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STATE DIVISION OF CONSERVATION

DEPARTMENT OF MINES, MINING AND GEOLOGY

GARLAND PEYTON, Director

THE GEOLOGICAL SURVEY

Bulletin No. 63

# STRATIGRAPHY, STRUCTURE, AND MINERAL RESOURCES OF THE MINERAL BLUFF QUADRANGLE, GEORGIA

By

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ATLANTA

1955

# LETTER OF TRANSMITTAL

**Department of Mines, Mining and Geology**

Atlanta, November 28, 1955

To His Excellency, Marvin Griffin, Governor

Commissioner Ex-Officio of State Division of Conservation

Sir:

I have the honor to submit herewith Georgia Geological Survey Bulletin No. 63, "Stratigraphy, Structure, and Mineral Resources of the Mineral Bluff quadrangle, Georgia," by Dr. Vernon J. Hurst who is a geologist with the Department of Mines, Mining, and Geology.

The manner of collecting and interpreting the data in this report is a new approach to mineral deposit investigation. It has already proven of value in Fannin County and will, it is anticipated, prove equally useful in other parts of the State.

Attention is particularly invited to Plate 6 which represents detailed petrofabric work done by a new technique and is the first of its kind to be published in English.

This bulletin fulfills a long-felt need for basic descriptive data on the petrography, stratigraphy, and structure of the Mineral Bluff-Blue Ridge area. It will be useful to geologists, mining engineers, and prospectors, and to all others who are interested in the complicated geologic history and mineral deposits of north-central Georgia.

Very respectfully yours,

Garland Peyton

GARLAND PEYTON
Director

# ABSTRACT

The 7½ minute Mineral Bluff quadrangle lies mainly in Fannin County, Georgia, but includes part of Cherokee County, North Carolina, and part of the Ducktown Basin of Tennessee.

The rocks of this area have been metamorphosed to medium-grade, and are nearly all of sedimentary origin. The oldest rocks belong to the Great Smoky group, a sequence of graywacke-type metasediments at least 15,000 feet thick, of probable pre-Cambrian age. The Great Smoky group is divided into four new formations: the Copperhill formation, the Hughes Gap formation, the Hothouse formation, and the Dean formation. Overlying the Great Smoky group with possible unconformity is a metasedimentary sequence 4000-6000 feet thick of probable Cambrian age. This sequence, oldest to youngest, is: Nantahala slate, Tusquitee quartzite, Brasstown formation, Murphy marble, Andrews formation, Nottely quartzite, and Mineral Bluff formation. The principal rock types of this sequence are black slate, feldspathic quartzite, metasubgraywacke, marble, and sericite schist.

In the Great Smoky group and in the younger series bedding, graded bedding, and scour channels are well preserved. Cross-bedding is preserved in the younger series.

The only rocks of recognized igneous origin are two sill-shaped epidote-amphibolite masses which originated either as intrusive sills or as submarine lava flows.

The folds trend NE-SW and are overturned to the NW. The major fold is the Murphy syncline, a bent fold, whose axis passes through Mineral Bluff and Blue Ridge. An anticlinorium of comparable size lies to the northwest. The main folds in the Copperhill area are second-order folds on the east limb of this anticlinorium. Axial plane flow cleavage is well developed to the northwest, weak toward the southeast; bedding-plane foliation is well developed in the Murphy syncline. Local crinkling of the bedding-plane cleavage correlates with crowding movements in the trough of the Murphy syncline. The intense crinkling of flow cleavage to the northwest is also incident to the tightening of earlier folds.

The metamorphic history includes two periods of deforma-

tion. During the first, the major folds and cleavages originated. During and partly after this period, the rocks completely recrystallized. Then followed the last period of deformation marked by the tightening of old folds, the formation of new folds on the limbs of old folds, the rumpling of the earlier cleavages, and local faulting. Subsequent to the last period of deformation the rocks were pervasively altered.

Rock and minerals that have been mined or might be amenable to future exploitation are marble, talc, iron ore, staurolite, kyanite, refractory schist, quartz, slate, quartzite, conglomerate and metagraywacke.

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# INTRODUCTION

## Previous Work; Purpose of This Investigation

The first geologic map of the Mineral Bluff quadrangle is that made by LaForge and Phalen (1913) under the general supervision of Keith. Their stratigraphic classification and structural views are the same as those which Keith had worked out earlier in quadrangles to the north and northeast (Knoxville, Asheville, and Nantahala quadrangles; 1895, 1904, 1907, respectively). According to these views, the rocks of the Murphy belt are younger than those of the Great Smoky group, and the Murphy belt is a complex syncline.

Since the work of LaForge and Phalen, a number of geologists have examined the Mineral Bluff area and have studied the same formations in other areas, with conflicting results. The Murphy belt has been interpreted as a fenster (Stose and Stose, 1944, p. 377), a syncline (Stose and Stose, 1949, pp. 286-291; King and Hadley, 1952; Furcron, 1953, pp. 36-37); an anticline (Jonas, 1932, p. 240); and a homocline (Van Horn, 1948, pp. 18-20). The Murphy marble has been placed at the base of Keith's sequence, near the middle, at the top, and in tectonic discontinuity with the enclosing strata.

The present investigation is an attempt to clear up the stratigraphy and structure of the Murphy belt and the flanking Great Smoky group. Specifically, an effort has been made to:

(1) determine the relative age and contact relations of the Murphy sequence and Great Smoky group;

(2) subdivide the Great Smoky group into mappable units;

(3) account for the difference in appearance of the rocks to the east and west of the Murphy belt;

(4) determine the structure of the Murphy belt, and the reason for the structural differences that exist between the Murphy belt and the areas to the west;

(5) inquire into the significance of the regional crinkling; and

(6) relate the ore-localizing structures of the Ducktown Basin to the geologic history of the Mineral Bluff area.

## Location and Size of Area

The Mineral Bluff Quadrangle (Fig. 1) is mostly in Fannin County, Georgia, but on the north it includes a narrow strip of Polk County, Tennessee, and Cherokee County, North Carolina. It is bounded by the parallels 34° 52′ 30″ and 35° 00″ N and the meridians 84° 15′ and 84° 22′ 30″ W. The principal towns are Copperhill, in Tennessee, and McCaysville, Blue Ridge, and Mineral Bluff, in Georgia.

The dimensions of the quadrangle are 7.1 (east-west) by 8.58 miles (north-south), an area of about 61 square miles.

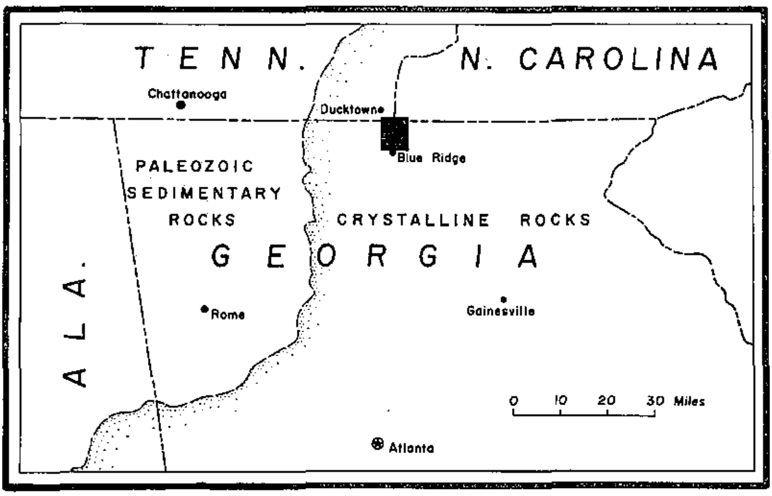



Fig. 1. Index map showing location of the Mineral Bluff quadrangle (black rectangle).

## Topography

The quadrangle is in the physiographic division which Keith called the Highland (see figure 2), and mainly within the subdivision known as the Ducktown Plateau (LaForge, Cooke, Keith, and Campbell, 1925) or the Toccoa Plateau (LaForge and Phalen, 1913).

The surface of the Plateau is everywhere rounded, with the summits of the knolls rising to a general elevation of 1600-1900 feet above sea level. All the streams are entrenched; relief is about 300 feet. The NE-SW ridge trend so characteristic of most of the Blue Ridge is conspicuously absent

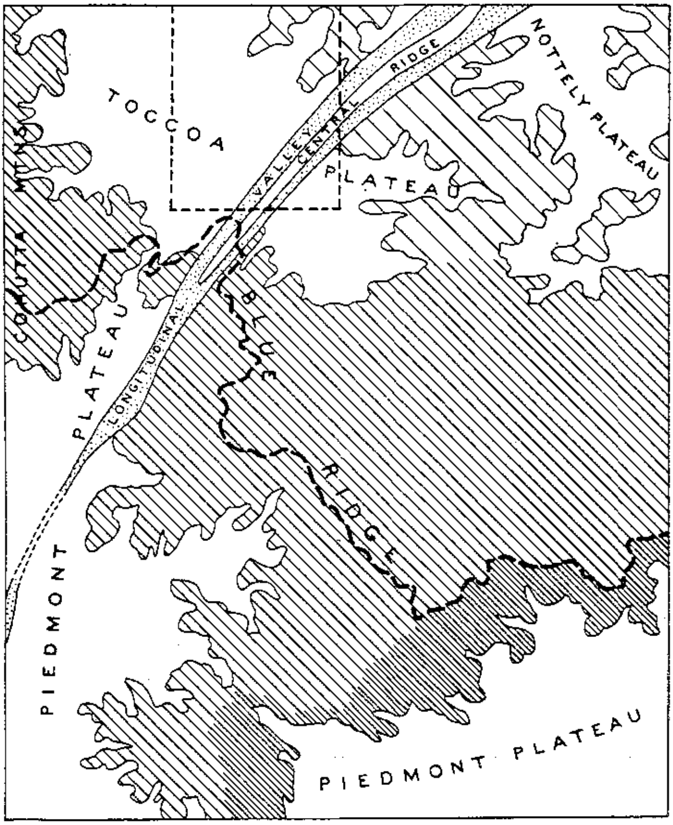



Fig. 2. Map of the topographic divisions of the Ellijay quadrangle. Medium ruling represents the higher mountain groups, with summits over 3000 feet; open ruling represents the lower mountain groups, with summits under 3000 feet; dense ruling, the Blue Ridge escarpment; unshaded areas, Piedmont and associated plateaus; stippled area, Longitudinal Valley. The boundaries of the Mineral Bluff quadrangle are dashed. (From LaForge and Phalen, 1913.)

within the Plateau proper. This trend is apparent, however, in the Longitudinal Valley Belt, which crosses the Ducktown Plateau in the vicinity of Mineral Bluff and Blue Ridge.

Seven mountains rise above the general level of the Plateau: Piney, Roundtop, Watson, Cordell, Franklin, Stewart, and Hughes. The highest is Roundtop, elevation 2453 feet.

**Drainage**

The Toccoa River, an entrenched, meandering stream, crosses the quadrangle from SE to NW. Its principal tributaries are Wolf Creek and Hothouse Creek, draining the northeastern part of the quadrangle, Sugar Creek, draining the southwest part, and Hemptown Creek, draining the southeast. The drainage pattern is dendritic over all the area except the Longitudinal Valley, where a trellised pattern prevails.

The Toccoa River becomes the Ocoee River in Tennessee.

## Field Work

The geologic mapping was done on aerial photographs, scale 1:12000. Foot traverses were carried out first along the roads, which provide most of the exposures, then across-strike at intervals of a few hundred feet between roads. All contacts were walked out.

Field work occupied 12 weeks during the summer of 1953 and one week in April, 1954.

## Laboratory Work

The laboratory work was done during 1953-54, partly at Johns Hopkins University and partly at the Geophysical Laboratory of the Carnegie Institution in Washington, D. C.

The petrographic study involved 125 thin sections and 50 rock slabs. Minerals were identified mainly by their optical properties, measured on the U-stage. To aid in the identification of K-feldspar 20 thin sections and 12 rock slabs were stained (Chayes, 1952). A Norelco Geiger-counter X-ray Spectrometer was used for identification of micas and garnets. Modal analyses were made with the point-count technique developed by Chayes (1940).

Thirty oriented thin sections were used in the petrofabric study. From seven of these sections, 13 density diagrams and one A.V.A. (Achsenverteilungsanalyse) were made.

## Acknowledgments

The Georgia Department of Mines, Mining, and Geology, under the direction of Captain Garland Peyton, sponsored the field work. Dr. A. S. Furcron, chief geologist for the State, participated in the initial reconnaissance, outlined previous work, pointed out the main stratigraphic units, and indicated features to be particularly studied. Wade Allen of Mineral Bluff helped to locate old marble quarries, lime pits, and talc prospects. Owen Kingman, chief geologist for the Tennessee Copper Company, made available company maps and critically discussed the stratigraphic and structural problems of the Ducktown Basin, which adjoins the Mineral Bluff Quadrangle on the north.

Much of the laboratory work was done at the Geophysical Laboratory of the Carnegie Institution, Washington, D. C. Special thanks are due to Hatten S. Yoder, Felix Chayes, and Gordon Davis. Dr. Yoder supervised the X-ray work on mica and garnet. Dr. Chayes instructed in feldspar staining and point count analysis; Gordon Davis helped with alkali determinations.

Special thanks are due also to the Faculty of the Geology Department of The Johns Hopkins University. Deeply appreciated are the unfailing interest and assistance of Prof. J. D. H. Donnay and Dr. Gabrielle Donnay; the perceptive criticism and friendly counsel of Prof. A. C. Waters; the encouragement, guidance, and continued help of Professors Ernst Cloos and Francis Pettijohn under whose supervision this study was made.

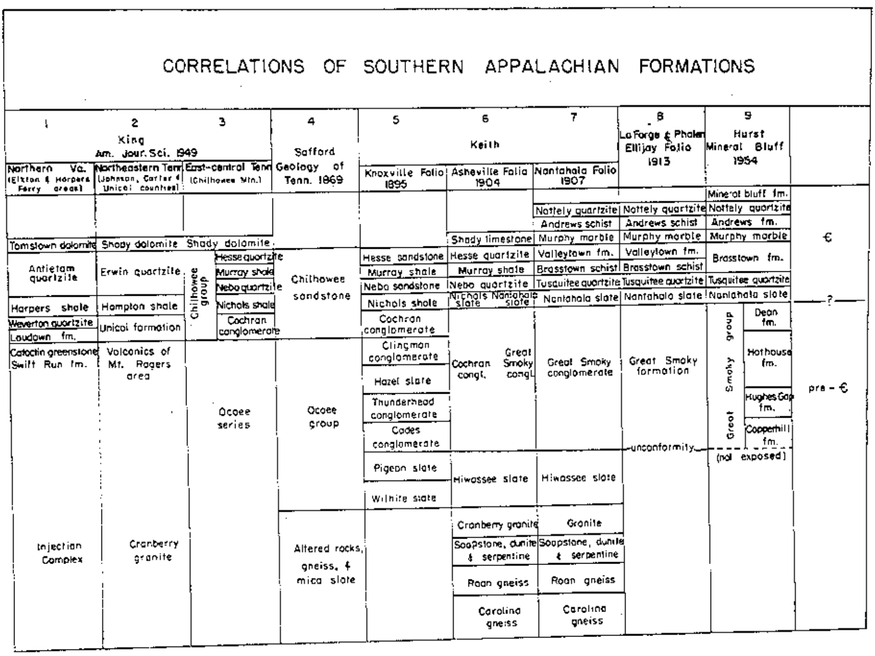
# METASEDIMENTARY ROCKS

## General Character and Stratigraphic Succession

The rocks in the Mineral Bluff quadrangle consist of metagraywacke or biotite gneiss, metaconglomerate, metaarkose, quartzite, staurolite-mica schist, garnet-mica schist, pseudodiorite, marble and slate, all of sedimentary origin. Relic sedimentary structures are well preserved, even in the mica schist and pseudodiorite.

The stratigraphic sequence is given in Plate 1. The formational names and boundaries are essentially those established by Keith in the Nantahala quadrangle (1907) and adopted by LaForge and Phalen in the Ellijay quadrangle (1913). The few major changes that have been made (compare columns, 7, 8, and 9 in Plate 2) are discussed in the descriptions of the formations.

PLATE 2




| Columns | Correlated by |
|---|---|
| 1-2-3-4 | King, 1949 |
| 4-5-6-7 | Keith, 1907 |
| 7-8 | LaForge & Phalen, 1913 |
| 8-9 | Hurst, 1954 |

### Age

Because fossils have not been found in the Mineral Bluff quadrangle, interpretations of age depend on lithologic correlation. The pertinent regional correlations are outlined in Plate 2. Cambrian fossils have been reported in the Shady limestone (Butts, 1940, p. 54); the Murphy marble (Crickmay, 1936, p. 1380); the Hesse, Erwin, and Antietam quartzites (Butts, op. cit., p. 40); and the Murray shale (Keith, 1895, p. 3). If the correlations of Keith are correct, all the rocks above the Tusquitee quartzite are of Cambrian age. Where the base of the Cambrian should be drawn is uncertain.

The position of the Cambrian-pre-Cambrian boundary is a familiar problem in Appalachian stratigraphy. Resser (1933, pp. 743-46) advocates placing the boundary at the bottom of the Antietam quartzite, the lowest fossiliferous horizon. Others regard the first major unconformity beneath the Antietam as a more logical boundary, but do not agree on the position of the unconformity. Stose and Stose (1944, p. 387) and King (1949, p. 513) favor the base of the Chilhowee group; Cloos (1951, pp. 25-28) the base of the Catoctin formation. Bloomer (1950, p. 781) believes the base of the Cambrian may be at any one of several points between the Swift Run formation and the Chilhowee group.

No unconformity below the Tusquitee quartzite has been recognized in the Mineral Bluff quadrangle; however, a distinct lithologic break exists at the base of the Nantahala slate. Poorly-sorted, "poured-in", graywacke-type sediments dominate below this break; above the break are black slates, better-sorted quartzites, and marble. These two lithologies represent different depositional environments (Pettijohn, 1949, pp. 242-43, pp. 252-55). In conformity with a widely-held view that the base of the Cambrian should be marked by a stratigraphic break, the Cambrian-pre-Cambrian boundary is provisionally drawn between the Great Smoky group and the Nantahala slate.

The formations are described below in order of age, beginning with the oldest.

### Great Smoky Group

The name "Great Smoky conglomerate" was used by Keith

(1907) for a thick sequence of interbedded conglomerates, graywackes, quartzites, schists, and slates in the Nantahala quadrangle. A similar sequence of rocks in the Ellijay quadrangle, which adjoins the southwest corner of the Nantahala quadrangle, was later regarded by LaForge and Phalen as equivalent to Keith's "Great Smoky conglomerate", and was mapped by them (1913) as the "Great Smoky formation". They estimated the thickness of the sequence at 5000-6500 feet. Detailed mapping in the Mineral Bluff quadrangle shows that this sequence is at least 15000 feet thick, and that it is divisible into four distinct lithologic units, each of sufficient stratigraphic importance to warrant a formational name. The sequence is therefore redesignated the Great Smoky group and divided into the Copperhill formation, the Hughes Gap formation, the Hothouse formation, and the Dean formation.

## COPPERHILL FORMATION

**Name and distribution.** The formation is named after the town of Copperhill, Tennessee, in the vicinity of which it is well exposed.

The formation underlies all of the northwest corner of the Mineral Bluff quadrangle. The eastern boundary of the formation passes through Franklin Mountain, just east of the mouth of Hothouse Creek, and just east of Stewart Knob. Along the boundary, metagraywacke and mica schist of the Copperhill formation are interbedded with staurolite-mica schist and metaquartz conglomerate which characterize the overlying Hughes Gap formation. The transition zone varies in thickness, but is usually a few hundred feet thick.

West of the Mineral Bluff quadrangle, in the Epworth quadrangle, and to the north in the Ducktown Basin, the Copperhill formation crops out extensively. The base of the formation is not exposed in the Mineral Bluff quadrangle, but may be observed in the Ducktown Basin, where it consists of thick, locally conglomeratic, metagraywacke beds underlain by staurolite-mica schist.

**Character.** About 60% of the Copperhill formation is metagraywacke, about 30% is mica schist; the remaining 10% is metaconglomerate, quartzite, and metaarkose. The metagraywacke and mica schist are interbedded throughout the formation. At places they alternate rhythmically, each bed of metagraywacke grading upward to mica schist (see figure 4). More commonly, the alternation is irregular, and succeeding beds differ in thickness. Metaconglomerate, quartzite, and metaarkose are prominent at only a few horizons. They are interbedded with, and usually gradational to, metagraywacke. The strata range in thickness from a fraction of an inch to 50 feet, but are mostly less than four feet thick. Metaconglomerate and metagraywacke beds are the thickest. The schist layers rarely exceed eight feet in thickness, commonly they are only a few inches thick.

Lithologic changes occur both laterally and vertically. Some beds lense-out within a distance of a few hundred feet. Other beds have wide lateral extent, particularly the thicker schists. A few horizons that are characterized by the predominance of a single rock type maintain their lithologic character for miles, even though the beds that mark the horizon vary greatly in number and thickness from place to place.

The interbedded relationship of these rocks, their composition, widespread graded bedding, and rounded pebbles of quartz, feldspar, and slate all clearly indicate a sedimentary origin. The metagraywackes, quartzites, and metaarkoses represent arenaceous sediments; the schists represent silts and clays. The range in mineral composition of most of the rocks is indicated by the stippled area in figure 3. Few of the schists contain more than 85% mica. Few of the metaarenites contain more than 40% feldspar or 75% quartz.

The coarser rocks are gray where unweathered. During weathering they change to a lighter color, as their feldspar is kaolinized and their biotite leached. The color of the schists depends to a large extent on their biotite content. A few schists are exceptionally dark because they contain fine, disseminated, opaque matter, either iron oxide or carbonaceous matter, sometimes both.

Metagraywacke. The metagraywacke is generally a light gray rock composed principally of quartz, feldspar, biotite, and muscovite, which can be recognized megascopically. Although much of the rock is massive, particularly in the thicker beds, as a rule it is distinctly foliated. The foliation planes, which are sometimes parallel to the bedding, are usually at an angle to the bedding and correspond to the flow cleavage in the interbedded schists. Graded bedding is extremely common (Fig. 4). Less regular variations in grain size occur both laterally and vertically. Compositional gradations from metagraywacke to mica schist and to metaconglomerate are frequent.

The texture of the rock stems partly from its clastic origin and partly from recrystallization. Original clastic shapes are usually preserved by the quartz and feldspar grains of medium size and larger. These grains are set in a granoblastic to schistose matrix of fine quartz and feldspar and fine to coarse mica. The metagraywacke has been called biotite gneiss. The term is appropriate where recrystallization has gone far enough to obscure original grain boundaries.

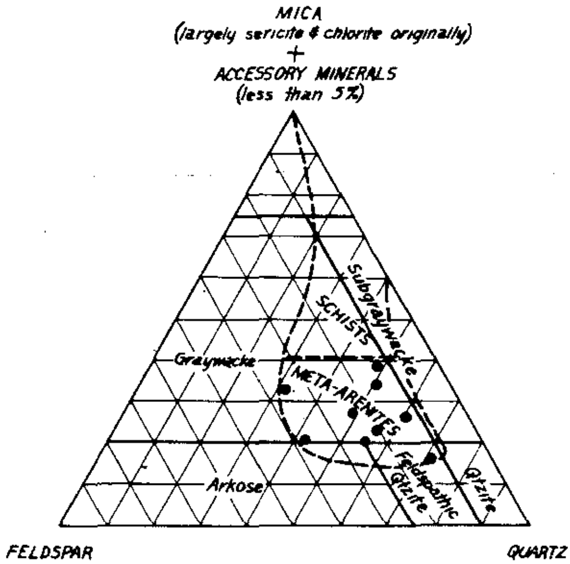



Fig. 3. Range in mineral composition of the rocks in the Copperhill formation. Dots represent typical metagraywacke specimens. Classificatory scheme from Pettijohn (1949, p. 227).

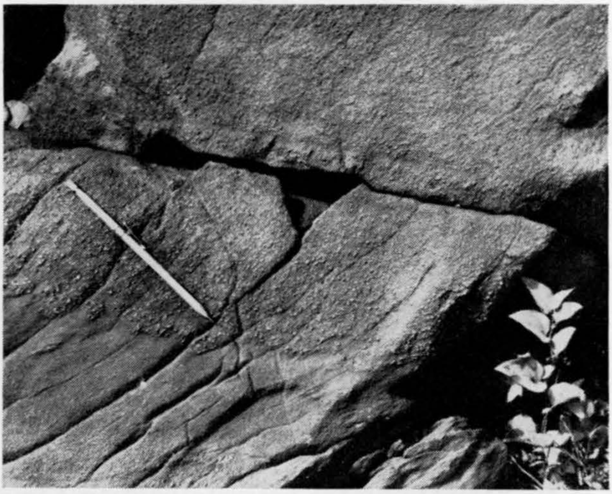


Fig. 4. Graded bedding in metagraywacke of the Copperhill formation.

Modal analyses of nine typical metagraywacke specimens are in table 1.

## TABLE 1

### Modal Analyses of Metagraywacke From the Copperhill Formation

| Specimen number | 2Z9 | $A5_b$ | 1M2 | $G5_b$ | 2D6 | 2T5 | 2S3 | IM3 | $N7_a$ |
|---|---|---|---|---|---|---|---|---|---|
| Quartz | 48.80 | 50.54 | 60.71 | 48.41 | 55.19 | 56.99 | 71.14 | 31.78 | 42.63 |
| Plagiclose | 24.74 | 15.53 | 12.89 | 13.41 | 25.27 | 20.29 | 12.43 | 22.75 | 26.86 |
| K-feldspar | | | | | | | | 13.66 | 11.43 |
| Biotite | 18.58 | 19.56 | 15.64 | 24.04 | 5.89 | 9.76 | 4.08 | 20.22 | 12.69 |
| Muscovite | 2.60 | 13.20 | 9.60 | 13.25 | 4.73 | 8.72 | 7.53 | 11.16 | 42.3 |
| Calcite | 4.14 | | | | 2.25 | 2.50 | 2.96 | P* | 0.80 |
| Apatite | 0.29 | 0.16 | 0.17 | | 0.46 | 0.43 | P | P | 0.20 |
| Black Opaque | 0.68 | 0.54 | 0.09 | 0.74 | 1.24 | 0.69 | 0.64 | 0.41 | 0.46 |
| Zircon | 0.09 | 0.23 | P | | P | P | P | P | |
| Chlorite | P | P | | 0.08 | 4.96 | 0.35 | 0.91 | | |
| Tourmaline | P | 0.08 | | | P | | 0.18 | | |
| Titanite | | 0.15 | 0.89 | 0.07 | | | | 0.10 | 0.69 |
| Epidote | | | | | | 0.26 | 0.09 | | |
| Garnet | P | P | | | | | | | P |

*Present in the rock, but not observed in this thin section.

Quartz is always the most abundant constituent. It occurs in fine to medium grains with interlocking boundaries (Figure 5). Its size varies with the texture of the rock and also with the size of the original clastic grains. Many of the smaller grains have recrystallized to about their original sizes. The coarser grains and pebbles have recrystallized to mosaics of smaller grains; only under uncrossed nicols are the outlines of the original grains plainly visible (Figure 6). Much of the quartz shows undulatory extinction.

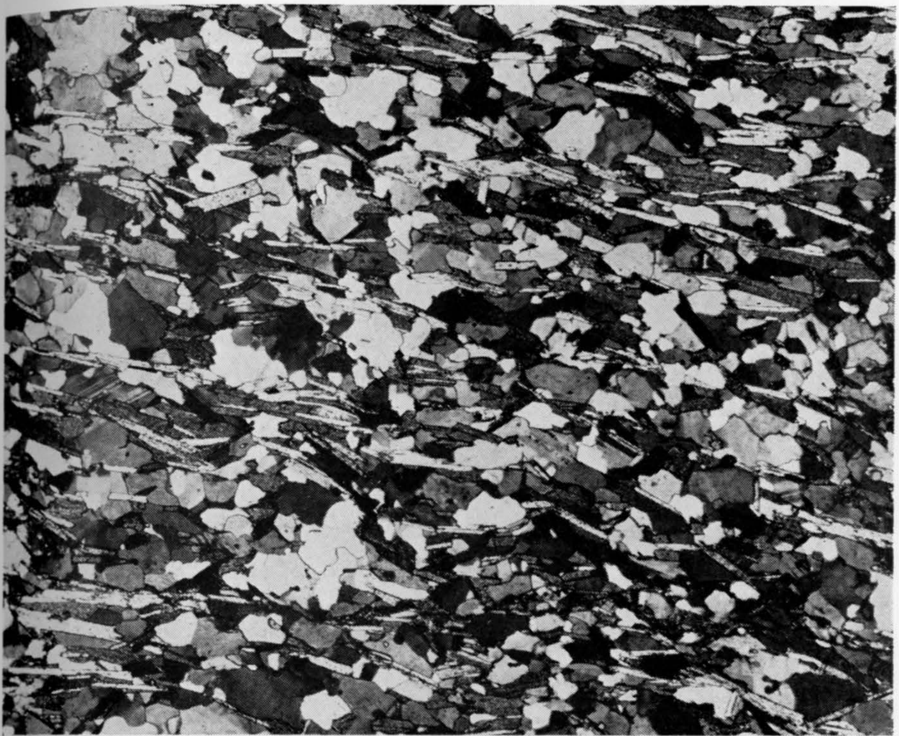
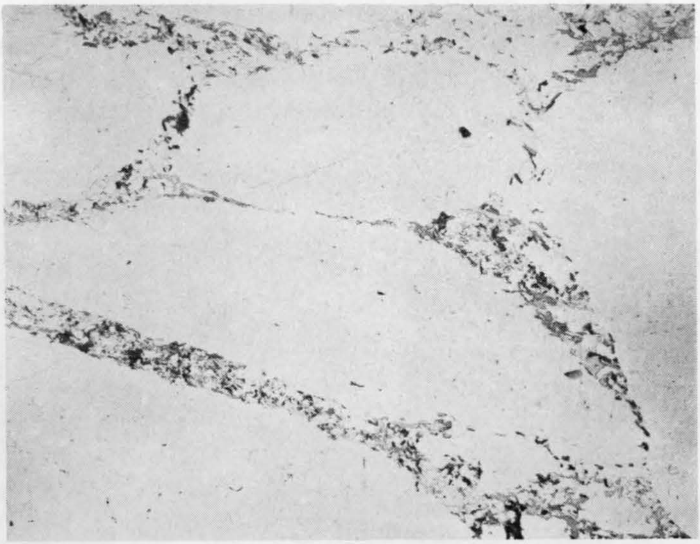
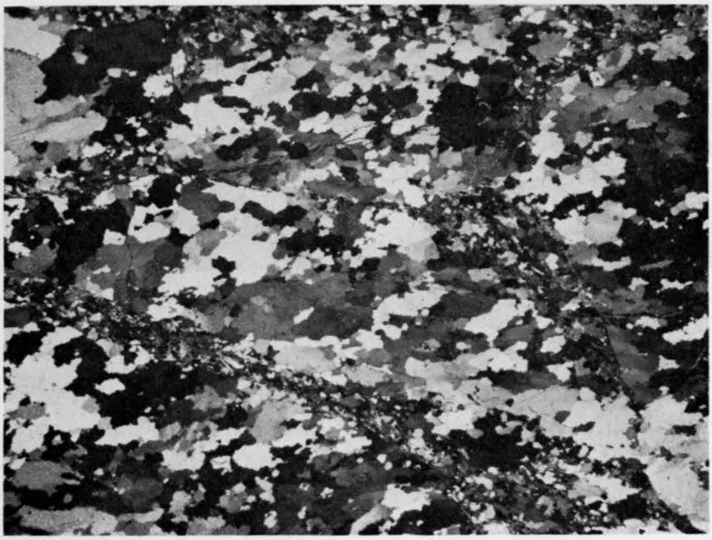


Fig. 5. Microphotograph, crossed nicols, X26. Metagraywacke, composed of quartz, oligoclase, biotite, muscovite, and accessory minerals. (Original grain shapes are visible only with uncrossed nicols.)

The plagioclase is dominantly oligoclase. It occurs in scattered, irregularly shaped grains that vary in size from silt-size particles to the coarsest grains present. Clastic shapes are better preserved by the feldspar than by the quartz. The plagioclase is commonly untwinned. Usually the grains have clouded areas or small patchy inclusions, the coarsest of which are chlorite, epidote, and tiny, black, opaque particles. Andesine is present in some beds, calcic albite in others; however, the anorthite content of the feldspar in a given bed usually varies less than 10%. Such compositional consistency over a large area of graywacke-type metasediments suggests that the feldspar has recrystallized.



A



B

Fig. 6. A, B, Microphotographs of metaconglomerate, X5.
A. Plain light, showing original pebble shapes.
B. Crossed nicols, showing the mosaics of small grains to which the pebbles, once crushed, have recrystallized.

K-feldspar is absent from much of the Copperhill formation, although locally abundant. Fifteen regular thin sections, 20 stained thin sections, and 12 stained rock slabs revealed K-feldspar only in the zone of conglomeratic metagraywacke which lies along and just to the east of Wolf Creek. The number of samples is too small to rule out the presence of K-feldspar in other parts of the Copperhill formation, but is sufficiently large to suggest that K-feldspar is restricted in occurrence and might therefore be used for intraformational correlation. In size and manner of occurrence the K-feldspar is similar to the plagioclase; much of it has the cross-hatch twinning of microcline.

Biotiote is well distributed through the rock in platy crystals, many of which have irregular shapes because of the interference of adjacent grains. In places the biotite plates are interleaved with plates of muscovite, but in general the biotite occurs in separate crystals. 2V ranges from 20 to 25 degrees. Pleochroism is strong, with X = yellowish gray, 5Y 8/2 (recorded range: colorless to 5Y 7/2), and Y = Z = moderate brown, 5YR 4/6 (recorded range: 5 YR 5/6 to 5YR 3/2). Large, dark pleochroic halos are common. The halos surround tiny zircon crystals, many of which are metamict, and small titanite crystals. The biotite is partly altered to chlorite.

Muscovite is nearly always less abundant than biotite, but is similar to the biotite in size, shape, orientation, and distribution.

At a few places the micas are unusually coarse: they are sieved porphyroblasts rather than small clean books, and are set in a matrix of quartz and feldspar that is also unusually coarse. At such places, recrystallization has gone far enough to obliterate all trace of clastic texture.

