# Geology of the Grandfather Mountain Window and Vicinity, North Carolina and Tennessee

*By* BRUCE BRYANT *and* JOHN C. REED, JR.

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*Structural and metamorphic history of rocks exposed in the largest window in the crystalline belt of the southern Appalachians, and a discussion of the significance of the window in interpretation of the tectonics of the region*

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# GEOLOGY OF THE GRANDFATHER MOUNTAIN WINDOW AND VICINITY, NORTH CAROLINA AND TENNESSEE

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## ABSTRACT

The Blue Ridge belt in northwestern North Carolina and northeastern Tennessee is composed chiefly of 1,000-million to 1,100-million-year-old metamorphic and plutonic rocks that have been thrust many miles northwestward across unmetamorphosed Cambrian(?) and Cambrian sedimentary rocks of the Unaka belt. The Blue Ridge thrust sheet is rooted on the southeast along the Brevard zone, a zone of strike-slip faulting along which metamorphic and plutonic rocks of the Inner Piedmont belt are juxtaposed with rocks of the Blue Ridge.

Near the southeastern edge of the Blue Ridge belt, the Blue Ridge thrust sheet is breached by erosion, and the rocks beneath are exposed in the Grandfather Mountain window, which is 45 miles long and as much as 20 miles wide; it is the only major window so far recognized in the Blue Ridge belt. The rocks exposed within it include 1,000-million-year to 1,100-million-year-old plutonic basement rocks, sedimentary and volcanic rocks of late Precambrian age, and an allochthonous tectonic slice of Lower Cambrian(?) and Cambrian sedimentary rocks identified with the Chilhowee Group and Shady Dolomite in the Unaka belt 20 to 30 miles to the northwest.

The Blue Ridge thrust sheet surrounding the Grandfather Mountain window consists largely of schist, gneiss, and amphibolite derived by metamorphism of sedimentary and volcanic rocks 1,000 to 1,100 m.y. ago, and of Cranberry Gneiss, a complex of migmatite and granitic rocks which underlies the metasedimentary and metavolcanic rocks and which probably formed during the same metamorphic episode. The Cranberry Gneiss is intruded by the Beech Granite, by aegirine-augite granite, and by quartz monzonite, all of which were emplaced during a late stage of or after the plutonic metamorphism. Stocks and dikes of Bakersville Gabbro of late Precambrian(?) age and small bodies of ultramafic rock, granodiorite, and pegmatite of early or middle Paleozoic age intrude the earlier Precambrian rocks. Although all these rocks may have been metamorphosed about 450 m.y. ago, the principal Paleozoic dynamothermal metamorphism occurred about 350 m.y. ago. At that time new medium-grade minerals, including staurolite, kyanite, monoclinic pyroxene, epidote, and calcic plagioclase, crystallized in the schist, gneiss, and amphibolite. During the late Paleozoic, most of the plutonic rocks were partly reconstituted to low-grade blastomylonitic and phyllonitic gneisses containing new biotite, albite, sericite, chlorite, actinolite, and epidote, whereas the overlying rocks were largely unaffected. The contact between low- and medium-grade rocks may be a fault.

Layering and foliation in rocks of the Blue Ridge thrust sheet dip away from the Grandfather Mountain window on all sides, and broad flexures in these structures plunge away from its northwest and northeast corners. Minor folds in both the low- and medium-grade rocks are of two generations: (1) tight and isoclinal folds having axial planes parallel to foliation and layering and axes trending in various directions in the plane of the foliation, and (2) later open folds and crinkles having steep axial planes and northeast-trending axes perpendicular to a well-developed northwest-trending mineral lineation. The early folds, which are possibly 350 m.y. old, perhaps formed during an early stage of thrusting and were themselves deformed during continued thrust movement. The later folds formed in a late stage of the thrusting.

In the northwest corner of the area, an intermediate sheet of partly metamorphosed Precambrian plutonic rocks occurs between the Blue Ridge thrust sheet and rocks of the Unaka belt in the Mountain City window. The Blue Ridge thrust sheet overrides both the intermediate sheet and the Mountain City window, in which rocks of the Chilhowee Group of Cambrian(?) and Early Cambrian age and the Shady Dolomite and Rome Formation of Early Cambrian age are exposed.

The basement exposed in the Grandfather Mountain window is composed principally of nonlayered granitic gneiss (Wilson Creek Gneiss) and coarse-grained augen gneiss (Blowing Rock Gneiss), both 1,000 to 1,100 m.y. old. The plutonic basement rocks are stratigraphically overlain by the Grandfather Mountain Formation, a sequence at least 20,000 feet thick of arkose, siltstone, shale, and conglomerate of late Precambrian age. The formation also contains tuffaceous rocks; flows of basalt, quartz latite, and rhyolite; and sills of diabase. The sediments were derived mainly from adjacent plutonic rocks but partly from volcanic rocks similar to those found in the formation and were apparently deposited in a rapidly subsiding basin. Diabase and felsic porphyry intrusives in the basement rocks are probably related to the volcanic rocks in the Grandfather Mountain Formation.

The main outcrop belt of the Grandfather Mountain Formation lies on the southeast limb of a large synclinorium overturned to the northwest. Medium- and small-scale folds are overturned to the northwest or west; they are isoclinal in the southeastern part of the belt and are more open in the northwestern part. In most of the outcrop belt, axes of minor folds are subhorizontal, and their axial planes strike northeast parallel to the trend of lithologic units. In the northern part of the outcrop belt, however, axes of minor folds plunge

northeast and axial planes strike north or northwest at a large angle to the trends of the lithologic units. These folds are evidently younger than, and superimposed on, the earlier major structure.

Pervasive cleavage, parallel to the axial planes of the minor folds in the upper Precambrian rocks, is parallel to cataclastic foliation in the underlying basement rocks. Strongly developed cataclastic lineation plunges southeastward on the cleavage and foliation planes. It is generally normal to the axes of the minor folds and is evidently in the *a* direction.

The rocks in the Grandfather Mountain window are of the same low metamorphic grade as the retrogressively metamorphosed rocks in adjacent parts of the Blue Ridge thrust sheet. Progressive metamorphism of the upper Precambrian rocks was concurrent with retrogressive metamorphism of the basement rocks. Typical metamorphic minerals in both groups of rocks are albite, microcline, epidote, actinolite, chlorite, and iron-rich muscovite. Basement rocks were converted to blastomylonitic and phyllonitic gneiss, phyllonite, and blastomylonite.

Metamorphism of the rocks in the Grandfather Mountain window apparently reached a thermal climax about 350 m.y. ago. The large synclinorium in the upper Precambrian rocks formed prior to that time, but most of the minor folds and the cleavage, cataclastic foliation, and lineation formed during the metamorphic episode 350 m.y. ago.

The Tablerock thrust sheet is a tectonic slice between the Blue Ridge thrust sheet and autochthonous rocks in the southwestern part of the Grandfather Mountain window. It is composed of Shady Dolomite of Early Cambrian age and several thousand feet of quartzite, arkosic quartzite, and phyllite of the underlying Chilhowee Group of Early Cambrian(?) and Early Cambrian age. These rocks are metamorphosed to the same grade as the rest of the window rocks, but their bedding and cleavage are approximately parallel with the cataclastic foliation in the Blue Ridge thrust sheet and are strongly discordant with structures in the underlying autochthonous rocks. Tight and isoclinal folds in rocks of the Tablerock thrust sheet have axial planes parallel with bedding and cleavage, and diversely oriented axes; superimposed on them are open folds having southeast-dipping axial planes and gently southwest-plunging axes which are approximately perpendicular to a well-developed mineral lineation. The geometry of these structures resembles that of the structures in the overlying Blue Ridge thrust sheet; both groups of structures probably formed and were rotated into their present orientation during thrusting. At the southwest end of the window, the Tablerock thrust sheet is overturned, broken by faults, and overridden by sheets of basement rock.

Layered gneiss, mica and sillimanite schist, amphibolite, and other associated metasedimentary and metavolcanic rocks in the Inner Piedmont belt are probably of late Precambrian or early Paleozoic age; they were metamorphosed in the early or middle Paleozoic. Where they have not been affected by later metamorphism related to the Brevard fault zone, they contain the apparently stable mineral pairs sillimanite-muscovite and epidote-diopside. Layering and foliation strike north and dip gently to moderately east. The axial planes of minor folds dip gently or moderately north or northeast, and axes plunge gently east, parallel to mineral lineation.

Migmatite and granitic rocks in the Inner Piedmont were apparently formed by recrystallization and partial anatexis of the layered rocks of the Inner Piedmont during the climax of high-grade regional metamorphism.

The Brevard fault zone is a narrow zone of strongly sheared and retrogressively metamorphosed rocks, including porphyroclastic blastomylonite gneiss, blastomylonite, and phyllonitic muscovite-paragonite schist. Most rocks in the zone were probably derived from the flanking rocks, but the paragonite-bearing schist is apparently an exotic tectonic slice. Foliation in the zone is steeply dipping or vertical.

Rocks of both the Inner Piedmont and Blue Ridge belts near the fault zone show well-developed polymetamorphic textures characterized by porphyroclasts of potassic feldspar, plagioclase, or muscovite in a groundmass of recrystallized biotite, garnet, epidote, and oligoclase-andesine. Locally, adjacent to the fault zone, the rocks are retrogressively metamorphosed in the chlorite zone.

A belt as much as 5 miles wide along the Brevard fault zone shows pervasive structural and metamorphic effects related to the faulting. The Piedmont rocks are overprinted by a subhorizontal northeast-trending cataclastic mineral lineation and a northeast-trending and southeast-dipping cleavage. Minor folds in the Piedmont rocks in this belt are subisoclinal, and their axes trend northeast, parallel to the mineral lineation. Folds become tighter and axial planes and cleavage become steeper as the Brevard fault zone is approached.

Southeast-plunging mineral lineation in the Blue Ridge and Tablerock thrust sheets swings abruptly clockwise adjacent to the Brevard and becomes parallel to and indistinguishable from the subhorizontal northeast-trending cataclastic lineation along the southeast side of the fault zone. This observation suggests that strike-slip faulting along the Brevard was contemporaneous with and mechanically related to northwest movement of the Blue Ridge and Tablerock thrust sheets.

A dike of unmetamorphosed Upper Triassic(?) diabase cuts the rocks of the Inner Piedmont, the Brevard fault zone, the Blue Ridge thrust sheet, and the Grandfather Mountain window.

The rocks of the Blue Ridge thrust sheet moved northwestward at least 35 miles over the Grandfather Mountain window after the close of the metamorphism 350 m.y. ago (Late Devonian) and before Late Triassic(?) time. Left-lateral strike-slip movement along the Brevard was concurrent with, but may have lasted somehat longer than, thrusting. Lateral displacement was greater than 135 miles.

# INTRODUCTION

## LOCATION AND PHYSICAL GEOGRAPHY

The Grandfather Mountain window in northwestern North Carolina exposes rocks beneath the Blue Ridge thrust sheet near the southeastern edge of the Blue Ridge belt (fig. 1). The window, which is 45 miles long and 20 miles wide, and adjacent areas included in this study comprise about 1,000 square miles in North Carolina and adjacent parts of northeastern Tennessee. The northwestern part of the area lies mainly in the Blue Ridge upland. The southeastern part includes the deeply dissected southeastern slopes of the Blue Ridge upland and extends into the Morganton basin, a northeastward extension of the Piedmont plateau partly enclosed by the Brushy

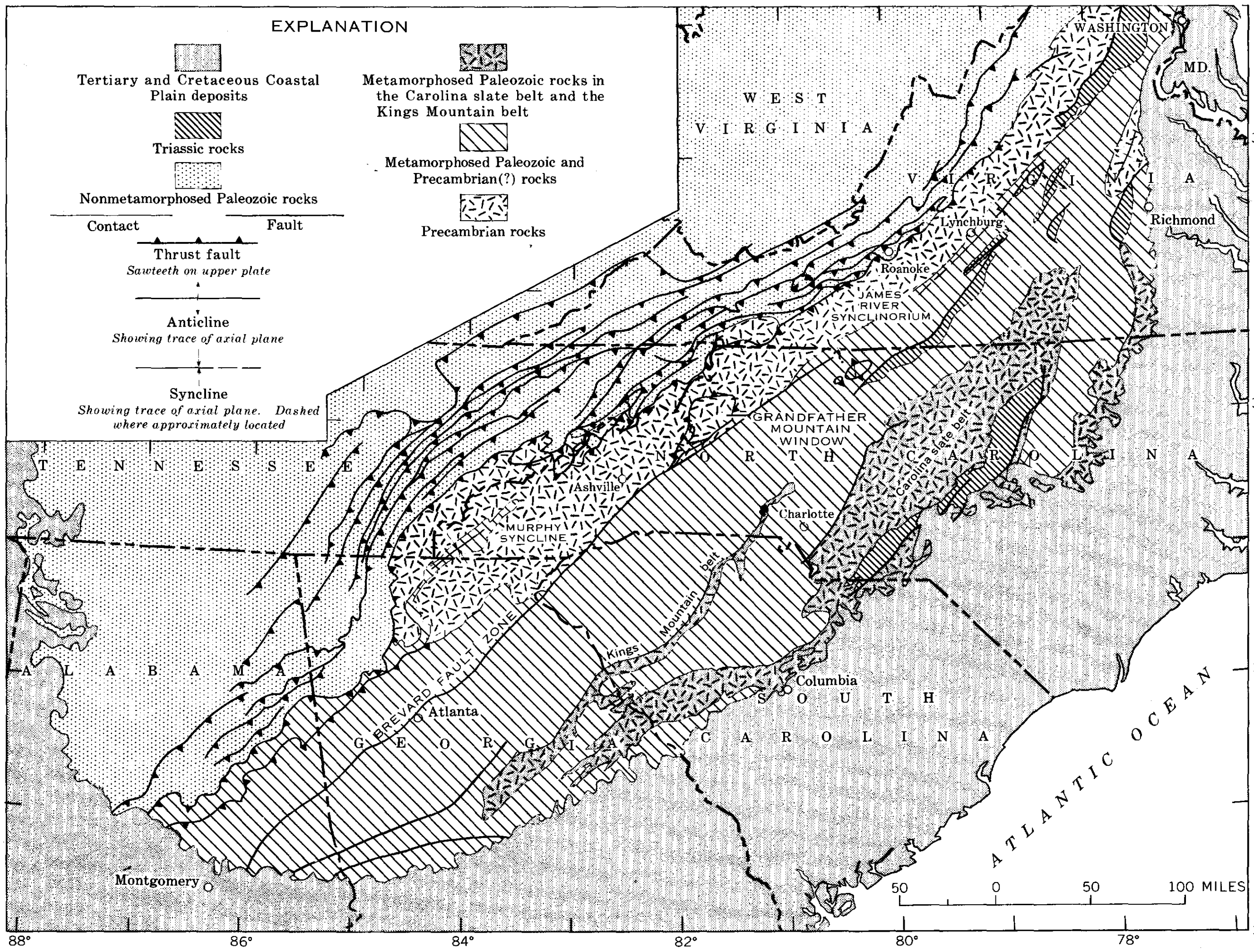



FIGURE 1.—Generalized geologic map of the southern Appalachian region showing the location of the Grandfather Mountain window. Generalized from the tectonic map of the United States (U.S. Geol. Survey and Am. Assoc. Petroleum Geologists, 1961).

Mountains on the northeast and by the South and Hickorynut Mountains on the south (fig. 2). Grandfather Mountain, from which the area takes its name, is the highest point (alt 5,939 feet) on the crest of the Blue Ridge.

The area includes all the Linville, Linville Falls[1], Blowing Rock, and Lenoir 15-minute guadrangles; adjacent parts of the Marion 15-minute quadrangle, the Little Switzerland, and Marion East 7½-minute quadrangles; parts of the Elk Mills and Sherwood 7½-minute quadrangles; and parts of the Maple Spring and Grandin 7½-minute quadrangles. Seven-and-a-half-minute topographic maps covering all the Linville, Blowing Rock, Linville Falls, and Lenoir 15-minute quadrangles are also available.

Several main highways (fig. 2) and a network of secondary roads and unmaintained roads and jeep trails make most of the area readily accessible. Even on the sparsely settled southeast flank of the Blue Ridge, it is difficult to find a point on a map more than 2 miles from a road.

Hilly and mountainous terrain of the Blue Ridge upland in the northwestern part of the area ranges from 2,500 to nearly 6,000 feet in altitude. The upland is drained by the South Fork of the New River, which flows northward into the Ohio River; by the Watauga, Elk, and North Toe Rivers, which flow westward into the Tennessee River; and by the Linville River, which flows southward to the Catawba River. Local relief ranges from a few hundred to 2,000 feet. Slopes are locally gentle, but large cliffs

[1] Called Table Rock quadrangle at an early stage in its preparation and referred to by that name in some earlier reports. Not to be confused with the Linville Falls 7½-minute quadrangle, which is the northwestern quarter of the Linville Falls 15-minute quadrangle and which is not mentioned below.

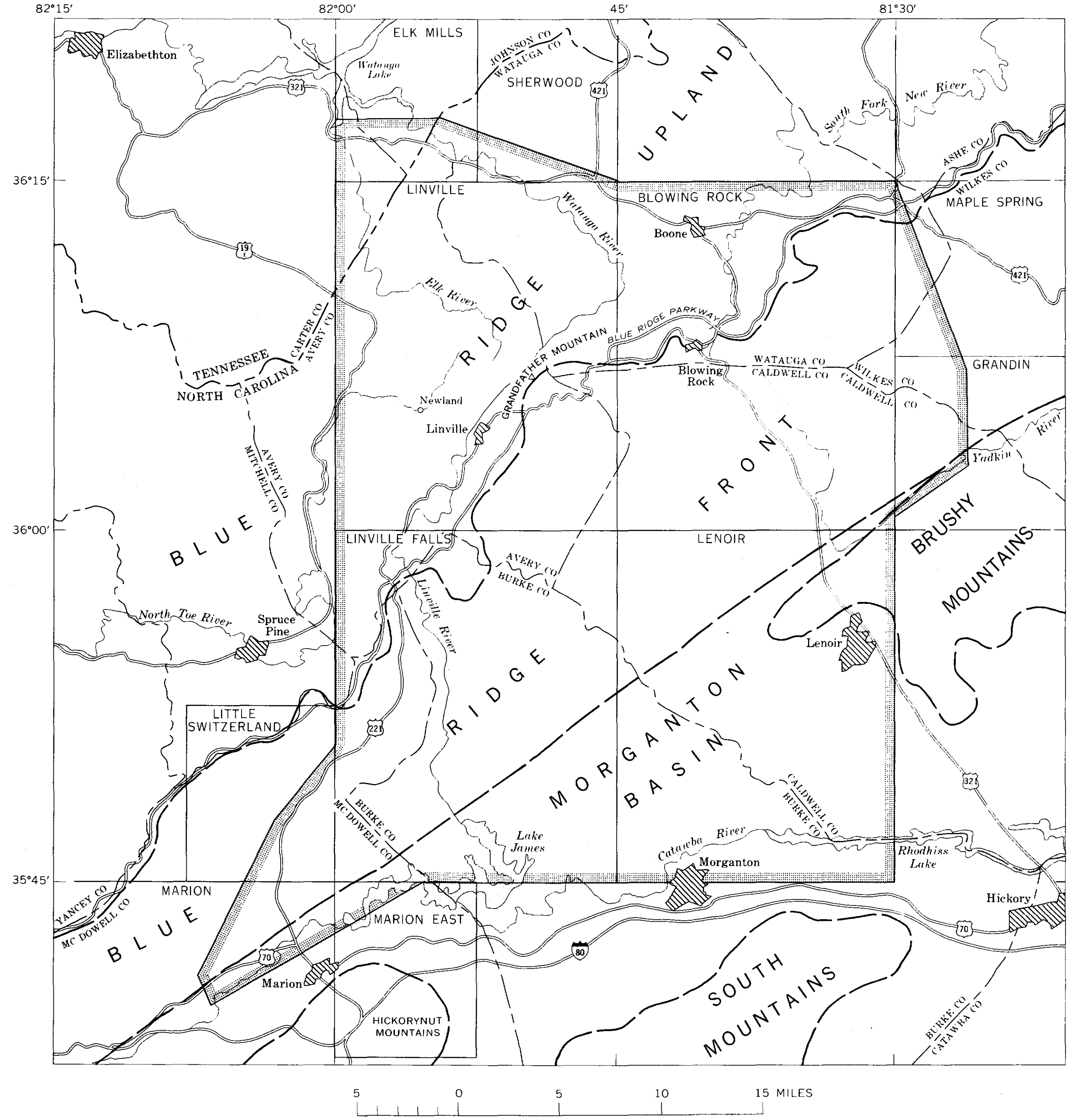



**FIGURE 2.**—Index map of the Grandfather Mountain area showing physiographic subdivisions, principal geographic features, political boundaries, and topographic quadrangles. Area of plate 1 outlined.

form the crests of some of the peaks, such as Grandfather Mountain. Outcrops of fresh rock are abundant, especially along the valleys of major streams. In some areas, however, the rock is deeply weathered and covered with colluvium.

The Blue Ridge upland has a moist temperate climate. Recorded annual precipitation ranges from 50 to 55 inches, but precipitation is probably greater locally. Mean July temperature in the valleys is 65° to 70°F; maximum summer temperatures seldom exceed 85°, and nights are cool. Minimum winter temperatures are as low as —10°F in the valleys and —20°F on the highest peaks, but periods of extreme cold are generally interspersed with milder weather, and snow cover seldom remains all winter. The mean temperature in January in the valleys is 34° to 36°F (figures from U.S. Weather Bureau, 1962). Northern hardwood forest covers most of the Blue Ridge upland, but fir and spruce grow locally on the higher peaks, particularly on Grandfather Mountain. Rhododendron grows luxuriantly in shady places; on sunnier slopes, laurel commonly forms the understory. Both laurel and rhododendron grow best on soils derived from siliceous rocks. Laurel is commonly called ivy by the local inhabitants, and rhododendron is referred to as laurel.

The principal towns of the Blue Ridge upland in the Grandfather Mountain area are Boone (site of Appalachian State Teachers College and county seat of Watauga County), Blowing Rock, Linville, and Newland (county seat of Avery County). The main sources of income are tourism and farming. Tobacco (mountain burley), cabbage, beans, and beef cattle are some of the chief products. Most farms are small, and many of those in the more remote hollows have been abandoned. There are a few small factories in the towns, but many people commute to work in the larger towns of the nearby Piedmont.

The Blue Ridge front constitutes the southeastern slope of the Blue Ridge upland. It is a steep, deeply dissected area about 10 miles wide which ranges in altitude from 3,500 to nearly 6,000 feet along the southeastern edge of the upland to 1,100 to 1,300 feet at the northwestern margin of the Piedmont. Average local relief is 1,000 to 1,500 feet, and streams draining the area have steep gradients. The headwaters of the Yadkin River drain the northeastern part of the belt, and tributaries of the Catawba River drain the southwestern part. The Linville River cuts through this belt in a spectacular gorge about 1,500 feet deep flanked by cliffs of quartzite (fig. 3). Outcrops are excellent along the streams, but deeply weathered rock is found on the ridge crests and on some of the steep slopes.

The Blue Ridge front is a zone of transition between the cooler climate and northern hardwood forest of the Blue Ridge upland and the warmer climate and southern hardwood forest of the Morganton basin. It is protected from the strong northwesterly winter winds that bring snow flurries on the Blue Ridge upland, but it receives the full effect of storms coming from the south and southeast. An understory of laurel is especially well developed on the quartzites, and rhododendron is found in shady places. Pine is abundant.

The Blue Ridge front was once thickly settled. Many farms in the more remote areas are now abandoned, but some subsistence farming is still done along bottoms of the major valleys. A large part of the front is in the Pisgah National Forest, and considerable logging is done. The Linville Gorge Wild Area, which includes the least accessible and most

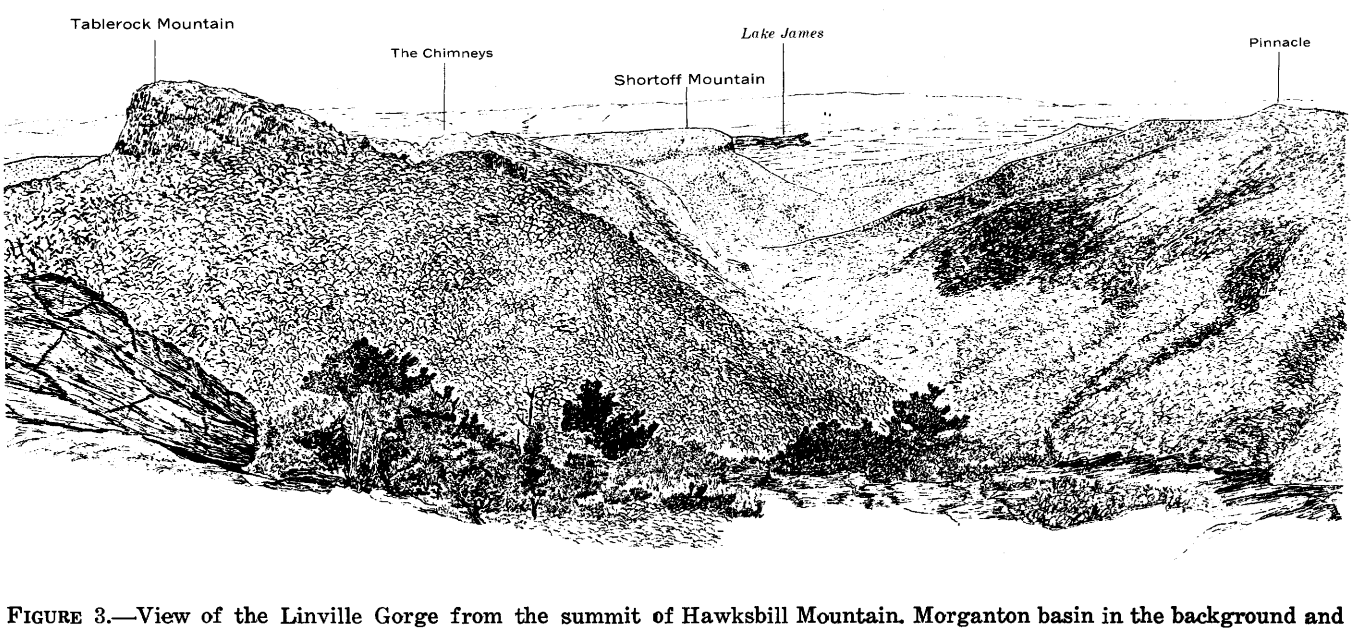



FIGURE 3.—View of the Linville Gorge from the summit of Hawksbill Mountain. Morganton basin in the background and South Mountains on the horizon. Rocks in the foreground are arkose of the Grandfather Mountain Formation. Prominent cliffs on Tablerock Mountain and Shortoff Mountain are quartzite of the Chilhowee Group.

scenic part of the valley of the Linville River, is one of the few designated wild areas in the eastern United States.

The surface of the Piedmont plateau in the Morganton basin lies at altitudes of 1,100 to 1,300 feet, more than 2,000 feet lower than most of the Blue Ridge upland. Small hills and ridges stand as monadnocks a few hundred feet above the plateau surface, and streams are incised as much as 200 feet into it, leaving a series of concordant flat-topped interfluves. In the northeastern part of the Lenoir quadrangle the southwest end of the Brushy Mountains stands as hills as much as 800 feet above the Piedmont surface. The Morganton basin is drained by the Catawba River and its tributaries. Parts of two large reservoirs are included in the area: Lake James, which is impounded by dams across the Catawba River and the Linville River west of Morganton, and Rhodhiss Lake, which is impounded by a dam across the Catawba northwest of Hickory (fig. 2).

Outcrops of fresh rock are scarce in most of the Morganton basin, but some occur along the sides of stream valleys and on the flanks of some hills and ridges that rise above the general level of the basin floor. Numerous exposures of weathered rock and saprolite are found in roadcuts, gullies, and along the shorelines of the large reservoirs.

The Morganton basin has hot summers and rather mild winters. Snow occasionally falls and temperatures as low as 0°F occur during some winters, but 50°F is also a common midwinter temperature. Mean temperatures for July are 76° to 78°F and for January, 41° to 43°F. Annual precipitation is about 50 inches (U.S. Weather Bureau, 1962). Pine and southern hardwood forests cover uncultivated parts of the Piedmont. Although laurel and rhododendron are scarce, many vines, including Virginia-creeper, cudzoo, and blackberry, add to the undergrowth.

Morganton and Lenoir, the county seats of Burke and Caldwell Counties, respectively, are the largest towns in the Grandfather Mountain area. Furniture and textile manufacturing are the principal industries. Some of the interfluves and many of the alluvial flats bordering the streams are farmed, but much of the northwestern part of the area is wooded.

## PREVIOUS GEOLOGIC INVESTIGATIONS

The unique nature of the rocks of the Grandfather Mountain area was recognized during the earliest geological reconnaissance in the southern Appalachians. Early maps (Maclure, 1817; Kerr 1875) show rocks of "transition" age in the Grandfather Mountain area amidst a terrane of gneisses classified as "Primitive" or "Huronian" in age.

In the last decade of the 19th century and the first decade of the 20th, Arthur Keith of the U.S. Geological Survey did much of the pioneer geologic mapping in the crystalline belt of the southern Appalachians. He mapped all the Grandfather Mountain area and wide expanses in the Blue Ridge to the north, west, and southwest. The Linville and Blowing Rock quadrangles compose the south half of the Cranberry 30-minute quadrangle (Keith, 1903); the Linville Falls and Lenoir quadrangles constitute the north half of the Morganton 30-minute quadrangle (Keith and Sterrett, 1954). The southwestern part of the Grandfather Mountain area is included in the Mount Mitchell 30-minute quadrangle (Keith, 1905).

In the Grandfather Mountain area, Keith mapped an extensive area of sedimentary and volcanic rocks of low metamorphic grade to which he assigned a late Precambrian and early Cambrian age. He believed that these rocks occupied a complex syncline bounded on the north and west by faults along which Precambrian plutonic rocks had overridden the younger rocks from three sides. This structural feature was reinterpreted on the geologic map of the United States (Stose and Ljungstedt, 1932) as a window in a major overthrust sheet, now called the Grandfather Mountain window (Stose and Stose, 1944, p. 383).

During reconnaissance for the geologic map of the United States, Jonas (1932) recognized that rocks of low metamorphic grade southeast of the Grandfather Mountain window were polymetamorphic and that they were continuous with similar rocks along strike to the northeast and southwest. She interpreted this belt of retrogressive rocks as marking the sole of a great overthrust continuous with the Martic overthrust of southeastern Pennsylvania.

George and Anna (Jonas) Stose are reported to have mapped the Grandfather Mountain window (Stose and Stose, 1944; 1950; Miser, 1962, p. 146), but their maps were never published and were not available to us during our work.

Geologic mapping in connection with mineral resource investigations in the Spruce Pine district west of the Grandfather Mountain area extends into the western parts of the Linville and Linville Falls quadrangles (Parker, 1946; Brobst, 1962).

A report on the geology of northeastern Tennessee by King and Ferguson (1960) covered the northwestern corner of the Linville quadrangle, and Hamilton (in King and Ferguson, 1960, p. 13–27) discussed the basement rocks in that area.

Topical studies by Eckelmann and Kulp (1956) and White (1950) have furnished information on various aspects of the geology of the Grandfather Mountain area.

Two small areas in the Grandfather Mountain window have been mapped in connection with theses (Bright, 1956; Goedicke, 1950), but the maps were not published.

Deposits of the following minerals and commodities in the Grandfather Mountain area have been discussed briefly in various reports: mica (Olson, 1944; Griffitts, 1953; Sterrett, 1907, 1910, 1923), iron (Nitze, 1893; Bayley, 1923; Kline and Ballard, 1948), gold (Nitze and Hanna, 1896; Bryson, 1936; Pardee and Park, 1948), sillimanite (Hash and Van Horn, 1951; Espenshade and Potter, 1960), limestone (Conrad, 1960; Loughlin and others, 1921; Watson and Laney, 1906; Lewis, 1893), and quartz (Mertie, 1959).

Various short reports and maps of parts of the Grandfather Mountain area have been published as our work progressed. These include Bryant (1962, 1963, 1965, 1966, 1967), Bryant and Reed (1961, 1962, 1970a, b), Reed (1964a, b), Reed and Bryant (1964b), Reed, Johnson, Bryant, Bell, and Overstreet (1961), Reed, Bryant, and Hack (1963), and Reed, Bryant, Leopold, and Weiler (1964). The map of the southwestern extension of the window was released to open file (Reed and Bryant, 1964a).

Economic geology of the Grandfather Mountain area is discussed in a separate short report (Bryant and Reed, 1966). Surficial deposits are not discussed in the present report as we have little to add to our preliminary reports (Bryant 1962; Reed 1964b) and two short topical papers (Reed and others, 1963, 1964). J. T. Hack has studied some aspects of the geomorphology of the area (1966).

The first draft of this report was completed in 1964; some references to important more recent work have been added as recently as 1969.

## PRESENT INVESTIGATION

### FIELDWORK AND ACKNOWLEDGMENTS

Bryant made a reconnaissance of the entire area in August 1956 and began geologic mapping in the Linville quadrangle in September 1956; Reed started fieldwork in the Linville Falls quadrangle in March 1957. Both authors spent about 10 weeks in the field each spring and fall in 1957, 1958, and 1959 and 3 to 4 months each fall and winter from 1960 until completion of fieldwork in February 1962. Each author spent approximately 24 months in the field, excluding time spent in reconnaissance in adjacent areas and in field conferences. Of this time, Bryant had field assistants about 16 months, and Reed, 13 months. We wish to acknowledge their help: C. E. Fritts, fall 1956 and spring 1957; F. G. Lesure and C. W. Spencer, fall 1957; William Van Horn and D. V. Lewis, spring 1958; C. E. Harris, Jr., and C. A. Shelby, fall 1958; R. L. Beck and K. E. Billeau, spring 1959; H. W. Sundelius, fall 1961; and D. B. Andretta, fall 1961 and winter 1962.

