the oil shale and marlstone. However, to explain why these laminae of mineral grains alternate with organic layers, as they do in the oil shale of the Green River formation, some periodic change of rate, either in the activity of the organisms or in the supply of the organic and inorganic constituents, must be postulated. Thus this hypothesis is not entirely adequate to explain the dual lamination of the varves without some such periodicity of supply, but it provides another mechanism by which to explain the occurrence of thin indefinite lenses of microgranular carbonates mixed with organic matter. Lenses of this sort obscure the varves in a few beds of moderately good oil shale. (See pl. 13, *B.*)

DIFFERENTIAL RATE OF SETTLING

Another hypothesis to account for the rhythmic alternation in the varves depends upon the difference in density between the organic matter and the carbonate grains.

The greater part of the organic matter presumably came down as a rain of minute planktonic organisms

[46] Barrell, Joseph, op. cit., pp. 827–829.

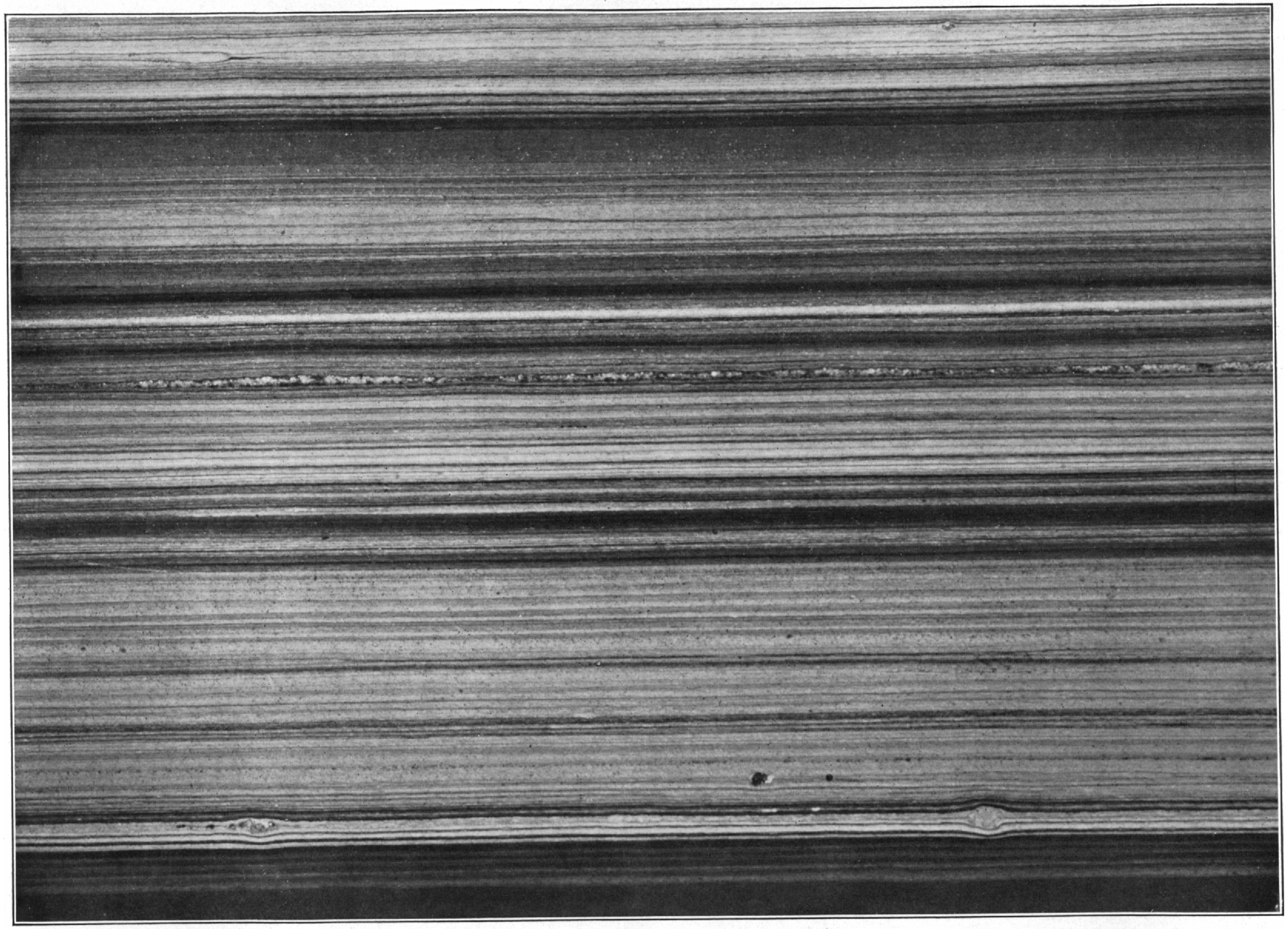
[47] Naumann, Einar, Die Bodenablagerungen des Süsswassers: Archiv für Hydrobiologie und Planktonkunde, vol. 13, pp. 100–186, 1921–22.

[48] Potonié, H., Die rezenten Kaustobiolithe und ihre Lagerstätten: K. preuss. geol. Landesanstalt Abh., vol. 55, pp. 9–36, 49–232, 1908.

[49] Nadson, S., Die Mikroorganismen als geologische Faktoren: Bot. Centralbl., vol. 96, pp. 591–593, 1904.

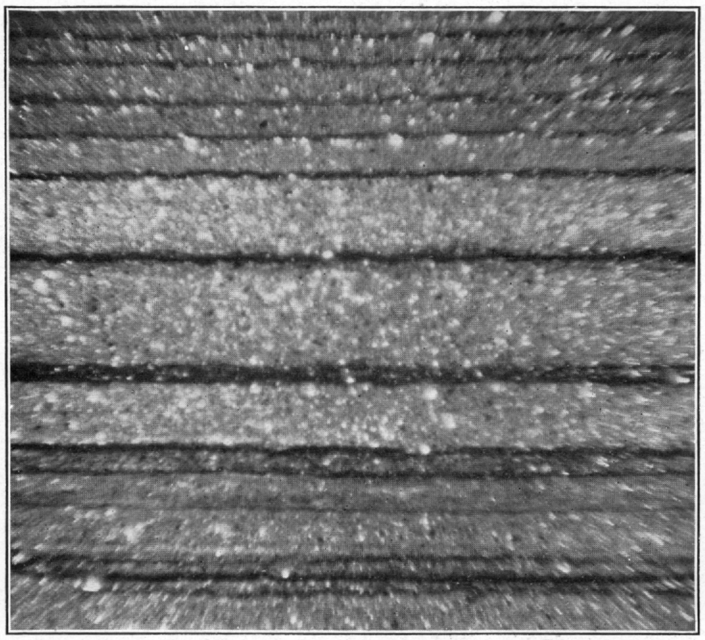
[50] Wesenberg-Lund, C., Lake lime, pea ore, and lake gyttja in Danish inland lakes: Dansk geol. Foren., No. 7, pp. 159–168, 1901.

[51] Nadson, S., op. cit., pp. 591–593.



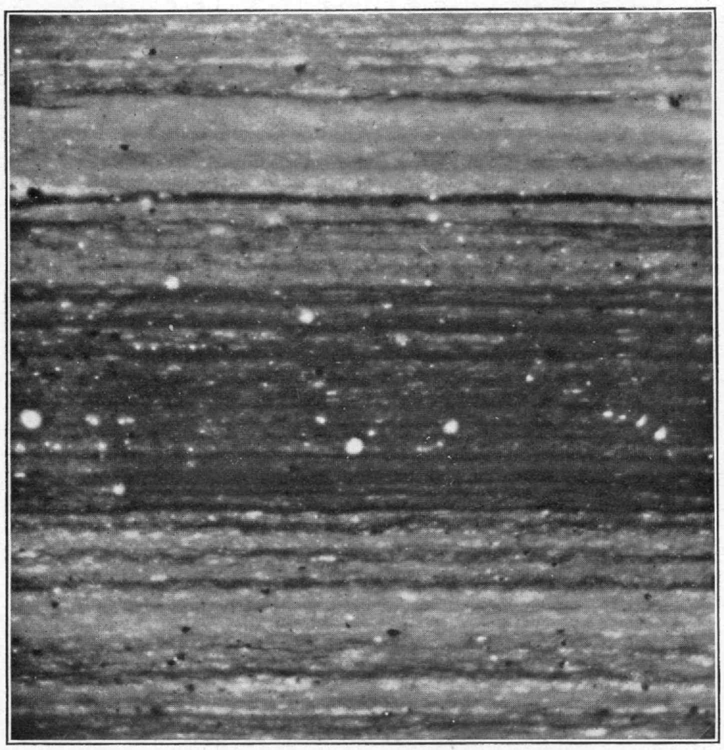
PHOTOGRAPH OF A POLISHED SPECIMEN OF ORGANIC MARLSTONE SHOWING TYPICAL VARVES OF THE GREEN RIVER FORMATION

The darker parts of the rock contain the most organic matter. At the base of the broad light-colored band that crosses the middle of the photograph the individual laminae of the varves can be most readily distinguished. Specimen from a bed about 1,800 feet above the base of the formation on Clear Creek, in sec. 9, T. 5 S., R. 98 W., Garfield County, Colo. Enlarged 4 diameters



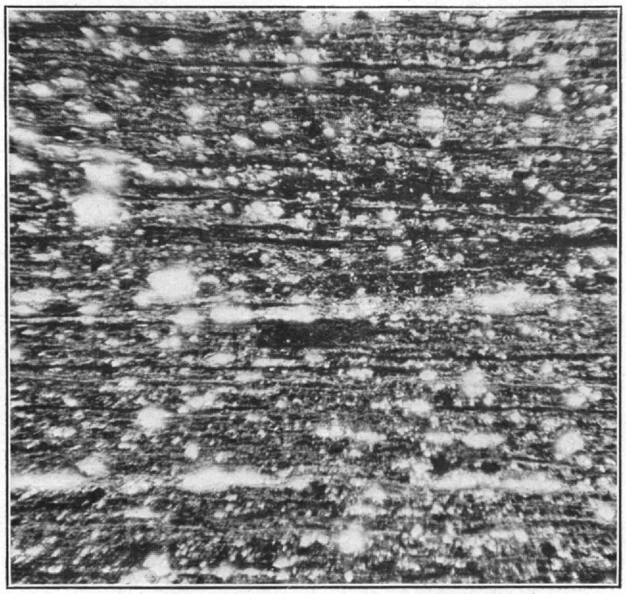
A. PHOTOMICROGRAPH OF A FINE-GRAINED LIMY SANDSTONE SHOWING THICK VARVES

The light bands are sandy and unusually thick. The dark bands consist largely of organic matter. Specimen from a bed in the upper part of the Green River formation at the junction of Parachute Creek and East Middle Fork of Parachute Creek in sec. 18, T. 5 S., R. 95 W., Garfield County, Colo. Enlarged 37 diameters



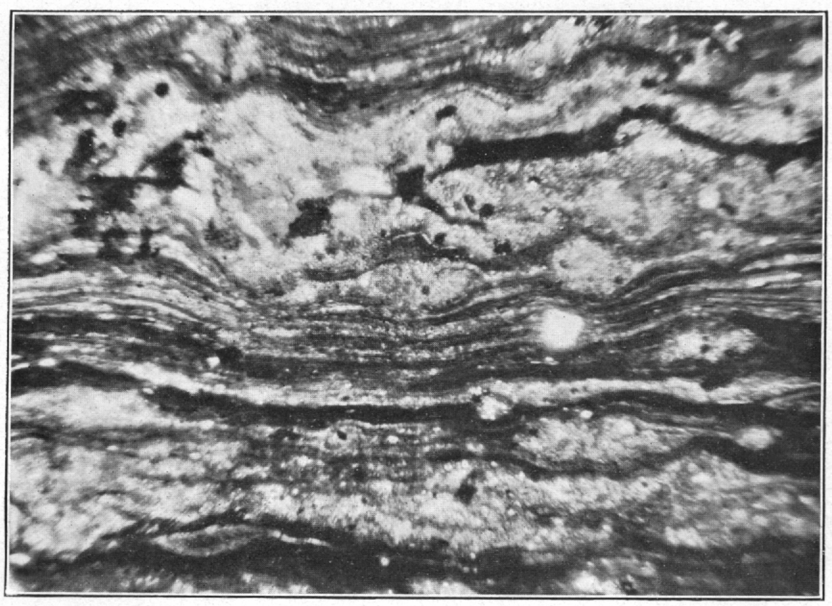
B. PHOTOMICROGRAPH OF A PART OF THE ORGANIC MARLSTONE SHOWN IN PLATE 11, ILLUSTRATING THE REGULAR SPACING OF THE DARK ORGANIC LAMINAE

The thin section was cut abnormally thick so as to bring out the more feebly colored organic layers. Enlarged 18 diameters



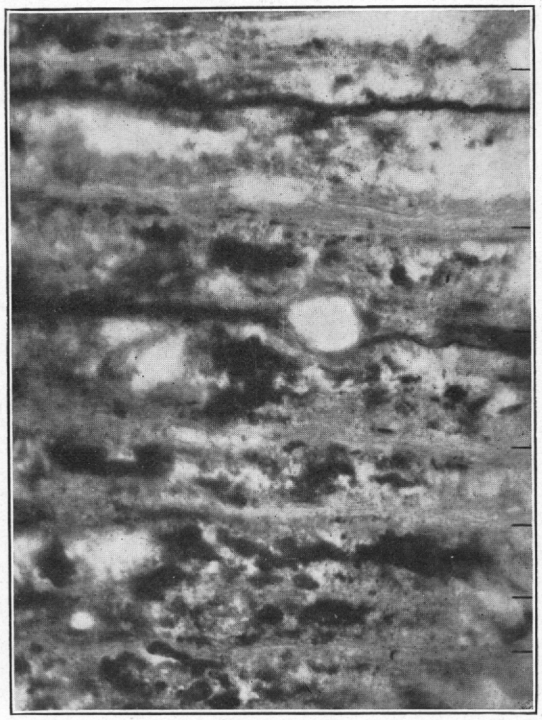
A. PHOTOMICROGRAPH SHOWING THIN VARVES IN A VERY RICH OIL SHALE

From the Mahogany Ledge near the head of Clear Creek, in sec. 9, T. 5 S., R. 98 W., Garfield County, Colo. The black parts are organic matter, and the light-gray areas are mineral grains. Enlarged 54 diameters



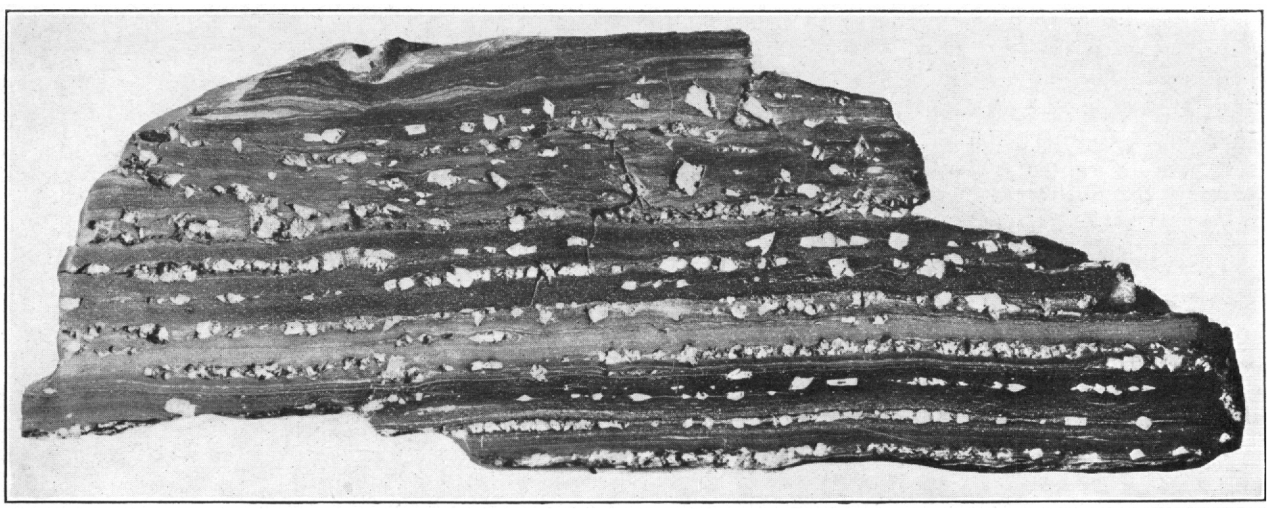
B. PHOTOMICROGRAPH OF A RICH OIL SHALE SHOWING GROUPS OF VARVES DISTORTED BY SMALL CONCRETIONARY LENSES OF CARBONATE GRAINS

Enlarged 54 diameters



A. PHOTOMICROGRAPH SHOWING DETAIL OF THE VARVES IN THE RICH OIL SHALE ILLUSTRATED IN PLATE 13, *A*

The organic laminae, indicated by marks in the margin, are themselves finely laminated. The mineral laminae contain considerable organic matter, but they are readily distinguished by their coarser grain and greater thickness. Enlarged 320 diameters



B. PHOTOGRAPH OF A POLISHED SPECIMEN OF MODERATELY RICH OIL SHALE

From Green River, Wyo. Shows regularly spaced layers of calcite-filled glauberite (?) cavities. Natural size which might have ranged from 1 or 2 microns to several millimeters in maximum dimension, though if an analogy with Lake Mendota in Wisconsin is significant it might be inferred that most of the organisms were less than 60 microns in diameter. Birge and Juday[52] define the nannoplankton as all the organisms which will pass through No. 20 silk bolting cloth (whose openings average about 60 microns in diameter) and state that in Lake Mendota these minute organisms constituted about five times as much organic matter as the larger net plankton.

It is reasonable to believe that the precipitation of carbonates accompanied the sedimentation of the remains of plankton organisms. And if the assumptions are made that organisms and carbonate grains began to settle from the same water stratum, that the particles of both sorts of material settled as spheres, that the carbonate grains averaged about 5 microns in diameter (their present size), and that the organisms averaged about 50 microns in diameter and had an average specific gravity of 1.05, then the variables of Stokes's law[53] show that the carbonate grains, despite their small size, must have settled many times more rapidly than the lighter organic matter. Consequently there would have been complete separation of the constituents into two layers even in shallow water. Actually, however, many plankton organisms are not spherical but flattened, and others have various appendages to facilitate floating; furthermore, such organisms immediately after death are subject to the attacks of other organisms that tend to disintegrate and decompose them as they settle toward the bottom. These two factors would combine to accentuate a density stratification. But if the deposition of both constituents was continuous then they must have accumulated together in an unstratified mixture after the period required for the first particles of the more slowly settling material to reach bottom. Clearly, differential rate of settling was a factor, yet as the varves consist of laminae rich in mineral matter alternating with laminae rich in organic matter it does not account for all the facts unless a periodic supply of either or both constituents is assumed.

SEASONAL CHANGE IN RATE OR KIND OF DEPOSITION

The third hypothesis advanced to explain the regular alternation of the material in the varves assumes a periodic change in rate either of the activity of microorganisms or of supply of organic and inorganic constituents.

In inland lakes there is a periodic production of aquatic organisms, which, if the diatoms are ignored, reaches its maximum in the summer. For the purpose in hand the diatoms can well be ignored, for none have been found in the Green River formation. The maximum growth of most other organisms occurs in summer and falls off to a negligible minimum in winter. In lime-rich lakes finely divided calcium carbonate is precipitated in summer but only slightly in winter; it may even be partly dissolved in winter. The maximum precipitation in summer may be ascribed to two principal causes—(1) the increased temperature of the water, which reduces the solubility of carbon dioxide and consequently the solubility of calcium carbonate, and (2) the activity of submerged green aquatic plants, which abstract carbon dioxide from the water and thereby also precipitate calcium carbonate. Hassack[54] and later Chambers[55] demonstrated how effectively algae can remove carbon dioxide from solutions of calcium bicarbonate. Powell[56] pointed out the value of plankton algae in precipitating carbonates from reservoir waters. Nauman,[57] Wesenberg-Lund,[58] Thienemann,[59] and others have emphasized the large production of aquatic organisms, both plant and animal, consistently found in lakes of moderate depth which contain an abundance of calcium in solution. Thus in certain modern lakes there is a pulsating supply of both organic matter and finely divided calcium carbonate available for sedimentation. Moreover, the pulses are seasonal, and as was pointed out on pages 96 and 97, they lead to the formation of varves remarkably similar to those of the Green River formation.

This similarity suggests the likelihood that the couplets of laminae in the rocks of the Green River formation were produced by a comparable pulsating supply of materials in Gosiute and Uinta Lakes.

If it is assumed that the greater part of the rainfall came in the winter and that the summers were relatively dry and warm, then the streams would have brought to the lake the greater part of their annual supply of dissolved and suspended mineral matter during the winter and spring, when the temperature of the lake water probably did not differ greatly from that of the streams. But during the summer the temperature of at least the surface layer of lake water must have risen, and the higher it became the more it must have favored the precipitation of carbonates and also the growth of minute aquatic organisms. Thus it seems that large supplies of both organic matter and carbonate particles would have been produced at the same time of the year. Yet probably the maximum pre-

[52] Birge, E. A., and Juday, C., The inland lakes of Wisconsin: Wisconsin Geol. and Nat. Hist. Survey Bull. 64, pp. 63, 93, 1922.

[53] Twenhofel, W. H., and others, Treatise on sedimentation, p. 40, 1926.

[54] Hassack, Carl, Über das Verhältniss von Pflanzen zu Bicarbonaten und über Kalkincrustation: Bot. Inst. Tubingen Untersuchungen, vol. 2, pp. 467–473, 1888.

[55] Chambers, C. O., The relation of algae to dissolved oxygen and carbon dioxide with special reference to carbonates: Missouri Bot. Garden Ann. Rept., vol. 23, pp. 178–192, 1912.

[56] Powell, S. T., The effect of algae on bicarbonates in shallow reservoirs: Am. Waterworks Assoc. Jour., vol. 2, pp. 703–708, 1915.

[57] Nauman, Einar, Die Bodenablagerungen des Süsswassers: Archiv für Hydrobiologie und Planktonkunde, vol. 13, pp. 115–117, 1922–23.

[58] Wesenberg-Lund, C., Lake lime, pea ore, and lake gyttja in Danish inland lakes; Dansk geol. Foren., No. 7, pp. 164–167, 1901.

[59] Thienemann, August, Seetypen: Naturwiss. Wochenschr., vol. 18, p. 345, 1921.

cipitation of carbonates occurred, as it does now in many lakes, rather early in the summer, while the temperature of the water was rising most rapidly toward the general high summer level. If another analogy with modern lakes is applicable, probably the peak in the plankton production came near the middle or even late in the summer. These conditions would have been about optimum for the deposition of varved sediments like those of the Green River formation. The carbonates, by reason of their greater density, would have fallen rapidly, whereas the organic matter would lag considerably and so the greater part of it would come to lie above the carbonate layer.

If these ancient lakes were deep enough to permit a thermal stratification of the water during the summer, and this seems probable for certain of their stages, then the separation of mineral and organic matter at the lake bottom would have been nearly complete, for the density of the cold lower layer should have been great enough to retard the rate of fall of small particles of organic matter. In certain modern lakes where the temperatures of the lower layers of water are lower and the densities therefore greater than they probably ever were in Gosiute and Uinta Lakes, this retarding effect is remarkable. Forel [60] says that in the Lake of Geneva the carapaces of various Entomostraca remain suspended for long periods as an opaque cloud at the thermocline—that is, the boundary between the hypolimnion, or lower stratum, in which the water is virtually stagnant, and the epilimnion, or surface stratum, in which the water is free to circulate. Brönsted and Wesenberg-Lund [61] found that during the summer stagnation of certain Danish lakes there was very little organic detritus above the thermocline, but below the thermocline there were vast quantities, which increased in amount downward. A free translation of their remarks upon the organic detritus in the hypolimnion follows:

Often the samples [of water] were colored light brown, and the detritus exceeded by far the living plankton. Very probably it becomes more or less dissolved in the deeper layers. It was frequently observed that the deepest water samples were colored light brown, and often the quantity of detritus was so great that it seemed as if the water sampler had been dragged along the bottom.

This much dissolved organic matter would probably make the water acid and thus tend to dissolve the carbonates already precipitated, but, as Kindle [62] points out in discussing the relations between the thermal stratification of McKay Lake, Ottawa, and the composition of the varves, the later increment of plankton débris would very likely protect the carbonates from this slightly acid water.

Plainly, however, the separation between the mineral and organic constituents in the varves of the Green River formation is not perfect; clay minerals, other clastic minerals, and carbonate grains are scattered through the organic layers, and organic matter occurs in the mineral-rich layers. It seems more reasonable, therefore, to postulate a more or less continuous sedimentation of mineral and organic constituents with first a peak in the production of the carbonates and then a peak in the production of plankton. Evidently it is possible that these peaks might be nearly or indeed quite coincident and still give rise to varved deposits, because of the different rate of settling of the two classes of material. That there may have been more than a single peak in the plankton production seems probable, to judge from the successive maxima of plankton in many modern lakes. Moreover, there is even a suggestion of this in the excessively thin laminae of differently colored organic matter that were observed within the organic parts of the varves of a few rich oil-shale beds. (See pl. 14, *A*.)

The hypothesis of seasonal change in the rate or kind of deposition advanced to explain the varves in the organic marlstone and oil shale appears to explain equally well the thicker varves in the beds of fine-grained limy sandstone. In fact, these varves even appear to provide independent evidence that the major part of the rainfall occurred at one season and at a period preceding the peak in the production of organic matter. The sharp contact between successive varves and the concentration of the sand and silt in the basal part of each varve seem to indicate that after a period of quiet, during which the organic matter accumulated and the finest mineral particles settled, there was a sudden influx of detrital material. This supply of detritus apparently started with maximum intensity and thereafter diminished until the organic matter began to accumulate, when it was barely perceptible. Accordingly, the structure of these varves suggests a period of considerable rainfall, during which erosion was relatively rapid and the streams were loaded, followed by a period of slackened rainfall and general absence of storminess, during which the production of plankton reached its maximum and the finest suspended material settled. This particular varve structure, however, might in itself be quite as well explained by assuming repeated storms or repeated influx of detrital material within a year if the periods of increased supply were separated by intervals long enough for the finest material to settle; but the transition within the same series from varves of this type to varves typical of the marlstone beds, in which the supply of organic matter seems probably to have been seasonal, argues against this

[60] Forel, F. A., Le Léman, vol. 1, p. 118, 1892.

[61] Brönsted, J. N., and Wesenberg-Lund, C., Chemisch-physikalische Untersuchungen der dänischen Gewässer: Internat. Revue der gesamten Hydrobiologie u. Hydrographie, Band 4, Hefte 5 and 6, p. 469, 1912.

[62] Kindle, E. M., The rôle of thermal stratification in lacustrine sedimentation: Roy. Soc. Canada Trans., vol. 21, sec. 4, p. 17, 1927.

explanation. Moreover, this transition appears to indicate that the organic part of the sandy varves is homologous with the organic part of the marlstone varves, and it therefore implies that the organic part of the sandy varves likewise represents a definite season.

CONDITIONS AFFECTING THE PRESERVATION OF THE VARVES

The preservation of varves, particularly varves like those in the beds of marlstone, whose bounding planes are almost geometrically perfect, implies certain requirements as to depth and circulation of the water and as to the population of the ooze at the time of deposition. It is evident that there could have been no wave action and only the most feeble currents in the bottom region of the lake. Currents in the surface layers of the water, on the contrary, were apparently necessary to supply the small but consistent quantity of clastic material (clay minerals, quartz, feldspar, etc.) found in the varved rocks. But this condition, in which the surface water circulated more or less but the deeper water did not, strongly suggests that the water of these ancient lakes was thermally stratified. Other evidence also favors this hypothesis. The original ooze of the varved rocks could not have harbored an active bottom fauna of worms, larvae, crustaceans, and the like, for surely these organisms would have destroyed the perfection of the lamination. Nipkow [63] found that only the ooze below the 90-meter (295-foot) contour in the Lake of Zurich was laminated. Above that level the water is oxygenated, and an active fauna keeps the bottom stirred up and thus prevents stratification.

Lenz [64] sampled the deposits of 12 different lakes in Holstein, each of which had a large production of plankton and consequently also thick deposits of putrescent organic ooze. Yet none of the deposits were laminated. He attributed this largely to the activity of worms, insect larvae, and similar organisms, which burrowed in the ooze, fed upon it, and thoroughly worked it over. Futhermore, the consistent presence of finely granular pyrite in the varves of the Green River deposits strengthens the notion that these deposits accumulated under reducing conditions such as prevail generally in the stagnant hypolimnion of modern lakes that are thermally stratified. One is tempted to go further and say not only that the water of these ancient lakes was thermally stratified but that the stratification was normal—that is, the epilimnion was warmer than the hypolimnion—and moreover that this stratification persisted and was not interrupted by spring and autumn overturns and the establishment of inverse stratification during the winter. In an inverse stratification the temperature of the epilimnion ranges between 0° and 4° C., whereas the temperature of the hypolimnion is 4° C., or very close to it. Spring and autumn overturns, during which all the water in a lake goes into circulation, seem to be precluded by the consistent perfection of the varves. Circulation reaching the lake bottom, it seems, would surely have distributed ooze whose constituent particles were so small, for, according to Lenz,[65] Thienemann found in the surface water particles of organic mud which were brought up from the bottom by the vigorous circulation at times of overturn.

The position of the thermocline in modern lakes varies so much and depends upon so many factors that its depth in these ancient lakes during the periods in which the varved deposits accumulated is difficult to estimate. The mean velocity of the wind is clearly an important factor determining the depth to the thermocline, for it governs the strength of the water circulation; and as these lakes were broad they must have felt the full effect of the winds. But according to Brooks,[66] the early part of the Tertiary period was not a time of great storminess, though apparently there were brief and somewhat violent squalls. Therefore, despite the great area of these lakes, it is perhaps unnecessary to postulate a deep thermocline so as to allow for a vigorous circulation in the epilimnion. Without continued circulation to distribute the sun's heat that reached the lake surface the epilimnion may conceivably have been very shallow and may have been measured in feet rather than tens of feet, as in many modern lakes in temperate zones. Thus, it seems, these ancient lakes, even though large and only 75 or 100 feet deep, may nevertheless have had a distinct thermal stratification. Clearly this is nothing more than a guess at what seems to be a reasonable minimum depth at which perfect lamination in ooze might have originated and have been preserved in bodies of water as large as the ancient Gosiute and Uinta Lakes.

CYCLES OF GREATER MAGNITUDE

Three cycles of greater average length than the varve cycle are suggested by fairly regular recurrent variations in the thickness of the varves and in the thickness and character of certain beds and by the fairly regular spacing of certain layers containing salt molds. The shortest of these cycles appears to correspond to the cycle of sun-spot numbers, and the longest suggests the cycle of the precession of the equinoxes, but the cycle of intermediate length corresponds with no well-established rhythm.

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[63] Nipkow, Fritz, Vorläufige Mitteilungen über Untersuchungen des Schlammabsatzes im Zürichsee: Zeitschr. Hydrologie, 1920 (reviewed by Hummel, K., Neues Jahrb., 1921, Band 2, p. 180).

[64] Lenz, Fr., Schlammschichtung in Binnenseen: Naturwiss. Wochenschr., vol. 18, pp. 327–329, 1921.

[65] Lenz, Fr., op. cit., p. 329.

[66] Brooks, C. E. P., Climate through the ages, pp. 57, 187, 224, New York, R. V. Coleman, 1926.

CYCLE OF SUN-SPOT NUMBERS

Recurrent groups of unusually thick varves are evident in the varve series that are 30 years or more long. (See fig. 15.) The average interval between these peaks is a little less than 12 years, but the length of individual intervals ranges from about 7 to about 18 years. The average interval between sun-spot maxima is a little more than 11 years, though the period ranges from about 7 to more than 16 years.[67] Plainly the correspondence between these two cycles is not perfect. Nevertheless, it is close enough to be suggestive and seems to merit consideration, especially as there are fairly good theoretical reasons for linking variations in the sun's energy with the physical conditions of the lakes that would affect the character of the varves. Brooks [68] has shown that there is a much closer correlation between sun-spot numbers and the levels of certain large lakes in central Africa than between the rainfall and the lake levels. This interrelation between variations in the sun's energy, of which the number of sun spots is an index, and the lake levels depends upon evaporation, which accounts for about 94 per cent of the water lost from the basin in a lake; and the shallower the lake and the broader its littoral zones the more pronounced this effect would be. It has already been pointed out in considering the origin of the varves that the rate of accumulation of both carbonates and organic matter varied directly with and depended chiefly upon the rise and fall of the water temperature. Hence, if at times of sun-spot minima these ancient lakes were abnormally low, they should also have been abnormally warm, and the rates of accumulation of carbonates and organic matter should also have increased accordingly, with the result that the varves for those years should be abnormally thick. In the same way varves corresponding to years of sun-spot maxima should be thinner than usual or perhaps should contain more clastic material because of the increased raininess. This explanation implies that if a real connection existed between solar activity and sedimentation during the Green River epoch, then the peaks on the curves shown in Figure 15 would correspond to intervals of sun-spot minima. Varves consisting predominantly of physicochemical deposits and of plankton débris, whose deposition depends largely upon the physical conditions within

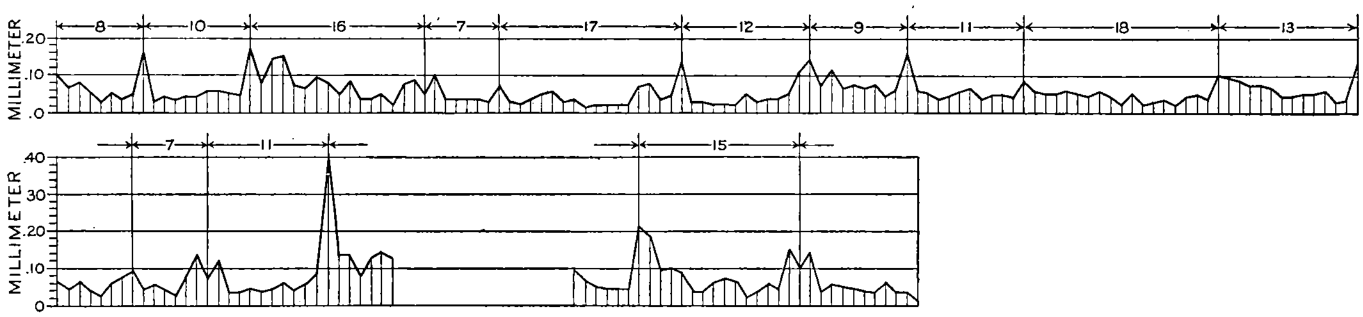


FIGURE 15.—Curves showing the recurrent peaks in the total thickness of the varves in an organic marlstone from Clear Creek, Colo. The numbers indicate the number of varves that separate the recurrent peaks.

of Lake Victoria, the other 6 per cent being the discharge from the lake. Brooks [69] says:

> The chief factor in the amount of evaporation, however, is not rainfall but solar conditions. The researches of W. Köppen and others have established beyond doubt that there is a close connection between sun-spot numbers and tropical temperatures, the latter being 1.1° F. higher at spot minima than at spot maxima. It is reasonable to conclude that the higher the temperature the greater the evaporation; hence at spot minima evaporation will be increased and the level of the lake will fall, while at spot maxima evaporation will be decreased and the level of the lake will rise. This relationship, as we have seen, is so intimate that it gives correlation coefficients of between 0.8 and 0.9.

Perfect correspondence would be represented by a coefficient of 1.0.

A change of level might be expected to have a distinct effect upon the mean temperature of the water in the lake itself, are clearly better suited to reflect the pulse of the solar cycle than varves consisting predominantly of clastic material, whose ultimate sedimentation can be connected with changes in solar radiation chiefly through the very indirect circuit of rainfall, erosion, transportation, and finally distribution within the lake. Thus, from purely theoretical considerations, the varves in the organic marlstone and oil shale of the Green River formation might be expected to show with more or less fidelity the effect of cyclic changes in solar energy. According to Brooks,[70] the opposition between sun spots and temperature is greatest in the Tropics, but diminishes and grows less regular in temperate latitudes. On the other hand, L. Mecking [71] found that the influence of the sun's activity upon the temperature at the earth's surface depends upon local geographic position and upon the season—for example, in the interior of North America the effect of the sun-

[67] Moulton, F. R., An introduction to astronomy, p. 383, 1916.

[68] Brooks, C. E. P., Variations in the levels of the central African Lakes Victoria and Albert: Great Britain Meteorological Office Geophys. Mem. 20, pp. 337-344, 1923.

[69] Idem, p. 343.

[70] Brooks, C. E. P., Climate through the ages, p. 102, 1926.

[71] Hann, Julius, and Süring, R. J., Lehrbuch der Meteorologie, p. 656, Leipzig, 1926.

spot numbers on the temperature is uncommonly strong.

Indications of the sun-spot cycle have been found in other varved sediments. Perfiliev [72] found in the varves of the black mud in Sakski Lake of northern Crimea evidence of cyclic changes whose average period was between 10 and 11 years. These he believes are to be correlated with the cycle of sun-spot numbers. Brooks [73] says that according to T. W. E. David the upper Carboniferous glacial varves of Australia appear to show an 11-year periodicity.

CYCLE OF THE PRECESSION OF THE EQUINOXES

Cyclic changes in the conditions of sedimentation of approximately the same average period as the resultant of the change of eccentricity of the earth's orbit and the precession of the equinoxes are indicated by a regular alternation of beds of oil shale and organic marlstone. These rhythmically alternating beds occur in four groups at three localities in Garfield County, Colo. One group is in a section measured near the head of East Middle Fork of Parachute Creek, in sec. 16, T. 5 S., R. 95 W.; the second group is in a section measured on the north side of Cathedral Creek, approximately in sec. 26, T. 3 S., R. 99 W.; and the third is about 50 feet lower in the Cathedral Creek section. These sections of the Green River formation are complete and were measured by the writer and his assistant, R. D. Ohrenschall, in 1925. The fourth group of rhythmically spaced beds is a part of a section measured by the assistants of Fred Carrol on Clear Creek in sec. 9, T. 5 S., R. 98 W., Garfield County, Colo., and submitted to the Department of the Interior at a hearing on oil shale held before the Secretary on December 1, 1926. The upper group of beds in the Cathedral Creek section, the group in the Parachute Creek section, and the group in the Clear Creek section may be precisely correlated, as they are situated respectively about 150, 160, and 130 feet above a certain thin bed that is locally known as the "Mahogany marker." This marker bed, heretofore generally regarded as a fine-grained sandstone, has been found to be a bed of zeolitized volcanic ash.[74]

The thickness of these regularly spaced oil-shale and marlstone beds multiplied by the average rate of accumulation of each kind of rock serves as an approximate measure of the cyclic changes in sedimentation. Most of the organic marlstone beds are about 6 feet thick, though the extreme range in thickness is from 3.8 to 8.8 feet. The oil-shale beds range in thickness from 0.6 to 3.0 feet. In a general way the thickness of the beds, particularly the oil shale, diminishes with increasing content of organic matter. However, the significance of these groups of beds lies in the regularity of the alternation and in the uniformity of character and relative thickness of the two distinctly different types of rock rather than in the absolute thickness of the beds. The estimated rates of accumulation for organic marlstone, moderately good oil shale, and very rich oil shale are given on page 99. Although these rates are plainly subject to rather large errors, as the thickness of the varves is probably not constant within the limits arbitrarily set for each kind of rock, they are probably the right order of magnitude. The time intervals represented by the sections of oil shale measured are given in the table below.

Groups of beds representing intervals of time suggestive of the precession cycle

Kind of rock	Oil yields per ton	Thickness of beds	Mean rate of accumulation per foot[b]	Interval indicated by each bed	Interval indicated for each cycle
	Gallons	*Feet*	*Years*	*Years*	*Years*
Parachute Creek, Colo.[a]					
Oil shale	28± 2	3. 0	4, 700	14, 000	26, 500
Marlstone	9± 2	6. 2	2, 000	12, 400	
Oil shale	28± 2	2. 3	4, 700	10, 800	25, 200
Marlstone	10± 2	7. 2	2, 000	14, 400	
Oil shale	25± 2	2. 3	4, 700	10, 800	22, 800
Marlstone	14± 2	6. 0	2, 000	12, 000	
Oil shale	33± 2	2. 0	4, 700	9, 400	27, 000
Marlstone	14± 2	8. 8	2, 000	17, 600	
Clear Creek, Colo.					
Marlstone	9. 8	6. 8	2, 000	13, 600	18, 500
Oil shale	35. 0	. 6	8, 200	4, 900	
Marlstone	11. 8	6. 9	2, 000	13, 800	23, 200
Oil shale	22. 9	2. 0	4, 700	9, 400	
Marlstone	11. 2	3. 8	2, 000	7, 600	16, 530
Oil shale	28. 1	1. 9	4, 700	8, 930	
Marlstone	10. 5	6. 1	2, 000	12, 200	25, 350
Oil shale	15. 1	2. 8	4, 700	13, 350	
Marlstone	11. 6	4. 8	2, 000	9, 600	22, 700
Oil shale	35. 2	1. 6	8, 200	13, 100	
Marlstone	10. 6	5. 4	2, 000	10, 800	21, 000
Oil shale	25. 5	2. 2	4, 700	10, 300	
Cathedral Creek, Colo. (upper group)[c]					
Oil shale	25± 5	1. 6	4, 700	7, 500	19, 500
Marlstone	12± 3	6. 0	2, 000	12, 000	
Oil shale	25± 5	1. 4	4, 700	6, 600	18, 600
Marlstone	12± 3	6. 0	2, 000	12, 000	
Oil shale	25± 5	1. 2	4, 700	5, 600	21, 600
Marlstone	12± 3	8. 0	2, 000	16, 000	
Cathedral Creek, Colo. (lower group)[c]					
Oil shale	25± 5	1. 7	4, 700	8, 000	21, 000
Marlstone	12± 3	6. 5	2, 000	13, 000	
Oil shale	35± 5	. 8	8, 200	6, 500	20, 500
Marlstone	12± 3	7. 0	2, 000	14, 000	
Oil shale	35± 5	. 5	8, 200	4, 100	16, 100
Marlstone	12± 3	6. 0	2, 000	12, 000	

Average length of cycle, 21,630 years.

[a] The oil yields for these beds were supplied by the Columbia Oil Shale & Refining Co. and are shown as uncertain merely because the samples distilled were collected about a quarter of a mile west of the place where the writer measured his section.

[b] For marlstone and related rocks yielding less than 15 gallons of oil to the ton the mean rate of accumulation is estimated at 2,000 years to the foot; for moderately good oil shale yielding 15 to 35 gallons, 4,700 years; and for rich oil shale yielding more than 35 gallons, 8,200 years.

[c] No distillation tests were made for these beds, and the yields indicated here were estimated by the writer.

[72] Perfiliev, B. V., Ten years of Soviet science, p. 403, Moscow, 1927.

[73] Brooks, C. E. P., Climate through the ages, p. 192, 1926.

[74] Bradley, W. H., The occurrence and origin of analcite and meerschaum beds in the Green River formation of Utah, Colorado, and Wyoming: U. S. Geol. Survey Prof. Paper 158, pp. 2-3, 1929.

Each sedimentary cycle is represented in these partial sections by one bed of oil shale and one bed of organic marlstone, and the average interval indicated is 21,630 years. The length of the precession cycle is between 25,000 and 26,000 years. However, there is a cyclic change in the position of the major axis of the eccentricity of the earth's orbit (revolution of the line of apsides) whose direction of change is opposed to the precession of the equinoxes. Gilbert[75] stated that "the resultant of the two motions has an average period of about 21,000 years. It is not absolutely regular but ranges ordinarily within 10 per cent of its mean value and exceptionally to 50 per cent above and below." This average figure is suggested by the average interval of 21,630 years for the sedimentary cycles represented by oil shale and organic marlstone as indicated in the table above. The alternation of these two kinds of rocks is what might reasonably be expected to result from the cyclic changes of climate produced by the precession of the equinoxes. If a distinct but yet not abnormally large eccentricity of the earth's orbit is assumed, the precession of the equinoxes should have produced during one half of its cycle a climate in which the summers were short and hot and the winters long and relatively cool and in the other half of the cycle a climate in which the winters were short and mild and the summers long and relatively warm. The mean annual temperature would have remained about the same, however, and the principal differences of climate in the two parts of the cycle would have been in the seasonal distribution of temperature. The greater the eccentricity the more marked would have been these seasonal effects. Spitaler has calculated, according to Brooks,[76] that at times of maximum eccentricity the seasonal extremes of temperature over continental interiors during one-half of the precession cycle would be considerably greater than the extremes of temperature at present, when the earth's orbit is more nearly circular, and the seasonal extremes during the other half of the cycle would be correspondingly diminished.

That part of the cycle in which the summers were short and hot and the winters long and relatively cool possibly would have favored the formation of relatively thin beds of rich oil shale. The precipitation of carbonates due to a relatively high temperature of the lake water would have been restricted to one season, and that a short one; accordingly the total thickness of beds formed under these conditions should be less than that of beds formed under conditions that favored precipitation of carbonates throughout a greater part of the year. Yet it is likely that the short hot summers might have been unusually favorable to large production of plankton, which would have given rise to an ooze abnormally rich in organic matter.

[75] Gilbert, G. K., Sedimentary measurement of Cretaceous time: Jour. Geology, vol. 3, p. 124, 1895.

[76] Brooks, C. E. P., Climate through the ages, pp. 117-119, 1926.

On the other hand, that part of the precession cycle in which the summers were long and warm and the winters short and mild apparently would have provided conditions favoring relatively rapid accumulation of organic marlstone. Under these conditions the temperature of the lake water should have been continually rather high, and this would have favored the precipitation of the carbonates furnished by the streams, whose water was probably somewhat cooler. Thus it seems likely that with a comparable or even smaller production of plankton the rate of accumulation of the ooze during this part of the cycle would have been more rapid than during that part with greater seasonal extremes. However, it should be borne in mind that the difference in thickness between the beds of oil shale and those of organic marlstone is doubtless accentuated and perhaps entirely explained by the greater compactibility of the organic material.

The abrupt change from one type of rock to the other in these partial sections of the Green River formation seems to be at variance with the effect that might be expected from the precession of the equinoxes. As the climate probably changes gradually with the precession, the climatic features characteristic of one extreme of the cycle must be transitional into those characteristic of the opposite extreme. And this suggests that such changes in sedimentary rocks as are controlled chiefly by the precession of the equinoxes should be transitional into one another. But the beds of oil shale and organic marlstone are sharply separated from each other. Also, in the Cretaceous rocks of Colorado, where a regular alternation of beds was interpreted by Gilbert [77] as probably an effect of the precession of the equinoxes, the changes from limestone to shale and the reverse are similarly abrupt. To correlate consistent abrupt changes of lithology like those just cited with the precession cycle it appears to be necessary to postulate that conditions favorable to the formation of one type of sediment, once established, are fairly stable until a certain critical point is reached, when they change rather abruptly to another set of conditions which are equally stable but which favor the formation of a rather different type of sediment. Too little is known of the conditions of sedimentation, however, to venture a guess as to the factors that might govern their stability or be critical for such minor changes in them.

CYCLE OF SALT DEPOSITION

A third cycle in the beds of the Green River formation is represented by regularly recurrent layers of calcite-filled glauberite (?) cavities. (See pl. 14, *B*.) These occur in a sample of moderately good oil shale collected by the writer in 1924 at Green River, Wyo., the type locality of the formation. The average interval between the salt layers is 3.4 millimeters, and

[77] Gilbert, G. K., op. cit., p. 122.

the varves in these intervals average 0.065 millimeter in thickness. The period of the cycle is therefore about 50 years. This corresponds to no well-established rhythm, though it suggests remotely the supposed climatic cycle whose average period is about 66 years. It is also interesting, though possibly without significance, to note that according to a graph compiled by Woolley[78] the two extreme low levels of Great Salt Lake, Utah, occurred about 45 years apart and were separated by a stage at which the level was little more than 15 feet above the lowest level recorded. Similar high stages at approximately the same interval are suggested by the graph for the periods before the first extreme low level and after the second. V. Kremser[79] found progressive temperature changes in northern Germany that suggest a cycle of about 50 years. Perfiliev,[80] however, found in the varved sediments of Sakski Lake, in northern Crimea, a clearly defined cycle of about 70 years.

DURATION OF THE EOCENE EPOCH

GREEN RIVER EPOCH

Before discussing the actual estimates of the time interval indicated by the Green River formation it seems worth while to consider how likely it may be that certain sections of this series of lake beds represent essentially continuous deposition. In any lake the position of wave base must, of course, control the rate of sedimentation within the compass of a certain marginal zone, whose width will depend upon the form ratio of the lake. Deposition and erosion must alternate in this zone, but the position of wave base can never control sedimentation over the entire area of a lake bottom. Even though waves reach bottom everywhere and stir up the mud it is inconceivable that more than a small proportion of the material so thrown into suspension could be carried over the lip of an outlet. On the contrary, it must settle again, so that in certain parts, presumably the deepest parts, of even very shallow lakes the sedimentary record must be complete. In deposits formed below wave base there can be no real breaks in sedimentation unless the lake dries up completely.