Calcite occurs sparingly in the metagraywacke, comprising up to 5% of the rock. It is twinned and usually irregular in shape, but in the strongly foliated rocks it may show flattening in the plane of the foliation. Usually it is disseminated through the rock, in contact with all the other minerals, and appears to be a primary constituent. At a few places, however, it has a noticeably spotty distribution and appears to be replacing other minerals. At such places, sulphides are most abundant in and near the calcite, and the feldspar and biotite near the calcite are more strongly altered than in other parts of the rock.

Small irregularly-shaped grains of sulphide are nearly always present, but rarely constitute more than 2% of the rock. The usual sulphide is pyrrhotite. Occasionally it is pyrite. Sometimes pyrite occurs with pyrrhotite.

Colorless to pale green chlorite is sparingly present in nearly all the metagraywacke, as well as in the other rocks. It occurs in books and shreds and in porphyroblastic aggregates of books. In contrast to the other micas, it shows little preferred orientation, except where it pseudomorphs biotite. Usually it is an alteration product of biotite. Often "dusted" through the chlorite is very fine opaque matter, presumably iron that was originally in the biotite and was not taken up by the pseudomorphing chlorite. The chlorite is pleochroic from colorless to pale blue-green, 10GY 6/2 or 5G 7/2. It is optically negative, with very small 2V, and low birefringence.

Tourmaline is usually present in the metagraywacke, as well as in all the other rocks. It occurs in scattered euhedral crystals which vary much in size but are rarely more than ½mm in diameter. It is strongly pleochroic with $\varepsilon$ colorless and $\omega$ grayish olive green, 5GY 3/2, to dusky green, 5G 3/2.

Other common accessory minerals are zircon, apatite, titanite, ilmenite, and garnet. The zircon is usually rounded and has darkened margins. Where it is in contact with or enclosed by biotite, there is a pleochroic halo. The apatite occurs in small colorless hexagonal prisms. The titanite is in small roundish crystals, some of which are lozenge-shaped. Titanite crystals are sometimes clustered about a black, opaque, leucoxenic mineral of irregular shape, probably ilmenite. The garnet is present as sieved porphyroblasts, usually very small. At some places they are numerous; at other places they are entirely lacking, although still present in the interbedded schists.

Mica schist. The mica schist in the Copperhill formation is composed of the same minerals as the metagraywacke. Differences between the two rocks stem largely from a textural difference in the original sediments. The schists were once finer-grained and contained a higher percentage of clay. Being less competent than the metagraywacke, the schists are strongly foliated and crinkled.

The beds of schist are usually thinner and darker than those of metagraywacke. Most of the schist is sparingly garnetiferous.

Varved schist. In the transition zone between the Copperhill formation and the Hughes Gap formation the schist changes in appearance and composition from the drab mica schist, typical of the Copperhill formation, to a vari-colored, abundantly garnetiferous schist. Concommittantly, thin pseudodiorite beds appear; upward, they become a prominent part of the lithology. Staurolitic schists make their appearance, first sparingly, but with increasing abundance upward. Usually the pseudodiorite is interbedded with the schist; to a less extent it is interbedded with metagraywacke. Near the middle of this transition zone are the "varved" schists illustrated in figure 7. Their total thickness appears to be less than 50 feet. The best exposures are (1) ¼ mile north of the Georgia-North Carolina line, on the west side of Stewart Mountain at an elevation of about 1800 feet and (2) on the southeast side of Stewart Knob, also at an elevation of about 1800 feet. Better exposures are found in the Epworth quadrangle to the southwest, at the same stratigraphic horizon.

The varves are mostly one to two inches thick. In type A, figure 7, fine- to medium-grained quartz-rich layers alternate

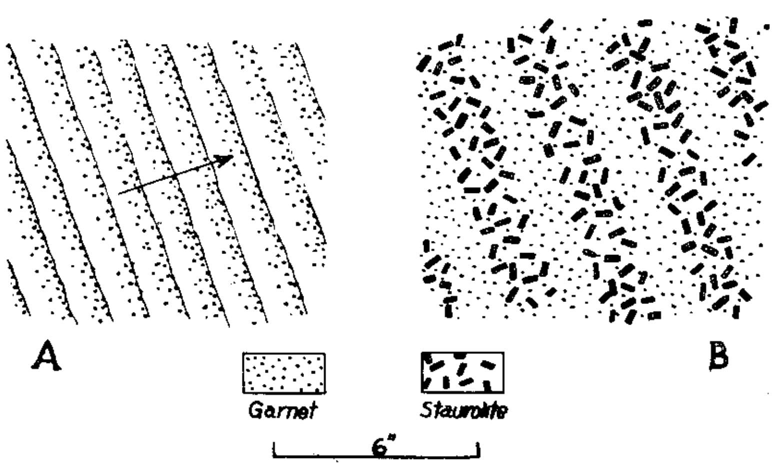



Fig. 7. Varved schists at the top of the Copperhill formation. Arrow indicates top of beds. See text for explanation.

with micaceous layers that contain abundant small garnets. Each garnetiferous layer has a sharp boundary on one side, and grades to a quartz-rich layer on the other side. The quartz-rich layers represent what were originally the coarser, basal parts of the varves; the garnet-rich layers represent the upper, fine-grained, clay-rich portions. These varves are good examples of "reversed" graded bedding (Shrock, 1948, p. 423) and are excellent top-bottom indicators.

In the type B varves, layers of garnet-mica schist alternate with layers of staurolite-garnet-mica schist. The direction of grading in the original sediment has usually been obscured by the staurolite porphyroblasts so that this type of varved schist rarely affords a top-bottom indication.

The varved schists are not continuously traceable for more than a few hundred feet because of poor exposure, but they recur at one stratigraphic horizon for a distance of at least 15 miles.

Metaconglomerate. Although small pebbles of quartz and feldspar occur sporadically throughout the Copperhill formation, thick beds of metaconglomerate are restricted to a few horizons. The metaconglomerate is interbedded with mica schist, and with coarse quartzite and metagraywacke into which it commonly grades.

The pebbles in the metaconglomerate are sub-angular to rounded, and are mostly less than one-half inch in length; rarely they exceed one inch in length. White quartz pebbles predominate, but feldspar pebbles are everywhere present. Locally, the pebbles of feldspar are more numerous than those of quartz. Where feldspar abounds, the pebbles are usually not as well-rounded as where quartz predominates. Pebbles and slabs of black slate up to five inches long occur in the coarsest beds.

The quartz pebbles are usually crushed and flattened parallel to the cleavage planes in the rock. They are also elongate parallel to the axes of the folds. Many pebbles are only one-third as thick as long. The feldspar pebbles are much less deformed.

A few metaconglomerate beds consist almost entirely of well-rounded, almond-size quartz pebbles. Typical beds, however, vary in composition from feldspathic metaconglomerate to conglomeratic metagraywacke. The term "metaconglomerate" is used in Plate 3 as a collective term for all the rocks in the strongly conglomeratic zones.

Metaarkose and quartzite. The metaarkose and the quartzite differ little from the metagraywacke and are hardly separable from it during mapping. Usually they are present not as separate beds but as particularly feldspathic or siliceous portions of metagraywacke beds. They have little significance as distinct rock types except that they are extremes of the compositional variation which characterizes the metagraywacke (see figure 3).

Pseudodiorite. Nodular masses one inch to three feet across consisting mainly of quartz, plagioclase, hornblende, and garnet occur widely in the metagraywacke of the Great Smoky group. They occur mainly along certain horizons as isolated, ellipsoidal masses embedded in the metagraywacke. They may be in any position within a bed, but their longest diameter is always roughly parallel to the enclosing rock's foliation. Some of the masses are irregular in shape, and may even have protuberances which crosscut bedding. Other masses are conformable sheets that range in thickness from a fraction of an inch to three feet and in length from a few inches to 25 feet. The sheet-like masses sometimes appear to have formed by the coalescence of several nodules. Because at a few places these quartz-plagioclase-hornblende-garnet rocks superficially resemble diorite Keith called them pseudodiorite.

The nodular pseudodiorite in the Great Smoky group has been described by LaForge and Phalen (1913, pp. 7-8), Emmons and Laney (1926, pp. 19-21), and Bayley (1928, pp. 118-121). LaForge and Phalen regarded the pseudodiorite as "having been formed in place by a complete recrystallization of portions of the original sedimentary rock." They believed that "percolating solutions played the chief part in the recrystallization." They deduced that the pseudodiorite formed during the closing stages of the last deformation because the nodules are only rarely foliated even when embedded in strongly foliated rocks. Bayley concurred with LaForge and Phalen in the manner of origin, but suggested that the nodules formed "under static conditions during the earlier stages of metamorphism." He explained the lack of foliation in the nodules as due, possibly, to their greater rigidity, which enabled them to resist the deformation to which other rocks yielded. A more definite mode of origin was proposed by Emmons and Laney who concluded that the nodules are metamorphosed calcareous concretions. The Emmons and Laney report contains excellent photographs of the shapes of the nodules and their zonal arrangement of minerals.

Pseudodiorite nodules from other localities have been described by Pettijohn (1940), Eskola (1932), and Sederholm (1928). Pettijohn (1940, p. 1849) has well summarized the arguments in favor of such structures being metamorphosed calcareous concretions: (1) their nodular form, (2) zonal structure, (3) occurrence in metaarenaceous rocks, (4) association of bytownite and other lime silicates with quartz, and (5) their high lime content.

Ross has stated (1935, p. 22) that at least part of the pseudodiorite in the Copperhill formation is "the result of thorough local recrystallization of the schistose county rock, with the addition of some new material from the outside." His principal argument is that some of the pseudodiorite occurs in vein-like, cross-cutting masses. As shown by the chemical analyses in Table 2, lime is the principal constituent that would have to be added to give the country rock the same composition as pseudodiorite. The nodules are more abundant in the Ducktown Basin, where carbonate veins occur, than away from the Basin at the same stratigraphic horizon. This abundance of nodules where lime is known to have been added is suggestive, but hardly impairs the evidence that the nodules are metaconcretions.

Bedded pseudodiorite. In the transition zone between the Copperhill formation and the Hughes Gap formation, pseudodiorite beds a few inches to 50 feet thick are interstratified with schist and metagraywacke. Similar pseudodiorite beds are common in the Hughes Gap and Dean formations. Mineralogically and texturally this bedded pseudodiorite is indistinguishable from the nodular pseudodiorite in the Copperhill formation. This bedded pseudodiorite, originally a sediment, is described in a later section.

**Thickness.** Accurate measurement is precluded by the scarcity of distinctive horizons and by structural complexity. The thickness is estimated to be 2000-5000 feet.

**Alteration.** Mineralogical alterations usually interpreted as the effects of retrogressive metamorphism are apparent in all

**TABLE 2**

**Chemical Analyses of Pseudodiorite Nodules and Metagraywacke**

| | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| $SiO_2$ | 61.77 | 73.90 | 68.13 | 66.65 |
| $Al_2O_3$ | 11.06 | 11.94 | 13.88 | 16.78 |
| $Fe_2O_3$ | 0.64 | 0.38 | 0.77 | 0.36 |
| FeO | 2.88 | 3.72 | 3.83 | 2.93 |
| MgO | 0.86 | 1.34 | 1.68 | 1.86 |
| CaO | 12.62 | 3.09 | 9.04 | 3.25 |
| $Na_2O$ | 2.25 | 2.84 | 0.62 | 4.29 |
| $K_2O$ | | 1.08 | 0.08 | 1.69 |
| $H_2O^-$ | 0.06 | 0.16 | 0.04 | 0.04 |
| $H_2O$ | 0.70 | 0.90 | 0.50 | 0.87 |
| $TiO_2$ | 0.64 | 0.65 | 0.70 | 0.39 |
| $P_2O_5$ | 0.18 | 0.12 | 0.16 | 0.11 |
| $CO_2$ | 6.01 | 0.14 | 0.27 | 0.49 |
| MnO | 0.30 | 0.10 | 0.19 | 0.07 |
| S | | 0.08 | | |
| Totals | 99.97 | 100.44 | 99.89 | 99.78 |

1. Pseudodiorite from Burra Burra mine, Ducktown district, Tenn. (after Emmons and Laney, 1926, p. 21, average of Nos. 3 & 4.)
2. Metagraywacke matrix of (1). (Idem, No. 5.)
3. Pseudodiorite from Abram series, Thunder Lake, Ontario (after Pettijohn, 1940, p. 1849, A.)
4. Paragneiss matrix of (3). (Idem, B.)

the rocks of the Copperhill formation: The principal alterations are sericitization of feldspar and chloritization of biotite.

## Hughes Gap formation

**Name and distribution.** The Hughes Gap formation crops out in a 0.9-1.2 mile wide belt that enters the Mineral Bluff quadrangle near Franklin Mtn and leaves the quadrangle in the vicinity of Gravelly Gap. Good exposures of the formation may be seen along the road that parallels the lower course of Sugar Creek, along the road between Union and the mouth of Hothouse Creek, and along the road which crosses the southern end of Stewart Mtn. In Hughes Gap, after which the formation is named, almost a complete section is exposed.

**Character.** The Hughes Gap formation is characterized by garnet-mica schist, staurolite schist, metaquartzconglomerate, quartzite, and pseudodiorite. In addition to these rocks, there are beds of metagraywacke and mica schist, particularly near the base of the formation. The strata range in thickness from a fraction of an inch to more than 50 feet. They are mostly less than 10 feet thick. Changes in thickness and lithology are common along strike.

Garnet-mica schist. Garnet-mica schist, interstratified with mica schist, quartzite and pseudodiorite, makes up thick zones in the Hughes Gap formation.

The principal minerals in the garnet-mica schist are muscovite, biotite, quartz, and garnet, which vary in proportion from bed to bed. Average specimens of the schist contain 60-70% mica, 20-30% quartz, and 5-10% garnet. Rarely, garnet constitutes as much as 30% of the rock. The accessory minerals are tourmaline, apatite, zircon, black opaque, and sometimes feldspar. A few spongy or stringy masses of staurolite are sometimes present. By increase in the size and proportion of these staurolite masses, the garnet-mica schists grade into staurolite-mica schists. Epidote is sometimes present when the schist contains little garnet. Chlorite is nearly always present.

The texture of the rock is shown by figure 8. Mica folia bend about garnet porphyroblasts as though thrust aside during the growth of the garnets. This imparts to the rock an "eyed" texture. Quartz is often concentrated in the corners of the eyes.

Muscovite is usually the dominant mica. Its books are six to twelve times as long as wide, and are interleaved with similarly elongate books of biotite. The muscovite is a ferriferous variety with 2V greater than 30°. The biotite is strongly pleochroic: X = very pale yellowish brown (5Y 8/4 to 10YR 8/2); Y = Z = moderate brown (10YR 5/8 to 5YR 4/6). Scattered through the biotite are small, dark, pleochroic halos about minute inclusions which were not positively identified but appear to be zircon.

Fig. 8. Microphotograph, nicols partially crossed, X7. Garnet-mica schist. The voids are marked by a V. The other large, clear areas are andesine grains.

The garnet is almandine. Its index of refraction (1.80) and unit cell dimensions (11.5A) are remarkably constant in all the mica schists. It occurs in pale red subhedral to euhedral metacrysts that rarely exceed three mm. in diameter. The centers of the metacrysts contain abundant quartz inclusions which probably represent material that could not be incorporated in the new-formed garnets. The margins of the crystals are usually clear. During the growth of the garnets, either the proportion of available silica in the groundmass of the rock decreased, or the physical environment changed in such manner that the garnets were able to expell the excess silica in the newly occupied areas. The latter possibility is favored because silica is usually concentrated in the groundmass near the garnets. Black opaque matter similar to that in the groundmass of the rock is included also in the garnets. In the groundmass, the opaque matter is aligned parallel to the foliation planes of the rock. In the garnets, the opaque matter is similarly oriented along planes, but these planes often make an angle with the foliation in the rock showing that the garnets have been "rolled" since their growth.

Most of the quartz is distributed through the rock in small tabular grains. Some of it is concentrated in pressure shadows about garnets, where it is in interlocking grains of irregular shape.

The tourmaline is green and strongly pleochroic. Its prisms are euhedral, mostly less than 0.1 mm. in diameter, and sparingly dispersed. Associated with the tourmaline are minute, colorless hexagonal apatite prisms.

The feldspar occurs in scattered, twinned grains many of which show crushing and lensing along the foliation planes. Commonly the feldspar is altered along fractures or cleavages. It ranges in composition from oligoclase to andesine.

Chlorite is nearly always present. It is pale green, faintly pleochroic, nearly isotropic. Occasionally it is interleaved with muscovite and biotite as though of simultaneous crystallization, but most of the time it is clearly an alteration product of the other micas. The chlorite pseudomorphs after biotite contain dark, pleochroic halos like those found in unaltered biotite. Commonly, the chlorite is porphyroblastic.

Small platelets, elongate grains, and irregularly shaped masses of black opaque matter are scattered throughout the schist. Occasionally crystal outlines are seen; rarely there are minute, geniculate twins. The black opaque matter occurs both interstitially and as inclusions in all the minerals, except the tourmaline, apatite, and feldspar. Most of it dissolves in HF acid and is therefore not carbonaceous matter. Mostly it is non-magnetic. An X-ray powder pattern of one hand-picked sample showed that part of the black opaque is rutile.

Staurolite schist. The staurolitic beds are restricted to certain zones in the Hughes Gap formation. Where zone boundaries transgress bedding, they do so by the interlensing of staurolitic and non-staurolitic beds. Within the zones, staurolite schists are interbedded with metaquartz conglomerate, quartzite, and with schists that differ from the staurolitic schists only by the absence of staurolite. The proportion and number of staurolite beds vary greatly from place to place. As an average, 30% of the beds contain staurolite. Usually the proportion appears higher because the staurolite crystals, being resistant to weathering, accumulate on and blanket the surface.

The staurolite-bearing schists include quartz-mica schist, garnet-mica schist, graphitic mica schist, sericite schist, and gradational types. At any one locality the staurolite may be found predominantly in one type of schist, or in two or more types that are interbedded with each other and with rocks that lack staurolite. The thickness of the beds is mostly less than three feet, rarely greater than ten feet.

The staurolite crystals vary in number, size, and color from bed to bed. They vary also in their degree of euhedralism, crystal shape, frequency of twinning, proportion of different twins, and degree of alteration; such variations are more noticeable, however, from area to area than from bed to bed. The usual color of the staurolite crystals, where they are not


Fig. 9. Microphotograph, plain light. X7. Staurolite and garnet porphyroblasts in a matrix of biotite (dark gray), muscovite (lighter gray), and quartz (lightest gray); voids are marked by a V.

altered, is dark red-brown to almost black. They are mostly one-half to two inches long and three-eights to three-fourths inches thick. They range from this size down to crystals too small to be recognized with the unaided eye. Regardless of size, the crystals are mostly well-formed. Only where the amount of included material exceeds 50% of the volume of

Fig. 10. Microphotograph, plain light X7. Staurolite, garnet, and biotite porphyroblasts in a fine matrix of muscovite and quartz.

the crystals do the boundaries become so notched that crystal outline is impaired. Most of the crystals are pseudo-hexagonal prisms, flattened perpendicular to the a-axis, and terminated by the basal pinacoid. At some localities the flattening is perpendicular to the b-axis. The frequency of twinning and the proportion of different twins are erratic. Generally, 15-75% of the crystals are twins; the 60°-twins are 10-50 times as frequent as the 90°-twins; and less than one out of every 500 twins is a trilling.

In the beds which have not undergone post-crystallizational movements, the rock matrix abuts without distortion against the staurolite porphyroblasts, which are randomly oriented. In most localities, however, there have been post-crystallizational movements. The micas in the matrix have been more or less plastered about the staurolite porphyroblasts, giving the rock a knotty appearance. The staurolite crystals have been partly oriented with their longest dimension in the plane of the foliation of the rock. Thin sections of such rocks often show that the inclusions in the staurolite crystals define a strong internal foliation which is at an angle with the foliation in the enclosing matrix, evidence that the crystals have undergone rotation since their growth.

Inclusions are very abundant in the staurolite crystals. The principal inclusion is quartz. It occurs in irregularly-shaped, elongate grains of all sizes up to about 0.6 mm. by 0.15 mm., and may occupy more than 50% of the volume of the staurolite. Slightly larger muscovite books constitute up to one percent of the included matter. The larger staurolite crystals often include small garnets. The garnets may be small or absent in the center of the staurolite crystals, but large and abundant in the margin. Other inclusions are black, opaque platelets or elongate grains, some of which are altered or opaque tourmaline. In the darkest staurolite schists, the opaque matter is largely carbonaceous. The optical properties of the staurolite vary from crystal to crystal, thus indicating some variation in chemical composition. The variation is greatest where crystals are few and poorly-formed and is least where the beds are thick and have a uniform texture. The staurolite is biaxial positive, with a measured range in 2V of 76°-88°. Pleochroism is moderately strong and slightly variable. Usually X is very pale yellowish gray (5Y 8/2); Y is pale yellowish gray (5Y 7/2); and Z is moderate yellow (5Y 7/6). Absorption: $Z > Y > X$.

The matrix of the staurolite schists consists mainly of mica and quartz. The mica is mostly a 2M muscovite whose books average about 0.03 mm. long by 0.08 mm. thick. The matrix is coarser when it is composed of biotite, muscovite, and quartz than when it is composed of quartz and muscovite alone (compare figures 9 and 10). In many of the schists, the matrix is fine muscovite and quartz, and the biotite occurs in porphyroblasts comparable in size to the garnet and staurolite (Fig. 10). Small, green tourmaline crystals are liberally sprinkled through the matrix. They are mostly less than 0.2 mm. long by 0.04 mm. thick, but are sometimes large enough to be easily seen by the unaided eye. They are strongly pleochroic from colorless to dusky green.

The staurolite schists are generally altered throughout the quadrangle. Locally, the staurolite crystals are completely altered to sericite pseudomorphs. The biotite is partly altered to a green pleochroic chlorite which is optically negative, and has a very small 2V. The garnet is slightly chloritized along fractures and crystal margins.

Bedded Pseudodiorite. The name pseudodiorite (Keith, 1913) has been applied both to the quartz-plagioclase-hornblende-garnet nodules in the Copperhill formation and to certain bedded rocks that are lithologically similar in the Hughes Gap and Dean formations. In both usages the term is a misnomer, because the rocks bear little resemblance to diorite; however, the term has been widely used, and a better term has not been suggested. For these reasons its use is continued.

The pseudodiorite in the Hughes Gap formation occurs in beds (see figure 11) that are mostly less than three feet thick, but rarely attain a thickness of more than 50 feet. It is interstratified with mica schist, garnet-mica schists, quartzite, staurolite schist, and occasionally with metagraywacke. The bedding planes are sharp; however, variations in texture and mineralogy are conspicuous within the beds. For the most part, these variations are planar and parallel to the bedding, and reflect differences in composition and grain size in the original sediment. Relic structures which prove the sedimentary origin of the bedded pseudodiorite are described on page 70.

The pseudodiorite is composed essentially of quartz, plagioclase, and hornblende. Epidote and garnet are nearly always present, and commonly are major constituents. The relative amounts of these five minerals differ greatly from bed to bed, as well as within the beds. Other minerals found in the pseudodiorite are listed in Table 3.

The quartz occurs in small interlocking grains that are irregular in shape and very variable in size. The plagioclase is in similar, but usually smaller, grains. Quartz and plagioclase constitute the light-colored, fine-grained matrix through which the porphyroblasts of hornblende and garnet are scattered. The plagioclase is commonly untwinned; when twinned, the lamellae are very fine. It is optically negative, with 2V near 80° and variable. The composition of bytownite is indicated both by the orientation of the optical indicatrix with respect to cleavage poles (Emmons composition curves, 1942) and by a 42°-45° maximum extinction in the zone perpendicular to (010).

## TABLE 3

### Modal Analyses of Typical Pseudodiorite Specimens

| Specimen number | 3HO | 3G3 | 1I9 | 1I1 |
|---|---|---|---|---|
| Quartz | 41.36 | 27.61 | 21.26 | 49.75 |
| Feldspar | 31.79 | 43.30 | 22.91 | 25.30 |
| Hornblende | 12.07 | 19.81 | 12.44 | 15.89 |
| Clinozoisite | 4.00 | 2.06 | 39.53 | 4.39 |
| Garnet | 4.00 | 3.61 | 1.97 | 1.58 |
| Titanite | 2.43 | 0.15 | 1.89 | |
| Calcite | 1.57 | 0.51 | | |
| Apatite | 0.57 | 0.59 | | 0.29 |
| Colorless mica | 0.29 | 0.59 | | 0.79 |
| Black opaque | 1.71 | 1.62 | | 1.22 |
| Chlorite | 0.21 | | | 0.65 |
| Zircon | | 0.15 | | 0.14 |

Randomly scattered through the quartz-feldspar matrix are sieved hornblende porphyroblasts whose length is usually less than 2mm., but may exceed one cm. Inclusions are num-

Fig. 11. Pseudodiorite (slightly boudinaged) interbedded with garnet-mica schist and staurolite-garnet-mica schist. Scale given by pick.

erous, sometimes constituting as much as one-third of the porphyroblast. Quartz is the principal inclusion, even when the matrix is dominantly feldspar. Minor inclusions are carbonate, a black opaque mineral, and rounded zircon. Clino-

Fig. 12. Metaconglomerate. The quartz pebbles are elongate in b, and are flattened in the plane of the cleavage, which is approximately at right angles to the plane of the picture.

zoisite and chlorite are found in the hornblende as alteration products. The hornblende is biaxial negative; 2V ranges from 78°-86°; the extinction angle $Z\wedge c$ ranges from 10°-16°. Pleochroism: X = grayish yellow-green (5GY 8/2) to pale olive (10Y 6/2); Y = dusky yellow-green (5GY 4/4) to grayish green (10GY 6/2); Z = pale green (5G 6/2) to moderate yellowish green (10GY 6/4). Absorption: $X<Z<Y$. These optical properties indicate a member of the actinolite-ferrotremolite series (Winchell, 1951, p. 433).

The garnet porphyroblasts are sieved and are about the same size as the hornblende porphyroblasts. Quartz is again the dominate inclusion, at times constituting as much as 50% of the porphyroblast. Other inclusions are carbonate, a black opaque mineral, and occasionally clinozoisite. The garnet is pale pink. It has a variable index of refraction, near 1.78.

The clinozoisite is mostly in prismatic crystals scattered through the matrix. In some rocks, it is a major constituent, and occurs in coarse, roughly equidimensional grains. Where it is an alteration product of hornblende, it occurs in irregular masses, some of which pseudomorph the sieved hornblende.

The clinozoisite is biaxial positive, with 2V greater than 80°. Its interference color is an anomalous blue. In plain light it is colorless.

The titanite is irregular in shape, leucoxenic, and scattered through the matrix. The carbonate grains are very irregular in shape and often twinned. Where hornblende has altered to clinozoisite, carbonate is often closely associated as though a secondary mineral. The zircon is well rounded, sparingly present. A pleochroic halo surrounds it when it is included in hornblende. Other minor constituents are apatite, which is in small hexagonal prisms, a black opaque mineral, and a colorless mica. The black opaque mineral, being a bronzy color in reflected light, is probably pyrite; its shape is irregular. Colorless to pale green chlorite, often in sheaf-like aggregates, is associated with hornblende and garnet as an alteration product; usually it is confined to fractures or to crystal boundaries. The colorless mica occurs in patches that are most abundant where hornblende has altered to clinozoisite. The patches are resolved to aggregates of tiny grains, or books, at high magnification. They have high birefringence (second order red and green), and are probably talc.

Accessory tourmaline is only locally present in the pseudodiorite.

Origin of the pseudodiorite. Local cross-bedding (see page 70) and occasional, tiny, well-rounded, zircon crystals show that the pseudodiorite is, at least partly, a metamorphosed sediment. The fine texture of the pseudodiorite's matrix and the absence of recognizable detrital shapes among the quartz and feldspar grains imply that it was originally fine-grained. (Detrital shapes are common in rocks interbedded with the pseudodiorite, also in the Murphy marble.)

In table 4, the bedded pseudodiorite is compared with bedded chert and average shale. Compositionally, the pseudodiorite differs from average shale mainly by a higher content of silica and lime, from bedded chert mainly by a higher content of alumina.

The bedded pseudodiorite in the Mineral Bluff quadrangle is probably metamorphosed argillaceous chert and/or calcarceous shale.

## TABLE 4

### Chemical Composition of Bedded Pseudodiorite, Bedded Chert, and Average Shale

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 |
|---|---|---|---|---|---|---|---|
| $SiO_2$ | 68.50 | 67.72 | 62.43 | 56.58 | 74.04 | 74.56 | 58.38 |
| $Al_2O_3$ | 14.64 | 12.29 | 15.57 | 18.56 | 9.88 | 3.23 | 15.47 |
| $Fe_2O_3$ | 3.79 | 0.66 | 0.49 | 2.93 | 0.44 | 5.23 | 4.03 |
| FeO | 3.77 | 2.87 | 4.02 | 2.05 | 3.48 | 0.00 | 2.46 |
| MgO | 1.14 | 1.80 | 2.93 | 1.72 | 2.56 | 3.60 | 2.45 |
| CaO | 6.46 | 11.00 | 12.00 | 16.13 | 7.91 | 6.26 | 3.12 |
| $Na_2O$ | 0.92 | 0.49 | 0.67 | 0.31 | 0.37 | | 1.31 |
| $K_2O$ | 0.55 | | | | | Trace | 3.25 |
| $H_2O$ | | 0.34 | 0.41 | 0.95 | 0.49 | | 5.02 |
| $TiO_2$ | 0.25 | 0.99 | 0.06 | 0.77 | | Trace | 0.65 |
| $P_2O_5$ | 0.00 | 0.25 | 0.25 | | 0.12 | 0.00 | 0.17 |
| $CO_2$ | | 0.69 | 0.22 | | | 5.85 | 2.64 |
| $SO_3$ | | 0.90 | 0.85 | | 0.63 | 1.30 | 0.65 |
| $ZrO_2$ | | | 0.10 | | 0.08 | | |
| C | | | | | | | 0.81 |
| Total | 100.02 | 100.00 | 100.00 | 100.00 | 100.00 | 100.03 | 100.41 |

1. Pseudodiorite specimen 3HO*. Analyst: L. H. Turner, Chief Chemist, Georgia Geological Survey.
2. Specimen 3HO*, composition calculated from a point count analysis.
3, 4, 5. Pseudodiorite specimens 3G3*, 1I9*, and 1I1*, respectively; compositions calculated from point count analyses.
6. Bedded chert, Fort Payne formation, Georgia (Hurst, 1953, p. 237, No. 1).
7. Average shale (Clarke, 1924, p. 631).
*See Table 3 for modal analysis of this specimen.

Metaquartzconglomerate. The conglomeratic zones in the Hughes Gap formation are usually about 50 feet thick, and contain more or less quartzite, as well as some schist, in addition to metaconglomerate. The thickest zones are 200-300 feet thick, and are usually traceable less than one-half mile. They lense-out by interleaving with beds of schist and quartzite. Some of the zones end abruptly.

Although some of the metaconglomerate in the Hughes Gap formation, particularly in the lower part of the formation, is similar to the metaconglomerate in the Copperhill formation, most of it is decidedly different. Whereas the metaconglomerate in the Copperhill formation is characterized by poor sorting, abundant feldspar pebbles, and a graywacke-type matrix, the metaconglomerate in the Hughes Gap formation is characterized by the preponderance of quartz pebbles, by a sandy matrix, by interbedding with orthoquartzite, and by occurrence in lense-shaped or more irregularly-shaped deposits. The quartz pebbles are well-rounded, sized and mostly less than an inch in diameter. The Hughes Gap metaconglomerate is typically oligomictic (Pettijohn, 1949, pp. 207-208) in contrast to the polymictic metaconglomerate of the Copperhill formation. The two types give evidence of very different conditions of transport and deposition.

Unusual pebbles occur in the metaconglomerate which crosses Hothouse Creek at the northeast base of Franklin Mtn, just north of the Mineral Bluff quadrangle. The pebbles are composed of quartz and tourmaline, with subordinate biotite and muscovite. The tourmaline, in grains and euhedral crystals of all sizes up to one mm. long, is sprinkled without noticeable orientation through the quartz, and constitutes up to 40% of the pebble. The apatite in the apatite-quartz pebbles is similar in size, distribution, and amount.

The pebbles at most localities have been elongated parallel (Fig. 12) to the local fold axes and flattened in the plane of the flow cleavage. Their rounded, sized, and sorted character is best seen at the localities where distortion has been slight.

Under the microscope, the pebbles are seen to be mosaics of small quartz grains which have interlocking or even sutured boundaries (see figure 6). The matrix is fine quartz and sericite, or quartz, biotite, and muscovite. Where the metaconglomerate tends toward the polymictic type, fine feldspar, often of more than one variety, as orthoclase and oligoclase, occurs also in the matrix. Small, well-rounded grains of zircon are locally abundant.

The metaquartz conglomerate readily disintegrates at the surface. The quartz pebbles weather free of the matrix and break down to a friable sand which is ideal for road base, backfill, and similar purposes.

Quartzite. The purest quartzites in the Hughes Gap formation are associated with metaquartz conglomerate (see figure 13). The beds rarely exceed five feet in thickness. The grain size ranges from fine to coarse. The largest grains have been deformed like the pebbles in the metaconglomerate.

Most of the quartzites are impure. Some of them contain hornblende and garnet as accessory minerals, others contain biotite, or biotite and muscovite. Their varietal minerals are usually the same as the essential minerals in the associated beds.