We wish to thank the inhabitants of the towns in which we stayed for their interest and assistance, and the landowners who, almost without exception, readily gave permission to do geologic mapping on their lands. Mr. Hugh Morton permitted us to use his toll road on Grandfather Mountain.

The investigation has been nurtured by discussion with many of our colleagues, especially J. B. Hadley, Warren Hamilton, W. B. Myers, R. A. Laurence, W. C. Overstreet, D. A. Brobst, F. G. Lesure, P. B. King, and J. T. Hack of the U.S. Geological Survey; John Rodgers of Yale University; S. W. Maher and G. D. Swingle of the Tennessee Division of Geology; S. G. Conrad of the North Carolina Department of Conservation and Development (now State Geologist of North Carolina); and H. S. Johnson, Jr., State Geologist of South Carolina.

W. M. Cady and D. W. Rankin made helpful suggestions concerning presentation of the data and conclusions given below.

### METHODS OF STUDY

Field mapping was done on 1:48,000-scale enlargements of the 15-minute topographic maps. In mountainous parts of the area, most contacts were traced in the field, but in the Piedmont, most contacts were interpolated between scattered exposures plotted on outcrop maps at a scale of 1:24,000. Contacts of surficial deposits were largely sketched from aerial photographs but were locally checked in the field.

Approximately 2,100 rocks were studied petrographically, and 103 samples were chemically analyzed. Modes of most of the analyzed rocks were determined by standard point counts and are given in the tables of analyses. Some of the analyzed rocks were so fine grained that the modes determined by point counts could not be reconciled with the analyses; these modes were discarded. Semiquantitative spectrographic determinations of trace elements were made on 27 of the analyzed rocks. The trace-element data are given in table 1 for reference, but they are not referred to further in this report.

Estimates of the average bulk composition of large heterogeneous rock units were made by counting 50

TABLE 1.—*Minor-element analyses of rocks from the Grandfather Mountain area*

[Analyses were determined by semiquantitative spectrographic methods by Paul R. Barnett, U.S. Geol. Survey, in 1959 for 1, 2, 7, 12, 13, 14, 15, and 17 and in 1961 for 4, 9, 10, and 19; and by John C. Hamilton, U.S. Geol. Survey, in 1961 for 3, 5, 6, 8, 11, 16, 23, 24, and 26 and in 1962 for 20, 21, 22, 25, and 27. Results are reported in weight percent to the nearest number in the series 1, 0.7, 0.5, 0.3, 0.2, 0.15, 0.1, and so forth, which represent the approximate midpoints of group data on a geometric scale. The assigned group for semiquantitative results will include the quantitative value about 30 percent of the time. D, barely detectable; ---, not determined]

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Table | 8 | 9 | 10 | 13 | 14 | | 15 | | | 16 | | | |
| Column in given table | 1 | --- | --- | 2 | --- | 2 | 4 | 8 | 1 | 4 | 6 | 1 | 2 |
| Field No. | GML–6 | GML–5 | GML–9 | GMB–4 | G | GML–8 | GML–7 | J–20–1 | GMB–2 | GML–13 | GML–12 | GML–3 | GML–1 |
| Lab No. | F–2589 | F–2588 | H–3424 | H–3264 | H–3434 | H–3423 | F–2590 | H–3427 | H–3263 | H–3262 | H–3425 | F–2418 | F–2416 |
| B | 0 | 0 | 0.003 | 0 | 0 | 0 | 0 | 0.003 | 0 | 0 | 0 | 0 | 0 |
| Ba | .007 | .015 | .007 | .15 | .005 | .1 | .07 | .3 | .07 | .07 | .07 | .03 | .015 |
| Be | .0007 | .00015 | .001 | .00015 | .0002 | .0007 | .0003 | .0007 | .0003 | .0003 | .00015 | 0 | 0 |
| Ce | .015 | .03 | 0 | .03 | .02 | 0 | 0 | .03 | .03 | .07 | 0 | 0 | 0 |
| Co | 0 | 0 | 0 | .0007 | 0 | 0 | .0003 | .002 | .0015 | .0015 | 0 | .003 | .007 |
| Cr | d | d | 0 | .0007 | 0 | .001 | .0003 | .003 | .007 | .007 | .002 | .007 | .007 |
| Cu | .00015 | .00015 | .0005 | .0015 | .0003 | .0003 | .00015 | .0002 | .003 | .015 | .0015 | .007 | .007 |
| Ga | .0015 | .0015 | .005 | .0015 | .005 | .003 | .0015 | .005 | .003 | .003 | .002 | .003 | .003 |
| La | .007 | .03 | .005 | .015 | .01 | .005 | 0 | .03 | .015 | .015 | 0 | 0 | 0 |
| Mo | 0 | .0003 | 0 | 0 | .0007 | 0 | 0 | .0015 | 0 | 0 | 0 | 0 | 0 |
| Nb | .003 | .003 | .015 | 0 | .01 | .002 | .0015 | .002 | .0015 | .003 | 0 | .0015 | .0015 |
| Nd | .007 | .03 | ------ | .015 | .015 | ------ | 0 | .03 | .03 | .015 | ------ | 0 | 0 |
| Ni | 0 | 0 | 0 | 0 | 0 | 0 | 0 | .003 | .003 | .0015 | 0 | .003 | .007 |
| Pb | .0015 | .0007 | 0 | .0007 | .005 | .0015 | d | .0015 | .0015 | .0015 | .005 | 0 | 0 |
| Sc | 0 | .0003 | 0 | .0015 | 0 | .0007 | .0003 | .005 | .0015 | .0015 | .0007 | .0015 | .003 |
| Sn | 0 | 0 | .002 | 0 | .001 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| Sr | .003 | .003 | .001 | .07 | .001 | .01 | .007 | .015 | .03 | .07 | .07 | .03 | .07 |
| V | 0 | 0 | 0 | .03 | 0 | .003 | .0015 | .002 | 0 | .015 | .003 | .03 | .03 |
| Y | .007 | .007 | .015 | .003 | .02 | .005 | .003 | .02 | .007 | .007 | .005 | .003 | .003 |
| Yb | .0007 | .0007 | .0015 | .0003 | .002 | .0007 | .0003 | .001 | .0007 | .0007 | .0005 | .0003 | .0003 |
| Zr | .03 | .03 | .07 | .015 | .05 | .03 | .03 | .02 | .015 | .03 | .02 | .015 | .007 |

| | 14 | 15 | 16 | 17 | 18 | 19 | 20 | 21 | 22 | 23 | 24 | 25 | 26 | 27 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Table | | 17 | | | | | 20 | | | 26 | | 27 | | 28 |
| Column in given table | 3 | 6 | 7 | 8 | 10 | 16 | 2 | 7 | 1 | 3 | 4 | 1 | 3 | 1 |
| Field No. | GML–2 | W–84 | GMB–6 | GMB–1 | 34(L) | GMB–5 | BR–71A | BR–71B | 28–1045 | 15–1061 | 21–2331 | 2583(L) | B | 2028–L |
| Lab No. | F–2417 | F–2419 | H–3426 | F–2591 | I–4057 | H–3265 | I–4058 | I–4059 | I–4061 | H–3433 | H–3432 | I–4060 | H–3431 | I–4062 |
| B | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| Ba | .015 | .003 | .15 | .15 | .1 | .07 | .1 | .07 | .15 | .15 | .01 | .07 | .07 | .07 |
| Be | 0 | 0 | 0 | .0003 | .0003 | .0003 | .0002 | .0002 | 0 | .0002 | .0005 | .0003 | .0002 | .0002 |
| Ce | 0 | 0 | .015 | .015 | .02 | .015 | 0 | 0 | 0 | <.02 | <.02 | 0 | <.02 | 0 |
| Co | .007 | .003 | .005 | .0007 | .0005 | 0 | .0015 | .0005 | .0007 | .0007 | 0 | .001 | .0007 | .0015 |
| Cr | .007 | .003 | .0007 | .00015 | .0003 | .00015 | .003 | .0005 | .001 | .0007 | 0 | .002 | .0007 | .002 |
| Cu | .0015 | .0015 | .002 | .0007 | .0007 | .0003 | .007 | .001 | .002 | .001 | .0005 | .0005 | .0007 | .001 |
| Ga | .0015 | .003 | .005 | .0007 | .003 | .0007 | .003 | .002 | .002 | .002 | .005 | .002 | .002 | .002 |
| La | 0 | 0 | .007 | .007 | .01 | .007 | .005 | 0 | .005 | .01 | .005 | 0 | .005 | 0 |
| Mo | 0 | 0 | .001 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| Nb | .0015 | .0015 | .002 | .0015 | .003 | .003 | 0 | 0 | .002 | .002 | .015 | 0 | .0015 | 0 |
| Nd | 0 | 0 | .015 | .007 | .015 | .007 | 0 | 0 | 0 | .015 | 0 | 0 | 0 | 0 |
| Ni | .007 | .003 | .001 | 0 | 0 | 0 | .003 | 0 | .0007 | 0 | 0 | .0007 | .0007 | .001 |
| Pb | 0 | 0 | .0015 | 0 | .002 | .0007 | .002 | .002 | .002 | .003 | .003 | .002 | .005 | .002 |
| Sc | .003 | .003 | .003 | .0015 | .0007 | 0 | .001 | .0005 | .0015 | .002 | 0 | .0015 | .001 | .002 |
| Sn | 0 | 0 | 0 | 0 | 0 | .0007 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| Sr | .03 | .03 | .07 | .015 | .02 | .007 | .07 | .05 | .05 | .02 | .0015 | .05 | .015 | .03 |
| V | .03 | .03 | .01 | .003 | .0015 | 0 | .02 | .003 | .01 | .005 | 0 | .01 | .005 | .02 |
| Y | .003 | .003 | .005 | .003 | .007 | .015 | .003 | .002 | .003 | .005 | .03 | .003 | .003 | .003 |
| Yb | .0003 | .003 | .0005 | .0003 | .0007 | .0015 | .0003 | .0002 | .0003 | .0005 | .003 | .0003 | .0003 | .0003 |
| Zr | .015 | .015 | .03 | .03 | .03 | .03 | .015 | .01 | .03 | .05 | .03 | .007 | .007 | .01 |

grains in each of a large number of thin sections. The grains were selected by random movements of the thin section on the microscope stage. For many rock units, modes estimated in this way are plotted in triangular diagrams to illustrate the range in composition and degree of heterogeneity of the unit. On these diagrams, the contours represent the percentage of points that fall within 1 percent of the area of the diagram.

Composition and optical properties of most individual minerals were not determined. However, plagioclase compositions were determined by measurement of extinction angles in thin section or of indices of refraction in immersion oils, and composition and optical properties of some of the iron-rich muscovitic micas were studied in some detail. All photomicrographs were taken with crossed polarizers except where noted. X-ray techniques were used to assist in some of the mineral identifications.

## GEOGRAPHIC LOCATIONS AND LOCATIONS OF TYPICAL OUTCROPS

In order to facilitate reference to specific geographic locations mentioned in this report, an arbitrary grid has been placed on the geologic map (pls. 1,2). Blocks in the grid are approximately rectangular areas 3.75 minutes of latitude and longitude in extent. They are lettered eastward from A through K and numbered southward from 1 through 10, beginning in the northwest corner of the map area. Whenever reference is made to a particular geographic feature, the grid area in which the feature lies is mentioned by letter and number.

In order to conserve space, locations of typical outcrops of the various map units are not described in the text, but are given in table 31 at the end of the report.

## PETROGRAPHIC NOMENCLATURE

In this report, mineral modifiers of petrographic names are listed in order of increasing abundance. For example, biotite-muscovite-quartz schist denotes a schist containing more quartz than muscovite and more muscovite than biotite. Many rocks in the Grandfather Mountain area have strongly developed cataclastic textures. The nomenclature used in describing these rocks is outlined in table 2. In most of the cataclastic rocks described herein, recrystallization and cataclasis were virtually synchronous.

The term "porphyroclast" is used to describe large mineral fragments or grains set in a finely granulated matrix. Porphyroclasts may be fragments or grains that are from an originally coarser grained, even-textured rock and that have escaped granulation, or they may have been phenocrysts or prophyroblasts in the original rock. In many rocks, the origin of the larger grains cannot be determined, but it is clear that they did not recrystallize during the latest cataclastic metamorphism.

TABLE 2.—*Classification of cataclastic rocks*

| Degree of recrystallization of matrix | Degree of cataclasis | | |
|---|---|---|---|
| | Rock granulated around grain boundaries | Rock consists of 90 to 10 percent porphyroclasts in fine-grained matrix | Rock contains less than 10 percent porphyroclasts |
| Unrecrystallized | Mortar gneiss | Mylonitic gneiss | Mylonite. |
| Recrystallized | Recrystallized mortar gneiss. | Phyllonitic gneiss (where recrystallized material is largely sericite). | Phyllonite. |
| Do | do | Blastomylonitic gneiss (where recrystallized material is largely quartz and feldspar). | Blastomylonite. |

The term "plutonic rock" refers to rocks that have or can reasonably be inferred to have had a medium- and coarse-grained allotriomorphic or hypidiomorphic granular texture. Figure 4 shows the classification of silicic plutonic rocks used in this paper. A plutonic metamorphism is defined as a metamorphic event in which large masses of plutonic rocks were emplaced, no matter what the mechanism of emplacement may have been.

Sedimentary rock names are used for arenaceous rocks even where they are of low metamorphic grade. Metamorphic rock names are generally applied to argillaceous of equivalent metamorphic grade.

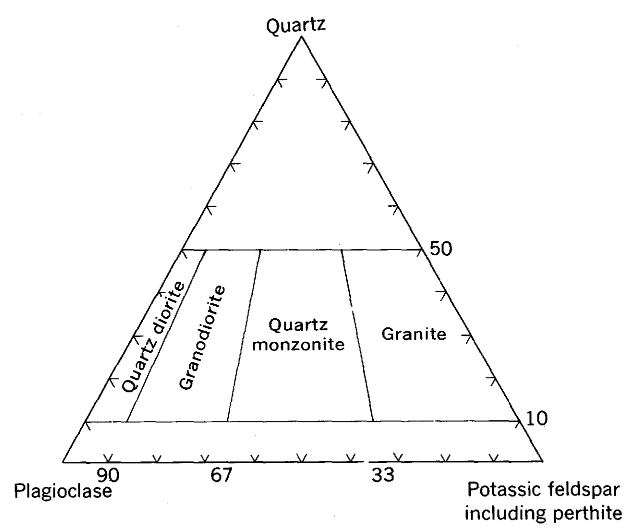



FIGURE 4.—Classfication of silicic plutonic rocks used in this paper.

Chlorites are classified using the system suggested by Albee (1962, p. 868).

## GENERAL GEOLOGY

The Blue Ridge in western North Carolina and eastern Tennessee is composed largely of gneiss, schist, migmatite, and granitic rock that are the products of metamorphism and plutonism that occurred at about the time of the metamorphism of the Grenville Series in the western Adirondacks a billion years ago. Stratigraphically overlying these rocks are thick sequences of sedimentary and volcanic rocks of later Precambrian age, deposited after the 1.000-m.y.-old event. These comprise the Ocoee Series, the Grandfather Mountain Formation, and the Mount Rogers Formation (pl. 3). All the Precambrian rocks were subjected to one or more episodes of thermal and dynamic metamorphism during Paleozoic time. During the later part of the Paleozoic and perhaps early Mesozoic, Precambrian crystalline rocks were thrust northwestward at least 35 miles over unmetamorphosed upper Precambrian and lower Paleozoic rocks in the Unaka belt (pls. 3, 4). The Grandfather Mountain window exposes metamorphosed Cambrian, upper Precambrian, and lower Precambrian rocks below the Blue Ridge thrust sheet many miles southeast of any other exposures of rocks of this tectonic level yet recognized in this part of the Blue Ridge.

During and after thrusting, the Blue Ridge thrust sheet was sundered by the Brevard fault, a large-scale strike-slip fault southeast of the Grandfather Mountain window. This movement juxtaposed an entirely different terrane of crystalline rocks of the Inner Piedmont belt with the Blue Ridge thrust sheet. The Inner Piedmont belt is composed of gneiss, schist, and granitic rocks of early Paleozoic or older ages, which underwent dynamothermal and plutonic metamorphism one or more times during the early and middle Paleozoic and which were retrogressively metamorphosed during movement along the Brevard fault zone. A diabase dike probably of Late Triassic age extends undisturbed across both the Brevard fault zone and the thrust fault bounding the Grandfather Mountain window.

The area described in this report includes all 470 square miles of the Grandfather Mountain window and adjacent parts of the Blue Ridge thrust sheet. It also includes nearby parts of the Mountain City window in the Unaka belt northwest of the Blue Ridge and parts of the Brevard fault zone and Inner Piedmont belt southeast of the Blue Ridge.

Because of their fundamental differences, the rocks of each of these major tectonic units are described and their structural and metamorphic histories are discussed separately before possible relations between tectonic units are considered. Discussions of the various tectonic units are arranged geographically from northwest to southeast.

## MOUNTAIN CITY WINDOW

Sedimentary rocks of Early Cambrian and Cambrian(?) age beneath the Blue Ridge thrust sheet are exposed in the Mountain City window in the northwestern corner of the Grandfather Mountain area (A–1, pl. 1). The tectonically lowest rocks in the Mountain City window in the Grandfather Mountain area are in the Doe River inner window (pl. 3) according to Rodgers (1953a). Tectonic slices of basement rocks and sedimentary rocks of Early Cambrian and Cambrian (?) age southeast of the Doe River inner window in the Grandfather Mountain area lie beneath the Blue Ridge thrust sheet and occupy an intermediate tectonic position between it and autochthonous rocks in the Doe River inner window.

The following descriptions apply to the sedimentary rocks exposed in the Doe River inner window in the Linville quadrangle. Basement rocks in the intermediate tectonic slices to the southeast are described with the rocks of the Blue Ridge thrust sheet. Sedimentary rocks in other parts of the Doe River inner window and in the intermediate tectonic slices were not examined petrographically during this study.

### ROCK UNITS

#### CHILHOWEE GROUP

Arenaceous rocks of the Chilhowee Group have well-developed detrital textures; the more argillaceous rocks are cleaved. Quartz is the dominant clastic mineral and is generally strongly strained. Microcline, biotite, and muscovite also occur as clastic grains. Biotite and muscovite clasts are locally altered to sericite. Clastic green tourmaline and zircon are widespread. Sphene, ilmenite or magnetite, epidote, and rutile are other accessory minerals. The rocks contain various amounts of matrix consisting of very fine grained sericite and chlorite. A few samples studied microscopically have a wide range of mica and chlorite content and contain less than 20 percent feldspar. These rocks contrast with the upper Precambrian rocks of the Grandfather

Mountain window but resemble the rocks of the Tablerock thrust sheet in their rather low feldspar content, lack of clastic plagioclase, and presence of clastic tourmaline. King and Ferguson (1960, p. 119) recorded 1,975 feet of beds of the Chilhowee Group in an incomplete section on Nowhere Ridge (area C–2, pl. 1).

#### UNICOI FORMATION

The Unicoi Formation (Keith, 1903) is composed of fairly coarse to fine-grained, light-gray, tan, and green arkosic quartzite, conglomeratic arkosic quartzite, vitreous quartzite, and darker gray more argillaceous beds. Thin dark-gray to black beds, a millimeter or less thick, are rich in heavy minerals. The lower part of the formation is cut out by a thrust fault. Some gray quartzite locally has many fractures containing iron oxide and pyrite. King and Ferguson (1960, p. 119) measured 1,010 feet of Unicoi on Nowhere Ridge.

#### HAMPTON FORMATION

The Hampton Formation (Keith, 1903) consists of thin-bedded gray shale, siltstone, and feldspathic quartzite. Sandy beds ½ to 2 inches thick are lenticular in some places. X-ray diffractometer study by Paul D. Blackmon, U.S. Geological Survey, of a sample of shale from the Hampton Formation from area C–1 (pl. 1) indicates that it is composed predominantly of mica, quartz, and chlorite.

#### ERWIN FORMATION

The Erwin Formation (Keith, 1903) consists of light-gray, tan, and greenish-gray vitreous quartzite and feldspathic quartzite, locally containing *Scolithus* and shaly partings and beds. A few quartz pebbles 1 cm (centimeter) in diameter are present. At the top of the Erwin Formation is an interval of light-green shale containing 1- to 3-foot-thick quartzite beds. This unit may correspond to the Helenmode Member as used by King and Ferguson (1960), but it is too thin to distinguish on the 1:62,500-scale geologic map (pl. 1). King and Ferguson (1960, p. 119) measured 605 feet of Erwin on Nowhere Ridge.

#### AGE

The Chilhowee Group generally lacks diagnostic fossils in northeast Tennessee (King and Ferguson, 1960, p. 36), although Keith (1903) reported some from the top of the Erwin Formation. West of the Great Smoky Mountains, Early Cambrian fossils have been found in the Murray Shale (Laurence and Palmer, 1963), which is equivalent to about the middle of the Erwin Formation in the Mountain City window. The rocks between the lowest diagnostic fossils and the base of the Chilhowee Group are considered to be Cambrian (?) in age by the U.S. Geological Survey.

### SHADY DOLOMITE

The Shady Dolomite (Keith, 1903) is locally well exposed in the Elk River valley (C–1, pl 1). It is a fine- to coarse-grained, thin- to thick-bedded dark- to light-gray, blue-gray, and white dolomitic limestone. Some of the gray dolomite has spots of coarse-grained white dolomite. The Shady Dolomite is about 1,200 feet thick. No fossils are known in the Shady Dolomite in northeastern Tennessee (King and Ferguson, 1960, p. 52), but fossils from the Shady in southwestern Virginia fix the age of the formation as Early Cambrian (Resser, 1938, p. 24–25).

### ROME FORMATION

The Rome Formation (Hayes, 1891, p. 143) is composed predominantly of dull-green to red siltstone and shale and contains few interbeds of white quartzite. Beds of dolomite and shaly dolomite are widespread according to King and Ferguson (1960, p. 53), but none were noted in the Linville quardrangle. X-ray diffraction study of a red shale from the Rome Formation just north of the Linville quadrangle by Paul D. Blackmon, U.S. Geological Survey, indicates that it is composed predominantly of mica having a slight amount of mixed layering, and quartz, hematite, and feldspar.

No fossils have been found in the Rome Formation in northeastern Tennessee (King and Ferguson, 1960, p. 53), but elsewhere it contains Early Cambrian fossils (Resser, 1938, p. 23–24).

## STRUCTURE AND METAMORPHISM

The sedimentary rocks in the Doe River inner window in the Grandfather Mountain area lie on the eastern limb of a syncline overturned to the west. They dip generally eastward beneath intermediate thrust slices near the sole of the Blue Ridge thrust sheet. Incompetent rocks of the Rome Formation are in subisoclinal folds overturned to the west. Cleavage parallel with the axial planes of minor folds is developed locally in argillaceous rocks.

In hand specimen the rocks seem to be unmetamorphosed, the shales contain mica and chlorite rather than clay minerals, and some of the mica has a slightly mixed layering. It is uncertain whether these rocks belong in the quartz-albite-muscovite-chlorite subfacies of the greenschist facies or the zeolite facies of regional metamorphism.

## BLUE RIDGE THRUST SHEET

The Blue Ridge thrust sheet surrounding the Grandfather Mountain window is a complex crystalline terrane composed of biotite-muscovite schist and gneiss, amphibolite, hornblende gneiss, and migmatite intruded by plutons of Precambrian granite and quartz monzonite, by stocks and dikes of upper Precambrian metagabbro and metadiabase, and by stocks, sills, and dikes of ultramafic rock, granodiorite, and pegmatite of early or middle Paleozoic(?) age. Both northwest and southeast of the window, thin slices of Cambrian(?) sedimentary rock are intercalated in the gneisses along subsidiary thrust faults.

West of the Grandfather Mountain window, biotite-muscovite schist and gneiss, amphibolite, and hornblende gneiss grade northward into granitic rocks. The transition occurs in a broad zone; layers and lenses of granitic material become increasingly abundant, and the rocks pass into layered migmatitic gneiss in which granitic layers predominate and finally into rudely layered and nonlayered granitic gneiss. A similar but narrower transition zone is found north of the Grandfather Mountain window, but in some areas (J–2, J–3, K–3, C–6, C–7, and C–8, pl. 1) mica and hornblende gneiss are in sharp contact with layered migmatitic gneiss.

Layering and foliation in rocks of the Blue Ridge thrust sheet generally dip away from the Grandfather Mountain window. West and north of the window is a belt of rocks of dominantly granitic aspect retrogressively metamorphosed to a low grade before or during the thrusting. Tectonically overlying rocks were metamorphosed to medium grade in the middle Paleozoic. Southeast of the window, especially in a narrow belt between the window and the Brevard fault, medium-grade rocks have been partly retrograded.

### ROCK UNITS

#### MICA SCHIST, MICA GNEISS, AND AMPHIBOLITE

Muscovite-biotite schist and gneiss and amphibolite make up the bulk of the Blue Ridge thrust sheet west and northeast of the Grandfather Mountain window (Keith, 1903, 1905; Brobst, 1962). The micaceous rocks were called Carolina Gneiss by Keith, a name which he applied to lithologically similar rocks in wide areas in the Blue Ridge and Piedmont (Keith and Darton, 1901; Keith, 1907b). He was uncertain whether they were derived from sedimentary or plutonic rocks. Keith called the associated hornblende-rich rocks Roan Gneiss and believed them to be metamorphosed diorites intrusive into the Carolina Gneiss. On Roan Mountain, about 6 miles west of the Grandfather Mountain area, many metamorphosed dikes of Bakersville Gabbro cut layered gneiss. These dikes were apparently miscorrelated by Keith with the amphibolite and hornblende schist interlayered with the mica schist and gneiss; this interpretation led him to the conclusion that the Roan Gneiss was intrusive.

Because the terms "Carolina Gneiss" and "Roan Gneiss" were used by Keith primarily as lithologic terms, they have been abandoned as stratigraphic names (Brobst, 1962).

Hornblendic and micaceous rocks are interlayered on all scales from fractions of an inch to a hundred feet, and all gradations between the two types are found. In the Grandfather Mountain area, large units of hornblendic rocks structurally underlie the large mica schist and gneiss units, but in the Spruce Pine area (Brobst, 1962) mappable amphibolite units are intimately interlayered with the micaceous rocks. The map pattern and smaller scale field relations in both areas indicate that the hornblendic rocks and the mica schist and gneiss are everywhere conformable. Where amphibolite is interlayered with mica schist and gneiss, it crops out much more commonly than the enclosing micaceous rocks, and the soil is reddish brown and is composed principally of small pieces of weathered amphibolite. The mapping of contacts between predominantly amphibolitic units and predominantly micaceous units is therefore extremely subjective in many areas.

#### BIOTITE-MUSCOVITE SCHIST AND GNEISS

The mica schist and gneiss are gray to light-gray and fine- to coarse-grained rocks containing muscovite flakes ranging from 0.5 to 8mm (millimeters) in diameter. The large muscovite flakes commonly give the rocks a glittery aspect. The rocks commonly contain light-pink to red garnet. Where the schist and gneiss are not interlayered with hornblendic rocks, they form light-colored soil containing flakes of hydromica. Schist and gneiss are gradational and intimately intercalated in layers that are fractions of an inch to a few tens of feet thick. Layers and lenses of amphibolite and a few thin layers of granofels and micaceous quartzite are intercalated with the mica schist and gneiss. The amphibolitic intercalations are most numerous adjacent to mapped bodies of amphibolite.

Parallel arrangement of the micas forms a well-developed foliation which is generally parallel with the compositional layering and wraps around fold noses, although in a few places it cuts through the layering on the noses of isoclinal folds (fig. 5).

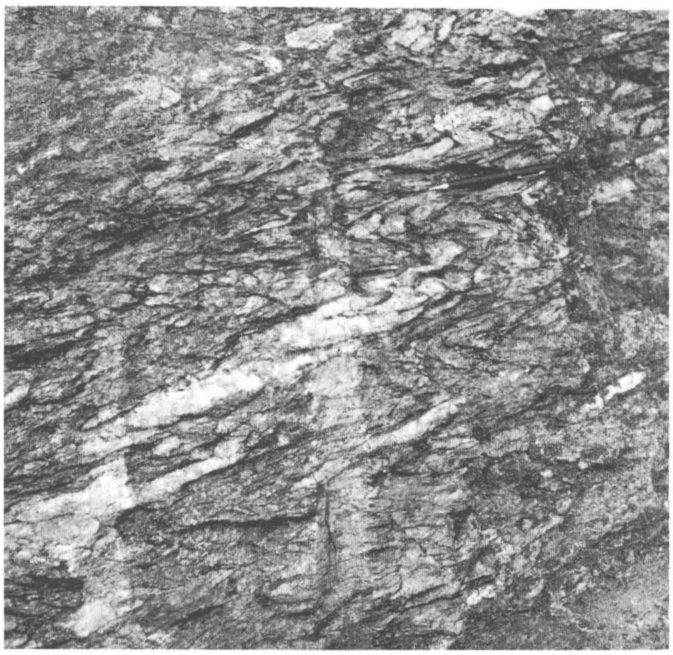


FIGURE 5.—Biotite-muscovite gneiss in the Blue Ridge thrust sheet. Foliation cuts layering in the noses of small folds. Roadcut on Blue Ridge Parkway south of Table Rock No. 2 triangulation station (area C–7, pl. 1).

Segregation knots, lenses, and stringers of quartz are especially numerous in areas J–2 and I–2 (pl. 1) and muscovite-bearing pegmatites are locally abundant in areas C–5 and C–6. Quartz stringers parallel with the foliation are as much as 1.5 feet thick and several tens of feet long. The quartz has a sugary texture that suggests that it has been granulated and recrystallized.

Kyanite occurs in the schist in a few places west of the Grandfather Mountain window. There the trail crosses Elk Ridge in area C–5 (pl. 1), float from a quartz-kyanite segregation contains blades as much as 6 inches long and 3/4 inch wide. Kyanite was also found north of the window in areas G–2 and H–2 (pl. 1), on the north side of Rich Mountain, on Doe Ridge, and on Howard Creek a mile west of the New River.

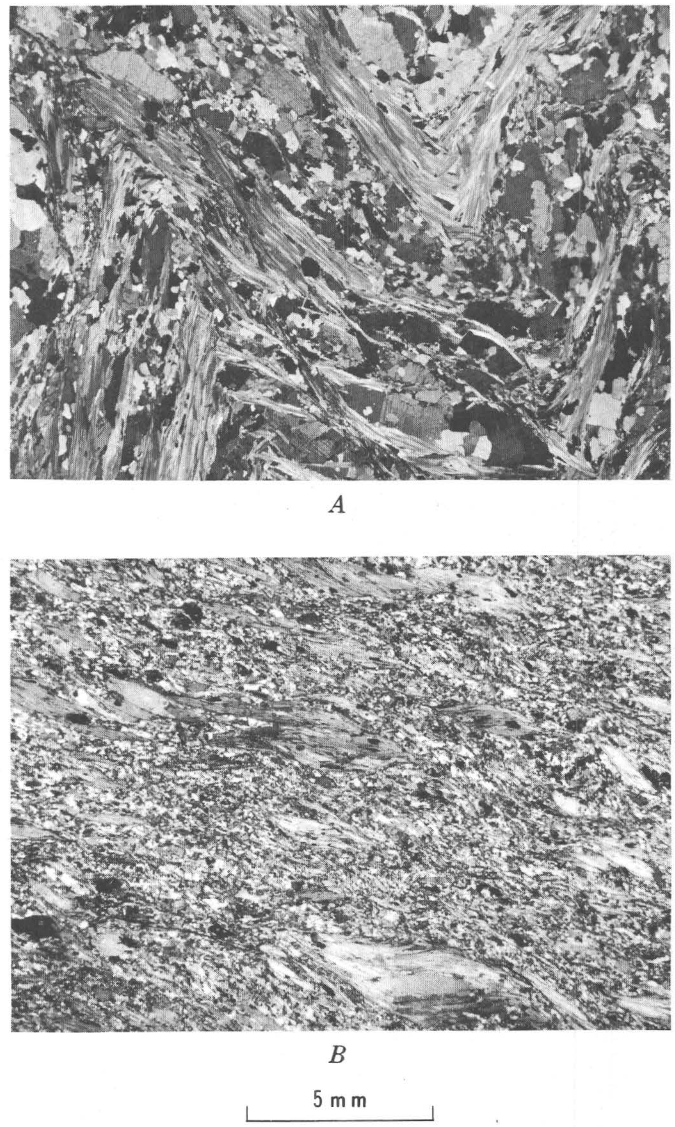



FIGURE 6.—Photomicrographs of biotite-muscovite schist and gneiss southeast of the Grandfather Mountain window. *A*, Garnet- and chlorite-bearing muscovite-quartz-plagioclase schist from southeast of Giles Knob (area J–4, pl. 1). Plagioclase is oligoclase. Microfolds have the same orientation and style as larger folds that are the result of deformation during movement along the Brevard fault zone. Section cut perpendicularly to mineral lineation. *B*, Chlorite- and garnet-bearing biotite-muscovite schist from 1.25 miles northwest of the mouth of White Creek (area E–9, pl. 1). Porphyroclasts of muscovite partly granulated and recrystallized in matrix of muscovite, biotite, plagioclase, and quartz. Plagioclase is partly altered and ranges from $An_{10}$ to $An_{30}$.

Southeast of the Grandfather Mountain window, especially near faults, the larger muscovite crystals are ovoid porphyroclasts resembling fish scales, which are in part converted to finer grained white mica (fig. 6*B*). In such areas, especially in the Linville Falls quadrangle, biotite and garnet are partly converted to chlorite. Northeast of the window, the schist is finer grained where it has been sheared along a few local fault zones.

The mica schist and gneiss generally have granoblastic and lepidoblastic textures, but near faults and shear zones, cataclastic and prophyroclastic textures predominate locally.