As the Green River formation consists predominantly of the five types of rock in which varves have been measured, the length of the Green River epoch can be approximated by applying the average rate of accumulation to the total amount of each kind of rock represented in a complete section of the formation. The section measured by the writer along Parachute Creek, in Garfield County, Colo., seems to be the most suitable for this purpose, as it probably furnishes a record of sedimentation in the deepest part of the basin, where deposition was nearly if not quite continuous. This section is about 2,600 feet thick and consists of about 7 per cent of fine-grained sandstone, about 76 per cent of marlstone and closely related rocks, about 13 per cent of oil shale that will probably yield between 15 and 35 gallons of oil to the ton, about 4 per cent of oil shale that will yield more than 35 gallons to the ton, and a negligible quantity of algal limestone and oolite. The rates of accumulation already assumed (see p. 99) are as follows: Fine-grained sandstone, 1 foot in 250 years; marlstone and related rocks, 1 foot in 2,000 years; oil shale yielding between 15 and 35 gallons to the ton, 1 foot in 4,700 years; and oil shale yielding more than 35 gallons to the ton, 1 foot in 8,200 years. According to these rates the Green River epoch is estimated to have lasted about 6,500,000 years. Probably the greatest errors in this estimate lie in assigning the beds to one or another of the types of varved rocks, whose respective rates of accumulation differ rather widely. The type designated marlstone, for example, though it consists dominantly of rocks closely similar to those containing varves, contains also beds of mudstone and ostracode limestone whose rates of accumulation are unknown but are here assumed to be equal to that of the varved marlstone. With allowance for these and other errors it is perhaps safe to say that the duration of the Green River epoch was between 5,000,000 and 8,000,000 years. This estimate is based, of course. on the assumption that the paired laminae are actually varves. Further study of the lamination and rhythmic alternations of beds in the vicinity of Parachute Creek would probably yield a much more accurate estimate.

The same method of estimating the total time required to deposit the Green River formation was applied to two other sections, and these seemed to indicate a much shorter period. One of these sections is a composite section whose basal part was measured on the divide between Douglas and East Salt Creeks, in western Garfield County, Colo., and whose upper part was measured on Cathedral Creek, in sec. 26, T. 3 S., R. 99 W. This section is about 1,800 feet thick and consists of about 15 per cent of sandstone, about 74 per cent of marlstone and related rocks, about 7 per cent of oil shale yielding between 15 and 35 gallons of oil to the ton, about 1 per cent of oil shale yielding more than 35 gallons to the ton, and about 3 per cent of algae reefs. The total time indicated is a little less than 4,000,000 years. The other section is also a composite section whose basal part was measured in Hells Hole Canyon, about in sec. 22, T. 10 S., R. 25 E., Uintah County, Utah, and whose upper part was measured in the canyon of White River, in sec. 27, T. 9 S., R. 25 E., also in Uintah County. This section is about 1,475 feet thick and consists of about 17 per cent of sandstone, about 76 per cent of marlstone, 2.3

[78] Woolley, R. R., Water powers of the Great Salt Lake Basin: U. S. Geol. Survey Water-Supply Paper 517, pl. 2, 1924.

[79] Hann, Julius, and Süring, R. J., Lehrbuch der Meteorologie, p. 656, Leipzig, 1926.

[80] Perfiliev, B. V., Ten years of Soviet science, pp. 402–403, Moscow, 1927.

per cent of oil shale yielding between 15 and 35 gallons to the ton, 1.7 per cent of oil shale yielding more than 35 gallons to the ton, and about .2 per cent of algae reefs. The total time indicated by this section is a little less than 3,000,000 years.

The relatively shorter time indicated by these thinner sections might be accounted for by assuming that in these localities either the varves are thinner or sedimentation was discontinuous. As the rocks are nearly identical with those occurring at several other localities where the varves have been studied there seems to be little reason to believe that the varves differ appreciably in thickness. But the sandstone in these sections is somewhat coarser, and this feature in itself suggests that scour and deposition may have alternated. Thus it seems a little more likely that these sections are thinner because of interruptions in sedimentation rather than because deposition was slower. The average rate of accumulation for the rocks of the Green River formation as indicated by these three sections is about 1 foot in 2,200 years.

WASATCH, BRIDGER, AND UINTA EPOCHS

The length of the whole Eocene epoch can be approximated by a method similar to that used for the Green River epoch if it is assumed that there are no significant breaks in sedimentation between the successive formations and if a rate of accumulation for the fluviatile deposits that make up the Wasatch, Bridger, and Uinta formations can be estimated. The assumption of essentially continuous deposition in these basins during Eocene time is well founded, though the evidence involves a general discussion of the stratigraphy of the Eocene formations in the Uinta and Piceance Creek Basins and hardly falls within the scope of this paper. These stratigraphic relations will, however, be treated in another and more comprehensive report.[81] An estimate of the mean rate of accumulation of the fluviatile deposits of the Wasatch, Bridger, and Uinta formations has a much less definite foundation than that for the Green River formation, yet even a crude estimate yields interesting results.

This rate of accumulation is deduced from figures given by Dole and Stabler[82] for the average loads of streams and from the inferred relation between the source of material and area of deposition in the Green River Basin of Wyoming. The estimate so obtained is then modified according to the relation between the rate of sedimentation in the ancient Gosiute Lake as computed from Dole and Stabler's data and the rate as indicated by the varves. The computation of the rate of accumulation for the fluviatile deposits is the same as that used on pages 98 and 99 in determining the rate for the lake beds, except that the area of sedimentation is taken as a little less than 17,000 square miles and the area exposed to erosion as a little more than 17,000 square miles, and, in addition, a rolling load of 15 per cent is assumed. Dole and Stabler's data include no estimate of the rolling load. Fortier and Blaney[83] estimate that about 20 per cent of the total normal load of the Colorado River passing Yuma, Ariz., is bed silt—that is, rolling load. This is probably too high for streams like those in the South Atlantic and eastern Gulf Coast States, and therefore 15 per cent is assumed as a more reasonable estimate. Even this may be too high, yet it seems better to overestimate rather than underestimate this factor, for an error in this direction will result in a more rapid rate of sedimentation and consequently a lower estimate of the length of the Eocene epoch. Further, a rock density of 2.0 is assumed as the average for all the rocks of the Wasatch, Bridger, and Uinta formations. The rock densities of nine specimens of Upper Cretaceous marine shale have recently been determined by P. G. Nutting for W. W. Rubey, both of the United States Geological Survey. These average 1.96. According to data compiled by White[84] the average rock density of 39 friable sandstones of Mesozoic and Tertiary age is 2.2. Probably these determinations are not strictly applicable to the Eocene fluviatile rocks considered here, yet they serve as a fairly reliable guide. As mudstone predominates in these formations the true average rock density is probably nearer that for shale than that for sandstone. For the purpose in hand, therefore, the assumed rock density of 2.0 appears to be reasonable. On the basis of this assumption and the others stated above the mean rate of accumulation of these rocks appears to be 0.057 millimeter a year. This estimate of the rate is probably too low, but it seems legitimate to apply to it a correction which is determined by the discrepancy between the probable annual increment of sediment supplied to Gosiute Lake as computed from Dole and Stabler's data[85] and the annual increment as indicated by the varves. (See p. 98.) The varves in the Green River formation indicate a mean rate of accumulation of 1 foot of rock in about 2,200 years, and this is nearly twice the rate indicated by the calculation on page 98 from Dole and Stabler's figures. Hence, if approximately this same relation is assumed to apply to the rate of accumulation of the Eocene fluviatile deposits, then it appears reasonable to double the calculated rate of 0.057 millimeter a year. Accordingly, 0.1 millimeter a year is taken as an approximation to the rate for the accumulation of the fluviatile rocks in the Uinta and Piceance Creek Basins. This is equivalent to 1 foot in about 3,000 years.

[81] Bradley, W. H., The origin and microfossils of the Green River oil shale in Colorado and Utah: U. S. Geol. Survey Prof. Paper — (in preparation).

[82] Dole, R. B., and Stabler, Herman, Papers on the conservation of water resources—Denudation: U. S. Geol. Survey Water-Supply Paper 234, p. 84, 1909.

[83] Fortier, Samuel, and Blaney, H. F., Silt in the Colorado River and its relation to irrigation: U. S. Dept. Agr. Tech. Bull. 67, p. 53, 1928.

[84] White, David, Gravity observations from the standpoint of the local geology: Geol. Soc. America Bull., vol. 35, pp. 214–215, 1923.

[85] Dole, R. B., and Stabler, Herman, op. cit., pp. 78–93.

The possible significance of the discrepancy between the rate of accumulation calculated from Dole and Stabler's data and the rate indicated by the varves may appropriately be pointed out here. As the rate indicated by the varves is the more rapid, it seems to imply either that the area being eroded was larger than that assumed for these calculations or that the rate of erosion was more rapid than in the streams of the South Atlantic and eastern Gulf Coast States, which were chosen as corresponding to the streams feeding Gosiute Lake. Of these two explanations the second seems the more likely, for a large part of each of these ancient streams was in the mountains flanking the basin and therefore probably had a steeper average gradient than the rivers of the southeastern United States. If the rainfall of the Gosiute Lake Basin during Eocene time was distinctly seasonal, that, too, would have favored more rapid erosion.

At first thought it may perhaps seem somewhat surprising that the rate of sedimentation in Gosiute Lake should have exceeded the rate of accumulation of the Eocene fluviatile deposits, especially as the fluviatile deposits are generally somewhat coarser grained. But this relation appears to be reasonable in view of the fact that the lake provided a very efficient settling basin, which retained practically all the material brought by streams, either in suspension or as rolled load. It also retained even the finest clay particles and apparently much of the dissolved mineral matter. Moreover, the lake bottom—that is, the area receiving sediments—was somewhat smaller for essentially the same drainage basin than the area of deposition during the fluviatile epochs. Then, too, sedimentation in the lake was more nearly uniform and continuous over all parts of the lake bottom, whereas the fluviatile deposition, as usual, was plainly more erratic. Streams deposit material relatively rapidly but locally and, in general, not continuously. Long intervals are likely to occur between successive increments of material at any particular place. Furthermore, slight changes in the conditions of the stream may bring about alternate erosion and deposition. Fluviatile sedimentation also is less efficient than lacustrine sedimentation. In perennial streams much of the finest suspended material and a large proportion of the dissolved mineral matter is carried far from its source. In arid regions with interior drainage the streams must, of course, deposit their entire loads. However, it is unlikely that the fluviatile deposits of the Wasatch, Bridger, and Uinta formations were formed in basins with interior drainage. From these considerations the writer concludes that if varves are found in these rocks of fluviatile origin they may be considerably thicker than those in the lake beds, yet the actual rate of accumulation of the fluviatile formation over a long period of time probably was distinctly less rapid. It is possible that a relatively slow rate of accumulation for fluviatile deposits may be more general than is commonly supposed.

According to the estimates given above, 1 foot in 3,000 years is assumed to represent the mean rate of sedimentation of all the Eocene formations of fluviatile origin of the Uinta and Piceance Creek Basins. Measurements by Gale,[86] Spieker and Reeside,[87] and the writer indicate that the average thickness of the Wasatch formation is about 3,600 feet. At the assumed rate of accumulation this would indicate a time interval of nearly 11,000,000 years. Osborn[88] gives the thickness of what the writer regards as the Bridger formation[89] in the Uinta Basin, Utah, as about 1,300 feet and that of the Uinta formation as about 600 feet. These indicate a time interval of nearly 4,000,000 years for the Bridger and nearly 2,000,000 years for the Uinta. In the subjoined summary of the Eocene epoch a factor of uncertainty of about 50 per cent is applied to estimates for these formations consisting of fluviatile deposits.

SUMMARY OF ESTIMATES

From the foregoing calculations based upon measurements of some of the beds of the different formations the length of the Eocene epoch may be estimated as follows:

Estimate of the length of the Eocene epoch

Formation	Time interval (million years)		
	Minimum	Calculated	Maximum
Wasatch	5. 25	10. 7	16. 25
Green River	5	6. 5	8
Bridger and Uinta	2. 75	5. 7	8. 50
	13	22. 9	32. 75
Barrell's estimate of the Eocene epoch based on radioactive minerals	20	23(mean)	26

It is interesting to compare this estimate with those based upon the lead-uranium ratios of radioactive minerals. Holmes[90] gives an age determination based upon a specimen of pitchblende from Gilpin County, Colo., which is late Cretaceous or early Eocene and whose calculated age is about 60,000,000 years. Barrell[91] based his estimates of the early Tertiary epochs upon the age determination of a radioactive

[86] Gale, H. S., Coal fields of northwestern Colorado and northeastern Utah: U. S. Geol. Survey Bull. 415, p. 82, 1910.

[87] Spieker, E. M., and Reeside, J. B., jr., Cretaceous and Tertiary formations of the Wasatch Plateau, Utah: Geol. Soc. America Bull., vol. 36, p. 440, 1925.

[88] Osborn, H. F., The titanotheres of ancient Wyoming, Dakota, and Nebraska: U. S. Geol. Survey Mon. 55, p. 96, 1929.

[89] Osborn refers to these beds in ascending order as Uinta A, Uinta B 1, and Uinta B 2; what the writer has called here the Uinta formation he calls the true Uinta or Uinta C.

[90] Holmes, Arthur, The age of the earth, p. 73, 1927.

[91] Barrell, Joseph, Rhythms and the measurements of geologic time: Geol. Soc. America Bull., vol.8, pp. 875-876, 884, 1917.

mineral from Ireland, which is of lower Oligocene or upper Eocene age and whose calculated age is about 31,000,000 years. With this as a tie point he estimated the relative lengths of the epochs on the basis of stratigraphic evidence and concluded that the length of the Eocene was between about 20,000,000 and 26,000,000 years. The very close correspondence between the average of the writer's estimates, a little less than 23,000,000 years, and the average of Barrell's estimates, 23,000,000 years, is doubtless merely a coincidence, for the evident errors in the writer's method of estimating are certainly of the order of several million years. Nevertheless, as the writer's estimate is wholly independent of those based on radioactive minerals it seems to provide an interesting though an obviously rough check.

CONTACT METAMORPHISM OF THE ROCKS IN THE PEND OREILLE DISTRICT, NORTHERN IDAHO

By Joseph L. Gillson

ABSTRACT

In the Pend Oreille district, Idaho, a section of sediments, part belonging to the Belt series, of Algonkian age, and part to the Cambrian system, was intruded and intensely metamorphosed by granodiorites of late Mesozoic or early Tertiary age. This metamorphism proceeded in three overlapping stages. During the intrusion of the igneous rocks a general recrystallization of the sediments took place, the siliceous rocks changing to adinoles near the contacts and to plagioclase-bearing rocks at greater distances and the limestones turning to marbles. Later, during the crystallization of the granodiorites, emanations carrying the so-called mineralizers were given off in quantity and formed high-temperature silicates in the sediments. Still later, after the igneous rocks had solidified and the temperature of the masses had become lower, sericite, chlorite, serpentine, magnetite, and sulphides were formed by replacement of the earlier minerals. Although no rare minerals were found in the contact zones, more than 50 minerals common to such zones were identified.

LOCATION AND GENERAL GEOLOGY

The Pend Oreille silver-mining district lies in Bonner County, in the panhandle of Idaho, and is adjacent to the south arm of Pend Oreille Lake, a large body of water lying in a tremendous glacier-cut trough. (See fig. 16.) An area about 15 miles wide and 20 miles long was studied by a party of the United States Geological Survey under Edward Sampson in the summers of 1921, 1922, and 1924, and in the support of this study the Idaho State Bureau of Mines and Geology kindly cooperated.[1]

The district is underlain by sediments of Algonkian age (the Belt series) and Cambrian age intruded by igneous rocks, the largest masses of which, at least, are of late Mesozoic or early Tertiary age. These igneous rocks, which crop out over about one-fifth of the land area, have already been described by the writer.[2]

The surface character of a part of the district and some of the geologic features are shown in Figure 17.

[1] Sampson, Edward, Geology and silver-ore deposits of the Pend Oreille district, Idaho: Idaho Bur. Mines and Geology Pamph. 31, 1928.

[2] Gillson, J. L., Granodiorites in the Pend Oreille district of northern Idaho: Jour. Geology, vol. 35, pp. 1-31, 1927.

THE PROBLEM

The intrusion of the igneous rocks was accompanied by metamorphism so intense that in few places are the sediments free from its effects. The study of the rocks of the district has shown that the process of igneous metamorphism was long continued, and

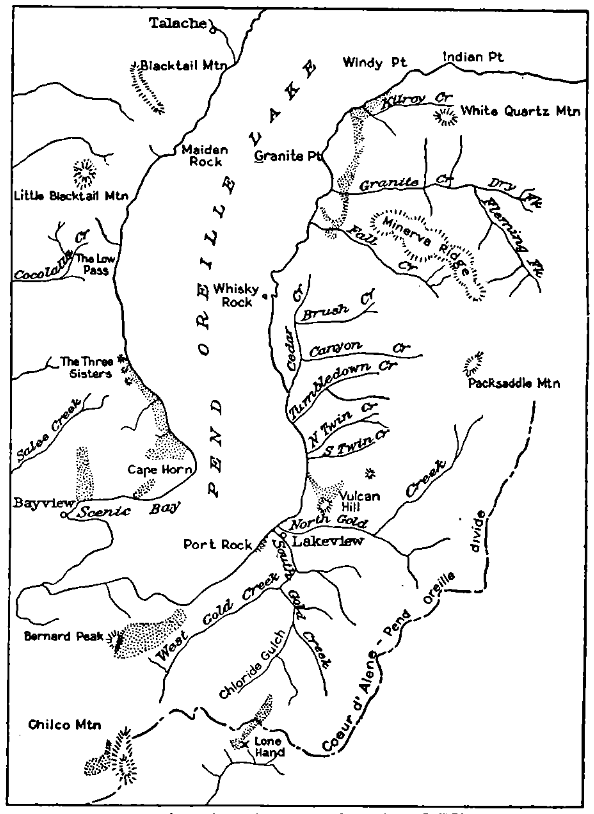


FIGURE 16.—Map of the Pend Oreille area, Bonner County, Idaho. Stippling indicates zones of intense contact metamorphism

that conditions progressively changed during the crystallization of the magma. Three periods or stages in the metamorphism can be recognized, and the chief purposes of this paper are to prove that these stages occurred and to describe them. The calcareous rocks were changed in a different manner from the noncalcareous rocks, and the difference indicates that the primary character of the sediments was an important factor in the metamorphism. Furthermore, considerable material was introduced during all three stages of the metamorphism.

THE ROCKS PRIOR TO METAMORPHISM

The section of Belt rocks in the Pend Oreille district is similar to that in the near-by Coeur d'Alene district, described by Ransome and Calkins.[3] The Pend Oreille section differs principally from that in the Coeur d'Alene district in that the rocks at the horizons of the Revett and St. Regis formations—that is, the rocks underlying the Wallace formation and overlying the Burke formation—are more siliceous, and a new name, Blacktail formation, has been given to them. The Belt rocks are chiefly argillites and quartzites of great thickness, and the strata differ only in color and in the proportions of argillaceous and siliceous material. They are affected by load metamorphism, and where not contact-metamorphosed they consist almost entirely of quartz and white mica. Under the microscope the quartz grains appear detrital and each is surrounded by a film of finely divided mica. In the more argillaceous beds the mica flakes are very abundant and tend toward a general orientation with their flat sides parallel to the bedding. The rocks are almost everywhere devoid of flow cleavage and preserve such sedimentary features as ripple marks, cross-bedding, mud cracks, clay galls, and rarely casts of salt crystals. The Wallace formation, which overlies the Blacktail, differs from the rest of the series in that it contains almost throughout a considerable amount of calcite, which is present in very peculiar structures, considered to be of algal origin. The Striped Peak, the uppermost formation of the series, is also locally calcareous.

Belt sediments.—The beds of the lowest formation, the Prichard, consist of dark-colored argillite and grayish quartzite. Some beds of the quartzite are fairly coarse grained. Many of the beds are closely laminated, owing to frequent alternations of sand and mud in deposition. The argillaceous layers are thin, black, and in many places very closely spaced.

The Burke formation, which overlies the Prichard, consists of thin layers of medium-gray fine-grained light-weathering quartzite interbedded with dark-gray siliceous argillite, the bedding surfaces of which are micaceous, are either green or nearly black, and exhibit conspicuous mud cracks. Some of the quartzite has a distinctly bluish tint on fresh fracture. Some beds are laminated; others are not. Some thick masses of heavy-bedded light-colored quartzite also occur.

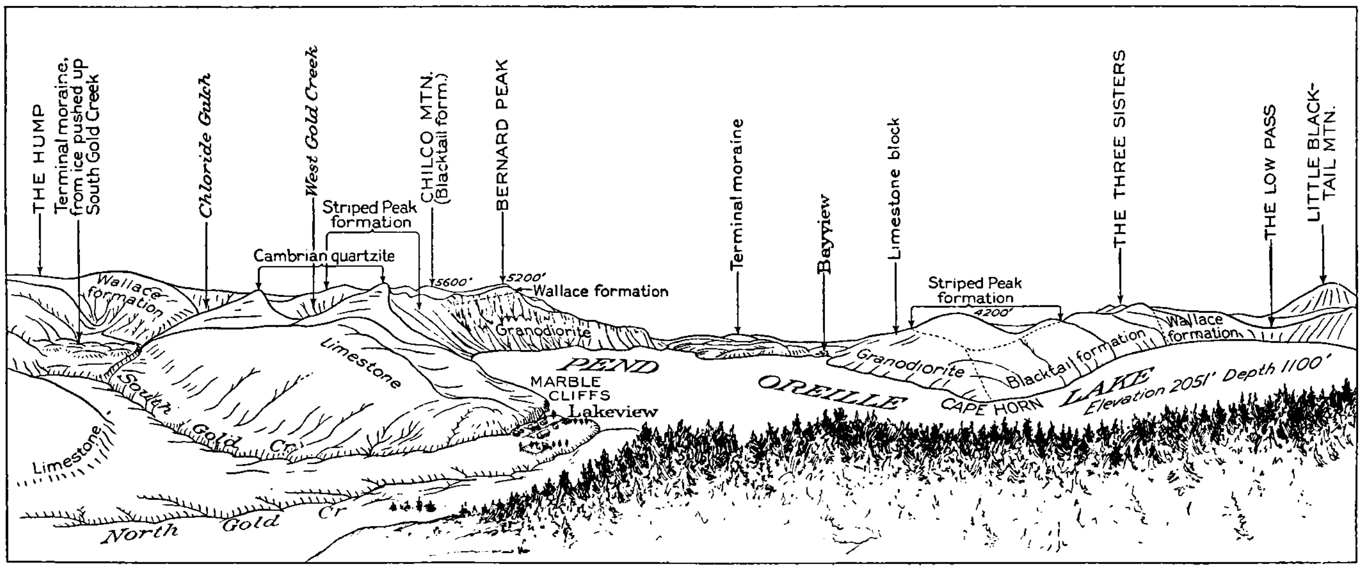


FIGURE 17.—Sketch from photographs taken from hill north of Lakeview, Idaho, looking west and south

The lower part of the Blacktail formation, which overlies the Burke, is dominantly a very massive bluish-gray quartzite that weathers nearly black and breaks into large rectangular blocks. On exposed faces across the bedding discontinuous grayish or purplish bands occur, abundant in some beds, rare or absent in others. The bedding surfaces are thin shaly partings, purple or maroon, lustrous, and conspicuously mud cracked and ripple marked. The upper part of the formation is more argillaceous, and the purple shaly partings are closely spaced. Many layers of the upper part are green but do not differ otherwise from the rest.

The lithology of the succeeding Wallace formation is varied, but two features appear in so many of the beds that its identification was usually simple. The argillaceous partings of thin-bedded quartzite expose a bedding surface of lustrous black, on which are con-

[3] Ransome, F. L., and Calkins, F. C., Geology and ore deposits of the Coeur d'Alene district, Idaho: U. S. Geol. Survey Prof. Paper 62, pp. 29-44, 1908.

A BLOCK OF THE WALLACE FORMATION, 10 BY 18 BY 24 INCHES IN SIZE, FOUND ON THE BEACH OF PEND OREILLE LAKE NEAR TALACHE, IDAHO

Shows the peculiar contorted cavities caused by the more rapid solution of the calcite bodies than of the rest of the rock. These cavities are believed to be of algal origin.

spicuous mud cracks. Also, the peculiar structures of calcite above mentioned are especially diagnostic. The calcite masses are fine grained, and most of them are elongated in two directions at right angles. The vertical cross section is the most striking. It shows a peculiar wavy band from 0.5 to 4 millimeters wide and from 0.5 to 6 centimeters high. Some cross sections resemble a question mark into which a few extra crooks have been put. Owing to the more rapid solution of the calcite these structures are represented on weathered surfaces by cavities, and by them the formation was generally distinguished. (See pl. 15.) These calcite masses modified the metamorphism to a different type from that prevailing in the surrounding noncalcareous materials, and in intensely metamorphosed beds they were seen as pseudomorphs. Argillaceous beds of the Wallace are black on fresh fracture but weather buff. A flinty green argillite is found at the bottom of the formation, and at the top the beds are closely laminated, similarly to those in the Striped Peak formation.

The Striped Peak, the top formation in the Belt series in the Pend Oreille district, is a thin-bedded fissile dark-gray argillite with interbedded white quartzite. Brick-red is the predominant color on exposed surfaces. Mud cracks and ripple marks, few of which are more than a centimeter in width, are conspicuous on the beds. The intercalation of many thin argillite and quartzite layers, 15 or 20 to the inch, is a characteristic feature of the formation. A considerable thickness of the Striped Peak differs, however, from the main type and is an unlaminated olive-drab graywacke. At least at one horizon the upper part of the formation contains irregular masses of calcite similar to those in the Wallace.

Cambrian sediments.—The Cambrian rocks comprise a massive conglomeratic quartzite, a thin-bedded friable shale, locally fossiliferous, and a thick limestone with several facies, fossiliferous at some horizons. The quartzite probably corresponds to the Flathead quartzite of Montana; the shale contains a small fauna of Middle Cambrian age. The fossils of one stratum of the limestone correspond with those of the Langston limestone of the Blacksmith Fork section in Utah, and a shale in the limestone may be correlated with the Spence shale member of the Ute limestone of Utah.[4]

THE METAMORPHISM

GENERAL FEATURES

Each sedimentary formation was found to be profoundly metamorphosed somewhere in the district, and each was nearly everywhere slightly metamorphosed. The zones of intense metamorphism, in which complete recrystallization had taken place with an almost total elimination of sedimentary characters, are relatively small, few being more than 200 yards wide on the surface. Marmarization of the limestones, however, was more widespread, and at some places marbles occur without near-by exposures of igneous rock.

The metamorphism of the Prichard, the Burke, the Blacktail, the noncalcareous part of the Striped Peak, and the Cambrian quartzite is so similar as to require no separate description. The metamorphosed calcareous rocks contain many features in common, although the Wallace and the calcareous part of the Striped Peak, being dominantly argillaceous, differ from the metamorphic product of the purer Cambrian limestone.

The igneous rocks that caused the exomorphism of the sediments are in no place severely endomorphosed. In a narrow zone close to the walls the intrusive granodiorite contains muscovite and has been rather intensely affected by sericitization and chloritization accompanied by the formation of considerable magnetite. Locally molybdenite and other sulphides formed in the endomorphic zone.

METAMORPHISM OF NONCALCAREOUS ROCKS

GENERAL FEATURES

The more intense metamorphism of the noncalcareous rocks has visibly changed them. They are commonly spotted, and crystals of biotite and muscovite a millimeter in diameter can be distinguished. The more argillaceous beds have become more resistant to erosion and make bold outcrops. The microscope shows that the old fabric of texture and structure has been very largely preserved, but the quartz and finely divided white mica have been replaced by oligoclase-albite with or without quartz. The feldspar crystals are of the same size as the former quartz grains and make a mosaic of interlocking anhedral grains. Irregularly distributed in this fabric are crystals of biotite, andalusite, tourmaline, apatite, zircon, and iron-rich chlorite and muscovite, most of which are of anhedral form and in general much larger than the quartz and feldspar. All were formed without regard to the boundaries of the feldspar and quartz and are clearly replacements of those minerals. Later finely divided sericite and chlorite were deposited by replacement of all these other minerals, and last of all magnetite, pyrite, and pyrrhotite were formed in the sericite and chlorite.

There were thus three rather well-defined periods of mineral formation—first a period of recrystallization and substitution of a sodic plagioclase for the mica and some of the quartz, later a period of formation of high-temperature silicates, and finally a period of sericitization with the formation of magnetite and sulphides.

Farther from the contact the changes that went on in the rocks differ only in degree from those just

[4] Walcott, C. D., Smithsonian Misc. Coll., vol. 53, pp. 191, 197, 1908.

described. The sodic plagioclase is present in subordinate amounts, and most of the interstitial mica of the load metamorphism remains. The biotite and iron-rich chlorite are present in small quantities, but magnetite and sulphides are more rare.

UNEQUAL DEVELOPMENT OF THE STAGES OF METAMORPHISM

The moderately metamorphosed rocks were not in general affected equally by the three stages of metamorphism. For example, a bed of white quartzite in the Wallace formation exposed on the shoulder of the spur on the north side of Dry Fork, near the mouth of Fleming Fork, was entirely recrystallized to an allotriomorphic aggregate of quartz and albite-oligoclase, the average grain size of which is about 0.05 millimeter. A few minute tourmaline prisms are seen in thin section, but there is no biotite, chlorite, sericite, or magnetite. Thus the rock was strongly affected by the first stage of metamorphism but hardly at all by the later stages.

Many other examples of this unequal development of the metamorphism could be given. The beds in Maiden Rock belong to the Blacktail formation and were recrystallized during the first stage, with the formation of much feldspar, but only a few rhombs of a late carbonate resulted at a later stage. The Blacktail beds at the summit of Chilco Mountain were not at all affected by the first stage, but biotite and tourmaline are abundant in them as replacement minerals. The Burke quartzite in the tunnel of the Phil Sheridan mine, on the north side of Granite Creek near its mouth, and in contact with the Granite Creek granodiorite at the lake shore, a short distance north of the mouth of the creek, was intensively metamorphosed only during the third stage of metamorphism. It contains pea-sized aggregates of late sericite, with chlorite, magnetite, pyrite, pyrrhotite, and some calcite. Only a little feldspar, biotite, apatite, and zircon were formed in earlier stages. At a point on the west side of the Chilco Mountain ridge the Blacktail quartzite was found to have been intensively metamorphosed during the third stage, so that the earlier metamorphic minerals, once rather abundant, were almost completely eliminated. This lack of relation of the stages of metamorphism affords one of the best proofs that the metamorphism took place in stages.

DISTRIBUTION OF MICROSCOPIC METAMORPHISM

Examples of beds that had not been metamorphosed were sought with care, in order to compare the entirely unmetamorphosed phases of each formation with the metamorphosed phases. During the early part of the field work it was not realized how widespread the metamorphism is, for except near the contacts of the igneous rocks the sediments are not visibly contact metamorphosed. When thin sections of rocks supposed to be unmetamorphosed were examined, it was found that the sodic plagioclase, biotite, chlorite, and minute prisms of tourmaline were of almost universal occurrence. Study showed, however, that these minerals were not of detrital origin, nor were they formed by regional load metamorphism. Where they occur the clastic texture of the rock has been somewhat obliterated, whereas in the rocks of the most typical clastic character these minerals are absent. Detrital feldspar occurs in some beds, but the grains are surrounded by a film of sericitic material. The sodic plagioclase occurs as fresh glassy crystals that interlock with one another or with the quartz. Foils of detrital biotite were found interstitially between clastic quartz grains; the metamorphic biotite has replaced the quartz grains, and a great many are larger than the average grain size of the rock. The tourmaline prisms have likewise replaced adjacent quartz grains. The tourmaline is remarkably widespread, being present at almost every point where specimens were collected. Calkins [5] found this universal distribution in the rocks of the Coeur d'Alene district, and he also considered it to be of metasomatic origin. Tourmaline was not found in rocks showing no other effects of metamorphism but was invariably accompanied by feldspar and biotite.

Proof that these minerals were not formed by recrystallization in the course of the load metamorphism that turned the old sediments into argillites and quartzites is afforded by the fact that such sericitic quartzites or siliceous argillites were shown by analysis to be low in soda, iron, and magnesia. Furthermore, so ubiquitous a distribution of tourmaline, if it was formed by recrystallization of an earlier detrital generation of tourmaline, demands the ultimate source to have been a metamorphic terrane, from which other metamorphic minerals should have been carried in with the tourmaline. No such minerals have been found as detrital or recrystallized grains in the rocks of the Belt series. The tourmaline must therefore indicate that igneous solutions have passed through these rocks.

Additional proof that the tourmaline, biotite, feldspar, etc., were formed by the action of solutions migrating from the intrusive granodiorite is afforded by the increasing abundance of the identical varieties of these minerals found in the same beds toward the igneous contacts.

Another mineral, rather widespread in the Belt rocks and also believed to have been formed by solutions from the igneous source, is a carbonate. A variety of ankerite was found to be the most abundant species, and the rhombic form the most characteristic of the occurrences. The carbonate was found in every

[5] Ransome, F. L., and Calkins, F. C., The geology and ore deposits of the Coeur d'Alene district, Idaho: U. S. Geol. Survey Prof. Paper 62, p. 101, 1908.

Belt formation except the Burke and is most abundant in the Wallace and in the Cambrian limestone. Because of the greater abundance in originally calcareous rocks it is probable that most of the carbon dioxide was derived from the sediment itself. As there was no apparent change in volume, some introduction of CO_2, Fe, and Mg must be supposed to make ankerite from calcite.

EXAMPLES OF INTENSE METAMORPHISM OF NONCALCAREOUS ROCKS

Contact of Prichard argillite with Granite Creek granodiorite.—The igneous contact is sharply exposed on the east shore of the lake north of Granite Point. The well-defined bedding of the sedimentary rock is obliterated for approximately 100 feet, and the hornfels is cut by many quartz veins. For 25 feet from the contact the rock is a dense aggregate of quartz and albite but contains also biotite, zircon, apatite, and ilmenite. The albite decreases in abundance away from the contact.

Chemical analyses of the adinole at the contact and of a sample of the argillite collected at some distance from it definitely prove that an increase in soda and a loss in potash took place during metamorphism. There was also an unmistakable increase in iron, phosphorus, and zirconia and probably in lime, magnesia, and titanium.

Analyses of metamorphosed Prichard argillite

[J. G. Fairchild, Analyst]

	Contact rock	Slightly metamorphosed rock [a]
Silica	66.95	69.06
Alumina	15.42	15.91
Ferric oxide	.63	1.21
Ferrous oxide	4.19	3.17
Magnesia	2.01	1.09
Lime	.62	None.
Soda	6.57	1.62
Potash	Trace.	3.96
Water (total)	1.96	1.98
Titanium oxide	1.00	.75
Carbon dioxide	None.	None.
Phosphorus pentoxide	.30	Trace.
Chlorine	None.	None.
Fluorine	None.	None.
Manganese oxide	.05	Trace.
Zirconium oxide	.07	[b].03
	99.77	98.78

[a] Contains organic carbon, probably graphitic.
[b] Accuracy of the zirconium determination said by Mr. Fairchild to be 0.005.

Contacts of Granite Creek granodiorite with Blacktail quartzite.—A contact of Blacktail quartzite with the Granite Creek granodiorite is well exposed on the east side of the lake near the mouth of Fall Creek. The sediment near the contact is spotted with black knots of biotite 1 to 2 millimeters in diameter or is streaked and mottled with biotite and muscovite aggregates. The igneous rock contains visible crystals of muscovite and is seen under the microscope to have been also sericitized and enriched in magnetite.

The sediment has a groundmass of anhedral albite-oligoclase grains and quartz. These grains are about 0.05 to 0.075 millimeter in diameter, except exactly at the contact, where they are twice this size. Replacing this groundmass are larger quartz grains and abundant irregularly shaped grains of biotite, muscovite, and andalusite, many of which are several times as large as those of the groundmass. Much of the andalusite is partly altered to a finely divided greenish mica. Minute subspherical grains of apatite, rutile, and zircon are widely scattered in the rock. Anhedrons of magnetite and ilmenite, about 0.05 millimeter in diameter and disseminated in the most random manner, were certainly the latest minerals to form.

On the west slope of Chilco Mountain the same formation is in contact with the Bayview batholith, but exposures are poor. The rock was recrystallized to an aggregate of quartz, microcline, and subordinate oligoclase-albite, with little increase in size of grain. Later larger crystals of muscovite, quartz, biotite, and andalusite formed, many of the grains of which are more than a millimeter in diameter. Minute crystals of zircon and blebs of an unidentified mineral having about the refractive indices and birefringence of pyroxene are abundant. Many of the andalusite grains are poikilitic, resembling a graphic intergrowth with quartz. In the later stages of the metamorphism the biotite was altered to chlorite with the separation of rutile needles, making sagenitic structures; the andalusite and feldspar were sericitized. In many of the large crystals of muscovite very peculiar spherulites of sericite formed, the individual needles of which are hairlike, and many are as long as 0.2 millimeter.

Intensely metamorphosed Striped Peak formation at Cape Horn.—Metamorphosed beds of the Striped Peak formation occur 1,700 to 2,000 feet above the lake on the south side of Cape Horn. The characteristic banding of the formation (as seen in section) is preserved, and the rock splits fairly well along the bedding. The dark bands are rusty brown; the light ones yellow-brown. Some of the light bands are marked with dark spots. The microscope shows that the three stages of metamorphism are well indicated. The groundmass consists of interlocking grains of quartz, microcline, and albite-oligoclase, in which larger and very irregular shaped crystals of biotite, muscovite, and tourmaline had developed at a later time. Small crystals of zircon and apatite are accessory constituents of the second stage of metamorphism. Magnetite, sericite, and chlorite formed later, as is indicated by the cross cutting of the grains and the replacement by them of biotite and feldspar. The amount of feldspar that had formed during the early period was relatively less than in other similarly severely metamorphosed rocks. The tourmaline is more abundant and occurs in larger grains than were found elsewhere in the district.

Contact of Cambrian quartzite with Bayview granodiorite.—A contact of the Cambrian quartzite with the granodiorite is exposed in the field east of the summer residence known as Dromore, northeast of Bayview. The sediment at the contact is a gray rock flecked with minute biotite scales and having an average grain size of 0.5 millimeter. In places it is faintly banded. The same sequence of metamorphism is indicated as in other contact zones. There was an early crystallization of albite-oligoclase with the quartz, forming an interlocking groundmass, the grains of which average about 0.1 millimeter in diameter. The feldspar is not uniformly distributed, making solid masses in some parts of the thin sections, and being absent from other parts. Biotite, muscovite, andalusite, and tourmaline formed abundantly; the tourmaline is found in small grains, the others in large ones. Many of the biotite foils contain zircon crystals surrounded by pleochroic halos, and many foils are also altered to thuringite, with the separation of rutile needles. Sericitization had been intense and widespread, attacking most successfully the feldspar and andalusite. Magnetite and ilmenite are also abundant; many of the grains of the ilmenite are partly altered to leucoxene.

METAMORPHISM OF CALCAREOUS ROCKS

GENERAL FEATURES

The mineralogy of the metamorphosed Wallace formation and of the Cambrian limestone is so strikingly different from that of the rocks above described that it indicates the importance of the original character of the rock in determining the nature of the minerals formed by metamorphism. Albite, andalusite, a biotite with very high index of refraction, tourmaline, and thuringite are characteristic of the metamorphism of the noncalcareous sediments, but in those in which calcite was originally present garnet, diopside, a magnesian biotite, amphibole, epidote, and titanite are abundantly developed, and scapolite, vesuvianite, olivine, chondrodite, topaz, and fluorite were found in a few places. On the other hand, zircon, quartz, apatite, sericite, chlorite, magnetite, and sulphides occur in both kinds of rock.

There were two distinct processes in the metamorphism of the calcareous rocks, and like the stages of metamorphism in the noncalcareous sediments the degree of development of one is more or less independent of the other. There was first marmarization of the limestones on a large scale, which eliminated the bedding and the carbonaceous material. Later came the introduction of material from the igneous rock, which formed the contact silicates above mentioned. This second process is conveniently divided into two stages on a rough basis of temperature, with sericite and chlorite marking the beginning of the later stage. There were thus three stages also in the metamorphism of the calcareous rocks.

EXAMPLES OF METAMORPHISM OF CALCAREOUS ROCKS

Metamorphosed Wallace formation.—The rock along the west shore of the lake, from Cape Horn northward under the Three Sisters, is all metamorphosed, but through much of this distance the exposures consist of the Blacktail formation. Farther north are found greenish and white quartzite beds, characteristic of the metamorphosed Wallace, the identity of which is established by the presence of the typical fretwork shown in section, which prior to metamorphism consisted of calcite structures of organic origin. The grains are very small, and the rock has a subconchoidal fracture. The green bands consist of a pale-green amphibole; the white bands are either of quartz and feldspar, quartz and scapolite, or diopside, which in some beds makes nearly solid masses. Zoisite, biotite, grossularite, titanite, zircon, apatite, and muscovite are accessory minerals.

The climb up the east side of Bernard Peak from West Gold Creek, at the south end of the lake, gives another instructive section of metamorphosed Wallace beds, there present as a capping over the Bayview granodiorite, which is exposed almost to the top of the tremendous cliff that rises abruptly from the shore. The same rock is also seen in the gap between Bernard Peak and Chilco Mountain. This rock is massive, is marked with white or white and green bands, and is notably heavy from the abundance of heavy silicates. The grain, however, is so fine that no minerals can be recognized with the hand lens. The mineral content is about the same as that of the rock along the shore under the Three Sisters, consisting of quartz, feldspar (both microcline and albite), diopside, zoisite, garnet, titanite, zircon, apatite, calcite, and magnetite. The pyroxene and amphiboles are distinctly later than the feldspar. Titanite is very abundant in some beds but occurs in minute grains. A pale-brown biotite makes up 30 per cent of some beds but is entirely absent from others.

Another area of intensely metamorphosed rocks is on the slope east of the Hewer mine, near the divide separating Chloride Gulch from the tributaries of the North Fork of Coeur d'Alene River, at the south edge of the district. No igneous rocks are exposed, but minute seams of pegmatite were found containing quartz, potash, feldspar, epidote, and magnetite. Beds at an altitude of 3,750 feet directly above the Hewer mine were studied. The rock is a dense greenish hornfels, consisting of a groundmass of quartz and microcline, in which are numerous finely divided prisms of the green amphibole, accompanied in some beds by a fibrous amphibole of lower refractive index.

Zircon, titanite, apatite, and magnetite are scattered widely through the rock.

In the metamorphic area just described but beyond the divide, in the North Fork drainage basin near the Lone Hand prospect, the difference between the minerals formed during metamorphism of calcareous and noncalcareous beds is convincingly demonstrated. In some laminae within a single thin section are quartz, sodic plagioclase, and andalusite, with sericite and chlorite—the typical mineralization of the siliceous sediment; whereas in adjacent originally calcareous laminae are epidote, diopside, amphibole, fluorite, titanite, and calcite. In some places the calcite forms pseudomorphs after the amphibole. During a mechanical separation of one unbanded specimen in bromoform, three-fourths of the powder sank, and of this fraction 95 per cent was amphibole.

Metamorphosed Cambrian limestone.—The Cambrian limestone is marmarized, or partly so, over large areas. At the lime quarries near Bayview, at the old quarries at the head of Cocolalla Creek and near Whiskey Rock, and in the headwaters of North Gold Creek the rock is marmarized but contains few silicates. The most interesting examples of more intense metamorphism were found in an included block of limestone on the south side of Cape Horn, in cliffs along the east shore between South Gold Creek and Port Rock, and at Vulcan Hill, east of Lakeview. These zones have a varied mineralogy and present the most interest to the mineralogist of all the contact zones in the district.

A number of prospects for ore on Vulcan Hill have made good exposures of the contact of the granodiorite and limestone. The sediment is a white marble in which many silicates and some ore minerals were locally developed so abundantly that the rock is entirely replaced by them. In some beds the rock is banded green and yellow and consists principally of garnet and augite but also contains vesuvianite, epidote, phologopite, amphibole, fluorite, quartz, sericite, chlorite, zeolites, magnetite, sulphides, and ankerite. In some closely banded shaly limestones the effect of alternating composition is again clearly demonstrated. Sodic plagioclase formed in the shaly layers, but at that stage of mineralization no silicate formed in the limestone beds, and the rock only recrystallized to a marble. Later biotite formed in the shale, and garnet, diopside, and amphibole formed in the limestone. The later sericite and chlorite formed in both without regard to the boundaries of the layers. In one bed a considerable quantity of corundum was found, apparently produced from the recrystallization of an impure bed in the limestone.

The garnet and pyroxene differ in composition in adjacent beds of the limestone and indeed in the same bed, as the garnets are zoned in some places. The difference in zoned garnets, however, is slight, but in adjacent beds grossularite and andradite are found, and diopside and augite occur at different horizons.

In the garnet zones on Cape Horn a few other minerals occur in subordinate quantity, topaz, bytownite, and allanite being identified microscopically. Some marble cliffs on the east shore of the lake, between the mouth of South Gold Creek and Port Rock, also yielded additional minerals. At the east end of a little beach 100 yards south of the mouth of the creek the rock is a light-gray and green, partly recrystallized but still bedded limestone. Besides ankerite, scapolite, spinel, muscovite, and chlorite, considerable chondrodite is present in small pale-yellow crystals, a few of which can be seen with a hand lens. Farther southwest, beyond a fault, the rock is more thoroughly marmarized and contains diopside, phlogopite, and tremolite and in the more shaly layers sodic plagioclase and quartz.

The independence of marmarization and silicate formation was well shown at the last series of outcrops. Marble occurs without silicates, and silicates are found in bedded limestone that is but little marmarized. Where silicates occur in marble they indicate by their boundary relations, so far as these can be interpreted, that they were formed later than the marmarization. From these facts it is inferred that the marmarization took place principally during the first effusion of solutions from the magma, mostly of water, which mixed with and heated water of connate or of surface origin, whereas the silicates formed later, after the advance of crystallization in the igneous rocks had caused a concentration of the emanations by adding more material in solution.

ENDOMORPHISM OF THE IGNEOUS ROCKS

The writer[5a] has elsewhere briefly described the endomorphism of the igneous rocks and has connected it, at least in a general way, with the action of the same solutions that slowly soaked through the body of the igneous rock and made a number of postconsolidation changes. At few points are the rocks conspicuously endomorphosed, and except for some changes visible only under the microscope the rocks appear the same at the sedimentary contacts as elsewhere. The addition of large flakes of muscovite, an intense sericitization and chloritization, and the enrichment of magnetite were the endomorphic changes. The muscovite is not pyrogenetic, for it has replaced earlier minerals. If its time of formation in the igneous rocks can be fixed as contemporaneous with that in the adjacent sediments, the time of their metamorphism in relation to the crystallization of the igneous rock is relatively well established. It seems to the writer that the time of crystallization of the muscovite must have been contemporaneous on the two sides of a sharp contact

[5a] Gillson, J. L., Granodiorites in the Pend Oreille district of northern Idaho: Jour. Geology, vol. 35, p. 20, 1927.

between igneous rock and sediment. This assumption furnishes a basis for the dating of the stages of metamorphism in the sediments.

TABULAR VIEW OF THE METAMORPHISM

The metamorphism of the rocks in the Pend Oreille area can be shown in tabular form as follows:

Metamorphism of rocks in Pend Oreille district, Idaho

Stage	Igneous rock	Noncalcareous sediments	Calcareous sediments
1	Magma molten	A general and uniform recrystallization, with obliteration of the undulatory extinction of the quartz, elimination of the sericite, and the formation of more or less feldspar.	Marmarization, usually with no silicates.
2	Magma crystallizing and margin probably solid.	Pneumatolytic emanations carrying the elements known as mineralizers. The minerals that formed are andalusite, biotite, cordierite, tourmaline, quartz, apatite, and zircon.	Pneumatolytic emanations carrying the elements known as mineralizers. The minerals formed are phlogopite, biotite, spinel, diopside, grossularite, iron-bearing garnet, epidote, vesuvianite, fluorite, quartz, topaz, zircon, apatite, tremolite, and actinolite.
3	Igneous rock solid and attacked by abyssal pneumatolytic and hydrothermal emanations. The pyrogenetic minerals replaced by the new minerals muscovite, sericite, chlorite, magnetite (or ilmenite), pyrite, quartz, ore minerals, and ankerite.	Abyssal pneumatolytic and hydrothermal emanations formed muscovite, sericite, chlorite, magnetite (or ilmenite), pyrite, pyrrhotite, and quartz.	Abyssal pneumatolytic and hydrothermal emanations formed sericite, chlorite, magnetite, pyrite, pyrrhotite, quartz, a series of metallic sulphides, drusy quartz, zeolites, ankerite, and stibnite.