Quartz-kyanite schist. Along the east side, or top, of the Hughes Gap formation there is a band of quartz-kyanite schist which can be followed, mainly by float, for a strike distance of about eight and one-half miles. The band has a maximum thickness of three feet. About 600 feet southeast of Union and about one-fourth mile northeast of where the band crosses Mill Creek, the band is an aggregate of randomly oriented, blue bladed kyanite crystals one to four inches long. At most other places the band is composed of quartz and kyanite, or is composed mainly of quartz, resembling a vein, in which there are masses of coarse kyanite. The kyanite blades are interleaved with plates of 2M muscovite, which is possibly an alteration product. Some of the compact kyanite masses

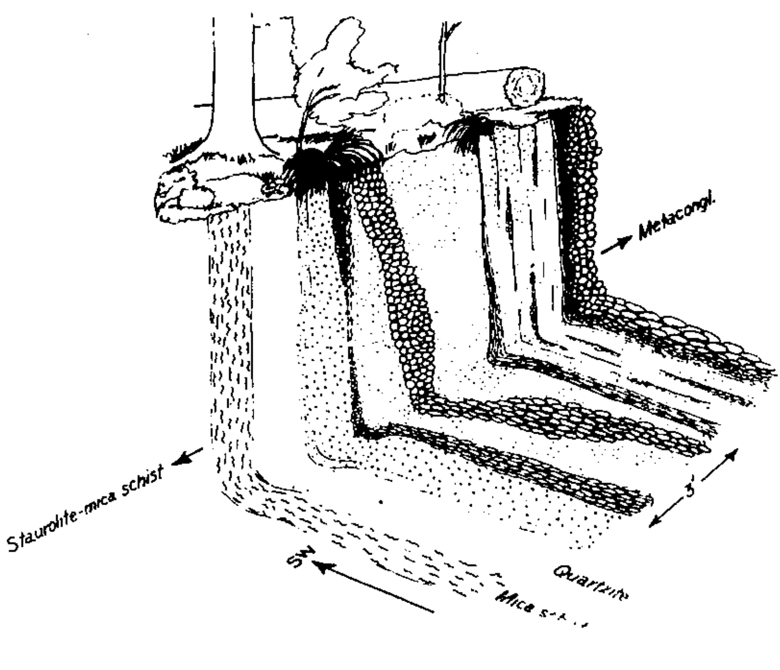



Fig. 13. Perspective sketch of road cut. Interbedded quartzite, metaconglomerate, and mica schist in the Hughes Gap formation. Top of beds to SE. (Size of pebbles exaggerated.)

contain small, poorly-formed corundum crystals. Because exposures are poor, little could be learned about the relationship between the quartz-kyanite schist and the silvery mica schists associated with it. However, the kyanite band is conformable to the bedding, and occurs at one stratigraphic horizon for a strike distance of more than eight miles; it is believed to be a metasediment.

**Thickness.** The present outcrop pattern is due partly to facies changes in the original sediments and partly to local structures. Because the effects of one are not always distinguishable from the effects of the other, the thickness of the formation could not be accurately measured. However, it appears that the repetition of beds is not common. An estimate of the thickness can therefore be obtained by multiplying the width of the formation outcrop by the sine of the average angle of dip. Such an estimate is 4000-6000 feet.

## HOTHOUSE FORMATION

**Name and distribution.** The Hothouse formation crops out in two belts. The largest is 1.8-2 miles wide; it trends northeast-southwest across the central part of the quadrangle. The second belt, parallel to the first, crosses the southeastern corner of the quadrangle. Each belt is on a limb of the Murphy syncline. The name of the formation is taken from Hothouse Creek, which flows for six miles along the central part of the larger belt.

**Character on the west limb of the syncline.** The principal rock types are metagraywacke and mica schist, which are interbedded throughout the formation. The other major rock types are metaconglomerate and quartzite.

The base of the formation is characterized by the predominance of mica schist. The thickness of the schist beds is usually less than two feet but sometimes more than 25 feet. The schist is fine-to medium-grained, and consists mainly of muscovite, biotite, and quartz. The mica content ranges from 40% to more than 90%, with muscovite the dominant variety. The biotite is typically red-brown. However, it has an atypical color when epidote is present. It is then strongly pleochroic in olive or green: X is pale yellow (5Y 8/2-7/4); Y and Z are olive (10Y 4/4) to greenish black (5G 2/1). The color of the biotite and its high 2V indicate that it is richer than usual in iron. Epidote and feldspar are common accessory minerals. The epidote is optically negative with a 2V of 70°-74°, hence also an iron-rich variety (Winchell, 1951, pp. 448-49). The feldspar is usually oligoclase or andesine. Potash feldspar occurs in some beds. Almandine garnets are sparingly present in crystals less than one-eighth inch in diameter. Typical textures of the schist are shown in figures 14 and 15. Modal analyses of epidote-bearing mica schist are listed in Table 5.

**TABLE 5**

**Modal Analyses of Epidote-Bearing Mica Schist in the Hothouse Formation**

| Specimen number | 2F1 | 3R4 |
|---|---|---|
| Quartz | 21.67 | 30.75 |
| Muscovite | 35.78 | 46.13 |
| Biotite | 23.03 | 10.15 |
| Feldspar | 7.17 | 3.92 |
| Epidote | 11.64 | 4.32 |
| Black Opaque | 0.71 | 4.42 |
| Tourmaline | P* | 0.10 |
| Apatite | P | 0.05 |
| Garnet | | 0.16 |

*Present in the rock, but not observed in this thin section.

The schists weather to fine, micaceous saprolite in which adjacent beds often have different colors. Dark yellowish orange (10YR 6/6), grayish red purple (5RP 4/2), and pale reddish brown (10R 5/4) are typical colors. This thin-banded, varicolored saprolite, which is very slippery when wet, is a characteristic feature of the base of the Hothouse formation.

The middle part of the formation consists mainly of interbedded metagraywacke and mica schist. The beds vary greatly in thickness from place to place, but tend to thicken upward in the section. This part of the formation is lithologically indistinguishable from the Copperhill formation.

The upper part of the formation is characterized by the abundance of metaconglomerate and quartzite and by the presence of a few thin beds of staurolite schist.

**Character on the east limb.** On the east limb of the Murphy syncline, the Hothouse formation looks different, although composed of the same rock types. The difference results from a coarser crystallization of the micaceous beds and from intense post-crystallization alteration. Even though both of these effects are discernable in the trough and on the west limb of the syncline, only on the east limb are they intense enough to greatly affect the megascopic appearance of the rocks. Both effects are discussed in later sections.

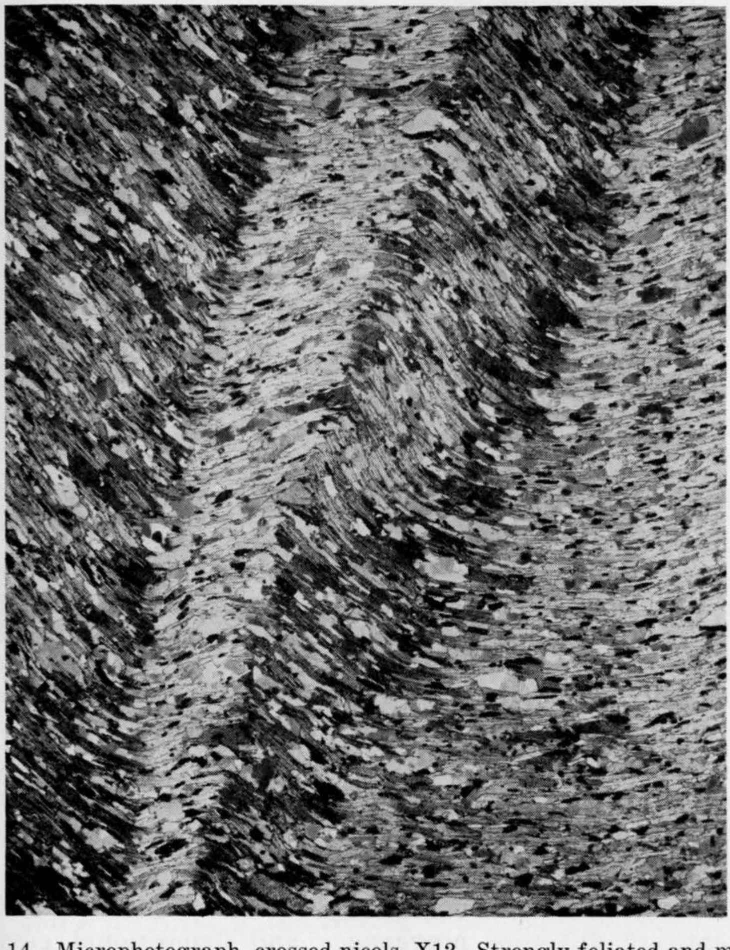


Fig. 14. Microphotograph, crossed nicols, X12. Strongly foliated and mildly crinkled quartz-mica schist.

**Thickness.** The outcrop width of this formation is at least one and eight-tenths miles. The dip of the bedding is steep, and no important repetition of beds was recognized. Accordingly the thickness of the formation is estimated to be 8000-11000 feet.

**Hothouse-Dean boundary.** The metasediments in the lower and middle parts of the Hothouse formation belong to the graywacke suite. They represent incompletely weathered and poorly sorted sediments which are indicative of rapid erosion, transportation, and deposition. When these sediments were produced, either the source area was rugged and had high

Fig. 15. Microphotograph, crossed nicols, X10. Strongly foliated and strongly crinkled mica schist, typical of the Hothouse formation on the west limb of the Murphy syncline.

relief, or the climate was rigorous, because strong indications of weathering are not found. The sediments were deposited in deep water, below wave base, as attested by the abundance of graded bedding and the absence of cross-bedding.

The metasediments in the Dean formation give evidence of an entirely different provenance and depositional environment. The quartzite and metaquartz conglomerate in this formation represent reworked, sized and sorted deposits of quartz, a mineral notably resistant to weathering. The associated staurolite schists represent fine-grained sediments unusually rich in iron and alumina, materials concentrated by

Fig. 16. Microphotograph, plain light, X10. Weakly foliated and uncrinkled mica schist, typical of the Hothouse formation on the east limb of the Murphy syncline.

prolonged weathering. These characters suggest for the sediments of the Dean formation either a source area of low relief or long and slow transportation. Deposition above wave base is indicated by the presence of cross-bedding.

The upper part of the Hothouse formation is a hybrid zone in which metagraywacke and feldspathic metaconglomerate, indicative of rapid erosion and transportation, are interbedded with metaquartz conglomerate and staurolite schist, indicative of weathering and slow transportation or reworking. This interleaving of "incompatible" types records important changes in the areas of supply: either there was a shift in the source, so that sediments began to arrive from new and weathered terrain, or there was a lowering of relief in the old source area and a decrease in the rate of deposition. In either case, the depth at which the sediments accumulated gradually decreased.

The hybrid character of the rocks between the Hothouse formation and the Dean formation makes the position of the boundary between them largely interpretative. The boundary has been drawn to include most of the hybrid rocks in the Hothouse formation and to make the Dean formation a clear-cut lithologic unit.

The hybrid zone changes in character along strike within the Mineral Bluff quadrangle. It is anticipated, therefore, that the relative thickness of the Hothouse and Dean formations and the manner of transition from one to the other may be radically different in other areas.

## DEAN FORMATION

**Distribution, name, and thickness.** The Dean formation crops out in a one-half mile wide belt on each limb of the Murphy syncline. The western belt extends along the west side of Dean Ridge, after which the formation is named, through Windy Ridge, and along the upper course of Sugar Creek. The eastern belt lies just to the west of Lake Toccoa. The thickness of the formation is 2500-3500 feet.

The Dean formation like the Hughes Gap formation, shows marked horizontal variations, abrupt vertical variations, and many breaks in lithologic sequence, some of which might be the result of erosion rather than non-deposition.

The best exposures are along Dean Ridge, north of Mineral Bluff, and along Georgia highway 5, northwest of Blue Ridge.

**Character.** Staurolite-mica schist, X-biotite schist, meta-quartz conglomerate, quartzite, and pseudodiorite are the principal rock types. They are interbedded with less amounts of gray slate of phyllite and sericite schist. Metagraywacke and metaarkose are locally present. Except for the abundance of X-biotite schist, and the absence of quartz-kyanite schist at the top, the Dean formation is lithologically indistinguishable from the Hughes Gap formation.

X-Biotite schist. One of the most distinctive rocks in the quadrangle is X-biotite schist, also called "spangled biotite gneiss" and "speckled schist". The rock has a light-colored, fine-grained, groundmass which is thickly studded with coarse, shiny black to dark brown biotite crystals (Fig. 17). The crystals are commonly one to three mm. across and unlike the usual biotite in metamorphic rocks, are stoutly tabular to equidimensional. Even when tabular, they lie at all angles to the foliation of the rock.

Fig 17. Interbedded pseudodiorite (left) and X-biotite schist (right). About ⅓ natural size.

Microscopically, the biotite is seen to be porphyroblastic. It includes small grains of quartz and is spotted by dark halos about tiny, highly birefringent inclusions. The pleochroism of the biotite at four localities is tabulated below:

1. X = very pale yellow (10Y 8/4)
   Y = light olive brown (5Y 5/6) Absorption: X<Z<Y
   Z = light olive brown (5Y 6/6)

2. X = very pale yellow (5Y 8/4)
   Y = Z = olive brown (5Y 5/4)

3. X = very pale green (5GY 8/2)
   X = Z = olive green (10Y 5/4)

4. X = very pale yellow (10Y 8/4)
   Y = dark yellow brown (10YR 2/4) Absorption: X<Y<Z
   Z = dark yellow brown (10YR 4/10)

The groundmass is a network of fine muscovite, or sericite, and small grains of quartz. The muscovite books are nearly always less than three-tenths mm. long and about one-fifth as thick as long. The quartz grains are irregularly-shaped to elongate, and are about the same size as the muscovite.

Sprinkled through the groundmass and included in the porphyroblasts are euhedral prismatic tourmaline crystals and black opaque, elongate masses. The tourmaline is strongly pleochroic, with $\varepsilon$ colorless and $\omega$ dark green (5GY 3/2 to 5G 4/2); the crystals are usually less than five-tenths mm. long and two-tenths mm. thick. Locally they are more than one mm. thick and constitute one to two percent of the rock. The opaque masses are about the same size as the muscovite. In reflected light, some appear leucoxenic, some bright red-brown, and others black. They dissolve in perchloric acid, and are mostly non-magnetic. Minute apatite prisms and small rounded zircon crystals are the other accessory minerals.

Chlorite is common along the margins of the biotite crystals (Fig. 18) and in microfractures, where it is clearly an alteration product. It is pleochroic in green, optically positive, and has a very small 2V. The chlorite is less ferriferous than the

Fig 18. Microphotograph, plain light, S29. X-biotite schist. Dark biotite porphyroblasts in a fine quartz-sericite matrix. The biotite has partly altered to chlorite (c) which is dusted by opaque matter.

biotite; where it pseudomorphs biotite, it is "dusted" with fine opaque matter which is probably the excess iron.

Most of the X-biotite schist contains euhedral garnet porphyroblasts one to three mm. in diameter. They often have sieved cores, with quartz the most abundant inclusion, and clear margins. Some beds of schist contain staurolite in addition to garnet. The staurolite porphyroblasts range from microscopic size to crystals more than two inches long and one-half inch thick.

The X-biotite schist is generally interbedded with impure quartzite and pseudodiorite.

Since the growth of the biotite porphyroblasts the schist in most localities has been deformed. Biotite folia have been "dragged" along microscopic shear planes, even minutely crinkled. Many of the crystals have undergone rotation. Some of the crystals whose cleavages were about parallel to the plane of shear have been partly "smeared-out". The garnet and staurolite crystals have been crushed. The intensity of the deformation varies greatly from place to place. The intensity of the mineralogical alteration which followed most of the deformation varies similarly. Both deformation and alteration are well displayed in fresh rock just west of the Toccoa dam. At this locality staurolite-biotite schist has been reduced to "greasy" sericite schist. The biotite remnants are still visible as dark chloritic "smears" in the rock, but the staurolite crystals have been altered to fine shimmer aggregates that are indistinguishable from the groundmass except in thin section (Figure 19). Formerly, staurolite schist was as abundant on the east limb of the Murphy syncline as on the west limb, but it is now much less prominent because of the strong alteration. The close similarity between the Dean formation on the east limb and that on the west limb is apparent only after microscopic study.

**Dean-Nantahala boundary.** Keith (1907, p. 4) and LaForge and Phalen (1913, p. 6) mentioned that staurolite schist and metaconglomerate occur in the lowest part of the Nantahala. These beds are now relegated to the Dean formation, and the Dean-Nantahala boundary is drawn at the base of the dark slate.

The contact relations between the Dean and the Nantahala

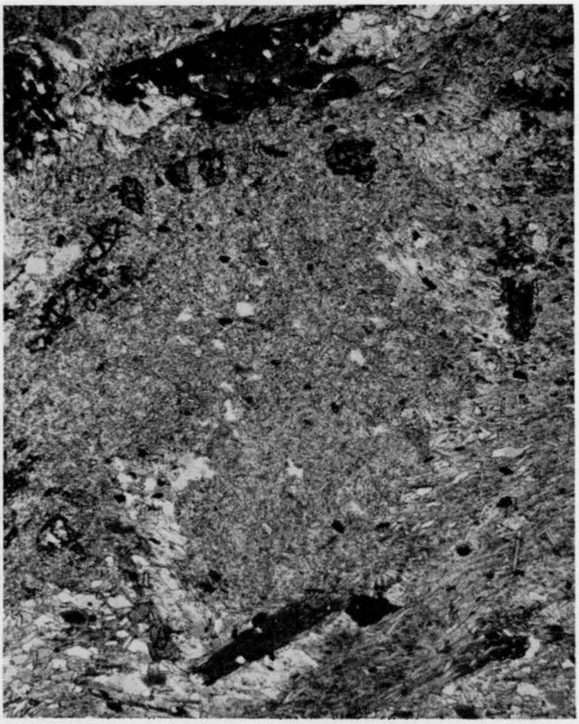


Fig. 19. Microphotograph, plain light, X30. In the center of the photograph is a shimmer aggregate, a mass of very fine sericite with staurolite remnants (dark, ragged grains) along the top and left side. The quartz grains in the shimmer aggregate represent the quartz inclusions that were in the staurolite from which the aggregate formed. The margins of the photograph show the fine quartz-sericite matrix surrounding the altered staurolite crystal. The biotite is marked by "b".

formations were investigated by Nuttall who concluded (1951, pp. 30-31) that the formations are gradational and conformable. Nuttall's conclusions are based on five sections across the Dean-Nantahala boundary in northern Georgia. Two of the sections are in the Mineral Bluff quadrangle.

Detailed mapping along 11 miles of the contact supports Nuttall's findings, though not all of his conclusions. The formations do appear to be gradational through a zone less than 20 feet thick, and concordant. The possibility of their being separated by an unconformity, however, as suggested by Furcron (1953, p. 36-37) is not ruled out. The evidence in favor of an unconformity is (1) the pronounced lithologic change across the contact, and (2) the changing character of the metasediments under the Nantahala. The evidence against a non-conformity is that the formations appear gradational, and concordant. The only evidence against **disconformity** seems to be the lack of conclusive evidence for one. With the Dean-Nantahala contact disturbed by local faulting and obscured by weathering, the presence or absence of a disconformity is hard to establish.

## NANTAHALA SLATE

**Name and correlation.** The formation was named by Keith for its fine exposures along the Nantahala River in Macon and Swain counties, North Carolina. The name was first used in the Asheville folio, 1904. Keith regarded the Nantahala as the metamorphosed equivalent of the Nichols slate.

**Distribution and thickness.** In the Mineral Bluff quadrangle the Nantahala slate crops out in two belts, one on each limb of the Murphy syncline. The northwest belt, topographically expressed as a low ridge, enters the quadrangle between Coles Crossing and Prospect Church. It extends southwestward through Dickey Mountain and along the northwest edge of Blue Ridge. The second belt enters the quadrangle along the high ridge east of Cutcane Creek; to the southwest this band narrows and bends toward the valley of Weaver Creek. The usual thickness of the formation is 1000-1800 feet. It is less than 500 feet thick where thinned by faulting east of Weaver Creek.

**Character.** The formation consists mainly of black to gray banded slates. It contains also a few dark-colored schists or phyllites and thin micaceous quartzites, particularly in the lower half of the formation. The schists are thin-bedded and fine-textured, and may contain small porphyroblasts of garnet and biotite. Ottrelite, mentioned as occurring in the Nantahala farther north (Keith, 1907, p. 4; Van Horn, 1948, p. 6) is very inconspicuous.

The banding in the slates is distinct, with individual bands ranging from paper-thin laminae to beds an inch or more thick. The lighter-colored bands are thicker, in general, than the dark-colored bands, and are more prone to pinch and swell. Some laminae pinch-out within a few inches; others are traceable for tens of feet without perceptible change.

The slates contain much disseminated organic matter, iron oxide, pyrite, and pyrrhotite. Only the iron sulphides are usually visible to the unaided eye. Typically they occur in discontinuous, ragged streaks or "smears" parallel to the foliation. The pyrite is rarely in cubes.

The principal minerals in the slate are biotite and quartz. The biotite is strongly pleochroic with X colorless, Y and Z light yellowish brown (10YR 6/8). It ranges in size from the smallest wisps to books 0.15 mm. long. The quartz grains are elongate and interlocking. The opaque matter is present as "dust", as irregularly shaped patches or streaks, and occasionally as small plates or cubes. Most of the "dust" is graphite. The other opaque matter is pyrite, pyrrhotite, magnetite, ilmenite, and leucoxene. Pyrite and pyrrhotite are largely restricted to the quartz-rich, coarser laminae, when the slate is strongly banded. Muscovite is common in some laminae, absent from others. When present, it is mingled with biotite from which it differs only in color. Small oligoclase grains occur with the quartz in a few of the coarser laminae. Irregularly shaped carbonate is usually disseminated through the fine-grained bands, where it appears to be a primary constituent. It occurs also in ragged patches in the coarse-grained bands. Unlike the biotite, which is elongate parallel to the bedding, the chlorite tends to be aligned parallel to the cleavage. The textural relations indicate that it is the youngest mineral in the slate and that it grew in many cases at the expense of biotite. Green tourmaline prisms up to one-tenth mm. across are scattered through the slate, in company with tiny, colorless apatite prisms. Garnet is sparingly present, very small, and poorly-formed.

The grain size is generally 0.02 to 0.15 mm. in the light-colored bands, mostly less than 0.02 mm. in the darker bands. The light-colored bands are always rich in quartz or in quartz and feldspar. The dark bands are usually rich in biotite or black opaque matter. However, some of the light and dark bands differ not in the proportion of light to dark minerals but in grain size.

The Nantahala slate on the east limb of the Murphy syncline looks the same, megascopically, as that on the west limb. Microscopically, there are differences: on the east limb, the grains are generally a little coarser, the biotite crystals show a stronger tendency to be porphyroblastic, and chloritization is more intense.

## TUSQUITEE QUARTZITE

**Distribution, name, and thickness.** The Tusquitee quartzite crops out in two bands that are roughly parallel, one on each limb of the Murphy syncline. Both bands are locally discontinuous, partly because of faulting and partly, it appears, because of non-deposition. The western band at two places has an unusual outcrop pattern which might be of depositional origin (see plate 3.)

The formation was named by Keith for its exposures in the Tusquitee Mountains of North Carolina. The thickness varies from 0 to 200 feet.

**Character.** The formation consists of white, thin-bedded, fine-to medium-grained, feldspathic quartzite, which is interstratified with paper-thin laminae and thin sandy beds of dark slate (Fig. 20). At a few localities there are thin beds of fine metaquartz conglomerate. The quartzite beds are mostly less than two feet thick, but vary greatly in number and thickness from one locality to another. By an increase in the number and thickness of its slate layers, the Tusquitee grades into the underlying and overlying dark slates.

The mineral composition of typical specimens of Tusquitee quartzite is given in Table 6. The quartz grains are completely recrystallized and usually have interlocking boundaries. The grains that are coarse or surrounded by mica preserve well-rounded detrital shapes. The feldspar grains preserve clastic shapes better than the quartz. The feldspar is often twinned and sometimes finely perthitic. It ranges in composition from albite to oligoclase. Except the fine perthite, no K-feldspar was observed in five stained thin sections. The muscovite, biotite, and chlorite occur as small intergranular books. The opaque matter is fine, intergranular, and partly leucoxenic. The zircon (?) and tourmaline have detrital shapes.

The quartzite in some localities appears to have been affected very little by metamorphism. There is no foliation, and the quartzite crumbles during weathering to a friable sand. In other localities the quartzite has been indurated to a massive, glassy rock with little indication of original sand grains.

Fig. 20. Interlayering of dark slate and feldspathic quartzite in parts of the Tusquitee quartzite. Top of beds to the right. About natural size.

## TABLE 6

### Modal Analyses of Tusquitee Quartzite

| Locality number | 2N8 | 2H6 |
|---|---|---|
| Quartz | 60.36 | 64.66 |
| Feldspar | 20.62 | 16.67 |
| Biotite | 4.44 | |
| Muscovite and chlorite | 14.13 | 18.40 |
| Zircon (?) and tourmaline | 0.09 | |
| Opaque | 0.36 | 0.27 |

## BRASSTOWN FORMATION

**Definition.** The beds between the Tusquitee quartzite and the Murphy marble were divided by Keith (Nantahala quadrangle, 1907) into two formations: The Brasstown schist and the Valleytown formation. Keith described the Brasstown schist as consisting of a lower member of banded slates and an upper member of ottrelite schists and/or banded slates.

He described the Valleytown formation as mainly mica schist and fine banded gneiss. The boundary between the formations is "very difficult to draw"; at places "individual layers of each formation cannot be distinguished from those of the other".

When LaForge and Phalen mapped the Ellijay quadrangle, they adopted Keith's sequence. Their description of the Brasstown schist is the same as Keith's. Their description of the Valleytown formation is "biotite schist, sericite schist, andalusite schist, and fine-banded, somewhat plicated mica gneiss or graywacke, with a few thin beds of quartzite, arkose, and fine conglomerate". In another locality the formation is "a nearly homogeneous mass of sericite mica schist and siliceous slate with some talcose material". In a third locality it is "siliceous mica slate, curly phyllite, or augen gneiss."

Detailed mapping in the Mineral Bluff quadrangle has revealed that Keith's sequence is not entirely applicable. The rocks LaForge and Phalen mapped as the Valleytown formation actually belong to four formations: the Brasstown formation, Andrews formation, Nottely quartzite, and Mineral Bluff formation.

As originally defined, the Valleytown formation is without a definite lower boundary and is lithologically indistinguishable from underlying beds. The original definition has not been improved by subsequent usage. The name is therefore not retained.

The rocks between the Tusquitee quartzite and the Murphy marble all conform to Keith's description of the Brasstown schist. They are renamed the Brasstown formation.

**Distribution and thickness.** Like the other formations in the Mineral Bluff quadrangle, the Brasstown formation crops out in two belts, one on each limb of the Murphy syncline. The western belt extends along the course of Young Stone Creek, Hogback Bend, and through the northwest side of Blue Ridge. The eastern belt is parallel with and just to the east of Cutcane Creek and Weaver Creek. The formation is 1200-1500 feet thick except where thinned by faulting in the vicinity of Weaver Creek.

**Character.** The base of the formation consists of dark gray to black banded slates whose lowest layers are interstratified with the underlying Tusquitee quartzites. Good exposures of the base can be seen along Georgia highway 5 on the northwest side of Blue Ridge, along a country road one-half mile southwest of Hogback Bend, and at the head of Weaver Creek between Blue Ridge and Toccoa Lake. At the last-named locality, which is near a fault, the base is black schist.

The remainder of the formation is dark slate or schist, often banded, in which there are numerous thin, lensing beds of sandy slate and gray quartzite. The top of the formation is calcareous.

The Brasstown rocks on the east limb of the Murphy syncline look coarser and more strongly banded than those on the west limb. The coarser aspect is attributable to abundant small porphyroblasts of biotite and garnet. Their scarcity and abundance in alternate layers emphasize the initial banding. Porphyroblasts are present also in the Brasstown rocks on the west limb, but there they are usually too small to be noticed megascopically. The coarsening of the micas on the east limb is discussed in a later section.

In thin section, the rocks at the base of the Brasstown formation are indistinguishable from the Nantahala slates. The rocks in the middle part of the Brasstown formation are similar, but generally contain less carbonaceous matter and more quartz.

The upper part of the Brasstown formation is well exposed in a quarry on the east side of Creaseman Branch Valley, six-tenths mile northeast of the Toccoa River. Stratigraphically, the quarry is about 125 feet below the Murphy marble. The rock in the quarry is a strongly banded slate or schist, which is sold as flagstone. The bands are generally one-eighth to two inches thick. The darker bands consist of abundant small biotite and garnet porphyroblasts in a fine matrix of quartz, carbonate, and andesine. The lighter-colored bands are composed of fine-grained quartz, andesine, very fine muscovite, carbonate, and sometimes fine biotite. The relative proportions of the minerals vary greatly from band to band. The mineral composition of the rock averaged from four thin sections, is:

| | |
|---|---|
| biotite | 27.74 |
| quartz | 25.86 |
| carbonate | 23.63 |
| andesine | 11.99 |
| garnet | 3.94 |
| muscovite | 3.43 |
| chlorite | 2.05 |
| black opaque | 1.37 |
| | 100.00% |

The biotite is coarse in the matrix, and usually fine in the garnets. It is pleochroic from colorless to moderate brown (5YR 5/6 to 5 YR 3/4). The quartz, andesine, and carbonate are in fine, elongate, interlocking grains. The garnets, most abundant in the biotite-rich layers, are filled with inclusions of quartz, carbonate, and to a less extent biotite. Their interiors are more crowded by inclusions than their exteriors. Through many of the garnets, the banding in the slate passes uninterruptedly, preserved by the inclusions. The banding is straight inside the garnets, even when it is rumpled in the matrix. Post-crystallization alteration is generally strong, but varies in intensity even in the hand specimen. Commonly one band is strongly altered whereas the adjacent band is completely unaltered. The end product of the alteration is a silvery white, fine-grained, quartz-sericite or quartz-carbonate-sericite schist more or less speckled by dark chlorite pseudomorphs after biotite and garnet.

## MURPHY MARBLE

**Name and correlation.** The name refers to Murphy, North Carolina. The formation was named by Keith (1907, p. 5) who correlated it with the Shady dolomite (1907, p. 11) of Cambrian age.

**Distribution and thickness.** The Murphy marble is confined to two northeast-trending belts, both of which are marked for almost their full length by a valley. The western belt extends along Young Stone Creek and Dry Creek; the eastern belt along Cutcane Creek, Creaseman Branch, and Weaver Creek. Outcrops being uncommon, the location of the marble is known mainly from old lime pits, drill holes, and wells.

The marble is more deformed than other rocks in the region. Its thickness varies from 0-250 feet. Some of this variation is the result of mechanical deformation, but some of it, as shown by the thin bedding and variable character of the formation, is a depositional feature.

**Character.** The marble is thin-bedded and mostly fine-grained. It is usually white, but may be blue, or blue and white banded, rarely pink. Interbedded with the marble are layers of slate and schist which vary in number and thickness from place to place. The basal marble beds commonly feather-out by interbedding with the topmost slates of the Brasstown formation. In this way, the marble can change in thickness more than 40 feet within a strike distance of 200 feet. Most of the pink marble and the coarse-grained, high-lime marble is found in this "pockety", basal zone. Higher, the marble is dolomitic, fine-grained, and generally white. The purest dolomite and the thickest beds appear to be in the central zone. The top of the formation is exposed only at the Campbell quarry on Cutcane Creek. There the marble is interbedded with basal schists of the Andrews formation.

A specimen of white, fine-grained, dolomitic marble from Cutcane Creek, 4/10 mile northeast of its confluence with Hemptown Creek, yielded a 1.4% insoluble residue composed of 2M muscotive, chlorite, and quartz, in about equal amounts. A thin section shows all three minerals scattered through the marble, the muscovite and chlorite in very small books, the quartz in small, irregularly-shaped masses. Optic angle measurements on seven of the muscovite books in the insoluble residue gave the following values: 30.3°, 21.6° 28.0°, 30.0°, 30.4° 28.0°, 31.0°. A specimen of fine-grained bluish gray dolomitic marble from the mouth of Young Stone Creek yielded a 0.1% insoluble residue composed of the same three minerals and stained dark by carbonaceous matter. Both specimens were digested in cold 1:4 HCl. Marble specimens from other localities contain streaks rich in pale brown mica, occasional pyrite, chalcopyrite, and garnet. Some specimens contain small clusters of radiating tremolite needles. In Cutcane Valley, where the marble is best exposed, tremolite-rich layers and "spotted" schists (metamorphosed mud pellets?) are common in the top half of the formation.

The chemical analyses in Table 7 shows typical variations in the composition of the marble.

## TABLE 7

### Chemical Analyses of the Murphy Marble

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 |
|---|---|---|---|---|---|---|---|---|---|---|---|
| MgO | 11.19 | 15.89 | 5.07 | 9.42 | 12.26 | 15.78 | 16.37 | 15.39 | 4.85 | 16.96 | 12.75 |
| CaO | 42.64 | 32.68 | 48.00 | 41.62 | 36.90 | 33.66 | 32.70 | 34.46 | 47.60 | 33.10 | 36.30 |
| Insoluble | 1.66 | 10.70 | 3.46 | 7.70 | 9.43 | 5.24 | 4.00 | 2.50 | 6.76 | 6.46 | 5.72 |
| $MgCO_3$ | 23.40 | 33.20 | 10.61 | 19.71 | 25.63 | 33.00 | 34.23 | 32.19 | 10.15 | 35.47 | 26.66 |
| $CaCO_3$ | 76.11 | 58.33 | 85.67 | 74.28 | 65.86 | 60.08 | 58.36 | 61.50 | 84.96 | 59.10 | 64.75 |
| Total carbonates | 99.51 | 91.53 | 96.28 | 93.99 | 91.49 | 93.08 | 92.59 | 93.69 | 95.11 | 94.57 | 91.41 |

1. White, medium-grained marble.
2. White, finely micaceous, dolomitic marble.
3. White, thinly bedded, medium- to fine-grained marble.
4. White, medium- to fine-grained marble.
5. White, coarse-grained marble.
6. Blue and white banded, medium-grained marble containing a few thin streaks of pink marble.
7. White, fine-grained marble.
8. White, fine-grained marble.
9. White, mostly fine-grained marble.
10. White, mostly fine-grained, dolomitic marble.
11. White, fine-grained marble.

Samples 1-5 collected from the head of Creaseman Branch, drill core; samples 6-10 collected on Mr. Lon Dean's property, Cutcane Creek, drill core; sample 11 collected from outcrop in Cutcane Creek about 1800 feet northeast of Hemptown Creek. Analyses by Dr. L. H. Turner, Chief Chemist, Georgia Geological Survey.

## ANDREWS FORMATION

A metasedimentary sequence 1400-1800 feet thick lies between the Murphy marble and the Nottely quartzite. The base of the sequence is a calcareous schist which corresponds to Keith's Andrews schist. The middle and upper sections of the sequence resemble the rocks which are above the Andrews schist in the Nantahala quadrangle, but were mapped by Keith as part of the Valleytown formation. The complete sequence is here called the Andrews formation. Although it contains rocks which were mapped erroneously by Keith and LaForge and Phalen as parts of other formations, its stratigraphic limits are the same as those which Keith specified for the Andrews schist.