Equant grains of quartz average 0.1 to 0.5 mm in diameter but locally reach 4 mm. Muscovite forms well-alined grains ranging from 0.1 to 5mm long and averaging about 2 mm. Where the rock has been crinkled musovite flakes form polygonal arcs around the folds (fig. 6*A*). Locally, crystallization lasted longer than movement, and some flakes have grown transverse to the foliation. In other places, deformation continued or took place after crystallization, and the larger muscovite grains are bent. Adjacent to faults southeast of the window and near the contact with the Cranberry Gneiss west and north of the window, the coarse-grained muscovite has been converted to ovoid porphyroclasts, some of which are bent. Southeast of the window, muscovite porphyroclasts have commonly recrystallized to aggregates of finer grained muscovite.

Biotite flakes are generally smaller and have apparently crystallized later than muscovite. They average about 0.5 mm in diameter, but some are as much as 2.5 mm in diameter. Biotite is synkinematic in some rocks but partly postkinematic in others. It is generally brown or greenish brown; in a few rocks it is brownish green or dark green. In the rocks containing porphyroclastic muscovite, the biotite is all fine grained and seems to have completely recrystallized.

Plagioclase grains average about 0.5 mm in diameter; locally, they range from 0.1 to 3 mm. Some of the larger grains contain inclusions of quartz, mica, and other minerals. Composition ranges from $An_{15}$ to $An_{35}$, except where the plagioclase has been saussuritized. Normal zoning, with as much as 10-percent range in anorthite content, is locally found. Altered plagioclase is found in shear zones and especially adjacent to the contact with the Cranberry Gneiss; locally, it is found southeast of the Grandfather Mountain window.

Garnet occurs as subhedral or euhedral grains and locally as anhedral grains or skeletal crystals as much as 1 cm in diameter. Sieve texture is widespread; a zonal or spiral (snowball) distribution of inclusions of quartz, plagioclase, muscovite, biotite, and opaque minerals is common. Some of the garnet has a light-red absorption; in a few rocks its core has a somewhat stronger absorption.

Anhedral to subhedral sieve-textured kyanite grains, locally as much as 1 cm in diameter, contain inclusions of garnet, biotite, quartz, staurolite, plagioclase, and rutile. Sieve-textured staurolite forms anhedral crystals 0.5 to 8 mm long that commonly contain inclusions of quartz, plagioclase, garnet, and rutile. The kyanite is partly replaced by muscovite in some rocks, and southeast of the window, relict grains of staurolite and kyanite occur in aggregates of sericite. Sillimanite included in a porphyroblast of muscovite was found in one specimen from a roadcut souheast of Mount Perion school (area J–2, pl. 1).

Light-green to green FeMg and MgFe chlorite occurs in small amounts after biotite and garnet.

Epidote occurs as anhedral to subhedral grains 0.05 to 0.5 mm in diameter, that commonly contain cores of allanite or metamict allanite and less commonly cores of zoisite or clinozoisite. Rarely, the epidote is iron rich.

Opaque minerals are magnetite, ilmenite, pyrrhotite, and pyrite and range from 0.02 to 2 mm in grain size.

Apatite and rounded zircon grains are widespread. Sphene, bluish-green to brownish-green tourmaline, rutile, and calcite are less abundant.

### COMPOSITION AND ORIGIN

The biotite-muscovite schist and gneiss were derived from interbedded graywackes and argillites. They generally contain more quartz than plagioclase and subequal amounts of biotite and muscovite (fig. 7). The total mica content ranges from 4 to 66 percent. The occurrence of kyanite and staurolite is not related to a high mica content in all cases. It may be partly controlled by the ratio of alkalis to alumina (see position of analysis 2, table 3, in fig. 7). The only volumetrically important minerals not plotted on the diagram are garnet and epidote. Garnet occurs in more than half of the thin sections examined, but in most of them it constitutes only a few percent of the rock. One specimen, however, contains 16 percent garnet. Epidote constitutes 2 to 16 percent of the rock in about a quarter of the samples.

Some of the chemical analyses of schist (table 3, analyses 1, 2, and 7) resemble those of some Precambrian lutites (Nanz, 1953, analyses 8, 10, and 13) and differ from the average shale (Clarke, 1924) in having higher $Al_2O_3$ and $Na_2O$ and lower CaO contents. Except for its higher iron content, analysis 1

TABLE 3.—*Chemical analyses, modes, and norms of biotite-muscovite schist and gneiss in the Blue Ridge thrust sheet*

[Analyses 1–6, by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey, 1962. Modes, by point counts; P, present but not intersected in counting. Analyses 7–10 are standard rock analyses by Ruth H. Stokes, U.S. Geol. Survey, 1952, furnished through courtesy of Donald A. Brobst. Major oxides and CIPW norms given in weight percent; modes, in volume percent]

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 |
|---|---|---|---|---|---|---|---|---|---|---|
| Field No. | ZD–60–1 | X–67–6 | ZB–54–2 | 4–586 | RE–58–2 | ZB–49–2–a | SP1 | SP9 | SP2 | SP8 |
| Laboratory No. | 160178 | 160181 | 160183 | 160186 | 160179 | 160184 | 51–1758CD | 51–1766CD | 51–1759CD | 51–1765CD |
| $SiO_2$ | 56.0 | 58.3 | 60.3 | 70.6 | 71.8 | 73.5 | 63.14 | 71.13 | 72.11 | 82.46 |
| $Al_2O_3$ | 17.5 | 22.2 | 16.1 | 13.9 | 13.3 | 11.8 | 16.30 | 13.14 | 13.50 | 8.60 |
| $Fe_2O_3$ | 6.0 | 2.1 | 1.5 | .71 | 2.6 | .50 | 2.86 | .98 | 1.77 | .19 |
| FeO | 6.3 | 4.2 | 7.2 | 3.7 | 3.1 | 4.1 | 4.26 | 3.34 | 2.52 | 1.76 |
| MgO | 2.4 | 2.1 | 3.0 | 1.5 | 1.4 | 1.6 | 2.62 | 1.51 | 1.76 | 1.51 |
| CaO | 1.4 | 1.0 | 3.1 | 2.2 | .98 | 1.6 | 1.91 | 1.95 | .99 | .74 |
| $Na_2O$ | 2.0 | 2.2 | 3.3 | 3.3 | 2.0 | 3.0 | 2.30 | 4.32 | 1.87 | 2.64 |
| $K_2O$ | 3.2 | 3.1 | 3.0 | 2.3 | 2.6 | 2.0 | 3.42 | 1.52 | 3.10 | 1.16 |
| $H_2O+$ | 2.2 | 2.0 | 1.1 | .92 | 1.0 | .91 | 1.74 | .61 | 1.26 | .42 |
| $H_2O-$ | .12 | .24 | .08 | .04 | .05 | .06 | .30 | .04 | .10 | .06 |
| $TiO_2$ | 1.8 | .87 | 1.3 | .75 | .86 | .76 | 1.06 | .86 | .70 | .38 |
| $P_2O_5$ | .36 | .39 | .35 | .17 | .17 | .22 | .11 | .22 | .18 | .06 |
| MnO | .23 | .12 | .11 | .10 | .08 | .08 | .15 | .07 | .07 | .03 |
| $CO_2$ | <.05 | <.05 | <.05 | <.05 | <.05 | <.05 | .03 | .02 | .01 | .01 |
| S | | | | | | | | .39 | | |
| Subtotal | | | | | | | | 100.10 | | |
| Less O for S | | | | | | | | .20 | | |
| Total | 100 | 99 | 100 | 100 | 100 | 100 | 100.20 | 99.20 | 99.94 | 100.02 |
| **Modes** | | | | | | | | | | |
| Quartz | 13 | 13 | 28 | 38 | 53 | 60 | | | | |
| Plagioclase | 10 | 17 | 31 | 36 | 10 | 15 | | | | |
| Muscovite | 37 | 41 | | 4 | 23 | 2.5 | | | | |
| Biotite | 12 | | 39 | 20 | 9 | 22 | | | | |
| Garnet | 17 | 8 | .7 | 1.0 | .2 | .5 | | | | |
| Epidote | 1.7 | | | | 2.0 | .5 | | | | |
| Chlorite | 5 | | | | .7 | .3 | | | | |
| Kyanite | | 6 | | | | | | | | |
| Staurolite | | .6 | | | | | | | | |
| Opaque minerals | 5 | .5 | .8 | .2 | 1.2 | .2 | | | | |
| Rutile | | P | | .2 | | | | | | |
| Zircon | | P | P | P | | P | | | | |
| Sphene | .5 | .7 | .8 | .2 | | .2 | | | | |
| Apatite | .1 | .1 | .3 | .3 | .2 | .2 | | | | |
| Tourmaline | P | | | | P | | | | | |
| Calcite | | | | | | P | | | | |
| Points counted | 805 | 1,011 | 606 | 600 | 600 | 603 | | | | |
| **CIPW norms** | | | | | | | | | | |
| Q | 24.72 | 27.31 | 14.95 | 33.80 | 45.49 | 40.48 | 27.27 | 32.62 | 44.20 | 57.90 |
| C | 9.06 | 14.34 | 2.63 | 2.39 | 5.82 | 2.32 | 5.69 | 1.42 | 5.72 | 1.82 |
| Or | 18.91 | 18.32 | 17.72 | 13.59 | 15.36 | 11.82 | 20.21 | 8.98 | 18.32 | 6.85 |
| Ab | 16.91 | 18.61 | 27.91 | 27.91 | 16.91 | 25.37 | 19.45 | 36.54 | 15.82 | 22.33 |
| An | 4.59 | 2.41 | 13.09 | 9.80 | 3.75 | 6.50 | 8.56 | 8.11 | 3.67 | 3.22 |
| En | 5.97 | 5.23 | 7.47 | 3.73 | 3.48 | 3.98 | 6.52 | 3.76 | 4.38 | 3.76 |
| Fs | 4.67 | 4.76 | 10.04 | 5.15 | 2.27 | 6.01 | 3.99 | 4.03 | 2.14 | 2.50 |
| Mt | 8.70 | 3.04 | 2.18 | 1.03 | 3.77 | .72 | 4.15 | 1.42 | 2.67 | .27 |
| Il | 3.42 | 1.65 | 2.47 | 1.42 | 1.63 | 1.44 | 2.01 | 1.63 | 1.33 | .72 |
| Ap | .85 | .92 | .83 | .40 | .40 | .52 | .26 | .52 | .43 | .14 |
| Cc | | | | | | | .07 | .04 | | |

1. Medium-gray schist containing garnets as much as 1 cm in diameter and muscovite as much as 4 mm long. A few white quartz-feldspar segregation lenses, less than 5 mm thick. Muscovite, partly in polygonal arcs and partly deformed. Olive-green biotite, as much as 3 mm long. Quartz and plagioclase ($An_{30}$), about 1 mm in diameter. Garnet includes quartz and opaque mineral and is altered along margins and cracks to sericite and chlorite. FeMg and MgFe chlorite from garnet and biotite. Mode, probably not representative of whole rock because of coarse grain size and irregular distribution of minerals. Probably contains more plagioclase and quartz and less mica. From 1,900-foot altitude on the east side of the 2,100-foot-high knob east of Carr Mountain (area J–5, pl. 1).
2. Light-gray schist containing kyanite and mica as much as 3 mm in diameter. Synkinematic muscovite and brown biotite, as much as 2 mm long. Plagioclase ($An_{20-25}$) in sieve-textured porphyroblasts, as much as 3 mm in diameter, that contain quartz and garnet. Quartz 0.1 to 0.3 mm in diameter. Subhedral garnet 0.15 to 0.3 mm in diameter. Bent and broken sieve-textured porphyroblasts of kyanite as much as 6 mm long are locally altered to sericite; they contain inclusions of quartz, garnet, biotite, and staurolite. Anhedral grains of staurolite, 0.5 to 3 mm in diameter, include quartz and garnet. From roadcut along North Carolina Highway 194, 0.9 mile N. 84° E. of Howard Creek Church (area H–2, pl. 1).
3. Dark-gray gneiss containing garnets as much as 5 mm in diameter and biotite as much as 2 mm in diameter. Quartz and plagioclase ($An_{25}$) grains, 0.2 to 0.7 mm in diameter; reddish-brown biotite 0.2 to 1 mm long. Garnet includes quartz and biotite. From roadcut along Blue

TABLE 3.—*Chemical analyses, modes, and norms of biotite-muscovite schist and gneiss in the Blue Ridge thrust sheet*—Continued.

Ridge Parkway at 3,450-foot altitude, 1,000 feet S. 80° E. of 3,670-foot-high knob southwest of Deep Gap (area J–2, pl. 1).

4. Fine-grained gneiss having granoblastic texture with an average grain size of 0.5 mm and nonoriented micas, 0.2 to 1 mm long. Dark-reddish-brown biotite; plagioclase is $An_{25}$. Subhedral granet, 0.3 mm in diameter. From roadcut along Blue Ridge Parkway 0.37 mile N. 74° E. of Table Rock No. 2 triangulation station (area C–7, pl. 1).
5. Medium-gray schist containing muscovite aggregates as much as 1 cm long. Muscovite aggregates, made up of individual grains 0.5 to 1 mm long, which form polygonal arcs; earlier foliation has been isoclinally folded with axial planes parallel with megascopic foliation of rock. Late kinematic and postkinematic brown biotite, less than 0.5 mm long. Granoblastic-textured quartz and plagioclase ($An_{15}$), 0.1 to 0.3 mm in grain size. MgFe chlorite. (From roadcut along U.S. Highway 321, 0.2 mile north of Stratton Creek (area I–5, pl. 1).
6. Light-gray gneiss containing micas as much as 3 mm long and quartz segregation stringers. Quartz, 0.1 to 0.5 mm in diameter; plagioclase ($An_{20}$), as much as 1.2 mm in diameter; and synkinematic brown biotite and muscovite, as much as 2 mm long. From roadcut along U.S. Highway 421, 0.35 mile northeast of road junction at spot altitude of 3,123 feet (area I–2, pl. 1).

The following descriptions of samples 7 through 10 were furnished by Donald A. Brobst. The rocks are rather typical of much of the Blue Ridge thrust sheet in the Grandfather Mountain area, but satisfactory modes of these specimens are not available.

7. Garnet-muscovite-quartz-biotite-plagioclase schist containing subordinate opaque minerals, epidote, chlorite, and accessory apatite and zircon. From roadcut on U.S. Highway 19E north of Spruce Pine on the south side of the bridge across the North Toe River in the Spruce Pine 7½-minute quadrangle, 0.43 mile S. 50° W. from the point where the Brushy Creek road leaves the west edge of area C–6 (pl. 1).
8. Muscovite-biotite-quartz-plagioclase gneiss containing accessory zircon and sphene. From small creek 1,500 feet N. 5° W. from 3,755-foot altitude on south ridge of Doublehead Mountain in the Carvers Gap 7½-minute quadrangle, about 1.5 miles west of the west edge of area C–5 (pl. 1).
9. Garnet-bearing muscovite-biotite-quartz-plagioclase schist, from roadcut on North Carolina Highway 80 at bridge across Rebels Creek, Bakersville 7½-minute quadrangle, about 10 miles west of Linville Falls quadrangle.
10. Biotite-quartz-plagioclase gneiss containing accessory opaque minerals, chlorite, apatite, and zircon. From roadcut 300 feet north of road junction at 2,867-foot altitude north of Plumtree, Carvers Gap 7½-minute quadrangle, 2,300 feet S. 63° W. from the point where Plumtree Creek road leaves area C–5 (pl. 1).

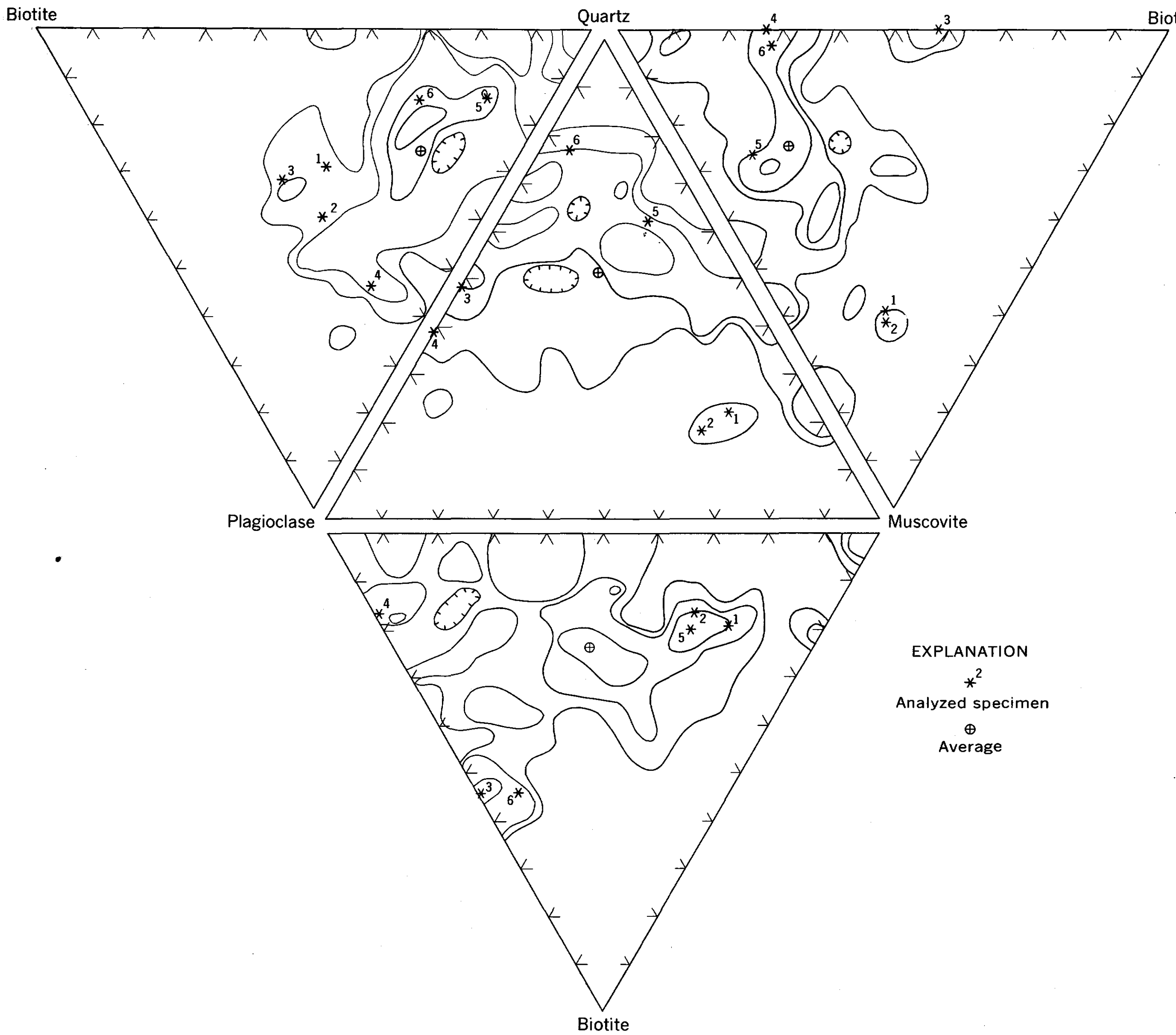



FIGURE 7.—Proportions of quartz, plagioclase, muscovite, and biotite in biotite-muscovite schist and gneiss in the Blue Ridge thrust sheet. Based on counts of 50 random grains in each of 70 thin sections. Contours 1.5, 3, 6, and 9 percent. Number of analyzed specimen refer to analyses in table 3.

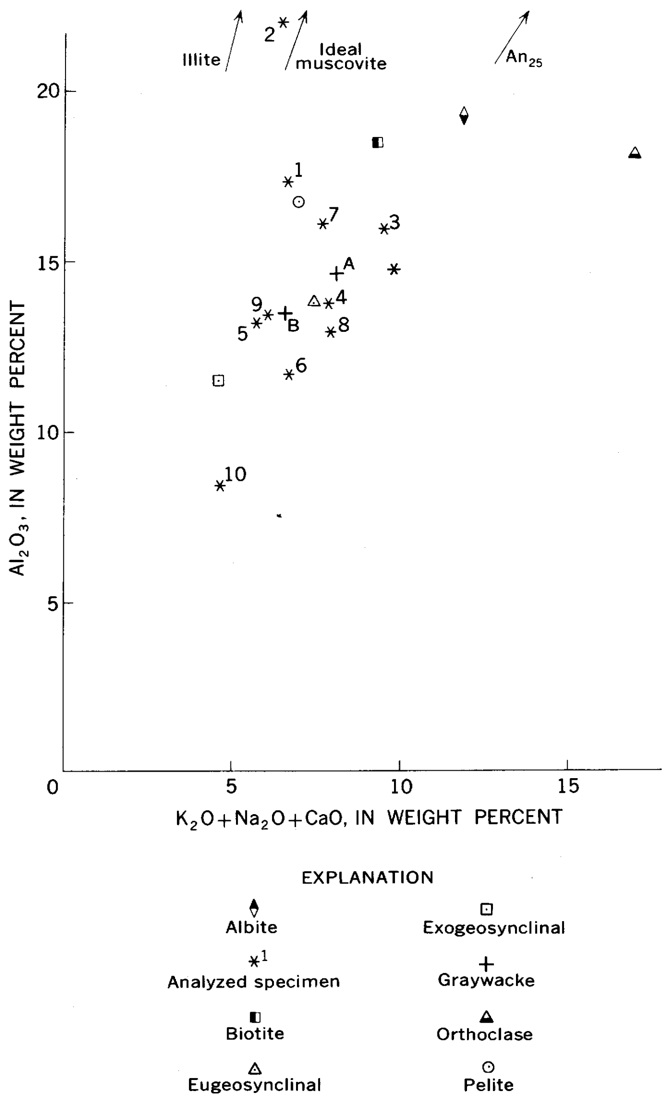



FIGURE 8.—Plot of weight percent $Al_2O_3$ against sum of $K_2O$, $Na_2O$, and CaO for analyzed biotite-muscovite schists and gneisses of the Blue Ridge thrust sheet. Numbers refer to analyses in table 3; unnumbered point, from table 5. Selected mineral compositions and averages of sedimentary rock types are shown. *A*, average analysis of graywacke, from Pettijohn (1957); *B*, average analysis of graywacke, from Pettijohn (1963); average analysis of pelite, from Shaw (1956); average analyses of eugeosynclinal and exogeosynclinal sandstone, from Middleton (1960).

resembles the average pelite of Shaw (1956). In figure 8, analysis 2 falls to the left of the muscovite line despite the plagioclase content of the rock. This position is compatible with the presence of kyanite and staurolite in the rock. The schists generally have more $K_2O$ than $Na_2O$ (fig. 9), whereas the gneisses, which contain less mica and more plagioclase, have less $K_2O$ than $Na_2O$. Some of the gneisses (analyses 4

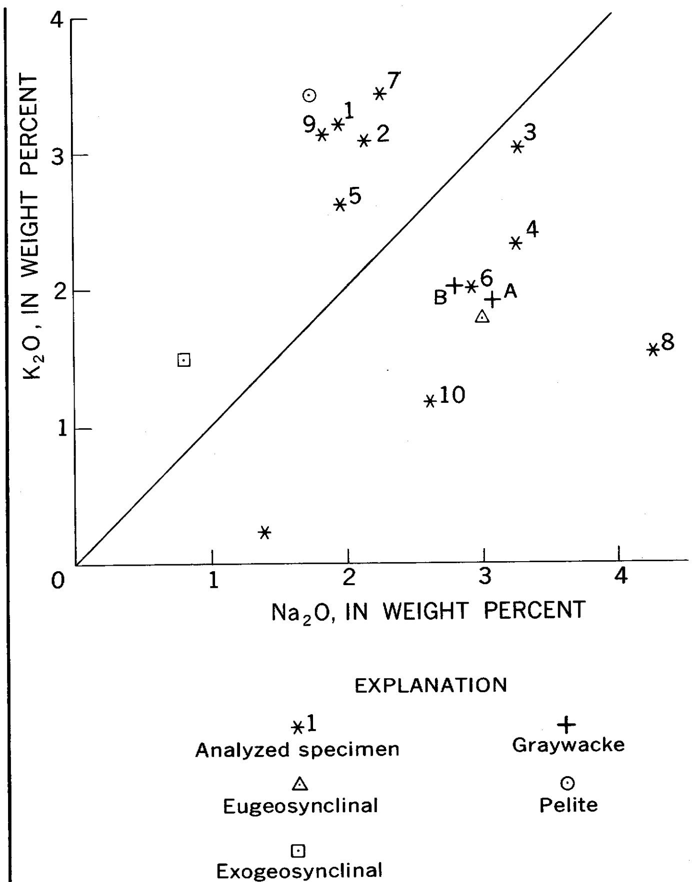



FIGURE 9.—$Na_2O:K_2O$ variation diagram for analyzed biotite-muscovite schists and gneisses in the Blue Ridge thrust sheet. Numbers refer to analyses in table 3; unnumbered point, from table 5. Selected averages of sedimentary rock types shown: pelite (Shaw, 1956), graywackes (Pettijohn, 1957, *A*; 1963, *B*), exogeosynclinal and eugeosynclinal sandstones (Middleton, 1960).

and 6, table 3) have chemical compositions typical of graywackes, but most of the gneisses of the Blue Ridge thrust sheet are richer in $SiO_2$ and $K_2O$ and poorer in CaO than average graywackes (Pettijohn, 1957, 1963; Middleton, 1960). The gneisses are not likely to have been derived from siliceous igneous rocks because they have lower alkali contents in relation to their silica contents than do most igneous rocks. This is reflected in the predominance of quartz over feldspar in their modes.

### AMPHIBOLITE AND HORNBLENDE SCHIST

Amphibolitic rocks interlayered with the mica schist and gneiss in the Blue Ridge thrust sheet are black, greenish black, or mottled black and white and are fine grained to coarse grained. Hornblende crystals generally lie along foliation planes parallel to the layering; they are commonly 1 to 3 mm long and may be as much as 5 mm long. In many rocks their long dimensions are parallel, forming a conspicuous lineation. Pinkish-red garnets, 0.5 to 2 mm in diameter, occur in many layers. Lenses and stringers of granoblastic quartz or plagioclase or both are generally less than an inch thick. The amphibolites contain muscovite-bearing plagioclase-rich pegmatites.

The amphibole is hornblende that has $Z \wedge c$ ranging from 14° to 22° and an absorption parallel to Z which is green to olive green in most rocks but ranges from pale green and bluish green to brownish green and light brown. In places, the hornblende grains contain inclusions of quartz, sphene, plagioclase, rutile, and opaque minerals.

Plagioclase is anhedral and has a grain size of 0.05 to 0.2 mm, except in a few coarse-grained quartz dioritic layers, where it is as much as 5 mm. It is generally calcic oligoclase and sodic andesine and locally shows normal zoning. In some places, especially near the contact with Cranberry Gneiss in areas C–5 and C–6 (pl. 1), it has been saussuritized. However, in these retrograded rocks the mafic minerals are very little altered. Locally, the plagioclase contains inclusions of quartz and hornblende.

Quartz ranges in grain size from 0.05 to 2 mm, and in some places is concentrated in thin segregation stringers. It locally includes hornblende and epidote.

Epidote and clinozoisite are 0.05 to 0.5 mm in grain size and range from anhedral to subhedral. Most of the epidote is rather iron poor; none is pistacite. Some crystals have allanite or zoisite cores. Most of the epidote belongs with the main assemblage, but some is derived by retrogressive alteration of plagioclase and, rarely, hornblende.

Light-pinkish-tan garnet occurs in subhedral to euhedral crystals 0.1 to 5 mm in diameter and locally shows sieve texture with quartz, plagioclase, and opaque minerals.

Diopsidic pyroxene is found in a few amphibolites.

Sphene is the most widespread accessory mineral and occurs as round ellipsoidal grains 0.03 to 0.5 mm long. Apatite is found as ellipsoidal to euhedral crystals 0.03 mm to 1.5 mm long. Rutile locally occurs as cores of sphene grains. Magnetite, zircon, pyrite, brown biotite, chlorite, and sericite are other accessory minerals.

### COMPOSITION AND ORIGIN

Although rocks of this unit vary rather widely in composition, most of them are amphibolites that are composed principally of hornblende and plagioclase and that contain small amounts of quartz and epidote. Some layers are hornblende schists that lack plagioclase and are richer in quartz than the amphibolites.

Chemical analyses of typical amphibolite (table 4, analysis 1) and hornblende schist (table 4, analysis 2) show that the schist contains less CaO, $Al_2O_3$, $K_2O$, MgO, and more $SiO_2$ and total iron than the amphibolite. The amphibolite has a composition similar to basalt, whereas the hornblende schist contains less $Na_20$ and $K_20$ than do most igneous rocks of its silica content, such as the average tholeitic andesite (Nockolds, 1954). Field relations indicate that neither rock can be a flow. The amphibolite was from a

TABLE 4.—*Chemical analyses, modes, and norms of amphibolite in the Blue Ridge thrust sheet*

Analyses 1 and 2, by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey, 1962. Analyses 3 and 4 by standard methods by Ruth H. Stokes, U.S. Geol. Survey, 1952 (Wilcox and Poldervaart, 1 8, table 5). Modes, by point count of 600 grains in each thin section; P, present but not intersected in counting. Major oxides and CIPW norms given in weight percent; modes, in volume percent]

| | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| Field No | 4–550A | FA–34–87 | SP7 | SP6 |
| Laboratory No | 160185 | 160182 | 51–1764CD | 51–1763CD |
| **Major oxides** | | | | |
| $SiO_2$ | 48.5 | 51.8 | 49.58 | 50.28 |
| $Al_2O_3$ | 15.3 | 12.6 | 16.11 | 16.11 |
| $Fe_2O_3$ | 2.5 | 4.7 | 1.33 | 1.46 |
| FeO | 8.8 | 11.3 | 8.42 | 8.65 |
| MgO | 6.5 | 5.2 | 7.42 | 7.76 |
| CaO | 11.4 | 8.9 | 11.47 | 9.99 |
| $Na_2O$ | 2.7 | 2.0 | 2.84 | 2.94 |
| $K_5O$ | .65 | .22 | .42 | .45 |
| $H_2O+$ | 1.0 | 1.3 | .79 | 1.43 |
| $H_2O-$ | .06 | .24 | .03 | .09 |
| $TiO_2$ | 1.9 | 1.5 | .99 | .54 |
| $P_2O_2$ | .28 | .24 | .19 | .18 |
| MnO | .20 | .22 | .17 | .18 |
| $CO_2$ | .10 | <.05 | .23 | .00 |
| S | | | .11 | |
| Subtotal | | | 100.10 | |
| Less O for S | | | .06 | |
| Total | 100 | 100 | 100.04 | 100.06 |
| **Modes** | | | | |
| Hornblende | 63 | 75 | | |
| Quartz | 1.5 | 15 | | |
| Plagioclase | 27 | .8 | | |
| Epidote | 3 | 5 | | |
| Biotite | | .3 | | |
| Garnet | 1.7 | P | | |
| Sphene | 2.8 | P | | |
| Opaque mineral | .2 | 3 | | |
| Apatite | .5 | .2 | | |
| **CIPW norms** | | | | |
| Q | | 10.26 | | |
| Or | 3.84 | 1.30 | 2.48 | 2.66 |
| Ab | 22.83 | 16.91 | 24.02 | 24.86 |
| An | 27.71 | 24.76 | 29.97 | 29.44 |
| Wo | 11.01 | 7.44 | 10.11 | 7.91 |
| En | 11.42 | 12.95 | 10.60 | 12.19 |
| Fs | 7.99 | 14.80 | 7.35 | 8.91 |
| Fo | 3.34 | | 5.52 | 5.00 |
| Fa | 2.57 | | 4.22 | 4.02 |
| Mt | 3.62 | 6.72 | 1.93 | 2.12 |
| Il | 3.61 | 2.85 | 1.88 | 1.03 |
| Ap | .66 | .57 | .45 | .43 |
| Pr | | | .21 | |
| Cc | .23 | | .52 | |

TABLE 4.—*Chemical analyses, modes, and norms of amphibolite in the Blue Ridge thrust sheet*—Continued

1. Amphibolite, in layers 0.5 to 1.5 feet thick in biotite gneiss. Well-alined drab-green hornblende is as much as 2 mm long; quartz and plagioclase ($An_{33}$) are as much as 0.5 mm in diameter. Plagioclase is locally saussuritized. Anhedral garnet as much as 1 mm in diameter includes hornblende, quartz, and plagioclase. From roadcut along Blue Ridge Parkway on east side of Humpback Mountain at contact of Cranberry Gneiss (area C–6, pl. 1).
2. Dark-gray-green schist containing a few thin quartz segregation stringers. Green hornblende, 0.1 to 2 mm long; zoned grains of epidote 0.05 to 0.5 mm in diameter have iron-rich cores; quart grain 0.01 to 0.2 mm diameter are concentrated in segregation stringers. Rock grades to biotite gneiss in outcrop. From roadcut along county road 0.7 mile S. 20° E. of road intersection at spot altitude of 3,123 feet on U.S. Highway 421 (area I–2, pl. 1).

Petrographic descriptions and locations of samples 3 and 4 furnished by Donald A. Brobst. No satisfactory modes of these specimens are available.

3. Garnet-hornblende-plagioclase gneiss containing accessory sphene and biotite. From outcrop 0.6 mile N. 13° E. of Slippery Hill Church (southeast corner of area C–4, pl. 1).
4. Plagioclase amphibolite containing subordinate quartz and sphene. From 200 feet S. 70° W. of summit of Copperas Bald in the Carvers Gap 7½-minute quadrangle. Copperas Bald is 0.4 mile N. 73° W. of Slippery Hill Church (area C–4, pl. 1).

layer less than 1½ feet thick intercalated with biotite gneiss, and the hornblende schist grades into biotite-muscovite gneiss; the amphibolite may have been a mafic tuff or a volcanic-derived sediment. The hornblende schist is too low in $K_2O$ and $Na_2O$ to have been derived from a normal mafic igneous rock; it may have been a dolomitic shale or a mixture of sedimentary and mafic igneous detritus.