SUMMARY OF THE EVIDENCE THAT THE METAMORPHISM WENT ON IN STAGES

1. The marmarization of the limestone and recrystallization of the siliceous sediments, with the formation of a sodic plagioclase and an elimination of the original potash mica, preceded the formation of all the other minerals, as shown by the mutual boundary relations and by the uniformly widespread distribution of the minerals formed by the early processes. The rock had been recrystallized to an allotriomorphic aggregate of even grain, and except in zones of most intense metamorphism the grain size is about the same as it was originally.

2. The second stage is represented by minerals the grains of many of which are much larger than those of the first stage and clearly replace them. In the noncalcareous sediments no order of crystallization of minerals of the second stage can be determined, and it is probable that they were in great part contemporaneous. In some of the garnet rocks, however, a definite sequence from pyroxene to amphibole was shown. Many of the minerals formed in the second stage are known to require a high temperature during their formation. The minerals are generally considered to be pneumatolytic, and many required the so-called mineralizers for their formation. The irregular distribution of these minerals in some beds and their grouping into spots and streaks imply that they resulted from solutions that passed through the solid rock with more difficulty than the first solutions.

3. The third stage is the only one whose results are to be found in both exomorphic and endomorphic zones, and this fact suggests that during the earlier stages either the igneous rock was not yet solid, or it was in equilibrium with the solutions that caused the metamorphism in the sediments. The later age of the minerals characteristic of the third stage is clearly proved by the boundary relations of the grains. The minerals of the third stage are varieties known to be late in most paragenetic sequences and to form at rather moderate temperatures, and their period of formation is generally spoken of as hydrothermal.

4. The points above given prove that the metamorphism of the sediments was progressive; but that the stages were to some extent independent of one another is proved by the finding of many places where the products of one of the stages have formed more abundantly than those of the others, without apparent regard to distance from contacts. Thus the three periods must be considered as overlapping.

MINERALOGY

No unusual species were found during the study. However, the species represented include a large proportion of those minerals generally considered to be commonly due to contact-metamorphic processes. Zircon had possibly not previously been considered a common contact mineral, but the writer has already published the evidence for knowing it to be here a contact mineral and has cited several other occurrences reported from contact zones.

At no point were large and well-formed crystals of any of the contact minerals found. In most specimens the grains are less than 1 millimeter in diameter, hence necessarily most of the study was made under the microscope. In addition to the examination of about 150 thin sections a considerable number of the specimens were crushed, the grains separated into fractions, and the properties determined by index liquids. Thus the constants of most of the minerals were determined from several specimens. The tabulation that follows may be of interest.

Minerals of the contact-metamorphosed rocks of the Pend Oreille district, Idaho

Mineral	Rock occurrence	Locality	Properties	Remarks
Sulphides				
Stibnite	Cambrian limestone	Baptist claims on Vulcan Hill.		Needlelike crystals in garnet rock.
Molybdenite	Packsaddle Mountain granodiorite.	McDonnah No. 2 claim, Vulcan Hill.		In endomorphosed zone.
Sphalerite	Cambrian limestone	Keno claims, Vulcan Hill.		Disseminated in limestone in small quantity.
Pyrrhotite	All rocks	General		Disseminated in small anhedral grains throughout metamorphosed rocks.
Pyrite	do	do		Disseminated in small crystals throughout metamorphosed rocks.
Haloids				
Fluorite	Wallace formation, Cambrian limestone.	Zones of intense metamorphism.		Not common. Seen only microscopically and in small quantity.
Oxides				
Quartz	All rocks	General		Found as veins in noncalcareous sediments and as a replacement mineral in calcareous rocks.
Corundum	Cambrian limestone	Keno Claim, Vulcan Hill.		Found in microscopic crystals in 1 specimen of an impure limestone.
Hematite	All rocks			Direct formation by contact metamorphism not proved.
Rutile	do	General		In minute quantities where found. Most abundant in the metamorphosed Blacktail beds at the mouth of Fall Creek.
Spinel	Cambrian limestone	Along southeast shore of lake and lower workings of Arcade group, Vulcan Hill.		Found in minute green crystals in thin sections of 2 specimens.
Magnetite	All rocks	General		One of the most widespread minerals due to contact metamorphism.
Ilmenite	do	do		Associated with magnetite and identified by chemical test.
Carbonates				
Calcite	Most of the rocks	Numerous places		Occurs mostly in the calcareous rocks and is there due to simple recrystallization. Also in subordinate quantities in a few specimens of other rocks.
Dolomite	Wallace and Striped Peak formations and Cambrian limestone.	do		Minute rhombs widespread in the Wallace formation and Cambrian limestone.
Ankerite	Wallace and Striped Peak formations, Cambrian limestone, and igneous rocks.	Widespread		Parallels dolomite in occurrence, but found also in veins. One of the last minerals in the metamorphic sequence.
Phosphates				
Apatite	All rocks	In zones of intense metamorphism. Rare but does occur in weakly metamorphosed rocks.	Biaxial in some specimens with a large optic angle.	
Silicates				
Microcline	All rocks but rare in the purely calcareous beds of the Cambrian limestone.	Many places		Disseminated in microscopic grains but is nowhere so abundant as the albite-oligoclase.
Albite to oligoclase albite.	All rocks except the purely calcareous beds of the Cambrian limestone.	General	Symmetrical extinction angles range from 5° to 13°, 10° being average.	Exceedingly widespread, forming the principal constituent of some adinoles.
Bytownite	Cambrian limestone	Overlying the granodiorite sill on Cape Horn.	Symmetrical extinction angles of 40°, β=1.570.	Found in microscopic crystals at one place.

Minerals of the contact-metamorphosed rocks of the Pend Oreille district, Idaho—Continued

Mineral	Rock occurrence	Locality	Properties	Remarks
Silicates—Continued				
Diopside and augite.	Only in calcareous rocks.	In all strongly metamorphosed beds of Wallace formation and Cambrian limestone.	β=1.690 to 1.710; color from white to green.	Always found in small anhedrons.
Hornblende	Wallace formation	In all strongly metamorphosed beds.	β=1.635 to 1.645, pleochroism weak, Z slate, X nearly colorless, c $\wedge$ Z about 25°.	In minute prismatic grains, forming felted masses.
Actinolite	Metamorphosed carbonate vein in old fault zone in the Wallace formation.	West side of NW. ¼ SW. ¼ sec. 3, T. 54 N., R. 2 W.	Strongly pleochroic in green tints.	Long needle-like crystals in quartz.
Tremolite	Cambrian limestone	At many places, most conspicuously in the cliffs along the southeast shore.	Clear white. β=1.620, c $\wedge$ Z 15°.	
Pyrope, grossularite andradite.	Only in calcareous rocks.	In most strongly metamorphosed beds.	Conspicuously zoned, alternately anisotropic at Vulcan Hill n=1.700 to 1.82. Colorless, or yellow to deep red.	Composition differs markedly in adjacent beds.
Olivine	Cambrian limestone	At one place on the south side of Cape Horn.	Optically +	In microscopic crystals, altering to serpentine.
Scapolite	Cambrian limestone and Wallace formation.	Baptist claim, Vulcan Hill; shore below Three Sisters.		Wernerite, found only in small crystals.
Vesuvianite	Cambrian limestone	Keno claim, Vulcan Hill.	ω=about 1.720; pleochroic in yellow tints.	Disseminated in microscopic anhedrons. One large crystal found.
Zircon	All rocks	General		In very minute crystals. See Am. Mineralogist, vol. 10, pp. 187–194, 1925.
Topaz	Cambrian limestone	Overlying sill on south side of Cape Horn.	α=1.620, γ=1.628	Microscopic crystals in one specimen of garnet rock.
Andalusite	All noncalcareous rocks	In all strongly metamorphosed zones.	Typical	In microscopic crystals many of which were more or less altered to finely divided mica flakes.
Zoisite	Calcareous rocks	Abundant in some beds of intensely metamorphosed Wallace, west shore below Three Sisters.	β=about 1.710	
Epidote	Endomorphosed igneous rocks and in calcareous rocks.	Many places	Typical	In microscopic crystals and anhedrons.
Allanite	Cambrian limestone	Over sill on south side of Cape Horn.		Rare accessory in one specimen.
Chondrodite	do	In cliffs along southeast shore south of South Gold Creek.	α=1.635, β=1.648, γ=1.665, X pale yellow, Y, Z colorless. 2V near 90°, optically +	Common at the one locality in microscopic crystals.
Muscovite	All noncalcareous rocks	General	2V about 25°–40°	Very widespread.
Lepidomelane	do	do	In noncalcareous sediments; β=1.640 to 1.655, 2V 0. Pleochroism to deep brown.	Do.
Biotite	Calcareous rocks	Typical in the Wallace on west shore below Three Sisters.	β=about 1.620. Pleochroism in pale tints of brown.	Abundant locally.
Phlogopite	Cambrian limestone	Cliffs southeast shore of lake and at Vulcan Hill.	α=1.560, γ=1.600, X yellow, Y, Z greenish to pale yellow, 2V 0.	Rare.
Sericite	All rocks	General		Distinguished from muscovite by minute size of flakes and belongs to a later generation.
Cordierite	Blacktail formation	North side of Cape Horn.	Identified by optic sign, pleochroism, and habit.	Found in only one specimen.
Tourmaline	All noncalcareous sediments.	General	ω=1.660 to 1.670, ϵ=1.630 to 1.635. Pleochroism from colorless to blue, green, or brown.	A microscopic constituent in all moderately metamorphosed beds. Absent from severely metamorphosed beds except locally.
Aphrosiderite	Many beds of the noncalcareous sediments.	do	β=1.620. Optically + also −	Widely distributed especially in the moderately metamorphosed beds.
Thuringite	do	do	β=1.645. Optically −	Do.

Minerals of the contact-metamorphosed rocks of the Pend Oreille district, Idaho—Continued

Mineral	Rock occurrence	Locality	Properties	Remarks
Silicates—Continued				
Antigorite	Cambrian limestone	Cliffs along southeast shore.	$\beta=1.555$; birefringence weak.	In veinlets or as minute masses in the limestone.
Chrysotile	do	Lime quarries at Bayview.	$\alpha=1.500$, $\gamma=1.520$	Asbestiform material along seams.
Heulandite	do	Upper adit of Arcade group Vulcan Hill.	$\alpha=1.495$, $\gamma=1.503$. Optically +, 2V small.	Crystals in open cavities with ankerite.
Unidentified zeolite.	do	do	n=low. Birefringence 0.015.	Found with heulandite.
Titanite	All calcareous rocks	General	Typical	Microscopic crystals locally abundant.

CONCLUSIONS

The conception that contact metamorphism went on in stages is simply a division of the long-continued process into overlapping periods. The idea of stages in contact metamorphism was perhaps first suggested by Spurr, Garrey, and Fenner [6] and has subsequently been advanced by several others. Umpleby [7] used the word "stages" in describing the metamorphism at Mackay, Idaho, and the idea is also expressed in the paragenetic tables of Eckerman.[8] The descriptions of the Edwards zinc mine by Smyth [9] and by Wade and Wandke [10] speak of stages in the metamorphism.

The knowledge that contact metamorphism was produced by a long procession of solutions that came out from the igneous rock has succeeded the old idea that it was a simple recrystallization of material already present, caused by the baking heat transmitted by conduction from an adjacent intrusive. This knowledge is of importance not only in itself but also because of the light that it throws upon the process of crystallization of the igneous rocks and the nature of the residual liquids given off during crystallization.

Although contact metamorphism was not a simple recrystallization, this study of the rocks in the Pend Oreille district shows that the nature of the original rock was a very important factor in determining the kind of product resulting from the metamorphism. It shows further that, except very locally, the amount of material introduced permanently was not great in proportion to the mass of the rock. The paragenetic sequence of the minerals due to contact metamorphism suggests that the progress of crystallization of the igneous rock and the resulting metamorphism was somewhat as follows:

1. During intrusion and early crystallization of the magma, which marked the first stage, steam carrying soda and some potash was given off in considerable quantity from the whole magma and caused the country rocks to be recrystallized. Limestones were turned to marbles, and sericitic quartzites and siliceous argillites became adinoles. Connate and surface water played a part in this recrystallization.

2. With the progress of crystallization of the igneous rock, which marked the second stage, the emanations given off from the residual liquid became richer in the so-called mineralizers. In addition iron, phosphorus, zirconia, and silica, probably magnesia, rare earths, and titanium, and possibly alumina and potash were also carried. Before this stage was over the igneous rock was solid at its borders, endomorphic muscovite had formed, and throughout the body of the igneous rock deuteric titanite, allanite, and probably apatite and zircon had crystallized by replacement.

3. After final consolidation of the main mass of the intrusive which introduced the third stage, when the temperature had become lower, hot waters rich in potash and iron and later in carbon dioxide, sulphur and base and precious metals were given off from greater depths. Sericite, chlorite, and magnetite formed both in the igneous rock and in the sediments and were followed by sulphides, carbonates, and zeolites.

[6] Spurr, J. E., Garrey, G. H., and Fenner, C. N., A contact-metamorphic ore deposit; the Dolores mine, at Matehuala, San Luis Potosi, Mexico: Econ. Geology, vol. 7, pp. 471–474, 1912.

[7] Umpleby, J. B., Geology and ore deposits of the Mackay region, Idaho: U. S. Geol. Survey Prof. Paper 97, p. 65, 1917.

[8] Eckerman, H. von, The rocks and contact minerals of the Mansjö Mountain: Geol. Fören. Förh., 1922, p. 343.

[9] Smyth, C. H., jr., Genesis of the zinc ores of the Edwards district, St. Lawrence County, N. Y.: New York State Mus. Bull. 201, p. 28, 1917.

[10] Wade, W. R., and Wandke, Alfred, A big zinc mine in New York State: Eng. and Min. Jour. Press, vol. 116, p. 96, 1923.

EARLY PLEISTOCENE GLACIATION IN IDAHO

By Clyde P. Ross

ABSTRACT

On and near Railroad Ridge, Custer County, Idaho, there are deposits of unconsolidated material, believed to be of glacial origin, which are out of harmony with the present topography. The evidence suggests that this material is the product of glaciation in early Pleistocene, possibly Nebraskan time and indicates that at that time the country was far less rugged than at present. Valley cutting of the order of 1,400 feet took place in the interval between this early glaciation and Wisconsin time, when the glacial cirques and U-shaped valleys that are prominent features of the present topography were formed. This deep cutting is in marked contrast with the erosion of 100 feet or less that has occurred in Recent time.

Similar though less complete evidence of ancient glaciation followed by deep erosion is known on Loon Creek, farther north, and on the Little Wood River, farther south. All three localities are close to the highest peaks in Idaho. Evidence of similar conditions in the Pleistocene epoch has been found by other investigators in Montana, Wyoming, and perhaps Washington. Although two widely separated periods of glaciation are recorded in northern Idaho, neither appears to be as old as that here described.

FOREWORD

It has long been known that the mountains of central and northern Idaho bear much evidence of former glaciation. The glacial phenomena commonly observed are confined to existing valleys and were doubtless produced late in the Pleistocene epoch. Recent work indicates that there was in addition a much earlier glaciation, which preceded the development of the present topography. The best evidence for this now known is found on and near Railroad Ridge, in unsurveyed Tps. 9 and 10 N., R. 16 E., south of the Salmon River, in the southern part of Custer County, but deposits that may have been produced during the ancient glaciation have been noted on Loon Creek, farther north, and on the Little Wood River, in Blaine County, farther south. It may be significant that all these localities, the only known places where there are traces of the ancient glaciation, are near the highest summits in Idaho.

RAILROAD RIDGE

DESCRIPTION

The smooth, almost level top of Railroad Ridge contrasts so sharply with the ragged pinnacles around it that it has long been an object of local interest. This ridge and much of the Custer quadrangle, in which it lies, were visited during the summer of 1928 by the writer and his assistants, Robert R. LeClercq and William D. Mark, in the course of a geologic study of this part of Idaho.

Railroad Ridge was so named because of the fancied resemblance of the ridge, viewed from a distance, to a railroad grade. Plate 17 shows the general appearance of the ridge, and, by way of contrast, the cirque cliff known as the Chinese Wall, a short distance to the west. As can be seen from Plate 16 the ridge trends nearly east, and its highest and smoothest part is 2 miles long and has an average width of somewhat less than half a mile. This portion slopes from an altitude of 10,400 feet above sea level near its west end to 9,400 feet at its east end. It therefore has an average grade of 10 per cent, which appears flat when contrasted with the cliffs and steep slopes on every side. From the high west end of the ridge there is an abrupt descent of over 1,400 feet to the lake at the head of Livingston Creek, and on the south there is a comparably sharp drop into Jim Creek. On the east and north the contrast with the surrounding area is less striking, as much of the country here is a dissected plateau whose surface lies more than 9,500 feet above sea level, as shown on Plate 16. (See also pl. 18, *A*.) The streams, however, are rapidly cutting into the plateau, and two of them, Lake and Silver Rule Creeks, are already cutting into Railroad Ridge. Plate 18, *A*, shows the extent to which Silver Rule Creek is incised below the plateau level. Much of the cutting of these streams took place in Pleistocene time, for the valleys of all the larger streams, including those just mentioned, have been glaciated in their upper reaches, as shown in Plate 19.

The top of Railroad Ridge is only about 400 feet below the highest peak in its immediate vicinity and about 1,400 feet below the summit of Castle Peak, 7 miles to the south. Castle Peak is the highest peak in the Custer quadrangle and one of the highest in Idaho, being exceeded in altitude only by Mount Hyndman, 12,078 feet, which is 30 miles southeast of it, and by a few peaks of intermediate height along the range between these two.[1]

[1] Since this was written Lee Morrison, of the U. S. Geological Survey, has found that one of the peaks in the Lost River Range is nearly 600 feet higher than this.

Railroad Ridge is remarkable not only for the topographic characteristics and relations sketched above but also for its geologic composition. Plate 16 shows the geology of this ridge and its vicinity. The boundaries on this map at a distance from the ridge are subject to revision in detail, but the essential features are correctly shown. A large area on Railroad Ridge and smaller areas both north and south of it are mapped as glacial detritus. As will be seen below, the proof of the glacial origin of this material is not absolute, but it is regarded as highly probable. The boundaries of these areas were traced with care, but it is possible that there are additional small areas. Numerous erratic boulders on Slate and Livingston Creeks, composed mainly of granitic rock, suggest the former presence of morainal deposits at high levels. Boulder Creek and the region south of it were not reached during the present study, but in the summer of 1929 Thomas H. Hite found extensive deposits of similar detritus on the ridges near the mouth of Boulder Creek and south of it.

Plates 17, *A*, 18, *B*, and 19, *A*, show that the deposit on Railroad Ridge is coarse, poorly sorted, and poorly rounded. Nearly all of it is composed of granitic rock like that at the head of Boulder Creek mapped on Plate 1. Near the base fragments of the rocks underlying this great mass of débris have been incorporated in it, but even here they are subordinate in size and amount to the granitic rock. The deposit is sufficiently compacted to form slopes steeper than would be maintained in loose gravel but contains little or no cementing material. The granitic rock composing most of the fragments has been sufficiently weathered to roughen the surface, but most of the boulders are firm and fairly fresh. The material composing the smaller detached masses is essentially similar to that in the main body described above. In all exposures noted there is little variation either in average coarseness or in degree of rounding of the material.

Railroad Ridge was referred to by Umpleby [1a] as one of the remnants of a postulated peneplain which he considered to be of Eocene age. This assumption, however, was based on observation from a distance of over 12 miles. The description of the ridge given above shows clearly that its surface can not be as old as Umpleby supposed.

AGE

There can be little doubt that this material is of early Pleistocene age. Its lack of consolidation, its freshness, and the fact that it is spread, in large part, over a surface carved on tilted lava and tuff beds of Tertiary (probably Miocene) age all indicate that it is geologically young, probably not older than Pleistocene. It appears from physiographic studies elsewhere in the general region, already presented in outline [2] and to be further discussed in later papers, that a mass of coarse detritus so situated can not be older than late Pliocene, but its lack of accord with modern drainage lines shows that it is older than the present erosion cycle. As the upper reaches of all the larger stream valleys in the high mountains show abundant evidence of late Pleistocene (Wisconsin) glaciation it is evident that detritus formed earlier than these valleys must be older than that formed in Wisconsin time. Direct evidence that the material is older than the Wisconsin glaciation is furnished by the fact that it forms the upper part of the glacial cirque wall near the head of Silver Rule Creek, as can be seen in Plate 19, *A*. Similarly the detrital mass is truncated by the upper part of the valley of Jim Creek, which has a cirque at its head and has the typical U-shaped cross section of a glaciated valley, as is evident from Plate 19, *B*.

ORIGIN

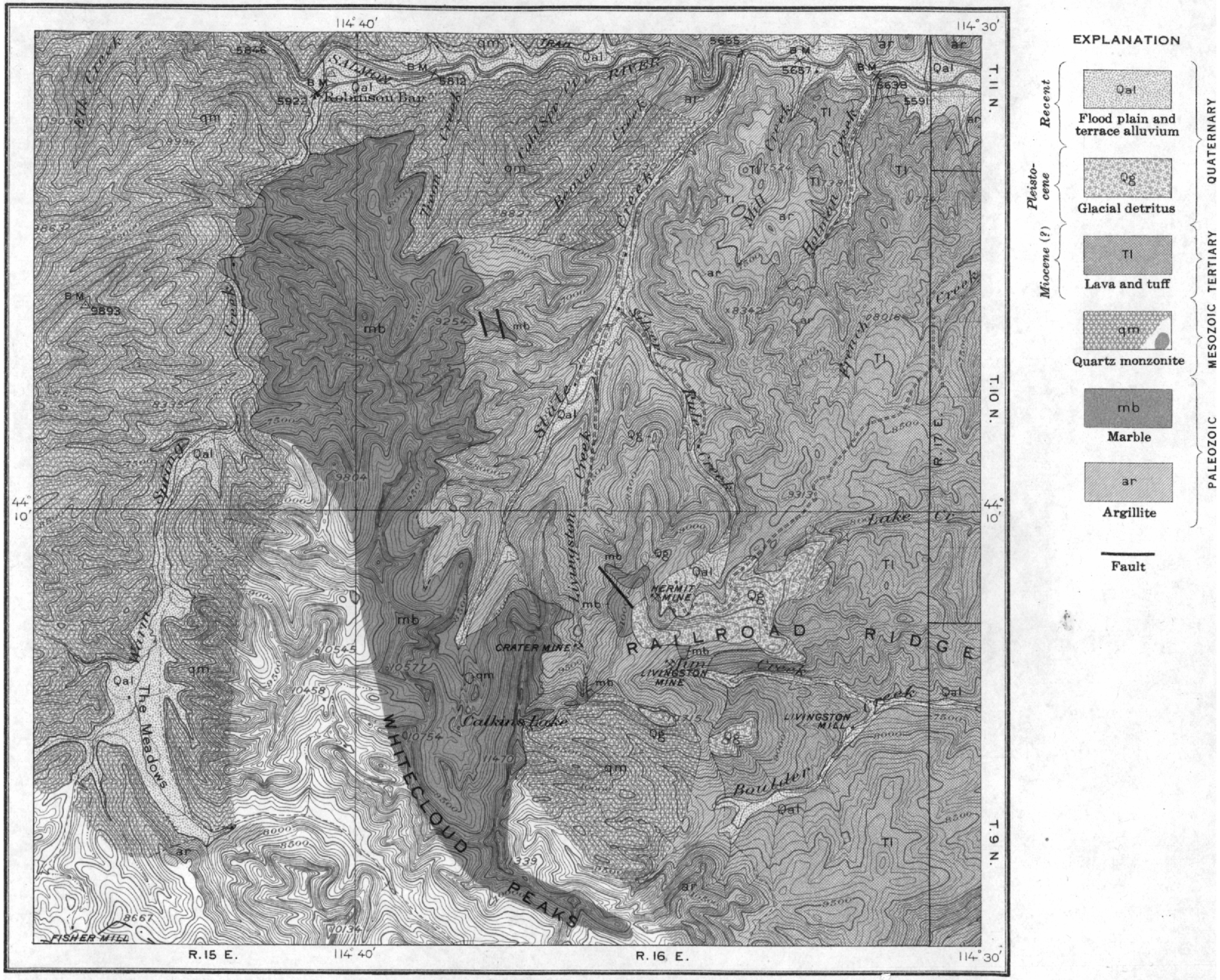
There is no positive way of distinguishing between many glacial deposits and the deposits of a torrential stream from the appearance of the material alone. The hypothesis of glacial origin fits the general relations of this material, however, far better than any other. The former presence in this location of a stream of adequate size and power to lay down such a deposit is almost inconceivable. Furthermore, the shape of the surface on which the material lies, the apparent lack of space for a drainage basin of adequate size at altitudes above the present position of the material, and the strictly localized source of most of it, all argue against a fluviatile origin.

The detritus lies on a gently inclined surface that shows no evidence of having been a part of a stream valley. Along the north side of Jim Creek and at the west end of Railroad Ridge the deposit of detritus is thin. Here it is evident that the surface on which this material was laid down is irregular in detail and contains small shallow undrained hollows. This surface is now covered with soil and fragmental material derived in part from the mass of detritus which lies above it, in part from the weathering of the bedrock, which is here mainly slate. Hence no opportunity was afforded at any locality visited to determine whether the bedrock surface is polished or grooved by a glacier.

Not only does the shape of the surface under the detritus on Railroad Ridge differ from that of a stream valley but the relations of the several patches of this material to one another and to the existing topography seem to preclude a fluviatile origin. Inspection of Plate 16 indicates that the country at the time the detritus was deposited was rolling and had only about

[1a] Umpleby, J. B., Some ore deposits in northwestern Custer County, Idaho: U. S. Geol. Survey Bull. 539, p. 15, 1913.

[2] Ross, C. P., Salient features of the geology of south-central Idaho: Washington Acad. Sci. Jour., vol. 18, p. 268, 1928.



Base from U. S. G. S. topographic map of Custer quadrangle

RECONNAISSANCE GEOLOGIC MAP OF THE AREA AROUND RAILROAD RIDGE, CUSTER COUNTY, IDAHO

Geology by C. P. Ross

Scale $\frac{1}{125000}$

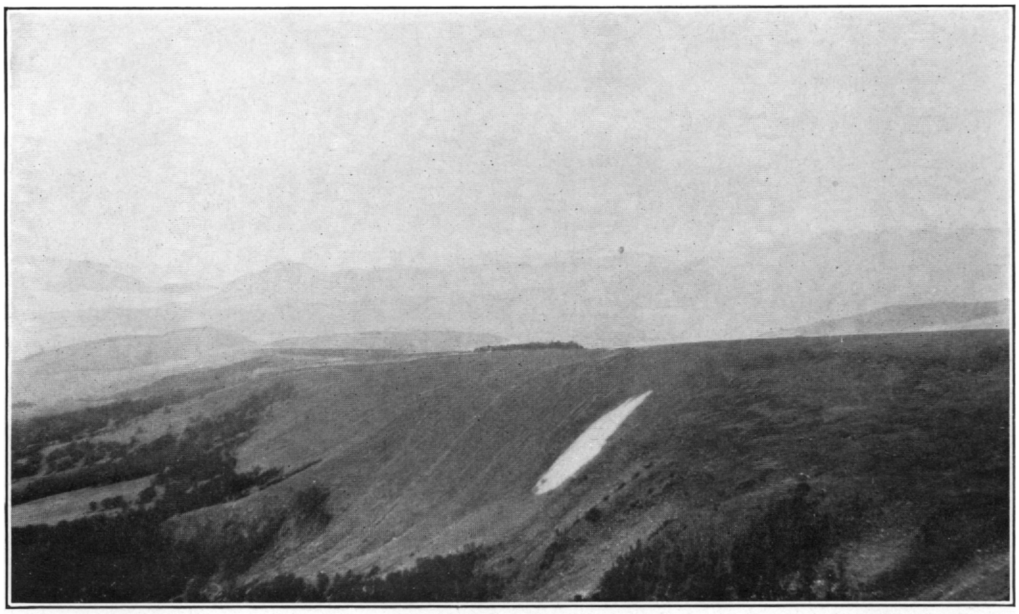
1 ½ 0 1 2 3 4 Miles

Contour interval 100 feet.

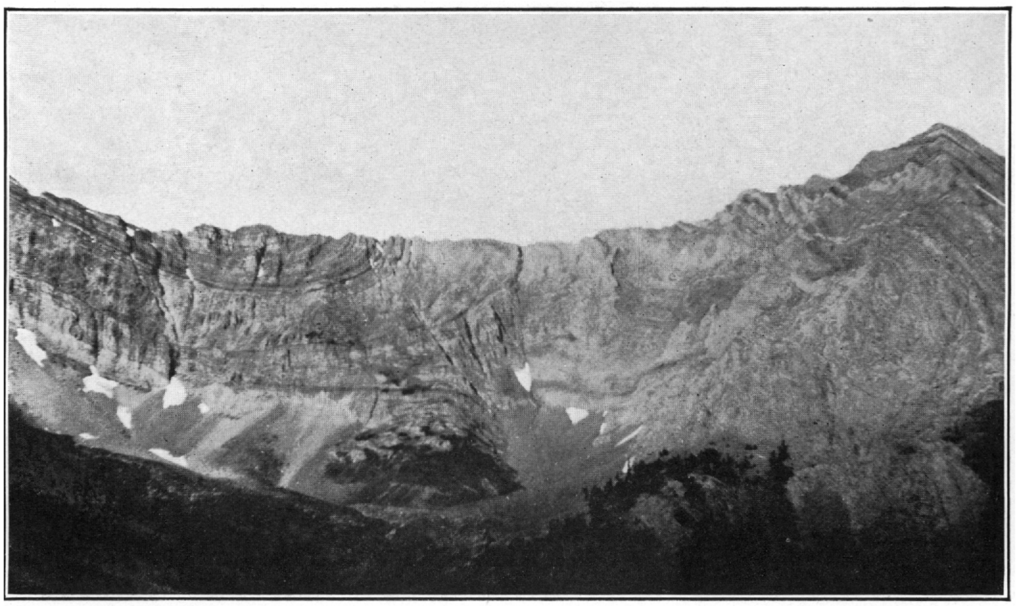
Datum is mean sea level.

1929

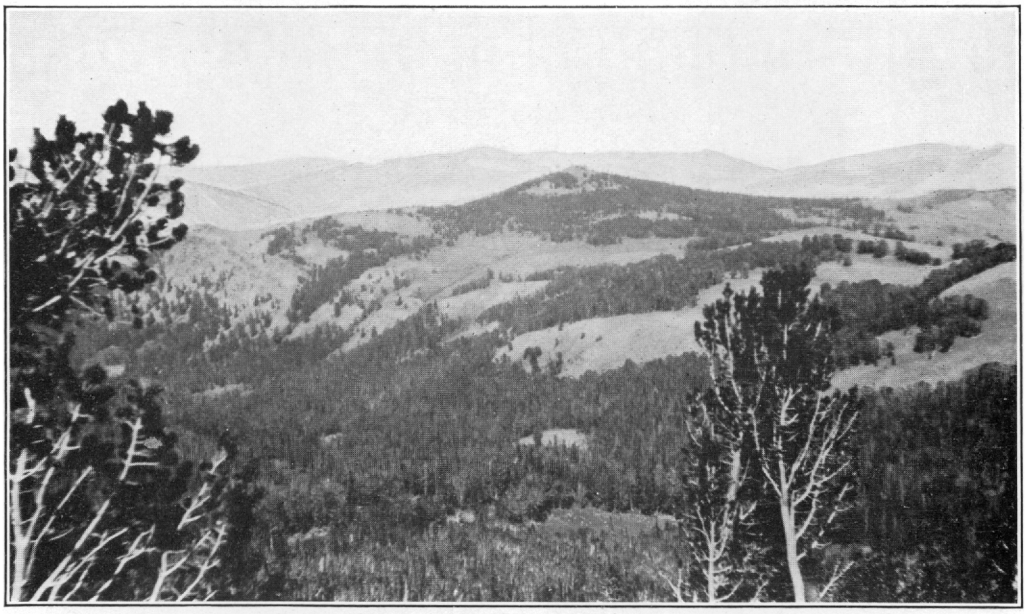
ENGRAVED AND PRINTED BY THE U.S. GEOLOGICAL SURVEY



A. RAILROAD RIDGE, IDAHO, LOOKING EAST FROM A POINT NEAR THE HIGHEST PART OF THE RIDGE



B. THE "CHINESE WALL" AT THE HEAD OF LIVINGSTON CREEK, IDAHO

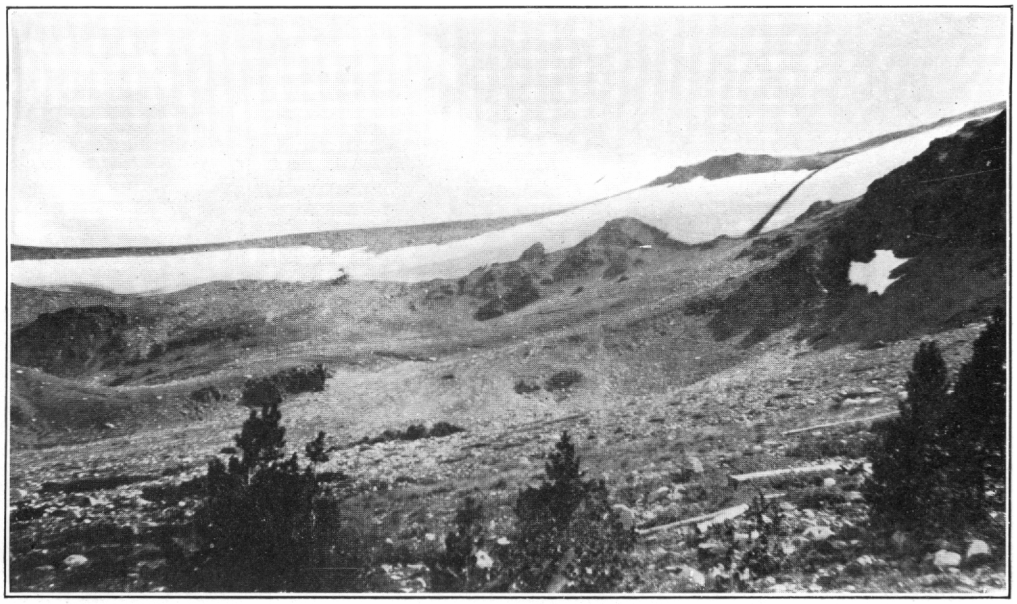


A. THE DISSECTED PLATEAU CARVED ON TERTIARY VOLCANIC ROCKS AND THE BASIN AT THE HEAD OF SILVER RULE CREEK, IDAHO

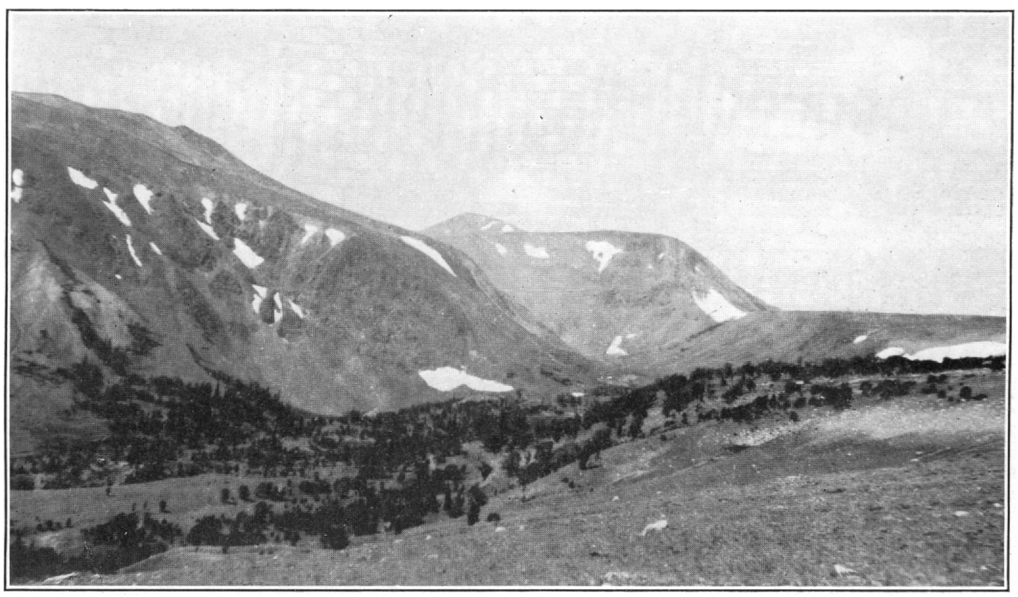
View northeast from a point just below the Hermit mine.

B. CLOSE VIEW OF THE DETRITUS COMPOSING RAILROAD RIDGE, IDAHO

Exposure is near the top of the ridge east of the Hermit mine.



A. CIRQUE CARVED IN DETRITUS AND PALEOZOIC STRATA AT THE HEAD OF SILVER RULE CREEK, IDAHO



B. THE HEAD OF JIM CREEK, LOOKING WEST FROM A POINT NEAR THE NORTHEAST END OF THE DETRITUS ON RAILROAD RIDGE, IDAHO

half as much relief as it has now. In order to present this fact more definitely an attempt is made in Figure 18 to restore the topography as it was at the time the detritus was being deposited. This figure is constructed on the basis of the distribution of the existing remnants of the detrital masses, the shape of the surfaces on which they rest, and the present topography. If possible later deformation is disregarded, it is evident that all parts of the surface not now protected by detritus are relatively lower than they were at the time of deposition. Hence, hills now higher than the existing detritus represent high ground at the time of deposition. This is the principal basis on which contours were interpolated on Figure 18 in areas not now covered by detritus. It is obvious that the detritus has suffered from erosion and was originally more extensive than it is now. Possibly it originally covered much of the area contoured in Figure 18. Its distribution with reference to the contours indicates that there is little correspondence between the location of the portions now remaining and the depressions of the old surface. Obviously the contouring in Figure 18 can not be accurate in detail. It is well within the bounds of possibility that a more pronounced depression existed along or near the course of Silver Rule Creek than is shown in that figure. The patches of detritus remaining on the ridge between Silver Rule and Livingston Creeks may have been deposited on the flanks of such a depression.

In other parts of south-central Idaho that have been studied by the writer there is evidence of minor faulting and of broad-scale uplift in the Quaternary period. The great amount of erosion in the region here considered, indicated by the contrast between the contouring on Figure 18 and that on Plate 16, is probably to be ascribed in part to differential uplift, with consequent rejuvenation of the streams. However, all the deformation of the rocks of which evidence was observed here probably took place long before the deposition of the detrital material under discussion. The Tertiary volcanic strata are warped but not intricately deformed. Hence it is improbable that any crustal deformation that may have taken place subsequent to the formation of the surface which Figure 18 is intended to portray was of such a character as to affect materially the accuracy of the restoration attempted in this figure. It appears from studies here and elsewhere in this general region that in late Pliocene or early Pleistocene time the region had comparatively low relief. Plate 18, *A*, shows a portion of this old surface, now considerably dissected. If due allowance is made for later erosion, this picture may be taken to give an idea of the topography existing at the time the material under consideration was being laid down—that is, the topography diagrammatically contoured in Figure 18.

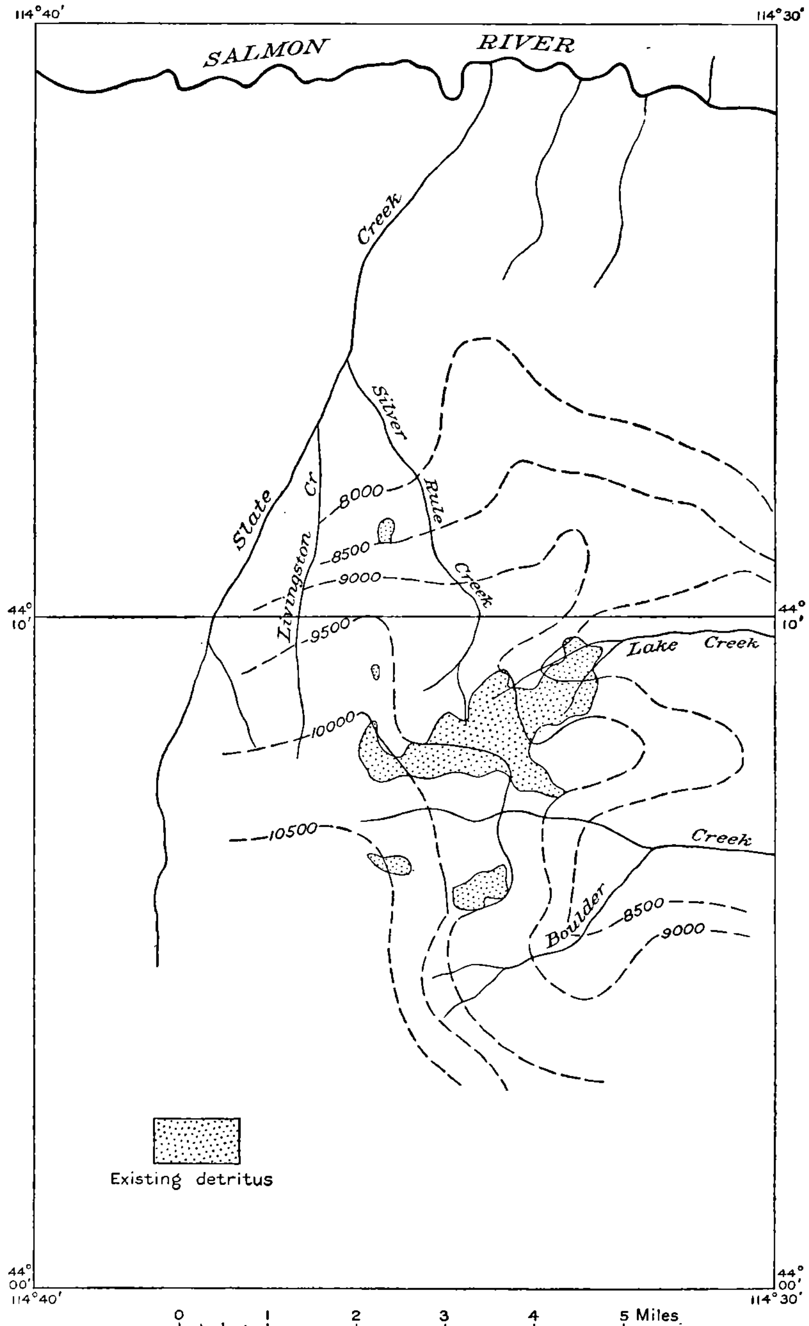


FIGURE 18.—Restoration of the surface on which the detritus on Railroad Ridge, Custer County, Idaho, was deposited

In general, it appears that the evidence afforded by the apparent shape of the surface on which the detritus was deposited is opposed to the theory that this material was laid down in ordinary streams. The evidence, both as to the general topography before the deposition of this material and as to the details of the shape of the surface that is now being uncovered along the borders of the detrital mass on Railroad Ridge, is much more nearly in accord with the conception of deposition through the agency of glaciation.

Further evidence against the conception of a fluviatile origin is furnished by the apparent lack of a drainage basin of sufficient size to feed a stream competent to transport and deposit so large a mass of gravel as that on Railroad Ridge. So far as can be seen from present exposures this material is unsorted and unstratified. Plate 18, *B*, shows the heterogeneity plainly and is typical of all the exposures seen. Such a mass of unsorted detritus, if of fluviatile origin, would be a fanglomerate and produced by a stream of considerable length and torrential velocity. Except under such abnormal conditions as are furnished by melting glaciers, or possibly by the sudden release of a large body of impounded water, a large stream could be formed only where there was a large area tributary to it. If this unsorted and poorly rounded material is of fluviatile origin, it was probably produced under arid conditions. Such conditions would decrease the area necessary, because the products of erosion would be dumped closer to their source and remain longer than where abundant water was available for transportation. Even under arid conditions, however, the area necessary would seem to be larger than appears to have been available. No other evidence of aridity in this region in early Pleistocene or late Tertiary time is known.

Railroad Ridge is at present one of the higher parts of the region, and the areas in the vicinity that rise above it are of decidedly small extent and the highest are close at hand. In order to conceive of a stream of nonglacial origin competent to deposit the detritus now present on Railroad Ridge it would be necessary to postulate the existence of a topography so different from that of the present time as to tax the imagination. The former presence of such a topography would imply pronounced orogenic movements, of which no evidence was obtained in the examination of this region. Glaciation, however, with its intensive erosion in circumscribed areas, could produce detritus comparable in amount to that under consideration by attack on an area far smaller than would be required by an ordinary stream.

The fact that nearly all the detritus is composed of granitic rock is likewise opposed to the theory of fluviatile origin. It is difficult to conceive of normal stream erosion, under any climatic conditions, being so selective in its action in a region of diverse rocks as to deposit such a mass of almost exclusively granitic gravel. Plate 16 shows that the only probable sources of this material are the rather small exposure south of Railroad Ridge and the large one west and northwest of it, which is part of the Idaho batholith proper. Both this part of the batholith and any detached masses beyond the limits of the geologic map (pl. 16) would seem to be too far away for gravel derived from them to be deposited on Railroad Ridge except under very special and improbable conditions. This objection applies to derivation by either fluviatile or glacial action. Furthermore, derivation from these sources implies a greater difference from existing topography than is probable. Hence the granitic rock south of Railroad Ridge in and near the headwater basin of Boulder Creek is regarded as the only probable source of the detritus. This mass, however, has an area of less than 6 square miles. The fact that this is the only probable source for by far the greater part of the detritus lends support to the argument in the preceding paragraph that there was insufficient area available for the accumulation of such a deposit as that here discussed by the agency of ordinary streams. Also it is surprising, if streams were the agents of accumulation, that tributaries failed to introduce material from the Paleozoic sedimentary rocks that are so abundant on all sides. Some of the detritus is 3½ miles from the edge of the granitic mass. A mountain glacier, however, derives much of its débris from erosion in the cirques at its head and hence would be more selective in its action than a stream with tributaries.

The hypothesis that the detritus on and near Railroad Ridge was derived from a glacier which has its head in the granitic rock near the present head of Boulder Creek is further supported by Figure 18. The highest part of the surface there contoured corresponds to the granitic mass at the head of Boulder Creek. With all due allowance for the inaccuracy inherent in the construction of this figure, this broad relation seems substantiated.

The character of the detritus on and near Railroad Ridge is in accord with the hypothesis of glacial origin, but the present shape of the mass on the ridge is not that of a moraine. The other detached masses have probably suffered too much from erosion for their original shapes to be recognizable. The mass on Railroad Ridge, as can be seen from Plates 17, *A*, and 19, *B*, and less vividly from Plate 16, is remarkably flat and smooth on top. This is the most striking feature of the ridge and suggests the leveling action of water rather than direct dumping from a glacier. Most moraines have irregular surfaces and rounded rather than flat tops. The mass on Railroad Ridge has large lateral dimensions in proportion to its thickness and hence most resembles a ground moraine.

Such moraines, however, have generally more or less hilly or undulating topography. If, therefore, the detritus on Railroad Ridge is of glacial origin, it has probably been somewhat smoothed by running water, possibly in part by water derived from melting of the glacial ice. The long time that has elapsed since its deposition would permit considerable smoothing of the original topography through the agency of weathering. The detached masses of similar material probably originally formed parts of the same plain, although it may be that some of them were parts of separate moraines or outwash plains deposited by the same or near-by glaciers.

LOON CREEK

One of the two other localities in Idaho where similar conditions have been noted is on the west side of Loon Creek near the Boyle ranch, in Custer County, 30 miles northwest of Railroad Ridge. This locality was examined in the course of a geologic study of the Casto quadrangle. The material is essentially similar to that on Railroad Ridge, being poorly sorted and rounded and composed mainly of granitic rock although resting on metamorphosed sedimentary rock. The boulders, which reach 10 feet in maximum diameter, are much more softened by weathering than those on Railroad Ridge. Such material was noted on the slopes of Loon and Trail Creeks up to an altitude of over 7,800 feet, which is about 2,000 feet above Loon Creek at the Boyle ranch, and it now extends down almost to stream level. The conclusion was reached that this detritus is probably a product of glaciation so much older than the Wisconsin that the record of it has been obliterated elsewhere in the vicinity. It appears to have been a moraine or similar body which has been undercut by Loon Creek, causing part of it to slide far down the slope.