All of the Andrews formation in the Ellijay folio is shown as Valleytown formation.

**Distribution.** On the east limb of the Murphy syncline, the Andrews formation crops out in a belt 2000-2500 feet thick between the Murphy marble and the Nottely quartzite. On the west limb nearly all of the formation is missing. Small sections of it are found along the west side of the Nottely quartzite between Mineral Bluff and Coles Crossing. Faulting is apparent at many places along this belt, even though exposures are not good. The Andrews formation is therefore believed to have been cut out by faulting.

**Character.** A calcareous schist containing nodules and thin lenses or beds of limonite marks the base of the formation. Being very susceptible to weathering, this part of the formation, like the underlying marble, occurs mostly in low, poorly exposed areas. Outcrops were seen in only three localities: One mile south of Toccoa River, on the west side of Weaver Creek Valley; three-fourths mile north of Toccoa River, on the west side of Creaseman Branch Valley; and at the Campbell quarry in Cutcane Valley, about one mile northeast of the confluence of Cutcane and Hemptown Creeks. At the first-named locality the limonite beds have been mined, but the open cuts are caved and the rocks so thoroughly weathered that little could be learned of the relation between the limonite and the enclosing schist.

The middle and upper parts of the formation are composed of fine quartz-sericite schist, sparingly garnetiferous, and gray slate or phyllite, some of which is faintly banded. The rocks are minutely and irregularly crinkled. In a few areas they are dark gray because of included, fine, carbonaceous matter.

The bedding planes in this formation are generally obscured by the strong foliation, the fine texture, and the uniformity of color.

## NOTTELY QUARTZITE

**Name and thickness.** Keith named the formation for its exposures along the Nottely River, near Culberson, North Carolina (1907). The thickness of the formation is 75 to 150 feet.

**Distribution and character.** One band of Nottely quartzite enters the quadrangle 0.4 mile south of Coles Crossing, and extends southwest through Mineral Bluff and Blue Ridge.

This band has been traced to the type locality. A second band lies 700-1500 feet southeast of the first band.

The Nottely quartzite consists of thin, white quartzite beds interstratified with dark slate laminae and thin quartz-sericite schists (Fig. 21). The thickness of the quartzite beds is usually less than two feet, but occasionally exceeds four feet. The beds vary greatly in number, as well as thickness, from one place to another. On the east side of the quadrangle the slate laminae are less numerous and the quartzite beds thicker than in the vicinity of Blue Ridge and Mineral Bluff.

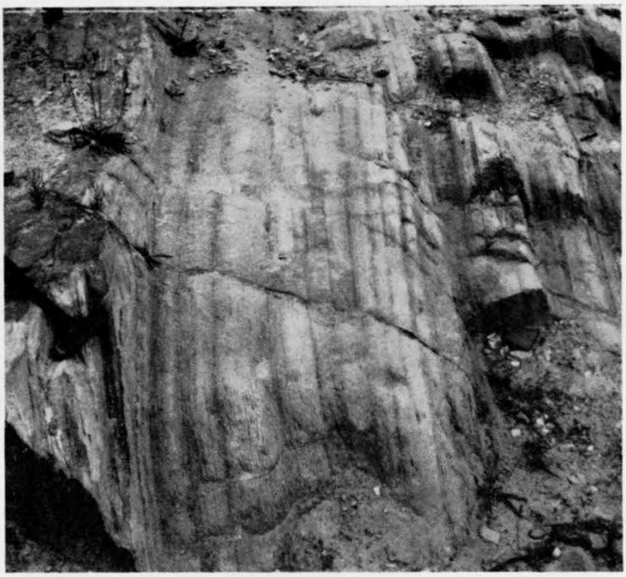


Fig. 21. Nottely quartzite, exposed at the school house in Mineral Bluff. Thin white quartzite beds interstratified with dark slate laminae.

Most of the quartzite is medium-grained, but commonly it is coarse, even conglomeratic. Most of it is feldspathic.

The Nottely quartzite closely resembles the Tusquitee quartzite, with which it has been confused. Microscopically, the two formations are indistinguishable.

## MINERAL BLUFF FORMATION

Overlying the Nottely quartzite is a fine-grained sequence which resembles the top of the Andrews formation. The sequence consists of quartz-sericite schist, slate or phyllite, and small amounts of graphitic schist and talcose schist. Metaconglomerate lenses, which help to distinguish the sequence from the Andrews formation, occur 100 to 200 feet above the Nottely quartzite.

This formation is well exposed in the vicinity of Mineral Bluff, after which it is named. Its gradational contact with the Nottely quartzite can be seen at the school house, on the east side of Mineral Bluff. Blue-gray slate is exposed in a flagstone quarry nearby. The graphitic schist and talcose schist, as well as the contact with the Nottely, are exposed northeast of Mineral Bluff along Georgia highway # 60. Slate and sericite schist, some of it containing small biotite and garnet porphyroblasts, crop out at many places along Dividing Ridge.

The Mineral Bluff formation occupies the trough of the Murphy syncline. The top of the formation has been removed by erosion; 300 to 800 feet remain.

**Summary of sedimentation characteristics, by formation.**

**Copperhill formation.** Arenaceous and argillaceous materials are intimately mingled. The feldspar content is high. Carbonate rocks are present only as nodules. The bedding is well-marked and rhythmic; graded bedding is typical; cross-bedding absent.

The metaconglomerates in the formation are characterized by the absence of very coarse debris, but still show a large range in sediment size. They have a high percentage of subangular feldspar pebbles, common black slate pebbles, and a graywacke-type matrix.

The original sediments were unweathered or incompletely weathered, poorly sorted, and were rapidly deposited below wave base. They were typical geosynclinal sediments of the graywacke suite, probably deposited in the axial portion of the geosyncline.

**Hughes Gap formation.** Marked by the predominance of metasediments high in iron and alumina, the materials usually concentrated by prolonged weathering. These metasediments are interbedded with what were originally thin cherts or limy shales (now pseudodiorites), quartz sandstones, and quartz conglomerates.

The metaconglomerates show fair to good sorting and sizing, pebbles almost entirely of well-rounded quartz, feldspar content low to nil, quartz-rich matrix. The metaconglomerates are oligomictic.

There are rapid, even abrupt, lithologic changes both laterally and vertically. All the characters of these metasediments suggest a weathered sediment supply, reworking, and final deposition above wave base, possibly in shallow water.

The transition from the graywacke-type metasediments of the Copperhill formation to the oligomictic metaconglomerates and iron-alumina-rich schists of the Hughes Gap formation takes place gradually by the interbedding of metasediments of both types. Varved schists occur at this horizon. The transition to the overlying Hothouse formation is also gradational.

**Hothouse formation.** Mostly typical graywacke suite metasediments. The base of the formation is characterized by the preponderance of what originally was silt and clay, now mica schist. Toward the top of the formation there is a gradual increase in average grain size and in the abundance of the metaarenites, also a change in metasediment type. The lithologic character of the top of the formation is notably variable along strike. Even so, the metasediments of the Hothouse and Dean formations are concordant and gradational.

**Dean formation.** Very similar to the Hughes Gap formation, but pseudodiorite is more abundant. Cross-bedding and scour channels again point to bottom current action and to deposition above wave base.

**Nantahala slate.** Mainly dark, banded slate containing pyrite and organic matter. The well-preserved laminations point to slow deposition in deep water. Although there are thin beds of metasubgraywacke and quartzite, the formation is typically of the euxinic or black shale facies.

Possibly there is a disconformity at the base.

**Tusquitee quartzite.** Thin beds of light-colored quartzite, feldspathic quartzite, and metasubgraywacke, all containing laminae of black slate. Abundant evidence of bottom current action, as truncated laminae and crude cross-bedding. Possibly this formation originated as a "shoestring" type of sand.

**Brasstown formation.** The base is black slate. Upward, metasubgraywacke and quartzite become common, interbedded with dark slate.

**Murphy marble.** Banded marble, gradational to the underlying and overlying formations by interbedding. Locally contains thin schist beds and scattered quartz sand. Mainly dolomitic.

**Andrews formation.** Mainly metaargillite and metasilt. Locally graphitic, or talcose, or finely banded. The base of the formation is marked by calcareous schists and thin beds of limonite.

**Nottely quartzite.** Very similar to the Tusquitee quartzite. Differs mainly by the smaller organic content of the interbedded slates.

**Mineral Bluff formation.** Similar to the top of the Andrews formation. Locally, there are prominent lenses of metaconglomerate and metaarkose, and zones of metasubgraywacke.

# METAIGNEOUS ROCKS

## The Newtown Sill

**Thickness and distribution.** An epidote amphibolite sill about 100 feet thick crosses the Toccoa River on the west edge of Copperhill. The sill can be followed northeastward for nearly two miles along the east limb of the Newtown anticline. Although broken by faults and locally attenuated, even missing in places, it can be followed around the nose of the NE-plunging Newtown fold and back along the west limb, almost as far as the Acid Plant (see Plate 3). The sill is notably thinner towards the northeast, and probably pinches out in this direction. In the opposite direction it thickens to 125 feet, and can be traced southwest of Copperhill for more than 12 miles, with almost continuous exposure.

**Field relations.** The sill appears to be conformable with the bedding in the enclosing rocks, even where they are tightly folded. The kind of rock in contact with the sill differs, however, from place to place. This might be regarded as evidence that the sill transgresses bedding except that lithologic change along strike is a characteristic of the enclosing rocks. The outcrop pattern in Plate 3 shows that if the sill does transgress bedding it does not depart far from one stratigraphic horizon.

Masses of the Great Smoky formation are found locally in the sill. A 12-foot band of metagraywacke and schist can be seen in the sill in a road cut between Copperhill and the Acid Plant. Bedding, cleavage, and crinkle axes in this band have roughly the same orientation as corresponding structures outside the sill. The band is less than 100 feet long. Similar included masses, the largest about 40 feet thick, were noted at eight places southwest of Copperhill. These masses are rarely traceable for more than a few hundred feet.

A pronounced coarsening of grain from the margins of the sill inward was observed in the Epworth quadrangle 300 yards SE of the mouth of Fightingtown Creek, at the top of the bluff, and again where the sill crosses a small creek just south of Fry. At most other localities the sill is coarse-grained throughout. Rarely, grain size is coarser along the margins than within the sill, but where this is true, faulting is evident.

Whether a consistent relationship exists between grain size and position in the sill could not be established because of marginal faulting and poor exposure.

The sill behaved as a competent layer during the regional deformation, becoming highly fractured and mashed. Numerous small faults developed along the contacts. Metamorphism preceded much of the movement.

**Megascopic description.** The fresh rock is dark greenish gray, heavy, and tough, and may be either fine- or coarse-grained, depending on the locality. The principal mineral is hornblende. Where the rock is coarse-grained, the hornblende is in interlocking, equidimensional to slightly elongate crystals which may exceed five mm. in length; where the rock is fine-grained, the hornblende occurs as slender prisms hardly visible to the unaided eye. Feldspar is the only other mineral readily recognized. Always much less abundant than hornblende, it is inconspicuous in fresh rocks, but becomes increasingly noticeable during weathering, when it changes from colorless to white.

**Microscopic description.** The hornblende is evenly distributed through the rock in pale green crystals with sieved interiors and irregular margins, as well as irregular shapes. The crystals, often twinned, show great variation in size. They range in diameter from a fraction of a millimeter, in the fine-grained rocks, to more than five mm. in the coarse-grained rocks. The hornblende is pleochroic with X colorless to pale green (5Y 7/2), Y dusky yellow green (5GY 5/2), Z grayish yellow green (5GY 6/2). Absorption is $X<Z<Y$. 2V ranges from 82°-86°, the extinction angles $Z \wedge c$ from 12°-18°. These optical properties indicate a member of the actinolite-ferrotremolite series (Winchell, 1951, p. 433).

The plagioclase crystals are small (mostly less than 0.2 mm. across) and usually twinned. Although a few of the larger crystals are lath-shaped, the majority have irregular shapes. The plagioclase is biaxial positive, with 2V about 88° and extinction $\perp$ (001) and (010) about 20°. Measurements on 27 crystals in three thin-sections indicate andesine, $An_{34\text{-}38}$.

## TABLE 8

### Modal Analyses of the Newtown Sill

| Locality number | 3Y3 | 5Y9 | 4U4 | 6W8 |
|---|---|---|---|---|
| Andesine | 19.48 | 21.91 | 21.16 | 22.80 |
| Hornblende | 58.90 | 64.09 | 74.88 | 69.50 |
| Clinozoisite | 10.20 | 3.54 | 2.31 | 4.20 |
| Titanite | 3.99 | 4.70 | 1.65 | 1.84 |
| Black Opaque | 3.60 | 2.97 | p* | 1.66 |
| Micaceous aggregates | 3.88 | P | | |
| Quartz | | 2.79 | | P* |

*Present in small amount

3Y3—Coarse-grained epidote amphibolite, collected one-half mile north of Newtown, Mineral Bluff quadrangle.

5Y9—Medium-grained epidote amphibolite, collected between Dunn Mill Creek and Patterson Creek, Epworth quadrangle.

4U4—Fine-grained epidote amphibolite, collected 300 yards east of the mouth of Fightingtown Creek, Epworth quadrangle.

6W8—Medium-grained epidote amphibolite, collected on south side of High Top Mountain, SW corner of Epworth quadrangle.

Strewn through the feldspathic portions of the rock, and occasionally included in the hornblende, are prisms and irregular grains of epidote. Nearly all of it has a blue interference color, variable 2V (72°-86°), positive optic sign, and inclined extinction, which identify it as clinozoisite. A few crystals have the negative optic sign and higher interference colors of common epidote.

Titanite is present as minute roundish crystals—a few of which are wedge-shaped—and as larger granular lumps. Titanite and a black, opaque mineral, either magnetite or ilmenite, occur together. They are spottily distributed.

Other minerals recognized in the sill but only sparingly or locally present are quartz, chlorite, pyrite, and pyrrhotite. The quartz is in small scattered grains and very small symplektitic intergrowths; usually it is not present. The chlorite is secondary after hornblende.

Parts of the sill in the vicinity of Copperhill contain microscopic, patchy aggregates of a fine, micaceous, highly birefringent mineral. This mineral was not identified, but textural relations show that it is an alteration product of hornblende. Probably it is talc.

Microscopic cataclastic effects—warped cleavages, bent prisms, segmented and displayed crystals—can be seen in nearly every thin section.

**Origin.** Specimens from the least metamorphosed portions of the sill, in the Epworth quadrangle, show that a part, at least, of the epidote amphibolite was derived from a fine-grained rock of uniform texture. Modal analyses (Table 8) show that the proportion of feldspar to dark minerals is remarkably regular regardless of present variation in grain size, and that the chemical composition of the rock is fairly uniform over a large area. In Table 9 the composition of the epidote amphibolite and of basic igneous rocks from which it might have been derived are compared. The epidote amphibolite is chemically indistinguishable from diabase, gabbro, and olive basalt.

Two possible modes of origin are suggested: (1) the intrusion of a diabasic or gabbroic sill into unmetamorphosed Great Smoky sediments; or (2) the subaqueous extrusion of thin

**TABLE 9**

**Chemical Composition of the Newtown Sill and Other Basic Igneous Rocks**

| | 1 | 2 | 3 | 4 | 5 | 6 |
|---|---|---|---|---|---|---|
| $SiO_2$ | 49.62 | 52.65 | 50.41 | 51.45 | 51.22 | 2.0 |
| $Al_2O_3$ | 15.87 | 16.23 | 15.14 | 18.67 | 13.66 | 6.0 |
| $Fe_2O_3$ | 1.75 | 0.51 | 2.71 | 0.28 | 2.84 | 6.0 |
| FeO | 7.90 | 8.21 | 7.95 | 9.04 | 9.20 | 3.0 |
| MgO | 7.29 | 6.64 | 6.57 | 6.84 | 4.55 | 2.0 |
| CaO | 12.86 | 11.34 | 11.30 | 10.95 | 6.89 | 2.0 |
| $Na_2O$ | 2.17 | 1.58 | 2.29 | 1.58 | 4.93 | 2.0 |
| $K_2O$ | 0.40 | 0.90 | 0.82 | 0.14 | 0.75 | 1.0 |
| $H_2O$- | 0.27 | 0.85 | 0.72 | 0.03 | | 0.6 |
| $H_2O$ | 1.59 | 0.48 | 1.01 | 0.34 | 1.88 | 2.0 |
| $TiO_2$ | 0.58 | 0.58 | 1.30 | 0.34 | 3.32 | 1.0 |
| $P_2O_5$ | 0.11 | 0.01 | 0.15 | 0.09 | 0.29 | 0.4 |
| $SO_3$ | 0.08 | | | | | |
| MnO | 0.20 | 0.15 | 0.17 | 1.47 | 1.25 | 1.4 |
| $CO_2$ | | | 0.07 | | 0.94 | |
| BaO | | | 0.03 | | | |
| Totals | 100.69 | 100.13 | 100.64 | 100.22 | 100.72 | |

1. Epidote amphibolite near Fry, Ga. (Emmons and Laney, 1926, p. 23).
2. Average undifferentiated diabase, six analyses (Edwards, 1942).
3. Olivine basalt (Holmes and Harwood, 1929).
4. Hypersthene gabbro (Hall, 1932).
5. Average spilite (Sundius, 1930).
6. Differences between percentages in first five columns are not to be regarded as significant if less than the figure in column 6 (Fairbairn, 1951).

basalt flows during the time the sediments were being deposited.

Although some of the metasedimentary inclusions in the epidote amphibolite are attributable to infolding, during regional deformation, this cause hardly accounts for the large, tabular masses near the center of the sill in the Epworth quadrangle. These masses represent sediments that were engulfed in the molten sill, if the first postulated origin is correct, or sediments deposited between subaqueous flows, if the second mode of origin is correct.

Neither origin is ruled out. However, the second is favored by: (1) the shape of the included masses; (2) the strict concordance between their bedding and the bedding in the rocks outside the epidote amphibolite; (3) the fine grain and uniform texture of the parent rock; and (4) the lath-shaped feldspar.

**Sill between Daly Creek and the Toccoa River.**

A basic sill which has a maximum thickness of two feet and a length of about 40 feet crops out between Daly Creek and the Toccoa River, near the center of the quadrangle. The rock composing the sill is dark greenish gray and medium- to coarse-grained.

**TABLE 10**

**Modal Analysis of the Sill Between Daly Creek and the Toccoa River**

| | |
|---|---|
| Hornblende | 78.21 |
| Bytownite | 9.40 |
| Clinozoisite | 0.34 |
| Black opaque | 4.53 |
| Titanite | 1.79 |
| Quartz | 3.68 |
| Chlorite | 1.97 |
| Biotite | 0.08 |
| | 100% |

The hornblende is in coarse, sieved crystals whose optical properties are the same as those of hornblende in the Newtown sill. The bytownite is optically negative, with a 2V of about 82°, and extinction $\perp$ (001) and (010) of 40°-44°. The quartz occurs mainly as small inclusions in the hornblende, the biotite as coarse, interstitial books. The chlorite is an alteration product of hornblende and biotite.

The sill weathers to dark red-brown saprolite easily confused with weathered pseudodiorite.

# QUARTZ AND PEGMATITE

Quartz veins ranging from stringers less than an inch thick to "blow-outs" more than six feet thick are found at many places in the Mineral Bluff quadrangle, although they are not numerous at any one place. They are usually less than 25 feet long. Some of the veins are parallel to the foliation in the enclosing rocks, and have been folded. Most of them cut across schistose rocks and are younger than the main period of deformation. They probably formed during the period of crinkling and alteration.

The orientation of the largest veins is shown in figure 22.

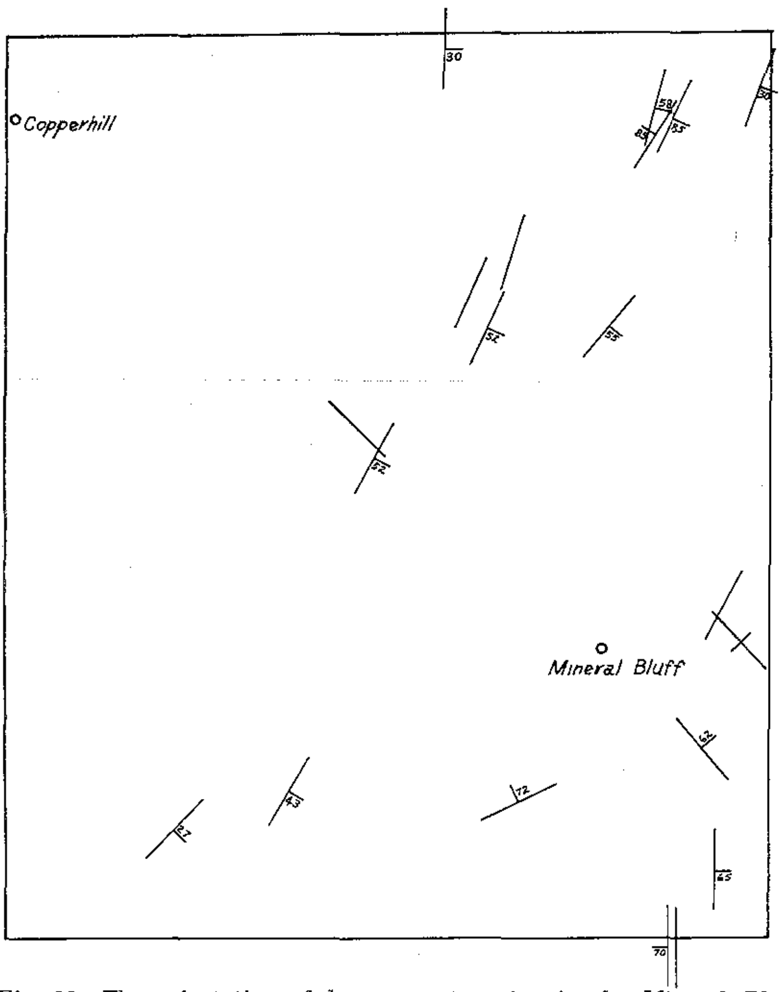



Fig. 22. The orientation of large quartz veins in the Mineral Bluff quadrangle.

Narrow stringers and small irregularly shaped masses that have been called pegmatites occur in the metagraywacke at a few places. These "pegmatites" are marked by a slightly higher quartz-feldspar content than the enclosing rock and by a slightly coarser grain. Their grain size rarely exceeds one cm. They are gradational to the enclosing metagraywacke and appear to be of segregational origin.

# STRUCTURE

## Preserved Sedimentary Structures

**Bedding.** At nearly every outcrop bedding is visible, shown by planar changes in composition, color, fabric, and grain size. The prominence of the bedding is due largely to strong, original, sedimentational differences, but it is also partly due to metamorphism, deformation, and weathering, which have accented rather than obscured the original bedding planes. Metamorphism has produced new textures and minerals that accentuate the bedding (see figures 7 and 17). Deformation has exaggerated competency differences (see figure 24). Where schist and fine-grained metagraywacke are interbedded, bedding planes that were hard to see prior to deformation are now plainly visible in the alternating crinkled and uncrinkled layers. Weathering has "etched" the rocks, and thereby brought out many small mineralogical and textural differences that are hardly apparent in the fresh rocks. Weathering has "retouched" the bedding planes with new color changes and with differences in the rate which the beds disintegrate.

**Graded bedding.** This structure is very common in the metasediments of the Great Smoky group. It is strikingly developed in some of the conglomeratic zones, where the gradation from fine metaconglomerate to schist may be repeated by as many as ten consecutive beds. The usual graded bedding is that shown in figure 4 and illustrated schematically in sketch No. 1 of figure 23. The lowest and thickest part of the bed is coarse metagraywacke; the upper part is either fine metagraywacke or mica schist. The manner of gradation from coarse to fine is that characteristic of graywacke sequences elsewhere. First there is a decrease in the proportion of coarse to fine material, then a decrease in the upper size limit of the grains. Where the fine-grained layer is missing, the size limits of the grains are the same from bottom to top, and grading is entirely the result of a decrease in the proportion of coarse to fine sediment.

The grading is hard to see in the fine-grained beds that are undeformed, but it is often conspicuous where there has been crinkling, as shown in figure 24. Crinkling was most

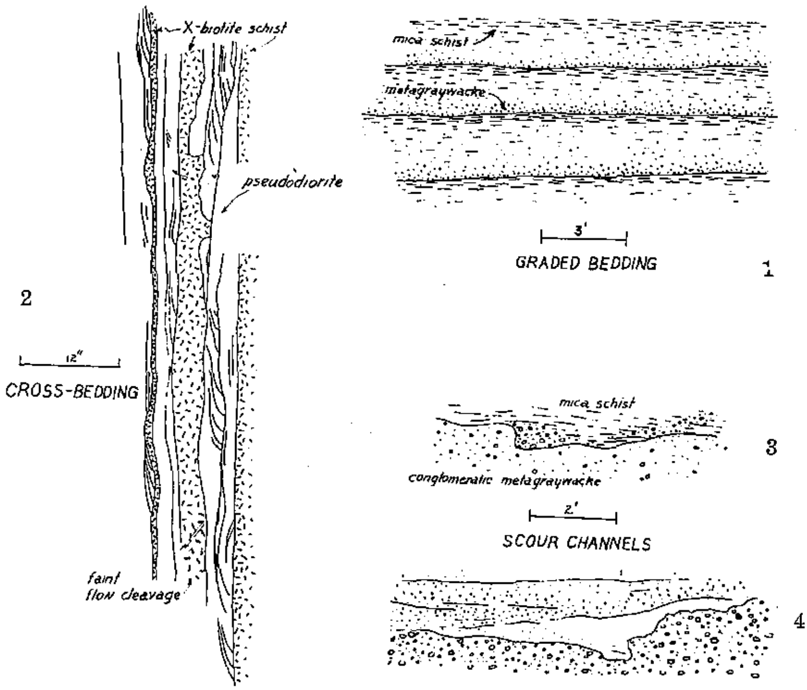



Fig. 23. Preserved sedimentary structures. 1. Typical graded bedding in the Great Smoky group. 2. Cross-bedding and scour channels at the base of the Dean formation. 3 & 4. Scour channels in the great Smoky group.

effective in the least competent, schistose layers, which represent the argillaceous tops of the original beds. The crinkling fades out on one side, which is toward the bottom of the bed, and is cut off abruptly on the other side, where there was once a sharp break between the fine-grained top of one bed and the coarser, basal portion of the overlying bed.

The thickness of the graded beds is usually less than one foot, but may exceed three feet. The direction of grading is always consistent where both grading and bedding are sharp, but where either is indistinct, grading is not reliable as a top-bottom indicator. Each graded bedding symbol on Plate 3 represents a minimum of five consecutive beds in which the direction of grading is consistent.

In the metasediments that overlie the Great Smoky group, graded bedding is much less common and poorly developed.

Fig. 24. Graded bedding accentuated by crinkling, top of beds to the left. The layers that were originally argillaceous are strongly crinkled; the crinkling fades out toward the basal, slightly coarser portions of the beds.

**"Reversed" graded bedding.** This structure is common in the varved schists in the transition zone between the Copper-

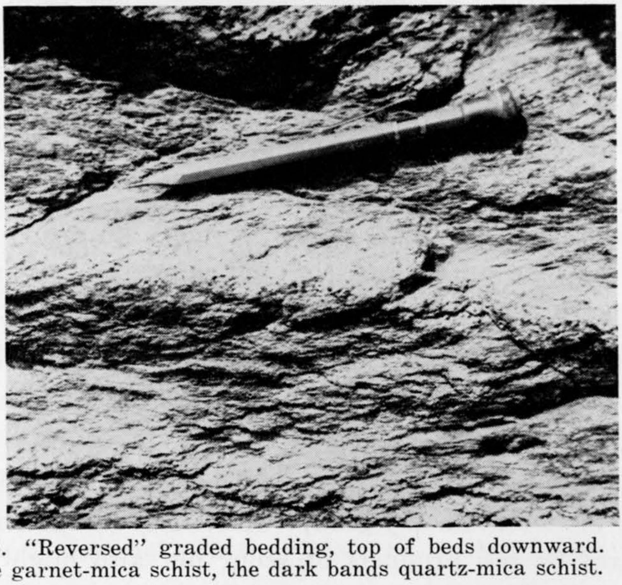


Fig. 25. "Reversed" graded bedding, top of beds downward. The light bands are garnet-mica schist, the dark bands quartz-mica schist.

hill formation and the Hughes Gap formation. It is also present in thinbedded schists of the Hothouse formation.

The reversal in grading is a result of metamorphism. The basal part of each bed, originally quartz-rich silt, has recrystallized to quartz-mica schist (the dark layers in figure 25); the top part, originally clay, has recrystallized to garnet-mica schist. The number of garnet metacrysts is related to the former percentage of clay. The direction of grading has been reversed by the growth of coarse garnets in what originally was the finer-grained part of the bed.

**Cross-bedding.** Although not observed in the metasediments that underlie the Hothouse formation, cross-bedding is present, sparingly, in the top of the Hothouse formation, and is common in some of the overlying formations.

Cross-bedding in pseudodiorite at the base of the Dean formation is shown in sketch No. 2 of figure 23. The pseudodiorite is interbedded with X-biotite schist. Some of the cross-bed laminations are thin seams of X-biotite schist. Both the cross-bedding and the scour channels in the central band of schist indicate that the top of the beds is to the right (west). This sketch was made at an abandoned aggregate quarry one mile east of Blue Ridge. On the east side of the quarry, pronounced grading in conglomeratic metagraywacke, belonging to the Hothouse formation, gives the same top-bottom direction.

The current laminations shown in figure 20 are characteristic of the Tusquitee and Nottely quartzites. Truncations of laminae provide occasional top-bottom indications. Excellent cross-bedding is preserved locally in thin layers of the Nottely quartzite, northeast of Mineral Bluff.

**Scour channels.** Sketches 3 and 4 of figure 23 illustrate the larger scour channels found at a few places in the metasediments of the Great Smoky group. Smaller bedding disruptions that might be the products of scour but also might be the products of regional deformation are fairly common. The origin of most of these structures is in doubt because the possible scour and fill materials are similar and deformation on a small scale is intense.

In the formations that overlie the Great Smoky group evidence of bottom scour is more positive but is still only rarely conspicuous. The usual evidence is truncated laminae. Small scour channels like those in sketch No. 2 of figure 23 are not rare.

## Secondary structures

### FOLDS

The folds all trend NE-SW. Most of them are asymmetric and overturned to the northwest. They have a variable plunge.

Only two major folds are found in the southeastern two-thirds of the quadrangle: the Morganton anticline, a small fold in the extreme southeast corner of the quadrangle, and the Murphy syncline, a large fold (synclinorium) whose axis passes through Mineral Bluff and Blue Ridge. In the northwest third of the quadrangle there are at least five major folds. All are secondary folds on the east limb of an anticlinorium that corresponds to the Murphy syncline in size. The axis of this anticlinorium lies in the Ducktown Basin, just northwest of the Mineral Bluff quadrangle.

**Morganton anticline.** The Morganton anticline is about one mile across and at least four miles long. It is an upright fold, plunging 25° to the southwest. In the vicinity of Lake Toccoa, distinctive beds of metagraywacke and staurolite schist can be traced with almost continuous exposure around the crest of the fold. As the dip of the bedding changes from steep on the limbs to flat-lying in the crest, the bedding planes undulate gently but show few, if any, secondary folds. To the northeast, along the paved road between Morganton and Hollywood, the rocks in the crest are mainly schist and are less regular in attitude: drag folds and crinkles are developed, and thin pseudodiorite beds are conspicuously boudinaged. The intensity of the boudinaging diminishes toward the east and west, on the limbs of the fold.

The flow cleavage in the Morganton anticline is visible only in the schist beds, and is not axial plane cleavage (see figure 27).

Most of the Morganton anticline is outside the Mineral Bluff quadrangle.

**Murphy syncline.** The structure that is here called the Murphy syncline was interpreted by earlier workers as an anticline, a homocline, and a fenster, as well as a syncline (see page 1). Its synclinal character is clearly brought out by the distribution and sequence of the formations in Plate 3 and by abundant top-bottom criteria. The complicated synclinal interpretation of Keith, however, does not apply.

The core of the syncline is a simple bent fold, overturned to the west, and faulted along the trough. The faults are reverse faults that dip toward the axis of the syncline. They are typical of bent folds, being a result of the crowding that occurs when the limbs of a fold are forced together (Cloos, 1937, especially figures 31, 36, and 37).

A few open cross-flexures with axes inclined 10°-30° to the trend of the syncline occur on the east limb. They are apparent in the sinuosity of the Nottely quartzite and of the metaconglomerate in the Dean formation. They are apparent also in the way the strike of the bedding is locally inclined to the strike of formational boundaries. A few small, tight folds are associated with the faults at the base of the Nottely quartzite on the west limb. Otherwise secondary folds are remarkably rare, being common only in the outer parts of the limbs. The Morganton anticline is a secondary fold on the east limb, three and one-half miles from the axis of the syncline. No secondary folds of comparable size were noted on the west limb. However, widespread crinkling appears $2\frac{1}{2}$ miles west of the axis of the syncline in the Hothouse formation. Drag folds appear farther out, in the top of the Copperhill formation, where the Murphy syncline grades into the anticlinorium.

The Murphy syncline is the largest fold in the quadrangle. The east limb embraces all the rocks southeast of Mineral Bluff; the west limb reaches as far as Harper Ridge.

Northeastward, the syncline extends for many miles, at least as far as the Nantahala River in North Carolina. Southwestward, it reaches to Ellijay, Georgia, and beyond. Its average width is 10-15 miles.

The principal cleavage in the Murphy syncline, as in the Morganton anticline, is not axial plane cleavage.

**Folds in the Copperhill area.** The extreme northwest corner of the quadrangle is a part of the Ducktown Basin, which was denuded by early mining operations. In this area, stripped and gullied by erosion, complicated folds can be readily deciphered.

The Coletown syncline (see Plate 3) is almost completely exposed. It is inclined to the northwest, plunges northeast, and is marked by many third- and fourth-order folds. The adjoining fold of the same size to the southeast is the Newtown anticline, also well exposed. An epidote amphibolite sill can be traced from the east limb of this fold around the northeast-plunging nose and back along the west limb. The next fold to the southeast is the Copperhill syncline. Both Copperhill, Tennessee, and McCaysville, Georgia, are located in its trough. Although this fold is not as well exposed as the first two, its synclinal character is clearly shown by the outcrop pattern and by graded bedding on both limbs. Farther southeast, so much of the surface is covered that the details of the folding could not be worked out. There are sufficient exposures to show, however, that the deformation plan of the Copperhill area continues southeastward almost to Harper Ridge. Along the top of the Copperhill formation the intensity of the folding diminishes and there is a transition to the kind of deformation that predominates in the Murphy syncline.