The interlayering of the hornblende-bearing rocks with mica schist and gneiss and the intergradation between the rock types suggest that many of the hornblendic rocks must be derived from sedimentary or tuffaceous rocks. However, some of the thicker, more uniform amphibolite layers may have been mafic flows or sills.

## GRANOFELS

Bluish-gray granofels layers superficially resembling quartzite are locally intercalated with schist, gneiss, and amphibolite in the Blue Ridge thrust sheet. The granofelses range from light-yellowish-brown rocks containing as much as 90 percent epidote and 10 percent quartz to gray rocks rich in plagioclase and pyroxene.

Pyroxene occurs in anhedral to subhedral grains 0.02 to 2 mm in diameter. It is light-green salite with $2V$ about 60°, $n_y$ 1.693, and $Z \wedge c = 42°$. Amphiboles in the granofelses tend to be lighter green than in the amphibolites and are probably less aluminous. Epidote is generally rather iron poor.

Plagioclase is generally more calcic than in the enclosing rocks; in one specimen it is labradorite. The one analysis (table 5) can hardly be representative of the wide variety of granofelses, but in general they probably have a higher CaO content in relation to $K_2O$ and $Na_2O$ than do the schist, gneiss, and amphibolite. They were probably derived from calcareous siltstones or sandstones.

TABLE 5.—*Chemical analysis, mode, and norm of granofels from the Blue Ridge thrust sheet*

[Determined by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey, 1962. Mode, by point count of 600 grains; P, present but not intersected in counting. Major oxides and CIPW norm given in weight percent; mode, in volume percent]

| | 1 |
|---|---|
| Field No | N–62–2 |
| Laboratory No | 160180 |
| **Major oxides** | |
| $SiO_2$ | 69.0 |
| $Al_2O_3$ | 14.7 |
| $Fe_2O_3$ | 1.2 |
| FeO | .74 |
| MgO | 1.1 |
| CaO | 8.0 |
| $Na_2O$ | 1.4 |
| $K_2O$ | .23 |
| $H_2O+$ | 1.4 |
| $H_2O-$ | .90 |
| $TiO_2$ | .88 |
| $P_2O_5$ | .26 |
| MnO | .01 |
| $CO_2$ | .05 |
| Total | 100 |
| **Mode** | |
| Quartz | 49 |
| Plagioclase | 32 |
| Epidote | 15 |
| Rutile | P |
| Zircon | .2 |
| Sphene | 1.0 |
| Apatite | .2 |
| Amphibole | 5 |
| **CIPW norm** | |
| Q | 43.03 |
| Or | 1.36 |
| Ab | 11.84 |
| An | 33.15 |
| Wo | 1.82 |
| En | 2.74 |
| Hm | 1.20 |
| Il | 1.58 |
| Tn | .11 |
| Ap | .62 |
| Cc | .11 |

1. Light-gray medium-grained amphibole-epidote-plagioclase-quartz granofels. Mosaic of quartz and plagioclase ($An_{55}$), 0.5 to 0.8 mm in grain size, epidote, 0.2 to 0.5 mm, and light-green amphibole, 0.4 to 0.5 mm. From roadcut on south side of Little Plumtree Creek, 0.85 mile S. 5° E. of Mount Pleasant Church, Linville quadrangle (area C–5, pl. 1).

## AGE

The mica schist, mica gneiss, and amphibolite of the Blue Ridge thrust sheet were considered to be of early Precambrian age by Keith (1903) because of their degree of metamorphism and their stratigraphic position below sedimentary rocks of presumed late Precambrian age. Stose and Stose (1949, p. 315) stated that these rocks are directly traceable into rocks that they mapped as Lynchburg Gneiss of late Precambrian age in the Gossan Lead district, Virginia, 40 miles northeast of the Grandfather Mountain area. However, Dietrich (1959, p. 58) expressed doubt that the rocks in the Gossan Lead district could be correlated with the Lynchburg Gneiss. He found that similar schist and gneiss in Floyd County, Va. are gradational with plutonic rocks of the Blue Ridge complex, which have an age of 1,000 to 1,100 m.y. in northern Virginia (Tilton and others, 1959, table 13). Hornblende at the Ore Knob copper deposit in the Blue Ridge thrust sheet about 10 miles northeast of the Grandfather Mountain area has a potassium-argon age of 1,130 m.y. (Thomas, 1963).

About 60 miles west of the Grandfather Mountain area, schist, gneiss, amphibolite, and migmatite and plutonic rocks are overlain by upper Precambrian rocks of the Ocoee Series (Hadley and Goldsmith, 1963).

Isotopic ages of zircon in layered gneiss at Deyton Bend on the northwest side of the Spruce Pine district are 950 and 1,270 m.y. The layered gneiss at Deyton Bend grades southward into typical schist and amphibolite of the Blue Ridge thrust sheet. Similar ages are found for zircon from layered gneiss from the intermediate sheet in the Mountain City window at Pardee Point (Davis and others, 1962, table 3). This gneiss is directly overlain by the Unicoi Formation of Cambrian (?) age.

These facts favor assignment of the mica schist, mica gneiss, and amphibolite of the Blue thrust sheet to middle Precambrian age in terms of the informal, local usage of the U.S. Geological Survey. The rocks are older than their metamorphism, which was 1,000 to 1,100 m.y. ago, and possibly younger than 1,270 m.y., which is the $Pb^{207}/Pb^{208}$ age of detrital-appearing zircon in the gneiss at Deyton Bend. In terms of present knowledge of the absolute time scale of the Precambrian of North America these rocks would be middle late Precambrian or Neohelikian (Stockwell, 1964, table 2).

Since the above was written, recent work by Rankin (1967, 1970), Hadley (1970), and Hadley and Nelson (1970) has reopened the question of the age of the metamorphic rocks of the Blue Ridge thrust sheet. Near the North Carolina-Virginia State line, Rankin has found good evidence of an unconformable relation between granitic basement rock and rock called the Ashe Formation (Rankin, 1969), which can be traced into the schist and amphibolite of the Grandfather Mountain area. In the Grandfather Mountain area, however, this contact does not appear to be an unconformity, for reasons stated below. Nevertheless, some of the metamorphic rocks in the Blue Ridge thrust sheet in the Grandfather Mountain area may be of late Precambrian age.

## MIXED ROCKS

A diverse group of rocks transitional between predominantly amphibolitic rocks and predominantly granitic Cranberry Gneiss crop out in a broad area west of the Grandfather Mountain window (area C–4, pl. 1) and in a narrow belt north of the window areas (G–2, H–2, I–2, J–2, pl. 1). Elsewhere these rocks are missing. The mixed rocks consist of interlayered and intergrading amphibolite, calc-silicate granofels, biotite-hornblende gneiss, hornblende-epidote-biotite gneiss, biotite-hornblende-plagioclase schist and gneiss, epidote-biotite-plagioclase schist and gneiss, and granitic gneiss ranging from quartz diorite to quartz monzonite. Plagioclase prophyroclasts as much as 2 cm long are widespread in the biotite-hornblende-plagioclase and epidote-biotite schist, and hornblende porphyroclasts as much as 1 cm long occur in some rocks. These rocks are mapped an a unit, the contacts of which are drawn at the first occurrence of layers of granitic rock in the amphibolite on one side, and at the place where granitic layers become dominant on the other side. Many bodies of metamorphosed Bakersville Gabbro occur among the mixed rocks, especially west of Newland in Area C–4 (pl. 1), but only the larger ones are mapped separately. Some may not have been recognized because of the difficulty of distinguishing the amphibolite from younger sheared and recrystallized metagabbro.

Typical exposures of the mixed rocks show well-developed compositional layering and a variety of rock types. In such outcrops, darker layers are amphibolite or biotite-hornblende gneiss, and lighter colored layers are quartz-feldspar gneiss. Locally the layering is sharp, but in other places it becomes diffuse and gradational. The gradation of dark-colored amphibolite to light-colored gneiss commonly takes place through biotite amphibolite, epidote-plagioclase-biotite schist, and epidote-biotite-plagioclase porphyroclast schist and gneiss. In many outcrops the more feldspathic layers contain larger feldspar and hornblende grains than do the amphibolitic layers. Other outcrops show a patchy and irregular distribution of light-colored rock rich in quartz and feldspar in darker rock more rich in biotite and hornblende. In some of the more granitic rocks the mafic minerals occur in irregular clots. Feldspar-rich rock also occurs in lenses, stringers, and crosscutting veinlets.

The unit contains bodies of nonlayered granitic rocks as much as several tens of feet thick. Contacts of these bodies are seldom exposed, but many are apparently gradational with the less granitic country rock.

Light-greenish-gray granofels forms layers ranging from less than an inch to several feet in thickness. The granofels is composed of various proportions of epidote, quartz, and, locally, hornblende. Rare layers of biotite-muscovite schist and gneiss with or without garnet were found in the unit. Granofels layers are well exposed in area C–4 (pl. 1) near Gooseneck Branch above U.S. Highway 19E.

Considerable vein quartz and pegmatite float is found in the outcrop area of the mixed rock, and a few pegmatites crop out. Several of those pegmatites on Bellvue Mountain (area C–4, pl. 1) have been prospected and mined for mica and feldspar.

All the mixed rocks are of medium metamorphic grade and are composed principally of quartz, calcic oligoclase, biotite, and hornblende.

The biotite is generally synkinematic, but crystallization locally outlasted deformation. However, the larger feldspar or hornblende crystals seem to be porphyroclastic rather than porphyroblastic (fig. 10*A* and *B*). Some finer grained, strongly sheared rocks are apparently mesozonal blastomylonites (fig. 10*C* and *E*). An unusual but interesting rock type contains round feldspar and hornblende grains as much as 2 cm in diameter in a fine-grained matrix of biotite, quartz, and plagioclase (fig. 10*C*). It resembles a clastic rock in outcrop and hand specimen, but none of the "clastic" grains are quartz. The most plausible interpretation of the rock is that it is a porphyroclastic blastomylonite.

Other rocks have mortar texture, bent feldspar and biotite, and strained quartz (fig. 10*D*). In some, oligoclase is altered to albite and epidote. These effects are found locally throughout the map unit.

Green and slightly bluish-green hornblende and dark-brown biotite are characteristic of the unit. Garnet has a light-red tinge. Epidote is commonly zoned; cores are allanitic or calcic, and rims are somewhat more iron rich. In most of the more granitic rocks, the potassic feldspar is fine-grained bleb-by-textured perthite. In a few granitic rocks, the potassic feldspar is nonperthitic microcline. Potassic feldspar-bearing rocks contain small amounts of myrmekite. One quartz monzonite contains pale-green diopsidic augite. Sphene, apatite, and magnetite are widespread accessory minerals. Rutile, zircon, pyrite, and ilmenite also occur.

The mixed rocks probably originated through incipient and local feldspathization of rocks similar to the adjacent amphibolite. Subsequent medium-grade metamorphism has destroyed textures attributable to replacement and converted porphyroblasts to porphyroclasts. Strongly developed layering may represent either sheared-out migmatitic layering or modified bedding. The most strikingly layered rocks are the most sheared. Less sheared rocks are generally more granitic and have a migmatitic aspect. This indicates that most, if not all, of the layering has been produced by shearing of migmatitic layering and is not relict bedding. The more sheared and layered parts resemble the more strongly layered parts of the adjacent Cranberry Gneiss, and the more granitic and less strongly layered parts resemble the more granitic parts of the Cranberry. The mixed rocks seem to be a migmatitic gradation zone between Cranberry Gneiss and schist, gneiss, and amphibolite, all of which were subsequently metamorphosed.

#### LAYERED GNEISS SOUTHEAST OF THE GRANDFATHER MOUNTAIN WINDOW

The Blue Ridge thrust sheet southeast of the Grandfather Mountain window is largely composed of interlayered mica gneiss, mica schist, plagioclase porphyroclast gneiss, and granitic rock. These rocks resemble the rocks of the mixed unit west and north of the window, but the unit southeast of the window contains only minor amounts of amphibolite and hornblende gneiss.

On its southeastern side the layered gneiss is in sharp contact with blastomylonite and phyllonitic schist and gneiss in the Brevard fault zone. This contact is exposed in a cut along the Clinchfield Railroad near Hankins (area C–10, pl. 1).

Contacts between the layered gneiss and bodies of mica schist and gneiss in the Blue Ridge thrust sheet southeast of the Grandfather Mountain window are gradational. An impression of the relationships can be obtained in the gullies on the north side of the valley of Licklog Branch (area J–5, pl. 1). There, large amounts of pegmatitic and granitic material in both units occur near the contact.

The rocks of the layered gneiss unit are muscovite-biotite-plagioclase-quartz gneiss and schist, plagioclase porphyroclast gneiss, blastomylonitic quartz dioritic gneiss, garnet-biotite schist, amphibolite, and pegmatite. Layers ranging from fractions of an inch to several inches in thickness are well developed.

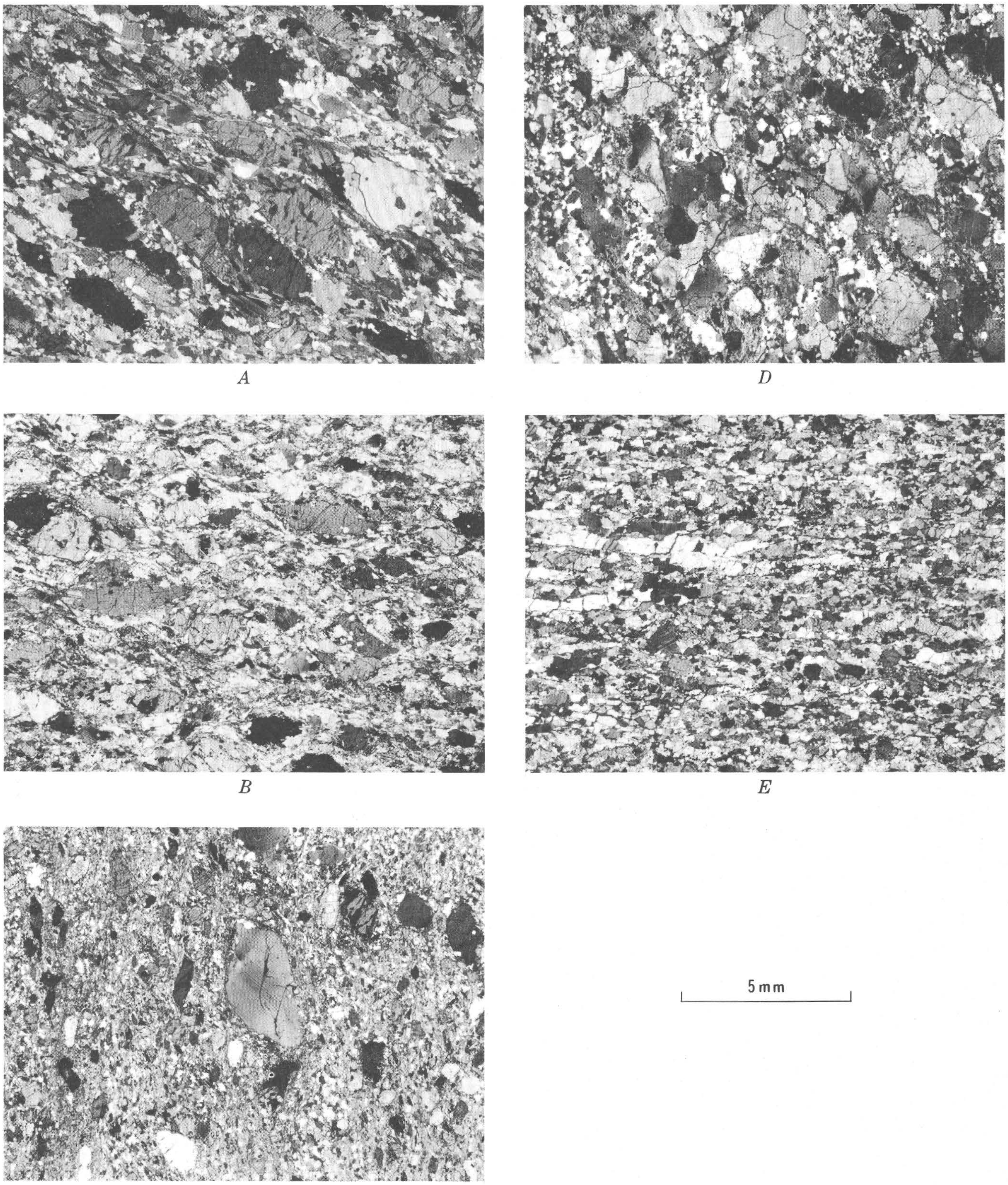

A
D
B
E
C
5 mm

FIGURE 10.—Photomicrographs of polymetamorphic mixed unit rocks. *A*, Hornblende porphyroclast gneiss from roadcut along North Toe River road on sharp curve west of Row Branch, Linville quadrangle (area C–4, pl. 1). Porphyroclasts of hornblende and quartz in a coarsely recrystallized groundmass of oligoclase, biotite, quartz, and epidote. Field relations suggest that this was a migmatitized amphibolite. Present texture is probably due to remetamorphism during the Paleozoic. *B*, Blastomylonitic biotite-hornblende gneiss from 4,000-foot altitude on north side of Bellvue Mountain, Linville quadrangle (area C–4, pl. 1). Porphyroclasts or hornblende and plagioclase in a matrix of recrystallized oligoclase, biotite, quartz, epidote, and hornblende. Outcrop contains gradations between this rock and amphibolite. *C*, Porphyroclastic blastomylonite of medium grade from east side of North Toe River south of Minneapolis, Linville quadrangle (area C–4, pl. 1). Porphyroclasts of hornblende and plagioclase in a groundmass of recrystallized oligoclase, quartz, biotite, and epidote. *D*. Biotite-feldspar-quartz gneiss from east side of the top of Spanish Oak Mountain, Linville quadrangle (area C–4, pl. 1). Grains of strained quartz, perthite with very fine grained blebby texture, and oligoclase surrounded by mortar and recrystallized mortar consisting of quartz, biotite, and feldspar. In the outcrop, layers and lenses of felsic and mafic composition have a patchy and irregular pattern. Section cut perpendicularly to mineral lineation. *E*, Blastomylonite from south ridge of Little Haw Mountain, Linville quadrangle (area C–4, pl. 1). Plagioclase is sodic andesine.

Granitic and pegmatitic material occurs in lenses and pods and locally in layers as much as 15 feet thick. All gradations between granitic rocks and pegmatites occur. Thin granitic layers grade into feldspar porphyroclast gneiss or schist. The coarser grained pegmatites have a grain size of 1 to 2 inches; they commonly contain deformed muscovite books. Accessory garnet is found in the granitic rocks. Pods and lenses of amphibolite a few feet in diameter and layers 1 inch to 10 feet thick occur in most outcrops, but nowhere does amphibolite compose an important part of the unit. One amphibolite pod was found inside a granite-pegmatite layer.

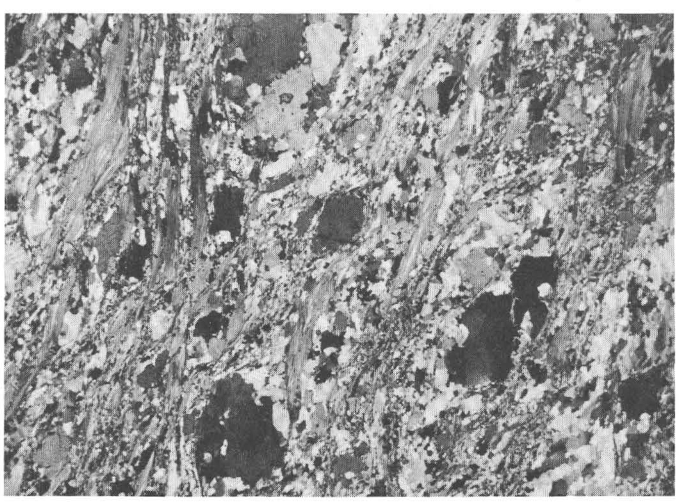



FIGURE 11.—Photomicrograph of biotite-muscovite-plagioclase porphyroclast quartz schist interlayered with gneiss southeast of the Grandfather Mountain window. Specimen from 1,540-foot altitude in Licklog Branch, Blowing Rock quadrangle (area J–5, pl. 1). Mica gneiss, amphibolite, and pegmatite are interlayered in outcrop. Porphyroclasts of sodic oligoclase and muscovite in matrix of biotite, muscovite, sodic oligoclase, and quartz. Some zones of mortar.

Most of the ricks have microtectures indicative of cataclasis and recrystallization (fig. 11). Porphyroclasts of feldspar and muscovite occur in a matrix of recrystallized quartz, muscovite, brown biotite, and plagioclase; the matrix generally has a grain size of 0.05 to 0.2 mm but locally is as coarse as 0.5 mm. A few specimens contain scattered porphyroclasts of biotite that average about 3 mm but that locally are as much as 7 mm in diameter. The amphibolites generally do not have polymetamorphic textures.

Most of the feldspar porphyroclasts are plagioclase, which is locally bent and in some rocks is partly or wholly saussuritized. Locally, the porphyroclasts are saussuritized, whereas in the groundmass, plagioclase is clear. Twinning in the plagioclase is poorly developed in some rocks, but albite and pericline twinning are found in others. A few rocks contain a few porphyroclasts of microcline or perthitic microcline. Occasional layers of quartz monzonitic composition contain numerous porphyroclasts of microcline with coarse-textured string perthite.

Muscovite porphyroclasts typically are wedge or lens shaped. They are commonly bent and in many places partly or wholly granulated and recrystallized into aggregates of finer grained muscovite.

The matrix consists of mosaic-textured quartz and feldspar and lepidoblastic muscovite and brown biotite. Quartz is locally strained. The new micas are synkinematic, and in some rocks crystallization lasted longer than movement.

Epidote makes up as much as 12 percent of the martix in some layers. Generally it is the same size as the other matrix minerals but locally is as much as 1.5 mm long.

Garnet occurs as anhedral to subhedral grains, in places partly altered to biotite, sericite, and chlorite, and as irregular sieve-textured grains containing inclusions of quartz, plagioclase, biotite, and opaque minerals. It is not clear whether the garnets are of the same age as the large mica and feldspar porphyroclasts or whether they are porphyroblasts. The fact that they are about the same size in the sheared rocks, as in rocks lacking cataclastic features, suggests that they might be inherited from the earlier stage in the evolution of the rock. If so, they must have been nearly in equilibrium during the later shearing and recrystallization.

Many of the rocks contain minor amounts of chlorite largely derived from garnet and biotite. It generally is a green FeMg variety, although some MgFe chlorite also occurs. Relict kyanite was found in a sericite aggregate in one rock, and similar sericite aggregates lacking relicts were found in a few other rocks. Other common accessory minerals are magnetite, ilmenite, zircon, sphene, and apatite. Tourmaline and allanite occur rarely.

A pod of calc-silicate rock sampled contains equal amounts of epidote and quartz and minor amounts of sphene, actinolite, and garnet.

The nonmigmatitic types (fig. 12) of layered gneiss contain less muscovite than most rocks of the

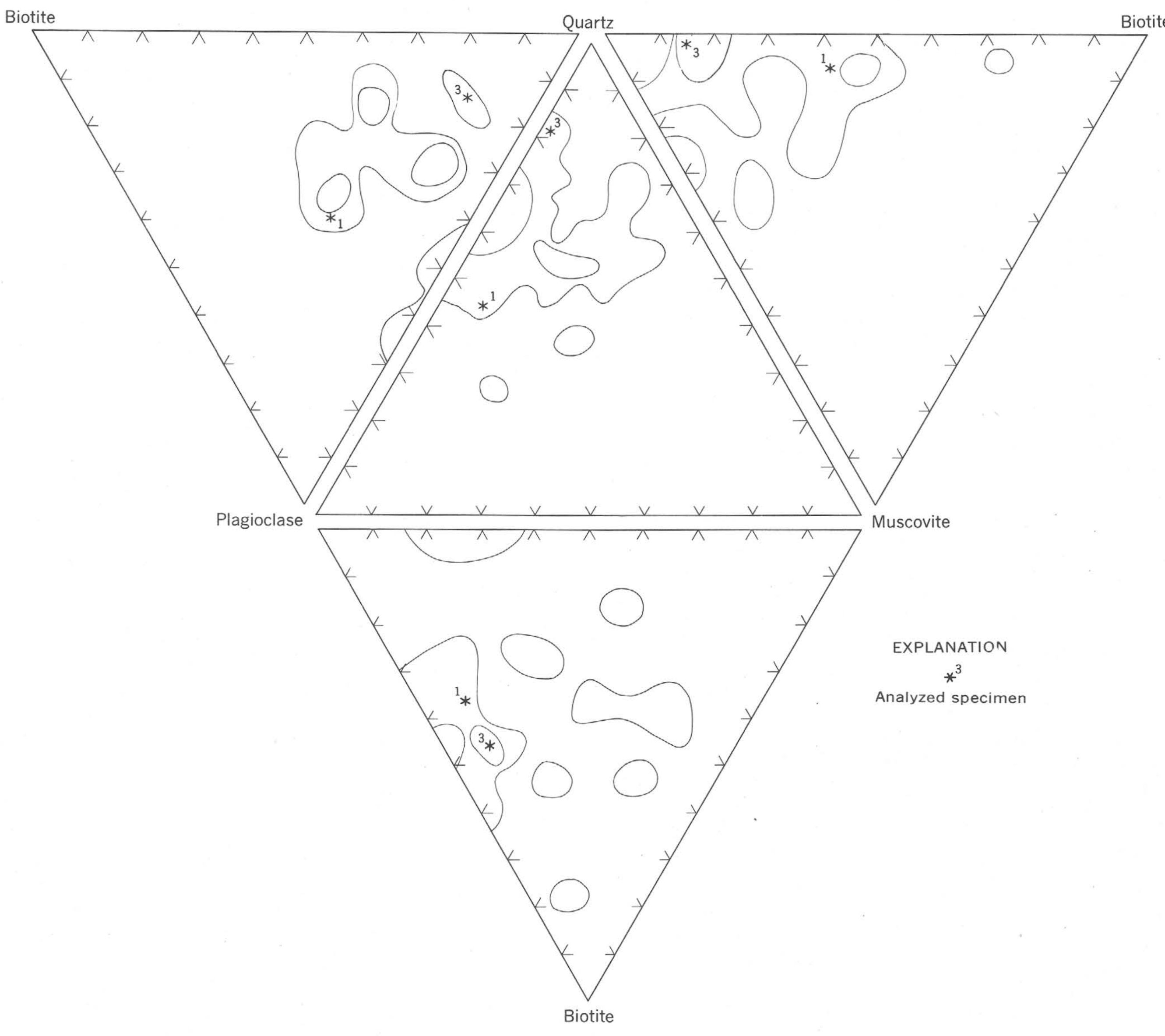



FIGURE 12.—Proportions of quartz, plagioclase, muscovite, and biotite in nonmigmatitic layered gneiss in the Blue Ridge thrust sheet southeast of the Granfather Mountain window. Based on counts of 50 random grains in each of 18 thin sections. Contours 6 and 11 percent. Number of analyzed specimen refers to analysis in table 6.

mica schist and gneiss unit (fig. 7), and their total mica content is somewhat less. However, the relative proportions of quartz and feldspar are similar. Granitic parts of the layered gneiss (fig. 13) contain less mica and more plagioclase than the ordinary gneiss. Most of them are leucocratic quartz diorite, but a few layers and pods of granodiorite and quartz monzonite occur.

Chemical analyses of nonmigmatitic samples (table 6, analyses 1 and 3) suggest that the original rocks ranged from impure quartzites to graywackes. A typical migmatitic rock (table 6, analysis 2) contains more $Na_2O$ and less $Fe_2O_3$ than nonmigmatitic gneiss with comparable $SiO_2$ content (table 6, analysis 1).

The layered gneiss southeast of the Grandfather Mountain window was apparently derived from less argillaceous rocks than the mica schist and gneiss to the west, northwest, and north. In addition, it has been subjected to plutonic metamorphism. Feldspar porphyroblasts (mostly plagioclase) are developed in the gneiss, and this porphyroblast gneiss grades into layers and lenses of leucocratic diorite, granodiorite, or quartz monzonite. These granitic rocks are locally coarse grained enough to be considered pegmatites.

The layered gneiss resembles some parts of the

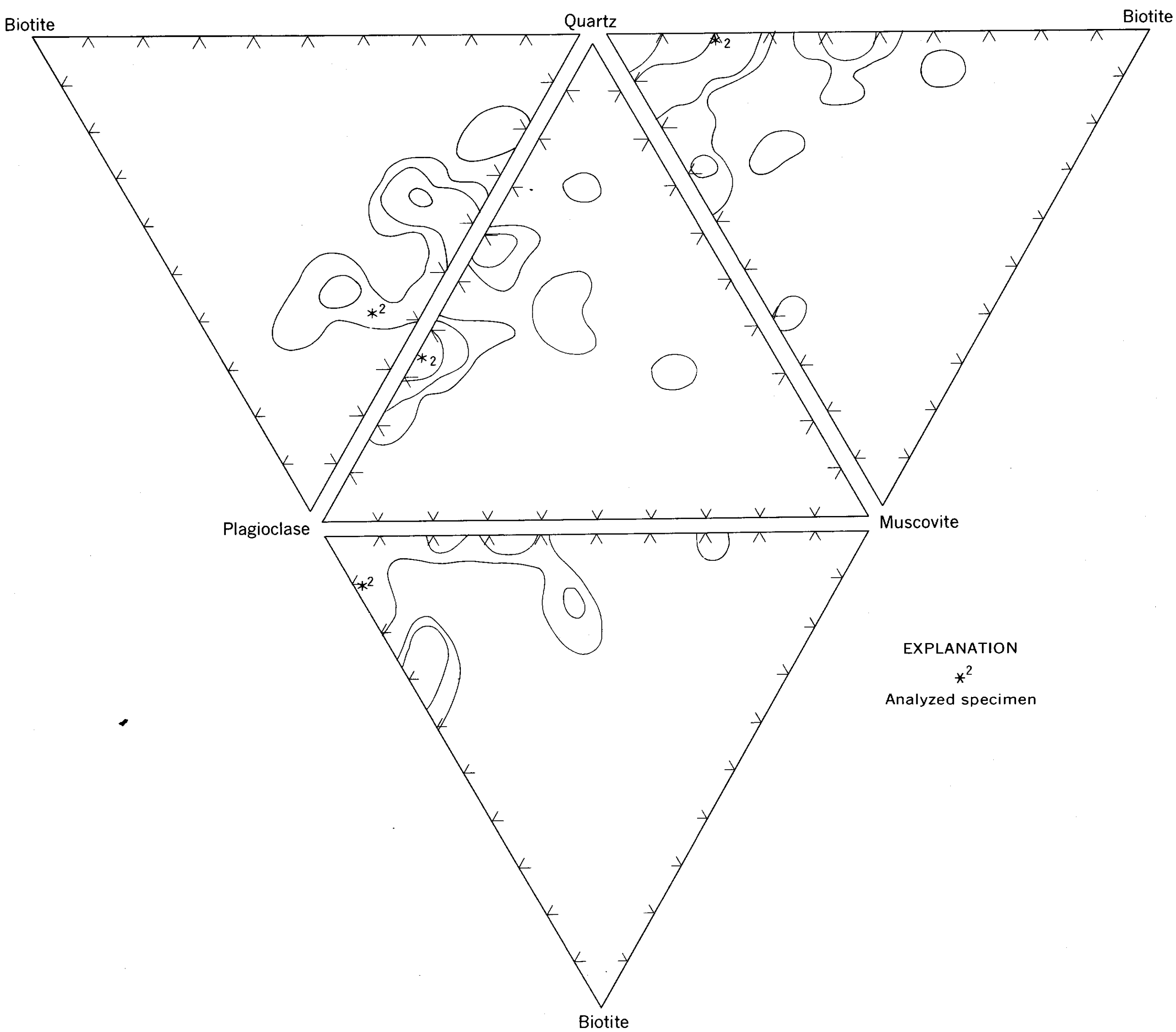



FIGURE 13.—Proportions of quartz, plagioclase, muscovite, and biotite in migmatitic layered gneiss in the Blue Ridge thrust sheet southeast of the Grandfather Mountain window. Based on counts of 50 random grains in each of 21 thin sections. Contours 5, 10, and 20 percent. Number of analyzed specimen refers to analysis in table 6.