LITTLE WOOD RIVER

Westgate [3] in his study of the Hailey quadrangle, which is immediately southeast of the Custer quadrangle, noted indications of glaciation older than that abundantly evident in the higher stream valleys of the region. He cites the presence of erratic boulders as much as 2 feet in diameter on the slopes above the Little Wood River at altitudes of 500 to 700 feet above the stream. Evidently the amount of old detritus remaining here is far less than in the two localities cited above, but in other respects the relations are similar.

SIMILAR CONDITIONS IN NEIGHBORING REGIONS

The localities above cited are the only ones in central Idaho where evidence of pre-Wisconsin glaciation is known, but such evidence has been found by several investigators in western Montana and Wyoming, northern Idaho, and eastern Washington. Alden [4] reports old glacial drift in Glacier National Park 1,000 to 1,300 feet above the bottoms of adjacent valleys. He believes that these valleys, which were themselves glaciated in Wisconsin time, were cut subsequent to the glaciation recorded by the old deposits. Thus conditions in Glacier National Park closely parallel those in the vicinity of Railroad Ridge, here described. Alden also finds evidence of old glaciation followed by 1,000 feet of downcutting in the Gros Ventre Valley, Wyo.[5]

Alden further believes [6] that there is strong presumptive evidence of early Pleistocene glaciation in the northern Rocky Mountains in general and emphasizes the marked erosion in pre-Wisconsin time in Montana in contrast to the comparatively small amount of erosion since then. He states that the depth of post-Wisconsin stream erosion in some of the mountain canyons in the region he describes is as great as 200 feet. Along the upper Salmon River in central Idaho recent downcutting is probably between 50 and 100 feet, partly in alluvium, partly in rock. Lindgren [7] stated that the depth of postglacial erosion by the Snake River on the western border of Idaho was limited to the removal of 300 feet of unconsolidated late Pleistocene sediments and an insignificant deepening of the early Pleistocene rock-cut channel.

Blackwelder [8] found similar evidence of old glaciation in western Wyoming and adjoining parts of Idaho, especially in and near the Teton Range and along the Buffalo Fork of the Snake River. This consists of glacial deposits, which he calls the Buffalo drift. Canyons have subsequently been cut through this drift and 200 to 1,000 feet into the underlying bedrock. The topography has been so much modified since the old glaciation as to obliterate most traces of morainal topography and to make it uncertain whether the glaciers were of valley type or parts of an ice cap. Blackwelder also found glacial drift of an intermediate stage which he designated the "Bull Lake stage." This drift, he thought, might correspond either to the Iowan or the early Wisconsin drift of the Mississippi Valley.[9]

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[3] Umpleby, J. B., Westgate, L. G., and Ross, C. P., Geology and ore deposits of the Wood River region, Idaho: U. S. Geol. Survey Bull. 814 (in press).

[4] Alden, W. C., Pre-Wisconsin glacial drift in the region of Glacier National Park, Montana: Geol. Soc. America Bull., vol. 23, pp. 687-708, 1912. Alden, W. C., and Stebinger, Eugene, Pre-Wisconsin glacial drift in the region of Glacier National Park, Mont.: Geol. Soc. America Bull., vol. 24, pp. 529-572, 1913.

[5] Alden, W. C., Landslide and flood at Gros Ventre, Wyo.: Am. Inst. Min. and Met. Eng. Tech. Pub. 140, pp. 12-13, 1928.

[6] Alden, W. C., Physiographic development of the northern Great Plains: Geol. Soc. America Bull., vol. 35, pp. 405-423, 1924; Pre-Wisconsin glaciation of northern Rocky Mountains [abstract]: Pan-Am. Geologist, vol. 43, pp. 371-372, 1925.

[7] Lindgren, Waldemar, The gold belt of the Blue Mountains of Oregon: U. S. Geol. Survey Twenty-second Ann. Rept., pt. 2, p. 585, 1901.

[8] Blackwelder, Eliot, Post-Cretaceous history of the mountains of central western Wyoming, pt. 3: Jour. Geology, vol. 23, pp. 328-333, 1915.

[9] See also Alden, W. C., Glaciation and physiography of Wind River Mountains, Wyoming [abstract]: Washington Acad. Sci. Proc., vol. 16, p. 73, 1926. Alden thinks that Blackwelder's "Bull Lake" stage may correspond to the Iowan stage of glaciation in the Mississippi Valley.

Anderson[10] has recently discussed the evidence of an early glaciation in northern Idaho. He thinks that this glaciation was more powerful than the last one and effected marked drainage changes and much erosion. He pictures the earlier glaciation as effected by thick glaciers which eroded deeply in valleys still existing, and he finds no such evidence of deep erosion subsequent to the early glaciation as has been noted in Montana and south-central Idaho. Also Anderson follows Bretz[11] in considering the earlier glaciation he describes as of either Iowan or early Wisconsin age.

Alden[12] regards the earliest glaciation in the Glacier National Park region as marking the beginning of Pleistocene time and probably contemporaneous with the Nebraskan stage of continental glaciation. In view of the similarity in conditions it is presumed that the early glaciation in south-central Idaho was contemporaneous with that in the Glacier Park region. Presumably the differences indicated for the early glaciation described by Anderson show that it was more recent than that described above.

The two better known stages of glaciation in eastern Washington[13] were of continental type and were probably more recent than the earliest glaciation in northern Montana and south-central Idaho. Leverett,[14] however, found glacial till at Cheney, southwest of Spokane, and also in the Puget Sound region, which he thought might be as old as Kansan. Perhaps this old till is to be correlated with that in south-central Idaho described above.

CONCLUSIONS

The conclusion seems inescapable that in early Pleistocene or late Pliocene time the topography of south-central Idaho was far less rugged than that of the present day or even of late Pleistocene time. Later erosion has incised valleys in the high country to depths of more than 1,000 feet, in striking contrast to the downcutting of a few score feet generally supposed to have taken place in the region since the Wisconsin glaciation. In some way considerable deposits of coarse, poorly sorted detritus were laid down in several localities in the higher parts of the region before the extensive downcutting took place. The available evidence indicates that these deposits are of glacial origin. Some, at least, were probably modified by the action of fluvioglacial water. The most probable alternative hypothesis is that they are products of torrential streams under more or less arid conditions, but this involves such profound orogenic and physiographic changes since the material was laid down as to be much less acceptable than the hypothesis of glaciation. The fact that similarly ancient glaciation followed by comparable erosion took place in Montana and Wyoming lends support to this conclusion.

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[10] Anderson, A. L., Some Miocene and Pleistocene drainage changes in northern Idaho: Idaho Bur. Mines and Geology Pamph. 18, pp. 1-28, 1927.

[11] Bretz, J. H., The age of the Spokane glaciation: Am. Jour. Sci., 5th ser., vol. 8, pp. 334-342, 1924.

[12] Alden, W. C., Physiographic development of the northern Great Plains: Geol. Soc. America Bull., vol. 35, p. 406, 1924.

[13] Pardee, J. T., Geology and mineral deposits of the Colville Indian Reservation, Wash.: U. S. Geol. Survey Bull. 677, pp. 47-53, 1918. Bretz, J. H., Glacial drainage on the Columbia Plateau: Geol. Soc. America Bull., vol. 34, pp. 573-608, 1923.

[14] Leverett, Frank, Glacial formations in the western United States: Geol. Soc. America Bull., vol. 28, p. 143, 1917.

THE FLORA OF THE FRONTIER FORMATION

By Edward Wilber Berry

INTRODUCTION

In 1917 the late F. H. Knowlton described a flora from the Frontier formation of southwestern Wyoming.[1] This description was based upon collections made in the vicinity of Cumberland, in Lincoln County, by John C. Frémont in 1843, by A. C. Veatch in 1906, by T. W. Stanton and F. H. Knowlton in 1908, by A. C. Peale in 1909, and by T. W. Stanton in 1913. Knowlton enumerated 25 species, representing 7 ferns, 1 equisetum, 1 monocotyledon, and 16 dicotyledons.

In the last few years N. H. Brown, of Lander, Wyo., has sent in several small collections of plants from the Frontier formation of the Wind River Basin, in Fremont County, Wyo. These collections were obtained within a 6-mile radius of Lander and have resulted in the addition of nine species to this flora, several of which are of especial interest and serve to give a much clearer evaluation of it in terms of chronology and environment than was before possible. The present contribution is devoted to a description of these additional species in the northeasterly extension of the formation and to a discussion of the age and environmental conditions indicated by the flora as a whole. I am much indebted to Mr. Brown for his sustained interest in this and other problems of western Wyoming.

The Frontier formation in its wider extent is a thick series of sandstone beds with a few conglomeratic lenses, with shale and coal beds, and containing, above the middle, beds of long, slender oysters (*Ostrea soleniscus*) and a small but characteristic marine fauna of Benton age. The plants from Lincoln County came from a thin whitish clay in a 30-foot light-colored clay shale of considerable lateral extent about 1,200 feet below the top of the Oyster Ridge sandstone member, which underlies the Kemmerer coal. This fine-grained matrix accounts for the excellent preservation of the plants described by Knowlton. The plant-bearing exposures of the Frontier formation in the Wind River Basin are the normal coarse basal sandstones, and, except for the coriaceous forms such as the *Nilsonia* and *Protophyllocladus*, the plants are poorly preserved.

[1] Knowlton, F. H., A fossil flora from the Frontier formation of southwestern Wyoming: U. S. Geol. Survey Prof. Paper 108, pp. 73–107, pls. 27–39, 1917.

It would seem that the Frontier formation as a whole and in a general way records a transition from continental swamp and river deposits through littoral deposits to shallow marine deposits. The beds near Lander from which the present collections were made form the basal sandstone of the Frontier as there identified and appear to me to be of continental origin and partly wind-laid, either in the lower part of a stream valley or in depressions of a beach ridge. I would not expect this phase to have been necessarily of any great extent or to have resulted in any unit thickness, but the predominance of coriaceous forms, the attitudes in which they were buried, and to a considerable extent the botanical character of the species identified all point in this direction.

CORRELATION OF FREMONT COUNTY WITH LINCOLN COUNTY FRONTIER

Although the more delicate species recorded from Lincoln County have not been found in Fremont County, especially most of the ferns, there can be no very great difference in the age of the plants from the two areas. This is proved by the presence in Fremont County of the following forms described by Knowlton as occurring in Lincoln County and not known from any other horizons:

- Anemia fremonti Knowlton.
- Cinnamomum? sp. Knowlton.
- Dryophyllum lanceolatum (Knowlton) Berry.
- Ficus fremonti Knowlton.
- Ficus? sp. Knowlton.

The fact that the plants found in the Frontier of Lincoln County are associated with representatives of Benton invertebrates, whereas in Fremont County the plants were found in the basal sandstone and the larger part of the Frontier above the plant horizon contains Niobrara invertebrates,[2] suggests that the Fremont County plants may be slightly younger than those from Lincoln County, although the range of the plants in the two areas rather points to just the opposite conclusion.

The matrix of the plant material of the Frontier formation in Fremont County varies in its lithology from a coarse friable massive grayish sandstone to a

[2] Stanton, T. W., personal communication.

finer more or less brownish or yellowish sandstone with a considerable intermixture of mudstone. In the finer material the leaf laminae are not parallel with one another nor usually flat but are more or less contorted—a condition usually ascribed to eolian sedimentation. The forms here described were associated with some fragments of fishbones, which are not precluded from occurrence in wind-blown deposits, because piscivorous birds can be relied upon to drop the bones on land surfaces near their fishing resorts or nesting places.

AGE INDICATED BY THE FLORA

In Knowlton's discussion of the age of the flora from Lincoln County he pointed out that, although all the plants were peculiar to the horizon and localities at which they were collected, their affinities seemed to be with post-Colorado forms. A part of his statement [3] is worth quoting in the present connection:

> The spleenwort (*Asplenium occidentale*) belongs to the same group as and is pretty closely related to an unpublished species from the Farmington sandstone member of the Kirtland shale (Montana group) of San Juan County, N. Mex. The shield fern (*Dryopteris coloradensis*) is strongly suggestive of an unpublished species from the Vermejo formation of the Raton Mesa region of Colorado and New Mexico. The *Anemia* is very closely related to and perhaps identical with a species common in the Eocene of Europe but also reported from the Montana at Point of Rocks, Wyo. Among the dicotyledons, *Cinnamomum hesperium* is hardly distinguishable, except in size, from *Cinnamomum wardii*, from the upper part of the Adaville formation at Hodges Pass, a few miles north of the Cumberland localities. *Dewalquea pulchella* and *Dryandroides lanceolata* are so close to *Dewalquea insignis* and *Dryandroides quercinea*, respectively, from the upper Senonian of Westphalia, that it is perhaps doubtful if they should be maintained as distinct. *Quercus stantoni* is very similar to *Quercus formosa*, also from the Senonian of Westphalia. *Ficus fremonti* agrees in shape with *Ficus proteoides* Lesquereux and in shape, size, and nervation with *Ficus lanceolata acuminata* Ettingshausen, as figured by Lesquereux, from the Dakota sandstone. This is the only species that appears to show any special likeness to forms from beds whose position is lower than that assigned to the plant beds at Cumberland.

The only one of the nine additional Frontier species discovered in the Wind River Basin that falls in with the preceding statement is the *Sequoia*, which is most like the Judith River remains referred to *Sequoia reichenbachi*. Of the remaining eight species the *Sabalites* indicates nothing with respect to age, and the other seven represent a distinctly pre-Colorado element which survived into Colorado time. The *Nilsonia* represents a type which had its origin in the Triassic, reached its maximum in the Jurassic, and died out during the Upper Cretaceous. No species as large or as characteristic as the Frontier form has been found in rocks as young as these.

The other six species were all described originally from the Dakota sandstone or recorded from it by Lesquereux, and the *Sterculia* is abundant in the eolian sandstones of the still earlier Cheyenne sandstone of southern Kansas. It is true that several of these species have been found to have a considerable time range and are not confined to pre-Colorado horizons. For example, the *Protophyllocladus* is found outside the Western Interior in beds as young as the Montana group, but its type occurrence was in the Dakota sandstone, which was its latest known occurrence in the West prior to its discovery in the Frontier. The *Phyllites crassipes* and *Ficus inaequalis* were similarly confined to the Dakota sandstone in the West but ranged to higher horizons in the Atlantic Coastal Plain and Greenland. If the indications of age furnished by these additions to the Frontier flora are integrated with the age indications, as quoted above, of the plants from Lincoln County, the result agrees perfectly both with the age indications of the Frontier invertebrates and also with our expectations of what a Colorado flora would be like. That is to say, it shows a facies of its own but consists in part of forms derived from the Dakota and others praenuncial of later Upper Cretaceous floras. This same statement is true with respect to floras of Colorado age from western Canada which I have studied: *Anemia fremonti* of the Frontier is also found in the Sukunka formation of British Columbia.

COMPOSITION OF THE FLORA

The flora from the Frontier formation now numbers 34 species, representing 24 genera in 16 families and 15 orders, as listed below.

- Arthrophyta:
  - Equisetales:
    - Equisetaceae:
      - Equisetum sp. Knowlton.
- Pteridophyta:
  - Polypodiales:
    - Polypodiaceae:
      - Tapeinidium? undulatum (Hall) Knowlton.
      - Microtaenia variabilis Knowlton.
      - Microtaenia paucifolia (Hall) Knowlton.
      - Dennstaedtia? fremonti (Hall) Knowlton.
      - Dryopteris coloradensis Knowlton.
      - Asplenium occidentale Knowlton.
      - Anemia fremonti Knowlton.
- Cycadophyta:
  - Williamsoniales:
    - Nilsonia mehli Berry, n. sp.
- Coniferophyta:
  - Sequoia reichenbachi (Geinitz) Heer.
  - Protophyllocladus subintegrifolius (Lesquereux) Berry.
- Angiospermophyta:
  - Monocotyledonae:
    - Arecales:
      - Arecaceae:
        - Sabalites sp.
    - Liliales?:
      - Smilacaceae?:
        - Smilax? coloradensis Knowlton.

[3] Knowlton, F. H., op. cit., p. 77.

Angiospermophyta—Continued.
- Dicotyledonae:
  - Myricales:
    - Myricaceae:
      - Myrica nervosa Knowlton.
  - Salicales:
    - Salicaceae:
      - Salix cumberlandensis Knowlton.
      - Salix frontierensis Knowlton.
  - Fagales:
    - Fagaceae:
      - Quercus stantoni Knowlton.
      - Dryophyllum lanceolatum (Knowlton) Berry.
  - Urticales:
    - Moraceae:
      - Ficus fremonti Knowlton.
      - Ficus inaequalis Lesquereux.
      - Ficus? sp., Knowlton.
      - Ficus? sp., Knowlton.
  - Sapindales:
    - Staphyleaceae:
      - Staphylea? fremonti Knowlton.
  - Malvales:
    - Sterculiaceae:
      - Sterculia towneri (Lesquereux) Berry.
  - Laurales:
    - Lauraceae:
      - Cinnamomum hesperium Knowlton.
      - Cinnamomum? sp., Knowlton.
  - Umbellales:
    - Araliaceae:
      - Aralia veatchii Knowlton.

Position uncertain:
- Dewalquea pulchella Knowlton.
- Phyllites ficifolius Knowlton.
- Phyllites grandifolius-cretaceus (Lesquereux) Berry.
- Phyllites cretaceus (Ettinghausen) Berry.
- Phyllites crassipes (Heer) Berry.
- Phyllites dentata Knowlton.
- Phyllites sp. Knowlton.

The Arthrophyta are represented by a single rather indifferently preserved species of *Equisetum*. The ferns furnish seven species in six genera, and all except the *Anemia* are confined to the region around Cumberland, in Lincoln County. These ferns are perhaps the most interesting element in this flora; the *Tapeinidium* is closely related to existing Malaysian and Polynesian forms; the two species of *Microtaenia* are unique and without any close relatives among existing davallioid ferns; the *Dennstaedtia* is not certainly determined, as the material is fragmentary. The genus is only sparingly represented in the geologic record and was not previously known from horizons earlier than the Fort Union. The *Dryopteris*, *Asplenium*, and *Anemia* are to be expected anywhere in the Upper Cretaceous and require no comment.

The cycads are represented by the remarkable species of *Nilsonia*, which has already been referred to. Coniferophytes constitute a very minor element in the flora, being represented by only a very few twigs of *Sequoia* and the more abundant phylloclads of the more or less problematic *Protophyllocladus*. Upper Cretaceous floras in general contain a relatively great variety of conifers, and of the two reasons that may be advanced for their seeming rarity in the Frontier—namely, unfavorable environmental conditions or lack of discovery—experience indicates that the latter is probably the true explanation and that we may look forward to finding eventually a much better representation of this class of plants.

The angiosperms comprise two monocotyledons and twenty-two dicotyledons. One of the monocotyledons is of uncertain affinities. The form which I have referred tentatively to *Sabalites* is certainly a palm, but the form which Knowlton referred tentatively to *Smilax* is not a *Smilax*, in my opinion, and there is some doubt even of its being monocotyledonous.

The dicotyledons are in no way remarkable but represent such genera as *Myrica*, *Salix*, *Dryophyllum*, *Ficus*, *Sterculia*, *Aralia*, *Phyllites*, and *Dewalquea*, which are common elements in most Upper Cretaceous floras, commencing with those of Cenomanian age. Several of the dicotyledons recorded from the Frontier formation are of rather doubtful value, notably the forms which Knowlton described as *Ficus?* sp., *Cinnamomum?* sp., and *Phyllites*.

ENVIRONMENTAL CONDITIONS INDICATED

The occurrence of numerous thick and widespread beds of coal in the Frontier formation is a conclusive indication of a humid climate at the time of their formation, and the known flora points in the same direction, especially that facies found in the clays in the vicinity of Cumberland. The plants found in Fremont County, although they do not indicate any aridity of climate, seem to me in part to indicate less humid conditions, which I would interpret as due not to a lessened rainfall but to their having grown on beaches or between dunes along a coast where insolation was high, winds were rather constant, and the sandy surface was apt to afford a less constant water table or one insufficient to keep pace with the increased evaporation due to insolation and wind. I regard the *Nilsonia*, *Protophyllocladus*, *Dryophyllum*, and *Sterculia* as being especially indicative of such an environment. Their coriaceous nature points in the same direction, and both *Dryophyllum* and *Sterculia* occur frequently as fossils in exactly this environment—*Dryophyllum* in beach and dune deposits in the sands of Aix-la-Chapelle, along the shores of the lower Eocene Paris Basin, and in the Mississippi Gulf embayment, and *Sterculia* in the Cheyenne and Dakota sandstones.

Knowlton concluded that the climate of Frontier time was tropical or subtropical, basing his conclusion on the indeterminate growth of *Tapeinidium* and the other davallioid ferns present and the occurrence of *Ficus* and *Cinnamomum*. As I have frequently pointed out, the universal tendency among paleobotanists is to demand tropical climates. If we consider the latitudinal range of the commoner Upper Cretaceous genera we either, like Knowlton, who followed Manson, accept a wholly impossible climatic control, or else, like Koppen and Wegener, we predicate a wandering pole. The number of plants as yet known from the Frontier formation is entirely too small to warrant any attempt to put forward an elaborate climatic discussion, but a few remarks are not inappropriate. In the first place, most discussions of climate that have been based on fossil organisms fail to differentiate between the geographic occurrence of the most closely related existing forms and the actual climate in which they live. Every country in the tropical zone is assumed to have a tropical climate, whereas, as a matter of fact, the altitudinal climatic zones in the Tropics may run up to those of Arctic conditions, as they do in all the Andean countries of South America. The plants most commonly thought of as tropical—these Frontier davallioid ferns, for example, or tree ferns in general—find their optimum modern conditions in temperate rain forests, not in tropical lowlands. We can only infer the climatic requirements of such wholly extinct Frontier genera as *Nilsonia* and *Protophyllocladus*, but of the Frontier genera that are represented in existing floras—namely, *Equisetum, Sequoia, Myrica, Salix, Quercus, Dryophyllum, Ficus, Staphylea, Sterculia, Cinnamomum,* and *Aralia*—some are decidedly temperate types, *Equisetum* and *Salix* extending into the Arctic zone, and none are out of place in a warm temperate climate.

As already stated, the Frontier plants are too few to afford the basis for a more conclusive analysis, but my inference would be that they indicate a warm temperate and not a subtropical or tropical climate.

SYSTEMATIC DESCRIPTIONS

Phylum CYCADOPHYTA

Order WILLIAMSONIALES

Genus NILSONIA Brongniart

Nilsonia mehli Berry, n. sp.

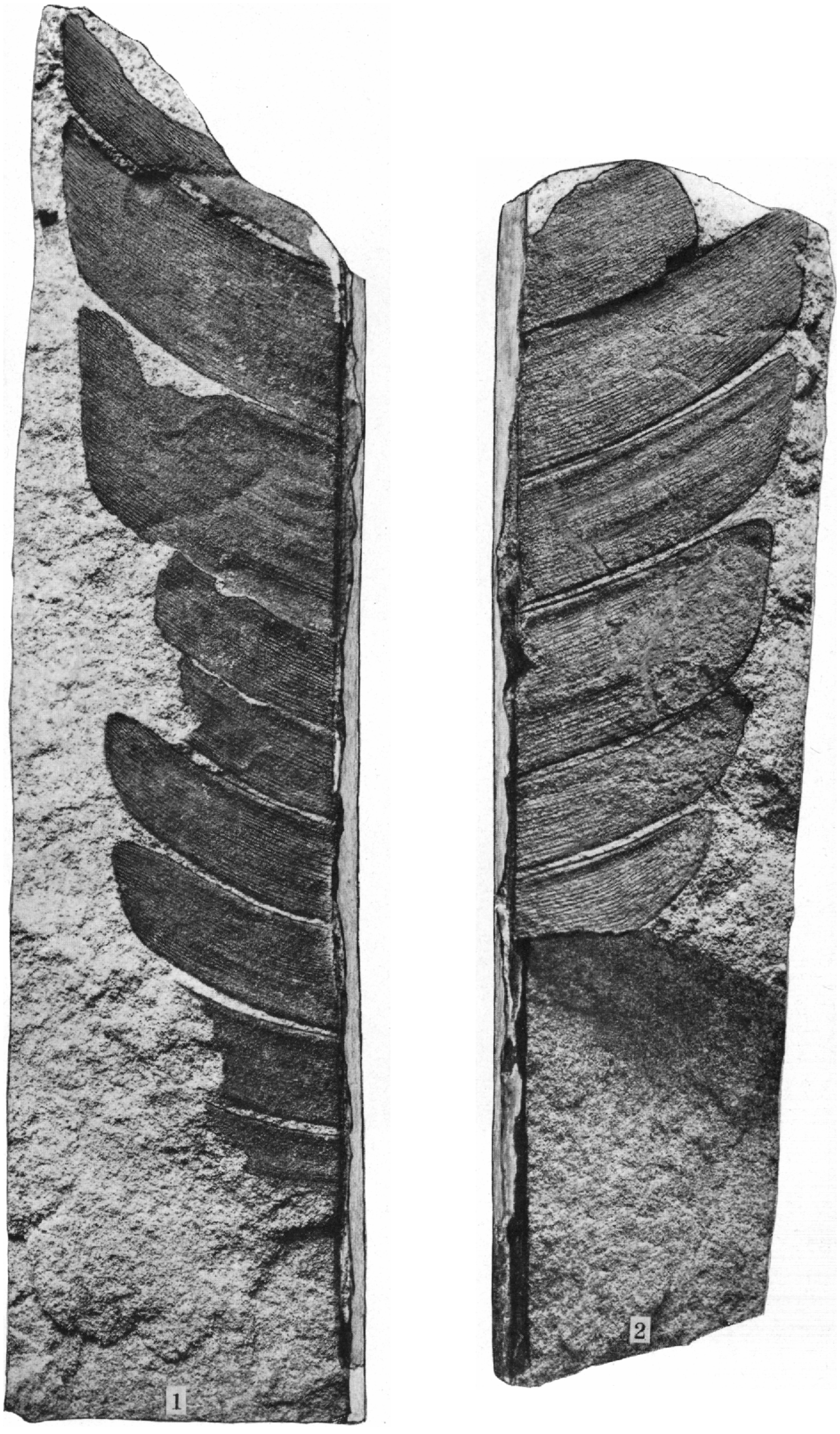
Plate 20

Petiole preserved for a length of 22 centimeters below the lamina. Something less than basal half of lamina preserved, amounting to 18 centimeters. Nine pinnules. Maximum width of lamina 9.5 centimeters. Pinnules falcate, the lower six *Pterophyllum*-like, uniformly 13 to 14 millimeters wide, increasing in length upward. The seventh pinnule is 3 centimeters wide, the eighth 2.5 centimeters, and the ninth similar but partly broken. The basal pinnule is pointed, the tip being close to the distal margin and the proximal margin forming an arc. The second pinnule is much less cut away below, and each succeeding one upward is more truncate, the outer edge of the wider pinnules being almost parallel with the rachis, the outer distal corner being obtusely pointed and the outer proximal corner being rounded off. The rachis is of medium size, about 4 millimeters in diameter as preserved. The pinnules are inserted in the center of its upper face. The veins are stout, immersed in the coriaceous substance, closely spaced, and parallel with one another and with the lateral margins of the pinnules, about 26 in number in the narrower pinnules and similarly spaced in the wider. Species named for Dr. M. G. Mehl, of the University of Missouri, who assisted Brown in the collection and shipment of the material.

This handsome species is much the largest known from so young a horizon. The exact locality is the NW. ¼ sec. 17, T. 6 N., R. 3 W. Wind River meridian. The genus appears in the Triassic, is especially characteristic of the Rhaetic and Oolitic, and may be said to be cosmopolitan in the Jurassic. It was less abundant but still widely distributed in Lower Cretaceous time, being recorded from all the continents except Africa. It became rare in the Upper Cretaceous but is represented by several undoubted species—three in Japan, two on Sakhalin Island, one in the Atane beds of western Greenland, and one in the Cenomanian of Bohemia. These other Upper Cretaceous forms are mostly small and more delicate, the Greenland form being the most robust but prevailingly entire instead of lobate. The form most similar to this Frontier species is *Nilsonia densinervis* (Fontaine) Berry,[4] of the Lower Cretaceous (Patuxent and Arundel formations) of the Atlantic Coastal Plain. This is somewhat narrower and often entire; when lobate the lobes are narrower and somewhat different in outline but essentially similar in venation.

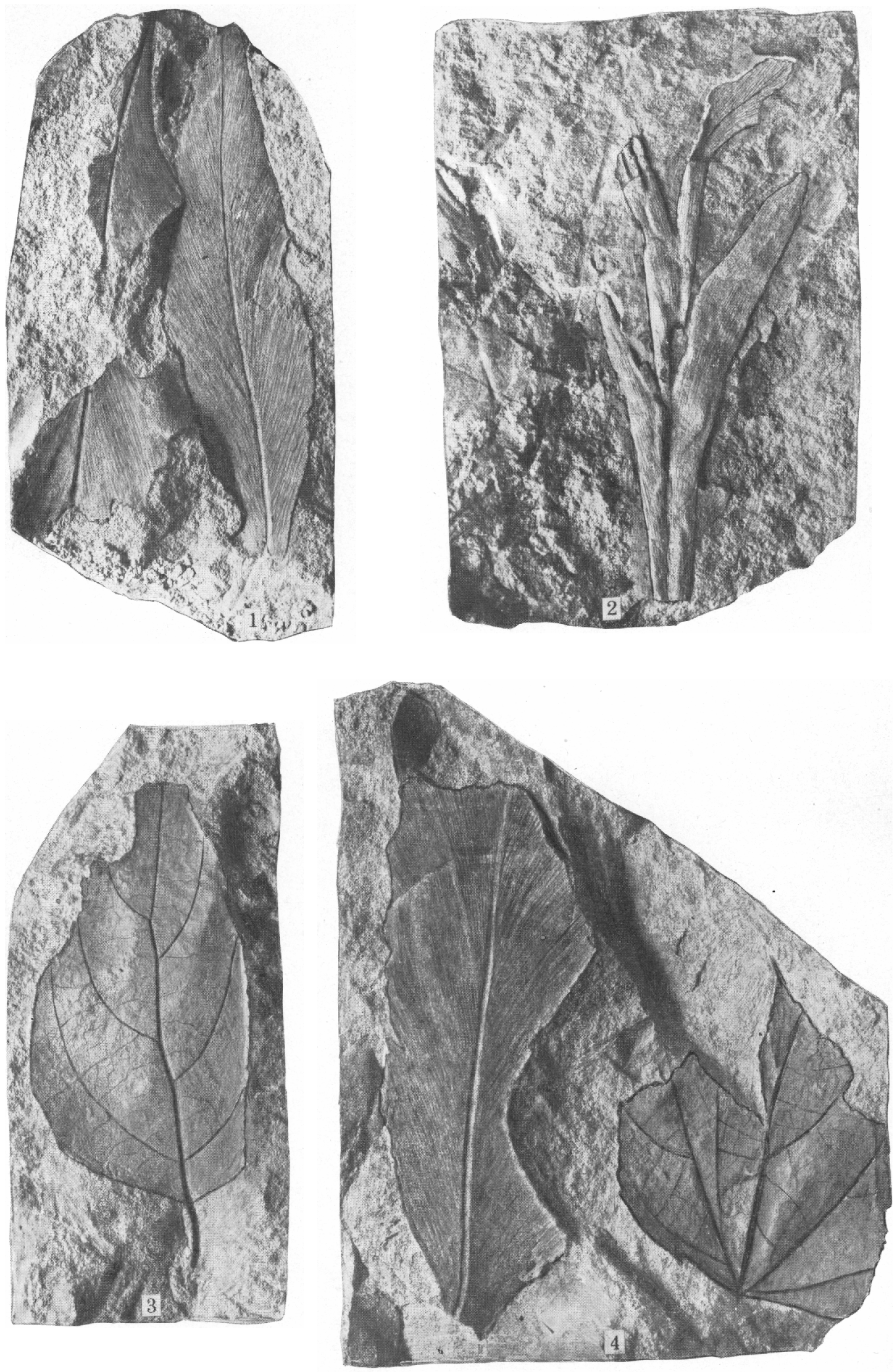
The genus was established by Brongniart in 1825 on material from the Rhaetic of Sweden and has been discussed at length by Saporta, Nathorst, Seward, and others. It may be characterized in the following terms: Frond coriaceous, elongate-lanceolate, entire or commonly more or less deeply pinnatifid by being split, usually to the rachis, into a number of more or less irregular segments which are contiguous, usually broad and truncate. Lamina attached to the upper surface of the rachis, the simple and parallel, equal lateral veins running almost or quite to the median line. In material showing only the under surface of the fronds the stout midrib is prominent and unsegmented specimens are scarcely distinguishable from *Taeniopteris* and allied forms; the segmented varieties approach *Anomozamites* or even some species of *Pterophyllum* in appearance.

[4] Berry, E. W., Maryland Geol. Survey, Lower Cretaceous, p. 362, pls. 57, 58, 1911.



FLORA OF THE FRONTIER FORMATION

Nilsonia mehli Berry, n. sp.



FLORA OF THE FRONTIER FORMATION

1, 2, 4, Protophyllocladus subintegrifolius (Lesquereux) Berry; 3, Ficus inaequalis Lesquereux. An undeterminable leaf of Ficus is shown associated with Figure 4.

Phylum CONIFEROPHYTA

Order PINALES

Family CUPRESSINACEAE

Genus SEQUOIA Endlicher

Sequoia reichenbachi (Geinitz) Heer

Sequoia reichenbachi Knowlton, U. S. Geol. Survey Bull. 257, p. 131, pl. 14, figs. 3–5, 1905.

There are a few rather poorly preserved coniferous twigs in the Frontier sandstones, and these are identical with the rather common coniferous twigs in the Judith River formation which Knowlton referred to this world-wide and protean species. I am not at all certain that they are identical botanically with other occurrences—in fact, the wide range, both in time and space, of *Sequoia reichenbachi* strongly suggests that it is a composite species.

The identity of the present specimens with the Judith River specimens furnishes an element in which the Frontier flora resembles younger floras.

Occurrence: Sec. 8, T. 32 N., R. 99 W., 6 miles south-southeast of Lander.

Family PHYLLOCLADACEAE?

Genus PROTOPHYLLOCLADUS Berry

Protophyllocladus subintegrifolius (Lesquereux) Berry

Plate 21, Figures 1, 2, 4

Protophyllocladus subintegrifolius Berry, Torrey Bot. Club Bull., vol. 30, p. 440, 1903; vol. 31, p. 69, pl. 1, fig. 5, 1904; New Jersey State Geologist Ann. Rept. for 1905, p. 139, 1906; Maryland Geol. Survey, Upper Cretaceous, p. 796, pl. 56, fig. 2, 1916; U. S. Geol. Survey Prof. Paper 112, p. 57, 1919; Johns Hopkins Circ. 199, pp. 89–91, fig. 6, 1907.

Kryshtofovich, Coll. Sci. Imp. Univ. Tokyo Jour., vol. 40, art. 8, p. 41, fig. 6, 1918.

Phyllocladus subintegrifolius Lesquereux, Am. Jour. Sci., vol. 46, p. 92, 1868; The Cretaceous flora, p. 54, pl. 1, fig. 12, 1874; U. S. Geol. Survey Mon. 17, p. 34, pl. 2, figs. 1–3, 1892.

Thinnfeldia lesquereuxiana Heer, Flora fossilis arctica, vol. 6, Abt. 2, p. 37, pl. 44, figs. 9, 10; pl. 46, figs. 11, 12 a, b, 1882.

Hollick, New York Acad. Sci. Trans., vol. 11, p. 99, pl. 3, fig. 6, 1892; New York Acad. Sci. Annals, vol. 11, p. 58, pl. 3, figs. 4, 5, p. 419, pl. 36, fig. 6, 1898; New York Bot. Garden Bull., vol. 2, p. 403, pl. 41, figs. 13, 14, 1902; U. S. Geol. Survey Mon. 50, p. 36, pl. 5, figs. 1–6, 1907.

Newberry, U. S. Geol. Survey Mon. 26, p. 59, pl. 11, figs. 1–17, 1895.

This species was discussed by me at some length in 1903 and 1907 and need not be redescribed in the present connection. It was described originally by Lesquereux, who referred it to the existing genus *Phyllocladus*. Heer subsequently recorded it from the Atane beds of western Greenland and transferred it to the genus *Thinnfeldia* of Ettingshausen. In 1903 I showed that it could not be related to *Thinnfeldia* and proposed the new genus *Protophyllocladus* for its reception. Before and since it has been shown to have a wide range in the earlier half of the Upper Cretaceous, having been recorded from Massachusetts, New York, New Jersey, Maryland, and Alabama in this country and from Russian Sakhalin in eastern Asia. Three additional species have been described—*Protophyllocladus lobatus* Berry, from the Magothy, Black Creek, and Ripley formations in Maryland, South Carolina, and Tennessee, respectively; and *P. lanceolatus* (Knowlton) Berry and *P. polymorphus* (Lesquereux) Berry, from the Eagle and Livingston formations of Montana. The species is exceedingly abundant in the recent collection from the Frontier formation and agrees in every detail with the Dakota sandstone types from Kansas and Nebraska. One specimen shows a possible tendency toward lobation, but it happens to be poorly preserved and indecisive. This habit is well marked in the Raritan material of the species and may therefore be considered to be without special significance. The somewhat younger Coastal Plain species, *P. lobatus*, is uniformly lobate. A tendency toward lobation is also seen in some specimens of *P. polymorphus*.

In view of the great range in size of the specimens from the Frontier formation there does not seem to be a single reliable character for differentiating *P. subintegrifolius* from *P. polymorphus* and *P. lanceolatus*, and I am inclined to think all three represent a single long-lived botanic species. The precise relationship of *Protophyllocladus* has never been settled. After having handled a large amount of material I am of my original opinion that it is undoubtedly coniferous and probably related to the existing genus *Phyllocladus*.

Occurrence: Sec. 8, T. 32 N. R 99 W., 6 miles south-southeast of Lander.

Phylum ANGIOSPERMOPHYTA

Class MONOCOTYLEDONAE

Order ARECALES

Family ARECACEAE

Genus SABALITES Saporta

Sabalites sp.

The recent collections contain very fragmentary specimens that undoubtedly represent the broken rays of a fan palm, which are tentatively referred to the genus *Sabalites*. They were obtained 6 miles south-southeast of Lander, in sec. 8, T. 32 N., R. 99 W.

Class DICOTYLEDONAE

Order FAGALES

Genus DRYOPHYLLUM Debey

Dryophyllum lanceolatum (Knowlton) Berry

Dryandroides lanceolata Knowlton, U. S. Geol. Survey Prof. Paper 108, p. 89, pl. 34, fig. 7, 1917.

This form, which is confined to the type locality in Lincoln County, Wyo., was identified by Knowlton as a species of *Dryandroides*, a genus usually referred to the family Proteaceae and supposed to be related to the genus *Dryandra* of the present Australian region.

Dryandroides has been recorded from a considerable number of regions in both Upper Cretaceous and Tertiary strata, but its botanic relationship has always been a matter of much difference of opinion among botanists. This is happily removed in the case of the present form by reason of the fact that it is not *Dryandroides* but belongs to the fagalean genus *Dryophyllum*. I have had the good fortune to handle a large amount of excellently preserved material of *Dryophyllum*, and this Frontier plant is surely a member of that genus.

Occurrence: Sec. 8, T. 32 N., R. 99 W., 6 miles south-southeast of Lander.

Order URTICALES

Family MORACEAE

Genus FICUS Linné

Ficus inaequalis Lesquereux

Plate 21, Figure 3

Ficus inaequalis Lesquereux, U. S. Geol. Survey Mon. 17, p. 82, pl. 49, figs. 6–9; pl. 50, fig. 3, 1892.
Berry, Torrey Bot. Club Bull., vol. 34, p. 194, pl. 12, figs. 2, 3, 1907; U. S. Geol. Survey Prof. Paper 112, p. 80, pl. 12, fig. 1, 1919.

This species was described by Lesquereux from the Dakota sandstone near Fort Harker, Kans., and was subsequently recorded by me from the Black Creek formation of North Carolina and the Tuscaloosa formation of Alabama.

As the Black Creek formation is considerably younger than either the Dakota or Tuscaloosa there is nothing remarkable in the fact that this species ranges as high as the Frontier formation in the Colorado group. It is not uncommon in the Wind River Basin, but, like most of the dicotyledons found there, it is not especially well preserved.

Occurrence: Sec. 8, T. 32 N., R. 99 W., 6 miles south-southeast of Lander.

Ficus fremonti Knowlton

Ficus fremonti Knowlton, U. S. Geol. Survey Prof. Paper 108, p. 87, pl. 34, figs. 4–6; pl. 35, figs. 4c, 5, 1917.

The type of this species was collected by Fremont in Lincoln County in 1843 and was referred to *Glossopteris* by Hall. It was re-collected by Knowlton and Stanton in 1908 at or near the type locality. It unquestionably represents a dicotyledon, but its reference to the genus *Ficus* is somewhat questionable.

Occurrence: I am able to add two new localities in Fremont County—near Westbrook quarry, southeast of Lander, in sec. 22, T. 33 N., R. 99 W. (collected by T. W. Stanton); 6 miles south-southeast of Lander, in sec. 8, T. 32 N., R. 99 W. (collected by N. H. Brown).

Ficus? sp. Knowlton

Ficus? sp. Knowlton, U. S. Geol. Survey Prof. Paper 108, p. 88, pl. 34, figs, 2. 3, 1917.

This form is based upon very incomplete and questionable material, and there is no conclusive reason for considering it to represent a *Ficus*. The type material came from Lincoln County, and it is included in the present report simply for the purpose of recording the two new localities in Fremont County.

Occurrence: Near Westbrook quarry, southeast of Lander, in sec. 22, T. 33 N., R. 99 W. (collected by T. W. Stanton); 6 miles south-southeast of Lander, in sec. 8, T. 32 N., R. 99 W. (collected by N. H. Brown).

Order MALVALES

Family STERCULIACEAE

Genus STERCULIA Linné

Sterculia towneri (Lesquereux) Berry

Sterculia towneri Berry, U. S. Geol. Survey Prof. Paper 129, p. 217, pl. 57, fig. 1; pl. 60; pl. 61, fig. 1, 1922.
Aralia towneri Lesquereux, U. S. Geol. and Geog. Survey Terr. Bull., vol. 1, p. 394, 1875 [1876]; Ann. Rept. for 1874, p. 349, pl. 4, fig. 1, 1876; Cretaceous and Tertiary floras, p. 62, pl. 6, fig. 4, 1883; U. S. Geol. Survey Mon. 17, p. 132, pl. 23, figs. 3, 4; pl. 31, fig. 1, 1892.
Hollick, New York Acad. Sci. Trans., vol. 16, p. 132, pl. 14, figs. 11, 12, 1897.
Berry, New York Bot. Garden Bull., vol. 3, p. 92, 1903.
Sterculia drakei Cummings, Texas Geol. Survey Third Ann. Rept., p. 210, fig. 8, 1892.
Knowlton, in Hill, Am. Jour. Sci., 4th ser., vol. 1, p. 213, 1895.
Sterculia snowii Lesquereux, U. S. Geol. Survey Mon. 17, p. 183. pl. 30, fig. 5; pl. 31, figs. 2, 3; pl. 32, figs. 1–4, 1892.
Hollick, U. S. Geol. Survey Mon. 50, p. 94, pl. 34, fig. 20, 1907.

Leaves of variable and often very large size, palmately two to seven lobed. The lobes are prevailingly conical and acuminate, occasionally widening somewhat medially and less acutely pointed, separated by generally open and rounded sinuses extending about halfway to the base. The angles that the lobes form with one another and the form of the sinuses vary with the number of lobes, as does also the character of the base, which ranges from truncate to decurrent. The median lobe is generally slightly wider than the others but may be smaller. The normal form is five-lobed. The texture is so coriaceous that these leaves are not rare in coarse sediments like those of the Dakota and Frontier sandstones or the eolian sands of the Cheyenne. The margins are entire. Length 8 to 20 centimeters; maximum width 6 to 24 centimeters. Petiole stout, usually broken away, 12 centimeters or more in length. Midvein stout, prominent; lateral primaries stout, basal or subbasal; secondaries thin, regularly spaced, camptodrome, usually more or less immersed in the leaf substance.

This is an exceedingly well marked species and, like most Sterculias, both ancient and modern, shows the characteristic variability of the genus. It was described originally from material collected in the Dakota sandstone of Kansas and occurs in the Tucumcari Mountains of New Mexico in beds referred to the Dakota. It is recorded from the Magothy formation of New Jersey, from rocks of the same age on Marthas Vineyard, and from the Cheyenne sandstone of Kansas. Only fragments have been found in the Frontier, but they are sufficiently characteristic to establish this species as a member of the Frontier flora.

Occurrence: Sec. 8, T. 32 N., R. 99 W., 6 miles south-southeast of Lander.

POSITION UNCERTAIN

Phyllites cretaceus (Ettingshausen) Berry

Artocarpidium cretaceum Ettingshausen, Kreideflora von Niederschoena: K. Akad. Wiss. Wien Sitzungsber., Band 55, p. 251, pl. 2, fig. 4, 1867.

Lesquereux, U. S. Geol. Survey Mon. 17, p. 86, pl. 50, fig. 7, 1892.

Although it is highly improbable that the leaf from the Dakota sandstone of Kansas is identical with the single fragment from the Cenomanian of Austria to which Ettingshausen gave this name, neither shows any features warranting reference to the Moraceae or suggesting any relationship with *Artocarpus*. The type material did not warrant any identification and should not have been described, and it may well be transferred to the noncommital genus *Phyllites* along with the leaf from the Dakota sandstone. The latter is complete, but I am unable to suggest its probable botanic affinity.

This is of interest chiefly as constituting another older element in the flora of the Frontier formation.

Occurrence: Sec. 8, T. 32 N., R. 99 W., 6 miles south-southeast of Lander.

Phyllites grandifolius-cretaceus (Lesquereux) Berry

Smilax grandifolia-cretacea Lesquereux, U. S. Geol. Survey Mon. 17, p. 40, pl. 46, fig. 3, 1892.

Much as I dislike the pseudogeneric term *Phyllites*, I can see no basis for referring this leaf to the genus *Smilax*, and there are a variety of unrelated existing genera that have leaves of this type. The species was based upon a single imperfect specimen from the Dakota sandstone of Kansas, which apparently was compared by its describer with pictures of fossil European forms in the works of Unger, Ettingshausen, and Heer and not with existing leaves.

A fragmentary specimen from the Frontier is exactly like Lesquereux's type except for being somewhat larger. If a considerable amount of material were available it would be worth the trouble of an extended search among existing forms to test the possibility of arriving at the true botanic affinity. In the absence of such material it is more scientific to face the fact that such specimens are undeterminable botanically.

Occurrence: Flank of Table Mountain, in the SE. ¼ sec. 8, T. 32 N., R. 99 W.

Phyllites crassipes (Heer) Berry

Juglans crassipes Heer, Schweizerische Gesell. Neue Denkschr., Band 23, p. 23, pl. 6, fig. 3, 1869; Flora fossilis arctica, Band 7, p. 27, pl. 61, fig. 4; pl. 65, fig. 9, 1883.

Lesquereux, U. S. Geol. Survey Mon. 17, p. 69, pl. 49, figs. 1–3, 1892.

Hollick, U. S. Geol. Survey Mon. 50, p. 55, pl. 9, figs. 3–5, 1907.

This species was described from the Cenomanian of Moravia and was subsequently recorded by its describer from the much younger Patoot beds of western Greenland. It was identified by Lesquereux in the Dakota sandstone of Kansas and by Hollick in the Raritan and Magothy formations of the Atlantic Coastal Plain.

It represents one of those fossil foliar types which are found to be common in unrelated existing genera and which entirely lack diagnostic generic features. There is not a shred of evidence that it is related to *Juglans*, nor is there any reason for supposing that the European, Greenland, and American specimens represent the same botanic species. On the other hand, the considerable geographic and geologic range is emphatically opposed to such a conclusion.

However, the recorded occurrences represent indistinguishable remains with which the material from the Frontier formation is identical. Beyond the evidence which it affords of a new type of plant in the Frontier flora I regard it as entirely worthless either botanically or geologically, and I much doubt if a clue to its relationship can ever be obtained.

Occurrence: Flank of Table Mountain, in the SE. ¼ sec. 8, T. 32 N., R. 99 W.

BORATE MINERALS FROM THE KRAMER DISTRICT, MOHAVE DESERT, CALIFORNIA

By WALDEMAR T. SCHALLER

SUMMARY

The development of the borate deposits of the Kramer district, in the Mohave Desert, Calif., received a strong impulse in 1926 by the discovery there of a large deposit of the new sodium borate, kernite. This deposit lies about 2½ miles east of the original discovery of colemanite and ulexite in the Suckow shaft. All the borates occur underground, there being no indication whatever at the surface of the existence of any borates below. The original discovery was made in drilling for water.