The oldest and strongest cleavage in the Copperhill area is an axial plane flow cleavage. It is now intensely crinkled and in many places rotated with respect to the bedding planes. A fracture cleavage has been superposed.

**Character of the folding.** In the Murphy syncline and Morganton anticline—particularly in the trough and crest—incompetent layers of slate and schist show compression (crinkling), whereas thin pseudodiorite beds, which are more competent, show tensional effects (pronounced boudinaging). Thick competent beds show only bending. Second-order folds are uncommon. Flow cleavage is visible, locally, but usually weak. The thickening and thinning of strata by shear is hardly apparent. These characteristics indicate that the Murphy syncline and Morganton anticline were produced mainly by bending (Cloos, 1937, pp. 55-56).

In contrast, the folds of the Copperhill area were produced by a combination of bending and shear. Deformation in this area is intense. Second-, third-, and fourth-order folds are common. Flow cleavage is strongly developed. The thicken-

Fig. 26. Third-order folds in thin-bedded metagraywacke and mica schist, one mile north of Harper Ridge. The schist beds show strong fracture cleavage. Width of photo is 4 feet.

ing and thinning of strata by shear and flowage is locally pronounced (Figure 26).

## CLEAVAGE

**Flow cleavage.** Over the entire quadrangle flow cleavage at an angle to the bedding has developed, but its strength varies from place to place. It is strong in the northwest corner of the quadrangle, less strong in the central part, only locally visible in the Murphy syncline, moderately strong again farther southeast.

Apart from the regional variation, the strength of the cleavage varies inversely with the relative competency of the beds. The cleavage is always best developed in mica-rich rocks, which are least competent. It is commonly visible in metagraywacke, which is more competent, very rarely visible in pseudodiorite. The angle between the cleavage and the bedding varies with competency, being smallest in the least competent beds, at a given position on the fold.

The cleavage is marked by the parallelism of micas and tabular grains of quartz (see Fig. 14). The shape of the quartz grains is a result of secondary flattening. Feldspar grains, also often show flattening, though not to the same extent as grains of quartz. Where not perceptibly flattened, the feldspar grains may still show preferred orientation (see the section on petrofabrics). Where the cleavage is particularly strong, the quartz grains have been flattened to very thin, sheetlike masses, and the feldspar grains sheared to pieces and strewn along the cleavage planes.

The regional orientation of the cleavage is shown on Plate 3. Axial plane cleavage prevails west of the Hothouse formation, where the folds partake of shear-fold characteristics. In the Morganton anticline, and possibly also in the Murphy syncline, which are bent folds, the cleavage is not the axial plane type. The changes that take place in cleavage-bedding relations from SE to NW across the quadrangle are readily seen in Section A-A′ and C-C′, at the bottom of the Plate 3 (see also figure 27).

**Formation of the flow cleavage.** Quartz pebbles in all parts of the quadrangle show that flattening has occurred in the plane of the cleavage. At some localities the pebbles also show elongation parallel to the fold axes, which lie in the cleavage plane. The flow cleavage appears to be an effect of elongation or flattening, accomplished by very small scale shears, rotations, and intramineral adjustments, in response to the forces that produced contemporaneous folding. A close genetic relationship between the cleavage and the folding is implied by the way the strength of the cleavage varies with the type of fold. That the cleavage originated partly by shear is suggested by the dependence of the cleavage-bedding angle on competency, by the orientation of minute shear planes in the rock (see the section on petrofabrics) and by abundant evidence of shear within the minerals themselves.

As a simplified approach to how the cleavage appears to

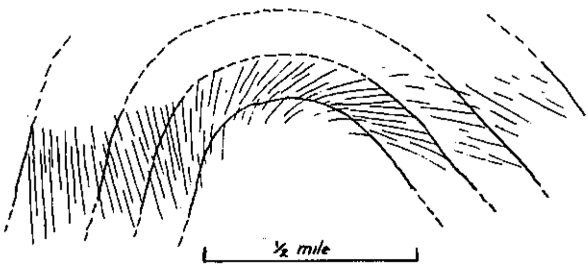



Fig. 27. Cross-sectional sketch showing cleavage-bedding relations in the Morganton anticline. East to the right.

have formed, the two-dimensional deformation of very fine-grained, bedded material is considered, briefly, in Plate 4. Deformation by shear is assumed. Bedding planes are vertical and perpendicular to the plane of the paper. The large open arrows are vectors representing the resultants of all stresses externally applied. The double-barbed arrows indicate componental movements, the single-barbed arrows relative movements along shear planes. Figures 30, 31, 32, and 33 show the deformation expected from stress applied exactly perpendicular to the bedding. With stress applied parallel to the bedding, the movements in figures 34 and 35 should result, if the bedding planes accommodate all the movement, or the movements in figures 36 and 37, if shear planes develop. The relative movements in figure 37 indicate external rotation.

In general, the resultant of all the stresses deforming a rock will be neither perpendicular nor parallel to the bedding but at some angle, as in figure 38. After shear planes develop, both internal and external rotation take place, as shown in figure 39, parts A and B. The relative movements along one set of planes are parallel and additive, those along the other set opposed. The net effect should be a single set of strong shear planes with a second set weak or not apparent, depending on the direction of the vectors, the competency of

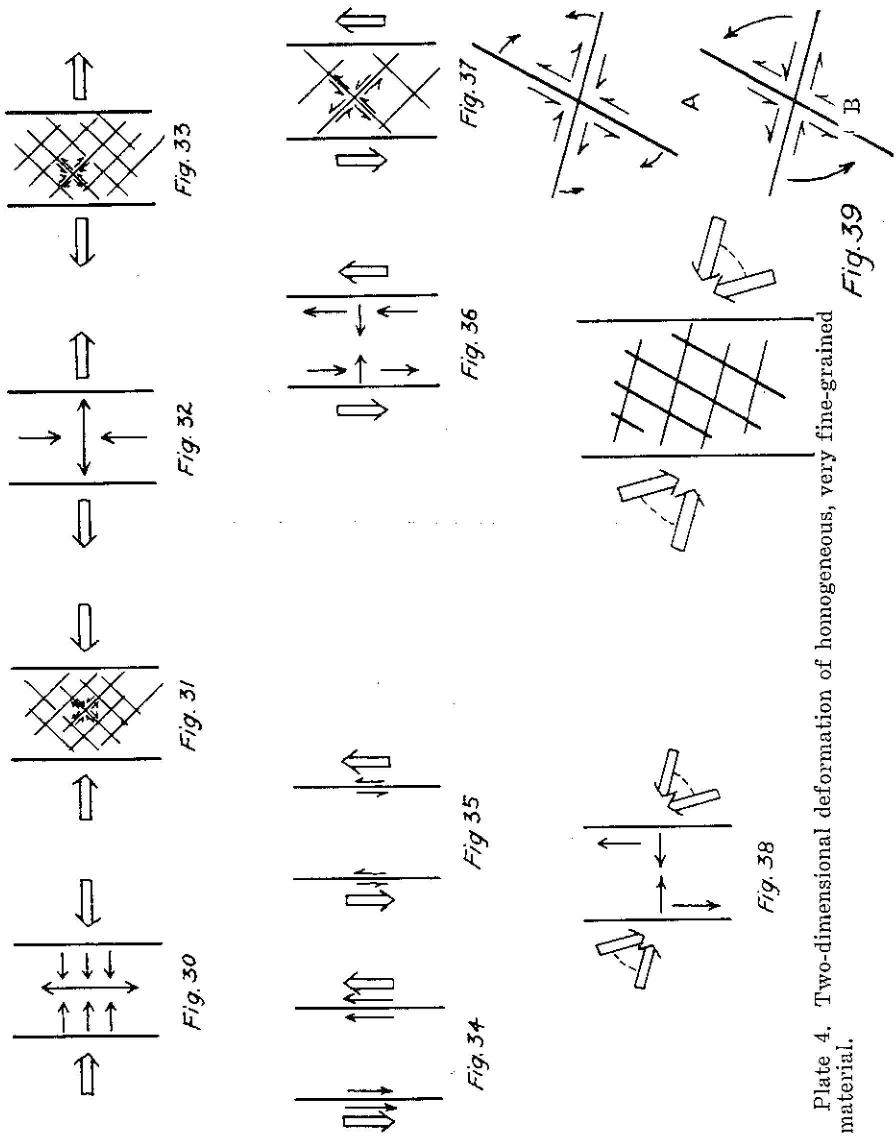



Plate 4. Two-dimensional deformation of homogeneous, very fine-grained material.

the rock, the efficacy of the bedding planes to accommodate slippage, the amount of deformation, and the time relations of deformation and recrystallization.

Shear planes resembling those of Plate 4 may be demonstrated in rocks with cleavage, where recrystallization has not obliterated the details of the cleavage-forming process. Examples are shown in Sander (1950, pp. 164-198) and in the section on petrofabrics below. In these examples the cleavage is not one of the shear planes. Before the cleavage can be related to the deformation patterns in Plate 4, inhomogeneity, a factor known to modify the ideal development of shear planes, must be taken into account.

The major inhomogeneity when fine-grained, thin-bedded rocks are considered at the meter scale is bedding. Between the meter and millimeter scales, the prominent inhomogeneities relate to the size, shape, identity, and distribution of the grains. At a still smaller scale, the most important inhomogeneities are anisotropies within the minerals.

In the $cm^2$- to $m^2$-domain, and larger, cleavage often transects bedding without deviation, and has a constant direction in beds of the same competency, even when they are separated by different layers. It is therefore evident that the inhomogeneities at this scale have not appreciably modified the cleavage-producing stresses.

At a much smaller scale, however, irregularities are apparent which show that the rock's gross distribution of stresses was, on a small scale, greatly modified. At this scale the cleavage no longer consists of parallel planes, as in figure 28, but of **trends** imperfectly aligned. The trends are due to elongate minerals and small interrupted shears. They diverge, anastomose, abut against cross-cutting trends, and vary in distinctness from one small area to another. These irregularities are traceable to the small-scale inhomogeneities, which made the local distribution of stresses very complicated. Shear planes like those in Plate 4 still tended to form, as they are often detectable in the fabric of the rock, but their ideal development was impeded by small scale "competency" differences, by stress-deflecting anisotropies, of which mineral cleavage and grain boundaries are examples, and by other factors, as crystal growth. The cleavage is usually not parallel to one of the shear planes of Plate 4, but to a plane of flattening that is partly, if not mainly, the result of small-scale shears.

The general behavior of each mineral during deformation can be deduced from the fabric and textural relations of the rocks (figures 40 and 6). The feldspar reacted sometimes by fracture but mainly by glide along and external rotation of its cleavage planes. The quartz reacted by fracture and recrystallization, or recrystallization flowage; the micas by passive external rotation and by slippage and growth along the cleavage planes.

Nearly all the rocks with strong foliation, in the Mineral

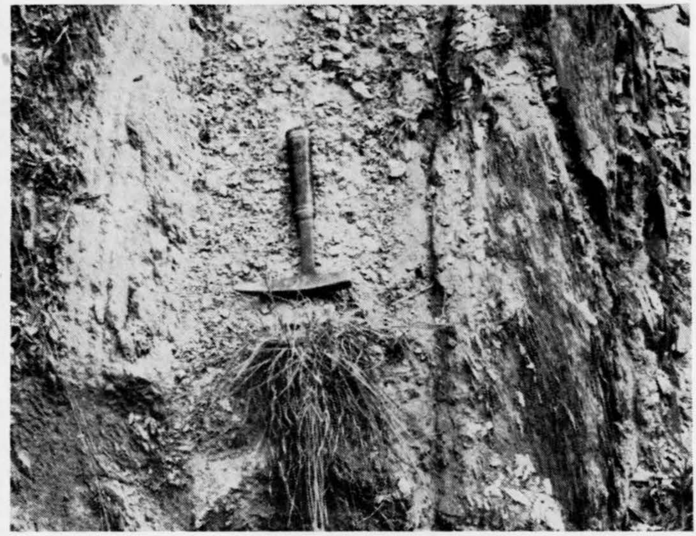


Fig. 28. Uncrinkled flow cleavage in mica schist, west limb of the Morganton anticline. Bedding is vertical; cleavage dips 75° to the right (SE). The schist is interbedded with metagraywacke.

Bluff quadrangle, were originally fine-grained. The finer the texture, the less total movement required to externally rotate tabular grains, like mica, into a common plane. The present strength of the cleavage may therefore bear some relation to the grain size at the time of deformation.

In the foregoing discussion, attention is directed mainly to

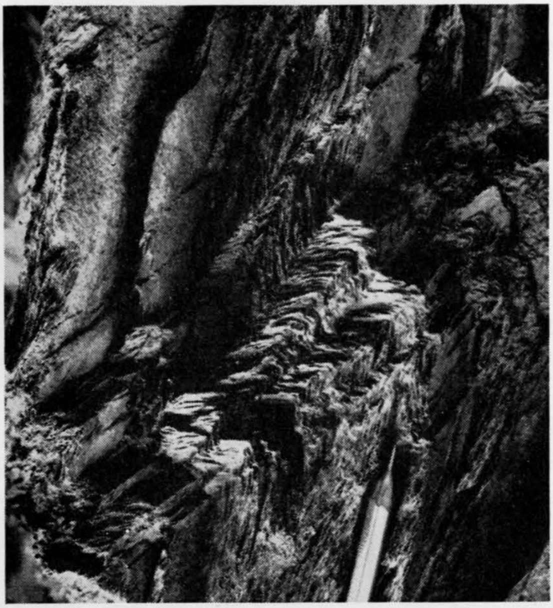


Fig. 29. Crinkled axial plane cleavage in mica schist; Copperhill area. The axial planes of the crinkles define a fracture cleavage which dips steeply to the left (SE). Bedding, not visible, dips steeply to the right (NW).

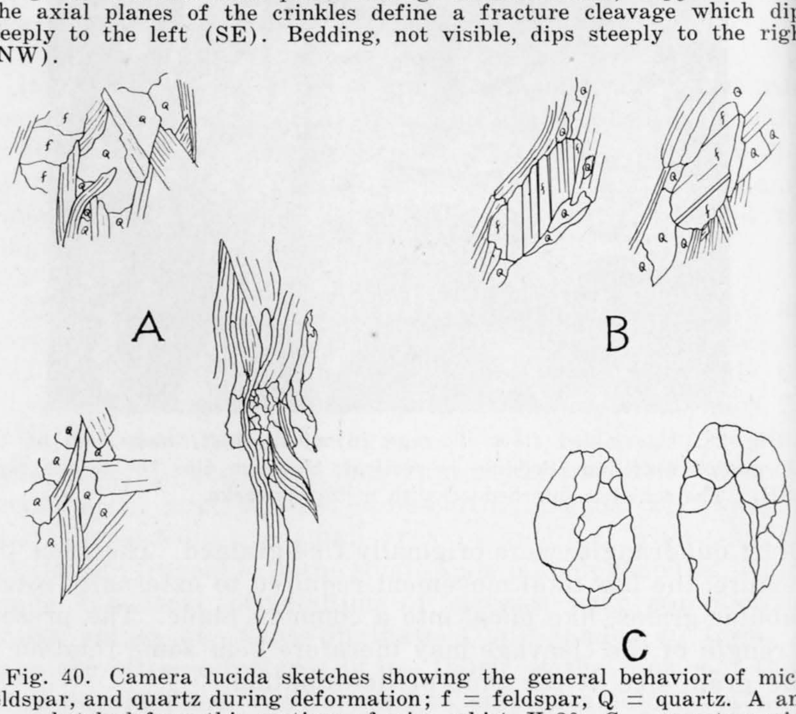



Fig. 40. Camera lucida sketches showing the general behavior of mica, feldspar, and quartz during deformation; f = feldspar, Q = quartz. A and B — sketched from thin sections of mica schist, X 30. C — quartz grains in metagraywacke, X20. Megascopic flow cleavage vertical in each case.

the role of shear in the formation of the cleavage. Other factors played a part, as recrystallization, which went on before, during, and after cleavage formation, but shear apparently played the dominant role. According to the fabric and textural relations, recrystallization was more effective in "annealing" and coarsening the grains than in orienting them.

**Bedding plane cleavage.** The slates and schists in the Murphy syncline possess bedding plane cleavage. It is due partly to fine compositional banding and partly to the preferred orientation of fine mica. To a less extent the cleavage is due to the orientation of tabular grains of quartz. A tabular shape characterizes the quartz only in the finer-textured bands, where mica is abundant. In the coarser bands, the quartz is more or less equidimensional. The tabular shape in the fine bands is probably due to recrystallization rather than to mechanical flattening. Probably the cleavage itself is the result, primarily, of mimetic recrystallization.

A faint flow cleavage at an angle to bedding is locally superposed on the bedding plane cleavage, particularly on the east limb of the Murphy syncline (see section B-B′ and C-C′, Plate 3).

In the Morganton anticline, bedding plane cleavage is dominant in some layers, flow cleavage at an angle to bedding dominant in others. The two kinds of cleavage are often present in adjacent beds.

**Fracture cleavage.** In the northwest part of the quadrangle a fracture cleavage consisting of close-spaced shear planes is visible in some of the competent beds. A fracture cleavage that is due to breakage and shear along the axial planes of crinkles is visible in some of the schists. The latter cleavage is irregularly developed and nearly always dips SE. The first cleavage is more regular and may dip either SE or NW depending on its position in the folds; its relation to bedding corresponds to that of the associated flow cleavage.

In the southeast part of the quadrangle, fracture cleavage is visible mainly in the incompetent beds. It is associated with crinkling where the beds show an earlier foliation, but it exists only as close-spaced shear planes where an earlier foliation is lacking.

The fracture cleavage in the competent beds and the flow cleavage in the incompetent beds are pre- to para-crystallizational. They appear to have developed at about the same time and in response to the same general distribution of stresses. The fracture cleavage in the incompetent beds developed later, and in the response to stresses which produced crinkling. The two fracture cleavages thus belong to different periods and are the result, in general, of different distributions of stress. They are not always separable in the field, but can usually be distinguished in thin sections, where their time relation to recrystallization can be observed.

In discussions of other areas, it is frequently mentioned that fracture cleavage associated with crinkling (strain slip cleavage, false cleavage, schübkluftung) may cross fold axes at an oblique angle and vary considerably in orientation, but that it is generally parallel to the axial planes of the major folds. This observation applies in the Mineral Bluff quadrangle. However, the usual interpretation, that the fracture cleavage is the imprint of tectonic factors of larger scale than those which determined the position of the folds and earlier cleavage, is questioned. In this area, the orientation of the fracture cleavage appears to depend mainly on the orientation of the earlier flow cleavage, and on the relative competence and thickness of the crinkled beds. Its rough parallelism with axial planes of the major folds appear to be a consequence of this dependence.

To illustrate, let Figure 51B represent a bed on the west limb of an upright syncline. The slip planes which form during crinkling will tend to parallel the cleavage surfaces that slope to the right, and will therefore be roughly parallel to the axial plane of the fold. The corresponding arrangement on the east limb of the hypothetical fold will be the mirror image of this sketch, with the fracture cleavage produced by the crinkling still roughly parallel to the axial plane of the fold. Variation in the orientation of the fracture cleavage will be introduced by differences in the original cleavage-bedding angle (the competence of the rock) and by the amount of rotation undergone by the cleavage planes (related to the competence of the rock, its thickness, and the intensity of the deformation). According to this explanation, the variation usually observed in the orientation of the fracture cleavage may originate with the cleavage and does not require postfracture cleavage movements.

## CRINKLES*

**Distribution.** The crinkles in the Mineral Bluff quadrangle are small flexures produced by the rumpling of earlier cleavage planes. The crinkling is most intense where earlier cleavage was best developed, in the northwest part of the quadrangle. Its intensity diminishes irregularly toward the southeast. In the top of the Hothouse formation on the west limb of the Murphy syncline, in the trough of the Murphy syncline, and on the west limb of the Morganton anticline crinkles occur only locally. They are minute and markedly inhomogeneous both in orientation and distribution. They correlate with drag along faults and with crowding movements in the trough of the syncline. In the crest of the Morganton anticline, farther SE, crinkling is again common but usually not intense. Because the rocks in this part of the quadrangle were mostly unfoliated or poorly foliated when the last deformation began, they yielded more by lamellar shear than by crinkling.

**Description:** The crinkles range in size from microscopic to more than 10 cm. across; their crest-to-crest distance is usually about three cm. The size of the crinkles is related to many factors: for example, grain size, "coarseness" of the cleavage, strength of the cleavage, proportion of quartz to mica, and especially small-scale inhomogeneities. The troughs and crests of the crinkles may be rounded or abruptly turned, the limbs straight or curved (see Plate 5). The angle between the limbs is highly variable. The crinkles are generally short; rarely their length/height ratio is as great as in figure 41. They terminate by pitching (Fig. 42). Where one crinkle vanishes, another appears overlapping it and continuing about the same amount of shortening. In this respect the crinkles resemble large-scale Appalachian-type folds. The axes of overlapping crinkles necessarily diverge. Where the crinkles are short and stout, the angle of divergence is large and a plane passing through two divergent axes is always steeply inclined to the crinkled plane. Divergence is increasingly

*Much of this discussion, especially the description of the three types of crinkles, is adapted from Sander (1948, pp. 148-158).

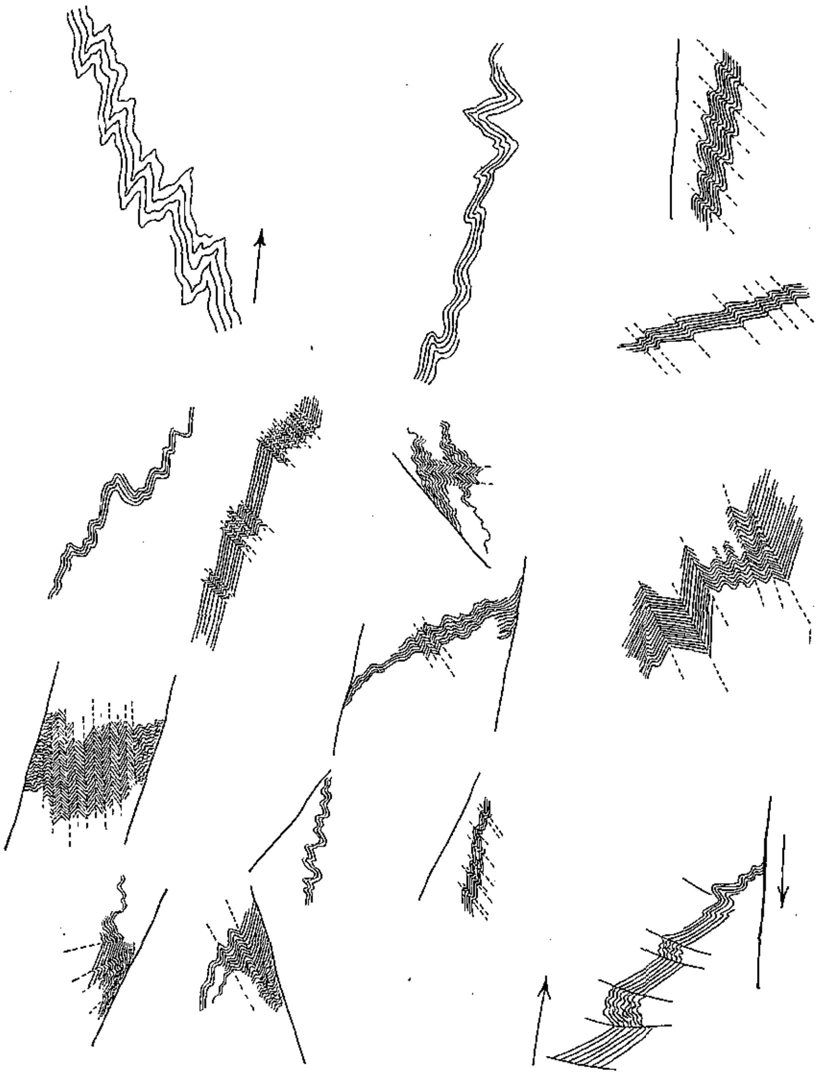


Plate 5. Sketches of crinkled flow cleavage

apparent in crinkles of smaller and smaller apical angle, i.e., in the tighter folds. Axial divergence thus relates to the crinkles' amplitude, crestal distance, and length, which are themselves related to the competence of the rock, the strength of the initial foliation, the size, shape, and character of the rock's inhomogeneities, and the stress distribution. Stated briefly, the divergence of the axes is traceable to inequalities in the rock or the deforming stress, with respect to the **a-c** plane (Sander, 1950, pp. 277-78).

True axial divergence, just discussed, is often easily confused with an **apparent** divergence which originates by oblique shear (Fig. 43). The shear planes, alternately cross-cutting and following the crinkled cleavage, may be inclined at any angle to the coordinate axes of the crinkles but are generally about perpendicular to the **b-c** plane and inclined about 30° to **b**. They are variably spaced, often five cm. or more apart, and have a fairly regular orientation within the m-domain. Unless the crinkling is simple enough that the trace of the broken cleavage can be seen on the oblique shear planes, as **t** in figure 43, apparent axial divergence may be mistaken for true axial divergence, or it may be erroneously inferred that there are two unrelated sets of crinkles. The presence of apparent axial divergence is sometimes revealed by a tendency of the crinkled mass to break into wedges.

Fig. 41. Lineation in *b*. Small crinkles in mica schist.

A late set of crinkles may be superposed on an earlier set (figure 44). Rarely three sets are superposed. Age relations are clearly evident when there are only two sets, but when there are three, the crinkling is so irregular that the time sequence is often obscure. Superposed crinkles are common in the Copperhill area and in the thick schist sequence along Hothouse Creek. The large-scale adjustments which produced

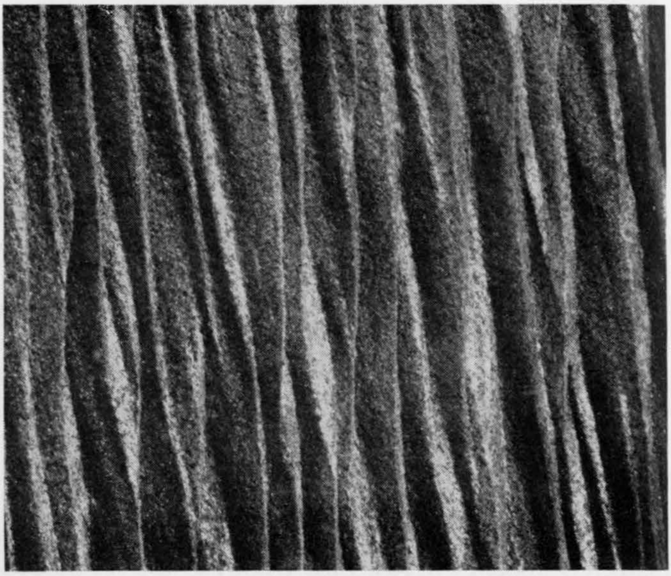


Fig. 42. Divergent crinkle axes in mica schist. Two-thirds natural size.

superposed crinkles in one area might be the same as those which produced oblique shearing in another.

The symmetry of the crinkles may be constant within a given outcrop or locality, and radically different from one locality to another. Symmetry is one of the most useful characters of the crinkles.

**Types of crinkles.** Where the outcrop is a plane roughly $\perp$ B, a few square meters in size, and shows crinkles of dm. to cm. size, the following types are again and again distinguishable:

Type 1. The crinkles show every combinable shape, position, and relative movement $\perp$ B. (The orientation of B may vary from the domain of one crinkle to another, but is constant in the $m^2$- domain.) The crinkles have no plane of symmetry. Their relative movements vary as much whether they are interpreted as shear folds, flexure folds, or folds due to both shear and flexure. Qualitatively, Type 1 crinkles are very significant. Their presence is clear evidence that there has been polytrope folding in the $m^2$-domain, moreover that the folding is a result not of homogeneous, laminar flow but of compression between more rigid "jaws", involving external rotation of the crinkled mass.

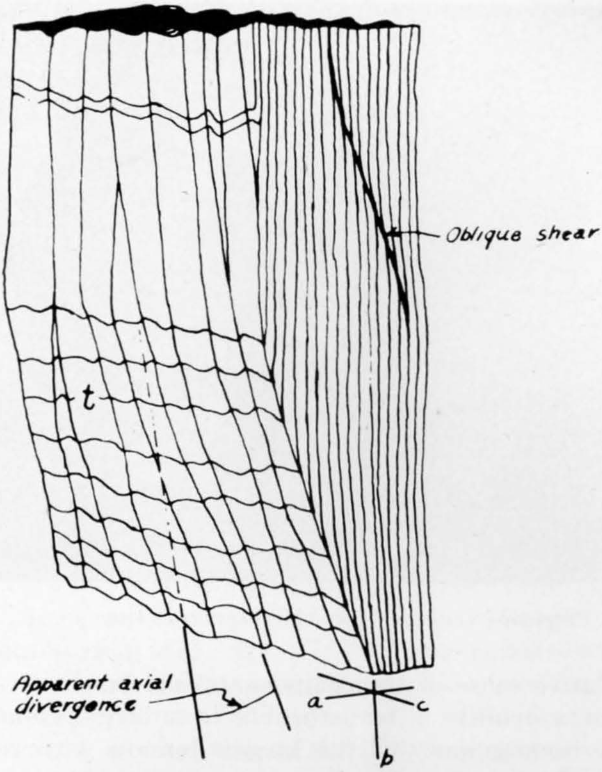



Fig. 43. Apparent axial divergence caused by oblique shear; $t$ marks the trace of a cleavage plane on an oblique shear surface; $a$, $b$, and $c$ are the coordinate axes of the crinkles.

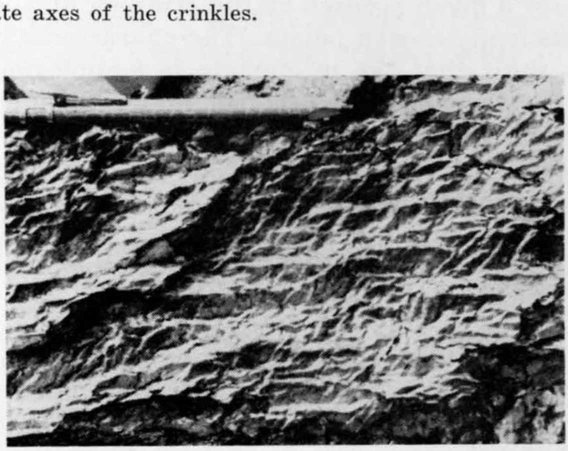


Fig. 44. A late set of crinkles (inclined 60° to the pencil) superposed on an earlier set (horizontal). Mica schist in the Copperhill area.

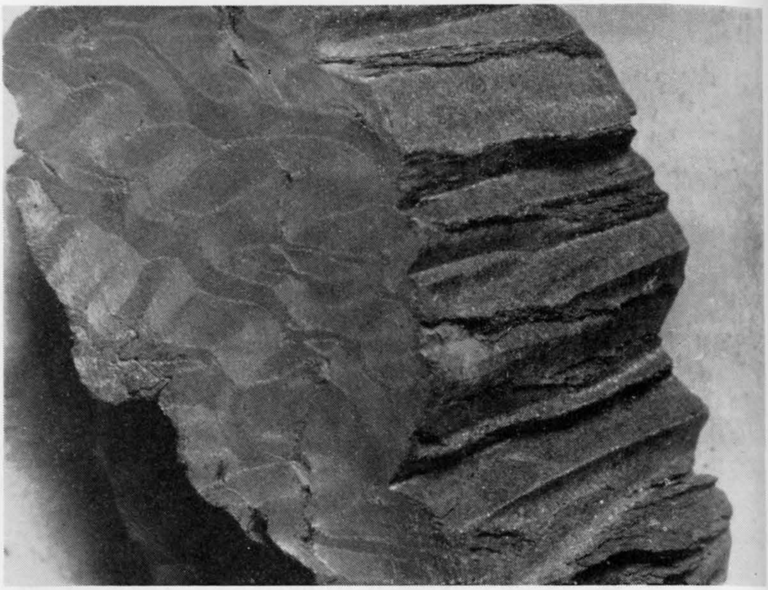


Fig. 45. Polytrope crinkles. Two-thirds natural size.

The relative sense of the componental movement $\perp$ B in the domain of a crinkle is transferable to a larger domain only when the homogeneity of the larger domain with respect to each movement sense has been proven. Homogeneity can be accepted when the movement sense has been shown to be the same, for a given position on the crinkles, at a large number of statistically-chosen points. The characteristics of Type 1 crinkles show that the $m^2$-domain is inhomogeneous with respect to movements in the crinkles. The componental movements determined from the individual crinkles are therefore not summable to the relative movements of the larger domain. If the initial orientation of the crinkled planes is not apparent, the relative movement in the dm-domain is ambiguous, and the usefulness of Type 1 crinkles is restricted to what they reveal about the **quality** of the deformation.

In the Copperhill area, the initial orientation of the crinkled planes is frequently determinable. It may be revealed by beds that were incompetent enough to develop faint flow cleavage but too competent to participate in the crinkling. The use of such beds can lead to error if fracture cleavage is mistaken for flow cleavage. The initial orientation is better determined from graded beds (figures 24 and 46) where, the intensity of the crinkling decreasing with increasing competency, the crinkled planes tend to straighten and swing around to their initial orientation toward the most competent layer.

Type 2. The crinkles have two planes of symmetry, one $\perp$ B (parallel to the **a-c** plane), the other parallel to B (parallel to the **b-c** plane, or axial plane). Such highly symmetrical crinkles develop by flexure. The movement they indicate in the domain of the crinkle and in the larger domain is compression perpendicular to the axial planes, regardless of whether the larger domain has a plane of symmetry parallel to B.

Such high symmetry in crinkles and in large folds is rare, or rarely preserved. Much more frequently the following type is encountered:

Type 3. The crinkles have only one symmetry plane, which is $\perp$ B. They are not variously oriented, as in Type 1, but similarly oriented (Fig. 47). As in Type 2, the componental movements determined from the individual folds are summable to the relative movement of the same sense in the larger domain. For this reason, and because of the wide occurrence of crinkles of this type, the decipherment of the sense of relative movement $\perp$ B in the individual crinkles is of particular interest.

**Relative movement $\perp$ B.** The effects of flexure and planar shear being often combined, the relative sense of the com-

Fig. 46. Two examples of graded beds with polytrope crinkles at the top. The dashed line indicates the orientation of the cleavage prior to crinkling. See text for explanation.

ponental movement $\perp$ B is not always easy to determine.