TABLE 6.—*Chemical analyses, modes, and norms of layered gneiss in the Blue Ridge thrust sheet southeast of the Grandfather Mountain window*

[Determined by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey, 1962. Modes, by point counts of 600 grains in each thin section; P, present but not intersected in counting. Major oxides and CIPW norms given in weight percent; modes, in volume percent]

| | 1 | 2 | 3 |
|---|---|---|---|
| Field No. | 19–1854 | 19–1855 | 23–2590 |
| Laboratory No. | 160166 | 160167 | 160168 |
| **Major oxides** | | | |
| $SiO_2$ | 64.5 | 65.0 | 80.5 |
| $Al_2O_3$ | 15.3 | 15.2 | 8.7 |
| $Fe_2O_3$ | 3.5 | 1.5 | 1.3 |
| FeO | 2.5 | 3.6 | 1.8 |
| MgO | 1.6 | 1.8 | 1.1 |
| CaO | 3.1 | 3.2 | .76 |
| $Na_2O$ | 3.4 | 5.2 | 1.5 |
| $K_2O$ | 3.0 | 1.2 | 1.3 |
| $H_2O+$ | 1.7 | 1.6 | 1.6 |
| $H_2O-$ | .26 | .14 | .22 |
| $TiO_2$ | .92 | .96 | .55 |
| $P_2O_5$ | .37 | .25 | .06 |
| MnO | .06 | .06 | .02 |
| $CO_2$ | <.05 | .20 | <.05 |
| Total | 100 | 100 | 99 |
| **Modes** | | | |
| Quartz | 29 | 28 | 70 |
| Plagioclase | 31 | 52 | 12 |
| Potassic feldspar | 3 | -- | -- |
| Biotite | 21 | 7 | 12 |
| Muscovite | 4 | .3 | 1.7 |
| Epidote | 9 | 5 | 2.8 |
| Chlorite | -- | 5 | 1.2 |
| Sphene | 1.2 | 1.5 | -- |
| Apatite | .7 | .5 | -- |
| Calcite | -- | .1 | -- |
| Opaque mineral | P | -- | 1.2 |
| Zircon | P | P | P |
| Garnet | -- | -- | P |
| **CIPW norms** | | | |
| Q | 25.15 | 20.12 | 63.09 |
| C | 1.71 | .60 | 3.59 |
| Or | 17.72 | 7.09 | 7.68 |
| Ab | 28.75 | 43.98 | 12.69 |
| An | 12.96 | 12.97 | 3.38 |
| En | 3.98 | 4.48 | 2.74 |
| Fs | .29 | 3.90 | 1.36 |
| Mt | 5.08 | 2.18 | 1.88 |
| Il | 1.75 | 1.82 | 1.04 |
| Ap | .88 | .59 | .14 |
| Cc | -- | .46 | -- |

1. Strongly foliated and thinly layered fine-grained biotite gneiss. Porphyroclasts of plagioclase (altered to albite), as much as 1.5 mm; perthitic microcline, 1 mm; and muscovite, 0.5 mm in grain size in a groundmass of recrystallized quartz, albite, microcline, brown biotite, muscovite, and epidote. Groundmass has a grain size of 0.05 to 0.3 mm. From outcrop on south side of Shooks Creek 0.3 mile west of its junction with the Linville River (area D–9, pl. 1).
2. Fine-grained very finely layered porphyroclastic biotite-muscovite gneiss. Porphyroclasts of plagioclase (altered to albite) as much as 3 mm in diameter and lenses of very strongly strained and incipiently brecciated quartz in a matrix of quartz, albite, brown biotite, epidote, and FeMg chlorite derived from alteration of biotite. Grains in matrix range from 0.05 to 0.3 mm in diameter. Zones of mortar are abundant. From east side of Linville River 0.4 mile N. 25° W. of bridge on North Carolina Highway 126 (area D–9, pl. 1).
3. Medium- to fine-grained porphyroclastic biotite gneiss. Scattered porphyroclasts of partly saussuritized plagioclase (about $An_{20}$) and muscovite as much as 3 mm in diameter in a groundmass of quartz, albite, brown biotite, epidote, and chlorite derived from alteration of garnet and biotite. From roadcut on west side of county road 0.75 miles S. 51° W. of road junction by South Mountain Institute (area D–9, pl. 1).

Cranberry Gneiss described below but contains fewer granitic layers than typical Cranberry.

The layered gneiss grades into biotite-muscovite schist and gneiss and is therefore essentially the same age. The plutonic features probably date from the first metamorphism 1,000 to 1,100 m.y. ago. Later metamorphism destroyed or modified plutonic textures and converted porphyroblasts to porphyroclasts.

## CRANBERRY GNEISS

The Cranberry Gneiss (Keith, 1903; Bryant, 1962) is a metamorphosed plutonic complex that composes a large part of the Blue Ridge thrust sheet in the Grandfather Mountain area and in areas to the northeast and southwest. Although the Cranberry is homogeneous in gross aspect, it is heterogeneous in detail. Keith's (1903) description of the formation in the Cranberry quadrangle is quite accurate, except that he termed it granite rather than gneiss. He recognized the cataclastic textures and structures, the presence of nongranitic layers, and the general increase in the degree of shearing and recrystallization from northwest to southeast. Eckelmann and Kulp (1956) showed that the rocks that Keith (1905) mapped as Henderson Gneiss northwest of the Brevard fault in the Mount Mitchell quardrangle are traceable into the Cranberry Gneiss. Our work confirms their conclusion (Reed, 1964b) but shows that the rocks mapped as Henderson northwest of the Brevard fault are not similar to Henderson Gneiss mapped by Keith in its type area southeast of the fault (Reed and others, 1961). North of the Grandfather Mountain area, Hamilton (1960) attempted to subdivide the Cranberry into informal map units, but we were unable to distinguish most of the lithologic varieties on the map either because of the small size of the individual rock bodies or because of their indistinct boundaries.

### CONTACT RELATIONS

The Cranberry Gneiss lies above the rocks of the Grandfather Mountain window along the Linville Falls fault on the west and north sides of the window. The contact with tectonically overlying amphibolite is gradational through the unit of mixed rocks already described and is nowhere completely exposed.

The sharp contact between Cranberry Gneiss and the overlying biotite-muscovite schist is well exposed along the Blue Ridge Parkway on the east side of Humpback Mountain (area C–6, pl. 1). There, the

Cranberry Gneiss consists of interlayered fine-grained dark-gray layered biotite schist, light-gray quartz diorite gneiss, and a few layers of amphibolite and calc-silicate granofels. Garnet amphibolite and biotite gneiss of the Cranberry are interlayered in 0.5-to 1.5-foot-thick layers through a zone 50 feet thick below the contact. The biotite-muscovite schist contains many sheared and foliated muscovite-bearing, plagioclase-rich pegmatites, some of which have been granulated and drawn out into trains of porphyroclasts. Discordance between the amphibolite and the underlying Cranberry Gneiss indicates at least local movement along the contact.

Amphibolite layers a few tens of feet thick and pods and stringers of pegmatites are widespread along the contact between the Cranberry Gneiss and biotite-muscovite schist in the Linville Falls quadrangle. The pegmatites are locally as much as a hundred feet thick. Near this contact the Cranberry Gneiss locally contains pods of amphibolite and biotite gneiss in which layering is at a large angle to foliation and layering in the surrounding gneiss (fig. 14). Amphibolite layers are also found along the contact east of the belt of mixed rocks in area J–2, plate 1.

### MEGASCOPIC DESCRIPTION

The Cranberry Gneiss consists predominantly of gray to light-gray and locally light-pink or pinkish-green layered and nonlayered granitic gneiss, but contains many dark-green to black nongranitic layers and lenses and scattered bodies of nonlayered hornblendic rock. The granitic parts range in composition from diorite to granite and have an average composition of quartz monzonite or granodiorite. Nongranitic layers are amphibolite, hornblende gneiss, epidote-biotite schist, and epidote-biotite-plagioclase schist and gneiss. Thin layers of calc-silicate granofels and quartzite occur locally. In many places, layering consists of white or pink gneissic quartz monzonite or granite intercalated with gray or green-gray, more biotitic layers of granodiorite or quartz diorite gneiss.

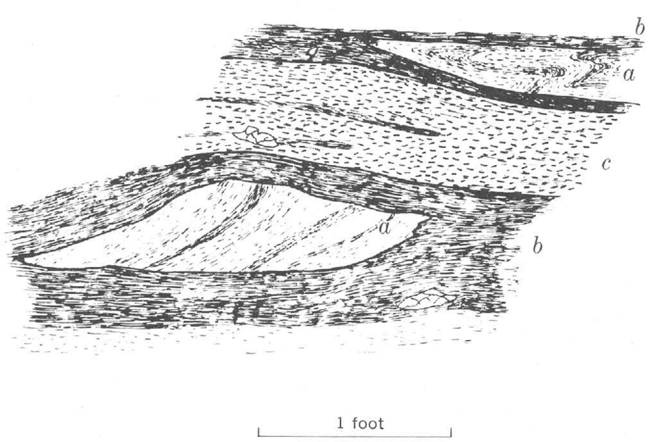



FIGURE 14.—Tectonic lenses of fine-grained biotite gneiss (*a*) in biotite schist (*b*) and biotite amphibole gneiss (*c*). Porphyroclasts of feldspar and small knots of pegmatite are scattered in the enclosing schist. Cranberry Gneiss, just below contact with coarse-grained mica schist and gneiss, 0.9 mile N. 55° W. of Linville Caverns, (area C–6, pl. 1).

Layering is most conspicuous near the contact with the tectonically overlying amphibolite and mica schist. It is least conspicuous northwest of the Grandfather Mountain window in areas C–2 and D–2 (pl. 1). The layering is on all scales from a fraction of an inch to tens of feet (fig. 15). In some exceptionally good exposures, individual layers can be traced continuously for several hundred feet. Layered gneiss locally grades into nonlayered gneiss, and some outcrops have a patchy pattern of granitic and nongranitic material, the latter in rod-shaped bodies elongated parallel to the lineation.

Cataclastic foliation formed by planar orientation of micaceous minerals is generally well developed in the Cranberry Gneiss. Lineation is also conspicuous; it is formed by strung-out aggregates of biotite or grains of feldspar or quartz and alined amphibole needles and biotite flakes. Foliation is parallel with layering in most places, but it cuts the layering on the noses of folds. Obvious cataclastic foliation is lacking only in the vicinity of Little Hump Mountain (area C–3, pl. 1), where the Cranberry Gneiss is of medium metamorphic grade.

The effects of the shearing that produced the foliation are variable; the rocks range from phyllonite and blastomylonite to slightly sheared mortar gneiss. Rocks containing abundant potassic feldspar or hornblende are generally less sheared than plagioclase-rich rocks. Much of the Cranberry is blastomylonitic gneiss containing conspicuous porphyroclasts of white or pink potassic feldspar in a greenish-gray matrix of quartz, feldspar, and mica. Quartz, plagioclase, or biotite also form porphyroclasts in some of the blastomylonitic gneisses. Some less sheared but retrogressively altered granodioritic or quartz monzonitic rocks have a distinctive mottled appearance caused by pink potassic feldspar and green plagioclase. These rocks, which resemble unakite, are most widespread north of Dark Ridge (areas C–2 and D–2, pl. 1).

FIGURE 15.—Conspicuously layered Cranberry Gneiss. Saprolite exposure in roadcut near Hines Gap about 0.5 mile west of Boone City limit, Blowing Rock quadrangle (area G–2, pl. 1). Overlain by typical colluvium.

Blastomylonites are fine grained, generally light-gray rocks and resemble felsic volcanic rocks, especially where they contain scattered feldspar porphyroclasts. Some blastomylonites have been enriched in magnetite but not as much as the phyllonites. Blastomylonites locally grade into unrecrystallized mylonites and breccias.

Silvery-gray, light-green, dark-green, dark-gray, or black phyllonites form distinct zones a few inches to 20 feet thick in many places. In other places, phyllonite grades into the country rock along or across strike. In the phyllonites, porphyroclasts of quartz 1 to 10 mm in diameter are common, and potassic feldspar porphyroclasts are scarce. Segregation lenses and stringers of quartz occur. In many places, the phyllonites have been enriched in magnetite or hematite and have been prospected for iron; a few zones north of Dark Ridge have been prospected for graphite. A phyllonite zone on Big Ridge (area D–2, pl. 1) contains fragments of granitic gneiss as much as 1.5 feet long in a matrix of phyllonite and could be called a pseudoconglomerate.

Dioritic and quartz dioritic rocks in the Cranberry Gneiss commonly have a distinctive gray to greenish-gray color, and some contain amphibole rather than biotite. They contain bodies of pegmatite as much as 3 feet thick. The pegmatite is bluish gray to white and consists of biotite, plagioclase, and quartz or quartz and potassic feldspar.

The Cranberry Gneiss contains white to pink pegmatite lenses, stringers, pods, and a very few dikes ranging from an inch to several feet thick. The pegmatites are sheared, but they are much more weakly foliated than the country rock because of their coarse grain size and the abundance of potassic feldspar and quartz. In the more mafic rocks, the pegmatites contain more plagioclase.

Quartz veinlets are widespread; locally, fluorite and chlorite accompany the quartz. Some veinlets formed before shearing ceased, but others apparently postdate shearing. Epidote forms segregation lenses and veinlets in some of the plagioclase-rich rocks. Seams of magnetite and hematite are found locally.

Coarse-grained granitic gneiss occurs in layers and lenses a few inches to at least a hundred feet thick. Some gneiss contains potassic feldspar porphyroclasts as much as 1 cm in diameter. These potassic varieties resemble the Beech Granite. A few bodies of nonlayered granitic gneiss were distinguished on the maps and described below; other similar bodies may be present, but they are too small or indefinitely defined to be mapped on the scale of pl. 1.

Some of the bodies mapped separately in the Cranberry Gneiss are alkalic granites. Smaller bodies of similar rock are widespread but are not distinguished on the map. The field relations of many of the alkalic rocks are obscure. Most of the alkalic rocks occur in a zone within a mile of the contact between the Cranberry Gneiss and the mixed rocks.

The most widespread lithology among the non-granitic layers is fine-grained, strongly lineated biotite schist in which the micas are about 0.1 mm in diameter. The amphibolite layers in the Cranberry Gneiss generally are sheared, and most of them contain biotite, although biotite is lacking in amphibolite layers where the Cranberry Gneiss is of medium metamorphic grade, as on Little Hump Mountain (area C–3, pl. 1).

In a few places, the Cranberry contains small bodies of nonlayered hornblende-rich rock, some of which were mapped as Linville Metadiabase by Kieth (1903). The best exposure is at 2,900-foot altitude on the Dark Ridge Creek road (near the boundary of areas C–2 and D–2, pl. 1). There the rock consists of various proportions of stubby randomly oriented prisms of hornblende as much as 15 mm long surrounded and cut by quartz-plagioclase veinlets having both sharp and gradational contacts. Contacts of the rock body are gradational and difficult to locate in many places, although locally they are sharp and sheared. The body of hornblendic rock is about 2,000 feet long and 500 feet wide. Smaller bodies of similar hornblende-rich rock were found along a fault zone on the west of the north fork of Left Prong (area C–4, pl. 1) and between the Linville Falls fault and the Beech Granite (area E–3, pl. 1). A body of somewhat less mafic rock is exposed along the Clingman Mine Branch road on Ward's Mountain (area D–2, pl. 1). It also has indistinct contacts. A thin mafic layer in the Cranberry Gneiss is mapped near Blevens Creek (area D–3, pl. 1). Some metamorphosed mafic intrusive rocks are younger than the plutonic metamorphism, but most are of the plutonic generation or older.

On Little Hump Mountain (area C–3, pl. 1) the Cranberry Gneiss is migmatitic. Hornblende, biotite, and garnet occur in clots, stringers, layers, and irregular lenses in granitic rock. Locally, layers of granitic pegmatite, with a grain size as much as 1 inch, grade to more mafic rock types. Gradations from amphibolite to granitic rock through biotite-plagioclase porphyroblast gneiss occur.

### PETROGRAPHY

Typically, the Cranberry Gneiss is an inequigranular blastomylonitic gneiss composed of grains of potassic feldspar, plagioclase, quartz, biotite, and amphibole that are a few millimeters in diameter in a matrix of recrystallized quartz, plagioclase, sericite, biotite, potassic feldspar, and amphibole, the matrix having a grain size of about 0.1 mm (fig. 16).

Blastomylonites have a mosaic texture and a grain size of about 0.1 mm. Phyllonites are very fine grained, and those lacking porphyroclasts cannot be distinguished from phyllites except by their gradational relations with recognizable cataclastic rocks. All gradations between slightly sheared and altered granitic rock through recrystallized mortar gneiss, blastomylonitic gneiss, and phyllonitic gneiss to blastomylonite and phyllonite are found in the Cranberry Gneiss (figs. 16, 17, 18). In a few places, the cataclastic rocks have not been much recrystallized and are mylonitic gneisses and mylonites. The non-sheared rocks generally have a granoblastic texture (fig. 17*A* and *B*).

Potassic feldspar is the most conspicuous porphyroclastic mineral (fig. 16*A*, *B*, *C*). It is generally microcline, but in the more potassic rocks it is patch or vein perthite (fig. 17*D*). The porphyroclasts are commonly bent and broken and healed by recrystallized quartz and albite (figs. 16*D*, 17*D*). Fragments commonly are strung out into pod-shaped aggregates parallel with the northwest-trending lineation. The potassic feldspar porphyroclasts contain inclusions of quartz, plagioclase, and biotite. Potassic feldspar in the recrystallized matrix is microcline.

Plagioclase is practically everywhere altered (fig. 17*C*) and in many places is completely converted to fine-grained albite and epidote. Porphyroclasts are generally saussuritized and show strain shadows and bent twin lamellae. Most of the altered and recrystallized plagioclase is $An_0$ to $An_5$. Unaltered plagioclase, which is rarely found, is oligoclase. Locally, especially near the contacts of the Cranberry Gneiss, the plagioclase is only partly altered and has a composition of about $An_{10}$.

Quartz typically occurs as mosaic-textured aggregates in the groundmass of the gneisses and in places forms porphyroclasts that are very strongly strained and partly brecciated (figs. 17*C*; 18*C*, *D*). In the least sheared rocks, the quartz shows more cataclastic effects than the accompanying feldspar. However, the predominant porphyroclastic mineral in phyllonitic gneisses and phyllonites is quartz (fig. 18*C*); thus, quartz appears to be more resistant when subjected to phyllonite-forming conditions than to when subjected to cataclasis not accompanied by chemical breakdown of feldspars, as in the formation of the blastomylonitic gneisses.

Biotite is found as 0.1-mm grains in the matrix and less commonly as party altered porphyroclasts. It is generally brown or greenish brown and tends to be concentrated in layers and lenses. Locally, the new

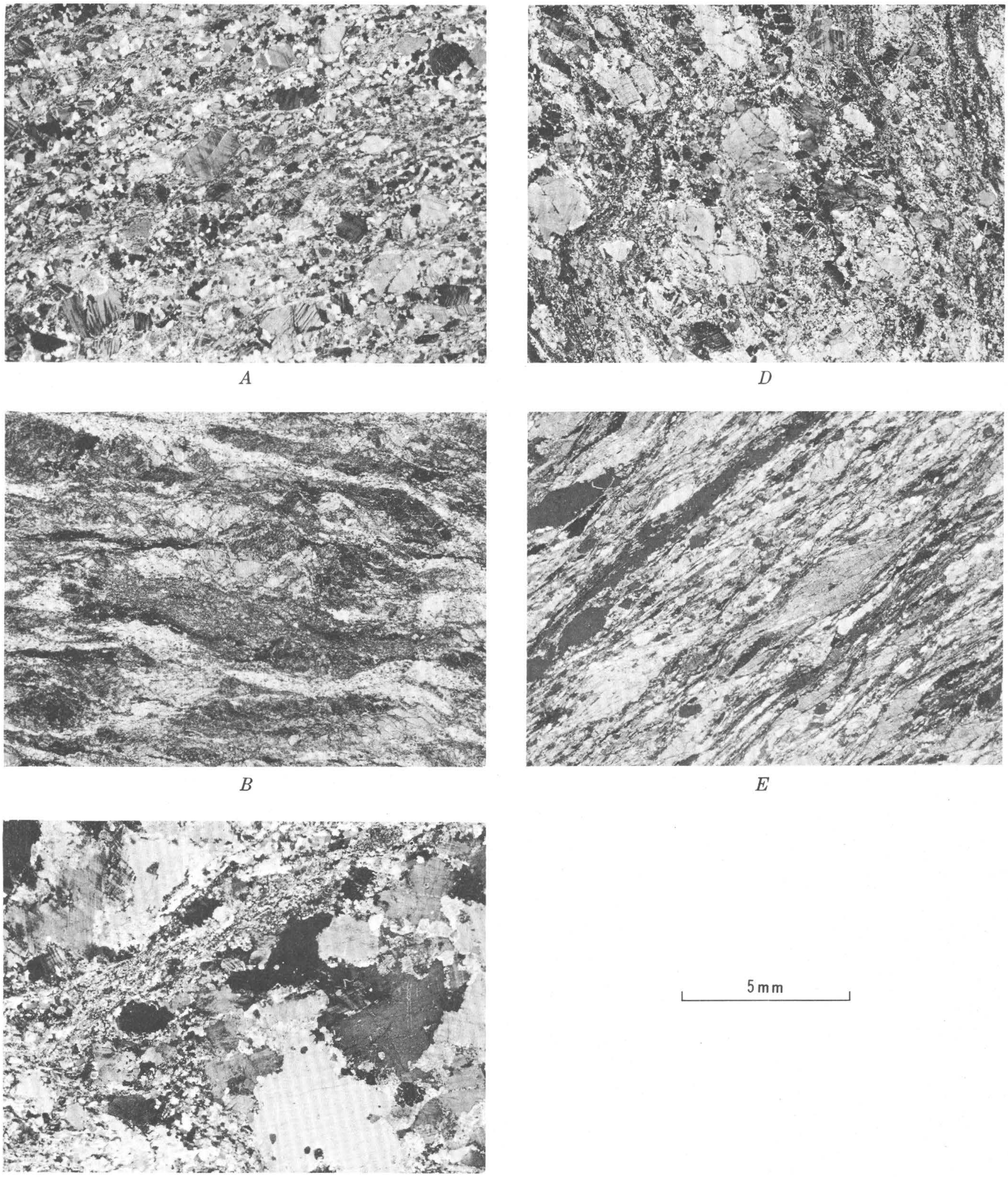

A
D
B
E
C
5 mm

FIGURE 16.—Photomicrographs of typical Cranberry Gneiss. *A*, Blastomylonitic sericite quartz monzonite gneiss from roadcut along Elk Creek just east of Triplett (area J–2, pl. 1). Porphyroclasts of microcline and saussuritized plagioclase in a matrix of recrystallized quartz, iron-rich muscovitic mica, albite, and microcline. Section cut parallel to mineral lineation. *B*, Blastomylonitic granite gneiss from ridge southwest of Smoky Gap (area C–4, pl. 1). Broken porphyroclasts of perthite in a matrix of recrystallized quartz. *C*, Blastomylonitic epidote-muscovite-biotite quartz monzonite gneiss from 3,360-foot altitude in valley south of Cranberry (area C–3, pl. 1). Rock fragments, rich in somewhat perthitic microcline, and porphyroclasts of microcline, and saussuritized plagioclase in a matrix of recrystallized quartz, albite-oligoclase, muscovite, biotite, and epidote. Section cut perpendicular to lineation. *D*, Blastomylonitic granite gneiss from hill south of Cranberry High School (area C–3, pl. 1). Porphyroclasts of microcline, a few small porphyroclasts of muscovite, and saussuritized plagioclase in a matrix of recrystallized quartz, sericite, albite, and epidote. Section cut parallel to mineral lineation. *E*, Biotite-chlorite-epidote-hornblende porphyroclast amphibolite from north ridge of mountain southeast of Cranberry (area C–3, pl. 1). From a layer in more felsic rock. Lenticular porphyroclasts of hornblende in a matrix of partly recrystallized hornblende, sodic oligoclase, clinozoisite, biotite, and chlorite. Section cut parallel to mineral lineation.

biotite reaches 1.5 mm in length. It is generally synkinematic, but locally crystallization lasted longer than deformation. Porphyroclasts are mostly reddish brown and sagenitic. They are commonly bent, and in some rocks they are elliptical in sections perpendicular to the foliation. In places they are partly altered to a green or pale-brown biotite or to chlorite. Porphyroclasts of biotite are most widespread in the quartz dioritic gneisses.

Sericite forms stringers and lenses, and bent porphyroclasts of muscovite are found in a few rocks. Some of the sericite is light green and probably iron rich. Sericite in two samples of phyllonite derived from rocks containing considerable plagioclase was examined by E. J. Young of the U.S. Geological Survey by X-ray diffractometer for the presence of paragonite, but none was detected. Most of the sericite is apparently muscovite.

Amphibole porphyroclasts (fig. 16*E*) have green to brown cores and blue-green or very light green rims that have the same extinction as the cores. The porphyroclasts are commonly lens shaped, and in the more amphibolite-rich rock they are locally surrounded by needles of recrystallized amphibole. Some of the alkalic granitic gneisses contain dark-green hornblende in the recrystallized matrix. In some rocks its extinction angle is as much as 34°, indicating that it may be kataphorite. Crossite was found in one rock which may be a small intrusive.

Chlorite is predominantly the FeMg type, although the MgFe type also occurs. It generally forms pseudomorphs after biotite. In a few of the more mafic rocks, chlorite apparently formed contemporaneously with biotite in the martix. Some rocks lack biotite, probably because of a deficiency of potassium; in these rocks, chlorite is the principal mafic mineral and is locally accompanied by recrystallized amphibole.

Garnet occurs both as euhedral crystals and as trains of fragments partly altered to chlorite, sericite, and biotite. It contains inclusions of quartz, epidote, opaque minerals, and sphene. Some garnet is tannish pink, but some blastomylonites contain a light-yellowish-brown garnet which shows anomalous birefringence.

Acmite occurs in equilbrium with the groundmass of recrystallized quartz, feldspar, and biotite in a layer of alkalic granite gneiss in more plagioclase-rich gneiss on the ridge west of Mast Gap and north of the Watauga River (area F–2, pl. 1). In another somewhat similar alkalic rock from near the head of the southwest fork of Cooper Brach (area C–3, pl. 1), partly altered pyroxene is aegirine-bearing augite with $Z \wedge c$ of 58°. Relict grains of light-green monoclinic pyroxene were found in one rock. Light tan monoclinic pyroxene occurs in the hornblende-rich rock on the Dark Ridge Creek road.

Fluorite occurs in the intergranular area and in veinlets cutting feldspar. It tends to be most abundant in potassic rocks and near zones of phyllonite. Green tourmaline is found in some phyllonites and granite gneisses. Carbonate, rutile, and stilpnomelane are other accessory minerals.

Opaque minerals are widespread and are especially abundant in phyllonite and blastomylonite. Magnetite, commonly in euhedral or subhedral grains, predominates. Other minerals in decreasing order of frequency are ilmenite, pyrrhotite, pyrite, hematite, graphite, and chalcopyrite.

Sphene occurs as round to euhedral grains and commonly rims opaque minerals. Zircon is generally in rounded grains, but is occasionally subhedral to euhedral. Some prismatic grains have rounded terminations. A few crystals are zoned and have overgrowths. Apatite ranges from round to euhedral. Epidote forms equant subhedral grains or very tiny

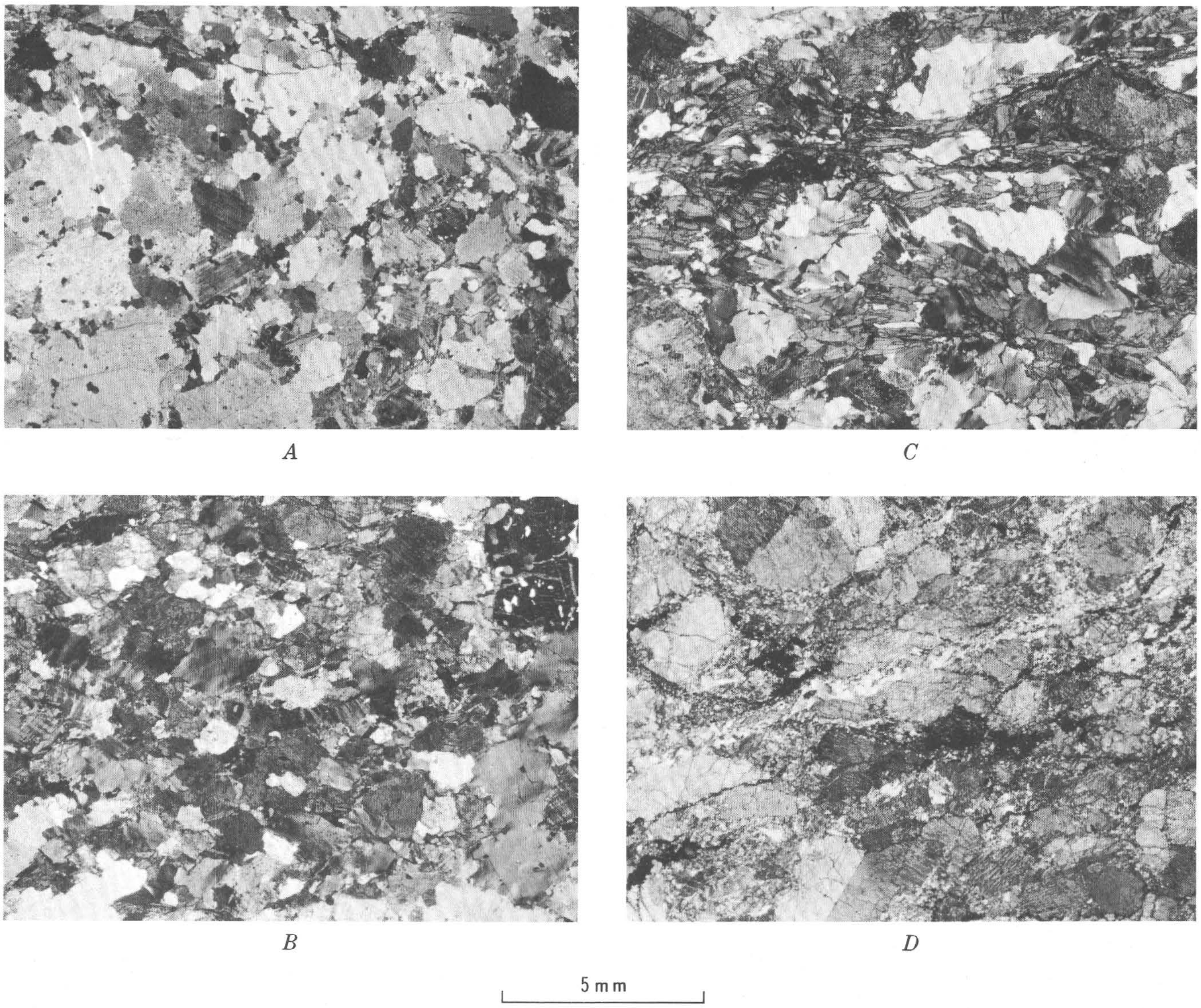



FIGURE 17.—Photomicrographs of less sheared Cranberry Gneiss. *A*, Biotite quartz diorite from boudin 2 feet thick in porphyroclastic biotite and amphibole gneiss. From roadcut on Blue Ridge Parkway on northeast side of Humpback Mountain (area C–6, pl. 1). Plagioclase, altered to albite. Quartz and plagioclase, somewhat bent. Coarse-grained granoblastic texture. *B*, Garnet-bearing quartz monzonite from roadcut in Dark Ridge Creek road near Tennessee-North Carolina State line (area C–2, pl. 1). Strained quartz and microcline. Oligoclase, locally altered to albite. Large garnet in upper right corner. Granoblastic-textured rock has been squeezed and slightly altered. *C*, Cataclastic biotite quartz diorite gneiss from ridge south of Double Knobs (area D–3, pl. 1). Strained quartz, saussuritized plagioclase, and bent biotite. *D*, Partly recrystallized mortar granite gneiss from just east of fault on hill north of White Oak Creek (area C–3, pl. 1). Vein perthite is broken and healed by fine-grained mosaic-textured quartz, epidote, magnetite, and sphene. Matrix also contains perthite fragments and some recrystallized albite.

grains still included in plagioclase. The iron content and zoning are variable. Allanite is rimmed by epidote in some rocks and is locally metamict.