The first sample of these borate minerals was received by the writer from Hoyt S. Gale, of Los Angeles, in the fall of 1926. In September, 1927, through the courtesy of Mr. Clarence M. Rasor, of the Pacific Coast Borax Co., an opportunity was had to visit the kernite deposit and collect material, which forms the basis of this report. The kindness of Mr. Rasor in allowing the collection and study of material is much appreciated.

The geologic setting of the kernite deposit is imperfectly known, but in general it is similar to that of the deposits a few miles west which have been described by Noble. The borate minerals, several hundred feet underground, occur in a complex clay series and are underlain by igneous rock, whether extrusive (and therefore of earlier age than the borates) or intrusive (and therefore of later age than the borates) is not known. Knowledge of this relative age of the lava would be of great help in understanding the genesis of the borate minerals.

The borate minerals found include colemanite, ulexite, borax, tincalconite, and two new borates, kernite and kramerite. The associated minerals are calcite, realgar, stibnite, and the clay minerals. Kernite was briefly described in a preliminary paper in January, 1927; kramerite is first described in this paper. Abundant and better material, recently collected by the writer, has furnished the basis for fuller description of kernite and also shows that some of the statements made in the original brief paper need to be corrected.

The composition of the two new minerals is as follows:

Kernite, $Na_2O.2B_2O_3.4H_2O$.

Kramerite, $Na_2O.2CaO.5B_2O_3.10H_2O$.

Kernite occurs in greater quantity and is of greater interest scientifically. It is the commercial mineral mined, existing as an immense deposit a hundred feet thick and many hundred feet in lateral extent. The deposit contains an estimated minimum of well over a million tons. Some of the crystals are of immense size, the largest one seen measuring 8 feet in thickness. Many of the crystals are several feet across. The mineral belongs chemically to the borax group but has only 4 molecules of water instead of 10, as in borax. The crystals are monoclinic, with the axial ratio $a : b : c = 1.5230 : 1 : 1.6989$, $\beta = 71° 08'$. Thirty-six crystal forms were observed, together with many additional vicinal forms. Two perfect cleavages break the mineral up into elongated rhombic prisms.

Kramerite occurs in small quantities as radiating prisms, averaging not over a millimeter in thickness. It is monoclinic, with the axial ratio $a : b : c = 1.1051 : 1 : 0.5237$, $\beta = 72° 16'$. Eight crystal forms were identified. In composition, kramerite is like ulexite but with less water. It was probably artificially made by Van't Hoff. The prisms of kramerite cut the clay, kernite, and borax and are of later origin.

Tincalconite, $Na_2O.2B_2O_3.5H_2O$, named by Shephard in 1878, is, as found, always secondary, forming by dehydration of borax and by hydration of kernite. It is a finely crystalline aggregate, existing only as a coating on the other borates.

These three sodium borates have been made artificially in the chemical laboratory of the Geological Survey in Washington. A comparison of the properties of a number of hydrates in the different series—the borax series, the colemanite series—has been made.

The origin of the deposit forms a separate problem in which the mineral relationships are the chief criteria. The geologic relations are so little known in detail that the processes of formation can only be suggested at this time. The kernite was perhaps formed by the fusion in its own water of crystallization of a previous accumulation of borax crystals, enough water being driven off to yield on solidification of the fused borax the 4-hydrate kernite. If no water could escape the resultant product was massive borax.

A list of all known boron minerals is added, with some critical remarks about their relationship and standing. A bibliography of the borate minerals of the Kramer district ends the paper.

LOCALITY

The Kramer borate district lies in southeastern Kern County, Calif., only a few miles west of the San Bernardino County line. The deposits of borate minerals are about 7 miles northwest of Kramer and 28 miles due east of Mojave. The first borate discovery was made in sec. 22, T. 11 N., R. 8 W., where, in 1913, colemanite and ulexite were found. The kernite deposit, first actively opened in 1926, lies on the boundary between sec. 24, T. 11 N., R. 8 W., and sec. 19, T. 11 N., R. 7 W. The deposits are only a few miles north of the main line of the Atchison, Topeka & Santa Fe Railway.

GEOLOGIC SETTING

The deposit of colemanite and ulexite in the western part of the Kramer borate district has been described by Noble [1] and Gale.[2]

The borate deposits lie in tilted and faulted beds of clay shale at least 100 feet beneath one of the gently sloping alluvial plains that form so characteristic a

[1] Noble, L. F., Borate deposits in the Kramer district, Kern County, Calif.: U. S. Geol. Survey Bull. 785, pp. 45-61, 1926.

[2] Gale, H. S., Borate deposits near Kramer, Calif.: Am. Inst. Min. and Met. Eng. Trans., vol. 73, pp. 449-463, 1926.

feature of the desert region. This alluvial plain of sand and gravel is bordered on the northeast by low hills of lava and tuff. A bed of basaltic lava underlies the borate deposits. The clay shale and the lava are of Tertiary age.

According to Noble, a bed of arkose sandstone overlies the borate-bearing shale in the Suckow shaft No. 2, 2½ miles west of the kernite deposit, where colemanite and ulexite are the only borate minerals present. The clay material containing lenses and nodules of colemanite, irregularly distributed, differs from the clay deposited upon the surface of Rogers Lake and other playas in the desert at the present time only in that it has been consolidated into shale, tilted, eroded, and buried. The clay matrix of the kernite deposit is more complex and is composed of a diverse set of materials. Noble considers the deposit in the region of the Suckow shaft to have been originally a playa deposit, the borate mineral ulexite forming in the mud of the drying lake and the colemanite being formed later by alteration of the ulexite, as postulated by Foshag.[3]

The bed of basaltic lava underlying the borate deposits is believed by Noble to have been poured out just before the shale was deposited and to be, indirectly, the source of the boron, which was derived from the hot springs and solfataras connected with the volcanic activity. A bed of igneous rock is likewise reported to underlie the kernite deposit.

According to the records, although the igneous rock beneath the borates was found wherever the borings were deep enough, the borate minerals were not encountered everywhere. They are extremely irregular in form and distribution, the structure of their Tertiary host rocks is exceedingly complex, and "the limits of the deposits can be determined only by underground exploration."[4] The report by Noble was written before the kernite deposit was discovered. The relations of the clay shale and the igneous rock to the kernite deposit are discussed under the heading "Origin of deposit."

The borate minerals known to occur in the Kramer district are as follows: Colemanite and ulexite, probably the most widespread, have been found wherever any borates occur. They are the minerals reported from the original discovery. Apparently they are the only borates found in any of the shafts and borings in sec. 22. At the kernite deposit, colemanite and ulexite occur in the clay above the kernite. Associated with the kernite are borax, tincalconite, and kramerite. Other minerals associated with these borates are the clay minerals, the rock minerals (feldspars, micas, etc.) found in the clay, calcite, stibnite, and realgar.

[3] Foshag, W. F., The origin of the colemanite deposits of California: Econ. Geology, vol. 16, p. 210, 1921.

[4] Noble, L. F., op. cit., p. 51.

BORATE MINERALS

COLEMANITE

The colemanite from the Kramer district yields good examples illustrating its derivation from ulexite, especially as small spherical masses of radiating crystals are embedded in ulexite and evidently derived therefrom, as indicated by Foshag. A specimen obtained above the kernite deposit, kindly presented by Mr. Rasor for study, shows the development of a group of compacted crystals of colemanite, as much as an inch thick, growing in the ulexite. This specimen is shown in Plate 22, *B*.

A number of specimens collected from one of the dumps in sec. 22 show clearly the nodular colemanite formed from ulexite, with a very small quantity of the ulexite still remaining. The centers of the nodules consist of compact colemanite in which are scattered bands of residual ulexite fibers. Around the massive center radiating bunches of colemanite fibers and plates have grown. Both the fibrous and the massive variety have the optical properties of colemanite, as the following tabulation shows:

Optical properties of colemanite

	As given by Larsen	Massive variety	Fibrous variety
α	1. 586	1. 586	1. 585
β	1. 592	1. 591	1. 590
γ	1. 614	1. 613	1. 614
Sign	+	+	+
$Z \wedge c$	6°	Very small.	5°
2E	97°	Large.	Large.

Most of the fibers have positive, though a few have negative elongation.

Colemanite does not seem to be very abundant in the eastern part of the field, where the kernite occurs. Only a few specimens, all associated with ulexite, were seen in the clay above the kernite.

ULEXITE

Ulexite occurs abundantly as compact fibrous veins in the clay shale. The familiar "cotton-ball" variety was not observed. Noble describes the ulexite found in sec. 22 as "compact." That occurring above the kernite is in beautiful veins of pure white material, affording a fine example of "satin spar." These veins may be several inches in thickness. Such an example, with crystals of colemanite developed in the ulexite, is shown in Plate 22, *B*. A set of smaller veins, with a beautiful "satin spar" luster, is shown in Plate 22, *A*. The relative position of these two specimens is not known, but they probably came from slightly different horizons, as the clay matrix shows considerable difference. But both are from the clay above the kernite.

A similar compact ulexite was reported from Lang, 55 miles to the southwest, by Foshag,[5] who states:

[5] Foshag, W. F., op. cit., p. 204.

It is not the ordinary "cotton-ball" type but is massive and fibrous. It has the appearance of cotton balls that have been consolidated by pressure. The ulexite occurs in irregular masses, more or less lenslike and surrounded by thin layers of clay. In structure these lenses are compact-fibrous, the fibers oriented in all directions in the centers and parallel at the peripheries.

Similar masses of ulexite are to be found in the Kramer district, but veins which have a straight parallel fiber structure, such as is shown in Plate 22, *A*, and which occur above the kernite probably did not form by the consolidation of "cotton balls" without going into solution and being redeposited. A curious growth of masses of ulexite, resembling in form a cone-shaped fungus, was also encountered. The shaft through the clay containing the ulexite, above the kernite, was boarded up at the time of examination, so that the relation of the ulexite-bearing clay to the kernite-bearing clay is unknown.

Examined in crushed fragments, the ulexite is seen to be intimately polysynthetically twinned, the traces of the twinning plane being parallel with the elongation of the fibers, which is generally negative, though a few fibers with a positive elongation were noticed. Y is approximately parallel with the elongation. The optical properties are the same as those of normal ("cotton-ball") ulexite.

Optical properties of ulexite

	As given by Larsen	Satiny ulexite
α	1. 491	1. 493
β	1. 504	1. 506
γ	1. 520	1. 519
Sign	+	+
$Y \wedge c$	0° to 23°	Variable, about 20°
2V	Moderate to large.	Large.

The density of this fibrous ulexite was determined to be 1.963.

The fine satiny ulexite shown in Plate 22, *A*, has the normal chemical composition, as shown by the following analyses:

Analyses of satin spar variety of compact ulexite.

	1	2	3
CaO	14. 06	14. 14	13. 85
Na_2O	7. 09	[a] 7. 06	7. 65
B_2O_3	42. 94	43. 12	42. 95
H_2O	35. 54	35. 68	35. 55
Insoluble	. 10	--------	--------
	99. 73	100. 00	100. 00

[a] By difference.

1. Analyst, W. T. Schaller; Kramer district, Calif.
2. Analyst, W. F. Foshag; Lang, Calif.
3. Calculated.

The analyses represent the figures for the established formula of ulexite, $Na_2O.2CaO.5B_2O_3.16H_2O$. By heating this ulexite, in a powdered form, in a sealed tube with a saturated solution of sodium chloride in the steam bath for a week, it was readily changed to the lower hydrate, kramerite, as described under the synthesis of that mineral.

KRAMERITE

CHARACTER OF MATERIAL

Groups of spherulites were noted as rather abundant near the floor of one of the drifts in the kernite mine. They proved to be a new mineral, a hydrous sodium-calcium borate, $Na_2O.2CaO.5B_2O_3.10H_2O$, of the same type of formula as ulexite ($Na_2O.2CaO.5B_2O_3.16H_2O$) but with less water. The name kramerite is given to this mineral, after the name of the district, which evidently was so named after Kramer, the nearest town.

Kramerite is in prismatic crystals, usually forming spherulites but also more rarely existing as single crystals or groups of prisms in more or less parallel position, or in radiating groups that do not form a complete spherulite. The diameter of the spherulites averages from 3 to 4 centimeters. The center is usually composed only of kramerite, but there is no sharp outer boundary, the prismatic crystals, in places considerably bent, projecting into the matrix for unequal distances. The longest individual crystal measured extended for 23 millimeters (about an inch), although still longer ones probably occur. The crystals are approximately a quarter to half a millimeter in average thickness.

The crystals penetrate indifferently both the clay matrix and the associated massive borax and highly cleavable kernite and seem to be the next to the last borate mineral of the deposit to form, tincalconite being the last. Although the striking spherulites were noticed at only one place in the mine, extending over a distance of but a few feet, crystals of kramerite probably are widely distributed throughout the deposit, for a few individual prisms or small radiating groups were noticed on many of the other specimens collected. The typical appearance is shown in Plate 23. The minute black spherulites of stibnite, a little calcite, and the clay are the only boron-free mineral associates noted. In the laboratory the borax of the specimens readily dehydrates to tincalconite in which the bright prisms of kramerite stand out, this mineral retaining its glassy luster and not becoming dull and white. On the specimens where the borax has become partly dehydrated, the center portion of the kramerite spherulites appears considerably darker against the dull-white background.

PHYSICAL AND OPTICAL PROPERTIES

Kramerite is colorless and transparent. It has a perfect prismatic cleavage, parallel to *m* (110), a hardness of about 2½, and a density of 2.141. The luster is vitreous, and the crystals remain bright and shining.

The mineral is monoclinic prismatic. The optic axial plane is parallel to the clinopinacoid b (010), Y $=b$ axis, Z $\wedge$ c axis $=12°$. Crystals lying on a (100), with parallel extinction, show part of one optic eye of the obtuse figure. Parallel to m (110), the extinction is about 14°. The sign is positive, and the axial angle is large. 2V (measured) 73°, 2E 126°. Dispersion $\rho > v$. The elongation of the crystals is positive.

The optical orientation of kramerite is shown in Figure 19. The refractive indices are $\alpha=1.515$, $\beta=1.525$, $\gamma=1.544$. C. S. Ross obtained α 1.514, $\beta=1.524$, $\gamma=1.543$. The birefringence is 0.029, and crushed fragments show brilliant colors. Kramerite is commonly not twinned, and the absence of twinning lamellae, so characteristic of ulexite, serves as an immediate criterion for its differentiation. An examination of crushed fragments, however, occasionally shows what might be simple twinning.

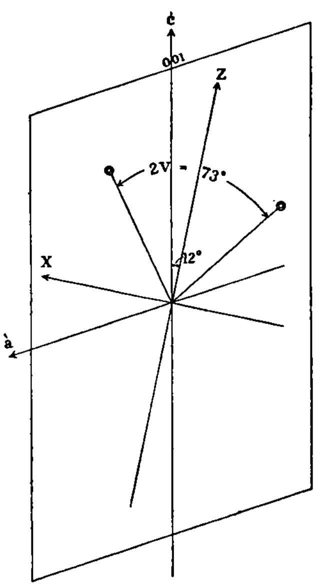


FIGURE 19.—Optical orientation of kramerite

CRYSTALLOGRAPHY

GENERAL DESCRIPTION

The prismatic monoclinic crystals are simple in their combinations and show all the unit forms except the base. Fourteen crystals were measured, but many more were examined under the microscope. No twins were seen. The crystals are all long prismatic in habit.

The forms present are

a (100)	m (110)	t (101)	p (111)
b (010)	e (011)	d ($\bar{1}01$)	o ($\bar{1}11$)

CALCULATION OF ELEMENTS

The axial ratios are $a : b : c = 1.1051 : 1 : 0.5237$, $\beta=72°\ 16'$, $p'_o=0.4976$, $q'_o=c=0.5237$, $e'=0.3197$. These elements were derived from the average measurements (ϕ and ρ) of m (110) and e (011). For m (110), the ϕ readings averaged 43° 32′, as follows:

Average of 24 most consistent readings, ranging from 43° 26′ to 43° 41′	43° 33′
Average value, on four crystals, where all four faces of m were measured accurately	43° 31½
Average for crystal 11 (readings 43° 31′, 43° 32′, 43° 31′, 43° 32′, all good)	43° 31½
Average ϕ	43° 32′

The total readings on m (110), exclusive of those due to the vicinal faces and those where the striations permitted only an approximate centering of the reflections, for 38 readings, ranging from 43° 03′ to 43° 56′, averaged 43° 29′.

The average of the 12 best measurements of good and fair reflections of the clinodome e (011) gave:

ϕ (from 31° 01′ to 31° 51′), 31° 24′
ρ (from 31° 26′ to 31° 36′), 31° 32′

From these average values—namely, ϕ (110) $=43°\ 32'$, ϕ (011) $=31°\ 24'$, and ρ (011) $=31°\ 32'$—the axial elements as given are calculated.

FORMS AND ANGLES

The measured and calculated angles for all the forms are as follows:

Measured and calculated angles of kramerite

Form	Measured		Calculated	
	ϕ	ρ	ϕ	ρ
	° ′	° ′	° ′	° ′
a (100)	90 00	90 00	90 00	90 00
b (010)	0 03	90 00	0 00	90 00
m (110)	43 32	90 00	43 32	90 00
t (101)	90 02	39 13	90 00	39 16
d ($\bar{1}01$)	89 56	9 53	90 00	10 05
e (011)	31 24	31 32	31 24	31 32
p (111)	57 12	44 08	57 21	44 09
o ($\bar{1}11$)	20 00	30 06	18 45	28 57

The individual measurements of the different forms are as follows, those measurements which, owing chiefly to striations, served only to identify the forms being omitted.

Measurements of a (100), kramerite

Crystal No.	Size	Reflection	Measured ϕ (calculated, 90° 00′)
			° ′
1	Medium	Fair	89 53
1	do	Poor	91 16
2	do	Good	89 56
2	Small	Poor	90 08
3	do	do	----
4	Large	do	88 50
4	do	do	89 29
5	Medium	Good	90 52
5	do	do	89 09
6	Large	do	90 00
6	do	Poor	----
7	Medium	do	90 11
8	Large	do	90 31
8	Medium	do	90 02
9	do	Fair	89 59
10	Large	Good	90 02
10	do	do	90 12
11	Medium	do	90 12
12	Large	Fair	89 08
12	Medium	Poor	90 27
13	Large	Fair	90 00
14	Small	Good	89 48

Measurements of b (010), kramerite

Crystal No.	Size	Reflection	Measured φ (calculated, 0° 00′) ° ′
1	Line face	Poor	[a] 0 07
1	do	do	
2	do	do	03
3	do	do	12
4	do	do	11
4	do	do	01
5	do	do	
5	Small	Fair	19
6	Line face	Poor	
6	do	do	49
7	do	do	
8	do	do	15
9	Small	Good	02
9	do	Poor	10
10	Line face	do	20
10	Small	do	01
12	do	do	02
13	do	do	
14	do	Fair	10
14	Line face	Poor	

[a] The positive or negative character of the deviation in angle from 0° 00′ is not indicated.

Measurements of m (110) and accompanying or replacing vicinal form, kramerite

Crystal No.	Size	Reflection	Measured φ (calculated, 43° 32′) ° ′	Vicinal form ° ′
1	Medium	Good	43 26	
1	do	do	43 37	38 17
1	Line face	Poor	43 11	
2	Medium	Fair	43 29	
2	Line face	Poor		
2	Medium	Good	43 56	
2	do	Poor	43 46	
3	do	Fair	43 37	
3	Small	Poor	43 16	
3	Medium	do	43 56	
3	Small	do		
4	Medium	do	42 44	40 18
4	do	do	41 44	41 06
4	do	do	43 05	40 32
4	do	Fair	43 32	
5	do	Good	43 12	
5	Small	Fair	43 39	
5	Medium	do	42 23	
5	Small	Poor		38 43
6	Medium	Good	43 31	
6	do	Fair	43 19	
6	do	Good	43 33	
6	Small	Poor	43 38	
7	Medium	do		
8	do	do		41 48
8	do	do		41 16
8	do	Good	43 32	
8	Small	Poor	43 12	
9	do	do		38 37
9	do	do		41 14
9	do	do	43 33	
9	Medium	Good	43 18	
10	Line face	Poor		39 23
10	do	do		39 13
10	Small	Fair	43 29	
10	do	Poor	43 26	
11	Medium	Fair	43 32	41 38
11	do	Good	43 31	
11	do	do	43 32	
11	do	do	43 31	
12	do	Fair	43 41	
12	do	Poor	43 37	
12	do	Fair	43 29	
12	do	do	43 33	
13	do	do	43 35	40 58

Measurements of m (110) and accompanying or replacing vicinal form, kramerite—Continued

Crystal No.	Size	Reflection	Measured φ (calculated, 43° 32′) ° ′	Vicinal form ° ′
13	Medium	Fair	43 32	
13	do	do	43 09	
13	do	do	43 36	39 38
14	Large	do	43 03	
14	Medium	Good	43 13	41 05
14	do	Fair	43 45	

Accompanying the reflection for *m* (110), especially if the face was striated and thereby somewhat rounded, there was, on many crystals, a distinct second signal, close to the one from *m* (110), which gave inconsistent readings ranging from 38° 17′ to 41° 48′. For eight of the fifteen occurrences of the vicinal prism, it accompanied the face of *m* (110); for the remaining seven occurrences it replaced *m* (110). On crystals Nos. 8, 9, and 10, where on each one there were two faces of *m* (110) and two faces of the vicinal form, each pair of faces was grouped around *b* (010). The vicinal form was always striated; if the prism face was not striated and gave a single distinct reflection its measurement was that of *m* (110).

Measurements of e (011), kramerite

Crystal No.	Size	Reflection	Measured φ (calculated, 31° 24′) ° ′	Measured ρ (calculated, 31° 32′) ° ′
1	Medium	Poor	32 08	31 45
1	Small	do	31 51	31 28
2	do	do	30 48	31 23
4	Large	Fair	31 49	31 36
4	do	do	31 43	31 36
5	do	do	31 51	31 29
5	Medium	do	31 10	31 34
6	Large	Good	31 19	31 30
6	do	do	31 34	31 32
7	Small	Fair	31 15	31 32
8	Large	do	31 30	31 26
8	do	do	31 25	31 26
9	do	Good	31 01	31 34
9	Small	Fair	31 01	31 34
10	Medium	Poor	31 50	31 36
10	Small	do	32 21	31 36
11	Large	Fair	31 12	31 36

Measurements of t (101), kramerite

Crystal No.	Size	Reflection	Measured φ (calculated, 90° 00′) ° ′	Measured ρ (calculated, 39° 16′) ° ′
1	Line face	Poor	89 47	39 00
2	Small	do	90 12	38 51
4	Large	Fair	90 19	39 16
5	Medium	Poor	89 09	39 30
7	Line face	None[a]	90 11	40 15
8	Large	Fair	90 14	39 07
9	do	do	90 00	39 19
11	Small	Poor	90 11	39 13
12	Line face	None		
13	Small	Poor	90 13	39 29

[a] Where no reflection of the signal was obtained, the measurements were made by maximum brightness.

Measurements of d ($\bar{1}01$), kramerite

Crystal No.	Size	Reflection	Measured ϕ (calculated, 90° 00′)	Measured ρ (calculated, 10° 05′)
			° ′	° ′
3	Small	Poor	89 18	10 04
5	do	do	92 20	9 14
6	Minute	None	90 01	11 20
7	do	do	90 11	9 22
11	do	do	89 55	10 04
13	do	do	90 13	9 11

Measurements of o ($\bar{1}11$), kramerite

Crystal No.	Size	Reflection	Measured ϕ (calculated, 18° 45′)	Measured ρ (calculated, 28° 57′)
			° ′	° ′
2	Line face	Poor	21 26	28 44
3	Minute	do	19 12	31 12
3	do	do	18 25	27 47
11	do	do	21 55	31 35
14	do	do	19 03	31 12

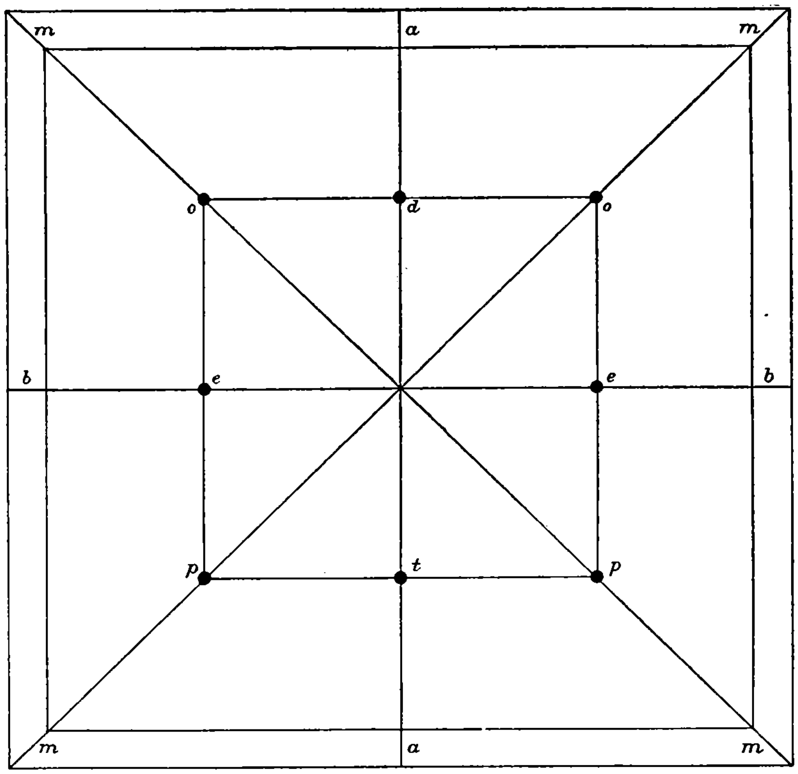


FIGURE 20.—Gnomonic projection of kramerite forms

Measurements of p (111), kramerite

Crystal No.	Size	Reflection	Measured ϕ (calculated, 57° 21′)	Measured ρ (calculated, 44° 09′)
			° ′	° ′
1	Medium	Poor	56 30	44 31
2	Small	do	56 46	43 54
2	Minute	do	57 28	43 42
3	Large	Fair	57 12	44 06
3	Minute	Poor	57 47	44 06
5	Large	Fair	57 35	44 11
5	Minute	Poor	56 42	44 29
7	Small	do	57 39	44 09
7	do	do	56 58	44 06
10	Medium	do	56 24	44 37
11	Large	Fair	57 20	43 56
11	do	Poor	57 36	44 08
12	do	Fair	57 26	44 18
12	do	Poor	57 37	43 55
13	Medium	do	56 35	44 10
14	Large	Fair	57 35	43 57

The gnomonic projection of the crystal forms of kramerite is shown in Figure 20. The projection has a striking isometric symmetry; this pseudoisometric symmetry is brought out still more clearly if in the axial ratios, as given, the *c* axis is doubled, or still better if the values for p_o' and q_o' are doubled. The pseudoisometric ratio of the crystallographic axes may then be written:

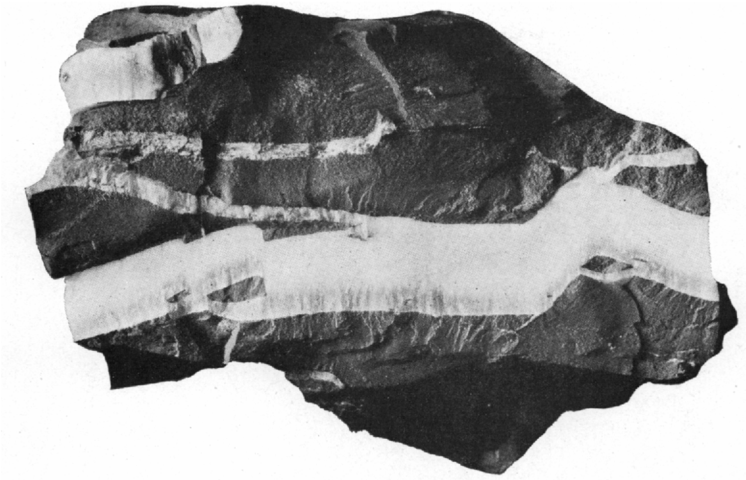
$$a : b : 2c = 1.1051 : 1 : 1.0474$$
$$2p_o' = 0.9052,\ 2q_o' = 1.0474$$

DESCRIPTION OF FORMS

The orthopinacoid *a* (100) is the largest form on the crystals of kramerite. On most crystals it is larger than the prism faces, though on a few it is of about the same size. Only rarely is it smaller than the prisms, occasionally being only a broad line face. On crystal 14, Figure 23, only one narrow face of *a* (100) is present. It is generally striated vertically and may be sufficiently rounded to give a row of reflections many degrees apart. The two faces of the form are not of equal size on all crystals. It is, however, the universal and dominant form on nearly all the crystals. When it and the prism faces are nearly of the same size, the crystal has a six-sided cross section (figs. 21, 22, 25, 26); when the faces are large, the crystal is flattened parallel to it (fig. 24).

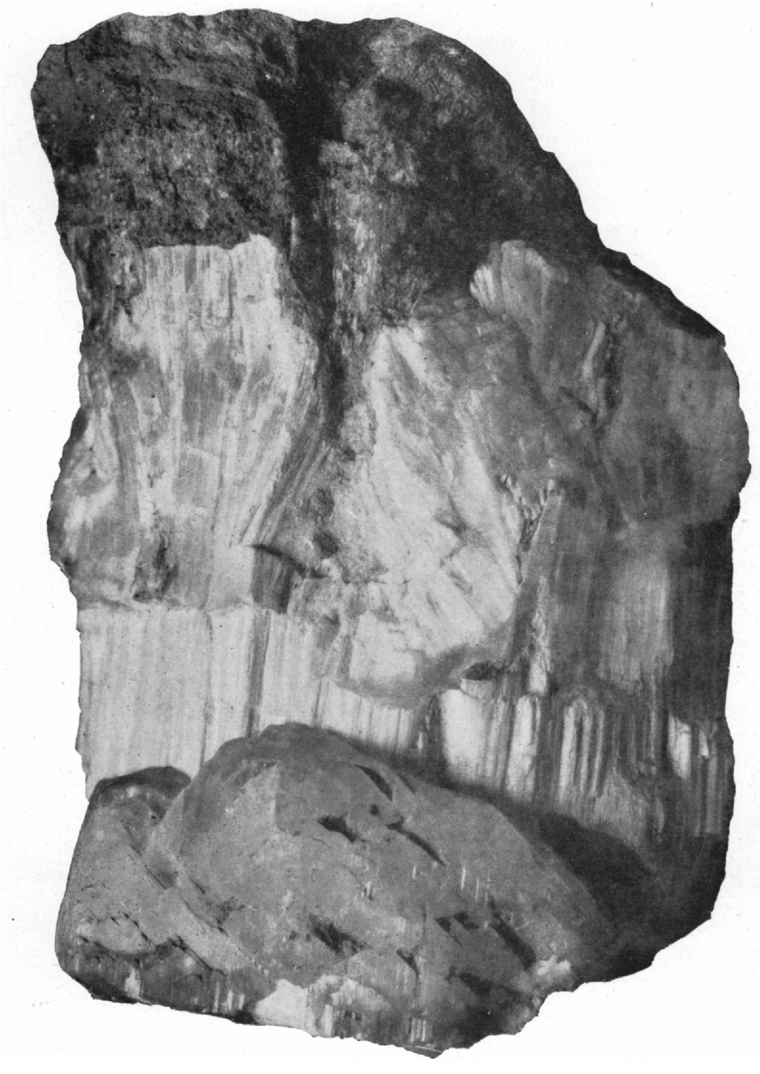
The clinopinacoid *b* (010) is a very narrow face, only rarely becoming nearly as large as the prisms. It is striated vertically and, if broader than a line face, is rounded over a considerable angle. A few crystals suggest that an imperfect cleavage may exist parallel to this form. It is present on all but one (No. 11) of the crystals measured. Reexamination of this crystal failed to show any face of this form.

The prism *m* (110) is universally present and is the second largest form. With the orthopinacoid it determines the habit of the crystals. Many of the faces give good single reflections; on some crystals the unit prism is accompanied by or replaced by a vicinal form, but a few degrees off, lying between it and *b* (010).



A. SEAMS OF ULEXITE IN CLAY ABOVE KERNITE DEPOSIT

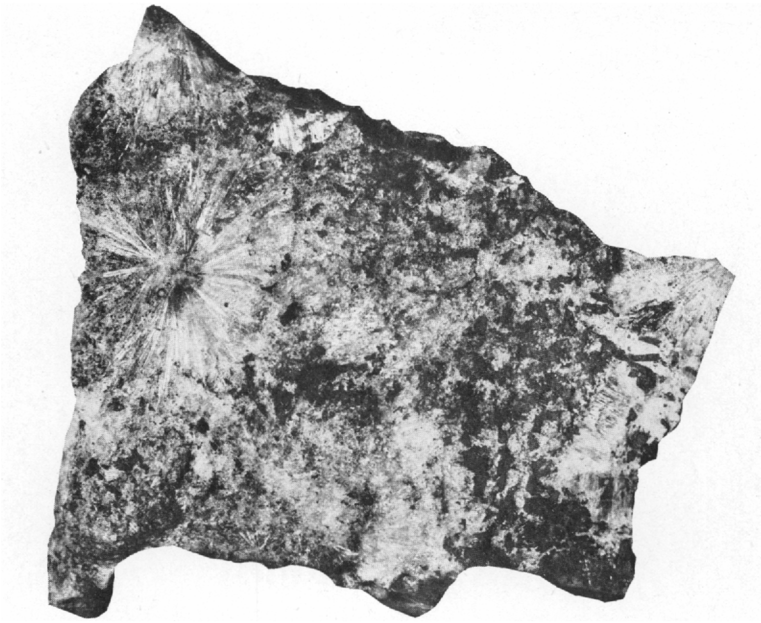
U. S. National Museum catalog No. 95831. Natural size.



B. COLEMANITE AND ULEXITE IN CLAY ABOVE KERNITE DEPOSIT

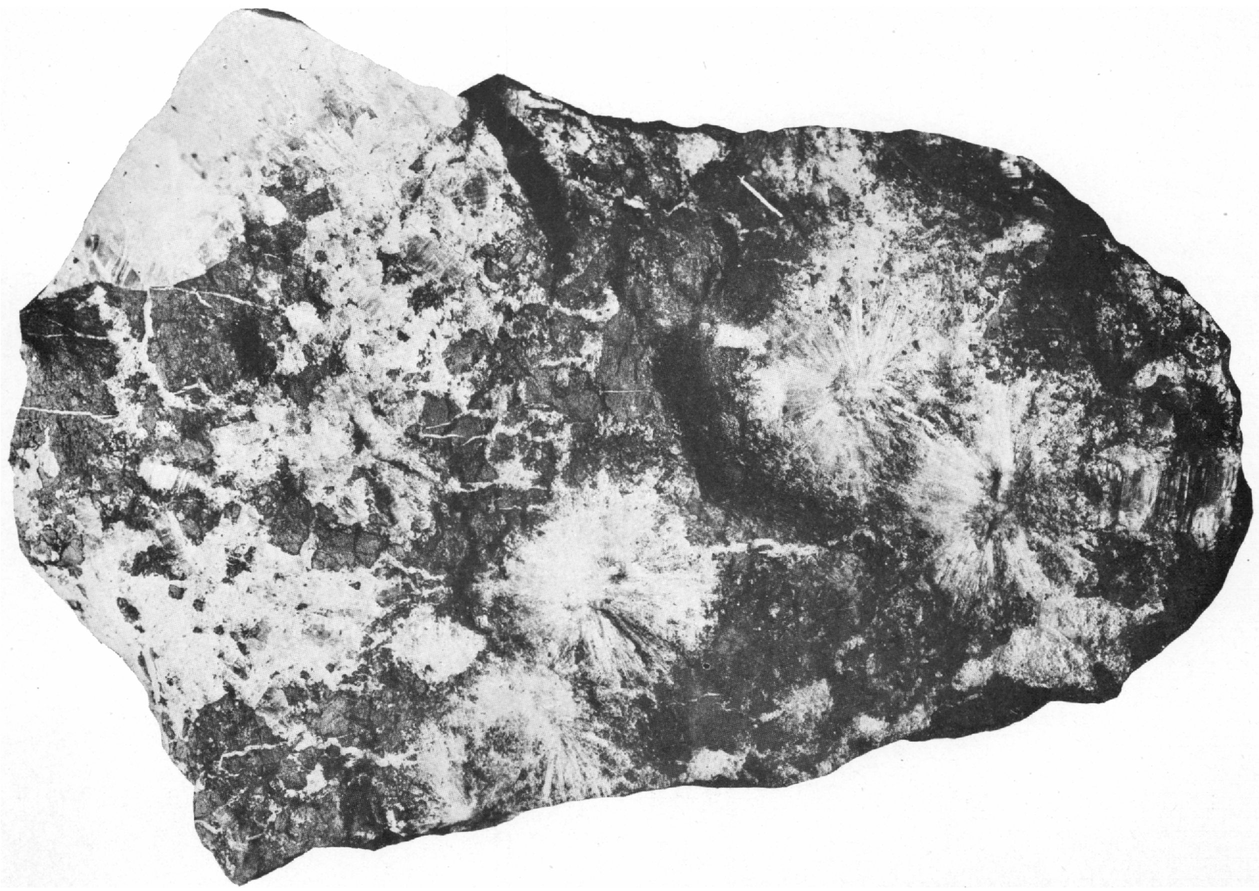
U. S. National Museum catalog No. 95837. Natural size.

ULEXITE AND COLEMANITE



A. RADIATING GROUPS OF KRAMERITE CRYSTALS IN CLAY

U. S. National Museum catalog No. 95832. Natural size.



B. RADIATING GROUPS OF KRAMERITE CRYSTALS IN CLAY ASSOCIATED WITH KERNITE AND BORAX

U. S. National Museum catalog No. 95833. Natural size.

KRAMERITE

Many of the faces of the unit prism are rounded, and nearly all are striated vertically. Occasionally, in a distorted crystal (fig. 23), one pair of the faces are very much larger than the other pair and the crystal is flattened parallel to the prism.

The clinodome *e* (011) is the dominant terminal form on some crystals (fig. 24), or combined with *t* (101) it determines the terminal habit. It also functions as a relatively minor form, as shown in Figures 25 and 26. It is present on all crystals measured except one. Although not shown in the drawing of crystal 14 (fig. 23), it is present as a minute face, giving no reflection. The measurements of *e* (011) were used for the calculation of the axial elements of kramerite.

zone [$\bar{1}01$]. Usually the faces of *d* ($\bar{1}01$) are very small, as on crystal 6, shown in Figure 24.

The positive pyramid *p* (111) forms the dominant terminal form on some crystals (figs. 21 and 23) and even where accompanied by large faces of other forms still asserts its dominant character (figs. 25 and 26). The two faces of the form are commonly not of equal size. (See figs. 23, 25, and 26.) Only rarely is any face of the unit pyramid minute.

The negative pyramid *o* ($\bar{1}11$), the least common of all the forms, is always very small in its development. The reflections were hardly discernible but serve to determine the form. Its largest observed occurrence is shown on crystal 11, Figure 26.

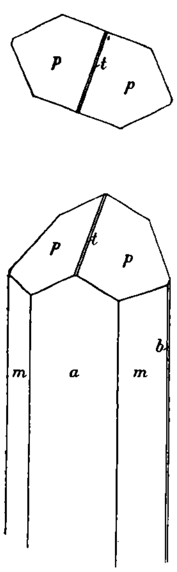


FIGURE 21.—Kramerite crystal 12. Dominant terminal form is *p* (111) with *t* (101) as long, narrow face. No other form present on crystal. All the measured crystals of kramerite are less than 1 millimeter thick

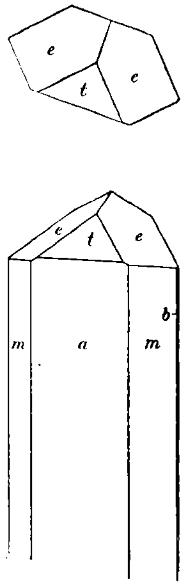


FIGURE 22.—Kramerite crystals 4 and 8. The clinodome *e* (011) is the dominant terminal form, with a large triangular development of *t* (101). Another crystal (No. 6, fig. 24), with *e* (011) the dominant terminal face, does not show *t* (101). Neither crystal 4 nor 8 shows any other form

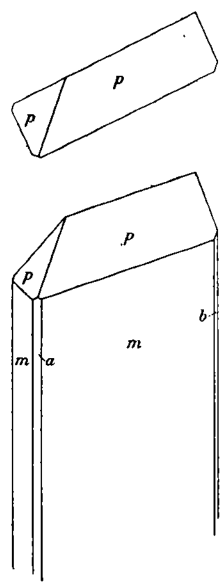


FIGURE 23.—Kramerite crystal 14. Distorted by unequal development of the prism faces, the crystal being flattened parallel to one pair. The clinopinacoid is larger than usual. The pyramid is the dominant terminal form, the crystal also having, though not shown in the drawing, single very minute faces of *e* (011) and *o* ($\bar{1}11$)

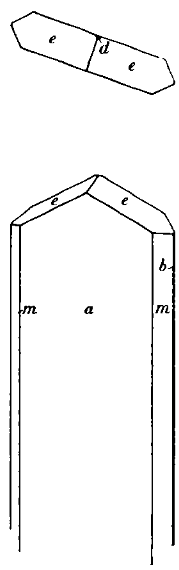


FIGURE 24.—Kramerite crystal 6. Flattened parallel to the orthopinacoid *a*, with *e* (011) the dominant terminal form and with a minute face of *d* ($\bar{1}01$)

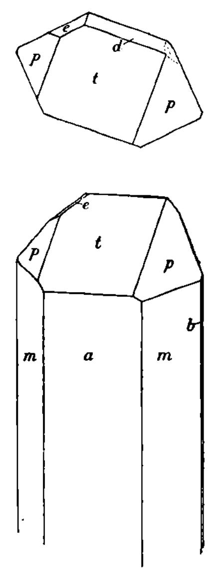


FIGURE 25.—Kramerite crystal 13. Slightly distorted crystal, showing large square face of *t* (101), with medium-sized faces of *p* (111) and smaller faces of *e* (011) and *d* ($\bar{1}01$). The only crystal on which the negative orthodome *d* reached any size. Shows all the kramerite forms except *o* ($\bar{1}11$)

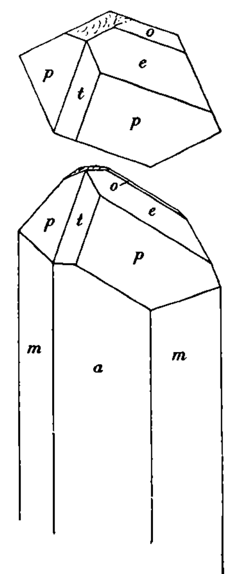


FIGURE 26.—Kramerite crystal 11. Another distorted crystal with *t* (101), *e* (011), *p* (111), and *o* ($\bar{1}11$). This face of *o* is larger than usual. The only crystal found not showing the clinopinacoid *b*. A minute face of *d* ($\bar{1}01$), not shown in the drawing, is present on the crystal

The positive orthodome *t* (101) varies considerably in size and relative dominance. It may be almost the largest terminal form, as in Figure 25, when it has a square shape due to *p* (111) being the other chief terminal form, or it is triangular if *e* (011) is the other chief terminal form. On other crystals the face becomes narrower (fig. 26), grading into a mere line face between the faces of *p* (111), as shown in Figure 21. It is present on about three-quarters of the crystals measured.

The negative orthodome *d* ($\bar{1}01$) is only about half as frequent in its occurrence as the positive dome. On crystal 13 (fig. 25) it is relatively large, the positive dome, with *p*, giving a long intersectional edge in the

The combinations observed on the 14 crystals measured are as follows:

Combinations on crystals of kramerite

Form	Percentage of occurrence	1	2	3	4	5	6	7	8	9	10	11	12	13	14
a (100)	100	*a*	*a*	*a*	*a*	*a*	*a*	*a*	*a*	*a*	*a*	*a*	*a*	*a*	*a*
m (110)	100	*m*	*m*	*m*	*m*	*m*	*m*	*m*	*m*	*m*	*m*	*m*	*m*	*m*	*m*
b (010)	93	*b*	*b*	*b*	*b*	*b*	*b*	*b*	*b*	*b*	*b*	--	*b*	*b*	*b*
e (011)	93	*e*	*e*	*e*	*e*	*e*	*e*	*e*	*e*	*e*	*e*	*e*	--	*e*	*e*
t (101)	71	*t*	*t*	--	*t*	*t*	--	*t*	*t*	*t*	--	*t*	*t*	*t*	-----
d ($\bar{1}01$)	43	--	--	*d*	--	*d*	*d*	*d*	--	--	--	*d*	--	*d*	-----
p (111)	71	*p*	*p*	*p*	--	*p*	--	--	--	*p*	*p*	*p*	*p*	*p*	*p*
o ($\bar{1}11$)	29	--	*o*	*o*	--	--	--	--	--	--	--	*o*	--	--	*o*

No measured crystal showed all the forms, but five crystals showed all the forms but one. Some of the faces were exceedingly minute, giving no reflection, and it is possible that some such forms missed recognition.

DESCRIPTION OF CRYSTALS

The crystals are all of long prismatic habit, the ratio of length to thickness being about 50 : 1. Many of the crystals are of nearly equal thickness in all horizontal directions; usually there is a slight flattening parallel to the orthopinacoid so that the crystal is a little longer in the direction of the *b* axis than in that of the *a* axis. A few are considerably flattened parallel to *a* (100) and rarely, by distortion, flattened parallel to one set of prism faces. (See fig. 23.) No twinning was observed on any of the crystals. The simple termination *e* and *t*, as shown in Figure 22, illustrates one type, as shown by crystals 4, 8, and 9. Another simple type is terminated dominantly by *p* (figs. 21 and 23), as shown by crystals 12 and 14. The drawings of Figures 25 and 26 are shown in the distorted condition in which the faces actually occur. Many of the crystals show a combination lying between the two drawings, with reference to the size and distribution of the terminal faces. Where only *e* is the chief terminal form, as in crystal 6 (fig. 24), the crystal is likely to be flattened parallel to the orthopinacoid.

CHEMICAL COMPOSITION

PYROGNOSTICS

When heated in a closed tube, kramerite decrepitates, giving off water, turns opaque white, swells slightly, and fuses imperfectly to a clinkerlike mass. The decrepitation is not as violent as with the compact fibrous ulexite, but the swelling is greater, ulexite showing hardly any swelling. When heated in the blowpipe flame, a crystal of kramerite readily fuses to a clear bubbly glass, giving a yellow flame. Kramerite is not soluble in water, either hot or cold, but is attacked by both.

ANALYSIS

The samples analyzed were prepared by crushing and hand picking. The results obtained with the first sample analyzed, in triplicate, are shown in the following table under "Hand picked, not washed." The prisms of kramerite penetrate all the associated minerals, and so small particles of borax may have remained in the sample, attached to or included between some of the prisms. Most of the glassy borax and all of the easily cleavable kernite was, however, readily removed. A small percentage of clay, both attached and included in the prisms, could not well be removed. The sample was dissolved in warm hydrochloric acid, which partly attacked the clay. For the B_2O_3 determinations (by titration), cold acid was used as a solvent.

A second sample was prepared after crushing and hand picking, by washing it for a few minutes with an excess of cold water, sufficient to dissolve any small quantity of borax that may have remained in the sample after careful picking.

The sample was small, and B_2O_3 was not determined in it. Both analyses yield closely the same formula, namely, $Na_2O.2CaO.5B_2O_3.10H_2O$.

Analyses of kramerite

[W. T. Schaller, analyst]

	Hand picked, not washed				Washed with cold water
	1	2	3	Average	
SiO_2, soluble in HCl	0.61	0.52	Not det.	0.57	------
Al_2O_3, soluble in HCl	.38	.39	0.49	.42	------
Insoluble in HCl	1.82	1.85	1.55	1.74	0.84
Na_2O	8.19	8.28	8.56	8.34	8.05
CaO	15.25	15.07	15.02	15.11	15.29
B_2O_3	49.24	49.52	49.30	49.35	[a] 50.30
H_2O (total)	25.44	25.23	25.52	25.40	25.52
	------	------	--------	100.93	100.00

[a] By difference.

A little potash was detected spectroscopically, but no chemical reaction was obtained for magnesia, sulphate, chloride, carbonate, or phosphate. The first sample, as analyzed, contained 2.73 per cent of anhydrous clay, which amounts to 3.10 per cent of clay, on the assumption of a water content of 12 per cent in the clay. Deducting this impurity and recalculating the analysis to a summation of 100 per cent gives the results obtained under 1 in the table below. Under 2 is given the analysis of the water-washed sample, with the insoluble clay deducted, recalculated to a summation of 100 per cent. In the third column is shown the calculated composition of the formula $Na_2O.2CaO.5B_2O_3.10H_2O$.