Figure 48 is an illustration. Sketch A represents the passive

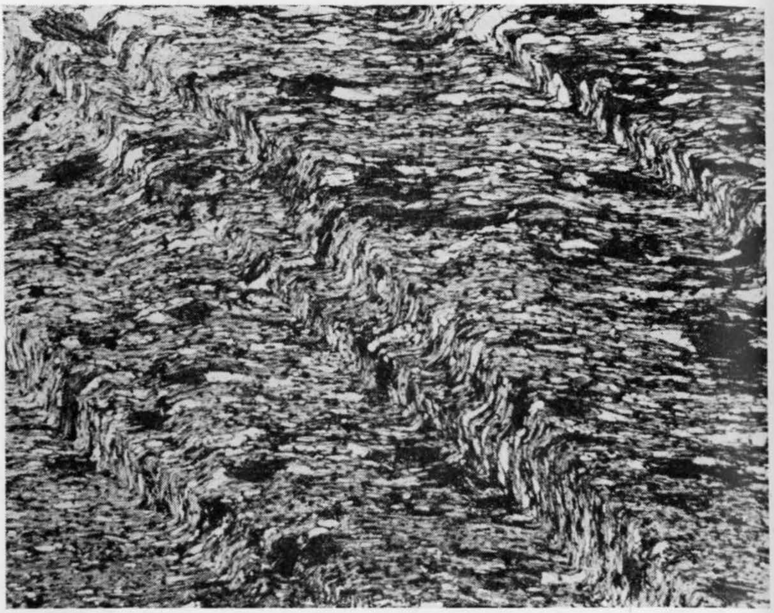


Fig. 47. Photomicrograph, uncrossed nicols X8. Crinkles with one plane of symmetry $\perp$ B (parallel to the plane of the photograph). Note the effects of inhomogeneities.

bending of a mechanically unimportant plane by shear at an angle to the plane. Sketch B shows the flexure of a mechanically effective plane, with shear mainly along the plane. The arrows indicate the direction of relative movement. Sketch C is a relationship observed in crinkled schist bounded by metagraywacke; the heavy dashed line represents the initial orientation of the cleavage planes as preserved in uncrinkled beds at the same outcrop. If the initial orientation of the cleavage were unknown, the relative movement of one bed with respect to another would be ambiguous, unless revealed by the relative movement along the cleavage planes.

When uni-directional laminar flow or shear along a single set of planes is involved, uncomplicated by flexure, the final configuration of a plane that was originally inclined to the direction of movement indicates the relative movement uniquely, as in Fig. 49. Ambiguity arises when the distribution of the shear planes is inhomogeneous, as in Fig. 50, or when the deformed surfaces were not planar at the beginning.

When laminar flow or shear along two sets of simultaneous planes is involved (Fig. 51), the direction of relative movement in a large domain cannot be uniquely deduced from the configuration of a deformed plane, but must be determined from a summation of the relative movements along the shear planes.

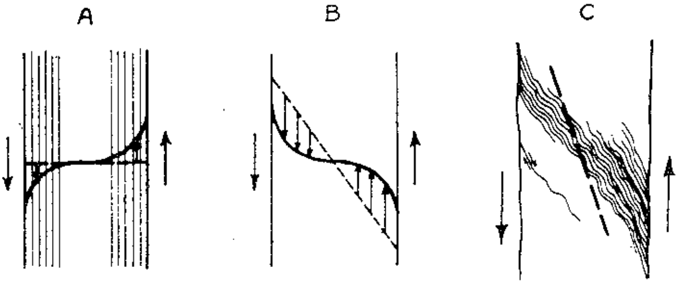



Fig. 48. A. Passive bending of a mechanically unimportant plane. B. Flexure of a mechanically effective plane. C. Flexure of cleavage in mica schist. The dashed line indicates the initial orientation of the deformed plane.

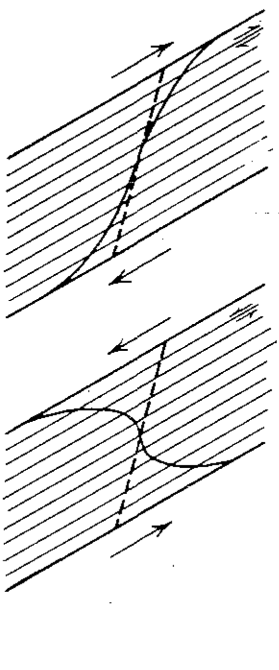


*Fig 49*

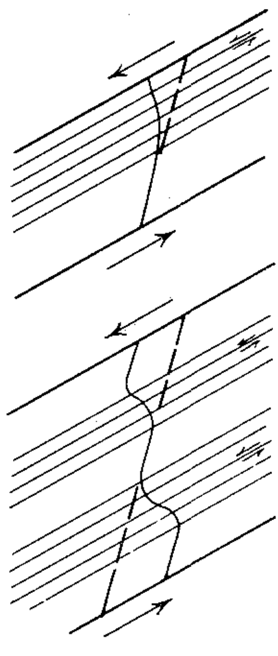


*Fig 50*

Fig. 49. Uni-directional shear along one set of homogeneously distributed planes.

Fig. 50 Uni-directional shear along one set of inhomogeneously distributed planes.

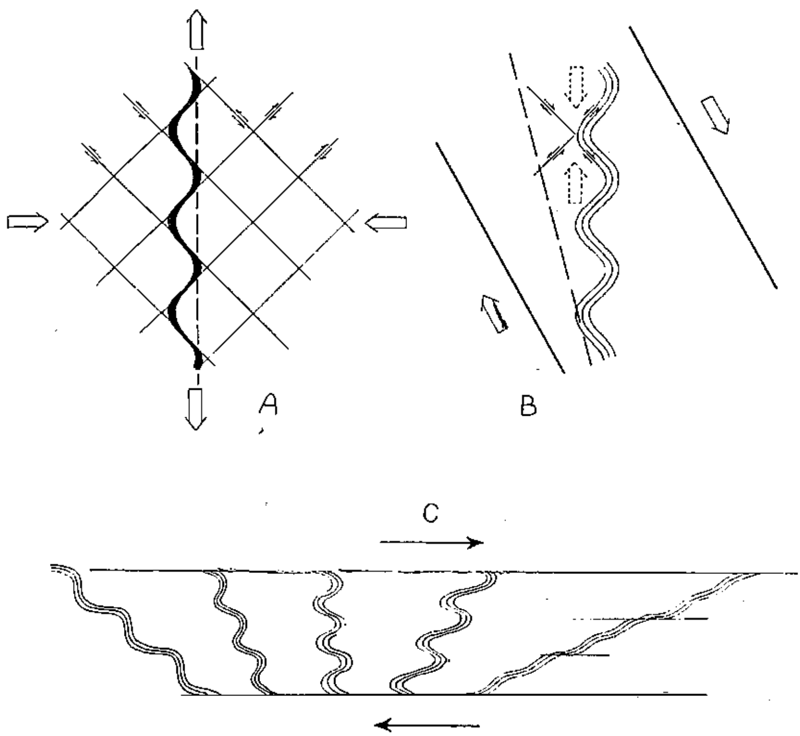



Fig. 51. A. Crinkling by simultaneous shear along two sets of planes. B. Crinkling by flexure of earlier planes. Large open arrows indicate componental movements, single-barbed arrows show relative movement in a smaller domain. C. Ideal configuration of homogeneous planes externally rotated by crinkling; successive stages left to right. Compare with Plate 6. Variation in the configuration of the crinkles is evidence of inhomogeneous deformation.

Crinkles can originate by planar shear, as in A (Figure 51), or by flexure, as in B. The deformation patterns in the two cases are similar, but the directions of relative movement on the limbs of the crinkles are reversed.

The crinkles in the Mineral Bluff quadrangle are mainly the result of flexure; however, their configuration is often complicated by shear along one or more sets of planes as in figures 50 and 51A. Figures 48-51 show that apparent "drag" in the configuration of the crinkled planes is not always a safe indication of relative movement.

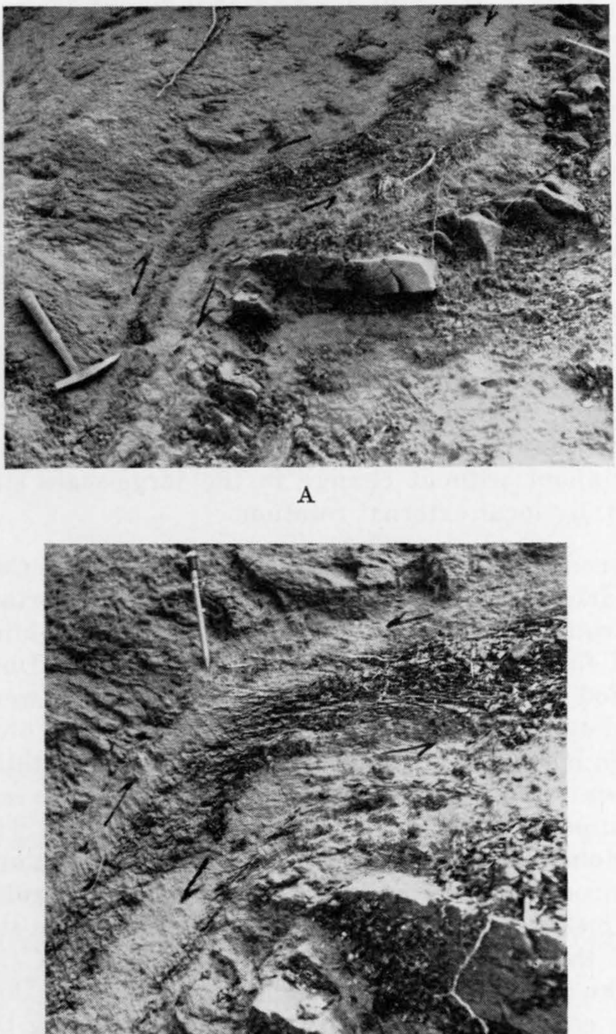
When the initial orientation of the crinkled planes can be determined, the relative sense of the componental movement $\perp$ B is at once apparent, regardless of the type of crinkle involved. When the initial orientation is unknown, relative movements in the large domain can still be obtained, from Types 2 and 3, by evaluating the roles of flexure and shear in the formation of the crinkles, deducing the relative movements in the domain of a few crinkles, and summing these movements.

**Origin and significance of the crinkling.** The formation of flow cleavage involves elongation in the cleavage plane; flexural crinkling of flow cleavage entails shortening in the same plane. The presence of flexural crinkles, then, is **prima facie** evidence that the deforming stresses shifted, with respect to the crinkled mass, after the formation of the cleavage. Such a shift might be caused by change in the regional stresses (as from one period of deformation to another) or might be brought about without change in the large-scale stress distribution, by local external rotation.

The predominance of polytrope crinkles in the Copperhill area clearly points to external rotation as an important cause of the crinkling. Evidence of rotation by drag folding (Fig. 52) and faulting is observable at many places. During the last period of deformation, when the large folds were tightened and overturned, when new folds developed on older ones, and when much of the local faulting originated, rotation must have been a major phenomenon. Each rotation represents an adjustment which shifted the stress distribution within the rotated domain. Small-scale adjustments were taken up largely by the incompetent schist layers which, already strongly foliated, adjusted by crinkling rather than shear. The structural trend of the crinkles coincides with the trend of the older folds. The intensity of the crinkling correlates with the degree to which earlier folds are disrupted. The crinkling therefore appears to be mainly an incident in the tightening of the earlier structures.

There is evidence that the Ducktown-type sulphide bodies were emplaced during and structurally controlled by the movements of the last period of deformation. The crinkles, being the most easily observed cues to the movements of this period, are therefore particularly significant.

Observable features pertinent to the interpretation of crinkling:



A

B

Fig. 52. A. External rotation and crinkling of flow cleavage by drag-folding. Crinkling occurs where the movement (arrows) is opposed to the cleavage. B. Enlargement of part of the drag fold in A. The bed on top is metagraywacke; the crinkled bed is mica schist; the fractured bed below is quartzite.

1. Symmetry of the crinkles.
2. Their average size, and size variation.
3. The configuration of the crinkles; whether their crests

are rounded or sharp, their limbs straight or curved; the apical angle; the relation of limb curvature to the axial plane.

4. $\frac{\text{Length}}{\text{height}}$ ratio.
5. Axial divergence. Angle of divergence and its orientation with respect to the crinkled plane.
6. Apparent axial divergence. Orientation, spacing, and size of the oblique shear planes.
7. Relationship between the crinkled planes and the bedding planes, especially where the crinkled beds grade to more competent layers.
8. Whether bedding plane faults separate the crinkled layers from the bounding, more competent rocks.

## FAULTING

Faults with a displacement of a few inches to a few feet can be observed in nearly all parts of the quadrangle. They are particularly common in the areas that show drag folding and crinkling. Faults with a displacement of more than 100 feet were recognized only in the trough of the Murphy syncline and on the southeast limb of the Copperhill syncline.

**Faults of the Murphy syncline.** LaForge and Phalen (1913) mapped three faults in the Murphy syncline: (1) the Young Stone Creek fault, through the west side of Blue Ridge, through Hogback Bend, and along the course of Young Stone Creek; (2) the Murphy fault, along the western marble belt which passes through Blue Ridge and Mineral Bluff; and (3) the Whitestone fault, along the eastern marble belt. The evidence given for each fault is good exposure at the type locality and the fact that the fault was "traced".

The Young Stone Creek fault is not shown on Plate 3, as no evidence for it was observed. Where LaForge and Phalen mapped the fault, no break in the stratigraphic succession is apparent. If the fault exists, it is a bedding plane fault.

The Murphy fault is well marked. Its position is defined by localized crinkling, distortion and small-scale shearing in the Nottely quartzite, and by a stratigraphic break at which about 1500 feet of metasediments are missing. Slickensided surfaces are exposed at many places between Blue Ridge and Mineral Bluff. Only to the northeast of Mineral Bluff, between the old talc prospects and the east side of the quadrangle, is the fault poorly exposed. The possibility that the Andrews formation is missing because of non-deposition rather than faulting is discounted because of the remarkable constancy of the metasediments of this belt toward the northeast and southwest, and because of the abundant evidence of faulting where the formation is missing.

Where the Tusquitee quartzite is cut off near the mouth of Weaver Creek, exposures are poor, but chunks of slickensided quartzite and curly schist in the residuum indicate the presence of a fault. The course of faulting to the northeast and southwest is marked by the thinning of formations and by zones of localized distortion. The discontinuous faults in this area are what LaForge and Phalen interpreted as a continuation of the Whitestone fault.

The faults of the Murphy belt should be represented not as large single breaks but as narrow zones in which there are many small faults. The small faults in these zones vary in attitude but nearly always dip toward the core of the syncline. The relative movements, sometimes determinable from drag, show upward and outward crowding of the core. The Murphy syncline being a bent fold, this type of faulting is the expected type. Although the long continuous breaks visualized by earlier workers are a fair representation at a large scale, they obscure the discontinuity and significance of the faults of this belt.

**Fault at the southeast side of the Copperhill syncline.** Unusually intense crinkling, local slickensiding, and discordant trends in the bedding indicate either a major fault or a very distorted zone. The amount of displacement is uncertain.

## BOUDINAGE

Boudinage is a common structure in this area, although rarely conspicuous. It is most common in the crest of the Morganton anticline (pseudodiorite boudins), the trough of the Murphy syncline (quartzite boudins), and in the Hughes Gap formation in the vicinity of Gravelly Gap (pseudodiorite and quartzite boudins).

Fig. 53. Boudinaged quartzite bed, bounded by quartz-sericite schist. Nottely quartzite, near Mineral Bluff. The boudins have been rotated into the cleavage plane.

In a plane $\perp$ to the fold axis, the boudins are elliptical or tabular with rounded ends, and are generally one to ten feet long. In a plane parallel to the bedding, they are roughly rectangular, with the long dimension usually about parallel to the fold axis.

Boudins originate by tension. When a series of competent and incompetent beds are folded (by bending), the competent beds are often segmented by fractures normal to the bedding. As the segments are pulled apart by the relative movement of higher beds over lower ones, the less competent beds flow into the gaps. All stages in the development of the boudins can be observed in the southeast part of the quadrangle. Their best development, in the trough and crest of the two main bent folds, accords well with their origin.

Many of the boudins have been rotated so that their long dimensions lie in the cleavage plane. (Fig. 53).

## STRIATION

The fault surfaces are usually striated. The striae are particularly noticeable where the fault surfaces are coated with black manganese oxide.

Striae on bedding planes are sometimes visible but rarely conspicuous.

## METAMORPHIC HISTORY

The metamorphic history of the southeast part of the Mineral Bluff quadrangle is somewhat different from that of the northwest part. The difference lies mainly in the extent and kind of deformation, and in the time relations between deformation and recrystallization.

Two periods of deformation are distinguishable; however, it is not clear whether they are separated by a long time interval or whether they are merely successive stages in one main event.

The major folds and cleavages originated during the first period. The main fold in the southeast was an open, bent syncline, uncomplicated by second-order flexures. In the trough of this fold bedding-plane foliation developed in thinly banded slates and phyllites. Flow cleavage developed locally in the least competent beds on the limbs. The main fold in the northwest was an anticlinorium. Although primarily a bent fold, it had marked shear fold characteristics. Strong axial plane flow cleavage formed in this fold in the incompetent layers; fracture cleavage with similar orientation developed in some of the more competent beds.

The micas in the rocks at the southeast side of the quadrangle are generally coarse and stout. They show by their textural relations that they grew prior to a period of regional deformation and that they were mainly unoriented at the time of growth. In contrast, the micas at the northwest side of the quadrangle are markedly tabular and strongly oriented. Their textural relations show that they grew during and after the first period of deformation. These facts may indicate that recrystallization began first at the southeast side of the quadrangle. An alternative explanation, one more in accord with the variation in the character of the deformation, is that strong penetrative stress prevailed in the northwest and not in the southeast during the micas' growth, which took place in both areas at roughly the same time. In either case, recrystallization was complete before the second period of deformation began.

During the second period of deformation old folds were tightened, new minor folds were produced, earlier cleavages were rumpled, and there was much local faulting. The effects of the late movements are more apparent toward the northwest than toward the southeast. This is due, at least in part, to the condition of the rocks when the late movements began. The incompetent beds in the southeast consisted largely of coarse interlocking mica and were poorly foliated; they yielded to the late movements by shear. Incompetent beds in the northwest, being strongly foliated, yielded instead by conspicuous crinkling. The tightening of the Murphy syncline developed faults in the trough and a few minor folds on the limbs. The scarcity of secondary folding due to drag is attributed to the bedding plane foliation, which may have helped to distribute the slippage as higher beds moved over lower ones. The tightening of the anticlinorium to the northwest, with its numerous second- and third-order folds, caused external rotation of large masses and consequent local reversals in stress, by which the strongly foliated incompetent layers were greatly deformed.

## POST-CRYSTALLIZATION ALTERATION

After the last period of deformation, the rocks were pervasively altered. Biotite, hornblende, and garnet were chloritized, staurolite sericitized, feldspar clouded, and sulphides shifted about in the rocks. So pervasive are the effects that they are visible in thin sections from every part of the quadrangle. Other alterations occurred at about the same time but are more local in their effects: these include epidotization, silicification, calcification, the introduction of sulphides, and the formation of talc.

**Chloritization.** Because chloritization affected minerals which are present in all the rocks, it is the most noticeable alteration. Biotite is the principal chloritized mineral. In the average rock one to five percent of the biotite has been altered. The percentage is higher toward the southeast side of the quadrangle.

The chloritization is closely associated with microfractures, cleavages, and crystal margins. Chlorite commonly pseudomorphs small biotite crystals. The pseudomorphs contain pleochroic halos, like the unaltered biotite, and are usually "dusted" by fine, opaque matter. The chlorite being less ferruginous than the biotite, this opaque matter probably is exsolved iron. Large crystals of biotite, garnet, and hornblende are altered to patchy or sheaf-like chlorite aggregates. The garnets alter only on their margins except when they are fractured. The chlorite is pleochroic in pale green, weakly birefringent, and has a small 2V. Its pleochroism and birefringence, hence its composition, varies with the mineral it replaces and with other factors that were not evaluated.

Large, unoriented chlorite porphyroblasts (Fig. 54) were noted in quartz-mica schist along the top of the Hughes Gap formation and in metagraywacke near the bottom of the Copperhill formation.

**Sericitization.** One of the principal minerals affected by sericitization is staurolite. The intensity of the alteration varies erratically, but is perceptible in staurolitic beds throughout the quadrangle. The alteration is particularly strong toward the southeast. In the Dean formation on the east limb of the Murphy syncline nearly all the staurolite has been converted to sericite (shimmer aggregates). The associated biotite has been largely changed to chlorite. The matrix of the rock is fine sericite and quartz. Sericitization of staurolite is strong in the Ducktown Basin, just to the northwest of the Mineral Bluff quadrangle. This alteration is not a surface phenomenon, as it is encountered hundreds of feet below the ground water level, in the Ducktown mines.

The sericitization of the staurolite began along crystal margins, cleavages, and fractures (Fig. 55). Much of the quartz that was included in the staurolite crystals remains in the sericite pseudomorphs. The included garnets have usually vanished; their positions are marked by patches of chlorite. The sericite is a 2M muscovite polymorph (Smith & Yoder, 1953). It is low in potash (Table 11) and probably high in water.

The conversion of staurolite to sericite requires potash from some external source. The chloritization of biotite (release of potash) is usually closely associated with the sericitization

## TABLE 11

### Potash-Soda-Lime Content* of Sericite in Pseudomorphs of Sericite After Staurolite

| | $K_2O$ | $Na_2O$ | $CaO$ |
|---|---|---|---|
| 1 | 6.3 | 1.3 | 0.08 |
| 2 | 6.8 | 1.1 | 0.1 |

1. Surface collection from Hackney's farm, near Blue Ridge, Georgia.
2. From Calloway B shaft, depth 410 feet, Ducktown Basin, Tennessee.

*Determined with flame photometer. The potash content of muscovite with the ideal composition $KAl_2(AlSi_3O_{10})(OH)_2$ is 11.8%.

of staurolite (reuse of potash). Both alterations require an external supply of water.

Sericitization is not restricted to the staurolitic beds. Large masses of fine-grained slate and phyllite in the trough of the Murphy syncline have been altered to silvery quartz-sericite schist. The alteration of these rocks was not investigated, microscopically, because unweathered specimens could not be obtained.

**Epidotization.** Only in the southeast corner of the quad-

Fig. 54. Microphotograph, plain light, X22. Unoriented chlorite porphyroblasts in quartz-mica schist (small garnet in the center of the photo graph).

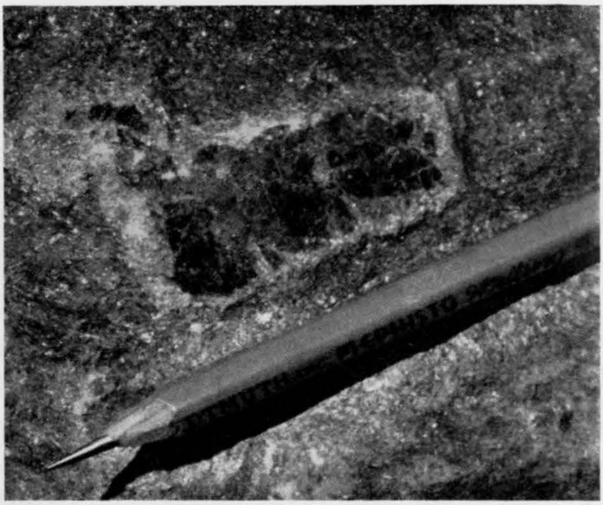


Fig. 55. Staurolite crystal, partly altered to sericite, in quartz-mica schist. Natural size.

rangle was this type of alteration observed. Alteration is strong three-quarters mile north of Morganton on the east edge of the quadrangle. Weaker alteration is apparent at many places on the shores of Lake Toccoa.

The megascopic effects of the epidotization of metagraywacke are a faint greenish tinge and a waxy cast. Rarely, the metagraywacke is cut by thin epidote seams. The epidotization of mica schist is more conspicuous. The altered schist is usually a duller shade and has suffered a pronounced loss of foliation.

## TABLE 12

### Modal Analyses of Epidotized Metagraywacke

| Specimen | 1A8 | $1A7_a$ |
|---|---|---|
| Quartz | 37.5 | 45.2 |
| Feldspar in symplektite | 26.6 | 31.8 |
| Other feldspar | 4.9 | 1.0 |
| Epidote | 13.0 | 11.4 |
| Biotite | 4.7 | 6.7 |
| Muscovite | 6.2 | 1.6 |
| Chlorite | 5.2 | 0.9 |
| Black opaque | 1.1 | |
| Titanite | 0.8 | 1.4 |

Modal analyses of two epidotized metagraywacke specimens are given in Table 12. The specimens were collected in a road cut six-tenths mile northeast of Hollywood. The epidote is biaxial negative with a 2V of 72-75°; it occurs in small irregularly-shaped masses of subhedral grains that are clustered about splayed remnants of biotite. Some of the biotite has altered to chlorite. Grouped about altered remnants of muscovite are quartz grains and small irregularly-shaped masses of feldspar. This feldspar is biaxial negative with a 2V of about 85°, but does not stain in cobaltinitrite solution. Some muscovite crystals are partly altered to chlorite. Where most of the mica has been altered, the matrix is fine symplektite.

The same mineralogical changes, biotite$\rightarrow$epidote and chlorite, muscovite$\rightarrow$chlorite and symplektite, are observed in the epidotized mica schist.

**Significance of the post-crystallization alteration.** The chemical reactions involved in the post-crystallizational alteration of the rocks may not be considered in terms of single mineralogical transformations. Each alteration process was dependent on the chemical environment, i.e., on minerals other than those which are obviously involved. The reacting constituents came from more than one source. Each alteration process represents an interplay of chemical exchanges which resulted in complete breakdown and reconstruction of some minerals and only minor structural adjustment in others. A formula obtained by balancing reactants and products thus may not represent what actually occurred for it does not take into account the unbalance which is permitted by latitude in the atomic structural requirements of the participating minerals.

Other than bulk composition, pressure, and temperature, the environmental factors which may affect mineralogic changes in rocks have not been experimentally investigated in the laboratory in a systematic fashion. Such factors as the previous history of the reacting materials, their atomic structural arrangement, and the catalytic effects of impurities are known to be important. Laboratory investigations have dealt so far with simple reactions at high temperatures (400°C and up). The data do not make it possible to interpret the P-T conditions of formation of many common metamorphic minerals. The data do show, however, that minerals like chlorite, sericite, regarded by some as good P-T indicators, may form under a wide range of physical conditions (Yoder, 1952, pp. 615-23), and that interpretations of P-T conditions based on the presence of these minerals alone are unreliable.

The alterations mentioned above are usually ascribed to retrograde metamorphism. This explanation stems primarily from the belief that sericite and chlorite are low-temperature minerals. Because it has been experimentally shown that sericite and chlorite can form at high temperatures as well as low, one of the principal reasons for regarding their formation as a retrograde process has been erased. Study of the altered rocks shows that the alteration might have been affected, so far as composition is concerned, simply by the addition of water. The water might have been introduced at the same temperatures and pressures as those at which the altered minerals originally grew.

Sericitization and chloritization in this area are regarded as evidence only that water pervaded the rocks. Textural relations show that this happened subsequent to the last period of deformation. The question of what temperatures and pressures prevailed at that time is left open.

# PETROFABRICS

The grain orientation of specimens from seven localities (Fig. 56) was investigated microscopically. The results are presented in a series of statistical diagrams, 1-14.

Diagrams 1-6 show the fabric elements of a metagraywacke specimen collected six-tenths mile east of Copperhill, Tennessee. The specimen comes from the east limb of a small anticline that is overturned to the northwest and faulted along the crest.

The quartz axes show a girdle ⊥ **b** with strong maxima somewhat irregularly disposed about **a** in positions I and IV (Fairbairn, 1949, p. 121). The plane of the diagram is not exactly ⊥ **b**. One irregularly-shaped maximum is inclined about 30° to the position of Maximum V, and has no symmetrical counterpart on the opposite side of the **a-c** plane. The relative movements along ③, ④, and ⑤ are related to crinkling, which was subsequent to the main recrystallization of the minerals (compare with Fig. 51B). Although the quartz has again recrystallized or at least healed its broken boundaries since crinkling, its optic orientation is probably related to the earlier movements, connected with elongation in the flow cleavage plane ②. The irregular shape of the maxima might be attributable to dispersion of the quartz axes during the period of crinkling.

The cleavage poles of the micas define a strong maximum ⊥ the flow cleavage (diagrams 2 and 3). The spread of the maximum in the **a-c** plane is due to crinkling, visible in Plate 7. The maximum of the biotite coincides with that of the muscovite. The biotite, however, is not as well oriented as the muscovite.

The oligoclase grains (diagrams 4, and 5, and 6) tend to be oriented with one of their cleavages parallel to ③ and ④. The grain rotation which could produce this orientation indicates greater movement than that shown by the slight displacements from which relative movements along ③ and ④ are deduced. This is one of the reasons for believing the orientation of the quartz and feldspar may be related to earlier movements. The elongation in the cleavage plane known to have occurred earlier (see section on cleavage) might have been accompanied by shear along planes of the same general orientation as ③ and ④ but with opposite movement sense. It is shown in the section on A.V.A. that planes of about the same orientation as ③ and ④ are an integral part of the total quartz fabric.

Diagrams 7 and 8 show the orientation of quartz in two metagraywacke specimens collected two miles apart in the northeast corner of the quadrangle. Both specimens come from nearly vertical beds on the west limb of the Murphy syncline. The diagrams show girdles that are $\perp$ **b** and inclined to the bedding. The maxima in 7 differ in position from those in 8; in neither case is the relation between s-planes and maxima clear.

The succeeding diagrams show the orientation of quartz in quartzite beds in the Murphy syncline. Diagrams 9, 12, and 13, representing the Tusquitee quartzite on the west limb of the syncline, are similar to each other in that they have a well-marked girdle $\perp$ to **b** and a strong maximum where the girdle intersects the cleavage. (The girdles show that **b** is not exactly $\perp$ the plane of the diagrams.) Diagram 14 represents the Nottely quartzite on the east limb of the syncline, and is similar to diagrams 9, 12 and 13. The specimen represented by diagrams 10 and 11 is Nottely quartzite from the hanging wall of a thrust fault. A girdle $\perp$ **b** is not apparent in this specimen. The large uncrushed quartz grains (No. 11) show a strong maximum $\perp$ **b** in the cleavage plane, like the quartz at other neighboring localities (Diagrams 9, 12, and 14). This maximum is not shown by the small quartz grains, whose strongest maximum is $\perp$ cleavage plane.

Slight discrepancies between the position of cleavage or bedding and the position of the principal quartz maxima are probably not significant, because the attitudes of cleavage and bedding are variable and were plotted from field measurements. Even where cleavage and bedding are visible in the thin section, they still could not be plotted more accurately than 5-15 degrees.

The quartz diagrams display these general features:

1. A girdle $\perp$ **b**.
2. A maximum of variable strength in or very close to the bedding plane and $\perp$ **b**.

3. Another maximum, usually strong, $\perp$ **b** and in the flow cleavage plane.

4. A slightly more regular distribution of maxima within girdles at the southeast side of the quadrangle than at the northwest side (compare diagrams 1, 8, and 14).

## INTERPRETATION OF THE DIAGRAMS

The quartz maximum $\perp$ **b** in the plane of the cleavage is interpreted as Maximum I and is correlated with elongation in the cleavage plane, normal to the fold axis.

The quartz maximum $\perp$ **b** in the bedding plane is correlated with the bedding plane foliation, which is well developed where this maximum is strong.

The more regular distribution of quartz maxima within girdles at the southeast side of the quadrangle coincides with the greater importance of bending as compared with shear in the deformation of this part of the quadrangle.

On the west limb of the Murphy syncline, 1.5-3 miles from the trough, the flow cleavage is roughly parallel to the axial plane of the fold. This cleavage fades out about 1.5 miles west of the trough and is replaced by a cleavage that dips east less steeply than the bedding. East of the trough, the flow cleavage is again about parallel to the axial plane of the fold. This anomalous relationship is partially clarified by specimens 2NS and $1X6_a$. The quartz axes in these specimens (Diagrams 9 & 13) define a strong maximum, in addition to the maxima associated with cleavage and bedding, which is normal to the fold axis and hard to explain in relation to visible s-surfaces. This maximum might represent an axial plane cleavage, preserved in the fabric although not megascopically visible. It might therefore mean that the axial plane cleavage to the east and west continues across this belt, preserved in the fabric, but obscured by a stronger cleavage which formed partly by fracture (thereby preserving the earlier orientation) and partly by flow. This interpretation agrees with field observations that the megascopic cleavage in this belt is locally associated with the axial planes of minute crinkles formed from the deformation of a weakly-visible earlier cleavage.

# EXPLANATION OF PETROFABRIC DIAGRAMS

The diagrams are equal area net projections on the lower hemisphere; contour counter one percent of the total area; distance between counts 0.5 cm. The diagrams are roughly $\perp$ the **b** fabric axis. The strike-dip symbol in the center gives the field orientation of the diagram. The strike is measured azimuthally from the north. The overturned symbol indicates that the dip is away from the observer.

The field location of each diagram is shown in figure 56.