The Cranberry Gneiss on Little Hump Mountain (area C–3, pl. 1) ranges in texture from granoblastic (fig. 19) to partly cataclastic and has apparently been recrystallized under medium-grade conditions. Perthitic potassic feldspar in this rock contains very thin elongate blebs of albite and is similar to the potassic feldspar in rocks of the mixed unit. Potassic feldspar replaces plagioclase, and myrmekite is found in adjacent plagioclase grains. Plagioclase

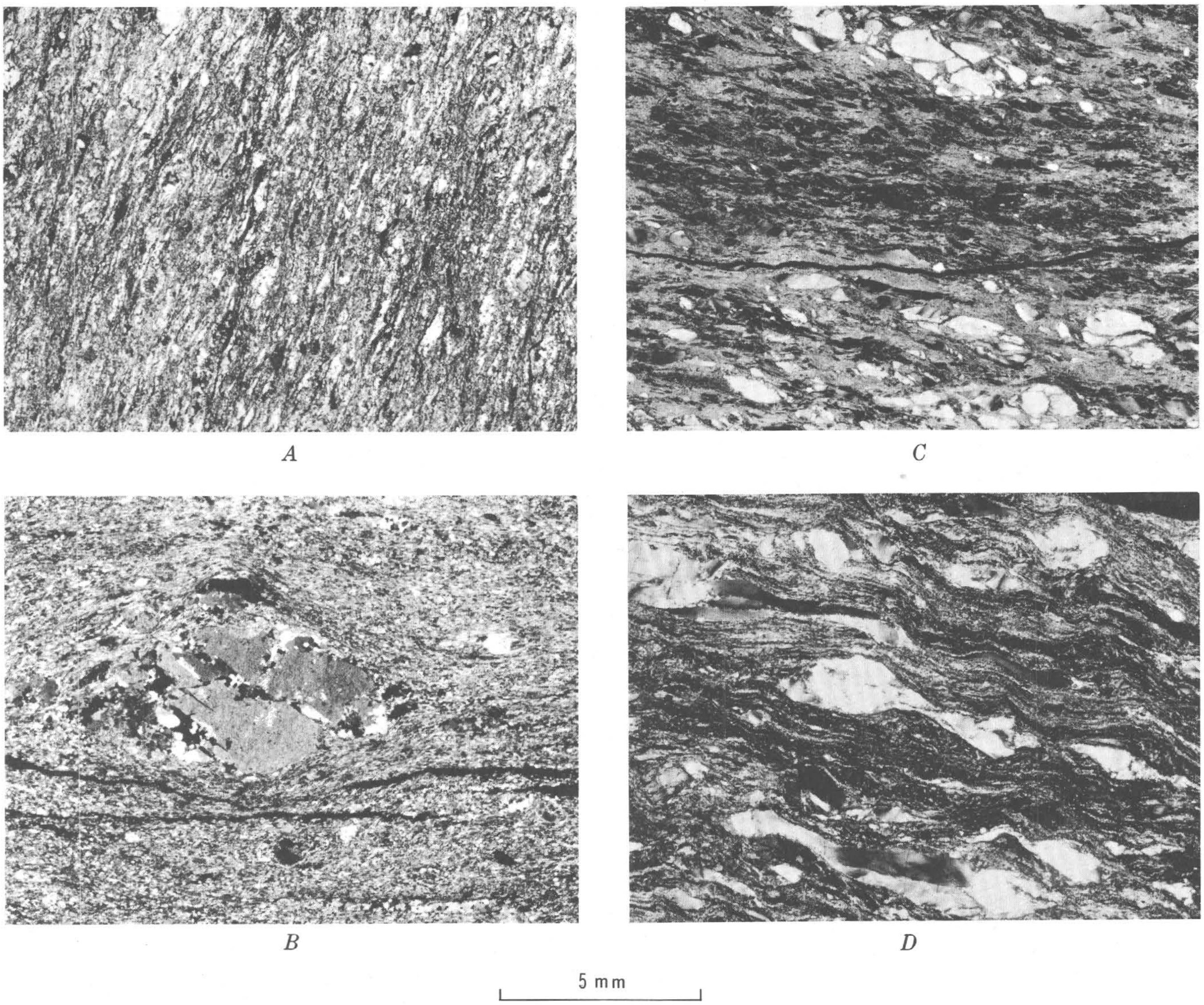



FIGURE 18.—Photomicrographs of blastomylonite and phyllonite in Cranberry Gneiss. *A*, Porphyroclastic blastomylonite from 3,390-foot altitude on south fork of Cooper Branch (area C–3, pl. 1). Scattered small porphyroclasts of microcline and biotite in a matrix of recrystallized quartz, sericite, albite, epidote, and biotite. Section cut nearly perpendicularly to mineral lineation. *B*, Porphyroclastic blastomylonite from roadcut on Blue Ridge Parkway west side of bridge across Linville River (area C–6, pl. 1). Large porphyroclast of saussuritized plagioclase (broken and healed by quartz), smaller porphyroclasts of saussuritized plagioclase and sagenitic biotite in a matrix of recrystallized quartz, albite, biotite, sericite, and epidote. Section cut parallel to mineral lineation. *C*, Chlorite-magnetite-sericite phyllonite containing quartz porphyroclasts from ridge northeast of head of Cooper Branch (area C–3, pl. 1). Probably derived from quartz diorite. *D*, Mylonitic gneiss from along the Stone Mountain fault at 3,740-foot altitude on south side of gully at head of Dark Ridge Creek (area D–2, pl. 1). Very elongate porphyroclasts of quartz and smaller porphyroclasts of biotite and hornblende in a mylonitic matrix of quartz, feldspar, sericite, and epidote. Section cut parallel to mineral lineation.

ranges from $An_{20}$ to $An_{32}$. One rock (fig. 19) contains hypersthene and a small amount of light-green monoclinic pyroxene that is partly converted to green hornblende. The occurrence of hypersthene suggests that at one stage in its history the Cranberry Gneiss may have had charnocitic affinities. In the pyroxene-bearing rock, aggregates of biotite and hornblende surround garnet and magnetite grains,

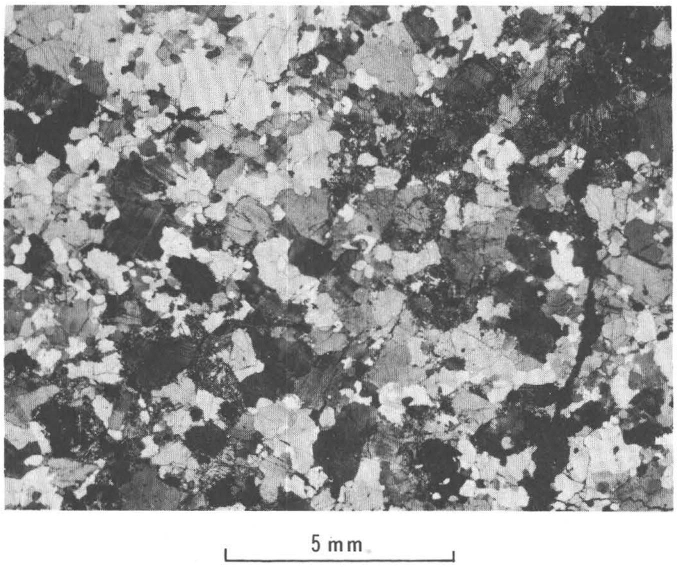



FIGURE 19.—Photomicrograph of slightly sheared Cranberry Gneiss recrystallized at medium grade. Hypersthene and potassic feldspar-bearing biotite quartz diorite in amphibolitic migmatite from 4,740-foot altitude on east ridge of Hump Mountain (area C–3, pl. 1). Granoblastic texture. Clots of fine-grained secondary biotite and amphibole in many places surround larger grains of garnet or magnetite. Amphibole rims pyroxene. Plagioclase is calcic oligoclase.

and fine-grained late biotite is found in other rocks in the Hump Mountain area. One rock contains some skeletal porphyroblasts of scapolite.

## COMPOSITION

Table 7 shows the average composition of some rock types in the Cranberry Gneiss. The rather small number of samples of quartz monzonite and granodiorite is due to sampling bias. Nongranitic layers and granites were oversampled, and typical granitic gneiss was undersampled. Within each type of granitic rocks, the content of mafic minerals is quite variable, although the averages show the usual tendency for diorites to be more mafic than granites.

Amphibolite and biotite schist were selected as typical end-member types of nongranitic layers in the Cranberry Gneiss. Most of the 56 samples of nongranitic layers that were petrographically studied fall between the two rock types. They generally contain more plagioclase than quartz, but gradations to more quartzitic layers do occur. Four samples contain more than 50 percent quartz accompained by various amounts of garnet, calcite, amphibole, plagioclase, and epidote.

Rocks classified as blastomylonite were derived from the various types of granitic rock. Some may have been silicified. They locally contain as much as 10 percent magnetite. Part of the iron for the magnetite may have been derived by destruction of biotite, which is lacking in these rocks, some iron may have been introduced.

Similarly the phyllonites were derived from various original rock types. In these rocks the feldspar has been hydrolized, and sericite and quartz have formed (Bryant, 1966). A significant increase in sericite is apparent in the average for this rock type in relation to the granitic rocks from which it was derived. The average does not show an increase in quartz content in relation to the granitic gneisses, although some samples contain as much as 50 percent quartz. The extra quartz generated by the reac-

TABLE 7.—*Average modes, in volume percent, of various rock types in the Cranberry Gneiss*

[Average of the number of 50-grain counts indicated in parentheses at head of column. P, present in half or more of samples. Difference between sum and 100 made up by accessory minerals]

| Mineral | Rock type | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|
| | Granite (27) | Quartz monzonite (16) | Gran-odiorite (19) | Quartz diorite (51) | Diorite (8) | Blasto-mylonite (19) | Phyllonite (27) | Amphibolite (14) | Biotite schist (8) |
| Quartz | 31 | 29 | 27 | 26 | 4 | 34 | 28 | 3 | 8 |
| Potassic feldspar | 47 | 24 | 13 | 2 | | 21 | 1 | | |
| Plagioclase | 10 | 29 | 41 | 49 | 64 | 12 | 1 | 22 | 15 |
| Biotite | 3 | 5 | 6 | 10 | 10 | 6 | 1 | 9 | 51 |
| Muscovite and sericite | 5 | 5 | 5 | 5 | 1 | 13 | 57 | 2 | |
| Epidote | 3 | 4 | 3 | 3 | 9 | 5 | P | 6 | 16 |
| Amphibole | P | 2 | 1 | 2 | 6 | | | 52 | 6 |
| Opaque minerals | P | P | P | P | P | 2 | 6 | 1 | 1 |
| Zircon | P | P | P | | | P | P | | |
| Sphene | P | | P | P | P | 1 | P | 3 | 2 |
| Apatite | | P | P | P | P | | | | |
| Chlorite | | | | | 2 | | 1 | 1 | 1 |
| Carbonate | | | | 1 | | | | | |
| Quartz and feldspar undifferentiated | | | | | | 5 | 1 | | |
| Fluorite | | | | | | | 1 | | |

tion was in part deposited in quartz segregation lenses, which were avoided during sampling of the phyllonites. Some quartz from the reaction may have contributed to silicification of the blastomylonite zones. Epidote minerals, which might be expected from alteration of plagioclase, are rare. Opaque mineral content of the specimens studied reaches a maximum of 32 percent. Much of it is magnetite, but hematite, ilmenite, sphalerite, and pyrite also occur. Redistribution of elements apparently was widespread during the formation of the phyllonites. Mafic silicates, principally biotite, were destroyed and could have furnished a source for iron, which was locally concentrated in the phyllonite zones. In some zones, the mica is green muscovite rich in ferric iron, which could have developed by a coupled reaction between oxidizing aqueous solutions and hydrolytic decomposition of feldspar. In relation to the granitic rock from which it was derived, the typical quartz-sericite phyllonite has lost calcium and sodium and gained potassium (Bryant, 1966).

### ORIGIN

The Cranberry Gneiss is a plutonic complex of the catazone as described by Buddington (1959). He especially pointed out the uncertainty of the origin and mode of emplacement of catazonal plutons (p. 714–715).

All compositional gradations between amphibolite, epidote-biotite schist, and granite are found in the Cranberry Gneiss. In the main body of the Cranberry Gneiss in the Grandfather Mountain area, shearing and partial recrystallization of low metamorphic grade have obscured original textural relations and have even modified some of the small-scale field relations indicative of origin. The widespread compositional layering, the occurrence of nongranitic layers in the gneiss, the erratic distribution of rock types, and the lack of well-defined contacts suggest that the Cranberry Gneiss is a complex of migmatite possibly containing melted or homogenized parts and small intrusive bodies.

Keith (1903) thought that some of the layers in the Cranberry Gneiss were inclusions in an intrusive mass and that some were the result of metamorphic differentiation, but stated that "the prevalent metamorphism * * * and the heavy forest cover make it difficult to obtain precise evidence of eruptive contacts with adjoining formations."

The possibility that some of the layers were produced by metamorphic differentiation during the low-grade cataclastic metamorphism cannot be dismissed. Indeed, more and less phyllonitic gneisses were produced during shearing, and some material was mobile at that time. However, the porphyroclasts of different minerals occur in differing proportions in the various layers, which indicates that most of the layering is inherited form the original plutonic complex.

Eckelmann and Kulp (1956) believed that the Cranberry Gneiss was a conformable sequence below the mica schist and amphibolite of the Spruce Pine district and that this underlying sequence had undergone metamorphism and varying degrees of feldspathization. They suggested that all the layering was inherited from sedimentary bedding. They obtained rounded zircons from the layered granitic rocks and euhedral ones from some of the massive coarse-grained nonlayered rocks. Their interpretation of the origin of the Cranberry Gneiss differs from ours in that we would emphasize plutonic processes, such as feldspathization, palingenesis, and anatexis and deemphasize the stratigraphic relations. We believe that the coarse-grained quartz-feldspar layers were produced by lit-par-lit granitization of schist, gneiss, and amphibolite, perhaps similar to the overlying rocks, rather than formed from beds of coarse-grained rocks of granitic composition. Amphibolite layers tended to be resistant and survived where granitization was rather weak. They were successively converted to epidote-biotite schist, epidote-biotite-plagioclase porphyroclast schist and gneiss, and dioritic and quartz dioritic layers with increasing feldspathization. At the same time that schist and gneiss layers were converted to rocks of granodioritic and quartz monzonitic composition, the layers were emphasized by shearing in many places. Locally, especially where granitic layers make up a small proportion of the rock or where the shearing was less intense, the pattern of granitic and nongranitic parts is more irregular and patchy.

The bodies of nonlayered hornblende-rich rock with indistinct boundaries transected by quartz-feldspar veinlets may have been derived from Precambrian ultramafic bodies to which silica and alkalies were added during the plutonic metamorphism. Amphibolite may form by reaction of serpentine with igneous gabbro and hornblende diorite (Cater and Wells, 1953, p. 102). Hornblendites, gabbro, and diorite in many respects similar to older mafic bodies in the Cranberry Gneiss have formed by granitization of ultramafic rocks in the Northern Cascade Mountains of Washington State (Crowder, 1959, p. 864, pl. 4).

Some of the more uniform granitic rock may have been mobilized and intruded during a late stage of the plutonic metamorphism. Certainly, the zircon data of Eckelmann and Kulp (1956, p. 314–315) support this. Subsequent shearing and metamorphism during Paleozoic time has obscured many of the original relations.

The contact between the Cranberry Gneiss and the overlying mica schist is discordant with amphibolite bodies and structural trends in the schist and gneiss both north and west of the window (pl. 1 and fig. 32). This discordance shows that there must be at least an unconformity between the Cranberry Gneiss and the overlying rocks. If the overlying rocks are younger, they must have been deposited against a basin margin having about 2 miles of relief in a distance of 6 miles to account for the apparent discordance, but the absence of conglomerates in the mica schist and gneiss seems to rule out this possibility.

It seems unlikely that all the layers and bodies of coarse-grained feldspar-rich rocks that had granitic textures before the Paleozoic metamorphism were sedimentary rocks, for we find no gneisses of similar composition with relict sedimentary textures. The development of the granitic texture and composition seem related. Consequently, we believe that the Cranberry Gneiss formed by metasomatic and magmatic processes and is younger than the nonplutonic rocks it invades.

The discordance between the Cranberry Gneiss and adjacent rocks indicates that either the Cranberry Gneiss is younger than the mica schist, as we have previously interpreted (Bryant, 1962; Bryant and Reed, 1962, p. 164), or the gneiss is in tectonic contact with the schist. The former conclusion certainly applies to the Cranberry at the west margin of the area that was not affected by the low-grade metamorphism.

Where the contact between the Cranberry and the mica schist is sharp, the rocks along the contact are strongly sheared but no more sheared than the main body of Cranberry Gneiss beneath the contact. Amphibolite layers 1 to 50 feet thick commonly occur along the contact. These layers and all or part of the mixed rocks north of the window might be in tectonic slices along a fault zone. On the other hand, it is difficult to place such a fault in some areas, particularly in areas C–4 and C–5 where digitations of amphibolite, mixed rocks, and mica schist and gneiss project well into the Cranberry Gneiss. Thus, the local evidence is inconclusive whether or not a major fault separates the Cranberry Gneiss from the overlying rocks.

If the data on the age of the schist, gneiss, and amphibolite of the Blue Ridge thrust sheet (Rankin, 1967, 1970; Hadley, 1970) that have been obtained since the above was written can be extended to the Grandfather Mountain area, we think that they support the hypothesis of a major thrust fault between Cranberry Gneiss in the basement and the Ashe Formation overlying it (Rankin, 1970), for units in the younger rocks rather than in the older rocks appear truncated along the contact. Such a relationship suggests that there may be Precambrian metamorphic rocks of two ages in the Blue Ridge thrust sheet in the Grandfather Mountain area: upper Precambrian rocks only metamorphosed during the Paleozoic occurring in a thrust sheet above a basement complex of schist, gneiss, amphibolite, migmatite, and granitic rock, which was metamorphosed 1,000–1,100 m.y. ago.

### AGE

Geologic relations and mineral ages indicate that the Cranberry Gneiss is of Precambrian age. In the eastern Great Smoky Mountains, 50 miles to the west, and in the Gossan Lead district, 30 miles to the northeast, similar plutonic rocks are overlain by rocks of late Precambrian age (Hadley and Goldsmith, 1963; Stose and Stose, 1957).

About 12 miles northwest of the Grandfather Mountain area at Pardee Point on the Doe River, rocks of the Unicoi Formation of Early Cambrian (?) age rest unconformably on plutonic rock (King and Ferguson, 1960, fig. 7) similar to but less sheared than the Cranberry Gneiss in the northwestern part of the Grandfather Mountain area. However, a major thrust fault separates the plutonic rocks at Pardee Point, which is in the intermediate sheet, from those in the Blue Ridge thrust sheet.

Isotopic ages of zircon, micas, and hornblende confirm the Precambrian age of the Cranberry Gneiss and indicate that it formed 1,000 to 1,100 m.y. ago.

## BEECH GRANITE

The Beech Granite (Keith, 1903) crops out in an east-trending belt northwest of the Grandfather Mountain window (pl. 1) and extends westward into Tennessee beyond the limits of the area mapped. The granite forms the summit and north slopes of Beech Mountain, the second highest peak in the Grandfather Mountain area. Hamilton (1960) included this rock among the rocks he described as "the complex of Lunsford Branch area."

The Beech Granite is a coarse-grained inequigranular white to light-pink cataclastic granite or quartz monzonite gneiss containing potassic feldspar crystals that are mostly 5 to 7 mm but are locally 25 mm in diameter. It typically forms large slabby outcrops and cliffs. Cataclastic foliation is obscure in many places, but lineation, which is conspicuous, is formed by aggregates of biotite and epidote and, in places, by drawn-out quartz and feldspar grains. Quartz lenses and veinlets along joints and irregular knots of quartz are widespread. Aplite veins are rare, and no pegmatite has been found. In a few places, the granite contains ellipsoidal inclusions, a few inches long, of a gneiss which is finer grained and more mafic than the granite. Inclusions of this type are exposed on the north side of Mill Creek at about 3,100-foot altitude (area C–2, pl. 1).

Near Skalley Branch (area C–3, pl. 1) are some outcrops resembling Cranberry Gneiss, but whether they are inclusions or roof pendants could not be determined. The granite is cut by zones of black to green and gray phyllitic rocks. Some of these rocks are plainly phyllonites because they contain porphyroclasts of minerals from the Beech Granite. Others contain a higher proportion of chlorite and (or) opaque minerals and are probably greenstones derived from mafic dike rocks.

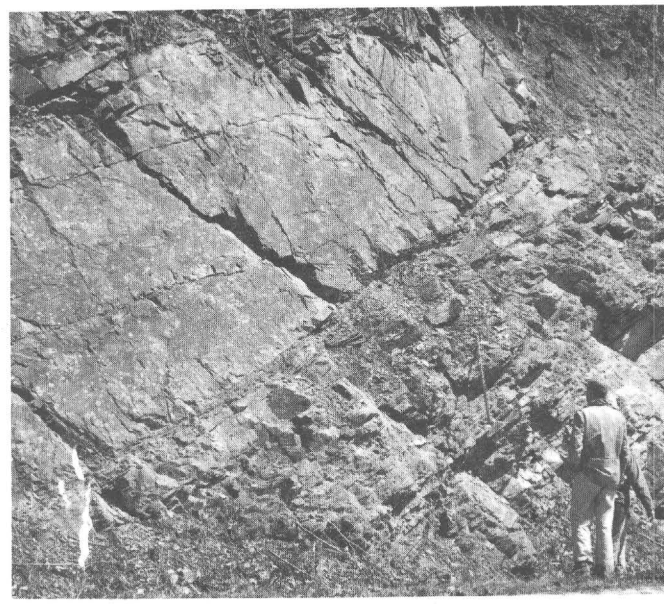


FIGURE 20.—Southern contact of Beech Granite. Contact between Beech Granite and Cranberry Gneiss on road northwest of Heaton (area C–3, pl. 1). Massive rock on left is Beech Granite, a homogeneous blastomylonitic granite gneiss; rock on right is Cranberry Gneiss, a heterogenous blastomylonitic layered gneiss. Amount of movement along contact is unknown.

### CONTACT RELATIONS

Contacts of the Beech Granite in the Grandfather Mountain area are sharp. In many places, especially along the southern contact, shearing and an unknown amount of movement have taken place. Such a sheared contact is exposed on the road northwest of Heaton (area C–3, pl. 1), where coarse- to medium-grained nonlayered gneissic cataclastic Beech Granite is in sharp and slightly discordant contact with layered Cranberry Gneiss. The Cranberry Gneiss near the contact consists of dark fine-grained layers of dioritic composition (now albite-actinolite-epidote-biotite schist) and gradations to coarse-grained light-colored blastomylonitic quartz dioritic gneiss containing white feldspar-rich pegmatite stringers and pods (fig. 20). The strongly sheared character of the underlying Cranberry Gneiss could be due to thrust faulting of the Beech Granite over the Cranberry or to the difference in mechanical properties between the homogeneous granite and the layered Cranberry Gneiss. Along the southern contact of the Beech Granite, cataclastic foliation is concordant or semiconcordant with the contact, the attitude of which may have been modified by movement parallel with the foliation.

Commonly, the Beech Granite is slightly finer grained within a few to a few tens of feet of the contact. In some places along the northern contact, the Beech Granite is poor in mafic minerals and contains considerable amounts of green iron-rich sericite. It appears crushed rather than strongly foliated.

The best exposure of the northern contact of the Beech Granite (fig. 21) is in Beech Creek (area D–2, pl. 1). There, greenish-gray porphyroclastic quartz monzonite to granodiorite gneiss containing aplite dikes and sills is in sharp contact with white to

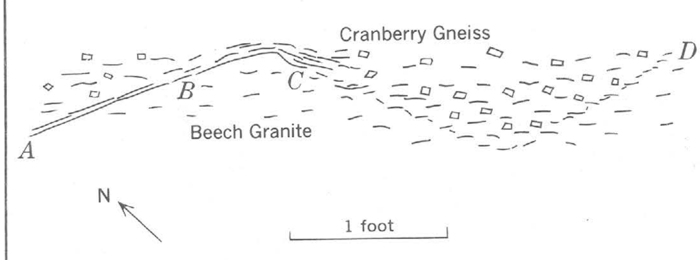



FIGURE 21.—Northern contact of Beech Granite exposed on a slab in Beech Creek (area D–2, fig. 4). Cranberry Gneiss is well foliated and contains potassic feldspar porphyroclasts as much as an inch long. Beech Granite is poorly foliated and medium grained next to the contact. Between *A* and *B* later movement along the contact has produced ¼-inch-thick well-foliated rock parallel to the contact in the Cranberry Gneiss, between *B* and *C* foliation has wrapped around irregularity in contact, and between *C* and *D* foliation has cut the contact and is obscure over a 1- to 2-inch interval.

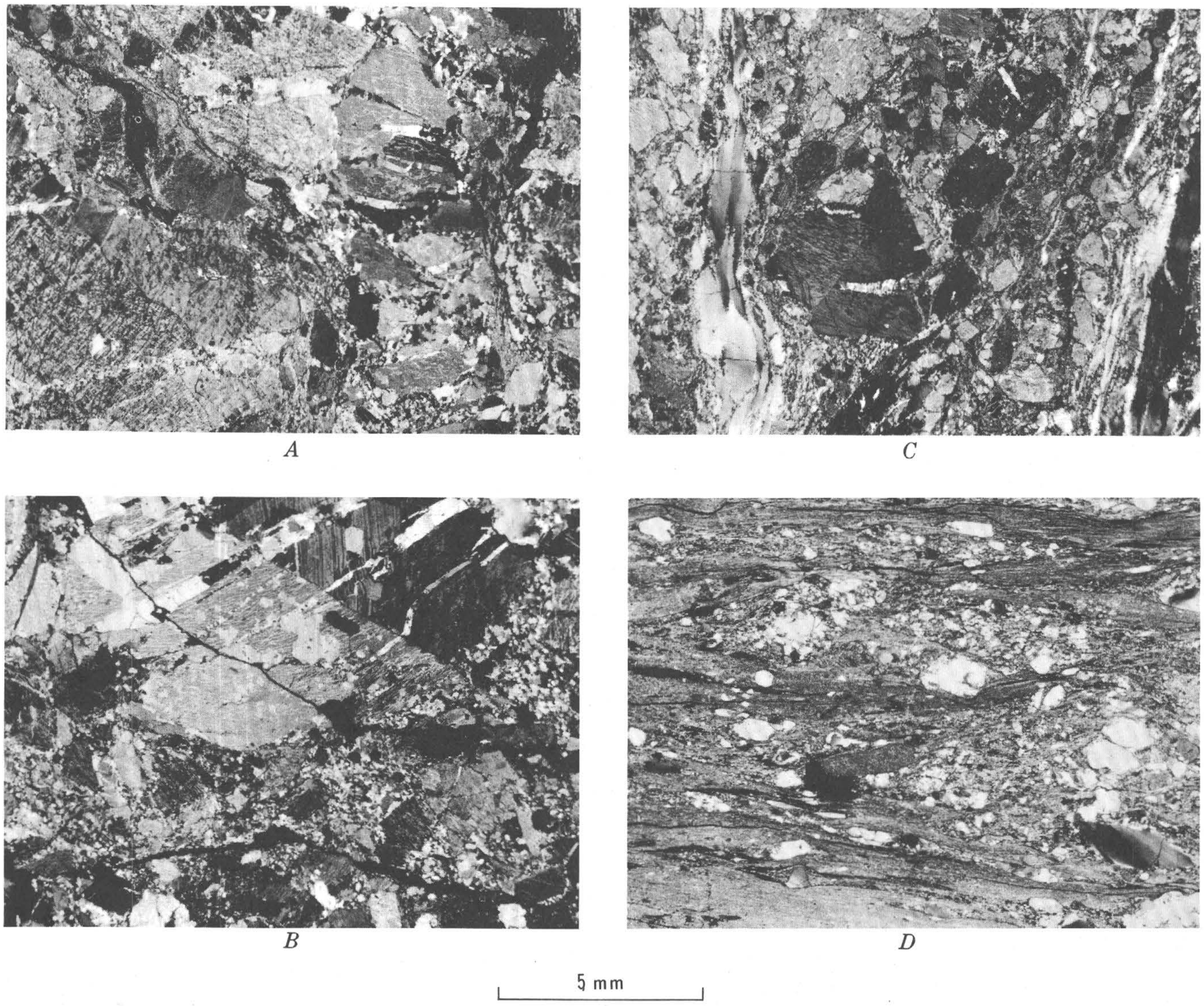



FIGURE 22.—Photomicrographs of Beech Granite and phyllonite. *A*, Coarse-grained recystallized mortar gneiss from Jones Falls (area C–2, pl. 1). Perthite broken and healed by quartz. Smaller grains of plagioclase altered to albite. Recrystallized mortar is mostly quartz but contains some albite, sericite, and biotite. Section cut subparallel to poorly developed lineation. *B*, Coarse-grained blastomylonitic gneiss from north side of Mill Creek valley (area C–2, pl. 1). Large broken porphyroclast of vein perthite containing patches of albite is healed by recrystallized quartz. Small plagioclase porphyroclasts altered to albite. Large plagioclase grain included in darker Carlsbad twin of potassic feldspar bent near quartz-filled fracture. Matrix of recrystallized quartz, albite, epidote, and biotite. Section cut at about 45° to mineral lineation. *C*, Coarse-grained blastomylonitic gneiss. From east side of Laurel Creek at 3,030-foot altitude (area C–2, pl. 1). Porphyrociast of vein perthite in a matrix of recrystallized quartz, albite, sericite, and biotite. Porphyroclasts of quartz are bent, broken, and strung out to form mineral lineation. Section cut parallel to lineation. *D*, Porphyroclastic phyllonite from about 3,300-foot altitude on north side of Big Pine Mountain (area C–2, pl. 1). Porphyroclasts of quartz, plagioclase (altered to albite), and potassic feldspar, in a matrix of sericite, quartz, and biotite. Section cut parallel to lineation.

light-gray Beech Granite. The contact dips 65° to 75° SW. The outcrop pattern of the contact in this area also indicates that it dips more steeply than the cataclastic foliation. A thin section from the contact in an adjacent roadcut shows that the Beech Granite is fine grained at its contact and that the foliation cuts the contact.

The Stone Mountain fault forms the contact of the Beech Granite along the northwestern margin of the Grandfather Mountain area.

PETROGRAPHY

The Beech Granite consists of numerous porphyroclasts of microperthite, a few porphyroclasts of quartz and albitized plagioclase, and rare porphyroclasts of biotite, in a fine-grained matrix of recrystallized albite, quartz, biotite, epidote, and sericite, generally having a grain size of 0.05 to 0.2 mm in diameter (fig. 22). The proportion of groundmass to porphyroclasts is highly variable. Some rocks have undergone only incipient crushing accompanied by formation of rims of mortar around quartz grains and by bending of biotite, whereas at the other extreme are blastomylonites and phyllonites in which no porphyroclasts remain.

The potassic feldspar is everywhere perthitic (fig. 22 *A*, *B*, *C*) and both string and patch varieties occur. Carlsbad twinning is widespread, and grid twinning is quite common but rather poorly developed. Plagioclase grains are locally included in the perthite. Many of the large crystals have been broken and healed by quartz (fig. 22*A*, *B*). The orange, pink color of much of the perthite seems to be due to numerous minute inclusions, which may be exsolved iron oxide.

Plagioclase occurs in grains as much as 5 mm in diameter and in groundmass size. Locally, plagioclase porphyroclasts are bent.

Quartz porphyroclasts are as much as 5 mm in diameter, and in some strongly sheared rocks, they have been stretched into rods as much as 2 cm long (fig 22*C*). In one specimen of slightly sheared granite from the western edge of the Linville quadrangle, quartz is interstitial to the feldspar. Quartz was generally the first mineral to break down during cataclasis; it recrystallized to mosaic-textured aggregates with a grain size of 0.05 to 0.2 mm. However, it was a resistant mineral under conditions leading to formation of phyllonite, and it occurs as porphyroclasts in that rock type (fig. 22*D*).

Greenish-brown, dark-green, and brown biotite is generally in aggregates accompanied in many places by epidote. Porphyroclasts as much as 2 mm in diameter are found in some of the less-sheared rocks.

Dull-green to brownish-green amphibole is found as a relict primary mineral in some of the Beech Granite south of Skalley Branch (area C–3, pl. 1); it is partly rimmed by a bluish-green or green amphibole having a greater extinction angle.

Zircon is the most common accessory mineral. Most zircon grains are prismatic, but others range from euhedral to round. Sphene and opaque minerals occur, and sphene rims the opaques in many places. Purple fluorite is widespread. Generally it is intergranular, but in places it is also found on fracture planes and in quartz segregations. It is especially noticeable near the margins of the granite along the Stone Mountain fault on the road down Dark Ridge and in the Long Ridge area. Metamict allanite is in places rimmed by epidote. Stilpnomelane in radiating aggregates, apatite, carbonate, and sericite are other accessories.

COMPOSITION AND AGE

Approximate modal compositions of the Beech Granite (fig. 23) fall rather uniformly in the granite field. One aplite veinlet (point 3, fig. 23) contains more plagioclase, but the potassic feldspar in that rock is less perthitic than in the wallrock and most of the specimens of Beech Granite examined. The specimen from the contact on Beech Creek (point 4, fig. 23) also contains only slightly perthitic microcline and plots near points calculated from analyses.

Chemical compositions of typical Beech Granite (table 8) are very similar to the average alkali granite of Nockolds (1954). Recalculation of the chemical analysis into ideal orthoclase and albite places that sample quite close to the eutectic for the system $NaAlSi_3O_8$-$KAlSi_3O_8$-$SiO_2$-$H_2O$ at pressure of 500 kg/$cm^2$ (kilograms per square centimeter) (Tuttle and Bowen, 1958, p. 75).

The Beech Granite probably intruded the Cranberry Gneiss after or during a late stage of plutonic metamorphism. No detailed intrusive relationships

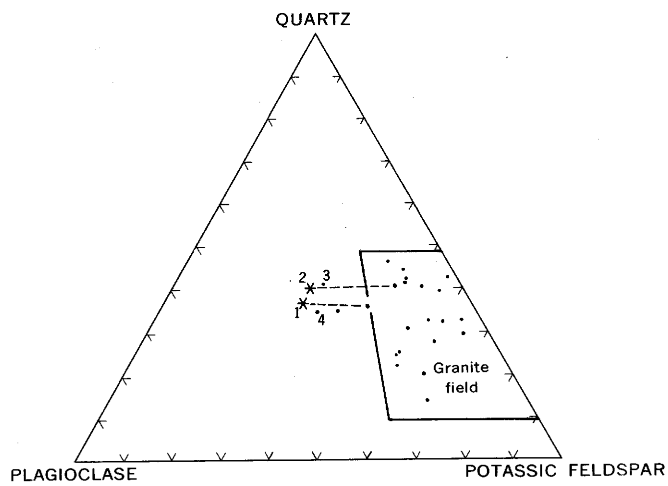



FIGURE 23.—Proportions of quartz, plagioclase, and potassic feldspar in Beech Granite. Perthite is plotted as potassic feldspar. Points 1 and 2 are calculated modes of analyzed samples (table 8) in terms of pure orthoclase and albite. Line connects calculated mode with measured mode; 3, aplite dike cutting granite; 4, Beech Granite at contact on Beech Creek. All points except 1 and 2 based on count of 50 random grains in each thin section. Granite field is outlined.

were found, but the gross map pattern, the sharp contacts, lack of layering, uniformity in composition, and perthitic character of the potassic feldspar support this interpretation. The original intrusive contacts have been modified by shearing during the Paleozoic metamorphism, which produced the present foliation, lineation, and texture of the Beech Granite.