Analyses of kramerite, with insoluble matter deducted

	Not washed	Washed	Calculated
Na_2O	8.53	8.12	8.83
CaO	15.45	15.42	15.98
B_2O_3	50.44	50.73	49.56
H_2O	25.58	25.73	25.63
	100.00	100.00	100.00

The ratios obtained from these analyses are as follows:

Ratios of analyses of kramerite

	Not washed		Washed	
Na_2O	0.1376	0.97 or 1×0.97	0.1310	0.92 or 1×0.92
CaO	0.2754	1.94 or 2×0.97	0.2748	1.93 or 2×0.96
B_2O_3	0.7247	5.10 or 5×1.02	0.7289	5.12 or 5×1.02
H_2O	1.4211	10.00 or 10×1.00	1.4294	10.03 or 10×1.00

These ratios clearly show the correctness of the formula $Na_2O.2CaO.5B_2O_3.10H_2O$.

The loss of water at different temperatures is as follows:

Loss of water of kramerite at increasing temperature

Temperature (°C.)	Loss (per cent)		Per cent of total water lost at each increase
	At each increase	Total	
107	5. 44	5. 44	22
122	2. 75	8. 19	11
150	2. 25	10. 44	9
179	4. 00	14. 44	16
200		14. 44	
280	5. 30	19. 74	21
334	2. 83	22. 57	11
Ignition	2. 66	25. 23	10
	25. 23		100

With a little over a fifth (possibly a fourth) of the water going off at 107°, the remainder of the water goes in many individual steps, approximating closely to a molecule at a time for many of the steps.

Although the mineral is not sufficiently soluble in water to give a clear solution, it is attacked by both cold and hot water, the amount of the differential solution being as follows:

Differential solution of Na_2O and CaO in kramerite, with water

[1 gram of mineral treated with 200 cubic centimeters of water over night]

	Na_2O		CaO	
	Per cent	Per cent of total	Per cent	Per cent of total
Soluble in cold water	1. 13	13	0. 47	3
Soluble in hot water	5. 30	62	9. 28	62
Soluble in hot HCl	2. 13	25	5. 27	35
	8. 56	100	15. 02	100

With cold water, more soda than calcium goes into solution, or else it goes faster; in hot water the same relative quantity of each dissolves. On the average about 70 per cent of the bases will be dissolved by hot water. About 1 gram of the powdered mineral was kept in a flask with 200 cubic centimeters of hot water on the steam bath (about 90°) for a week, but complete solution was not effected.

The establishment of the composition of kramerite determines definitely the existence of two hydrates of this type of compound—namely,

$Na_2O.2CaO.5B_2O_3.10H_2O$, or $H_{10}NaCaB_5O_{14}$, kramerite.
$Na_2O.2CaO.5B_2O_3.16H_2O$, or $H_{16}NaCaB_5O_{17}$, ulexite.

Apparently the lower hydrate prepared by Van't Hoff [6] is a third member of this group. But as stated in the next paragraphs, the supposedly 8-hydrate found by him—namely, $Na_2O.2CaO.5B_2O_3.8H_2O$—is probably the 10-hydrate and identical with kramerite.

Ulexite is stable up to about 60°, according to Van't Hoff, breaking down above this temperature into sodium borate and calcium borates (pandermite or colemanite, depending on the other salts present and character of inoculation).

SYNTHESIS

The lower hydrate prepared by Van't Hoff was formed by heating a mixture of two parts of ulexite and one of borax to a temperature slightly over 60°. The experiment was undertaken in an attempt to make franklandite, which was supposed to have the composition expressed by the formula $2Na_2O.2CaO.6B_2O_3.15H_2O$ (but which was shown by Van't Hoff to be identical with ulexite). He obtained a new compound, in needlelike crystals, to which, on the basis of analysis, he ascribed the formula $Na_2O.2CaO.5B_2O_3.8H_2O$.

In repeating the experiment as described by Van't Hoff except that the sealed tube was heated in the steam bath (about 90°) instead of being heated only slightly over 60°, a saturated solution of sodium chloride was introduced into the mixture of two parts of ulexite and one of borax. After a week's heating the insoluble mass in the tube (the soluble part being brilliant octahedral crystals of $Na_2O.2B_2O_3.5H_2O$—artificial tincalconite) was well washed with water and then air dried. It consisted of a finely fibrous uniform mass, whose optical properties showed that it was artificial kramerite, as follows:

	Artificial product	Kramerite
α	1. 517	1. 515
γ	1. 542	1. 544
Elongation	+	+
Extinction	15°	15°

An analysis of the air-dried artificial preparation yielded a total water content of 26.04 per cent, of which 4.45 per cent was given off on heating to 107° for three hours, leaving 21.59 per cent given off at higher temperature, a content practically identical with that found by Van't Hoff after washing and drying (21.2 per cent). As Van't Hoff gives no details as to how he dried the sample nor at what temperature, it is believed that his

[6] Van't Hoff, J. H., Franklandite und eine neue, dem Boronatrocalcite verwandte Verbindung: Preuss. Akad. Wiss. Sitzungsber., 1907, p. 303.

preparation was artificial kramerite but that on drying before analysis, some of the water was driven off.

The analysis of the artificial kramerite yielded the following figures.

Analysis of artificial kramerite

[W. T. Schaller, analyst]

	Artificial	Calculated
Na_2O	9. 17	8. 83
CaO	15. 87	15. 98
B_2O_3	[a] 48. 92	49. 56
H_2O	26. 04	25. 63
	100. 00	100. 00

[a] By difference.

KERNITE

OCCURRENCE

Kernite has been found only on the east side of the Kramer borate field in Kern County, close to the San Bernardino County line. The mineral was first obtained from the Discovery shaft of the Pacific Coast Borax Co.'s mine, known as the Baker deposit, in sec. 24, T. 11 N., R. 8 W. A later shaft, the Osborne shaft, also in kernite, is 382 feet east of the Discovery shaft, in sec. 19, T. 11 N., R. 7 W. Apparently no kernite was found below the earlier-discovered deposits of colemanite and ulexite lying a few miles to the west, chiefly in secs. 14 and 22, T. 11 N., R. 8 W.

When visited in September, 1927, the workings were reported by Mr. Rasor to show a thickness of the kernite deposit of a little over 100 feet and a lateral extent in two directions of at least 600 to 700 feet each way, the deposit not being penetrated in several places. The horizontal dimensions given are therefore a minimum.

The deposit consists of clay in which are embedded seams and crystals of kernite. (See pl. 27.) The clay, described on page 165, is bedded and faulted and is probably composite in character. The extreme abundance of kernite, its large crystals, its perfect cleavage, and its clearness and transparency render the occurrence one of almost unique character as well as of striking beauty. For many feet kernite is the only borate mineral seen. In places it is so abundant that the clay matrix becomes almost negligible in amount. Although the writer has no figures available to determine the quantity of kernite present he would estimate that not less than 75 per cent of the deposit is formed of this striking mineral. Large bodies of massive borax were encountered, but the proportion of borax to kernite is very small. With a thickness of 100 feet and a lateral extent in two directions of 600 feet each, a rectangular body having a kernite content of 75 per cent would contain at least 1,600,000 short tons of kernite. This figure would be equivalent to about 2,000,000 short tons of borax, as 1 ton of kernite, when dissolved in water, will upon crystallization yield 1.39 tons of borax. That the deposit is probably much larger is suggested by the published statement [7] that

> It appears that an important extension of the richest part of these deposits lies south of the property now being operated, and this new deposit is evidently under independent control. The discovery was made by borings and is situated about three-quarters of a mile southwest of the original shaft, in the northeast corner of sec. 24. According to authentic information, these borax deposits reveal a thick section of particularly pure kernite in the new area referred to, and a mining shaft is being rapidly sunk to develop this new deposit.

This same article states that the average daily shipment of crude kernite ore of high purity is 200 tons. It is shipped to Wilmington, near Los Angeles, where it is dissolved, the clay is filtered off, and the residue is recrystallized into borax.

According to report, the top of the kernite deposit is about 300 feet below the surface. Above the kernite the clay, as seen in dump specimens, is rich in seams of fibrous ulexite of a fine satin-spar appearance. Such an occurrence is shown in Plate 22, *A*. The ulexite seams, though apparently abundant, are only a few inches thick at most, to judge from the specimens collected from the dump. Colemanite also occurs with the ulexite above the kernite but apparently is not very abundant. A fine specimen presented to the United States National Museum by Mr. Rasor is shown in Plate 22, *B*. Mr. Rasor states that both colemanite and ulexite were found for a few feet below the kernite, but neither colemanite nor ulexite was seen anywhere in the kernite deposit by the writer. A bed of igneous rock is reported to lie beneath the borates, as in the occurrences a mile or two to the west.

The minerals associated with the kernite are few, comprising the minerals in the clay, borax, tincalconite, kramerite, calcite, realgar, and stibnite. In most of the deposit no other borate in quantity is associated with the kernite, but it is probable that very small quantities of borax, tincalconite, and kramerite are to be found throughout the deposit. A thin film of white opaque tincalconite coats some of the kernite crystals and also forms on many of the cleavage surfaces. Though not present in large amounts, it is very widespread in its occurrence.

Genetically the most interesting association of kernite and other borates occurs where the irregularly bounded crystals of kernite are embedded in massive borax. A sufficient number of such occurrences were noted to show that the relation is not unique. Such kernite crystals embedded in borax as were collected are irregular in shape (see fig. 33) and show no definite crystal faces, all the faces of *c* (001) and *a* (100) observed being cleavage faces. It is very probable, however, that crystals bounded by true faces (not

[7] Eng. and Min. Jour., vol. 125, p. 551, 1928.

cleavage) are present in the borax. The sides corresponding to the faces of the clinodome *e* (011) have a rough, dull appearance just like many of the kernite crystals embedded in the clay matrix. These kernite crystals in the borax reach a thickness of several inches. In the presence of borax kernite specimens (in the Washington laboratory) readily alter on the surface to tincalconite. This alteration is described below in connection with the chemical composition. In general appearance the kernite crystals in the borax matrix are identical with those in the clay matrix. When collected, the borax of these specimens was bright and glassy, but within a few months, in the Washington laboratory, the borax alters to tincalconite and part of the kernite undergoes a similar alteration, though very much more slowly.

In the mine kernite is most abundant as large crystals (pls. 24, *A*, and 27), though veins of cleavable kernite (pl. 24, *B*), as well as irregular masses, are likewise very abundant. The largest crystal noted, near the Discovery shaft, measured 8 feet (*c* axis) by 3 feet. Single crystals 2 to 3 feet thick are very common. As seen in the face of the tunnels, the exposed surface of the crystals is bounded by alternating faces of the cleavages *c* (001) and *a* (100). (See fig. 32.) By careful removal of the clay matrix it might be possible to obtain large crystals of kernite bounded by natural faces. One of the specimens of clay matrix contains abundant crystals averaging about 1 centimeter in thickness. By careful picking and removal of the clay, natural faces of forms in the orthodome zone—either *c* (001), *a* (100), or *D* ($\bar{1}$01)—can be seen to be present. The faces of the other terminal form—namely, the clinodome *e* (011)—are always dull and rounded. In fact, if only the kernite crystals embedded in the clay were available for the determination of the crystallographic elements, they could not be accurately determined. Fortunately, some excellent small crystals were found completely embedded in the large crystals of kernite, and these small crystals were removed by very careful cleaving of the large crystals and found to be bounded by smooth and brilliant faces whose measurements furnished the required angles for the calculation of the crystallographic constants. Small embedded crystals of kernite are also developed in some of the fracture zones of the larger cleavage pieces. The veins of kernite, several inches thick and 10 feet or more long, show an apparent fibrous structure either normal or slightly inclined to the walls. No definite genetic relation was established between these veins and the individual crystals, but many of the large irregular-shaped well-crystallized masses, showing consistent cleavage throughout, have what appear like vein feeders. Many of the crystals show no connecting or feeding veins and appear to be isolated in the clay, but the kernite is so abundant that it is difficult to show conclusively that these crystals are not connected to other masses or crystals by smaller veins. (See pl. 27.)

Much of the clay is richly seamed with veins of kernite, in both parallel and divergent groups. Many of the kernite veins follow the bedding of the clay, but many also cut across the bedding at all angles and swell and pinch and are most irregular in shape and size. "Eyes," disconnected or connected by the merest threadlike veins, such as are typical of quartz veins in schist or gneiss, were not seen. The kernite veins are commonly either nearly horizontal or have a small inclination. Vertical veins are not common.

Many veins of kernite have begun to form in the clay, with but a small percentage of the borate mineral present. At other places the kernite is more abundant, and many of the veins consist of nearly pure kernite, with a very small quantity of included clay. A set of such parallel veins in the clay is shown in Plate 24, *B;* the upper vein contains a minimum of clay, but clay greatly predominates in the lower veins. All the kernite veins in the specimen are nearly parallel, and the kernite itself is all in parallel position. It is believed that if the process of kernite formation had continued, practically all the clay would have been removed and a single vein of pure kernite would have resulted.

The fibrous structure shown by the veins is due to the excellent cleavages possessed by the mineral, which when well developed greatly simulate a fibrous structure. The cleavage of kernite is parallel to faces all lying in the orthodome zone. It is perfect parallel to two forms, the base *c* (001) and the orthopinacoid *a* (100), and imperfect parallel to the rear or negative unit orthodome *D* ($\bar{1}$01). These cleavages have a strong tendency to break down a crystal into a mass of fine fibers resembling a mass of tremolite asbestos. In places on the floor of the tunnels there will be a mass of such fibers of considerable thickness, and unless one breaks up a mass of solid kernite himself he finds it hard to believe that such a mass of fine fibers could develop from a solid piece of the mineral. The two cleavages are so perfect that very thin hairlike fibers several inches long can readily be cleaved off.

The general appearance of the large crystals, as seen in the mine (pl. 27, *B*), is illustrated by Plate 24, *A*, which shows a crystal nearly 9 centimeters high and 6 centimeters wide. The front is formed by alternating cleavages of *c* (001) and *a* (100), and the two rough faces of the clinodome *e* (011) show in part. Large crystals of similar shape are very abundant in the mine. Many of these have a distinct 6-sided appearance, suggesting that the clinopinacoid *b* (010) is present as a large face, although on the very small measured crystals, embedded in a large crystal, the clinopinacoid is a rare form and present only as a line face. On the large crystals seen in the mine the top angle between the two sloping faces is distinctly greater than 90°

and was estimated to be about 110°, which agrees with the calculated angle of 116° 14′ for *e* (011) ∧ *e*′ (0$\bar{1}$1).

Another mass of kernite is shown in Plate 25, *A*. This mass represents the irregular shape of much of the kernite in the mine, for where it is in contact with the clay matrix there is usually no definite crystal face present. This specimen shows a number of other relationships, for the mass of kernite contains a number of small veins of borax cutting right through the kernite. The relations, as shown in Plate 25, *A*, are illustrated and described in Figure 27. The seam of banded clay probably represents a harder and more resistant layer, which was difficult to replace by the borate minerals. A similar but smaller layer of such harder clay is seen on the extreme right of the specimen. These harder layers contain a little borax, whereas on each side of the larger layer, in the mixture of borax and clay, the borax is present in the largest quantity. It is in the kernite specimen here shown that the first small embedded crystals were found,

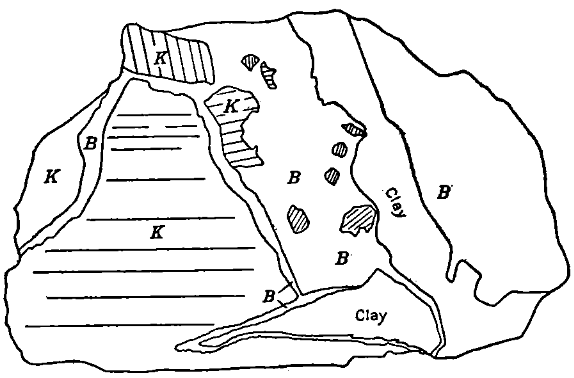


FIGURE 27.—Relations of kernite and borax. Explanatory of Plate 25, *A*. The seams of borax cutting kernite (K) are marked B. They are now coated by its dehydration product, tincalconite. The seam of clay, containing a little borax, which is harder and more compact than the other clay, is well shown in Plate 25, *A*

but since then such crystals have been observed in several other masses of the kernite.

PHYSICAL AND OPTICAL PROPERTIES

Kernite possesses a number of separation directions of varied perfection. Two cleavages are perfect, parallel to the base *c* (001) and the orthopinacoid *a* (100). A third direction of fair cleavage is parallel to *D* ($\bar{1}$01). On some large cleavage fragments there is a suggestion of a poor cleavage after *e* (011), and a few fragments were found on which a cleaved surface parallel to this clinodome was observed. Several pieces also show a distinct fracture—it can hardly be called cleavage—after the clinopinacoid *b* (010), which yields irregular and not plane surfaces, some of which are fairly parallel to the *b* face.

The cleavage after the base *c* (001) is perhaps the best developed and easiest obtained. Many of its surfaces yield mirror-like reflections on the goniometer. Many of these cleavage surfaces also show a multitude of lines parallel to the intersection edge *c* (001) ∧ *e* (011). These may cover the entire surface of the *c* cleavage and suggest traces of planes of separation or of gliding or warping (or possibly twinning) parallel to *b* (010). They resemble the traces of parting planes seen on crystals of stibnite. Warping in this zone, *c e b*, is common for the kernite cleavage fragments, the pieces being warped or bent for several degrees; one long cleavage piece (19 centimeters long and 1 centimeter thick) was warped through an angle of nearly 4°. Thin cleavage pieces (after *c* (001)) can be bent through a small angle without breaking, quickly springing back on release of pressure. The difference between the *c* and *a* cleavages can be readily seen in pieces fractured parallel to *b* (010). On these pieces the cleavage after the base *c* is developed as innumerable closely spaced lines, giving almost a striated appearance to the surface, whereas the cleavage after *a* may show only as a few—perhaps three or four—well-developed lines.

The cleavage after *a* (100) is likewise perfect, and many surfaces yield similar mirror-like reflections. There is, however, more of a tendency to develop small fibers on this cleavage than on that after the base. The cleavage surfaces of *a* do not show the parting lines so characteristic of the basal cleavage.

The cleavage after *D* ($\bar{1}$01) is decidedly fibrous. Whereas plates or sheets can be readily split off parallel to *D* ($\bar{1}$01), their surfaces are not smooth and mirror-like but strongly fibrous, owing in part perhaps to the simultaneous development of the *a* and *c* cleavages. Most of the occurrences of the form *D* ($\bar{1}$01) recorded on crystals 1 to 12 are cleavage faces.

The separation after *e* (011) can be classed only as very rare and not well developed, but the resultant faces are fairly plane and resemble a cleavage rather than a fracture. The separation after *b* (010) is better developed than that after *e* (011) and often shows on the *a* and *c* cleavages as strong cracks, but the resultant faces are never plane but always uneven, so that this separation should be classed as a fracture rather than a cleavage.

The multitude of lines on the basal cleavage, parallel to the intersection edge *c* (001) ∧ *e* (011), may result from warping, or gliding planes, or twinning (after *e* (011)). They can be produced by pressing a piece of kernite in the direction of the *b* axis. In fact, by clamping a cleavage fragment between two pieces of soft wood in a vise and gradually increasing the pressure, not only can these lines be readily developed, but the kernite fragment is so rigid that the two ends are forced into the wood for a considerable distance. By pressing or rolling a cleavage piece between the fingers in a direction at right angles to the *b* axis the mineral is easily broken into innumerable fibers, all three cleavages in the orthodome zone developing.

The hardness of kernite is 2½ but varies slightly on the different faces and in different directions. Calcite readily scratches kernite, and kernite readily cuts gypsum. The average density obtained by suspending 12 clear fragments in diluted bromoform was 1.911, and a second group of 12 clear fragments gave an average of 1.904. Slight variation in the fragments is caused by the presence of minute traces of clay, of a minute quantity of adhering tincalconite, of the abundant inclusions of negative crystals (probably filled with some liquid), and of incipient cleavage cracks holding some air. The average of the two results, or 1.908, is probably close to the true density. This figure is slightly lower than the value first given, but being made on more abundant material, almost free from clay, is considered more accurate. Kernite is colorless, transparent. The white opaque appearance shown by some pieces is due to the presence of a film of tincalconite. The luster is vitreous but on a fibrous cleavage surface slightly satiny. The streak is white.

The optical axial plane is normal to the clinopinacoid *b* (010), X and Y lying in the plane of symmetry. The mineral is definitely negative, determined by the bar on a section normal to an optic axis. Sections parallel to *a* (100) show an inclined interference figure, of negative character. Therefore, *b* axis $= Z$, $Bx_o \perp (010)$. On a section parallel to the clinopinacoid *b* (010), $X \wedge 001$ cleavage $= 37\frac{1}{2}°$; $X \wedge c$ axis (*a* cleavage) $= 70\frac{1}{2}°$. Therefore a positive orthodome, with a ρ angle of $70\frac{1}{2}$, would be normal to the acute bisectrix. On the *a* cleavage the axial plane is tilted 20°; on the *c* cleavage the tilt is $52\frac{1}{2}°$. The *a* cleavage will therefore show the interference figure inclined to one side, whereas the *c* cleavage does not yield a figure. The optical orientation of kernite is shown in Figure 28.

Measurement of the refractive indices by the oil-immersion method gave the following results:

Refractive indices of kernerite

	Schaller	Ross	Henderson	Average	Artificial kernite
α	1.454	1.454	1.455	1.454	1.455
β	1.471	1.472	1.473	1.472	1.472
γ	1.487	1.489	1.488	1.488	1.487
B	.033	.035	.033	.034	.032

The axial angle, 2V, was measured as 80°. The angle in air, 2E, is 142°. Dispersion distinct, $\rho > v$.

CRYSTALLOGRAPHY

GENERAL DESCRIPTION

From the small cleavage fragment first received from Mr. Gale it was concluded [8] that kernite was orthorhombic and that the cleavage was prismatic. When more material was received it was noticed that the two pairs of cleavage faces showed considerable differences, and it was soon seen that kernite is monoclinic and that the cleavages belong to forms in the orthodome zone. This conclusion was independently arrived at by Dr. C. Hlawatsch,[9] of Vienna, who also called attention to the marked difference of the etch figures on the two cleavages. The specimens of kernite that had been distributed were cleavage pieces and showed an elongation parallel to the cleavages. This direction, originally chosen as the *c* axis when the mineral was thought to be orthorhombic with prismatic cleavage, is the direction of the *b* axis of the monoclinic crystals, and although the cleavage pieces are elongated in this direction, the crystals themselves are not, being more nearly equant (pl. 24, *A*), with generally a slight elongation in the vertical direction. An examination of the outlines presented by many small crystals embedded in the clay matrix, as given in Figure 33, shows that elongation of the crystals parallel to the *b* axis is very rare, the common elongation being in the direction of the *c* axis, as also illustrated in Figure 32, which shows, too, how a cleavage fragment would lie horizontally in the crystal and have an elongation normal to the elongation of the crystal itself.

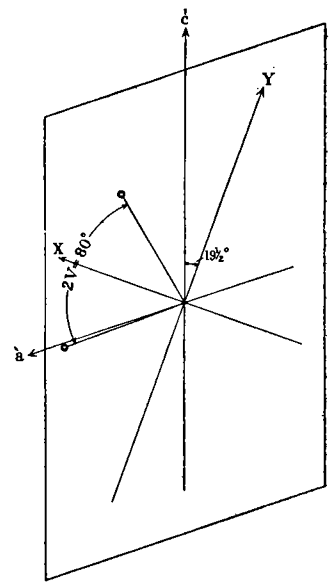


FIGURE 28.—Optical orientation of kernite

The large crystals seen in the mine have the appearance of the one shown in Plate 24, *A*, with various modifications. An attempt to show the actual appearance of a kernite crystal unusually elongated parallel to the *c* axis and embedded in the clay matrix is given in Figure 32, where no actual crystal faces are present but the exposed front side of the crystal is bounded by alternating cleavage faces of the basal pinacoid *c* (001) and the orthopinacoid *a* (100). Both of these cleavages are so nearly perfect that it is difficult to handle a piece or crystal of kernite without developing one or both of them.

Several of the cleavage fragments received later from Mr. Gale showed a number of narrow line-face forms, all in the orthodome zone, but no other terminal faces could be found. The material collected by the writer contained several pieces which showed small embedded crystals of kernite in the larger cleavage fragments. These small crystals were carefully removed, and a few terminated crystals were obtained. The measure-

[8] Schaller, W. T., Kernite, a new sodium borate: Am. Mineralogist, vol. 12, p. 24, 1927.

[9] Personal communication.

ments of these small terminated crystals afforded a basis for deciphering the crystallography of kernite. The perfect cleavages of the mineral made it extremely difficult to remove and to handle these small crystals. On the other hand, it is doubtful if they could have been obtained from the inside of the larger masses but for this perfect cleavage. The crystal measurements were made, therefore, on a composite lot of cleavage pieces showing only very narrow line faces as natural crystal faces, on cleaved fragments of terminated crystals, and on a very few complete crystals.

It was soon seen that the crystals of kernite were bounded by faces in only two zones—one, the dominant zone, rich in faces and heavily striated, and the other zone, at right angles to it, showing essentially only one large form, with four faces all around the crystal, not striated. The cleavages were readily seen to lie in the striated zone. Goniometric measurements showed that many forms were present in the striated zone, that the single form seen in the second zone was the dominant one and the few other forms found in this zone were represented only by very narrow faces. No other forms not shown in one of these two zones were observed on any of the crystals.

A total of 20 crystals or cleavage fragments were measured. Of these, Nos. 1 to 12 were cleavage pieces that showed forms additional to the cleavage faces of *c* (001) and *a* (100). Fragment No. 9 was part of No. 8, and fragment No. 12 was part of No. 11. Nos. 13 and 14 were cleavage fragments of a crystal, and Nos. 16, 17, and 18 were cleavage pieces of one crystal which broke or cleaved into several pieces on removal from its kernite matrix.

Measurements were therefore made on three complete crystals (Nos. 15, 19, 20), two other crystals separated into two (Nos. 13 and 14) and three pieces (Nos. 16, 17, 18), respectively, and 10 cleavage fragments. Where it is necessary to refer to fragments 13 and 14 collectively, they will be called crystal 13, and fragments 16, 17, and 18 will be called crystal 16. The crystallography of kernite, with regard to the development and relationship of crystal forms and faces, and the accompanying crystal drawings are therefore based on only five complete crystals. Several more such embedded crystals were seen, but they showed no essential differences from those measured.

Some of the crystal drawings represent an idealized type of combination, rather than a single crystal measured, owing in part to the fact that most of the crystals were somewhat distorted and showed differences in combinations of forms and development of faces in front and rear, and owing also to the strong striations, embayments, and irregular development that most of them showed.

The crystals of kernite can be referred to two fundamental types, both of the same general habit, which are illustrated in Figures 29 and 30. The only difference between the two drawings is that in Figure 29 the dominant form is the orthopinacoid *a* (100), whereas in Figure 30 the dominant form is the rear negative dome *D* ($\bar{1}01$). The only other forms shown are the base *c* (001) and the clinodome *e* (011). A combination of these two developments, showing both *a* (100) and *D* ($\bar{1}01$) is shown in Figure 31, wherein is also illustrated another characteristic—namely, that one basal pinacoid shows its greatest development in a direction normal to that of the other basal pinacoid. Either one of these three combinations, modified by other forms, represents the habit of all the crystals measured, although none of the measured crystals were as simple in their combinations as those shown in Figures 29, 30, and 31. Figures 35, 36, and 40 represent variations and modifications of Figure 29; Figure 37 represents variations and modifications of Figure 30; and Figures 38 and 39 show various modifications of the habit and combination shown in Figure 31.

It is very probable, however, that many of the large crystals of kernite in the mine will show a simple combination, such as is depicted in Figures 29 to 31. The

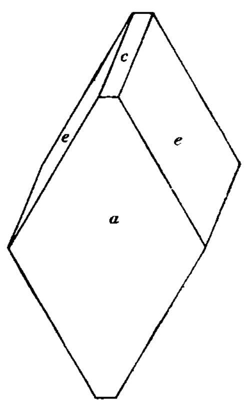


FIGURE 29.—Fundamental simple type of kernite crystal, with large development of *a* (100). Other forms are *c* (001) and *e* (011)

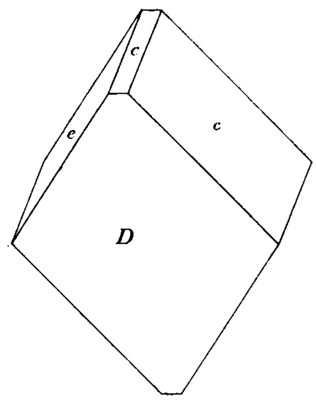


FIGURE 30.—Fundamental simple type of kernite crystal, with large development of *D* ($\bar{1}01$)

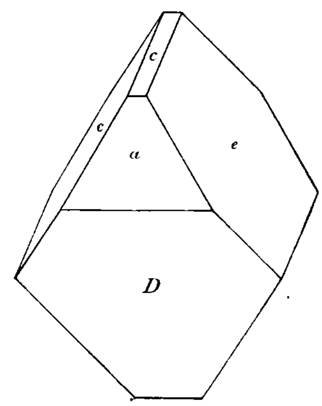
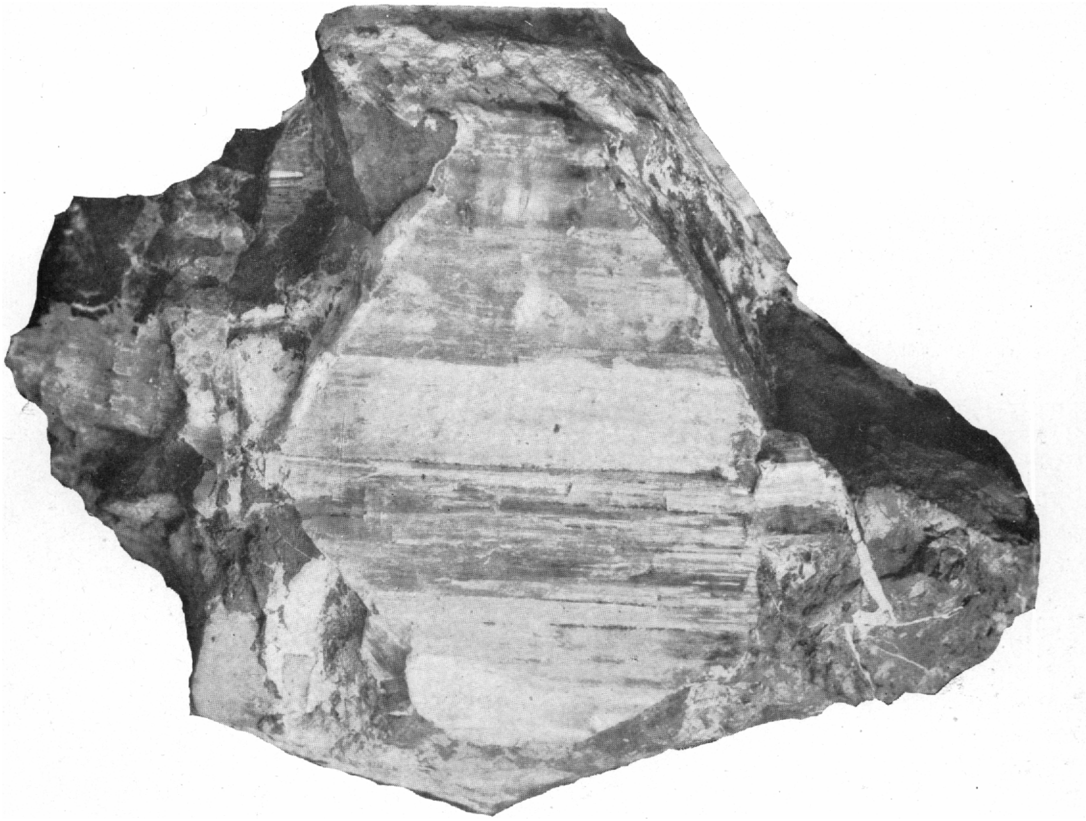


FIGURE 31.—Combination of the two types shown in Figures 29 and 30. Forms *a* (100), *c* (001), *e* (011), *D* ($\bar{1}01$)



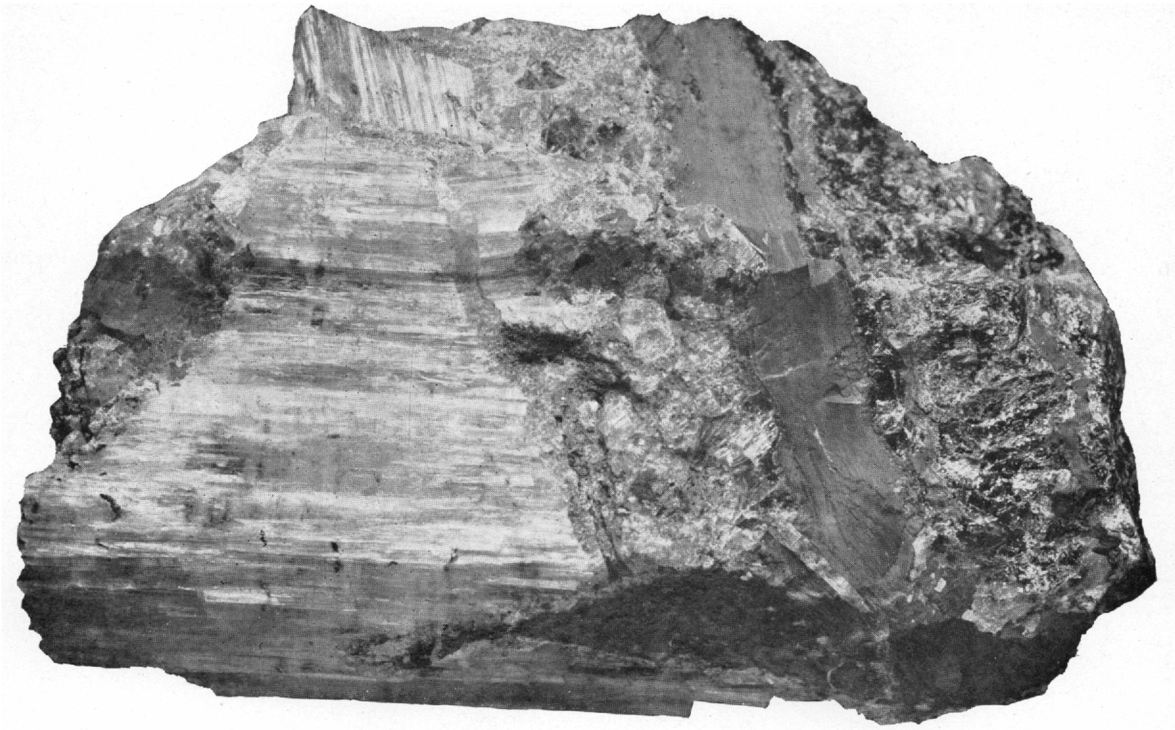
A. KERNITE CRYSTAL IN CLAY

U. S. National Museum catalog No. 95834. Natural size.

B. SEAMS OF KERNITE IN CLAY

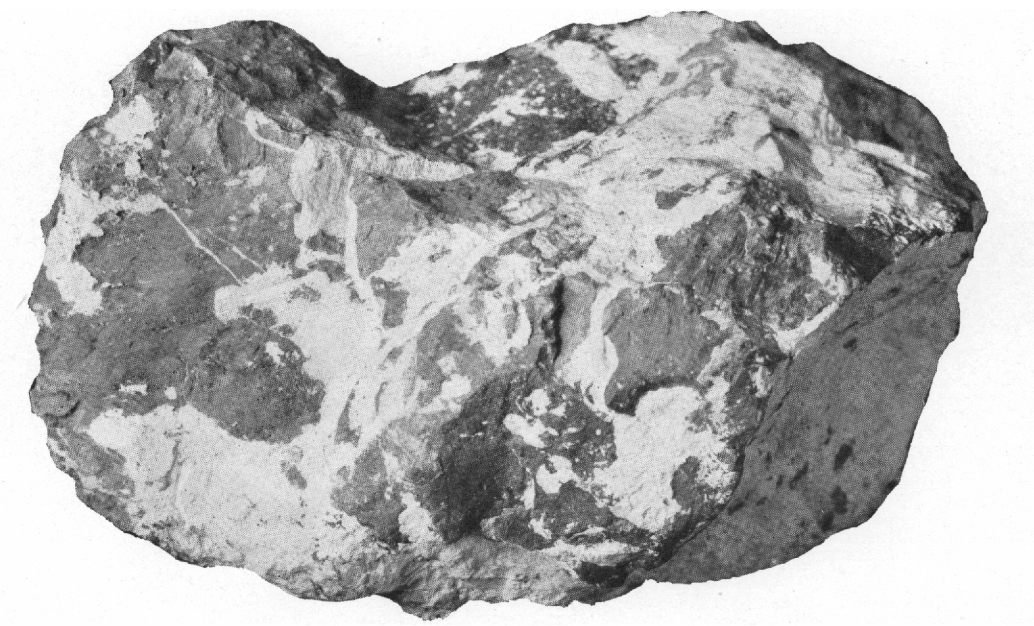
U. S. National Museum catalog No. 95835. Natural size.

KERNITE



A. KERNITE WITH SEAMS OF BORAX AND TINCALCONITE

Relationships explained by Figure 27. U. S. National Museum catalog No. 95835. Natural size.



B. TINCALCONITE

Back side of specimen illustrated in Plate 24, *A*. U. S. National Museum catalog No. 95834. Natural size.

KERNITE WITH BORAX AND TINCALCONITE

actual form habit of the large crystals embedded in the clay is very difficult to determine, as in breaking them out the perfect cleavages parallel to *c* (001) and to *a* (100) readily develop and the real crystal face forming the external boundary of the crystal is lost. The crystal shown in Plate 24, *A*, probably had a form combination shown by either Figure 29, 30, or 31. Owing to this development of the two perfect cleavages the exposed surface of the crystal in the clay shows many steplike alternations of the two cleavages (*a* and *c*), together with a highly striated surface, such as is shown in Figure 32, which is an attempt to illustrate the appearance of such a crystal embedded in the clay matrix. All the faces lettered *c* and *a* are cleavage faces. The two clinodomes, *e* (011) and *e'* ($0\bar{1}1$), are probably represented, but both their surfaces are uneven and dull. The crystal shown in Figure 32 is also considerably elongated parallel to the *c* axis, an occurrence not at all rare.

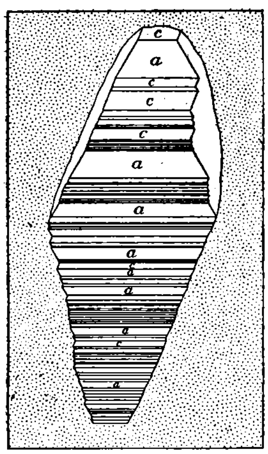


FIGURE 32.—Appearance of kernite crystal in clay matrix, showing the alternation and repeated cleavage faces of *c* (001) and of *a* (100)

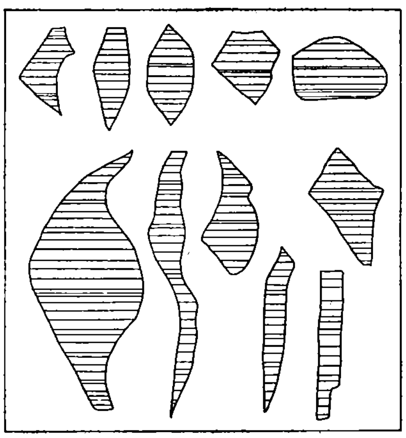


FIGURE 33.—Various shapes outlined by kernite crystals embedded in the clay matrix

Many of the crystals in the clay matrix are bounded by rounded surfaces, so that as viewed on the broken surface bounded by the cleavages, the crystals appear of various shapes and configurations. Figure 33 shows a collection of such shapes as were observed on a specimen of the clay matrix containing many of these small kernite crystals, averaging from about 1 to 3 centimeters high. Some of these outlines suggest the presence of a large face of the clinopinacoid *b* (010), especially the sketch in the lower right-hand corner of Figure 33, but none of the crystals measured showed any large development of this crystal form.

The crystals measured, Nos. 13, 15, 16, 19, and 20, averaged several millimeters in size. Thus crystal 15 is about 2 millimeters high, 1½ millimeters wide, and 1 millimeter thick; crystal 19 is about 2½ by 2 by 1 millimeter; crystal 20 is 3 millimeters high. The crystal whose cleaved fragments form Nos. 16, 17, and 18 was probably at least 5 millimeters high, and after the crystallographic measurements were made a crystal embedded in kernite was found that is about 1 centimeter high.

All the measured and observed crystals are simple and not twinned, but on a large specimen 2 feet thick partial development of crystal faces shows twins of kernite (after *e* (011)). (See fig. 41.)

Kernite being monoclinic, the orthodome zone remains fixed in any orientation that might be chosen. But the crystal can be revolved on the *b* axis into any position, and as only one other form, lettered *e* in the drawings, reaches any development, the allocation of this form determines the orientation. No conclusive reason could be found for decisively making either of the two perfect cleavages the basal pinacoid. The only important form present on these crystals outside of the orthodome zone lies in the zone *b* (010) $\wedge$ one of the perfect cleavages. This form then becomes either the unit prism or the unit clinodome. Both orientations were considered for a while, and it was finally decided to choose the clinodome, chiefly because the habit of the crystals could be best shown in drawings in this position, and the appearance in the mine—as shown in Plate 24, *A*, for example—could then be better compared with the drawings. The simplicity of the indices of the crystal forms in the two orientations remains the same. The change simply amounts to an interchange of the basal or orthopinacoids, with a revolution of the crystal through 180°, so as to keep the angle $001\wedge100$ less than 90°.

CALCULATION OF ELEMENTS

Although many crystal forms are present in kernite, a total of 36 having been determined, the crystallographic elements were calculated from the measurements of only three forms. The finding of more and perhaps better material and the measurement of its angles may give more accurate figures than those here presented. The measurements of the angle between the two perfect cleavages have given consistent results, and so have readings on the natural faces of *c* (001). The average of the 25 best readings of ρ, finally taken, is 18° 52′. This average is derived from 2 readings of 18° 49′, 3 of 18° 50′, 3 of 18° 51′, 10 of 18° 52′, 5 of 18° 53′, 1 of 18° 54′, and 1 of 18° 55′. This is the same value (or more accurately, its complement) originally obtained and given as the prismatic cleavage angle $m\wedge m = 71°\ 08'$, when kernite was considered orthorhombic. The cleavage angle ($001\wedge 100 =$ complement of ρ (001)) was also again measured directly on 10 separate cleavage fragments, the minutes thus found being 9, 7, 7, 7, 10, 6, 8½, 7, 8, 8, an average of 7¾—that is, 71° 07¾′, or a ρ value of 18° 52¼′. The angle β (or μ if β is taken as greater than 90°) for kernite is then 71° 08′, or ρ (001) = 18° 52′. The fair or good reflections from all the cleavage and natural faces gave a ρ value for c (001) ranging from 18° 45′ to 18° 56′. The average of all such fair to good readings is 18° 51′+.

Many less good readings were obtained on e (011) than on c (001), because the faces were not as smooth and brilliant. The crystals were all measured in the orientation where e (011) was taken as the unit prism, so that the figures here given represent the ϕ value—that is, the angle between e and the clinopinacoid b (010), when e is taken as the unit prism (110). Some of the faces of e also gave two reflections, which reached a maximum difference of half a degree. However, by proper observation and adjustment, the following measurements of the angle $e \wedge b$ (010), representing e taken as (110), were obtained:

°	′	°	′	°	′	°	′
31	35	31	43	31	53	32	07
31	35	31	44	31	54	32	08
31	36	31	45	32	01	32	10
31	38	31	46	32	02	32	10
31	38	31	48	32	04	32	12
31	40	31	51	32	05	32	12
31	41	31	53	32	06		

The properly weighted average of these 27 measurements is 31° 53′. As the angle $001 \wedge 010$ is 90°, the complement of this average, or 58° 07′, is the angle c (001)$\wedge e$ (011). This value, with the ρ value of 18° 52′ for c (001), gives a basis for calculating the c axis of kernite, which is found to be 1.6989.

The ρ angle for the rear negative unit dome D ($\bar{1}01$), in the position where e is taken as the unit prism, was found to average 31′ 12°, based on the following 13 measurements:

°	′	°	′	°	′	°	′
30	56	31	11	31	12	31	17
30	58	31	11	31	14	31	22
31	09	31	11	31	15	31	26
31	11						

The average of all measurements, ranging from 29° 54′ to 31° 40′, is 31° 10′, or nearly the same. The average of the best readings of large faces or cleavages, giving good reflections, ranging from 30° 58′ to 31° 26′, also gives 31° 12′. The ρ angle for D ($\bar{1}01$), with e as (110), is then 31° 12′, and the angle c (001) $\wedge$ D ($\bar{1}01$) in this position becomes (31° 12′+18° 52′) 50° 04′. In the orientation finally chosen, with e as (011), the faces c and a are interchanged, so that the angle 50° 04′ becomes that of D ($\bar{1}01$) $\wedge$ a (100), and ρ D ($\bar{1}01$) in the correct position is (90° 00′−50° 04′) 39° 56′.

With the values for β and the c axis already determined, the a axis is readily calculated from the ρ value (39° 56′) of D ($\bar{1}01$) and found to be 1.5230.

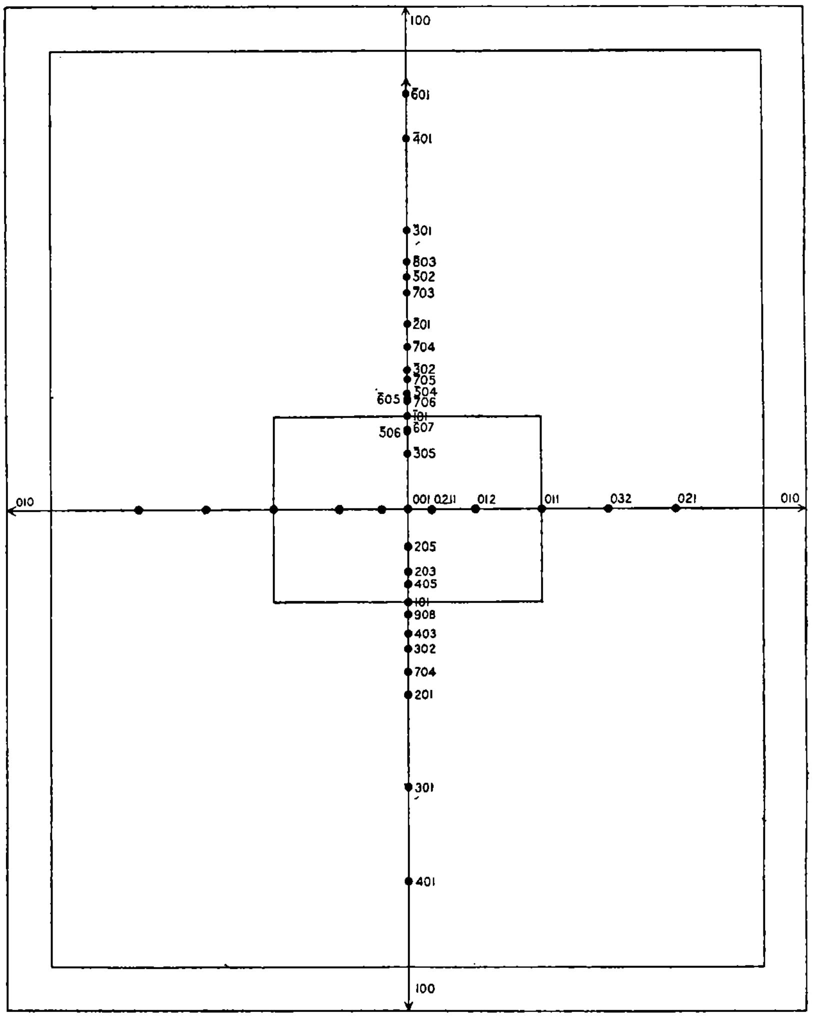


FIGURE 34.—Gnomonic projection of crystal forms of kernite

The axial elements for kernite then are $a : b : c = 1.5230 : 1 : 1.6989$, $\beta = 71° 08'$, $p'_o = 1.1788$, $q'_o = 1.6989$, $e' = 0.3417$.