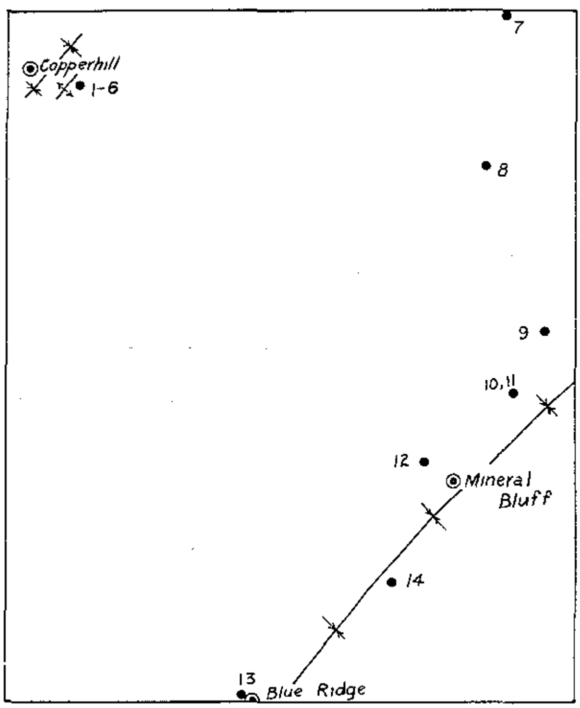



*Fig 56. Location of petrofabric diagrams.*

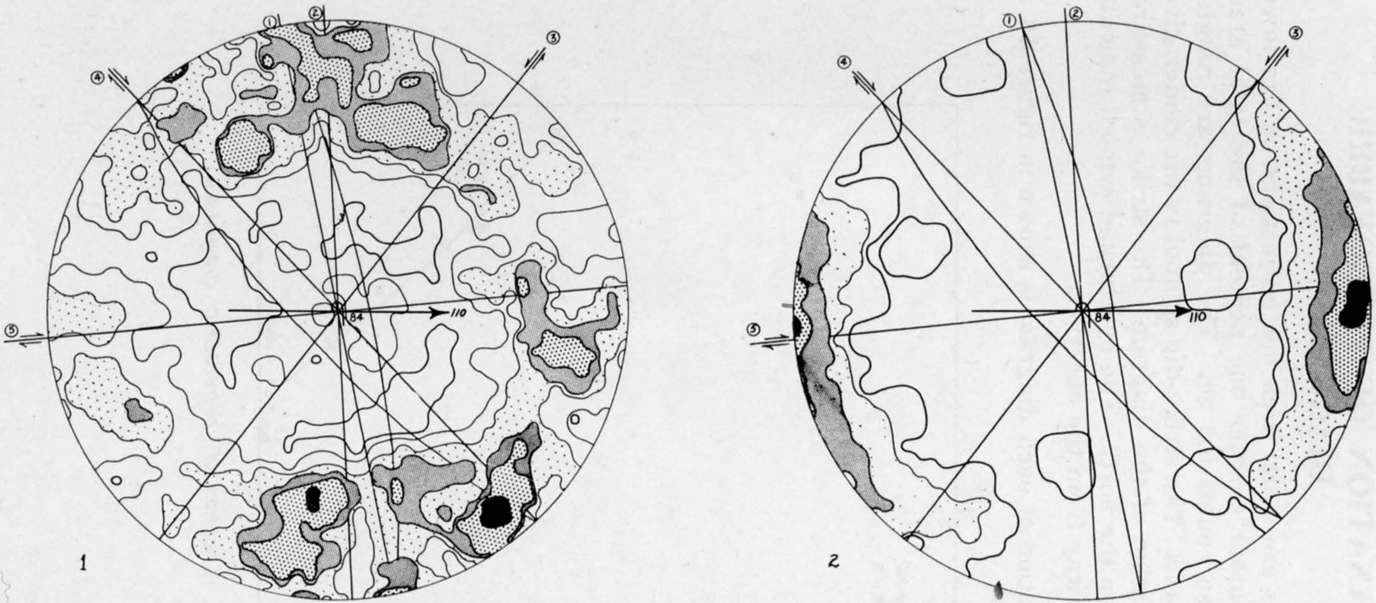
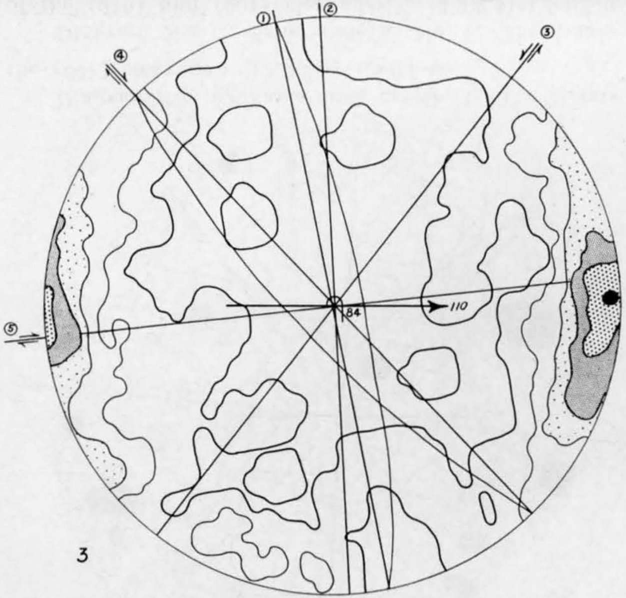
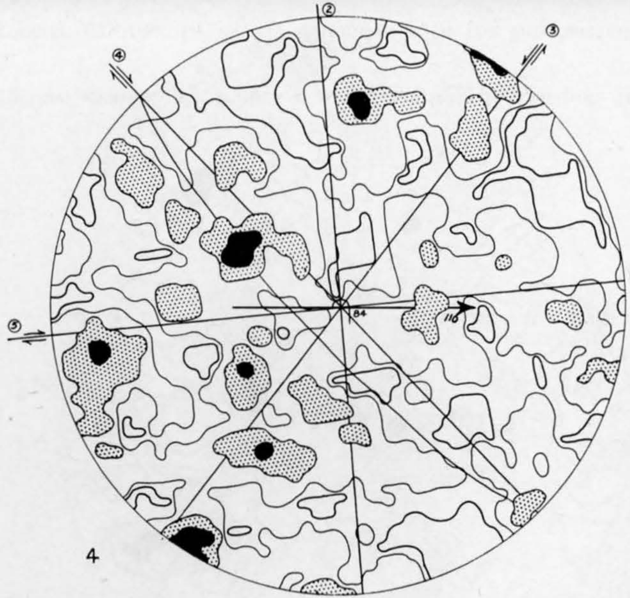



Diagram No. 1. Slide $G5_b$. Metagraywacke. 447 quartz axes. 3-2-1.5-1-0.5-0 ① is bedding. ② is strong flow cleavage. ① and ② cannot be distinguished in the thin section, although readily separable in the hand specimen. ③ and ④ are faintly visible as lines along which there are concentrations of quartz grains; relative movements were deduced from occasional broken and offset opaque platy grains and from dragged mica folia. ④ has roughly the same orientation as the megascopic fracture cleavage in associated beds. ⑤ is faintly visible as the axial planes of microcrinkles, which originated by the rumpling of ②.




3

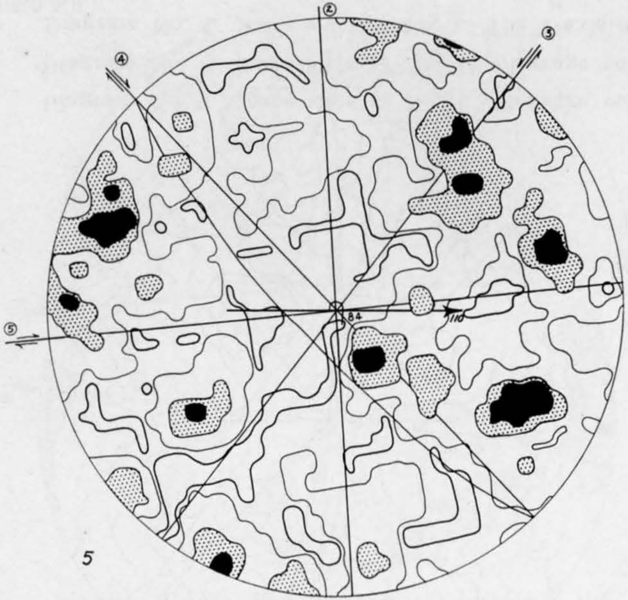



4

Diagram No. 2. Same slide as No. 1. Cleavage poles of 169 muscovite grains. 15-10-4-1-0 (max. 18.3%).

Diagram No. 3. Same slide as No. 1. Cleavage poles of 318 biotite grains. 15-10-6-3-1-0 (max. 16.3%).

Diagram No. 4. Same slide as No. 1. The ϒ-axis of 210 oligoclase grains. (ϒ is almost ⊥ (010) cleavage). (3.3-2.5)-1.5-0-5-0.



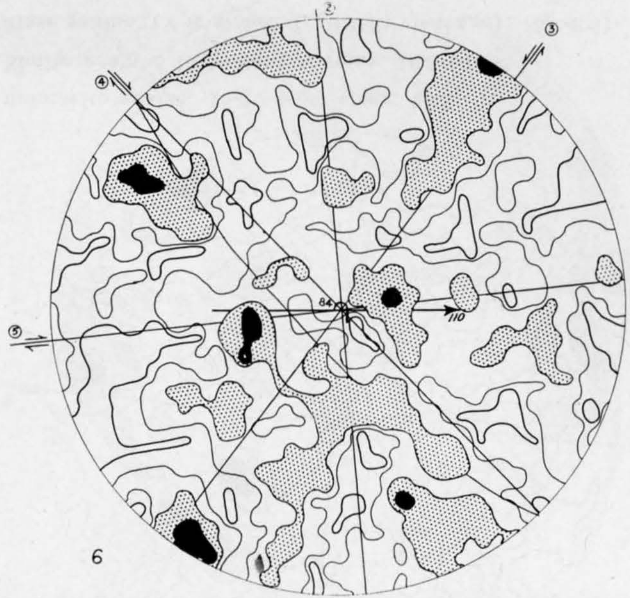
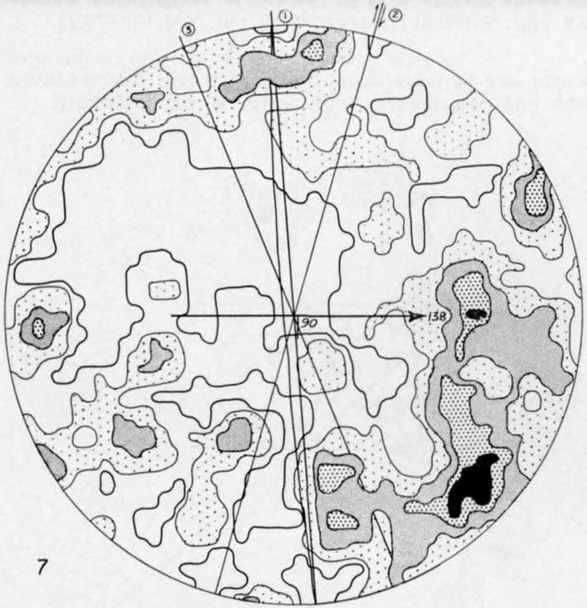




Diagram No. 5. Same slide as No. 1. The $\beta$-axis of 210 oligoclase grains. ($\beta$ makes a small angle with the pole to the (001) cleavage.) (3.3-2.5)-1.5-0.5-0.

Diagram No. 6. Same slide as No. 1. The $\alpha$-axis of 210 oligoclase grains. ($\alpha$ nearly coincides with the intersection of the (010) and (001) cleavages.) (3.3-2.5)-1.5-0.5-0.



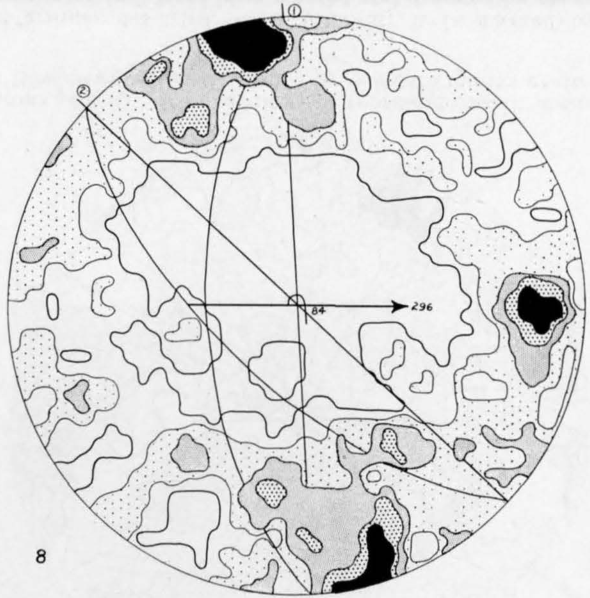
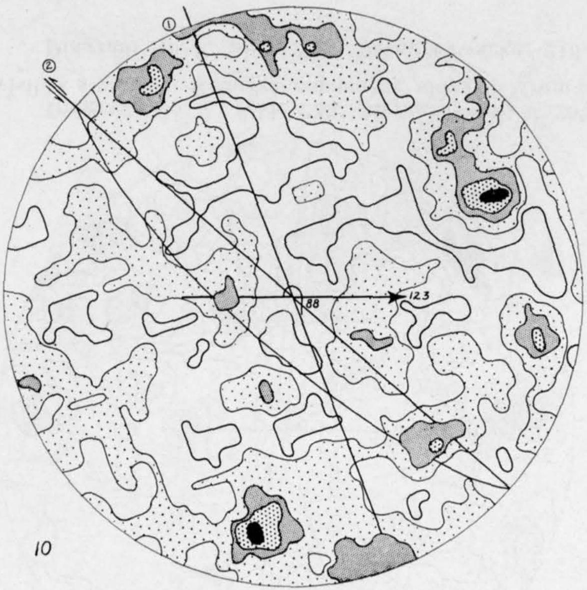




Diagram No. 7. Slide 2Z9. Metagraywacke. 207 quartz axes. 4-3-2-1-0 (max. 4.5%) ① is bedding. ② and ③ are faintly visible s-planes. Relative movement along ② from bent mica. ③ is stronger than ②.

Diagram No. 8. Slide 2T5. Metagraywacke. 213 quartz axes. 4-3-2-1-0 (max. 6.5%). ① is bedding. ② is flow cleavage.



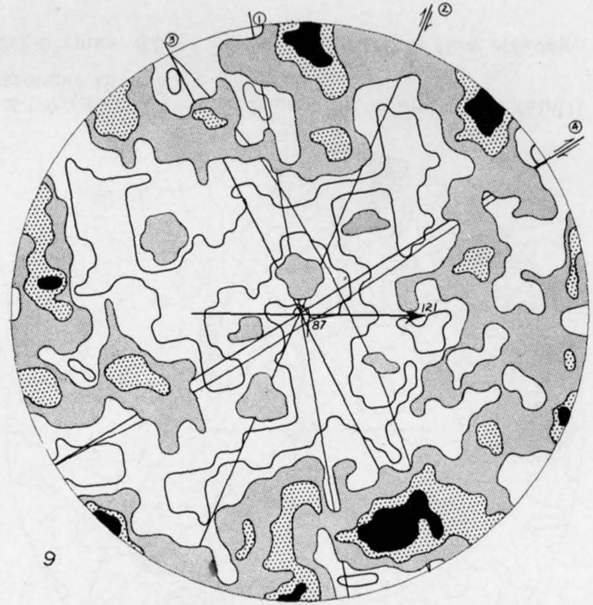
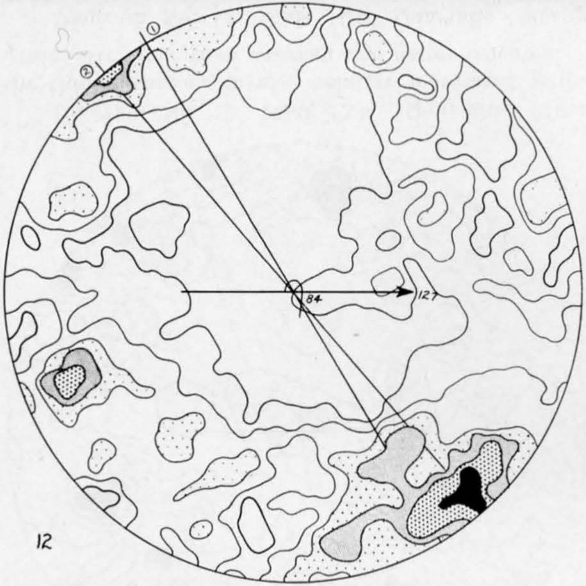




Diagram No. 9. Slide 2N8. Quartzite. 203 quartz axes. 3-2-1-0 (max. 4.0%). ① is bedding. ② represents small shear planes along which feldspar twin lamellae are broken and offset. ③ is flow cleavage, made visible by elongate quartz grains and mica folia; it is older than ② and ④.

Diagram No. 10. Slide 2N0. Quartzite. 203 axes of small quartz grains. 4-3-2-1-0. ① is bedding; it is marked by strong parallelism of micas. ② is a strong shear cleavage; the micas oriented in ① have been rotated and dragged by shear along ②.



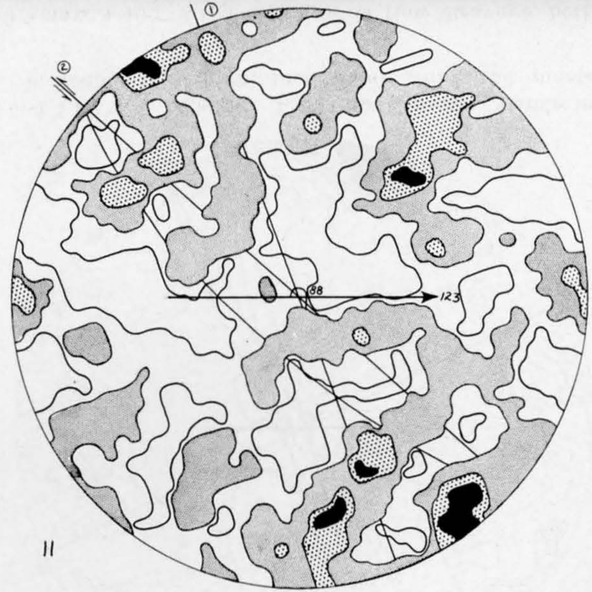
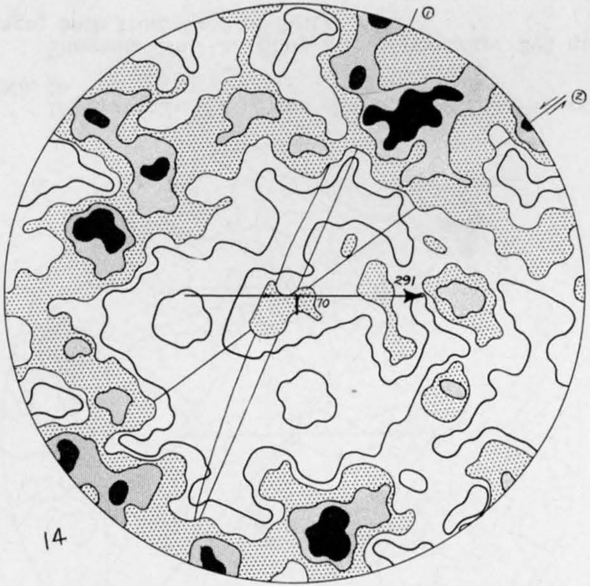
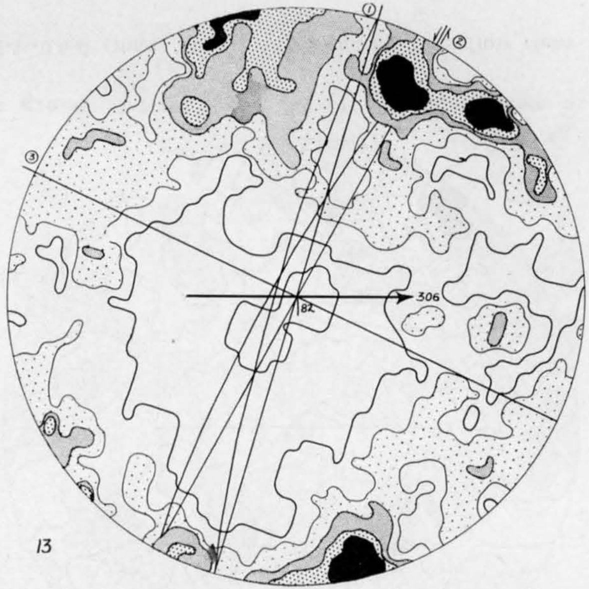




Diagram No. 11. Same slide as No. 10. 164 axes of large quartz grains. 3-2-1-0 (max. 3.5%). ① and ② same as No. 10.

Diagram No. 12. Slide 2H6. Quartzite. 202 quartz axes. 6-4-3-2-1--0.5-0 (max. 7.0%). ① is bedding, ② is flow cleavage; both from field readings.



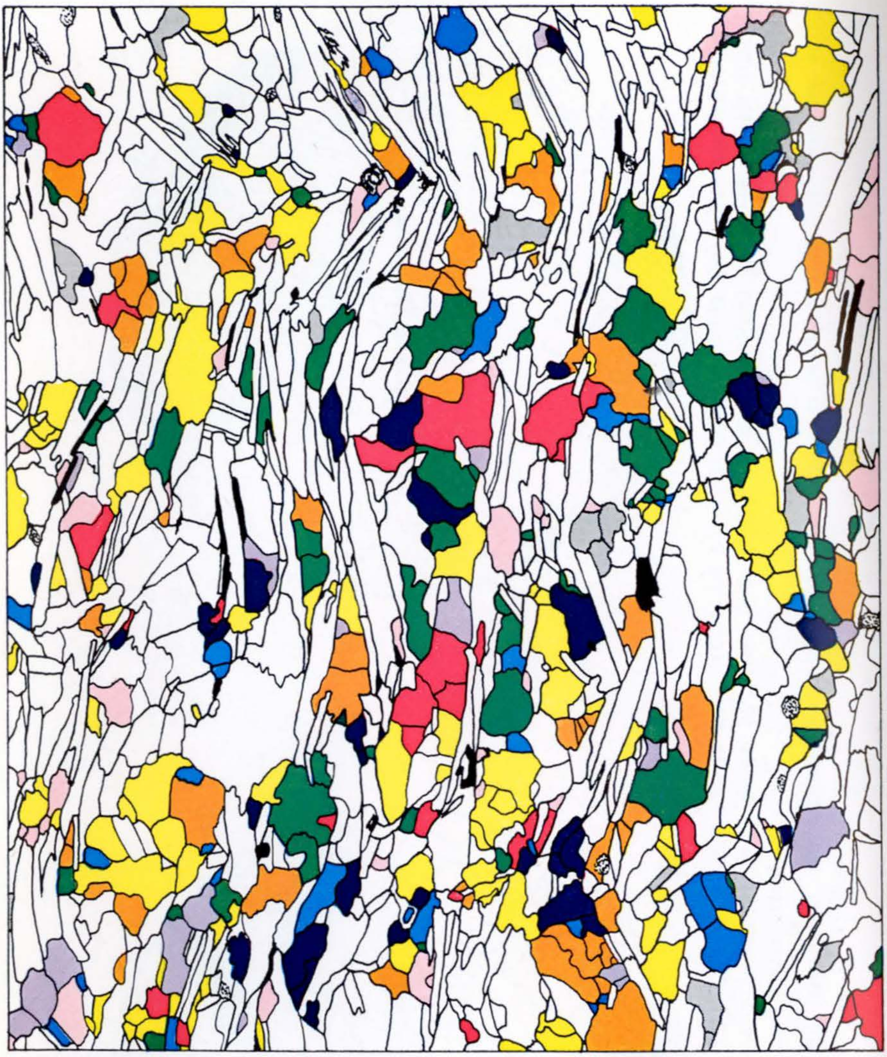

14




13

Diagram No. 13. Slide $1X6_a$. Quartzite. 212 quartz axes. (6-4)-3-2-1-0. ① is bedding. ② is flow cleavage, visible in the thin micaceous layers; relative movement well shown by drag. ③ is visible as straight lines along which the quartz grains are finer than average and better oriented.

Diagram No. 14. Slide 1Q6. Quartzite. 200 quartz axes. 3-2-1-0 (max. 4%). ① is bedding, ② is flow cleavage, both field readings.



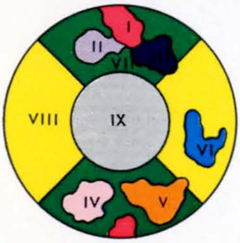



Plate 6. A.V.A. Metagraywacke, 0.6 mile east of Copperhill, Tenn.; section 1B; 447 quartz axes; magnification x 44. Uncolored grains: 318 biotite, 169 muscovite, 210 oligoclase, 9 chlorite. Apatite stippled.

## A. V. A.

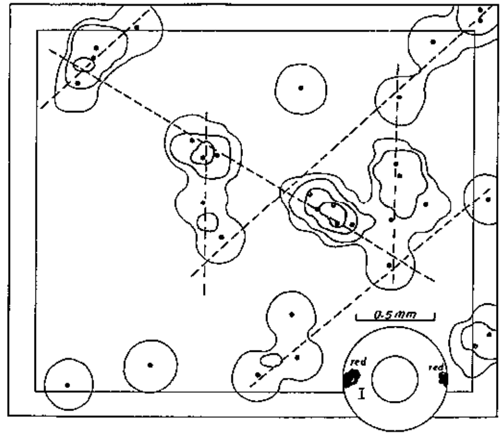
The a-c plane of one specimen, $G5_b$, was investigated by A. V. A. (Achsenverteilungsanalyse). This investigation required more than 200 hours, including the time used to draft the diagrams.

A complete A. V. A. diagram is a composite representation of all spatial data pertaining to the orientation, size, shape, and distribution of the grains in the investigated domain. The homogeneity-inhomogeneity of the fabric with respect to the orientation and shape of the grains is made especially clear. For a description of the method and a discussion of the results that may be obtained see Ramsauer (1941) and Sander (1950, pp. 39-42; 161-217).

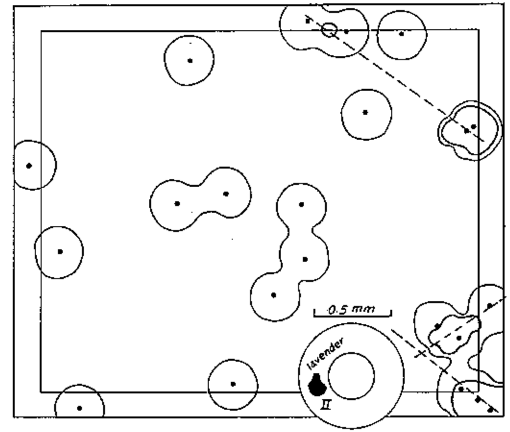
Plate 6 is a partial A. V. A. diagram, as it does not show the orientation of the feldspar grains. Their orientation is presented separately in Diagrams 4-6. The quartz grains (colored grains) are divided into nine trend-groups **(Richtungsgruppe)**. The division is based on maxima and minima in Diagram 1. The orientation of these groups in relation to the fabric is given in the small reference net at the bottom of Plate 6. The groups are distinguished by colors.

The grains of trend-groups I-V and VII-VIII range in size from large to small. The grains of VI and IX range from intermediate to small. The grain boundaries are mostly "fractured" and smooth in I, smooth to irregular and interlocking in IV, irregular and interlocking in II, III, V, and IX, irregular and interlocking, even sutured, in VII and VIII. Most of the large grains are roughly equidimensional. The smaller grains are more often inequant. The long dimension of the inequant grains is generally parallel to the distribution-banding **(Feinlagen)**. The grains of each trend-group are inhomogeneously distributed.

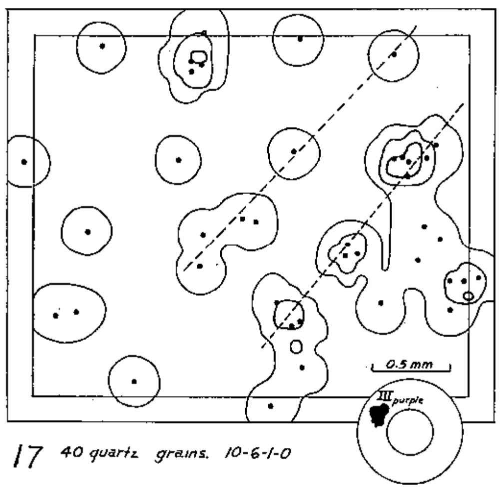
A. V. Diagrams (Achsenverteilungsdiagrammen) 15-22 show the areal distribution of the trend-groups separately. Each dot marks the position of a grain which has the orientation indicated by the dark area in the inset net. The dots have been contoured, using a counter whose area is about one percent of the total area, with a distance between counts is 0.5 cm. The contouring largely eliminates subjective impressions



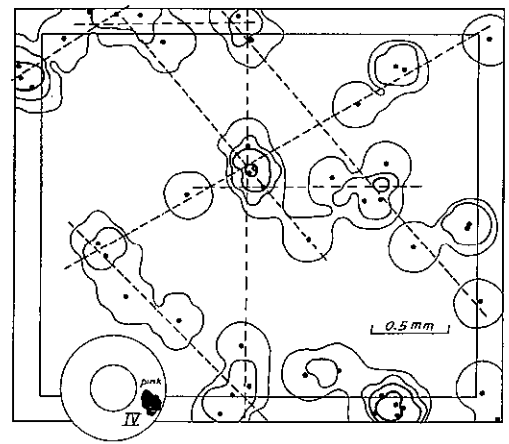

15 32 quartz grains. 11-8-4-1.5-0



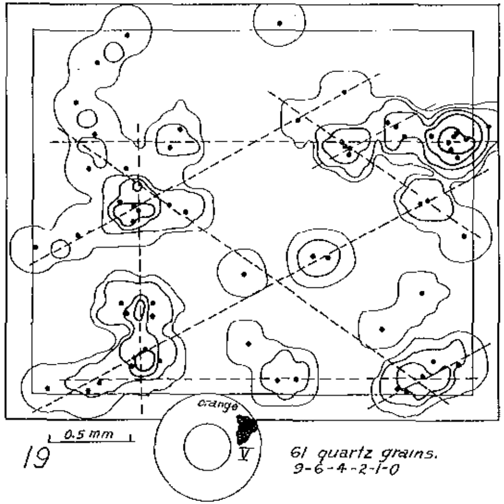

16 22 quartz grains. 7-2-0



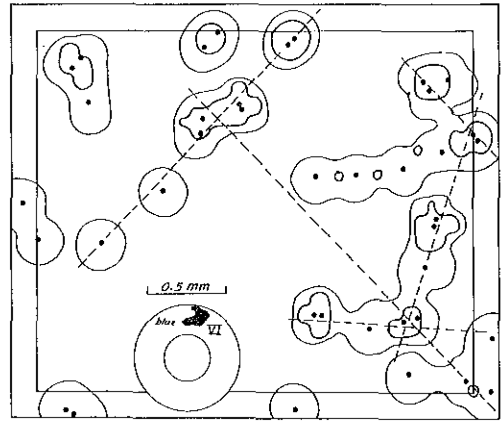

17 40 quartz grains. 10-6-1-0



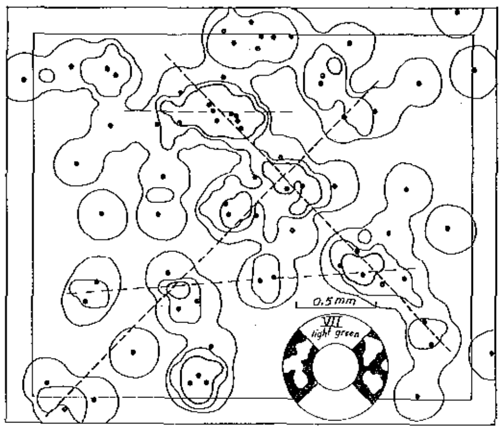

18 45 quartz grains 8-5-3-1-0



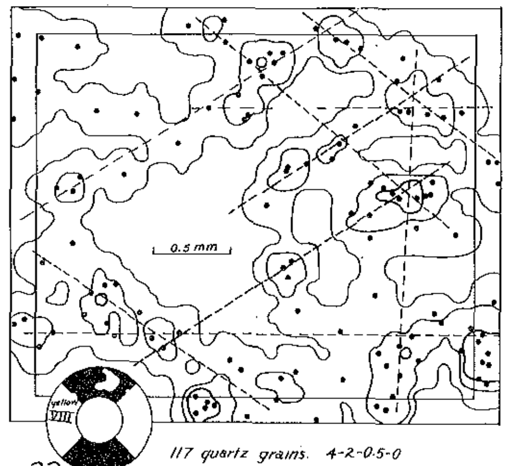

19 61 quartz grains. 9-6-4-2-1-0




20 39 quartz grains. 5-1.5-0




21 71 quartz grains. 3-2-1-0




22 117 quartz grains. 4-2-0.5-0

which might otherwise influence the evaluation of the inhomogeneities.

The **Feinlagen** (in this case, the fine banding due to the inhomogeneous distribution of the grains of one orientation) are marked by dashed lines. Ways to objectively derive the **Feinlagen** are discussed by Sander (1950, pp. 40-41).

Four sets of **Feinlagen** comprise the major fabric inhomogeneity. Because all the trend-groups reflect this inhomogeneity, its origin is not necessarily syngenetic with the present orientation of the quartz, but may be. All the trend-groups may or may not have participated actively in the origin of the **Feinlagen:** for example, a banded distribution of eight groups might geometrically require a banded distribution of the ninth. By systematically superimposing the A. V. Diagrams, the **Feinlagen** that might be a consequence of geometry can be separated from those which would only be genetically significant, if the **Feinlagen** and the quartz orientation are syngenetic. The **Feinlagen** of groups III, IV, V, VII, and VIII belong to the last category.

The superimposition of the A. V. Diagrams, for example 21 and 22, shows further that concentrations of grains belonging to one trend-group, in this example VII, are intermingled with grains belonging to a very different group, VIII. This means that the process whereby the orientation of VII originated was operative at the same place in the fabric as that whereby VIII was oriented.

A. V. A. demonstrates that no one quartz maximum is associated with any one plane or area in the rock. Instead, the quartz-orienting mechanism, for the grains of each maximum, was operative throughout the rock. The **Feinlagen** are an integral part of the quartz fabric; whether or not they are syngenetic with the orientation of the quartz is indeterminate.

## INTERPRETATION

The crinkling was subsequent to the flow cleavage, involved little movement in this specimen, and was largely a mechanical process (see the bent micas and broken grains in Plate 6). It is therefore concluded that crinkling did not produce the **Feinlagen.**

The **Feinlagen** are not easily explained as a depositional fabric. The main decipherable event which occurred between the time of deposition and the time of crinkling was the imposition of flow cleavage, or recrystallization under stress. Accordingly, the origin of the **Feinlagen** was probably coincident with the origin of the flow cleavage. The oblique **Feinlagen** are interpreted as shear planes developed during the formation of the cleavage by elongation $\perp$ **b** in the cleavage plane.

The relative movements along ③ and ④ in Diagrams **1-6** are related to the crinkling and, if the above interpretation is correct, are the reverse of the relative movements which accompanied the orientation of the quartz.

# ECONOMIC GEOLOGY

The following section is a brief description of rocks and minerals that have been mined or might be amenable to future mining in the Mineral Bluff quadrangle.