TABLE 8.—*Chemical analyses, modes, and norms of Beech Granite*

[Analysis of sample 1 determined by standard rock analysis by Dorothy F. Powers, U.S. Geol. Survey, 1959. Minor-element analysis of sample 1 given in table 1. Sample 2 analyzed by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey, 1963; F and Cl determined by Vertie C. Smith, U.S. Geol. Survey. Modes, by point counts; P, present but not intersected in counting. Major oxides and CIPW norms given in weight percent; modes, in volume percent]

| | 1 | 2 |
|---|---|---|
| Field No. | GML-6 | O-59-1 |
| Laboratory No. | F-2589 | 161255 |
| **Major oxides** | | |
| $SiO_2$ | 74.61 | 74.2 |
| $Al_2O_3$ | 12.82 | 12.4 |
| $Fe_2O_3$ | .49 | 1.2 |
| FeO | 1.44 | 1.0 |
| MgO | .08 | .19 |
| CaO | .79 | 1.2 |
| $Na_2O$ | 3.68 | 3.2 |
| $K_2O$ | 5.12 | 5.2 |
| $H_2O+$ | .21 | .66 |
| $H_2O-$ | .04 | .04 |
| $TiO_2$ | .14 | .21 |
| $P_2O_5$ | .01 | .04 |
| MnO | .04 | .04 |
| $CO_2$ | .08 | .41 |
| F | ---------- | .12 |
| Cl | ---------- | .01 |
| Total | 99.55 | 100 |
| **Modes** | | |
| Quartz | 32 | 36 |
| Potassic feldspar and perthite | 39 | 40 |
| Plagioclase | 20 | 11 |
| Sericite | 4 | 6 |
| Biotite | 4 | 2.8 |
| Calcite | P | 2.4 |
| Sphene | ---------- | .6 |
| Allanite | P | .2 |
| Epidote | P | ---------- |
| Opaque mineral | P | .6 |
| Zircon | P | P |
| Apatite | P | P |
| Fluorite | P | ---------- |
| Stilpnomelane | P | ---------- |
| Points counted | 706 | 672 |
| **CIPW norms** | | |
| Q | 31.12 | 34.25 |
| C | 0.00+ | .71 |
| Or | 30.25 | 30.72 |
| Ab | 31.12 | 26.99 |
| An | 3.35 | 2.34 |
| Hl | ---------- | .02 |
| En | .20 | .47 |
| Fs | 2.08 | .57 |
| Mt | .71 | 1.74 |
| Il | .27 | .40 |
| Ap | .02 | .10 |
| Fr | ---------- | .24 |
| Cc | .18 | .93 |

1. Medium-grained gray cataclastic biotite granite gneiss with conspicuous lineation formed by aggregates of fine-grained biotite. Bent and broken porphyroclasts of microcline, microcline microperthite, and saussuritized plagioclase as much as 2 mm in diameter in a matrix of recrys-

TABLE 8.—*Chemical analyses, modes, and norms, Beech Granite*—Continued

tallized quartz, albite, potassic feldspar, and subordinate green biotite and sericite having a grain size of 0.05 to 0.1 mm in diameter. Accessory calcite, allanite, stilpnomelane, epidote, fluorite, zircon, and opaque mineral. From quarry in area C-2 (pl. 1) at 3,950-foot altitude on south side of Timbered Ridge just east of road between Heaton (area C-3) and Whaley (area D-2).

2. Coarse-grained cataclastic granite gneiss with light-pink potassic feldspar porphyroclasts as much as 1 cm in diameter. Porphyroclasts of string perthite, as much as 5 mm; quartz, 4 mm; and bent plagioclase (altered to albite) 1 mm in diameter in a matrix of recrystalized quartz, sericite, albite, potassic feldspar, subordinate brown and greenish-brown biotite and calcite, and accessory allanite, sphene, opaque mineral, zircon, and apatite. From roadcut on east shoulder of Beech Mountain at 4,150-foot altitude near south contact of Beech Granite south of Oliver Hollow (area E-2, pl. 1).

## QUARTZ MONZONITE GNEISS

Several elongate bodies of nonlayered crushed and gneissic quartz monzonite crop out north of the Grandfather Mountain window within a mile of the Linville Falls fault. They generally form better outcrops than the adjacent more mafic and plagioclase-rich Cranberry Gneiss. The contacts of the quartz monzonite bodies are strongly sheared in most places, and in places phyllonite is found along them. The bodies of quartz monzonite gneiss may be either tectonic lenses along faults in the Blue Ridge thrust sheet or intrusive bodies in the surrounding layered Cranberry Gneiss. They do not closely resemble rocks exposed along the southeast side of the Grandfather Mountain window, as might be expected if they were tectonic slices.

The quartz monzonite gneiss is a light-pink to gray coarse-grained rock which ranges from well foliated to massive. Pink potassic feldspar crystals form porphyroclasts as much as 2 cm long which have been localled fractured and the fractures filled with quartz. Various amounts of greenish-gray plagioclase are seen in some outcrops. Biotite occurs sparingly, both as porphyroclasts and as recrystallized flakes. Purple fluorite occurs on shear planes and as disseminated blobs as much as 2 cm in diameter. Pyrite and rarely chalcopyrite are visible. Epidote and quartz segregation veinlets occur.

The rocks are mainly recrystallized mortar gneisses but range to blastomylonite gneisses. The recrystallized mortar consists of quartz, albite, sericite, and microcline 0.05 to 0.2 mm in grain size. Quartz, plagioclase, microcline, and microcline microperthite grains are bent, broken, and healed by quartz. In the more strongly sheared rock, only the potassic feldspar remains as porphyroclasts. Proportions of plagioclase and potassic feldspar are variable, and the rocks mapped as quartz monzonite gneiss include some granodiorite and granite. Large plagioclase grains have been albitized. All the biotite in many of the rocks has recrystallized, and it is generally green or light brown. Some rocks contain subhedral grains of sphene as much as 1.5 mm long.

Opaque minerals are magnetite, ilmenite, and pyrite. Fluorite, zircon, apatite, epidote, allanite, chlorite, carbonate, and stilpnomelane are other accessory minerals.

### AEGIRINE-AUGITE GRANITE GNEISS

Small bodies of coarse-grained gray nonlayered granite gneiss were mapped in the Cranberry Gneiss near Crossnore (area C–5, fig. 4) and east of Boone (area H–2, pl. 1). These bodies are poorly exposed, and their contacts with the surrounding layered gneiss are covered, but judging from float the contacts are sharp.

The granite gneiss is moderately well foliated. Perthitic feldspars as much as 1 cm in diameter have finer grained mafic minerals between them.

The large feldspars are porphyroclasts of perthite and antiperthite in a groundmass of recrystallized quartz, albite, microcline, and mafic minerals (fig. 24). Quartz, albite, and stilpnomelane are found in fractures in the feldspar porphyroclasts. Mafic minerals are dark-brown biotite, pyroxene, and amphibole. The pyroxene is aegirine or aegirine-augite. It occurs as porphyroclasts as much as 1.5 mm in diameter and as small grains in the groundmass, but whether these grains are fragments or have recrystallized is uncertain. The amphibole is derived from proxene. It has $2V=35°$, $Z \wedge c=18°$, $Z$=dark green, Y=bright green, and X=light brownish green and is probably hastingsite. Other accessories are zircon, sphene, allanite, magnetitie, ilmenite, tourmaline, and carbonate.

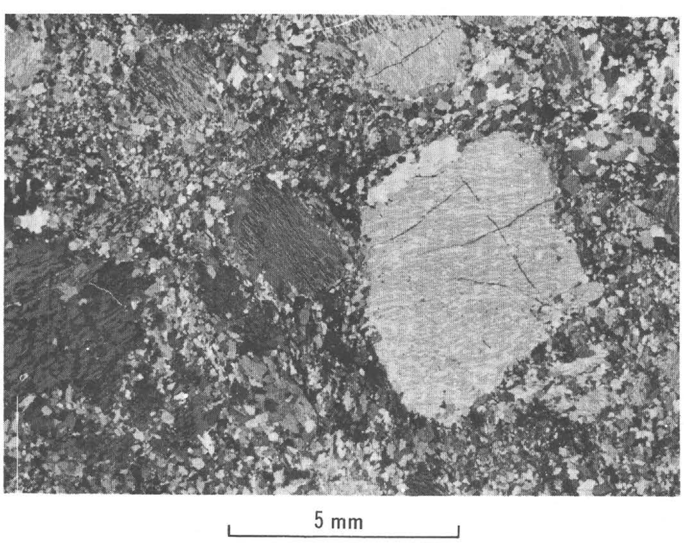



FIGURE 24.—Photomicrograph of aegirine-augite granite gneiss from quarry 0.7 mile northwest of Crossnore (area C–5, pl. 1). Porphyroclasts of vein perthite and antiperthite in groundmass of recrystallized quartz, microcline, albite, and subordinate aegirine-augite, biotite, hastingsite, and stilpnomelane. Analyzed specimen, table 9.

Approximate modes for rocks of this unit (fig. 25) fall into the granite field if perthite is counted as potassic feldspar. However, the perthite is coarse grained enough to count the two phases separately. When this is done, the modes fall near the composition calculated for the analyzed specimen.

Chemical analysis (table 9) of rock from the Crossnore quarry most nearly resembles average ferrohastingsite granite (Nockolds, 1954). Compared with the Beech Granite, the granite is poorer in $SiO_2$ and richer in $Na_2O$, FeO, and combined $Na_2O$ and $K_2O$.

Eckelmann and Kulp (1956) pointed out that the body near Crossnore lacks layering and contains euhedral zircon; they therefore suggested that it is intrusive into the Cranberry Gneiss.

### BAKERSVILLE GABBRO

The term "Bakersville Gabbro" was first used by Keith (1903) in referring to the rocks on Hump Mountain (areas C–3 and C–4, pl. 1) which he correlated with similar rocks near Bakersville, about 13 miles to the southwest. He believed that these rocks were unmetamorphosed and regarded them as being of Triassic age. Subsequently, however, he did not map Bakersville Gabbro in the type locality west of the Grandfather Mountain area because he thought that it occurred only as small irregular bodies (Keith, 1907a). He probably confused Bakersville Gabbro with Roan Gneiss, the name applied to the

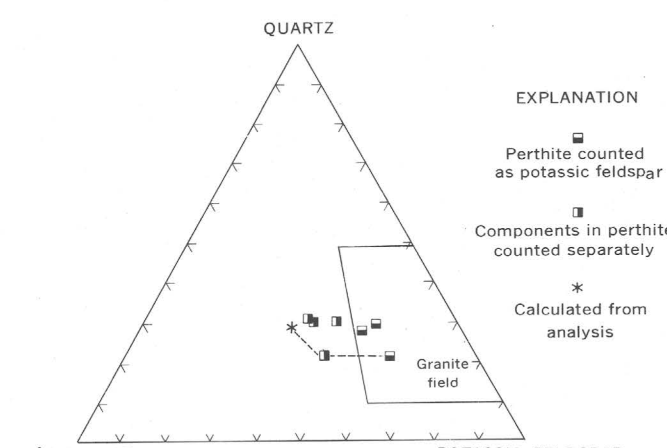



FIGURE 25.—Proportions of quartz, plagioclase, and potassic feldspar in aegirine-augite granite gneiss. Granite field outlined. Leaders connect composition calculated on basis of analysis (table 9) with modes determined by point count. Other points determined by count of 50 random grains in each thin section.

TABLE 9.—*Chemical analysis, mode, and norm of aegirine-augite granite gneiss*

[Standard rock analysis by Dorothy F. Powers, U.S. Geol. Survey, 1959. Mode, by point count, 709 points counted; P, present but not intersected in counting. Minor-element analysis of this rock is given in table 1. Major oxides and CIPW norm given in weight percent; mode, in volume percent]

| Field No. | GML-5 | Field No. | GML-5. |
|---|---|---|---|
| Laboratory No. | F-2588 | Laboratory No. | F-2588. |
| **Major oxides** | | | |
| $SiO_2$ | 71.24 | $K_2O$ | 5.51 |
| $Al_2O_3$ | 13.63 | $H_2O+$ | .13 |
| $Fe_2O_3$ | .64 | $H_2O-$ | .05 |
| FeO | 2.78 | $TiO_2$ | .28 |
| MgO | .04 | $P_2O_5$ | .04 |
| CaO | 1.14 | MnO | .09 |
| $Na_2O$ | 4.12 | $CO_2$ | .08 |
| | | Total | 99.77 |
| **Mode** | | | |
| Quartz | 21 | Hastingsite | 1.3 |
| Microcline, perthite, and antiperthite | 54 | Stilpnomelane | 1.6 |
| | | Opaque mineral | P |
| Albite | 18 | Zircon | P |
| Aegirine-augite | 2.9 | Sphene | P |
| Biotite | 2.3 | Allanite | P |
| **CIPW norm** | | | |
| Q | 22.62 | Fs | 4.28 |
| Or | 32.55 | Mt | .93 |
| Ab | 34.84 | Il | .53 |
| An | 2.43 | Ap | .10 |
| Wo | 1.02 | Cc | .18 |
| En | .10 | | |

NOTE.—Description of analyzed specimen follows: Gray cataclastic granite gneiss. Porphyroclasts of perthite and antiperthite as much as 8 mm in diameter in a groundmass of recrystallized grains 0.06 to 0.6 mm in diameter of quartz, microcline, and albite. Subordinate aegirine-augite, brown biotite, hastingsite, and stilpnomelane and accessory zircon, sphene, allanite, and opaque mineral. From quarry in area C-5 (pl. 1) 0.7 mile N. 38° W. of Crossnore (area D-5).

amphibolites of the region, because he considered the Roan Gneiss to be composed of diorite or gabbro in addition to amphibolite, hornblende, schist, and hornblende gneiss. He also believed that the Roan Gneiss cut the Cranberry Gneiss, but feldspathic layers associated with the Cranberry are found in the hornblende gneiss, and the Bakersville cuts both Cranberry and hornblende gneiss.

Bayley (1923) recognized that the Bakersville Gabbro in the vicinity of Cranberry has been metamorphosed, but he regarded it as a part of the Roan Gneiss because, following Keith, he considered the Roan to be metamorphosed gabbro and diorite.

Kulp and Poldervaart (1956) recognized that the Bakersville Gabbro cuts the Cranberry Gneiss but has been metamorphosed, thus furnishing excellent evidence for the polymetamorphic history of the rocks of the Blue Ridge thrust sheet.

Wilcox and Poldervaart (1958) made a comprehensive study of the Bakersville Gabbro near its type locality; they believed that it occurred solely as dikes and sheets and that the dike swarm did not extend east of Roan Mountain. However, fieldwork by D. A. Brobst (oral commun., 1960) indicates that dikes of Bakersville Gabbro are found between Roan Mountain and the Linville quadrangle.

In the Grandfather Mountain area, the Bakersville Gabbro consists of metamorphosed gabbro, diabase, and basalt which occur in dikes and stocks in a belt extending from Hump Mountain (areas C-3 and C-4, pl. 1) to Spanish Oak Mountain (area C-4, pl. 1). Many of the bodies contain inclusions or septa of granitic gneiss and amphibolite. The best exposures are on the steep sides of Little Hump Mountain where many dikes and larger bodies cut Cranberry Gneiss and rocks of the mixed unit. The dikes trend N. 30°-60° W. and are generally vertical. Some have chilled margins; others are foliated parallel with their margins. The contact of the metagabbro body on Hump Mountain is drawn where dikes apparently make up more than half the rock. Thus, there is much dike material outside the contact and much country rock inside it. In the interior of the body little feldspathic gneiss or amphibolite is found. This metagabbro body is similar to that mapped by Brobst (oral commun., 1960) west of Bakersville. Irregular bodies of metagabbro southeast of Hump Mountain do not have many dikes associated with them.

A few small bodies of metagabbro occur in the Cranberry Gneiss north of the Grandfather Mountain window. Dikes and sills a few inches to several tens of feet thick of fine-grained metamorphosed mafic rocks in the Cranberry Gneiss and Beech Granite may be related to the Bakersville Gabbro. Some of these bodies are exposed in roadcuts along U.S. Highway 19E near the North Carolina-Tennessee State line (area C-3, pl. 1).

The metagabbro is a dark-gray to black rock containing primary labradorite, monoclinic pyroxene, and opaque minerals. The rock has been recrystallized to varying degrees. It is massive to schistose and is locally porphyritic. In the completely recrystallized rocks, hornblende, garnet, calcic oligoclase-sodic andesine, monoclinic pyroxene, biotite, and opaque minerals are the principal constituents.

The Bakersville Gabbro has porphyritic, diabasic, and ophitic textures (fig. 26). All the igneous textures have been modified to some extent by recrystallization. Most of the rocks have relict porphyritic textures that have been partly converted to grano-

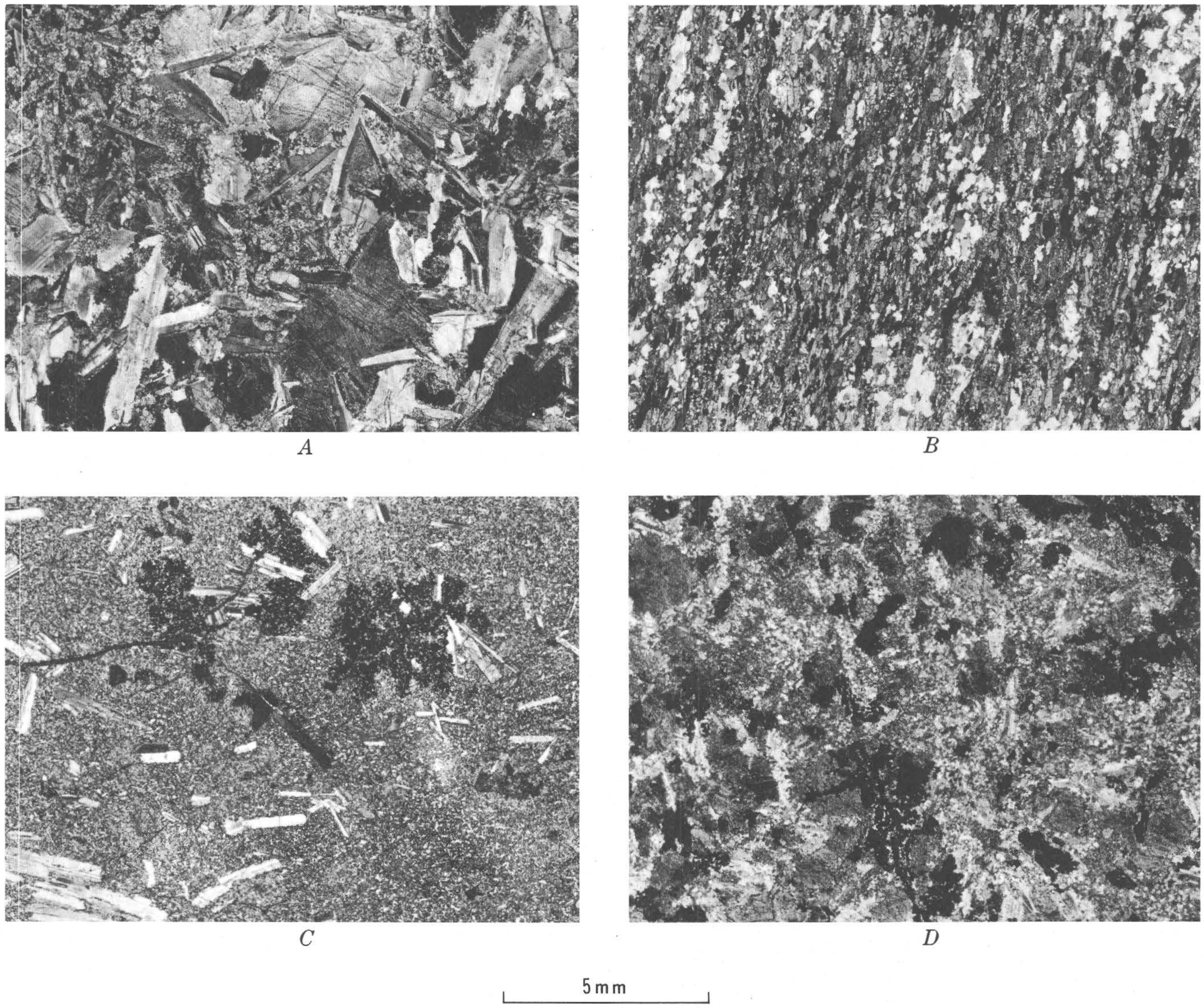



FIGURE 26.—Photomicrographs of Bakersville Gabbro. *A*, Ophitic texture in slightly altered gabbro from interior of dike exposed in roadcut on Little Horse Creek Road where road crosses creek (area C–4, pl. 1). Laths of sodic labradorite, partly enclosed in large monoclinic pyroxene grains. Pyroxene, rimmed by light-green actinolitic hornblende; some pyroxene recrystallized. Garnet locally replaces plagioclase; plagioclase, locally altered. Small amount of hypersthene. *B*, Amphibolite from sheared part of same dike as *A*. Nematoblastic hornblende and sodic andesine. *C*, Porphyritic basalt from 4,950-foot altitude on west ridge of Hump Mountain (area C–3, pl. 1). Phenocrysts of sodic labradorite, in matrix of sodic andesine, hornblende, and pyroxene. Garnet porphyroblasts replace groundmass. *D*, Biotite-garnet-pyroxene metagabbro from 4,200-foot altitude on ridge south of road opposite Bellvue Church (area C–3, pl. 1). Plagioclase, recrystallized to oligoclase, although some lath-shaped crystals remain. Monoclinic pyroxene, partly altered. Garnet porphyroblasts and irregular blobs of magnetite.

blastic textures during recrystallization. Most of the mafic minerals have been converted to hornblende, garnet, and biotite, and much of the plagioclase has been reconstituted, but some phenocrysts of plagioclase as much as 5 cm long are still recognizable.

Where the porphyritic rocks have been strongly metamorphosed, the phenocrysts have been converted into aggregates of plagioclase and garnet. A few of the rocks have relict basaltic textures. One body northeast of Chestnut Ridge (area D–3, pl. 1) contains feldspar laths as much as 2 cm long and has diabasic texture.

Wilcox and Poldervaart (1958) described similar rocks in the Bakersville Gabbro in the Bakersville-Roan Mountain area. Some of the orthoamphibolites they described are identical to gneissose or schistose Bakersville Gabbro from the western part of the Linville quadrangle. Other rocks described by them may be orthoamphibolites derived from older rocks. They attributed the textural variations of the Bakersville Gabbro to variations in the amount of water available during the recrystallization. In the Grandfather Mountain area, more completely recrystallized gabbro has gneissose structure; therefore differences in intensity of shearing may also have been a factor in the textural variations.

In one exposure on the road up Little Horse Creek (area C–4, pl. 1), all gradations from coarse-grained nonfoliated metagabbro to garnet amphibolite occur (fig. 26 *A* and *B*). The amphibolite contains lenses of feldspar representing the former phenocrysts. In exposures on the steep slope west of Tucker Hollow (area C–3, pl. 1), gradations from coarse-grained metagabbro in the center of dikes to amphibolite at the margins can be seen.

Fine-grained hard skarnlike rocks with pink and green mottled appearance form a distinctive type of metamorphosed Bakersville Gabbro. They consist of plagioclase and interstitial pyroxene. The pyroxene is partly altered to hornblende, and anhedral skeletal porphyroblasts or aggregates of garnet irregularly replace the earlier minerals and produce the mottled aspect.

Plagioclase forms 12 to 52 percent of most of the rocks. It is mostly calcic and best retains its shape in fine-grained diabasic-textured rocks in which it forms laths as much as 0.5 mm long. Plagioclase as calcic as $An_{68}$ is found in some rocks. Well-developed normal zoning with as much as 30-percent variation in An content from core to rim is found in some of the less altered grains. Plagioclase phenocrysts have been partly decalcified without destruction of their identity. Epidote minerals locally occur in the phenocrysts. In other places, the phenocrysts have altered to garnet and mosaic-textured aggregates of calcic oligoclase or sodic andesine grains 0.05 to 0.2 mm in diameter (fig. 26*D*). Where shearing has been strong, these aggregates have been made into lenses and strung out, until, with further shearing and recrystallization, the new plagioclase becomes evenly distributed. The lath-shaped igneous plagioclase is more complexly twinned than the granoblastic plagioclase.

Pyroxene forms 0 to 35 percent of the rocks; it is mostly monoclinic, although a minor amount of hypersthene was found in one specimen. The least metamorphosed rocks contain more pyroxene than the more altered ones (fig. 26*A*). Pyroxene forms phenocrysts as much as 6 mm in diameter and equant grains 0.01 to 0.1 mm in diameter. The phenocrysts have $Z \wedge c$ ranging from 40° to 44° and are probably augite and diopsidic augite. Generally, they are largely altered to hornblende and biotite and are filled with many tiny opaque inclusions. Much of the pyroxene occurs only as relicts in aggregates of hornblende grains. The smaller equant grains of pyroxene lack inclusions and have a light-green tinge. In many rocks they seem to be metamorphic, although in some, such as the skarnlike rocks, evidence of their origin is inconclusive.

Amphiboles in the Bakersville Gabbro range from green to olive green, brownish green, dull green, and rarely to bluish green. Amphibole forms 3 to 74 per cent of the rocks. It seems to have been derived mainly from pyroxene. Rocks with granoblastic texture contain amphibole and no pyroxene. The amphibole generally occurs as equant grains 0.1 to 0.2 mm in diameter, but in the more metamorphosed gabbros it is found as alined needles as much as 2 mm long (fig. 26*B*). It surrounds many of the pyroxene grains and in places has a porphyroblastic habit and includes grains of quartz, plagioclase, opaque minerals, and rutile. $Z \wedge c$ ranges from 14° to 18° but is generally 15° to 16°.

Garnet occurs in skeletal crystals and aggregates as much as 5 mm in diameter (fig. 26*C*), which form a trace to as much as 32 percent of some rocks. It has a light-red hue and is commonly sieve textured and contains inclusions of plagioclase, pyroxene, hornblende, and opaque minerals. As the content of hornblende increases, the content of garnet decreases.

Magnetite, lesser amounts of ilmenite, and minor amounts of pyrite, occur as irregular grains as much as 1 mm long and in amounts ranging from a trace to 14 percent.

Brown to orange-brown biotite in flakes 0.2 to 1 mm long forms as much as 20 percent of the rocks. It is derived from garnet and hornblende.

Epidote and chlorite tend to be in and adjacent to late shear zones. Quartz, sphene, and apatite are common accessory minerals, and rutile is a less common one.

Three small bodies of rock entirely surrounded by Cranberry Gneiss are mapped as Bakersville Gabbro, but they differ from the gabbro bodies elsewhere and from each other. The one southwest of the Cranberry High School (area C–3, pl. 1) probably was originally porphyritic and contains pigeonite (now altered to green and colorless amphibole), a few small albite porphyroclasts (relict phenocrysts?), and much very fine groundmass composed of epidote, albite, and sericite. The body east of Heaton (areas C–3 and D–3, pl. 1) has more porphyroclasts of plagioclase and partly altered pyroxene and has a fine-grained blastomylonitic groundmass. The body east of the east fork of Curtis Creek (area D–3, fig. 4) is relatively unsheared and has a very coarse diabasic texture in which the saussuritized plagioclase laths are as much as 15 mm long, and the interstitial titanaugite as much as 5 mm in diameter is partly altered to amphibole.

Small dikes of mafic rock possibly related to the Bakersville Gabbro cut the Cranberry Gneiss and Beech Granite. Most of them are very fine grained and consist of biotite, albite, sphene, actinolite, epidote, chlorite, and opaque minerals. In some dikes, the plagioclase forms laths as much as 2 mm long.

Chemical analyses of Bakersville Gabbro from the Bakersville-Roan Mountain area indicate that the rocks represent a differentiated suite of olivine basalts (Wilcox and Poldervaart, 1958, p. 1351 and table 5). Older analyses from the Cranberry-Hump Mountain vicinity (Bayley, 1923, p. 44; Clarke, 1900) are quite similar, but rock and locality descriptions are inadequate to determine beyond doubt that they came from Bakersville Gabbro rather than older amphibolites. As Wilcox and Poldervaart were unable to establish any chemical differences between Bakersville Gabbro, orthoamphibolite, and para-amphibolite north of the Spruce Pine district, it is not surprising that completely metamorphosed Bakersville resembles some of the older amphibolites.

### QUARTZ PORPHYRY

Thin dikelike bodies of aphanitic light-green to white quartz porphyry too small to map are exposed in the Blue Ridge thrust sheet and the Grandfather Mountain window. They are lenses or tabular masses 10 inches to 50 feet thick, in places parallel to and in places cut by the cataclastic foliation. Some seem to be associated with faults, such as the ones north of Dark Ridge Creek at the southwest end of Horse Ridge (area C–2, pl. 1) and south of Liberty Hill school (area G–2, pl. 1). Others seem to be dikes, such as the ones east of Whaley (area D–2, pl. 1) and north of Aho (area H–3, pl. 1).

The quartz porphyry consists of euhedral to subhedral embayed phenocrysts of quartz as much as 3 mm in diameter in a groundmass of quartz and sericite having a grain size ranging from 0.01 to 0.2 mm. In many rocks the sericite is alined forming foliation. No feldspar was identified with certainty in most of the specimens, but one contains lath-shaped potassic feldspar with Carlsbad twinning. Accessory minerals are biotite, sphene, magnetite (ilmenite?), apatite, and zircon.

The rocks contain various proportions of quartz and sericite, and in several, sericite is dominant. Analysis of a typical quartz porphyry having volcanic texture (table 10) shows that it has lower CaO, $Na_2O$, and $K_2O$ contents than volcanic rocks of equivalent $SiO_2$ contents. Evidently the rock has been partly silicified.

Some of the rocks megascopically resembling quartz porphyry lack any well-defined relict texture and could be silicified blastomylonites.

The quartz porphyry bodies may be related to the felsic volcanic rocks of the Grandfather Mountain Formation.

### CHILHOWEE GROUP

#### SLICES NORTH OF THE GRANDFATHER MOUNTAIN WINDOW

Quartzite, feldspathic quartzite, and quartz pebble conglomerate occur as tectonic inclusions along several faults within the Cranberry Gneiss within a mile of the Linville Falls fault and locally along the Linville Falls fault itself. They also are found along faults of the Stone Mountain family near the Mountain City window.

The slices range from a few feet to as much as 1,000 feet thick; many of the thicker lenses are well exposed and form prominent ridges. The ridges support thick growths of rhododendron and are mantled with quartzite float which makes it difficult to determine the exact width of the quartzite lenses. In some places, phyllonite derived from the granitic gneisses is found adjacent to the quartzite slices, but in other places, the degree of shearing in the gneisses adjacent to the quartzites does not significantly differ from that of the surrounding rock. The width of many of the slices has been exaggerated somewhat

TABLE 10.—*Chemical analysis of quartz porphyry*

[Standard rock analysis by C. L. Parker, U.S. Geol. Survey, 1961. Results given as major oxides in weight percent. Minor-element analysis given in table 1]

| Field No. | GML–9 | Field No. | GML–9. |
|---|---|---|---|
| Laboratory No. | H–3424 | Laboratory No. | H–3424. |
| $SiO_2$ | 78.13 | $H_2O+$ | 1.64 |
| $Al_2O_3$ | 12.94 | $H_2O-$ | .10 |
| $Fe_2O_3$ | 1.24 | $TiO_2$ | .12 |
| FeO | .80 | $P_2O_5$ | .01 |
| MgO | .15 | MnO | .01 |
| CaO | .00 | $CO_2$ | .00 |
| $Na_2O$ | .10 | Cl | .00 |
| $K_2O$ | 4.29 | F | .17 |
| | | Total | 99.63 |

NOTE.—Description of analyzed specimen follows:
Light-greenish-gray rock in dike(?) 30 feet thick in Cranberry Gneiss. Nearly euhedral quartz grains as much as 1.2 mm in diameter. Groundmass of quartz 0.05 to 0.2 mm and sericite 0.01 to 0.03 mm in grain size. Groundmass quartz is partly in spheroidal growths filled with sericite inclusions. Approximate composition: quartz, 64 percent; sericite, 36 percent. From roadcut at 3,270-foot altitude along road east of Whaley (area D–2, pl. 1). Norm, not computable.

on the geologic map (pl. 1); some of the thinner ones are not shown. The slices typically pinch and swell, and some are reduced to a string of separated lenses connected by zones of phyllonite. Bits of quartzite only a few inches thick occur in the phyllonite. Where the quartzite slices or the phyllonite zones are no longer traceable, the faults are dropped on the map, although they may extend farther.

The quartzites are thin bedded to thick bedded, light greenish gray, light green, white, gray, and, less commonly, bluish gray and purplish gray. They contain beds of green sericite phyllite and dark-gray beds rich in heavy minerals. Clastic feldspar grains are light pink. Conglomerates contain quartz pebbles as much as 5 inches long. Feldspar and fine-grained gray to red volcanic(?) rock fragments occur sparingly in the conglomerates. Clasts of muscovite are locally visible. The rocks commonly have numerous veinlets of segregation quartz. Most of the veinlets seem to be younger than the main shearing, but some are older. Cleavage is generally parallel to bedding and subparallel or parallel to the attitude of the slices themselves. Clastic grains and pebbles are elongated in a northwesterly direction. In one specimen, quartz pebbles are drawn into flattened rods 1 cm thick, 2 cm wide, and 10 cm long.

Very locally, the quartzite is enriched in iron and contains abundant disseminated magnetite octahedra or hematite. On the ridge west of Whitehead Creek (area D–3, pl. 1), quartzite and adjacent phyllonite have been enriched in iron. Pyrite also occurs there and locally elsewhere in the quartzites.

Most of the rocks have well-preserved clastic textures. They consist of round or subround clastic grains of quartz, microcline, and rarely muscovite and plagioclase, in a matrix of recrystallized quartz and sericite (fig. 27). Where the rocks are extensively sheared or mineralized, they are difficult to distinguish from blastomylonite or phyllonite derived from the Cranberry Gneiss.