FORMS AND ANGLES

A total of 36 forms were determined to be present on the kernite crystals and fragments measured. These forms include the 3 pinacoids, 5 clinodomes, 11 positive orthodomes, and 17 negative orthodomes. Only forms belonging to the two dome zones are present, there being no prisms and no pyramids. The distribution of the forms is well shown in the gnomonic projection in Figure 34. Each of the forms was observed on at least two crystals. The forms determined, with their average measured and calculated angles are shown in the table below.

Forms and angles of kernite

Form		Measured		Calculated	
Letter	Symbol	ϕ	ρ	ϕ	ρ
		° ′	° ′	° ′	° ′
c	001	90 00	18 52	90 00	18 52
b	010	0 13	90 18	0 00	90 00
a	100	90 00	90 00	90 00	90 00
f	0.2.11	See table of measurements below.		47 53	24 44
g	012			21 55	42 29
e	011			11 22	59 57
h	032			7 38	68 45
i	021			5 45	73 41
j	205	90 00	38 49	90 00	39 07
k	203	90 00	47 52	90 00	48 26
l	405	90 00	52 01	90 00	52 06
d	101	90 00	56 35	90 00	56 40
n	908	90 00	59 04	90 00	59 03
p	403	90 00	61 52	90 00	62 24
q	302	90 00	64 20	90 00	64 38
r	704	90 00	67 19	90 00	67 25
s	201	90 00	69 45	90 00	69 40
t	301	90 00	74 23	90 00	75 32
u	401	90 00	77 56	90 00	78 49
A	$\bar{3}05$	90 00	20 08	90 00	20 05
B	$\bar{5}06$	90 00	32 47	90 00	32 39
C	$\bar{6}07$	90 00	33 42	90 00	33 46
D	$\bar{1}01$	90 00	39 56	90 00	39 56
E	$\bar{7}06$	90 00	45 52	90 00	45 57
F	$\bar{6}05$	90 00	47 10	90 00	47 01
G	$\bar{5}04$	90 00	48 38	90 00	48 32
H	$\bar{7}05$	90 00	52 28	90 00	52 37
J	$\bar{3}02$	90 00	54 49	90 00	54 58
K	$\bar{7}04$	90 00	59 46	90 00	59 51
L	$\bar{2}01$	90 00	63 36	90 00	63 37
M	$\bar{7}03$	90 00	67 35	90 00	67 27
N	$\bar{5}02$	90 00	69 02	90 00	69 00
P	$\bar{8}03$	90 00	70 23	90 00	70 21
Q	$\bar{3}01$	90 00	72 29	90 00	72 37
R	$\bar{4}01$	90 00	76 56	90 00	77 07
S	$\bar{6}01$	90 00	81 49	90 00	81 33

As the clinodomes were measured in the orientation where they functioned as prisms, their true ϕ and ρ values were not measured; instead, the angle *c* (001)$\wedge$*okl* was measured. These measurements are shown in the table below, in which the numbers of the crystals are given at the left of the individual measurements. The values in parentheses are given as a matter of record; they are not used in obtaining the average values. Measurements of *e* (011) are given on page 152.

Measurements of clinodomes, kernite

[Angle = (001)$\wedge$(*okl*)]

	f (0.2.11)	*g* (012)	*h* (032)	*i* (021)
	° ′	° ′	° ′	° ′
Calculated	16 18	38 48	67 29	72 43
Average of measurements	16 26	38 47	67 32	72 18
Individual measurements	15—16 19	8—38 03	9—67 20	8—72 03
	16—16 25	10—(36 15)	14—67 44	9—72 33
	17—16 33	15—39 23		12—(70 28)
		17—38 44		
		18—38 56		

DESCRIPTION OF FORMS

The combinations observed on the 20 fragments measured are shown in the following table. The fragments 8 and 9, 11 and 12, 13 and 14, and 16, 17, and 18, belonging respectively to the same crystal, are grouped together, the letter of the crystal being repeated for each fragment on which it was determined. Fragments 1 to 12 are simply small cleavage pieces which happened to show faces other than the cleavages. Nothing is known as to the combination of the crystal from which they were cleaved. The measurements on the cleavage fragments had been made before the embedded complete crystals were found. Similarly, fragments 13 and 14 and fragments 16, 17, and 18 are, respectively, only part of the whole crystal. In removing the embedded crystals from the kernite matrix, parts of crystals 13 and 16 remained attached to the matrix, and other small pieces were lost. So the comparison of the relative abundance of the different crystal forms, as given below, is based only on five crystals, Nos. 13, 15, 16, 19, and 20, and is necessarily incomplete.

The two pinacoids *c* and *a*, the clinodome *e*, and the negative unit orthodome *D* are the only forms present on all five crystals. Five other forms occur on four of the crystals—namely, the positive orthodome *d* and the negative orthodomes *B*, *E*, *G*, *H*. A single form, *i* (021), was not found on any of the last five crystals, and three other forms, *h* (032), *k* (203), and *l* (405), were present on only one of the five crystals. All the other forms occurred on two or three of these crystals.

Combinations observed on measured fragments and crystals of kernite

Letter	Symbol	1	2	3	4	5	6	7	8, 9	10	11, 12	13, 14	15	16, 17, 18	19	20
c	001	As cleavage faces										*c c*	*c*	*c c c*	*c*	*c*
a	100											*a a*	*a*	*a a a*	*a*	*a*
b	010	.	.	.	.	.	.	.	*b b*	*b*	. .	. .	.	*b* . *b*	.	.
f	0.2.11	.	.	.	.	.	.	.	. .	.	. .	. .	*f*	*f f* .	.	.
g	012	.	.	.	.	.	.	.	*g* .	*g*	. .	. .	*g*	. *g g*	.	.
e	011	.	.	.	.	.	.	.	*e e*	*e*	*e e*	*e e*	*e*	*e e e*	*e*	*e*
h	032	.	.	.	.	.	.	.	. *h*	.	. .	. *h*	.	. . .	.	.
i	021	.	.	.	.	.	.	.	*i i*	.	. *i*	. .	.	. . .	.	.
j	205	.	.	.	.	.	.	.	. .	.	. .	. .	.	*j j* .	.	*j*
k	203	.	.	.	*k*	.	*k*	.	. .	.	. .	. .	.	. . .	.	*k*
l	405	.	.	.	.	.	.	.	*l* .	.	. .	*l* .	.	. . .	.	.
d	101	.	.	.	*d*	.	.	.	. .	.	. .	. *d*	*d*	*d d* .	.	*d*
n	908	.	.	.	.	.	.	.	. .	.	. .	*n n*	.	. . .	.	*n*
p	403	.	.	.	.	.	.	.	. .	.	. .	*p p*	.	. . .	.	*p*
q	302	.	.	.	.	.	.	.	. .	.	. .	. .	.	*q q q*	.	*q*

Combinations observed on measured fragments and crystals of kernite—Continued

Letter	Symbol	1	2	3	4	5	6	7	8, 9	10	11, 12	13, 14	15	16, 17, 18	19	20
r	704	.	.	.	.	.	.	.	. .	.	. .	*r r*	*r*	. . .	.	.
s	201	.	.	.	.	.	.	.	. .	.	. .	. .	*s*	*s s* .	.	.
t	301	.	.	.	.	.	.	.	. .	.	. .	*t t*	.	*t* . .	.	.
u	401	.	.	.	*u*	.	.	.	. .	.	. .	*u* .	.	. . .	.	*u*
A	$\bar{3}05$	*A*	.	.	.	.	.	*A*	. .	.	. .	*A A*	.	. *A A*	.	.
B	$\bar{5}06$	.	*B*	.	.	.	.	.	. .	.	. .	*B* .	*B*	. . .	*B*	*B*
C	$\bar{6}07$	.	.	.	.	.	.	.	. .	.	. .	. .	*C*	. . .	.	*C*
D	$\bar{1}01$	*D*	*D*	*D*	.	*D*	.	*D*	*D D*	*D*	*D D*	*D D*	*D*	. . *D*	*D*	*D*
E	$\bar{7}06$	*E*	.	.	.	.	*E*	.	. .	.	. .	*E* .	*E*	. . .	*E*	*E*
F	$\bar{6}05$	.	.	.	.	.	.	.	. .	.	. .	. .	*F*	. . .	*F*	*F*
G	$\bar{5}04$	.	*G*	.	*G*	.	.	.	. .	.	. .	. *G*	*G*	. . .	*G*	*G*
H	$\bar{7}05$	.	.	.	.	.	*H*	.	. .	.	. .	. .	*H*	*H* . .	*H*	*H*
J	$\bar{3}02$	.	.	.	.	.	.	.	. .	.	. .	. .	.	. . .	*J*	*J*
K	$\bar{7}04$	.	.	.	.	.	.	.	*K* .	.	. .	. .	*K*	*K* . .	*K*	.
L	$\bar{2}01$	*L*	*L*	*L*	*L*	.	.	.	. *L*	.	. .	. .	*L*	. . .	.	*L*
M	$\bar{7}03$	.	.	.	.	.	.	.	*M* .	.	. .	. .	.	. . .	*M*	*M*
N	$\bar{5}02$	.	.	.	.	.	.	.	. .	.	. .	. .	.	. . .	*N*	*N*
P	$\bar{8}03$	.	.	.	.	.	.	.	. .	.	. .	. .	*P*	. . .	*P*	*P*
Q	$\bar{3}01$	.	.	.	.	.	.	.	. .	.	. .	. .	*Q*	. . .	*Q*	.
R	$\bar{4}01$	*R*	.	.	.	.	*R*	.	. .	.	. .	. .	*R*	. . .	*R*	*R*
S	$\bar{6}01$	.	*S*	.	.	.	.	.	*S* .	.	. .	*S* .	*S*	. . .	.	*S*

The orthopinacoid *a* (100) is probably the largest form on any of the crystals, though *e* (011) and *D* ($\bar{1}01$) are on some crystals nearly as large. It varies considerably in size and on some crystals is little more than a series of repeated broad striae or line faces, forming with the other faces of the orthodome zone a continuous set of rounded faces, heavily striated horizontally. A few of the large faces of *a* are not heavily striated, but usually where there is a set of rounded and striated domes running together the beginning of the striae can be seen on the faces of the orthopinacoid.

The clinopinacoid *b* (010) is a very insignificant form for the crystals measured, although it may possibly reach a large development on some of the large crystals embedded in the clay matrix. On all its observed occurrences it was a line face, between the two faces of the clinodome *e* (011) and gave poor reflections. The measurements are as follows:

Measurements of b (010) kernite

Crystal No.	Measured ϕ (calculated 0° 00′)		Measured ρ (calculated 90° 00′)	
	°	′	°	′
8	−0	35	90	01
9	−2	38	90	00
10	−0	35	90	00
16	+2	11	91	10
18	+0	32	90	18

The basal pinacoid *c* (001) is a medium-sized form, usually several times as long as wide, with the direction of elongation at right angles for the upper and lower faces. It is usually striated parallel to the orthodomes but nevertheless is smooth and brilliant, affording the best reflection of all the forms. The excellent cleavage naturally likewise affords perfect reflections.

The unit clinodome *e* (011) belongs, with *a* (100), to the largest and dominant forms of kernite. If many complete crystals were available for observation, it might be found that *e* (011) was the largest form. The faces of *e* have a tendency to be slightly uneven and to lack the high luster of the natural faces of *a* and *c*. The form is at places slightly striated parallel to its intersection with *c* (001). Rather characteristically, it gives two reflections, with a difference as great as half a degree, one of them generally being better and brighter than the other.

The other clinodomes are represented only by long, narrow faces—broad line faces—giving distinct but poor reflections. The angles measured have already been given. The form system in the zone *c* (001) $\wedge$ *b* (010) is simple, except for the form *f* (0.2.11), which was noted on one crystal (No. 15) and two fragments (Nos. 16 and 17) of another crystal (No. 16). The consistent measured angles are close to the calculated angle and considerably apart from the calculated values for (015) = 17° 49′ ($001 \wedge 0kl$) and for (016) = 15° 00′.

The positive orthodomes are characteristically very narrow faces, practically line faces, giving poor but distinct reflections. Only a very few of the faces are any broader than line faces. As a group they are not as large or common as the negative orthodomes. The measurements of all the faces of the positive orthodomes are presented in the table below, in which the numbers of the crystals are given at the left of the individual measurements.

Measurements of ρ angle of positive orthodomes, kernite

	j (205)		*k* (203)		*l* (405)		*d* (101)	
	°	′	°	′	°	′	°	′
Calculated	39	07	48	26	52	06	56	46
Average of measurements	38	49	47	52	52	01	56	35
Individual measurements	16—39	27	4—48	12	8—51	46	4—(54	39)
	17—38	53	6—47	29	13—52	16	14—57	05
	20—38	07	20—47	54			15—56	28
							15—56	20
							16—56	40
							16—(57	54)
							17—56	24
							20—(58	10)

Measurements of ρ angle of positive orthodomes, kernite—Con.

	n (908)	*p* (403)	*q* (302)	*r* (704)
	° ′	° ′	° ′	° ′
Calculated	59 03	62 24	64 38	67 25
Average of measurements	59 04	61 52	64 20	67 19
Individual measurements	13—59 03 14—59 06 20—59 04	13—61 19 14—62 11 20—61 48 20—62 12	16—63 29 17—63 29 18—65 27 20—64 56	13—67 39 14—67 22 15—66 56

	s (201)	*t* (301)	*u* (401)
	° ′	° ′	° ′
Calculated	69 40	75 32	78 49
Average of measurements	69 45	74 23	77 56
Individual measurements	15—70 28 16—69 46 17—69 02	13—75 06 14—73 55 16—74 07	4—77 21 13—77 57 20—78 29

The difference in prominence between the positive and negative orthodomes can perhaps be shown by the similar difference in the occurrence of the two unit forms. The positive dome *d* (101) is present on but 5 crystals, and only 6 faces were measured, whereas the negative dome *D* ($\bar{1}01$) occurred on 13 crystals and 16 faces were measured. Also the positive section of the orthodome zone did not contain the large number of vicinal forms which are present in the negative section of the zone.

In discussing the symbols of the positive domes, according to the well-known procedure of Goldschmidt in applying the law of complication, the zone, from *c* (001) to *a* (100), yields:

$$\begin{array}{lccccccccccccc} \text{Form} & c & j & k & l & d & n & p & q & r & s & t & u & a \\ \text{Symbol} & 0 & \frac{2}{5} & \frac{2}{3} & \frac{4}{5} & 1 & \frac{9}{8} & \frac{4}{3} & \frac{3}{2} & \frac{7}{4} & 2 & 3 & 4 & \infty \end{array}$$

Splitting the zone at 1 and considering the two parts separately, we have:

$$\begin{array}{lccccc} \text{Form} & c & j & k & l & d \\ \text{Symbol} & 0 & \frac{2}{5} & \frac{2}{3} & \frac{4}{5} & 1 \\ \frac{v}{1-v}= & 0 & \frac{2}{3} & 2 & 4 & \infty \\ \frac{v}{2}= & 0 & \left[\frac{1}{3}\right] & 1 & 2 & \infty \\ N_2= & 0 & \frac{1}{2} & 1 & 2 & \infty \end{array}$$

$$\begin{array}{lccccccccc} & d & n & p & q & r & s & t & u & a \\ & 1 & \frac{9}{8} & \frac{4}{3} & \frac{3}{2} & \frac{7}{4} & 2 & 3 & 4 & \infty \\ v-1= & 0 & \frac{1}{8} & \frac{1}{3} & \frac{1}{2} & \frac{3}{4} & 1 & 2 & 3 & \infty \\ N_3= & 0 & \left(\frac{1}{8}\right) & \frac{1}{3} & \frac{1}{2} & \left[\frac{2}{3}\right] & 1. & 2 & 3 & \infty \end{array}$$

The form *j* should have the symbol $\frac{1}{2}$ = (102) instead of $\frac{2}{5}$ = (205), in order to fit into the series N_2. However, the ρ angle for (102) is 42° 56′, a figure 4° different from the measurements. The zone simply shows slight disturbance caused by the (second) intersecting clinodome zone. In the second part of the zone, the form *n* (908), so close to *d* (101), is obviously extra. However, the measurements are so remarkably consistent and close to the calculated value that the form is accepted, especially as it occurs in addition to *d* (101) and does not simply replace that as a vicinal form. The form *r* gives $\frac{3}{4}$ instead of the $\frac{2}{3}$ of the normal series. This raises the question whether its symbol should not be (503), for which the ρ angle is 66° 32′. Two of the three measurements, however, agree so much better with the indices (704) that they are taken as correct for the form *r*, even though they are obviously more complex.

The only positive dome shown in any of the crystal drawings is *d* (101), which is shown as narrow faces in Figures 38 and 39 and, what is unusual, as a relatively broad face in Figure 35.

The negative orthodomes have, in general, a large effect in determining the general habit of the crystals. The negative dome *D* ($\bar{1}01$), either by itself or in conjunction with a series of forms of fairly complicated indices, makes up a large part of the surface of the crystals. The large face *D* is invariably striated and is very likely to show both convex and concave forms. Especially prone is it to repeat itself all over the crystal as minute striae and to develop a striated zone, which on the goniometer gives a continuous solid band of reflections. On other crystals, however, the set of faces accompanying *D* give sharp and separate reflections, which are easily and accurately measurable. The faces of *D* commonly alternate with those of *a* (100), as shown in Figures 36 and 37. The associated domes of more complex indices may be so abundant, large, and repeated as practically to displace the unit form. (See fig. 40.)

As shown in Figure 40, *D* may be replaced by other domes, here chiefly *P* and *K*, and this repeated alternation of fine line faces will constitute practically the whole front surface of a crystal. A similar occurrence of striated and alternating faces of *D* is shown in Figure 39.

Although most of the negative domes, like the positive ones, are very narrow faces, practically line faces, a few are much larger faces, or rather, perhaps, sets of faces, or sets of repeated alternating striae, such as are shown, for example, in Figure 36, where the large faces of *N* and *H* are sets of striae, most of which belong to *N* and *H*, or in Figure 40, where again the large faces of *K, P, N, H, E* are really a composite set of striae, composed predominantly of faces of the form shown. They will also form a set of fine lines, which give sharp and distinct reflections for each form, but in which the individual forms are hardly recognizable, as shown in Figure 38 for *P, Q, R, S*. On this crystal, for comparison, the faces of *K, H,* and *G* are broader and distinct, not verging nor repeated.

The measurements of the negative orthodomes, except for *D* ($\bar{1}01$), which have already been given, are listed below. The numbers of the crystals are shown at the left of the individual measurements.

Measurements of ρ angle of negative orthodomes, kernite

	A ($\bar{3}05$)	*B* ($\bar{5}06$)	*C* ($\bar{6}07$)	*E* ($\bar{7}06$)
	° ′	° ′	° ′	° ′
Calculated	20 05	32 39	33 46	45 57
Average of measurements	20 08	32 47	33 42	45 52
Individual measurements	1—20 17	2—32 37	15—33 48	1—46 02
	7—20 34	13—32 06	20—33 36	6—46 07
	13—20 30	15—33 19		13—46 35
	14—20 02	19—33 11		15—45 43
	17—19 54	20—32 44		19—45 33
	18—19 32			20—45 14

	F ($\bar{6}05$)	*G* ($\bar{5}04$)	*H* ($\bar{7}05$)	*J* ($\bar{3}02$)
	° ′	° ′	° ′	° ′
Calculated	47 01	48 32	52 37	54 58
Average of measurements	47 10	48 38	52 28	54 49
Individual measurements	15—47 17	2—48 26	6—52 17	19—54 49
	19—47 14	4—48 33	15—52 15	20—54 48
	20—46 58	14—49 14	16—52 33	
		15—48 27	19—52 44	
		19—48 58	20—52 33	
		20—48 48		

Measurements of ρ angle of negative orthodomes, kernite—Contd.

	K ($\bar{7}04$)	*L* ($\bar{2}01$)	*M* ($\bar{7}03$)	*N* ($\bar{5}02$)
	° ′	° ′	° ′	° ′
Calculated	59 51	63 37	67 27	69 00
Average of measurements	59 46	63 27	67 35	69 02
Individual measurements	8—59 41	1—63 04	8—67 55	19—68 53
	15—60 23	2—63 38	19—67 25	20—69 11
	16—58 51	3—63 13	20—67 26	
	19—60 10	4—(64 28)		
		9—63 52		
		15—63 41		
		20—63 14		

	P ($\bar{8}03$)	*Q* ($\bar{3}01$)	*R* ($\bar{4}01$)	*S* ($\bar{6}01$)
	° ′	° ′	° ′	° ′
Calculated	70 21	72 37	77 07	81 33
Average of measurements	70 23	72 29	76 56	81 49
Individual measurements	15—70 12	15—72 32	1—76 24	2—82 24
	19—70 31	19—72 25	6—(78 42)	8—(83 01)
	20—70 27		15—77 29	13—81 25
			19—77 09	15—81 37
			20—76 41	20—(79 45)

The discussion of the negative domes follows, the gnomonic projection showing that a division can also be made at *L* ($\bar{2}01$):

Form	*c*	*A*	*B*	*C*	*D*		*E*	*F*	*G*	*H*	*J*	*K*	*L*		*M*	*N*	*P*	*Q*	*R*	*S*	*a*
Symbol	0	$\frac{3}{5}$	$\frac{5}{6}$	$\frac{6}{7}$	1		$\frac{7}{6}$	$\frac{6}{5}$	$\frac{5}{4}$	$\frac{7}{5}$	$\frac{3}{2}$	$\frac{7}{4}$	2		$\frac{7}{3}$	$\frac{5}{2}$	$\frac{8}{3}$	3	4	6	∞
$\frac{v}{1-v}=$	0	$\frac{3}{2}$	5	6	∞	1	$\frac{7}{6}$	$\frac{6}{5}$	$\frac{5}{4}$	$\frac{7}{5}$	$\frac{3}{2}$	$\frac{7}{4}$	2	2	$\frac{7}{3}$	$\frac{5}{2}$	$\frac{8}{3}$	3	4	6	∞
$\frac{v}{3}=$	0	$\frac{1}{2}$	$\frac{5}{3}$	2	∞	$v-1=0$	$\frac{1}{6}$	$\frac{1}{5}$	$\frac{1}{4}$	$\frac{2}{5}$	$\frac{1}{2}$	$\frac{3}{4}$	1	$v-2=0$	$\frac{1}{3}$	$\frac{1}{2}$	$\frac{2}{3}$	1	2	4	∞
$N_2=$	0	$\frac{1}{2}$	[1]	2	∞	$\frac{v}{1-v}=0$	$\frac{1}{5}$	$\frac{1}{4}$	$\frac{1}{3}$	$\frac{2}{3}$	1	3	∞	$N_3=0$	$\frac{1}{3}$	$\frac{1}{2}$	$\frac{2}{3}$	1 .	2	[3]	∞
						$N_3=0$	$\left(\frac{1}{5}\right)$	$\left(\frac{1}{4}\right)$	$\frac{1}{3}$.	$\frac{2}{3}$	1 ..	3	∞								

The form $B(\bar{5}06)$ should be ($\bar{3}04$), yielding 1 in the normal series N_2. The ρ angle for ($\bar{3}04$) is 28° 29′, 4° different from the measurements. The two forms corresponding to $\frac{1}{5}$ and $\frac{1}{4}$, *E* and *F*, are both extra in the series agreeing with the normal series N_3. *E* was measured six times and *F* three times; the measured angles agree well with the calculated angle. The form *S*, to fit perfectly in the normal series N_3, should give 3 instead of 4. The value of 3 would correspond to a symbol of ($\bar{5}01$), the calculated ρ angle of which is 79° 47′. Although a single face on crystal 20 gave a reading agreeing within 2′ of this value, the face is nevertheless referred to ($\bar{6}01$), as the other measurements of this steepest dome all give higher values, and two of them agree well with the angle for ($\bar{6}01$).

It is difficult to show the extreme value of such a discussion of crystal forms without going into great detail, showing the results first obtained, discussing them, and then by comparison with the normal series trying out the better agreeing figures and thus finally correcting, if necessary, the indices first chosen. It must suffice here to state that the discussion has been, as it always is in crystallographic description, of the greatest value in correctly deciphering the proper indices for the various forms.

VICINAL FORMS

It is with considerable diffidence that a series of vicinal negative domes are here listed. The writer is in general opposed to the listing of vicinal forms with high indices. One of the chief reasons against their presentation here is that so few complete crystals were available that it can not be satisfactorily determined that they really are a salient feature of the crystallography of kernite. Three other considerations, however, have outweighed this objection. The first is that they are present on all the complete crystals measured (Nos. 15, 19, and 20). The second is the consistency of their angular measurements and the close agreement of the measured and calculated values. Compare for example, the set of measurements of the group of the last five vicinal forms, beginning with those for ($\bar{1}7.0.6$). Or compare the two measurements for ($\bar{1}2.0.5$) or the three measurements for ($\bar{1}7.0.11$) or for ($\bar{3}1.0.23$). The third reason for giving them is that they are nearly always present in addition to the form of simpler indices to which they are vicinal. On the goniometer their signals, though faint and described as poor, are as good and definite as the signals for the accepted forms. In fact, during the measurements, no difference was noted between the signals for the faces later found to be vicinal and those which gave relatively simple indices. Moreover, the table of measurements of the negative domes (p. 156) and the list of combinations show that the accepted forms with very simple indices (not higher than the figures of normal series N_3), such as $(\bar{3}02)$ and $(\bar{3}01)$, are much rarer than those of semivicinal indices, such as $(\bar{3}05)$, $(\bar{3}06)$, $(\bar{7}06)$, and $(\bar{5}04)$. Also a number of indices belonging to the normal series N_3, such as $(\bar{1}03)$, $(\bar{1}02)$, and $(\bar{2}03)$, are missing. The greater number of negative domes lie between $D(\bar{1}01)$ and $a(100)$ and similarly, all the vicinal forms except one lie between $D(\bar{1}01)$ and $a(100)$. It does not simplify matters to change the position of the unit dome. Neither was any other orientation found that would simplify the indices, the general morphology of kernite, with the large dominant faces *a D e* and the two excellent cleavages *c* and *a*, being kept in mind.

Should an abundance of complete crystals of kernite be available sometime and carefully measured, a fuller understanding of the complex indices of the vicinal forms might be attained. For the present, the results obtained are presented as found. To one interested in theoretical questions they may serve as the basis of discussion.

The vicinal forms found, with their occurrence and angles, are given here. As in the preceding tables, the numbers of the crystals are shown at the left of the individual measurements.

Measurements of ρ angle of vicinal negative domes, kernite

	$(\bar{21}.0.22)$	$(\bar{21}.0.20)$	$(\bar{14}.0.13)$	$(\bar{11}.0.10)$
	° ′	° ′	° ′	° ′
Calculated	38 05	41 52	42 52	43 41
Average of measurements	38 03	41 49	42 48	43 48
Individual measurements	15—38 03 20—38 03	8—41 56 15—41 39 20—41 58	6—43 13 15—42 47 19—42 44 20—42 26	15—43 42 16—44 17 18—43 43 20—44 00

	$(\bar{9}07)$	$(\bar{21}.0.16)$	$(\bar{31}.0.23)$	$(\bar{17}.0.11)$
	° ′	° ′	° ′	° ′
Calculated	49 34	50 19	51 17	55 57
Average of measurements	49 32	50 29	51 17	56 00
Individual measurements	14—49 14 19—49 46 20—49 36	19—50 34 20—50 23	15—51 20 19—51 17 20—51 15	4—55 57 19—56 08 20—55 55

	$(\bar{13}.0.8)$	$(\bar{27}.0.13)$	$(\bar{13}.0.6)$	$(\bar{12}.0.5)$
	° ′	° ′	° ′	° ′
Calculated	57 34	64 36	65 41	68 06
Average of measurements	57 35	64 38	65 32	68 05
Individual measurements	2—58 07 6—57 48 15—57 15 19—57 32 20—57 11	4—64 28 19—64 36 20—64 49	19—65 39 20—65 24	19—68 08 20—68 02

Measurements of ρ angle of vicinal negative domes, kernite—Con.

	$(\bar{17}.0.6)$	$(\bar{34}.0.11)$	$(\bar{23}.0.7)$
	° ′	° ′	° ′
Calculated	71 33	73 09	74 11
Average of measurements	71 36	73 09	74 10
Individual measurements	2—71 57 15—71 34 18—71 13 19—71 44 20—71 32	15—73 23 19—73 03 20—73 01	19—74 03 20—74 17

	$(\bar{17}.0.5)$	$(\bar{29}.0.8)$	$(\bar{15}.0.4)$
	° ′	° ′	° ′
Calculated	74 45	75 44	76 13
Average of measurements	74 47	75 48	76 18
Individual measurements	1—74 37 8—(75 13) 9—74 52 15—74 43 19—74 43 20—74 55	15—75 52 19—75 42 20—75 50	1—76 24 19—76 17 20—76 13

The three complete crystals measured, Nos. 15, 19, and 20, showed most of these vicinal forms. They were carefully remeasured, with special attention to the many reflections yielded by the faces in the orthodome zone. Although parts of this zone were so rounded as to give simply a continuous band of reflected light, most of the zone gave poor but very distinct single reflections. These were abundant and close together and yet sufficiently distinct to give a single undistorted reflection for each one measured. In order to show what the measurements yielded, those for these three complete crystals are given below in their entirety. Only those readings that were similar for at least two crystals are considered at all. Where but a single angular value was obtained, it is given as a matter of record, but no attempt at interpretation is offered. All these measurements may become of importance when a large suite of kernite crystals are available for accurate measurements.

Measurements of ρ angles of all negative orthodomes on kernite crystals Nos. 15, 19, and 20

[Vicinal forms not lettered]

	No. 15		No. 19		No. 20		Occurrences on other crystals
	°	′	°	′	°	′	
B $(\bar{5}06)$	33	19	33	11	32	44	Nos. 2, 13.
C $(\bar{6}07)$	33	48	----		33	36	
					34	31	
	35	05					
					37	27	
$(\bar{21}.0.22)$	38	03	----		38	03	
					39	26	
D $(\bar{1}01)$	39	56	39	52	39	52	
	40	40					
$(\bar{21}.0.20)$	41	39	----		41	58	No. 8.
$(\bar{14}.0.13)$	42	47	42	44	42	26	No. 6.
$(\bar{12}.0.10)$	43	42	----		44	00	Nos. 16, 18.
	44	33					
E $(\bar{7}06)$	45	43	45	33	45	14	Nos. 1, 6, 13.
	46	23					
	46	33					
F $(\bar{6}05)$	47	17	47	14	46	58	

Measurements of ρ angles of all negative orthodomes on kernite crystals Nos. 15, 19, and 20—Continued

[Vicinal forms not lettered]

	No. 15		No. 19		No. 20		Occurrences on other crystals
	°	′	°	′	°	′	
G $(\bar{5}04)$	48	27	48	58	48	48	Nos. 2, 4, 14.
$(\bar{9}07)$	-----		49	46	49	36	No. 14.
$(\overline{21}.0.16)$	-----		50	34	50	23	
$(\overline{31}.0.23)$	51	20	51	17	51	15	
H $(\bar{7}05)$	52	15	52	44	52	33	Nos. 6, 16.
J $(\bar{3}02)$	-----		54	49	54	48	
$(\overline{17}.0.11)$	-----		56	08	55	55	No. 4.
$(\overline{13}.0.8)$	57	15	57	32	57	11	Nos. 2, 6.
K $(\bar{7}04)$	60	23	60	10	-----		Nos. 8, 16.
					61	06	
L $(\bar{2}01)$	63	41	-----		63	14	Nos. 1, 2, 3, 4, 9.
					64	18	
$(\overline{27}.0.13)$	-----		64	36	64	49	
			65	02			
$(\overline{13}.0.6)$	-----		65	39	65	24	
					66	18	
M $(\bar{7}03)$	-----		67	25	67	26	No. 8.
$(\overline{12}.0.5)$	-----		68	08	68	02	
					68	46	
N $(\bar{5}02)$	-----		68	53	69	11	
			69	24			
P $(\bar{8}03)$	70	12	70	31	70	27	
$(\overline{17}.0.6)$	71	34	71	44	71	32	Nos. 2, 18.
Q $(\bar{3}01)$	72	32	72	25	-----		
$(\overline{34}.0.11)$	73	23	73	03	73	01	
					73	47	
$(\overline{23}.0.7)$	-----		74	03	74	17	
$(\overline{17}.0.5)$	74	43	74	43	74	55	Nos. 1, 8, 9.
$(\overline{29}.0.8)$	75	52	75	42	75	50	
					76	13	
			76	44			
R $(\bar{4}01)$	77	29	77	09	76	41	Nos. 1, 6.
					77	51	
					79	17	
	80	10					
S $(\bar{6}01)$	81	37	-----		79	45	Nos. 2, 8, 13.

DESCRIPTION OF CRYSTALS

All the measured crystals have essentially the same habit, varying only in detail on account of the large development of one crystal form at the expense of other forms. The crystals range from equidimensional ones to those which are slightly elongated in the vertical direction. The large development of the dome faces gives some of the crystals a wedge-shaped appearance. Rarely a set of repeated faces furnishes a rounded striated surface of relatively large extent which causes the crystal to be somewhat flattened.

The simple combinations *c*, *a*, *e*; *c*, *D*, *e*; *c*, *a*, *D*, *e*, which furnish the key habit of the kernite crystals and which the large crystals in the mine commonly show, are given in Figures 29, 30, and 31. In Figures 35 to 40 the attempt has been made to illustrate the general appearance of the crystals, giving only the dominant forms with but a few modifications, rather than to portray accurately the true appearance of the crystal, with all its faces and striations and distortions. Narrow striae of some of the forms are repeated on the crystals many more times than are shown in the drawings. Many rounded concave and convex surfaces, covered by such striated faces, are present, but only a few are shown. The two sides of the same crystal are not developed with the same forms or with equal development of the face of the same form, so that the appearance of the two sides of the same crystal may show considerable difference. The drawings are in part idealized, together with such distortion as seems to be characteristic of the kernite crystals.

In Figures 29 to 32 and 35 to 40 the guide line and angle point *W* have been somewhat shifted from the conventional standard position. In the standard position the face *e′* $(0\bar{1}1)$ lies back of the guide line and would not show in the upper portion of the clinographic projection. The guide line consequently has been shifted up on the left side of the gnomonic projection a distance equal to that of *e′* $(0\bar{1}1)$ from the standard position, so that *e′* $(0\bar{1}1)$ lies midway between the standard and the shifted guide line. Moreover, the guide line has been shifted down on the right side of the projection so that it passes through the center of the projection. Several trials were made with the guide line in various positions, and the one finally adopted seems to illustrate best the appearance of the crystals in clinographic projection.

The more complicated crystals are developed as a crystallographic composite from the simple types and combinations shown in Figures 29, 30, and 31. Thus Figure 35, of which crystal 16 is an example, is a distorted modification of Figure 29. The top base is much broader than the bottom base, a feature not rare for kernite, and the direction of greatest length lies at right angles for the two faces of this form. Crystal 16 shows a broad face of *d* (101) and the largest face of *b* (010) observed on all the crystals. Figure 36 is a modification of the simple type of combination *c*, *a*, *e* (shown in fig. 29), where the front face of *a* (100) contains many occurrences of the negative domes, appearing repeatedly. In the center is a large concave area formed chiefly of *N* $(\bar{5}02)$ but containing additional negative domes not shown in the drawing. One side of crystal 20 shows such a development. The other side of crystal 20 (fig. 37) shows a similar development of forms, except that here the dome *D* $(\bar{1}01)$ is the prominent form, *a* (100) being very much diminished in importance, though the essential forms are *D* and *a*, with the negative domes, except the unit one, of very minor importance. This side of crystal 20 therefore represents a modification of the simpler type *c*, *D*, *e*, shown in Figure 30.

A modification of the simple type *c*, *a*, *D*, *e* (fig. 31) is represented by crystal 15, the two sides of which are shown in Figures 38 and 39. In Figure 38 the two faces of *e* (011) are drawn in symmetrical development. It is rather rare to find them so nearly equal in size. In Figure 39, showing the other side of crystal 15, the two faces of *e* (011) are drawn unequally developed, a condition that is rather common. In consequence the two drawings do not accurately portray crystal 15 but are so drawn as to serve as types, illustrating the various modifications and distortions observed in the kernite crystals. Figure 38 shows the usual size of the faces of d (101), as does also Figure 39, and in $(\bar{5}04)$, H $(\bar{7}05)$, and K $(\bar{7}04)$. Figure 39 has the simple form D $(\bar{1}01)$ replaced in its upper portion by repeated alternations of striated faces of L $(\bar{2}01)$ and D $(\bar{1}01)$, together with other forms, and also shows broader faces of E $(\bar{7}06)$, F $(\bar{6}05)$, G $(\bar{5}04)$, H $(\bar{7}05)$, and J $(\bar{3}02)$.

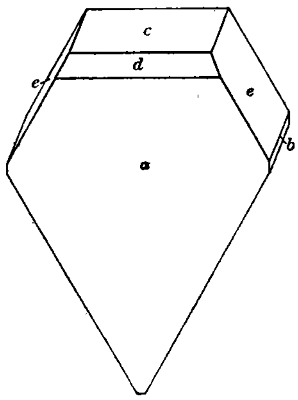


FIGURE 35.—Kernite crystal 16. Modification of the simple type c, a, e, with unusually broad face of d (101) and the largest face of b (010) observed. The two faces of c (001) are unequally developed. The actual crystal is about 5 millimeters high

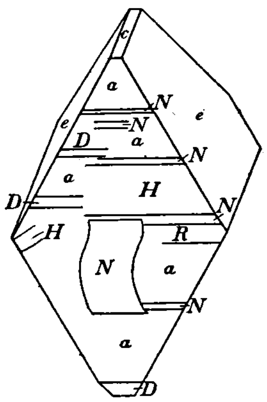


FIGURE 36.—One side of kernite crystal 20. Another modification of the simple type c, a, e, with many negative domes appearing repeatedly as broad and narrow striations and with a large concave face of N $(\bar{5}02)$ accompanied by other negative domes not shown in the drawing. H $(\bar{7}05)$, N $(\bar{5}02)$, R $(\bar{4}01)$

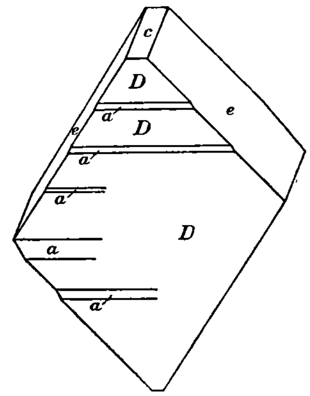


FIGURE 37.—The other side of kernite crystal 20, showing a modification of the type c, D, e, with alternations of a (100) and many negative domes, most of which occur as narrow faces on striations and are not shown in the drawing. Crystal 20, the two sides of which are shown in a simplified manner in Figures 36 and 37, is about 3 millimeters high

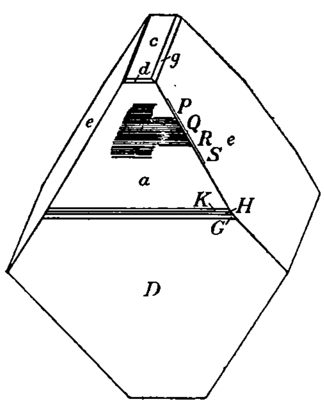


FIGURE 38.—One side of kernite crystal 15, showing the general type and only the more prominent negative domes P $(\bar{8}03)$, Q $(\bar{3}01)$, R $(\bar{4}01)$, S $(\bar{6}01)$, K $(\bar{7}04)$, H $(\bar{7}05)$, G $(\bar{5}04)$. Other forms are d (101), g (012). The two faces of e (011) are drawn in ideal symmetry, a condition not often encountered

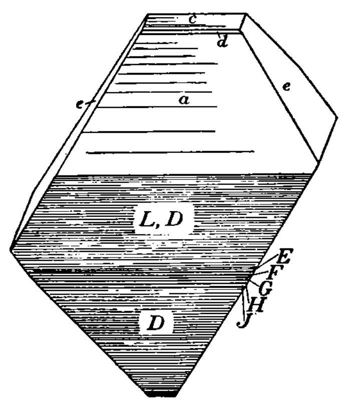


FIGURE 39.—The other side of kernite crystal 15, with the two faces of e (011) in unequal development, a more common condition than the symmetrical arrangement of Figure 38. The two drawings represent types of form development rather than a strict copy of the crystal measured. The unit positive dome d (101) is shown, with a striated orthopinacoid. The lower portion is a repeated combination of D $(\bar{1}01)$ and L $(\bar{2}01)$, with the broader faces of E $(\bar{7}06)$, F $(\bar{6}05)$, G $(\bar{5}04)$, H $(\bar{7}05)$, and J $(\bar{3}02)$. The lower portion is again essentially D $(\bar{1}01)$. Both these figures again show the change in position of greatest elongation of the two faces of the basal pinacoid c (001). Crystal 15 is about 2 millimeters high

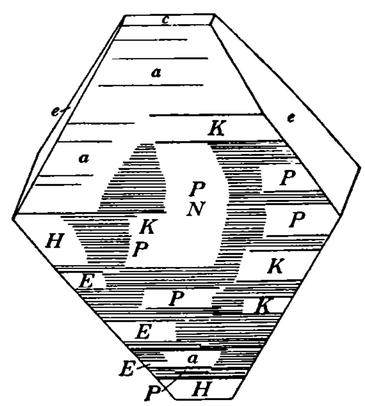


FIGURE 40.—One side of kernite crystal 19, with relatively large faces of K $(\bar{7}04)$, P $(\bar{8}03)$, N $(\bar{5}02)$, H $(\bar{7}05)$, and E $(\bar{7}06)$. Although these faces are not striated in the drawing, actually they are finely striated and merge into the more coarsely striated parts, as shown in the drawing, where the individual faces can not be identified. Crystal 19 is about 2½ millimeters high

addition the average size of the clinodomes other than e (011), here represented by g (012). In the center of a (100) is a concave striated hollow, bounded by faces of P $(\bar{8}03)$, Q $(\bar{3}01)$, R $(\bar{4}01)$, S $(\bar{6}01)$, often repeated. The center of the crystal shows the three faces of G The true representation of these repeated and striated faces would entail much more work than is justified.

Crystal 19 is represented by Figure 40, where the upper half of the front surface is composed largely of the orthopinacoid a (100), with small repeated striae of some of the negative domes. The lower portion is a heavily striated composite of negative domes, some of which are represented by relatively broad faces.

TWINNING

None of the complete crystals measured showed any twinning, but on one of the larger compact masses there were groups of built-up individuals at about 60° that strongly suggested twinning. A large group of kernite, about 2 feet long, collected for exhibition in the National Museum, shows several such twin structures, and measurements of their angle and etching of the faces show that they are twinned with *e* (011) as the twinning plane. Measurement of their relative position gave a value of 61°, corresponding to the angle *c* (001) : *e* (011) of 58° 07′. The two faces of *a* (100) remain in the same plane; the two

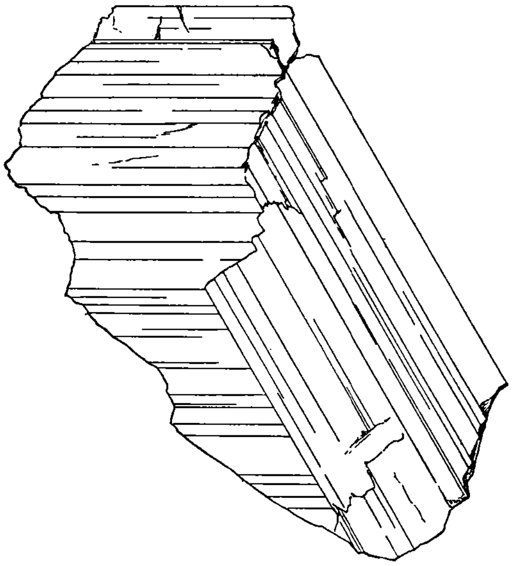


FIGURE 41.—Sketch of twin of kernite

faces of *c* (001) lie in the clinodome zone of the untwinned crystal. Figure 41 is a sketch of the appearance of such a twin grouping.

ETCHING EFFECTS

The two perfect cleavages of kernite easily develop characteristic etch figures, which show that the cleavages do not belong to the same crystal form. The etchings also serve admirably for the identification of the cleavages. These etchings were called to the writer's attention by Dr. C. Hlawatsch, of Vienna, who said that they showed that kernite could not be orthorhombic in symmetry, as stated in the preliminary report.

After several modifications of treatment had been tried out, it was found that excellent etchings could be obtained if a cleavage fragment of kernite was immersed in hot water for about 10 seconds. Longer treatment covers the entire surface with etchings which run together and are not as distinct. In fact, a cleavage fragment of kernite appears to become dull white when left in water, either hot or cold, for some time, owing to the multiplicity of the etchings. Hot water is the most practicable liquid to use if the etchings are developed for purposes of orientation of the cleavage fragment and are desired quickly. Cold water gives just as good results, but the etchings are developed much more slowly. When immersed in cold water, the etch figures are large enough to differentiate in about 5 minutes, and a 10-minute immersion suffices to develop them to a satisfactory size, so that their shape and symmetry can be seen.

In general, the etchings on the basal pinacoid *c* (001) are square in outline, and those on the orthopinacoid *a* (100) are triangular, pointing upward. The basal pinacoidal cleavage also shows the fine parting lines, normal to the edge (001) $\wedge$ (100), as described in connection with cleavage. Although groups of etchings on both faces can be seen to run in all directions,

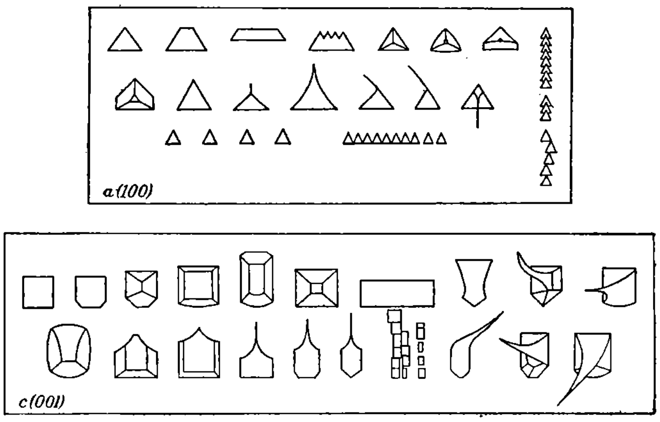


FIGURE 42.—Etch figures artificially developed on the cleavage surfaces of *c* (001) and *a* (100) of kernite, by immersion in hot water for 10 seconds. On the left side are shown the more simple types

there is a tendency for those on the basal pinacoid to form chains normal to the edge (001) $\wedge$ (100), whereas the chains of etchings developed on the front pinacoid *a* (100) have a general tendency to run parallel to the intersection edge (001) $\wedge$ (100).

All the etch figures, free from beaks, have a plane of symmetry parallel to *b* (010) and clearly indicate monoclinic symmetry. Many of the square etchings on the base *c* (001) have their base more modified than their top, so that there clearly is no horizontal plane of symmetry to them. The various types observed, which were not specially studied, are shown in Figure 42.

Not a few of the etch figures have developed considerable beaks, which, though in general vertical, run in all directions, as shown particularly on the right side of the lower sketch of Figure 42. The writer had no opportunity of making any detailed study of these etchings, but the ease with which the etchings can be obtained renders kernite excellent material for such a study.

If a very dilute acid is used, or even if a drop or two of dilute acid is added to the water in which the kernite is immersed, the shape of the etching on c (001) is changed from a square outline to long, thin ovals, of a not very definite shape, with rather characteristic small and very short beaks growing out of the middle on each of the sides. The general appearance of these etch figures as developed by dilute HCl is shown in Figure 43. With very dilute H_2SO_4, similar figures were obtained, but the ovals were somewhat broader.

The very dilute acid solution produces triangular etchings on a (100) similar to those produced by water, except that the sides of the triangle are rounded instead of being straight. Horizontal chains are very characteristic. The appearance of these etchings is shown in Figure 43.

Etch figures are just as readily obtained on the other surfaces of kernite. Cleavage plates after D ($\bar{1}01$), a third cleavage of fair quality, on immersion in hot water for 10 seconds, develop etch figures resembling a flaring gas jet. The surface of this

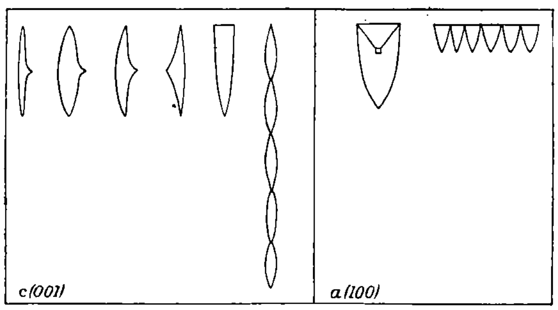


FIGURE 43.—Etch figures artificially developed on cleavage faces of kernite by immersion in very dilute acid

cleavage is not as plane as those of the cleavages a (100) and c (001), and the etchings are not as distinct. They also clearly show only the vertical plane of symmetry, parallel to b (010). The general appearance of these etchings is shown in Figure 44.