## IRON

The base of the Andrews formation contains nodules, lenses, and thin beds of limonite. Whether the limonite represents a ferruginous sediment or epigenetic iron introduced along faulted zones is not clear. At the best exposures (just east of Cox Cemetery, Weaver Creek Valley, and just west of Campbell quarry, Cutcane Creek Valley) the limonite conforms to the bedding, and appears to represent a ferruginous sediment. This view is strengthened by the regular recurrence of the mineral at one stratigraphic horizon. However, the amount of iron varies greatly along strike, and it is certain that secondary processes connected with weathering and the movement of ground water have played some part in the present condition and local distribution of the mineral.

At one place the limonite has been found in sufficient quantities to be mined. On the west side of Weaver Creek valley, about one mile southwest of where Weaver Creek empties into the Toccoa River, mining was carried on prior to the Civil War, when the ore was used in the Hemptown forge near Morganton, and during World War I, when the ore was shipped to LaFollette, Tennessee. Altogether, more than 25 car loads of ore were produced. The ore-bearing zone is about 60 feet wide at the surface. Mr. Harry Quintrell of Morganton says that in 1907 or 1908 he investigated the possible downward extension of this ore by drilling several holes east of the ore outcrops. He inclined the holes about 45° to the northwest, expecting to intercept the southeast dipping ore zone at a depth of 75-100 feet, but encountered no ore in any of the holes. This property, which is in lot 265, 8th District, was described by Haseltine (1924) who called it the Conley property.

Northeast and southwest of this property, in contiguous quadrangles, many small but minable bodies of limonite have been found in the Andrews formation.

## KYANITE

The distribution of kyanite is shown on Plate 3. Its manner of occurrence is discussed under **quartz-kyanite schist,** page 34.

The best showings are one mile southeast of Hughes Mountain on the property of Mr. A. D. Lewis. The prospect pits on Mr. Lewis' property are all small; no kyanite has been marketed.

## MARBLE

The marble is confined to two narrow belts, a western belt marked by the course of Young Stone Creek and Dry Creek, and an eastern belt marked by Cutcane Creek, Creaseman Branch, and Weaver Creek. The two marble bands presumably come together at a depth of 2-3000 feet in the trough of the Murphy syncline.

The western band is not continuous near the surface. Marble is exposed in the Toccoa River bed at the mouth of Young Stone Creek and has been encountered in drill holes at two other places: the valley just west of Mineral Bluff, and the northeast edge of Blue Ridge. Probably, marble occurs near the surface at very few other places in this belt, which is along the course of a major fault.

The eastern band appears to be continuous, though its thickness varies. Exposures are found in the bed of Weaver Creek, 500 feet southwest of Toccoa River; in the bed of Cutcane Creek, 1800 feet northeast of Hemptown Creek; and at the Campbell quarry on Cutcane Creek. At other places the marble is covered by 3-20 feet of overburden. Its position, however, is well known from old lime pits and drill holes. Recent drilling by Paul Campbell on Creaseman Branch and Cutcane Creek has provided much information about its extent, attitude, and character.

The thickness of the band ranges from 0-250 feet; its usual thickness is about 120 feet. The band dips 40-90 degrees southeast. Interbedded with the marble are thin layers of schist and slate, the number and thickness of which vary rather erratically along strike. Generally, they are most numerous at the bottom and the top, where the marble grades into the underlying and overlying formations. At the Campbell quarry, on Cutcane Creek, they are abundant at the base, or east side, of the marble through a thickness of about 20 feet, scarce in the overlying 40-60 feet, abundant in the next 5-25 feet (where a slate bed is locally 4 feet thick), and relatively scarce from there to the top of the marble. The interleaving of beds of marble and slate or schist produces changes of as much as 40 feet in the total thickness of the marble within a strike distance of a few hundred feet. Thus some of the changes in thickness are the result of original depositional conditions, not faulting, and some of the gaps in the marble outcrop in other quadrangles are probably the result of non-deposition.

For further description of the marble, see page 51.

**Campbell quarry.** About 1.6 miles northeast of Mineral Bluff in Cutcane Creek valley, on the property of Mr. Lon Dean, marble is being quarried by the Campbell Stone and Lime Company of Benton, Tennessee. The marble is crushed for use as agricultural lime, manufactured sand, terrazo chip, roofing stone, and aggregate. The capacity of the plant is about 80 tons per hour. The quarry site was prospected in 1954 with a diamond drill, and opened early in 1955. Cutcane Creek, which formerly flowed along the east side of the marble at the quarry site, has been diverted through a ditch about 50 feet to the east. A dirt levee has been built around the quarry to keep out surface water. Seep water is collected in a sump and pumped out. The crushing plant is about 400 feet southwest of the quarry. A gravel road has been completed running northwest from the plant across Dividing Ridge to the paved Mineral Bluff-Murphy highway.

## METAGRAYWACKE

Quarries for crushed stone have been opened in metagraywacke at two places, which are indicated on Plate 3. Metagraywacke makes good stone for road surfacing, but because of its hardness and massive character is not as easily quarried as other types of rock. The metagraywacke quarries are in areas where other suitable kinds of rock are not available, and all the stone produced has been used locally.

## QUARTZ

From many parts of the quadrangle, as well as adjacent areas, residual boulders have long been collected and sold to the Tennessee Copper Company for use as flux. A few large quartz veins have been mined. As the price of quartz has always been low, production has been intermittent and on a small scale. In 1951 the local collection of quartz stopped, when the Tennessee Copper Company began using quartz sand from South Carolina.

## QUARTZITE AND METACONGLOMERATE

These rocks have largely disintegrated near the surface to friable sand and pebbles. The extent of disintegration, which is caused by weathering, varies along strike as well as downward. At most places the weathered rocks can be mined to a 25 foot depth with a bulldozer and can be used without further crushing for fill and road surfacing. At some places mining can be carried on to a depth of 50 feet or more.

One large quarry has been opened in the Nottely quartzite on the west side of the Mineral Bluff-Murphy highway, 1.2 miles northeast of Mineral Bluff. The quartzite was used in County road work.

The large quartzite and metaconglomerate beds in the Dean formation have not been used, probably because they are not conspicuous, but they could be mined at many places for fine gravel and sand.

## REFRACTORY SCHIST

Some of the early furnaces used to smelt Ducktown ores were lined with a refractory schist, locally called soapstone, which was mined about one mile southeast of Hughes Mountain. One small quarry is on the east bank of Mill Creek, just north of the kyanite belt. Another old pit is about one-half mile to the northeast. The schist is vari-colored, has a greasy appearance, and can be sawed like talc. It is composed almost entirely of fine 2M muscovite. X-ray diffraction patterns of the powdered schist reveal no talc. The rock is not soapstone, but sericite schist. No further use has been made of it except in local fireplaces.

## SLATE

### Wade Allen's Quarry

Three hundred yards south of Mr. Wade Allen's home in Mineral Bluff a 30′ x 20′ opening 12′ deep has been made in the end of a slate ridge. The slate is greenish gray (5G 6/1) where fresh, dark gray (N3) to grayish black (N2) on weathered surfaces, and contains small, scattered pyrite crystals. The slaty cleavage strikes about N20E and dips 55 degrees southeast, and is cut by four sets of joints: N40W, 50NE; N8W, 68E; N8W, 85W; N75W, 85S. The quarry has been worked intermittently over a period of many years. Most of the quarried slate has been used as flagstone.

### Wes Ray's Quarry

South of Mineral Bluff 1.2 miles, on the east side of Creaseman Branch, a small flagstone quarry has been opened in the slate at the top of the Brasstown formation. The slate, which is strongly banded, breaks readily into thin sheets many feet across. For a description of the slate, see pages 50-51.

The quarry is owned by Mr. Wes Ray, but is now leased by Mr. Oscar Henry.

### County Road Quarry

Crushed rock used for many years to surface the roads in Fannin County has come from a quarry in the Nantahala slate 600 yards north of Coles Crossing, east side of the Mineral Bluff quadrangle. The quarry is owned by Mr. Arthur Ross, Sr. The quarry foreman is Mr. Roy Harper. About 1200 cubic yards of crushed rock are being produced per month and sold to the County.

## STAUROLITE

In the manufacture of cement, the Lehigh Portland Cement Company, Bunnell, Florida, uses staurolite as a source of alumina and iron (Nordberg, 1953). The chemical formula of staurolite is $Fe(OH)_2.2Al_2SiO_5$. Thus about 49 per cent of the pure mineral is alumina ($Al_2O_3$) and about 17 per cent is iron oxide (FeO). The Lehigh Company uses about 30 pounds of staurolite per barrel of cement. They obtain the staurolite from the Humphreys Gold Corp., Starke, Florida, where it is a by-product of an ilmenite plant.

A large amount of staurolite is available in Fannin County. The distribution of staurolite-bearing zones is shown on Plate 3. Portions of the zones contain as much as 10 per cent staurolite by weight. All the raw materials needed for the production of Portland cement—high-lime marble ($CaCO_3$), quartzite ($SiO_2$), staurolite ($Al_2O_3+Fe_2O_3$), and high-iron slag (Tenn. Copper Co.)—are available within a radius of 10 miles.

Aside from possible future use of staurolite in the manufacture of cement, present exploitation is possible on a small scale. The staurolite crystals are well-formed and exhibit a variety of interesting twins. They compare favorably with crystals from other famous staurolite localities. Small quantities of the crystals have been picked up and sold to mineral collectors for many years, but only a small part of what could be sold through proper use of advertising.

## TALC

The known talc prospects in this quadrangle are closely related to faulted zones where post-crystallization alteration processes have been operative in the Murphy marble. Future prospecting for talc should be guided by surface, geological work which can locate most of the faulted zones and determine whether they are attended by the characteristic mineralogical alteration.

### Nichols Prospect*

In 1905, Mr. W. T. S. Dickey sank a shaft close to the Louisville and Nashville Railroad and about 300 yards southwest of the depot in Mineral Bluff. Reportedly, the shaft encountered good talc, but the operation was stopped by a section foreman who alleged that Dickey was encroaching on Company property. In April, 1905 Mr. Dickey requested permission of the Railroad Company to resume mining, but no further work was done. Because the shaft is located on the west side of the railroad, further work would have carried the shaft beneath the tracks.

At about the same time talc was being mined a short distance to the southwest. The old workings begin where the wagon road crosses the railroad halfway between Mineral Bluff and the Toccoa River and extend about 600 feet to the southwest. They consist of inclined shafts and open cuts, which are now largely filled. Their depth is not known. The talc was mined from thin beds up to four feet thick along the eastern edge of the Nottely quartzite.

All the specimens of talc now exposed around the workings are gray or bluish-gray and schistose but hard. Some of it contains quartz grains, and is gradational to talcose quartz schist. Some of it contains pyrite.

On strike with these workings, on a small knoll just south of the Toccoa River, showings of talc have been noted (Hopkins, 1914, p. 238) but not prospected.

### Hall and Howard Prospect

On the bluff north of Hemptown Creek, just west of the Mineral Bluff-Murphy highway, Hall and Howard of Chattanooga, Tennessee, dug a prospect trench about 130 feet long and 10-15 feet wide along the west side of the Nottely quartzite. According to report, Hall and Howard moved some talc from this trench to their washer at Mineral Bluff. At present no talc can be seen in place.

### Hyde-Steins-Ray Prospect

The workings are 1.5 miles northeast of Mineral Bluff and about 300 yards west of the Mineral Bluff-Murphy highway. They are along the west side of the Nottely quartzite, which forms a low ridge. The old mines and prospects are on property now owned by Lester Hyde, Willard Steins, and Henry Ray. The property line between Hyde and Steins follows the center of the ridge southward to a small transecting branch; Steins owns the property west of the line. Ray's property is south of the branch.

Talc was mined from this locality about 1909 and was shipped to D. M. Stewart of Chattanooga, Tennessee, who manufactured gas tips, crayons, powder, etc. from it. The workings follow the faulted zone along the west side of the Nottely quartzite for a distance of about 800 feet. The thickness of the talc ranges from 0-10 feet, with an average thickness of about four feet.

The first prospect on the northeast end of the old workings is a pit about 25 feet long, 10 feet wide, and 8 feet deep. Its original depth is unknown. The dump contains fragments of talcose quartzite, yellow schistose talc, and white talc.

About 530 feet southwest of this pit there is an inclined shaft, now filled to within 15 feet of the surface. Southeast to northwest, the following section is exposed: 30 feet of massive bedded quartzite dipping 50 degrees southeast; 6 inches of thin-bedded quartzite; 9 inches of thin-bedded talcose quartzite; 6 inches of coarse-grained quartzite; 34 inches of thin-bedded quartzite with a few thin talc layers; 8 feet of slaty brown iron-stained talc (some layers are light gray, others nearly white); the footwall is not exposed.

About 30 feet farther southwest there is a trench 150 feet long and parallel to the quartzite, which exposes the hanging wall. Sixty feet southwest of the terminus of this trench there is a second trench which is 130 feet long; it terminates near the transecting branch which is the northeast boundary of Mr. Ray's property. The trenches are 4-15 feet deep. The talc zone that was removed was about five feet thick.

A shaft at the southwest end of the second trench was begun about 1909 and carried to a depth of 20 feet. In 1932 Harry Quintrell and others carried it to a depth of 112 feet for the Cohutta Talc Company. At the 65 foot level there is a 40 foot drift to the northeast. According to Mr. Quintrell, the talc is about 6 feet thick in the shaft; the talc is gray and contains pyrite to the 80-foot level, dark and slaty from the 80-95 foot level, and tough, massive, and possibly suitable for the manufacture of pencils from the 95-foot level to the bottom of the shaft.

South of the shaft, on Mr. Ray's property, there is a large open cut in the quartzite. This cut, which is 60 feet long, 15 feet deep, and 8 feet wide, was made by Stewart. According to report, talc was removed from this opening locally to a depth of 80 feet, and the workings were back-filled.

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*The following descriptions abstracted from "Talc Deposits of the Murphy Marble Belt, Georgia" by A. S. Furcron and Klefton H. Teague, unpublished manuscript in the Georgia Geological Survey files.

On the old W. A. Hyde place talc has been prospected near the spring about 250 feet south of the old Hyde home. These old prospect pits are 600 feet northwest of the Hyde-Steins-Ray prospects. Fragments of white talc may be dug from the dump, but not much prospecting was done here.

### Ross Property

This property is on the west side of the Mineral Bluff-Murphy highway, and about two miles northeast of Mineral Bluff. It adjoins the Lester Hyde property on the southwest. Talc occurs as float 100 feet west of the Ross home on a cultivated hillside, but has not been prospected.

### Cutcane Creek

Several talc deposits have been discovered and prospected along Cutcane Creek northeast of the Mineral Bluff quadrangle. Although none were encountered along this belt within the quadrangle, the area still offers opportunity for prospecting.

# BIBLIOGRAPHY

BAYLEY, W. S. (1928) Geology of the Tate quad., Ga.: Ga. Geol. Surv., Bull. no. 43.

BLOOMER, R. O. (1950) Late pre-Cambrian or Lower Cambrian formations in Central Virginia: Am. Jour. Sci., vol. 248, pp. 753-783.

BUTTS, CHARLES (1926) Geology of Alabama, The Paleozoic rocks of Alabama: Ala. Geol. Surv., Special report 14, pp. 49-61.

——————, 1940) Geology of the Appalachian Valley in Virginia: Va. Geol. Surv., Bull. 52, part 1.

CHAYES, FELIX (1949) A simple point counter for thin section analysis: Am. Min., vol. 34, pp. 1-11.

——————, 1952) Notes on the staining of potash feldspar with sodium cobaltinitrite in thin section: Am. Min., vol. 37, pp. 337-340.

CLARKE, F. W. (1924) Data of geochemistry: U. S. Geol. Surv., Bull. 770.

CLOOS, ERNST (1937) The application of recent structural methods in the interpretation of the crystalline rocks of Maryland: Md. Geol. Surv., vol. 13, pp. 1-100.

——————, (1951) Washington Co.: State of Md., Dept. of Geol., Mines and Water Resources. pp. 1-333.

CRICKMAY, GEOFFREY W. (1936) Status of the Talladege series in Southern Appalachian stratigraphy: Geol. Soc. Am. Bull., vol. 47, pp. 1371-1392.

EDWARDS, A. B. (1942) Differentiation of the dolerites of Tasmania: Jour. Geol., vol. 50, p. 465, no. 1.

EMMONS, R. C. (1942) The universal stage: Geol. Soc. Am., Memoir 8.

——————, and LANEY, F. B. (1926) Geology and ore deposits of the Ducktown Mining District, Tenn.: U. S. Geol. Survey Professional Paper no. 139.

ESKOLA, PENTTI (1932) Conditions during the earliest geological times as indicated by the Archean Rocks: Suomalaisen Tiedeakatemian Toimituksia, Sarja A. Nid. 36, no. 4, pp. 5-74.

FAIRBAIRN, H. W. (1949) Structural petrology of deformed rocks: Addison-Wesley Press, Inc.

—————— and other (1951) A cooperative investigation of precision and accuracy in chemical, spectrochemical, and modal analysis of silicate rocks: U. S. Geol. Surv., Bull. 980, Plate 1.

FURCRON, A. S. (1953) Comments on the geology of the Ellijay quad., Ga.-N. C.-Tenn.: Ga. Geol. Surv., Bull. 60, pp. 32-40.

HALL, A L. (1932) The Bushveld igneous complex of the central Transvaal, South Africa: Geol. Surv. Memoir 28, p. 310, No. 1.

HASELTINE, R. H. (1924) Iron ore deposits of Georgia, Bull. No. 41, Ga. Geol. Survey.

HOLMES, A. and HARWOOD, H. F. (1929) The tholeiitic dikes of the North of England: Mineral Mag., vol. 22, p. 16.

HOPKINS, O. B. (1914) Asbestos, Talc, & Soapstone deposits of Georgia, Bull. No. 29, Ga. Geol. Survey.

HURST, VERNON J. (1953) Chertification in the Ft. Payne formation, Georgia: Ga. Geol. Surv., Bull. 60, pp. 215-238.

JONAS, A. I. (1932) Structure of the metamorphic rocks of the Southern Appalachian: Am. Jour. Sci., 5th Ser., vol. 24, pp. 228-243.

KEITH, A. (1895) U. S. Geol. Surv. Atlas, Knoxville Folio: No. 16.

————, (1907) U. S. Geol. Surv. Geol. Atlas, Nantahala Folio: No. 143.

————, (1913) Production of apparent diorite by metamorphism: Geol. Soc. Am., Bull., vol. 24, pp. 684-85.

KING, P. B. (1949) The base of the Cambrian in the Southern Appalachians: Am. Jour. Sci., vol. 247, pp. 513-530, 622-645.

LA FORGE, LAURENCE and PHALEN, W. C. (1913) U. S. Geol. Surv. Geol. Atlas, Ellijay Folio: No. 187.

————, COOKE, W., KEITH, A., and CAMPBELL, M. R., Physical Geography of Georgia: Geol. Surv. of Ga., Bull. 42, pp. 93-114, 1925.

NORDBERG, BROR (1953) Lehigh manufactures cement from coquina and staurolite residue, Rock Products, Aug., pp. 130-149, 202-204.

NUTTALL, BRANDON D. (1951) The Nantahala-Ocoee contact in North Georgia: Unpublished Master's Thesis, Univ. of Cincinnati.

PETTIJOHN, F. J. (1940) Archean metaconcretions of Thunder Lake, Ontario: Geol. Soc. Am., Bull., vol. 51, pp. 1841-1850.

————, (1949) Sedimentary Rocks.

RAMSAUER, H. (1941) Achsenverteilungsanalysen und Quartztektoniten: Dissertation, Innsbruck.

RESSER, C. E. (1933) Preliminary generalized Cambrian time scale: Geol. Soc. Am. Bull., vol. 44, pp. 735-756.

ROSS, CLARENCE (1935) Origin of the copper deposits of the Ducktown type in the southern Appalachian region: Geol. Surv. Professional Paper 179.

SANDER, BRUNO (1948) Einführung in die Gefügekunde der geologischen Körper, erster Teil, 409 pp.

————, (1950) Einführung in die Gefügekunde der geologischen Körper, zweiter Teil, 409 pp.

SEDERHOLM, J. J. (1928) On orbicular granites: Commission Geol. Finlande, Bull. 83, pp. 82-83.

SHROCK, ROBERT R. (1948) Sequence in layered rocks: McGraw-Hill Publishing Co., 507 pp.

SMITH, J. V. and YODER. H. S. (1953) Theoretical and X-ray studies of the mica polymorphs (abstract): Geol. Soc. Am., Bull., vol. 64, p. 1475.

STOSE, G. W. and STOSE, A. J. (1944) The Chilhowee group and Ocoee series of the southern Appalachians: Am. Jour. Sci., vol. 242, pp. 367-390, 401-416.

————, (1949) Ocoee series of the southern Appalachians: Geol. Soc. Am. Bull., vol. 60, pp. 267-320.

SUNDIUS, N. (1930) Geol. Mag., vol. 67, p. 9.

VAN HORN, E. C. (1948) Talc deposits of the Murphy marble belt: N. C. Div. Min. Resources, Bull. 56, 54 pp.

WINCHELL, A. N. (1951) Elements of Optical Mineralogy, Part II, 4th ed.

YODER, H. S. (1952) The $MgO—Al_2O_3—SiO_2—H_2O$ system and the related metamorphic facies: Am. Jour. Sci., Bowen vol., pp. 569-627.

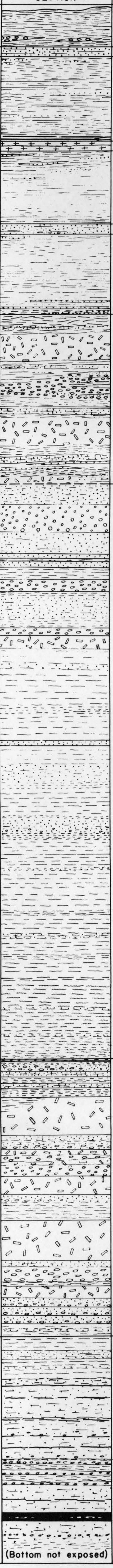
# INDEX

# INDEX – Continued

# INDEX – Continued

# INDEX – Continued

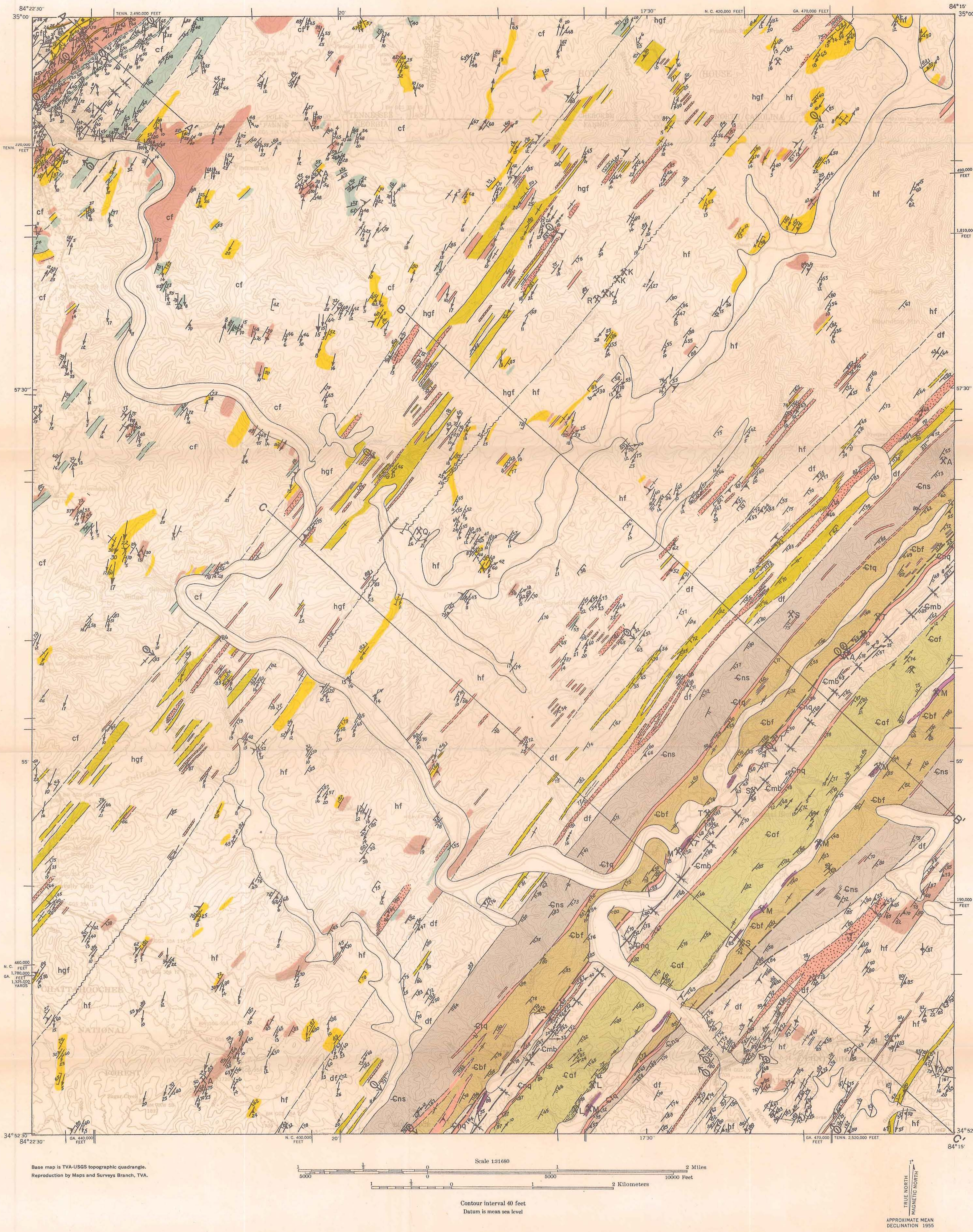
# GENERALIZED SECTION OF THE META-SEDIMENTARY ROCKS OF THE MINERAL BLUFF QUADRANGLE

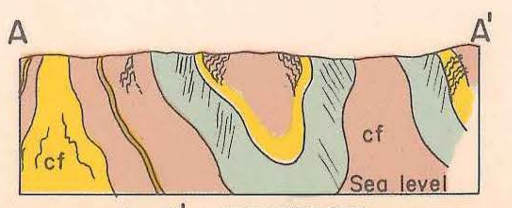
| SYSTEM | | FORMATION NAME | SYMBOL | COLUMNAR SECTION | THICKNESS (FEET) | CHARACTER OF ROCKS |
|---|---|---|---|---|---|---|
| CAMBRIAN | | Mineral Bluff formation | Ꞓmb |  | 800+ | Sericite schists containing small, scattered garnets; gray to gray-green slates; some talcose schist and graphitic schist. Quartz conglomerate and meta-arkose lenses 100'-200' above the Nottely quartzite. |
| | | Nottely quartzite | Ꞓnq | | 75-150 | White orthoquartzite, commonly feldspathic. Contains dark slate laminae and thin quartz-sericite schists. |
| | | Andrews formation | Ꞓaf | | 1400-1800 | Fine sericite schists, sparingly garnetiferous; gray slates or phyllites. Calcareous schists and thin limonite beds at the base. |
| | | Murphy marble | Ꞓmm | | 75-250 | Thick-bedded, fine- to medium-grained, white, blue, and blue and white banded marble. |
| | | Brasstown formation | Ꞓbt | | 1200-1500 | Dark slates and schists containing sandy layers and lensing micaceous quartzites. Calcareous biotite schists at the top. Usually black slate or graphitic schist at the base. |
| | | Tusquitee quartzite | Ꞓtq | | 20-200 | Thin-bedded, light-colored, feldspathic quartzite in which there are discontinuous, paper-thin laminae and thin sandy beds of dark slate. |
| | | Nantahala slate | Ꞓns | | 1000-1800 | Black to gray banded slates. Occasional dark schists and thin, micaceous quartzites, particularly near the bottom. |
| ? | | | | | | |
| PRE-CAMBRIAN | Great Smoky Group | Dean formation | df | | 2500-3500 | Staurolite-mica schists, X-biotite schists, quartzites, meta-conglomerates, and "pseudo-diorite" beds. Some dark schist or slate, meta-arkose, and meta-graywacke. |
| | | Hothouse formation | hf | | 8000-11 000 | Interbedded meta-graywacke, quartzite, and meta-conglomerate at the top. Downward these beds become less and less prominent. The base is principally mica schist. |
| | | Hughes Gap formation | hgf | | 4000-6000 | At the top are thin-bedded sericite schists, usually garnetiferous, and a 1-3' bed of quartz-kyanite schist. The remainder of this form. is lithologically the same as the Dean formation. |
| | | Copperhill formation | cf | (Bottom not exposed) | 2000-5000 | Massive beds of meta-graywacke or biotite gneiss, mica schist, meta-conglomerate, micaceous quartzite, and meta-arkose. Occasional dark slates.<br>Epidote-amphibolite sill (shown in black). Masses of mica schist are in the sill, locally. |

# GEOLOGIC MAP
of the
# MINERAL BLUFF QUADRANGLE

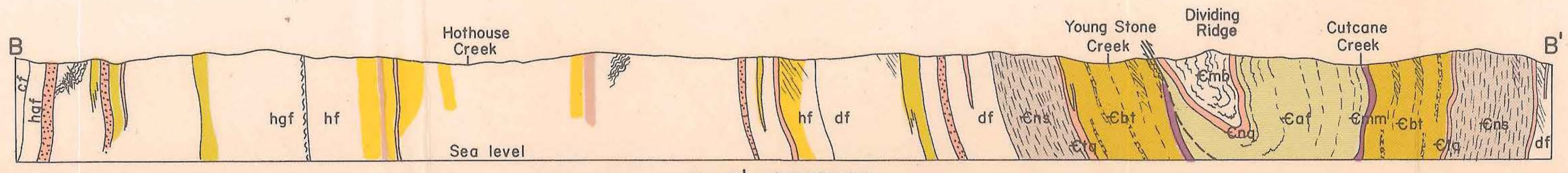
by
Vernon J. Hurst

1954




A-A' SECTION

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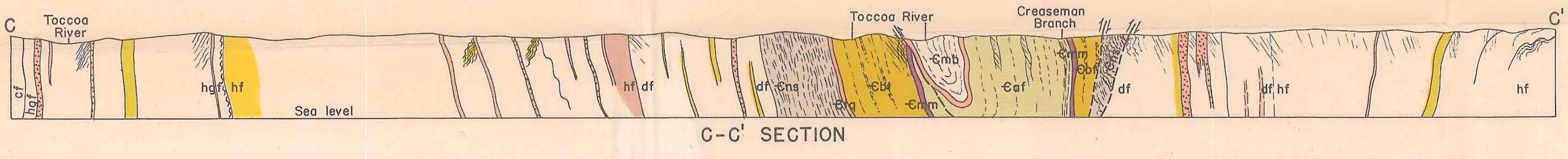


C-C' SECTION

## EXPLANATION

### META-SEDIMENTARY ROCKS

CAMBRIAN

Ꞓmb

**Mineral Bluff formation**
Sericite schists containing small, scattered garnets; gray to gray-green slates or phyllites; some talcose schist and graphitic schist. Conglomerate lenses 100'-200' above the base.

Ꞓnq

**Nottely quartzite**
White orthoquartzite, commonly feldspathic; contains dark slate laminae and thin-bedded quartz-sericite schists.

Ꞓaf

**Andrews formation**
Sericite schists, sparingly garnetiferous; gray slates or phyllites; calcareous schists and thin limonite beds at the base.

Ꞓmm

**Murphy marble**
Fine- to medium-grained, white, blue, and blue and white banded marble.

Ꞓbf

**Brasstown formation**
Dark slates and schists containing sandy layers and thin, lensing, micaceous quartzites. At the top is calcareous biotite schist. At the base is usually black slate, or black graphitic schist.

Ꞓtq

**Tusquitee quartzite**
Thin-bedded, light-colored, feldspathic quartzite in which there are discontinuous, paper-thin laminae and thin, sandy beds of dark slate.

Ꞓns

**Nantahala slate**
Black to gray banded slate. Occasional dark schists and thin, micaceous quartzites, particularly near the base.

PRE-CAMBRIAN ?

Great Smoky Group

df

**Dean formation**
Staurolite-mica schist, X-biotite schist, quartzite, meta-quartz conglomerate, and "pseudo-diorite" beds. Some dark schist or slate and meta-graywacke.

hf

**Hothouse formation**
Dominantly meta-graywacke, meta-conglomerate, and quartzite at the top. Downward these beds become less and less prominent. The base is principally mica schist.

hgf

**Hughes Gap formation**
At the top are thin-bedded sericite schists, usually garnetiferous, and a 1'-3' bed of quartz-kyanite schist. The remainder of the formation is similar to the Dean formation.

cf

**Copperhill formation**
Meta-graywacke or biotite gneiss, mica schist, meta-conglomerate, micaceous quartzite, and meta-arkose. Occasional dark slates.

Symbols used independently of formational boundaries

Dominantly meta-graywacke

Dominantly mica schist

Conglomeratic meta-graywacke

Staurolite-mica schist

Quartzite

Meta-quartz conglomerate

Quartz-kyanite schist

Black slate or schist

### META-IGNEOUS ROCKS

Epidote amphibolite

### STRUCTURE SYMBOLS

- Lithologic contact, dashed where approximate or gradational
- Strike and dip of bedding
- Bedding vertical
- Bedding horizontal
- Strike and dip of flow cleavage
- Flow cleavage vertical
- Crinkled flow cleavage which dips less than the bedding
- Crinkled flow cleavage which dips more steeply than the bedding
- Strike and plunge of crinkle axes
- Crinkle axes vertical
- Crinkle axes horizontal
- Axis of major anticline
- Axis of major syncline
- Strike and plunge of drag folds
- Strike and dip of fracture cleavage
- Fracture cleavage vertical
- Strike and dip of S-plane, either bedding or another cleavage
- S-plane vertical
- S-plane horizontal
- Strike and dip of foliation which is parallel to, or makes a low angle with, the bedding; slaty cleavage
- Foliation vertical
- Strike and plunge of elongate pebbles
- Strike and dip of fracture cleavage; strike and plunge of crinkle axes
- ① Graded bedding; coarse to fine in direction of arrow
- ② X-bedding; top of beds in direction of arrow
- ③ "Reversed graded bedding"; top of beds in direction of arrow
- Fault, dashed where approximately located

### ⚒ QUARRIES, MINES, & PROSPECTS

| | |
|---|---|
| K kyanite | M marble |
| Q quartz | T talc |
| L limonite | S slate |
| R refractory | A aggregate |