On fracture surfaces parallel to the clinopinacoid b (010) hot water produces etch figures similar to those shown in Figure 44. This surface, being parallel to the plane of symmetry, develops etchings of a different symmetry than the cleavages in the orthodome zone (normal to the clinopinacoid), but their symmetry belongs to the monoclinic system. All the observed etchings indicate monoclinic symmetry for kernite, and nothing was seen to suggest definitely any deviation from holohedral symmetry.

Kernite contains many planes, both straight and curved, in which lie an abundance of liquid inclusions and negative crystals. Some of these are sharply angular, and others appear considerably rounded. The general direction of elongation of the negative crystals follows the b axis, but some are at right angles to it, and others extend in diverse directions.

CHEMICAL PROPERTIES

PYROGNOSTICS

Kernite readily fuses before the blowpipe flame, with expansion, to an opaque white cauliflower mass. On further heating, it fuses to a clear glass (borax glass). When heated in a closed tube, it expands, curls up, cleaving into numerous fibers, and expands considerably more to a large white mass, finally condensing and fusing to clear borax glass. It is slowly soluble in cold water, a clear piece appearing dull white in a short time, owing to the formation of innumerable etch figures on the surface. A piece 1 centimeter long and half as thick will not be completely dissolved on standing in cold water over night. Kernite is readily soluble in hot water and in acids. The solution, on evaporation of the water, yields 1.39 times as much borax.

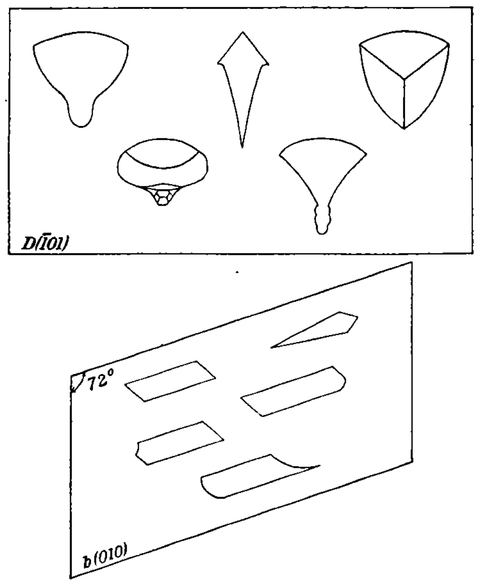


FIGURE 44.—Etch figures on kernite artificially developed on the cleavage surface parallel to D ($\bar{1}01$) and on the clinopinacoid b (010). Produced by immersion in hot water for 10 seconds

ANALYSIS

The chemical composition is as follows:

Analysis of kernite

[W. T. Schaller, analyst]

	Analysis	Calculated	Artificial
Na_2O	22. 63	22. 66	22. 65
B_2O_3	50. 76	51. 02	[a] 50. 80
H_2O	26. 50	26. 32	26. 55
	99. 89	100. 00	100. 00

[a] By difference.

The ratios of the analysis of kernite are $Na_2O : B_2O_3 : H_2O = 365 : 729 : 1{,}472$, or $1.00 : 1.99 : 4.02$, giving the formula $Na_2O.2B_2O_3.4H_2O$, or $H_8Na_2B_4O_{11}$. Nearly half the total water is given off up to 110°, and only three-quarters of the total water is lost up to 200°, there being no swelling of the mineral at this temperature. The water lost at different temperatures is:

Water lost on heating kernite

	Per cent	Per cent of total
Up to 110°	10. 88	41
From 110° to 200°	9. 08	34
At ignition (above 200°)	6. 45	25
	26. 41	100

Of the water lost at 110°, 82 per cent was reabsorbed in balance case in 1½ days.

ALTERATION

Kernite that is free from any other borate (borax, for example) and that has been kept in the laboratory for nearly two years, exposed to the air, is perfectly stable, no cloudiness being developed. But if any borax (and possibly other borates) is attached to the kernite, it partly alters to opaque white tincalconite in a few months. The progress of the alteration can readily be observed in crushed fragments under the microscope. The tincalconite coats the kernite and also develops in cleavage cracks, gradually replacing the kernite. Several clear pieces of kernite show the development of small rounded areas of tincalconite along a cleavage plane in the interior of the kernite. The change of kernite ($Na_2O.2B_2O_3.4H_2O$) to tincalconite ($Na_2O.2B_2O_3.5H_2O$) consists chemically simply in the addition of one molecule of water. This change, however, is mechanically very destructive, for a large specimen of kernite, 2 feet long, fresh and solid when collected, is in less than a year splitting to pieces, owing to the development of tincalconite along its cleavage planes.

SYNTHESIS

Many futile attempts were made to prepare kernite artificially from borax. Fusions of borax in open and closed tubes, with the expulsion of part of the water, invariably yielded only the 5-hydrate, artificial tincalconite. Crystallization in the presence of various other salts, of which a long list was tried, yielded no success. Very gradual evaporation of a solution of borax near 100° yielded only fine crystals of the 5-hydrate. R. C. Wells, of the Geological Survey laboratory, became interested in the problem and finally succeeded in making a preparation from borax that yielded almost pure artificial kernite, only a few crystals of artificial tincalconite being associated with it.

The method used was to heat borax in a closed tube to about 150° for several days in an electric furnace, the outside end of the tube being bent down so that the water driven off from the borax was condensed and collected here and did not return to the heated borax. Under these conditions, the heating was continued for several days; the furnace was cooled, and then the water driven off was allowed to flow back onto the heated borax, the furnace reheated, and the tube again introduced and kept at 150° for several days. When cooled and examined, the resultant product was all crystalline and consisted essentially of small kernite crystals with only a very small quantity of the 5-hydrate. The optics and the chemical composition of the artificial product, as already given, agree with that of kernite.

The small crystals, averaging between a fifth and a tenth of a millimeter in size, have the appearance shown in Figure 45, being considerably more elongated in the direction of the *b* axis than the natural crystals. The faces of *a* (100) and *c* (001) are the dominant ones, with those of *e* (011) the only essential terminal ones. Faces of *d* (101), *D* ($\bar{1}01$), and *f* (0.2.11) are present

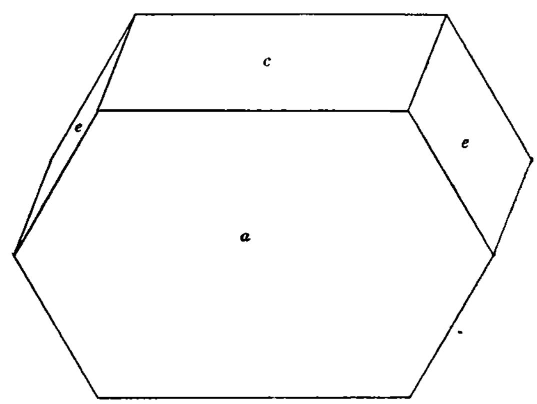


FIGURE 45.—Crystal of artificial kernite, about a fifth of a millimeter long

as line faces, and broader faces of the pyramid (322), between *a* and *e*, were noted on several crystals. It is curious that the artificial crystals should show pyramid faces, totally lacking on the natural crystals. The measurements on the artificial crystals yielded the following angles:

Measurements on artificial crystals of kernite

Angle	Measured		Calculated	
	°	′	°	′
c (001)∧*a* (100)	71	15	71	08
c (001)∧*e* (011)	57	57	58	07
c (001)∧*f* (0.2.11)	16	30	16	18
a (100)∧*e* (011)	80	16	80	10
c (001)∧*d* (101)	37	40	37	48
c (001)∧*D* ($\bar{1}01$)	58	49	58	48
a (100)∧ (322)	43	01	43	04

An earlier experiment, made by Mr. Wells under similar conditions, except that the water driven off was not allowed to run back on the heated borax, yielded a crystalline product whose analysis showed it to be the 3-hydrate.

Analysis of artificial 3-hydrate

[W. T. Schaller, analyst]

	Analysis	Ratio	Calculated
Na_2O	23.16	0.95 or 1	24.88
B_2O_3	[a] 55.64	2.03 or 2	58.86
H_2O	21.20	3.03 or 3	19.26
	100.00		100.00

[a] By difference.

The sample analyzed was not completely crystallized, small particles of an amorphous material being intermingled. The optical properties (both β and γ being higher than 1.50) showed that no kernite or higher hydrate was present.

The bearing of the success attained in making artificial kernite is discussed under heading "Origin of deposit."

TINCALCONITE

OCCURRENCE

The name tincalconite was given by Shephard [10] to a pulverulent and efflorescent borax from California, containing 32 per cent of water. Although no further description is given, it is obvious that his material was the 5-hydrate of the borax series, $Na_2O.2B_2O_3.5H_2O$ identical in composition with "octahedral borax." A sample of Shephard's original tincalconite, now in the United States National Museum, is identical with the tincalconite from the Kramer district.

Tincalconite was found, not very abundant but rather widespread, coating both borax and kernite. As it forms readily from the 10-hydrate, or borax, by partial dehydration, it probably occurs at many places in the Mohave Desert of southern California and can probably be found as a coating on many of the borax deposits of the desert region. Microscopic examination of these films and included cleavage fragments of kernite definitely shows the order of formation, the tincalconite being later than and derived from the kernite. Tincalconite derived from and coating borax did not seem to be nearly as abundant in the mine, although the dull-white film was seen at a few places in the mine coating some of the massive borax. In the Washington laboratory the samples of borax quickly dehydrate to tincalconite, the glassy borax becoming covered with a dull-white layer in a few months. The appearance of tincalconite as it develops naturally in the Washington laboratory from borax is well illustrated in Plate 25, *B*. The kernite, too, if any borax is present, readily alters to tincalconite in the laboratory.

Tincalconite is the natural equivalent of what has long been known as "octahedral borax." This 5-hydrate crystallizes in the hexagonal-rhombohedral system, but the combination of six rhombohedra and two bases greatly resembles an octahedron, and hence this compound has been (wrongly) called octahedral borax. The *c* axis was determined by Arzrumi on artificial crystals to be 1.87. The forms are *c* (0001), *r* $(10\bar{1}1)$, *d* $(01\bar{1}2)$. No natural crystals of tincalconite were observed.

[10] Shephard, C. U., Soc. Min. Bull., vol. 1, p. 144, 1878.

PHYSICAL AND OPTICAL PROPERTIES

The natural material was very unsuitable for determination of its properties, as it occurs only as a fine-grained crystalline powder. In appearance it is dull white, although artificial crystals are vitreous in luster, colorless, and transparent. The artificial crystals show no cleavage on breaking but a hackly fracture with an occasionally poorly developed conchoidal surface. The specific gravity of the artificial crystals was determined as 1.880. This is considerably higher than that given in the literature (1.815), which was determined by Payen [11] over a hundred years ago!

The artificial crystals of tincalconite are uniaxial positive, $\omega = 1.461$, $\epsilon = 1.474$, $B = 0.013$. These values are the average of three determinations as follows:

Indices of artificial tincalconite

	Schaller		Ross	Average
	1927	1928		
ω	1.460	1.462	1.461	1.461
ϵ	1.473	1.476	1.474	1.474

The mean index $\frac{(2\omega + \epsilon)}{3}$ is 1.465. The natural mineral is so fine grained that accurate determination of refractive indices is difficult. The mean index was measured on one sample as 1.463, and Shephard's original material gave the same value.

CHEMICAL COMPOSITION

PYROGNOSTICS

The pyrognostics of artificial tincalconite are similar to those of other members of the borax group.

ANALYSIS

A number of samples of tincalconite were analyzed. Of these, only the two (Nos. 1 and 2) that coat kernite were collected as such in the mine. The other three samples have developed in the Washington laboratory on borax.

Samples 1 and 2 coat kernite and were present on the kernite when collected, though it is possible that the alteration of kernite to tincalconite has continued after the specimens were brought to Washington. Sample 1 is a fine-grained crystalline mass, and the powder analyzed contained a very few minute fibers of kernite. Sample 2 was prepared from a single

[11] Payen, A., Jour. chimie méd., vol. 3, p. 594, 1827.

cleavage piece of what was originally all kernite but had changed to tincalconite, a true pseudomorph, as the cleavage shape of the kernite was retained. This piece was crushed, and as much kernite as possible was removed, the remaining sample being the typical fine-grained crystalline powder so characteristic of this pseudomorphous tincalconite. Many of the minute cleavage fibers of kernite have been completely changed to tincalconite. The total quantity of kernite remaining in the sample analyzed was small, the impurities in the sample being determined as consisting of 1.32 per cent (of the original sample) of clay (ignited) and 2.11 per cent of kernite (not ignited). In the table below this analysis is given as made (2a) and repeated with the clay and kernite deducted (2b).

Sample 3 was a fine-grained crystalline mass of tincalconite coating massive borax and developed after the samples reached Washington. A very few small pieces of borax were contained in the sample analyzed.

Sample 4 was collected from seams in the clay matrix, associated with a little kernite. The seams, when collected, were probably borax, which changed to tincalconite in the Washington laboratory. The associated clay (insoluble in water) gave a loss on ignition of 11.85 per cent. The sample analyzed was very finely crystalline, with a few pieces of borax and a very few fibers of kernite. These associated minerals stand out strikingly in a slide of the analyzed powder of tincalconite, as the individual crystal units of kernite and of borax are very large in comparison to those of the fine-grained tincalconite. Several fibers of kernite completely altered to tincalconite are present. Much of the tincalconite is so fine grained as to appear almost opaque under low magnification.

Sample 5 is a borax crystal (natural or artificial?) from Searles Lake, Calif., naturally dehydrated in the Washington laboratory.

The samples analyzed gave no reaction for any appreciable lime, magnesia, sulphate, chloride, or carbonate.

Analyses of tincalconite from California

[W. T. Schaller, analyst]

	1	2a	2b	3	4	5	Calculated
Insoluble in cold water	0.48	3.43	-------	0.37	0.90	0.25	-------
Na_2O	20.80	19.97	20.67	21.28	20.72	21.40	21.29
B_2O_3	47.52	47.00	48.67	47.26	47.19	47.26	47.80
H_2O (total)	30.59	29.75	30.81	30.78	30.93	31.01	30.91
	99.39	100.15	100.15	99.69	99.74	99.92	100.00

As can readily be seen, the analyses of tincalconite agree closely with the calculated composition. The ratios are as follows:

Ratios determined from analyses of tincalconite

	1	2b	3	4	5
Na_2O	0.99	0.97	1.01	0.99	1.01
B_2O_3	2.01	2.03	1.99	2.01	1.99
H_2O	5.01	4.98	5.02	5.09	5.05

The analyses and the resultant ratios clearly show that the formula of tincalconite is $Na_2O.2B_2O_3.5H_2O$.

The loss of water on a sample of artificial tincalconite is as follows:

Water lost on heating artificial tincalconite

	Per cent	Per cent of total
Up to 110°	16.03	50
From 110° to 200°	10.82	34
At ignition	5.31	16
	32.16	100

There was no reabsorption of the water lost up to 110° on letting the sample stand in the balance case over night.

SYNTHESIS

Tincalconite can readily be made by boiling a solution of borax until crystallization ensues or by letting a hot strong solution cool to not below 60°. Crystals of tincalconite also readily form by melting borax in its water of crystallization, in either a closed or an open tube. Tincalconite represents the stable form of $Na_2O.2B_2O_3.nH_2O$ above 60° or above about 35° in saturated salt solutions.

BORAX

The borax of the deposit does not show many features of strictly mineralogic interest. It is, however, of the greatest importance in considering the origin of the deposit of borate minerals.

Most of the borax associated with the kernite is massive, showing no crystal outline. (See pl. 26, *A*.) Several bodies of such massive borax, as large as an ordinary room, or even larger, were encountered. In appearance the massive borax resembles some massive greasy quartz or even more the massive compact elaeolite.

Included in the massive borax are a few imperfect and distorted crystals of kernite, spherulitic groups of kramerite, and sharply angular fragments of the clay matrix abundantly intersected and cut by minute veins of borax, which either run across the clay fragments or only partly cut into them. One specimen contained minute shredlike realgar, giving the borax specimen an orange-colored mottled appearance. More rarely the borax is developed as porphyritic crystals, a centimeter or two long, in the clay matrix, as shown in Plate 26, *B*. The borax readily dehydrates to white opaque tincalconite in the laboratory.

The optical properties of this borax, so far as determined, show no difference from those already determined. The refractive indices of borax given in the literature show close agreement, as follows:

Refractive indices of borax

	α	β	γ
Dufet	1.4467	1.4694	1.4724
Descloizeaux	1.447	1.470	1.473
Tschermak	1.4468	1.4686	1.4715
Kohlrausch	1.4463	1.4682	1.4712
Average	1.4467 1.447	1.4691 1.469	1.4720 1.472

The borax from the Kramer district gave similar indices ($\alpha=1.446$, $\beta=1.468$, $\gamma=1.472$). A rather striking diagnostic property of borax, as observed under the microscope, is that sections normal to an optic axis pass through characteristic pinchbeck-brown to blue-gray colors instead of extinguishing.

ASSOCIATED MINERALS

The minerals, aside from other borates, associated with the borates above described are very few in number. They are, as determined, the clay minerals, calcite, realgar, and stibnite. The clay minerals comprise those of the clay itself and the associated rock minerals resulting from the disintegration of various rocks. The clay matrix, as the host of the kernite has been called, is a complex unit, of diverse origins. There are streaks and bands of hard clay in the softer material (such as are shown in pl. 25, *A*) and coarse and fine material, all aggregated together into the clay matrix. Several thin sections of different varieties of the clay matrix suggest that most of the clay mineral is beidellite.[12] One thin section shows uniform areas of fine-grained material, mixed with agglomeratic material, some of which suggest original porphyritic lava, with well-defined flow lines, and others suggest débris from granitic rocks, as for example the fragments of microcline and black tourmaline. Feldspars and micas, are abundant in parts of the clay. Considerable opaque black material (carbonaceous?) is irregularly scattered throughout the clay, as is also more or less calcite.

The clay carrying the seams of ulexite (pl. 22, *A*) lying above the kernite is far more homogeneous. It is much finer grained and resembles beidellite. The matrix of the colemanite-ulexite specimen (pl. 22, *B*) obtained above the kernite is coarser grained and resembles the clay of the kernite matrix. In spots there is abundant biotite, somewhat altered. Some of the clay resembles in structure the basalt lying below the colemanite-ulexite in the Suckow shafts, several miles to the west, especially in what seems to be a retention of the trachytic structure.

The calcite occurs in small aggregates of rhombohedral crystals and also as small masses of no definite shape scattered through the coarser clay, especially where this has the appearance of having been drived from an altered volcanic rock. The realgar was observed inclosed in one specimen of borax as fine shreds, giving the borax a mottled orange-colored appearance. A mass of cleavable realgar with a coarse radiating columnar structure, about half a centimeter across, was also observed in the clay associated with the black-appearing kernite. When heated in a glass tube, the realgar readily sublimed and partly oxidized to white arsenious oxide, which condensed as transparent, colorless, isotropic octahedra. Stibnite occurs characteristically as small spherulites, 1 millimeter or less in diameter, in the clay and in the borate minerals. It is present in greatest quantity in the specimens showing the radiating kramerite, and although in places it is concentrated so that a square centimeter of surface may show two or three spherulites, its total quantity in the deposit is very small. Most specimens show no sulphide mineral whatever.

ORIGIN OF DEPOSIT

In discussing the origin of the borate minerals of the Kramer district, one is greatly handicapped by the lack of quantitative knowledge of the geologic conditions. These have been described for the deposit just west of the kernite by Noble. For the Suckow shaft No. 2, in sec. 22, and probably the surrounding ones, the geologic formations comprise essentially an overlying alluvium (Quaternary), about 100 feet thick, which rests unconformably on the eroded surface of the upturned edges of the Tertiary sedimentary rocks. These, beginning at the top, consist of 75 feet of pale-greenish clay shale, somewhat sandy and containing volcanic ash, followed by 40 feet of an arkose sandstone with volcanic ash, which in turn is underlain by about 60 feet of the borate-bearing clay shale. Below this is lava, at least 20 feet thick. The borate-bearing clay shale with nodules of colemanite irregularly distributed through it, is everywhere mashed,

[12] For a description of the clay minerals, see Ross, C. S., and Shannon, E. V., The minerals of bentonite and related clays and their physical properties: Am. Ceramic Soc. Jour., vol. 9, pp. 77-96, 1926.

broken, and slickensided—a condition similar to that of the clay matrix of the kernite 2 miles to the east.

Noble considers that the contact of the borate shale and the lava is a depositional contact, the lava being a flow upon which the shale was deposited. The reported vesicular character of the top of the lava is adduced as evidence that it was a surface flow. In none of the shafts in the immediate vicinity of the Suckow shafts was any kernite or borax reported.

The colemanite and ulexite bearing clay of the Suckow shafts is probably similar to the clay carrying the same minerals which overlies the kernite. It is possible that the kernite-carrying clay is of the same formation, only thicker. Possibly this increased thickness is only local, and the kernite deposit may be the site of a former depression filled with a borax lake upon which the ulexite and colemanite bearing clays were deposited. Igneous rock is reported to underlie the kernite, just as lava underlies the other borate minerals in the Suckow shaft.

The discussion of the origin of the kernite deposit is based on a number of features, which first will be mentioned briefly and then treated at greater length. The general theory of the formation of the kernite here proposed is the one suggested by Mr. Thomas Cramer, chief chemist of the Pacific Coast Borax Co., at Wilmington, Calif. It is essentially that the kernite has formed by the fusion, chiefly in its own water of crystallization, of a preexisting accumulation of borax. The excess of water given off, permeating the overlying clay beds, may have dissolved the ulexite present, very likely as "cotton balls," and recrystallized it into the satinlike fibrous veins, a specimen of which is shown in Plate 22, *A*.

The features characterizing this deposit which may have had a decided influence in the formation of kernite are as follows:

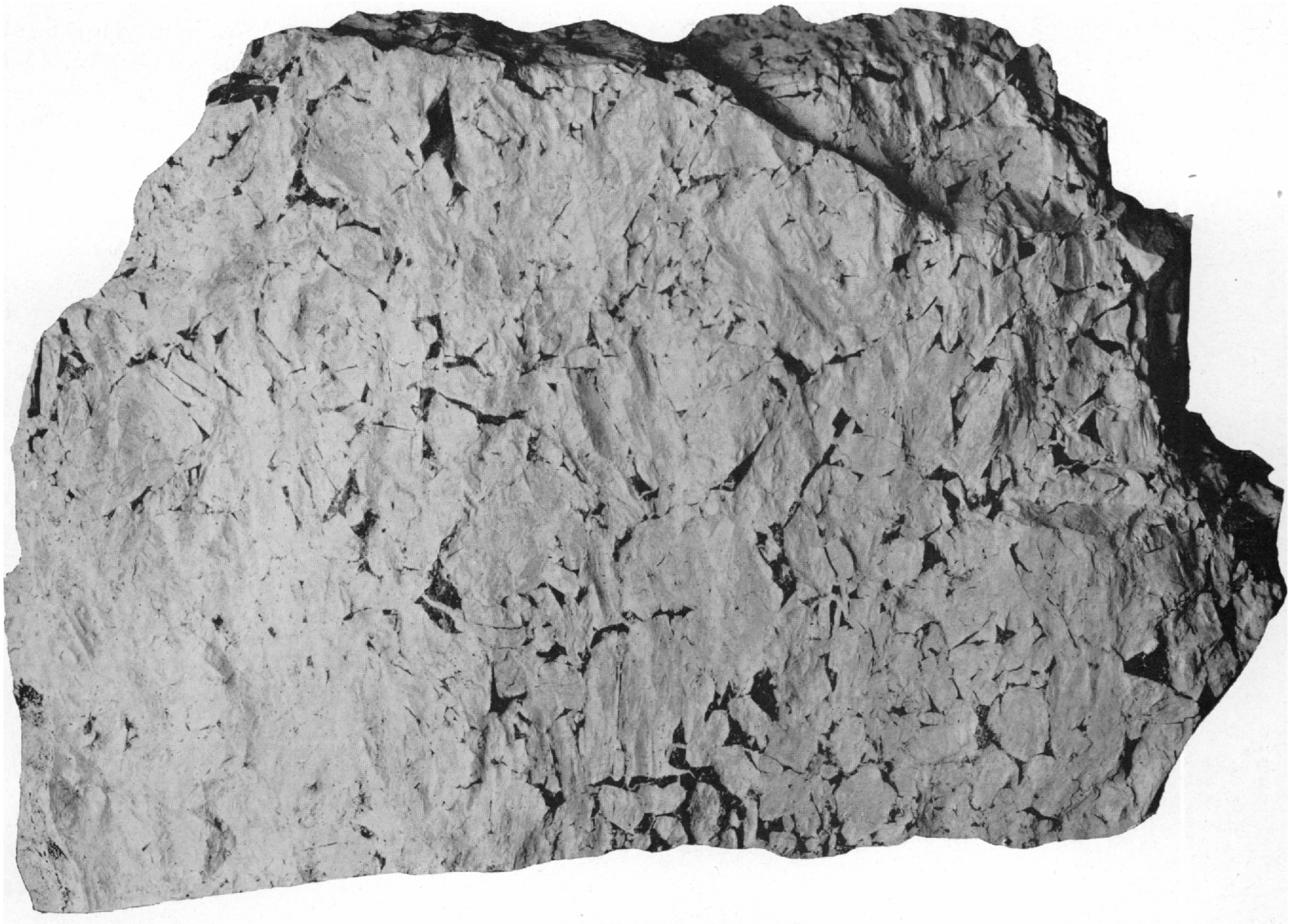
1. No other deposit of kernite is known, so that the assemblage of conditions was probably very unusual.
2. Kernite is absent in the western part of the field.
3. The difficulty of artificially making kernite shows that the conditions of its formation are limited.
4. Large bodies of massive borax are present.
5. Small veins of borax cut the kernite.
6. Kernite crystals are embedded in massive borax.
7. Igneous rock underlies the deposit.
8. Stibnite and realgar are present.
9. Calcium borates are scarce.
10. Ulexite veins occur above the deposit.

These features are considered to be of importance with reference to the origin of the kernite deposit and are discussed below.

1. The fact the kernite is a new mineral, never before found and not made artificially, even though the borax-water system has been studied in the laboratory, indicates that it is the result of a combination of unusual conditions. Were it a common product, easily made artificially, it would have been found in many localities, especially in the borate regions of California and Nevada. Apparently no single item of the conditions would yield kernite naturally, but the combination of several such unusual conditions as existed in this region has produced enormous quantities of this most interesting and valuable mineral. One can not help wondering what other unusual mineral associations lie buried in other parts of the Mohave Desert.

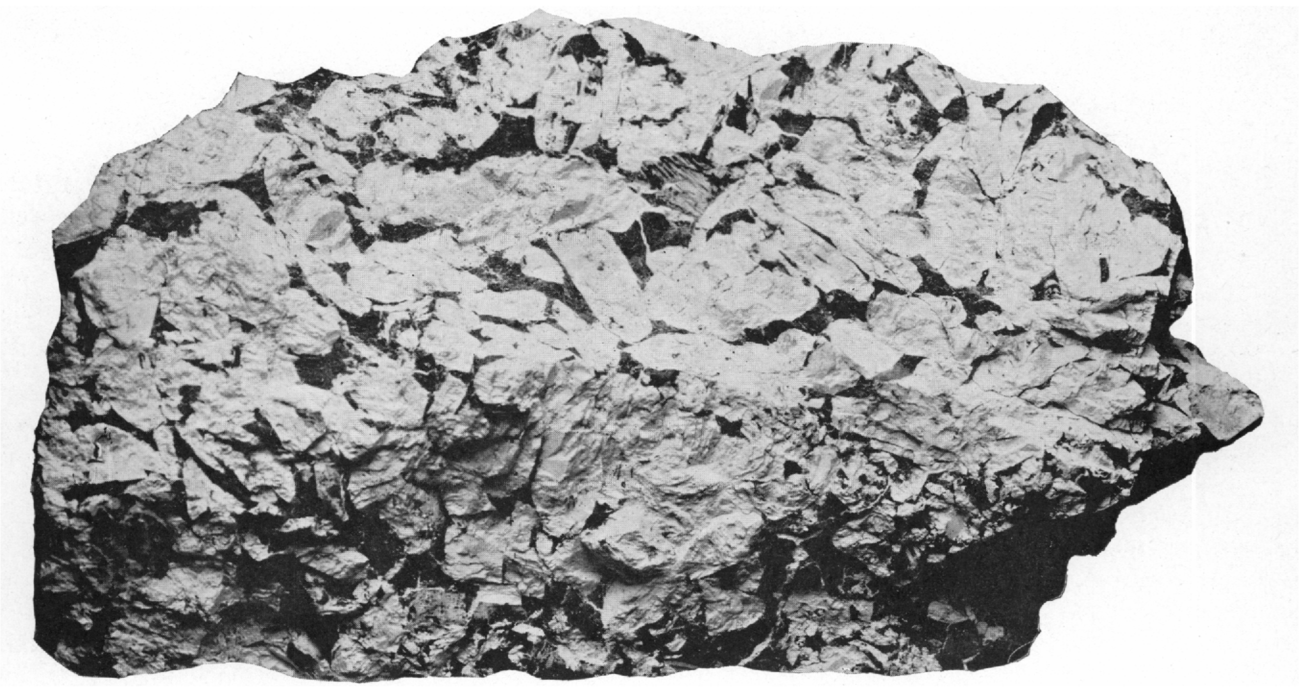
2. The absence of kernite in the western part of the Kramer district, where both ulexite and colemanite are present in abundance, suggests that an additional factor to those favoring the formation of ulexite and colemanite must have been present. The same conditions that yield these two calcium borates would not yield kernite, and it is considered doubtful if there is any genetic relation between kernite and ulexite. The concentration of the kernite in the eastern part of the field suggests also that this side of the field had a different topographic expression, and that there was at one time a depression here wherein was concentrated sodium borate, probably only as borax, long before the overlying clay containing ulexite was laid down. A detailed topographic map, with plottings of the logs of all the shafts and drill holes of the region, would be needed before a definite conclusion could be reached regarding the existence of a depression in the eastern part of the field.

3. The difficulty of making kernite artificially shows that the conditions usually existing in natural regions containing borates would not suffice to produce kernite. The experiment of Wells, in which the 3-hydrate was obtained at 150°–160° by heating borax in a closed tube, shows that the upper limit of stability of the 4-hydrate kernite lies near 150°–160°. The continuous production of the 5-hydrate by boiling a solution of borax and also by melting borax in its water of crystallization in both open and closed tubes, and the experiments of the writer in duplicating Wells's experiment but obtaining only the 5-hydrate, all suggest that the upper limit of stability of the 5-hydrate and consequently the lower limit of stability of the 4-hydrate is considerably above 100° and very likely not much below 150°. This leaves a very narrow range of stability for the 4-hydrate. Wells's success in obtaining kernite artificially by allowing the water distilled off to flow back on the sodium borate and reheating and redistilling off part of the water shows that pressure and the presence of some excess water are probably needed for the formation of the 4-hydrate. If the kernite was formed by the fusion of the preexisting borax in its own water of crystallization, then the clay matrix probably acted as a retardant for the free and ready escape of all the excess water, and it may be that the presence of such excess water not only helped but was essential in the production of the 4-hydrate, just as in Wells's successful experiment.



A. MASSIVE BORAX SUPERFICIALLY ALTERED TO TINCALCONITE

U. S. National Museum catalog No. 95838. Natural size.



B. CRYSTALS OF BORAX IN CLAY SUPERFICIALLY ALTERED TO TINCALCONITE

U. S. National Museum catalog No. 95839. Natural size.

BORAX AND TINCALCONITE

A. LARGER KERNITE CRYSTALS BETWEEN LAYERS OF SHALE

B. AGGREGATE OF LARGE KERNITE CRYSTALS

OCCURRENCE OF KERNITE IN THE BAKER DEPOSIT, PACIFIC COAST BORAX CO., KRAMER DISTRICT, CALIFORNIA

Photographs by W. F. Foshag, U. S. National Museum. One-fifteenth natural size.

4. The presence of large bodies of massive borax, such as are illustrated in Plate 26, *A*, indicates the form that an aggregate of borax crystals, when fused, would take, provided the escape of any water was prevented by an impervious surrounding clay. These large bodies of massive borax, showing practically no crystal outline, are not the form taken by borax crystallizing out of solution. It seems therefore that the assumption is justified that at one time large quantities of borax existed here in a molten state. If the escape of all water was prevented, only massive borax would result on cooling; if some water escaped, a lower hydrate would have to form; and if conditions were just right and the necessary quantity of water escaped, then kernite might readily form. The clay with the porphyritic crystals of borax, as shown in Plate 26, *B*, a rather rare occurrence, might represent the original borax crystals embedded in the clay or, more probably, a recrystallization of the massive borax into distinct crystals. The nonuniformity of the clay, with the presence of harder (and more impervious?) layers, as shown in Plate 25, *A*, strongly supports the idea that in parts of the deposit the escape of any water from any molten borax might well have been hindered.

5. The small veins of massive borax, apparently cutting kernite (pl. 25, *A*), may be a later effect, though they may be only apparent veins and actually may represent residua of the fused borax from which no water could escape.

6. The kernite crystals embedded in massive borax offer one of the most interesting associations for speculation about their origin. If kernite was derived from the fusion and partial dehydration of borax—an assumption considered to represent the conditions but in no way proved—then the explanation of this association becomes simple. Parts of the fused borax lost enough water to permit the formation of kernite, which started to grow as a solid, the consequent loss of volume being readily taken up by the still fused borax, which later crystallized as such in a massive form where no further escape of water could take place.

7. If the underlying igneous rock was extrusive and the original borax lake was deposited on it after it cooled, then a source of heat to fuse the deposit must be looked for. If, on the other hand, the igneous rock was intrusive and later than the original borax lake, then it would be the source of the heat that melted the borax crystals, whose partial dehydration then yielded kernite.

8. The presence of stibnite and realgar strongly suggests the underlying igneous rock as a source, these minerals probably being formed in the fumarole stage, after the molten rock had cooled considerably. In the absence of any knowledge of the extrusive or intrusive character of the igneous rock, the possible effect it may have had on the development of kernite must remain an unknown factor.

9. The scarcity of any calcium borates directly associated with the kernite shows that neither colemanite nor ulexite is directly connected with the formation of kernite, for the lime borates represent a fixed product. For example, in the formation of colemanite from ulexite, the lime borate colemanite is the fixed end product. The borax of the original (theoretical) deposit may represent the washed-out sodium borate from ulexite altering to colemanite, but it is doubtful whether the possibly higher deposits of ulexite to the west were ever extensive enough to furnish sufficient sodium borate to yield the immense deposit of kernite, though they may well have contributed. The source of the supposed original borax represents no greater difficulty than the origin of the borax in any of the borax lakes in the arid Southwest.

10. The ulexite veins (pl. 22, *A*) in the clay above the kernite deposit are believed to be later in their formation than the original borax deposit that changed to kernite. It is possible that the water escaping during the supposed dehydration of the original borax deposit entered the overlying clay and recrystallized any "cotton ball" ulexite that may have been present into the satinlike fibrous veins now found.

The theory of the origin of the kernite, as expressed to the writer by Mr. Thomas Cramer, seems to offer the best explanation. It is essentially as follows: A borax lake formerly occupied the area of the Kramer borate district. By evaporation of the water, borax crystallized out. Borax crystals may have been concentrated at the east side by wind (as suggested by Mr. Cramer), or there may have been a local depression at the east side into which the concentrated borax solution flowed, leaving only a relatively minor quantity of borax at the west end (in the region of the Suckow shafts). This borax was covered by mud and clay, with probably layers of volcanic ash. On top of this borax-bearing clay was deposited additional clay, also containing borates, more probably ulexite. This clay bed would correspond to the borate clays of the west end of the district. On top is the borate-free valley fill (sand and gravel). Heat from some source—the igneous rock if intrusive—melted the buried borax deposit in its own water of crystallization, a temperature of about 150° being reached. Enough of the water was driven off so that on cooling the 4-hydrate crystallized as kernite. Locally, where the escape of water was prevented by impervious clay, the fused borax solidified into massive borax. The excess of water driven off from the fused borax entered the overlying ulexite-carrying clay and recrystallized the original "cotton ball" ulexite into the fibrous veins now found.

BORON MINERALS

The following list comprises all the known boron minerals, in part arranged in systematic groups.

WITHOUT WATER

Avogadrite. $KF.BF_3$. A little caesium may be present, $(K, Cs)F.BF_3$.
Jeremejevite. $Al_2O_3.B_2O_3$. The name is here given, as by Dana, in the German transliteration from the Russian; the correct English transliteration, according to the standard rules of the Library of Congress and the Royal Society, is eremeyevite. Eichwaldite is considered to be a dimorphous form of the same compound.
Rhodizite. $4(H, Na, K, Cs, Rb)_2O.4BeO.3Al_2O_3.6B_2O_3$.
Nordenskioeldite. $CaO.SnO_2.B_2O_3$.
Ludwigite. $4MgO.Fe_2O_3.B_2O_3$. Synonyms, collbranite, magnesioludwigite.
Vonsenite. $4FeO.Fe_2O_3.B_2O_3$.
Pinakiolite. $4(Mg, Mn)O.Mn_2O_3.B_2O_3$.
Boracite. $5MgO.MgCl_2.7B_2C_3$. Synonym, stassfurtite. Iron boracite contains a little iron.
Warwickite. $6MgO.FeO.2TiO_2.3B_2O_3$. Synonym, enceladite.

WITH WATER; WITHOUT SILICA

Sassolite. $B_2O_3.3H_2O$.
Larderellite. $(NH_4)_2O.4B_2O_3.4H_2O$.
Paternoite. $MgO.4B_2O_3.4H_2O$.
Kaliborite. $K_2O.4MgO.11B_2O_3.9H_2O$. Synonyms, hintzeite, heintzite.
Bechilite. $CaO.2B_2O_3.4H_2O(?)$. Synonym, borocalcite. The existence of this mineral is doubtful.
Kernite. $Na_2O.2B_2O_3.4H_2O$. Synonym, rasorite.
Tincalconite. $Na_2O.2B_2O_3.5H_2O$.
Borax. $Na_2O.2B_2O_3.10H_2O$. Synonym, tincal.
Kramerite. $Na_2O.2CaO.5B_2O_3.10H_2O$.
Ulexite. $Na_2O.2CaO.5B_2O_3.16H_2O$. Synonyms, boronatrocalcite, cryptomorphite, franklandite, and many others.
Colemanite. $2CaO.3B_2O_3.5H_2O$.
Meyerhofferite. $2CaO.3B_2O_3.7H_2O$.
Inyoite. $2CaO.3B_2O_3.13H_2O$.
Hydroboracite. $CaO.MgO.3B_2O_3.6H_2O$.
Priceite. $5CaO.6B_2O_3.9H_2O$ or $4CaO.5B_2O_3.7H_2O$. Synonym, pandermite.
Pinnoite. $MgO.B_2O_3.3H_2O$.
Szaibelyite. $2MgO.B_2O_3.H_2O$. Synonyms, boromagnesite, camsellite(?), and possibly ascharite, which has been regarded as containing only two-thirds molecule of H_2O.
Camsellite. $2MgO.B_2O_3H_2O$. Probably the same as szaibelyite. Camsellite from California is supposed to contain some silica.
Sussexite. $2(Mn,Mg)O.B_2O_3.H_2O$.
Ascharite. $2MgO.B_2O_3.\frac{2}{3}H_2O(?)$. According to Van't Hoff's work, ascharite has the same formula as szaibelyite. It needs restudy.
Hambergite. $4BeO.B_2O_3.H_2O$.
Fluoborite. $6MgO.B_2O_3.3((OH),F)$.
Luenebergite. $3MgO.B_2O_3.P_2O_5.8H_2O$.
Sulphoborite. $6MgO.2SO_2.2B_2O_3.9H_2O$.
Hulsite. $12FeO.2Fe_2O_3.SnO_2.3B_2O_3.2H_2O$. Paigeite differs in containing a smaller proportion of tin but may be essentially the same.
Cahnite. $4CaO.As_2O_5.B_2O_3.4H_2O$.
Lagonite is a mixture of sassolite and limonite.

WITH SILICA

The formulas of many of the borosilicates are complex and uncertain. The minerals are arranged below alphabetically and not in systematic groupings.

Axinite. $2Al_2O_3.2(Fe, Mn)O.4CaO.H_2O.B_2O_3.8SiO_2$.
Bakerite. $8CaO.6H_2O.5B_2O_3.6SiO_2(?)$.
Cappelenite. Ba, Y, H, B, Si, O.
Caryocerite. Th, Ce, La, Nd, Pr, Y, Ca, H, F, B, Si, O.
Danburite. $CaO.B_2O_3.2SiO_2$.
Datolite. $2CaO.H_2O.B_2O_3.2SiO_2$.
Dumortierite. $8Al_2O_3.H_2O.B_2O_3.6SiO_2$.
Grandidierite. $11(Al, Fe, B)_2O_3.7(Mg, Fe, Ca)O.2(H, Na, K)_2O.7SiO_2$.
Homilite. $2CaO.FeO.B_2O_3.2SiO_2$.
Howlite. $4CaO.5H_2O.5B_2O_3.2SiO_2$.
Hyalotekite. $16(Pb, Ba, Ca)O.F.2B_2O_3.24H_2O$.
Kornerupine. $8MgO.6(Al, B)_2O_3.7SiO_2(?)$.
Manandonite. $7Al_2O_3.2Li_2O.12H_2O.2B_2O_3.6SiO_2$.
Melanocerite. Ce, Nd, Pr, La, Y, Ca, H, F, B, Si, O.
Sapphirine. $5MgO.6(Al, B)_2O_3.2SiO_2(?)$.
Searlesite. $Na_2O.2H_2O.B_2O_3.4SiO_2$.
Serendibite. $3Al_2O_3.2CaO.4MgO.B_2O_3.4SiO_2$.
Tourmaline. Al, Mg, Fe, Mn, Ca, Na, K, Li, H, F, B, Si, O. A group name including several species.
Tritomite. Th, Ce, La, Nd, Pr, Y, Fe, Ca, H, F, B, Si, O.
Vesuvianite. Al, B, Fe, Ca, Mg, H, Si, O. The variety wiluite contains the largest percentage of boron (as much as 4.2 per cent) but other varieties, from widely separated localities, contain smaller amounts.

This list of boron minerals contains several series of hydrates of the same compound—the kernite-tincalconite-borax series, the colemanite-meyerhofferite-inyoite series, and the kramerite-ulexite series. Their properties being known, a comparative study may be made.

The sodium borate series will be first considered. Evidences of the existence of artificial hydrates of $Na_2O.2B_2O_3$, other than the 5-hydrate and 10-hydrate, in a well-defined crystalline phase were not found in the literature. Lescoeur [13] mentions literature references to the 2, 3–4, and 6 hydrates, but these show no data of value on which the existence of such hydrates could be based. The 3-hydrate and 4-hydrate, for example, was amorphous and prepared at 100°. The 6-hydrate found by Bechi [14] as an incrustation at Tuscany, Italy, gave on analysis a composition agreeing in water content with the 6-hydrate, but Bechi gives no description whatever regarding the physical state or homogeneity of the material. An explanation as a mixture of borax with a more or less dehydrated borax (tincalconite?) is more probable.

The anhydrous amorphous compound belonging to this series, with zero water content, the well-known "borax glass," was found to have a density of 2.36 (determinations given in the literature are 2.371, 2.373,

[13] Lescoeur, H., Annales chimie et phys., 7th ser., vol. 9, pp. 543-546, 1896.
[14] Bechi, C., Jahrb. Chemie, vol. 7, p. 867, 1854; abstract in Am. Jour. Sci., 2d ser., vol. 19, p. 120, 1855.

and 2.5, the last obviously in error). The refractive index was determined to be 1.513 (1.513 for glass obtained from kernite, 1.513 for glass obtained from artificial tincalconite, and 1.513–1.514 for glass obtained from artificial borax). The value 1.515 is given in the literature. From the values of the density (d) and mean refractive index (n) the specific refractive index of B_2O_3 can be calculated, using the formula $\frac{n-1}{d}=K=\frac{k_1p_1}{100}+\frac{k_2p_2}{100}+\frac{k_3p_3}{100}$, k_n being the specific refractive index for the radicle whose percentage in the particular substance is given by p_n. The same values for Na_2O (0.181) and for H_2O (0.34) are used, taken from Larsen's tabulation.[15]

The values for the series of sodium borates then are as follows:

Properties of sodium borates

	d	n	k (B_2O_3)
Isotropic borax glass, $Na_2O.2B_2O_3$	2. 36	1. 513	0. 217
Kernite, $Na_2O.2B_2O_3.4H_2O$	{1. 91	1. 471	. 228
	{1. 95	1. 471	. 218
Tincalconite,[a] $Na_2O.2B_2O_3.5H_2O$	1. 88	1. 465	. 217
Borax,[a] $Na_2O.2B_2O_3.10H_2O$	1. 72	1. 463	. 217

[a] Artificial.

Two values are given for kernite, depending on the gravity. A density of 1.91 was determined, whereas a density of 1.95 would give a more consistent value for k. It is possible that the multitude of inclusions and negative crystals present in kernite have a lowering effect on the density of the whole mass. On the other hand, the constant figure 0.217 obtained for $k(B_2O_3)$ from the other members of the series may be accidental. The effect of the varying figure for the density on the calculated value of k may be shown for borax, where the specific refractive value, as calculated, will vary about 0.004 for each 0.01 difference in the density as here shown. Using the value of 1.463 for n, we find that the value of $k(B_2O_3)$ in borax varies according to the density, as follows:

$$\begin{array}{ll} d=1.70 & k=0.2256 \\ d=1.71 & k=0.2212 \\ d=1.72 & k=0.2169 \\ d=1.73 & k=0.2127 \end{array} \quad \begin{array}{l} \Delta=0.0044 \\ \Delta=0.0043 \\ \Delta=0.0042 \end{array}$$

The rather constant value of the mean index for the three hydrates (the differences being but slightly greater than the error of determination) is very striking. The density decreases, fairly regularly, with the increase in water content. The density (1.815) given in the literature for the 5-hydrate (artificial tincalconite) gave a discordant value for k. The density was therefore redetermined (heavy solution, bromoform and carbon tetrachloride) and found to be 1.88. On looking through the literature, only this one determination of the density of the 5-hydrate could be found,[16] and this was published over a hundred years ago!

The calcium borates form a series of three minerals, as follows:

Properties of calcium borates

	d	n	$k(B_2O_3)$
Colemanite, $2CaO.3B_2O_3.5H_2O$	2. 423	1. 597	0. 217
Meyerhofferite, $2CaO.3B_2O_3.7H_2O$	2. 12	1. 532	. 211
Inyoite, $2CaO.3B_2O_3.13H_2O$	1. 875	1. 505	. 215

Apparently, the slightly discordant value of 0.211 for k for meyerhofferite suggests that the value given for its density is not quite accurate. In this series the variation in the mean index corresponds with the variation in the density. Why the mean index in the sodium borate series is so nearly constant is not known.

The sodium-calcium borate series comprises kramerite and ulexite:

Properties of sodium-calcium borates

	d	n	k (B_2O_3)
Kramerite, $Na_2O.2CaO.5B_2O_3.10H_2O$	2. 141	1. 528	0. 217
Ulexite, $Na_2O.2CaO.5B_2O_3.16H_2O$	1. 963	1. 506	. 213

The average value of k for B_2O_3, using the values for Na_2O, CaO, and H_2O as given by Larsen, is close to 0.216.

Although comparisons of crystallographic axes of different minerals may lead to unwarranted deductions, it is worth while to call attention to such similarities as exist. Thus, for example, the axial ratios of kramerite are close to those of borax, and by fractionally changing those of colemanite and kernite, a similarity can be shown for four of these borates, all monoclinic:

$$\begin{array}{llll} \text{Colemanite} & \frac{4}{3}a:b:c & =1.0357:1:0.5430 & \beta=69^\circ\ 53' \\ \text{Kernite} & \frac{2}{3}a:b:\frac{1}{3}c & =1.0153:1:0.5663 & \beta=71\ \ 08 \\ \text{Borax} & a:b:c & =1.0995:1:0.5632 & \beta=73\ \ 25 \\ \text{Kramerite} & a:b:c & =1.1051:1:0.5237 & \beta=72\ \ 16 \end{array}$$

[15] Larsen, E. S., The microscopic determination of the nonopaque minerals: U. S. Geol. Survey Bull. 679, p. 31, 1921.

[16] Payen, A., Jour. chimie méd., vol. 3, p. 594, 1827. Payen also gives 1.74 as the density of borax.

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