Quartz clasts are commonly strongly strained, and the larger ones are partly granulated and recrystallized into mosaic-textured aggregates (fig. 27*A*). In some rocks they are extremely flattened and elongated. Microcline clasts are fractured and healed by quartz. A minor amount of the potassic feldspar is a finely textured perthite resembling that in the felsic volcanic rocks of the Grandfather Mountain Formation. Clastic grains of muscovite and altered biotite(?) as much as 1 mm long are found locally. Clastic grains of tourmaline and zircon average about 0.2 mm in length. Ilmenite also occurs as clastic grains. The matrix of recrystallized quartz and well-alined muscovite has a grain size ranging from 0.01 to 0.2 mm. Some of the muscovite is light green in thin section and is similar to the iron-rich muscovite of the Grandfather Mountain Formation (see below). That mineral accounts for the green

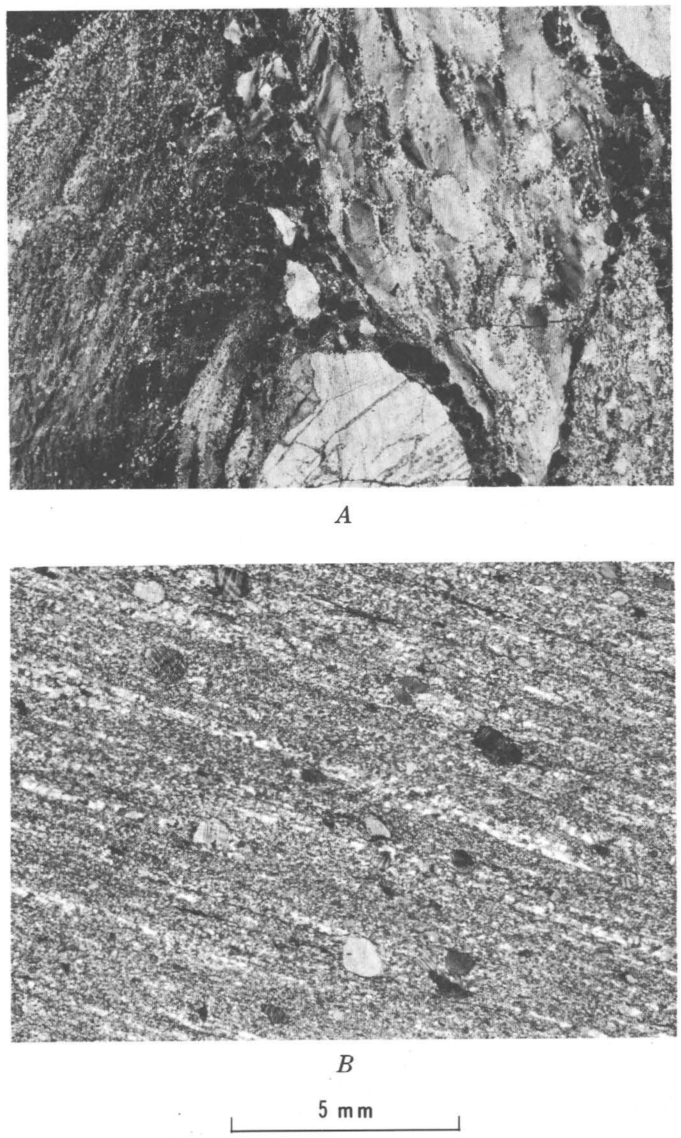



FIGURE 27.—Photomicrographs of rocks of the Chilhowee Group in slices along faults. *A*, Sheared quartz-pebble conglomerate from slice about 300 feet thick along a subsidiary of the Linville Falls fault, 3,515-foot altitude on old road up north fork of Cooper Branch (area C–3, pl. 1). Quartz pebbles in various stages of granulation and recrystallization into mosaic-textured aggregates. Matrix of magnetite, iron-rich muscovite, and zircon. *B*, Biotite-bearing sericitic arkosic quartzite from roadcut on North Carolina Highway 80 near Lake Tahoma dam (area A–10, pl. 1). From slice about 250 feet thick along Linville Falls fault.

color of many of the rocks. Recrystallized ilmenite, magnetite, and hematite are also found. Other accessory minerals are sphene, epidote, and apatite.

Figure 28*A* shows the compositions of sedimentary rocks from the tectonic slices. Most of these rocks are shaly quartzites and quartzites. A few are feldspathic quartzites, and even fewer contain substantial amounts of green sericite and would be considered graywacke in some classifications. However, even the mica-rich rocks are rather light colored. Dark-colored rocks generally contain more opaque minerals, which make up as much as 25 percent of a few rocks.

### SLICES SOUTHEAST OF THE GRANDFATHER MOUNTAIN WINDOW

Thin slices of quartzite and feldspathic quartzite are also found on the southeast side of the Grandfather Mountain window from near Collettsville (area H–6, pl. 1) southwest to the end of the window. They occur along the Linville Falls fault and are intercalated along subsidiary faults in the adjacent gneisses. Slices are also found along the Linville Falls fault on the west side of the window southwest of Lake Tahoma (area A–10, pl. 1). The slices range from a few inches to 40 feet thick, and some can be traced for half a mile along strike. All the slices within the Blue Ridge thrust sheet lie within a few hundred feet of the trace of the Linville Falls fault. Those in the Wilson Creek Gneiss are generally near mapped subsidiary faults, but a few probably mark faults that were not mapped.

The quartzite is a fine-grained white, gray, or light-green sugary rock containing dark partings of heavy minerals and dark-green parting of sericite. Bedding and cleavage are parallel and are conformable with the foliation in the enclosing rocks. Small clastic grains of feldspar are visible in some outcrops, and in a few places small quartz pebbles are present.

The quartzite consists of a mosaic of fine-grained quartz 0.05 to 0.5 mm in grain size, fine-grained green iron-rich muscovite, and minor amounts of feldspar. Angular to subrounded clasts of microcline (fig. 27*B*) and less common microperthite generally form 5 to 10 percent of the rock. In a few specimens from the southwestern extension of the window, small flakes of brown biotite are associated with the sericite. Accessory minerals are magnetite, ilmenite, tourmaline, zircon, and sphene.

The quartzite in most of the slices closely resembles the quartzite of the Chilhowee Group of the Tablerock thrust sheet in mineral composition (compare fig. 28*B* with 59*A* and *B*), including even the presence of clastic tourmaline. The slices are interpreted as having been carried up from the buried part of the Tablerock thrust sheet along the Linville Falls fault and subsidiary faults.

Quartzite in a few of the slices in the Lenoir quadrangle lacks the distinctive clastic texture and tourmaline of most of the rocks elsewhere; it resembles sericite quartzite of uncertain origin found at scattered localities in the Wilson Creek Gneiss.

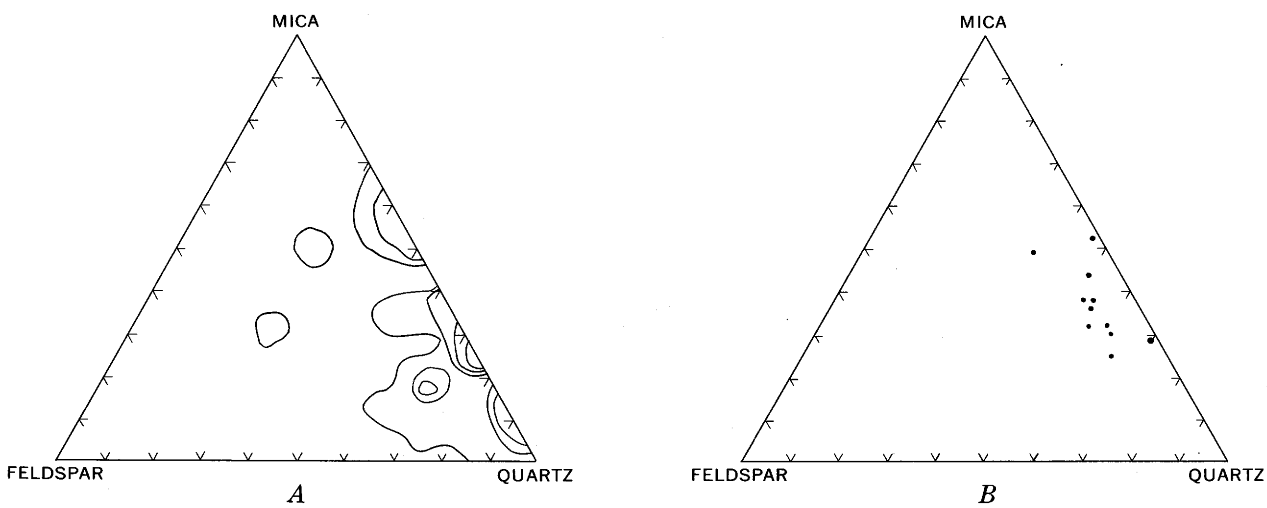



FIGURE 28.—Proportions of quartz, feldspar, and mica in rocks of the Chilhowee Group in tectonic slices in the Blue Ridge thrust sheet. *A*, In the Blue Ridge thrust sheet between the Grandfather Mountain and Mountain City windows. Based on a count of 50 random grains on each of 23 thin sections. Contours 4, 9, 13, and 17 percent. *B*, Along Linville Falls fault on the southeast side of the window. Based on count of 50 random grains in each thin section.

## CORRELATION

Rocks in the tectonic slices resemble those of the Chilhowee Group in the Mountain City window and the Tablerock thrust sheet in the Grandfather Mountain window in their lack of clastic plagioclase and content of clastic tourmaline. In these respects they differ from sandstones in the Granfather Mountain Formation. Conglomerate is more abundant in the slices north of the Grandfather Mountain window than in the upper part of the Unicoi and younger formations of the Mountain City window and in the incomplete(?) section of the Chilhowee Group in the Tablerock thrust sheet.

Rocks in the tectonic slices are more micaceous than the Chilhowee rocks in the Tablerock thrust sheet (compare fig. 28 with fig. 59). Some of the slices may have been derived from lower units or different facies of the Chilhowee than are present in the exposed part of the Tablerock thrust sheet.

## ULTRAMAFIC ROCKS

Small bodies of variously metamorphosed ultramafic rock are found in the Blue Ridge thrust sheet west and north of the Grandfather Mountain window. These are part of a belt of similar rocks extending along the western part of the Appalachian belt from Alabama to Newfoundland (Pratt and Lewis, 1905). All the ultramafic bodies are less than 400 feet long (table 11), and some may have been overlooked. Most of these bodies have been prospected for anthophyllite or soapstone, and the prospects are shown on the geologic maps.

The ultramafic rocks are massive, crudely foliated, or schistose. The rocks are light green to very dark green, greenish gray, or black and contain various proportions of olivine, enstatite, antigorite, anthophyllite, talc, tremolite, Mg-rich chlorite and minor amounts of chrysotile, chromite, pyrite, and carbonate. The olivine-rich rocks weather tan and serpentine-rich rocks, a milky white.

Keith (1903) assigned the ultramafic rocks an Archean age because of their metamorphism. On the basis of evidence from the northern and central Appalachians, Pratt and Lewis (1905, p. 159) suggested that the age of the ultramafic rocks might be Paleozoic, but they pointed out that such an age was not established for the bodies in North Carolina. In a later publication, Lewis (1921, p. 111) concluded that there may be two ages of ultramafic rocks but that both were Paleozoic. Hadley and Goldsmith (1963) found no ultramafic bodies in the upper Precambrian rocks of the Ocoee Series in the eastern Great Smoky Mountains, although they found numerous ones in the underlying rocks, some adjacent to the basal contact of the Ocoee Series. They suggested that the ultramafic rocks are of early Paleozoic age. Kulp and Brobst (1954) noted that pegmatites of the Spruce Pine district cut one ultramafic

TABLE 11.—*Occurrences of ultramafic rocks in the Blue Ridge thrust sheet*

| Location | Approximate dimensions (feet) | Trend | Country rock | Mineralogy | Structure |
|---|---|---|---|---|---|
| Bellvue Mountain (area C-4). One of two bodies mapped together on plate 1. | 400 by 350 | N. 25° W. | Amphibolite | Olivine, enstatite, antigorite with minor chrysotile, talc, Mg-rich chlorite, magnetite, chromite, carbonate. | Crude foliation parallels that in country rock. |
| 0.35 mile N. 27° E. of Mount Pleasant Church (area C-5) at asbestos mine shown on plate 1. | 320 by 60 | N. 60° W. | do | Enstatite, talc, antigorite, olivine, anthophyllite and accessory carbonate and pyrite; relatively talc-rich at margin. | Schistose at margin. |
| Roadcut along road southwest of Golden Creek in area C-5, 1.5 miles S. 44° W. of center of village of Crossnore (area D-5). | Unknown | Unknown | Cranberry Gneiss | Antigorite in core; talc and asbestiform amphibole at margin. | Do. |
| Roadcut along road south of Squirrel Creek 0.2 mile N. 59° W. of Mount Pleasant Church (area C-5). | 20 wide | Unknown | Amphibolite | Mg-rich chlorite, talc, anthophyllite | Schistose. |
| South slope of Hawshore Mtn. (area C-5) at asbestos prospect shown on plate 1. | 300 by 50 | N. 65° W. | do | Tremolite, talc, enstatite, antigorite, olivine, Mg-rich chlorite. | Foliated and lineated where rich in talc. |
| 0.8 mile N. 77° W. of Mount Pleasant Church (area C-5) at asbestos prospect shown on plate 1. | 160 by 50 | N. 10° E. | do | Talc, asbestiform amphibole | Unknown. |
| 0.8 mile N. 81° E. of Turkey Knob (area I-2) at soapstone prospect shown on plate 1. | Unknown | Unknown | do | Talc, tremolite, Mg-rich chlorite, opaque minerals. | Do. |
| 0.6 mile N. 89° E. of Turkey Knob (area I-2) at soapstone prospect shown on plate 1. | 170 by 80 | N. 30° E. | do | Talc, amphibole, antigorite, chlorite | Do. |
| 1.1 mile N. 78° W. of Trivett Gap (area G-2) at asbestos prospect shown on plate 1. | Unknown | Unknown | do | Asbestiform amphibole | Do. |
| 0.5 mile N. 77° W. of summit of Rich Mtn. (area G-2) at asbestos prospect shown on plate 1. | Unknown | Unknown | do | Anthophyllite, talc | Do. |
| 1.0 mile S. 19° E. of Green Valley Church (area C-6). | Unknown | Unknown | Mica schist? | Tremolite, chlorite | Schistose. |
| Along Camp Branch 0.7 mile S. 26° W. of school in village of Triplett (area J-2). | 150 by 250 | N. 35° W. | Cranberry Gneiss | Tremolite, Mg-rich chlorite, asbestiform amphibole, opaque minerals. | Unknown. |

body on the west margin of the district; and therefore, that body is older than 350 m.y., the minimum age of the pegmatite (see below).

In the Grandfather Mountain area we have no direct evidence of the age of the ultramafic rocks. None are known in the basement rocks of the Grandfather Mountain window. Therefore, their absence in the upper Precambrian rocks in the window shows nothing about their age. They are older than the latest medium-grade regional metamorphism of the rocks in the Blue Ridge thrust sheet.

### GRANODIORITE AND PEGMATITE

Irregular stocks, sills, and pods of medium- to coarse-grained white granodiorite (Spruce Pine Alaskite of Hunter and Mattocks, 1936) and pegmatite are found in the schist, gneiss, and amphibolite of the Blue Ridge thrust sheet west, north, and southeast of the Grandfather Mountain window. This area is the east margin of the Spruce Pine pegmatite district. These rocks are best exposed in the valley of Brushy Creek (areas C–5 and C–6, pl. 1) and in the valley of Plumtree Creek (area C–5, pl. 1). They are not found in the adjacent Cranberry Gneiss or in the rocks of the Grandfather Mountain window. They are also absent from the Blue Ridge thrust sheet rocks south of White Oak Branch (areas C–7 and C–8, pl. 1), except near the contact between the Cranberry Gneiss and the mica schist and gneiss. Very few of these rock bodies occur in the mica shist and gneiss north of the window, except in the Deep Gap-Stony Fork area (area J–2, pl. 1), where they occur with biotite quartz monzonite.

Because of their economic importance the pegmatites and "alaskites" of the Spruce Pine district have been studied by many geologists (Sterrett, 1901, 1910, 1923; Maurice, 1940; Olson, 1944; Parker, 1946, 1953; Brobst, 1962; Lesure, 1968). Most of the pegmatite bodies are too small to delineate at a scale of 1:62,500. Concordant bodies range from 1 inch to about 100 feet in thickness; most are less than 30 feet thick. The larger bodies are generally granodiorite, but they also contain pegmatite and all gradations between the two rocks. Larger pegmatite bodies are commonly flanked by small satellitic stringers and lenses. In the larger and more mica-poor bodies, the granodiorite and pegmatite are weakly foliated, but cataclastic foliation is conspicuous in smaller bodies and along the margins and in restricted zones within larger bodies (fig. 29). Zoning is not well developed in most pegmatites in the Grandfather Mountain area.

The smaller pegmatite bodies are generally more foliated and have lineation formed by alined micas. This lineation is parallel to lineation in the wallrocks. In some larger pegmatites, lineation is conspicuous only near or at the margins, or in restricted shear zones. Lesure (1959) noted foliation and lineation in pegmatite throughout the Spruce Pine district but found them to be best developed near the margins of the district.

Inclusions of schist are found in the granodiorite and locally in the pegmatite. The pegmatite and granodiorite bodies generally have sharp contacts; where they are in contact with amphibolite, a thin selvage of biotite schist is commonly developed.

The granodiorite and pegmatite contain plagioclase, quartz, perthitic microcline, muscovite, and biotite. In the more strongly foliated bodies, porphyroclasts of plagioclase, microcline, quartz, and muscovite are set in a matrix of coarsely recrystallized quartz, plagioclase, microcline, and muscovite. Size of the porphyroclasts is variable. Many of the larger quartz and feldspar grains in the granodiorite are 0.5 to 1 inch in diameter. Perthite crystals in pegmatite are as much as 4 feet long. Muscovite forms books as much as 2 feet in diameter, but most flakes are only 1 to 2 inches in diameter. Muscovite books are commonly bent and ruled. Muscovite flakes 1 to 3 mm long, the same size as those in the adjacent mica schist, occur on foliation planes; fine-grained light-green sericite occurs on widely spaced shear planes. In many places, the last two types of muscovite form a northwest-trending lineation that is parallel to that in the country rock. Biotite forms books and flakes as much as 3 inches in diameter. Red garnet as much as 3 inches in diameter is a common accessory mineral. Apatite, epidote, and allanite are minor accessory minerals, and a host of rarer minerals has been reported from the pegmatites of the Spruce Pine district (Maurice, 1940, p. 59).

Thin sections show large porphyroclasts of plagioclase, microcline, quartz, and muscovite surrounded by mosaics of coarsely recrystallized plagioclase, quartz, microcline, and flakes of muscovite (fig. 30). The quartz and feldspar grains are as much as 2 mm in diameter. Plagioclase porphyroclasts are bent and locally broken. Some grains have a weak normal zoning. Plagioclase in different pegmatite bodies ranges from $An_{11}$ to $An_{27}$ Quartz rarely forms porphyroclasts but commonly forms coarse-grained mosaics, especially in some quartz-muscovite parts of the pegmatites.

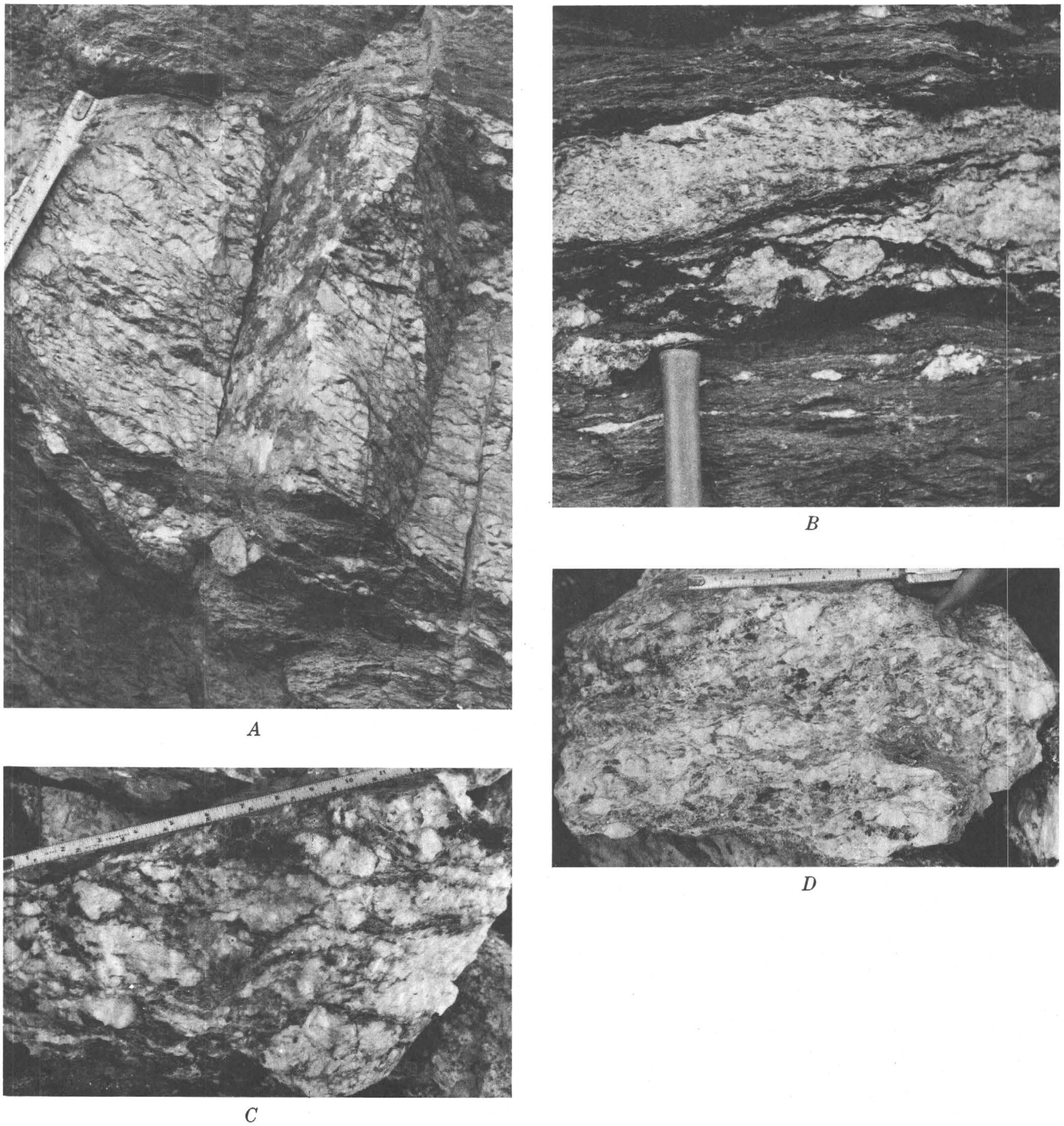



FIGURE 29.—Sheared pegmatite in the Blue Ridge thrust sheet. *A*, Boudin of foliated pegmatite in biotite-muscovite schist on Blue Ridge Parkway near contact between schist and Cranberry Gneiss on west side of Humpback Mountain (area C–7, pl. 1). Note feldspar porphyroclasts in schist adjacent to pegmatite. *B*, Sheared pegmatite and feldspar porphyroclasts in schist. Same locality as *A*. *C* and *D*, Foliated pegmatite from gneissose zone exposed in the Slippery Elm Mine, north side of Plumtree Creek west of Fall Branch (area C–5, pl. 1). Thin-section study shows that porphyroclasts of plagioclase and muscovite occur in a groundmass of recrystallized quartz and plagioclase which resembles the surrounding schist and gneiss in grain size.

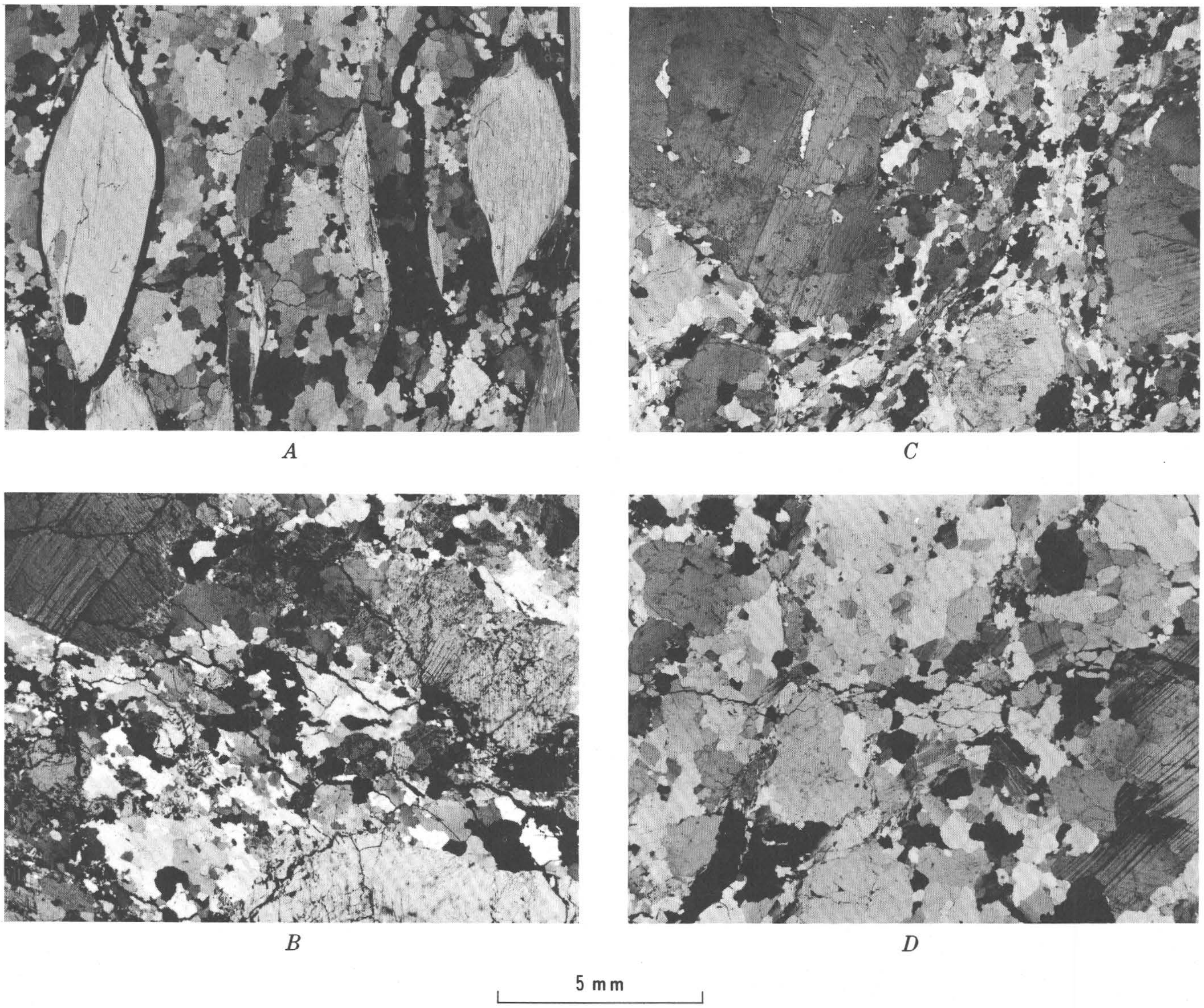



FIGURE 30.—Photomicrographs of granodiorite and pegmatite in the Blue Ridge thrust sheet. *A*, Muscovite-quartz pegmatite from ridge of Rich Mountain 0.4 mile east of summit (area G–2, pl. 1). Elliptical porphyroclasts of muscovite in a mosaic of recrystallized quartz. (Some separation in section, especially around porphyroclasts.) *B*, Foliated pegmatite from Bellvue Mountain (area C–4, pl. 1). Porphyroclasts of oligoclase in a matrix of recrystallized quartz and oligoclase (feldspar, slightly weathered). Accessory microcline.). *C*, Foliated pegmatite from a layer 4 feet thick on south side of Doe Hill Mountain (area C–6, pl. 1). Porphyroclasts of oligoclase in a matrix of recrystallized quartz, oligoclase, and muscovite. Section perpendicular to mineral lineation. *D*, Quartz diorite from Mill Race mine near Brushy Creek (area C–6, pl. 1). Oligoclase porphyroclasts in a coarse-grained mosaic-textured matrix of quartz and oligoclase. Porphyroclasts are bent and broken.

Muscovite porphyroclasts are elliptical (fig. 30*A*) and commonly are bent. Many are partly recrystallized into aggregates of muscovite 0.5 to 2 mm long.

Microcline is perthitic an dseems to replace plagioclase in less-sheared rocks.

### COMPOSITION AND AGE

Most specimens of nonpegmatitic rocks studied in thin section are oligoclase quartz diorites containing few or no mafic minerals, but some are granodiorites and quartz monzonites (fig. 31). Most of the rocks of this unit are too coarse grained for thin sections to be representative, and field estimates indicate that most are granodiorite. Brobst (1962) estimated that the average mineral composition of "alaskite" and pegmatite in the Spruce Pine district is oligoclase,

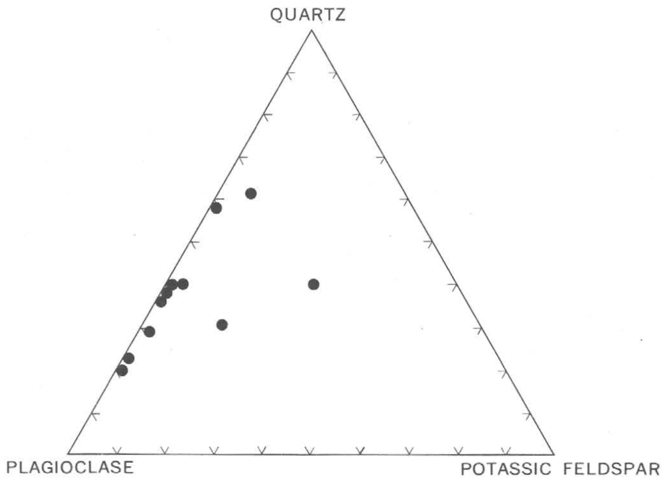



**FIGURE 31.**—Proportions of quartz, plagioclase, and potassic feldspar in light-colored granodiorite and pegmatite. Based on count of 50 random grains in each thin section.

40 percent; quartz, 25 percent; perthitic microcline 20 percent) and muscovite, 15 percent. Chemical analyses of rocks from near Sprune Pine cited by Hunter (1940) and Olson (1944) show that the "alaskite" there is higher in $SiO_2$ and $K_2O$ and lower in CaO than most granodiorites but has more CaO, $Al_2O_3$, and $Na_2O$ and less $K_2O$ than most alaskites. No information on the mineralogy of the analyzed samples is available, however. The mineralogy of rocks of this unit in the Grandfather Mountain area indicates that they may contain more $Na_2O$ and less $K_2O$ than the analyzed specimens.

The granodiorite and pegmatite have been partially sheared and recrystallized under the same conditions as the wallrocks during the latest metamorphism. Keith (1903) recognized the polymetamorphic character of the wallrocks and concluded that the pegmatites had been intruded after one metamorphism but before the latest metamorphism. He mentioned that they "were thoroughly crushed and drawn out by the second deformation and retain in many places only a fraction of their original coarseness." He believed they were of Archean age.

Maurice (1940) recognized three sets of textures in the pegmatities: magmatic, late magmatic, and cataclastic. He stated that "in the majority of the pegmatites the magmatic crystallization was brought to a close by stress resulting in secondary texture * * *" (compare Maurice's figs. 1, 2, 6, and 7 with our fig. 30). On the basis of a chemical age determination on uraninite, he suggested that the pegmatites were of Paleozoic age.

Kulp and Poldervaart (1956, p. 339) stated that "the pegmatites and alaskite are believed to have been placed during the later phases of the second regional metamorphism * * * since there is no evidence of deformation or metamorphism of these rocks" but added "near the contacts, however, exposures may show faint to distinct foliation which parallels the foliation of the country rocks." Eckelmann and Kulp (1957, p. 1123) stated that "the formation of the pegmatites appears to have concluded the metamorphic history of the region."

Many isotopic age determinations have been made on minerals from the pegmatites of the Spruce Pine district; all samples have come from west of the Grandfather Mountain area. Table 12 summarizes these results. All methods give ages in fair agreement at about 350 m.y. It has been widely assumed that this is the date of emplacement of the pegmatites. However, because micas from the pegmatites and the country rocks give the same results, Long, Kulp, and Eckelmann (1959) recognized that the date of 350 m.y. "probably represents the date of the height of the last regional metamorphism of the area." Whether the age determinations on the minerals from the pegmatites indicate the date of the last regional metamorphism of the rocks of the Blue Ridge thrust sheet or the date of emplacement of pegmatites is unknown, for the effects of metamorphism on the contents of the various isotopes of lead and uranium in the uraninite are uncertain. The granodiorite and pegmatite may be synorogenic, virtually the same age as the determinations indicate, or they may be older.

Pegmatites in the Blue Ridge thrust sheet north and southeast of the window have not been dated, but their mineralogy and structural habit is similar to pegmatites of the Spruce Pine district, and they are probably correlative.

Lesure (1968) postulated that the pegmatites in the Blue Ridge belt formed during Paleozoic regional

**TABLE 12.**—*Ages of minerals from pegmatites of the Spruce Pine district*

| Mineral | Age (Million years) | Type of determination | Number of determinations | Source |
|---|---|---|---|---|
| Uraninite | 310–365 | Chemical | 8 | Eckelmann and Kulp, 1957, table 6 (summary of published determinations). |
| Uraninite | 322–375 | Isotopic | 3 | Eckelmann and Kulp, 1957, table 5. |
| Samarskite | 300–405 | $Pb^{207}/Pb^{206}$ | 4 | Do. |
| Muscovite | 334–348 | K:Ar | 3 | Long, Kulp, and Eckelmann, 1959, table 2. |
| Uraninite | 370–420 | Isotopic | 2 | Davis, Tilton, and Wetherill, 1962, table 4. |
| Muscovite | 335 | K:Ar | 1 | Do. |
| Muscovite | 375 | Rb:Sr | 1 | Do. |
| Microcline | 385 | Rb:Sr | 1 | Do. |
| Muscovite | 328–348 | Rb:Sr | 3 | Deuser and Herzog, 1962, table 2. |
| Biotite | 311 | Rb:Sr | 1 | Do. |

metamorphism of at least kyanite grade rather than by differentiation from a large underlying batholith because of their simple mineralogy, lack of rare elements that might be expected in late differentiates of a large granitic batholith, and their localization in rocks of at least kyanite grade.

### QUARTZ MONZONITE AND PEGMATITE IN THE DEEP GAP AREA

A swarm of small intrusive bodies of granitic rock and pegmatite of variable composition cuts mica schist and gneiss in the Blue Ridge thrust sheet in northeastern part of the Blowing Rock quadrangle. Dikes, sills, and lenses 1 to 60 feet thick locally make up about 50 percent of the bedrock.

The rocks are generally crudely foliated, especially those which are mica poor. One leucocratic granitic sill 8 feet thick contains concordant stringers of pegmatite ½ to 2 inches thick and has 4 inches of pegmatite at its margins. Xenoliths of wallrock were seen in one pegmatite dike. Some bodies contain partings, indistinct lenses of schist, and numerous feldspathic lenses. Gradations between granitic rock and pegmatite occur. Some quartz and feldspar grains in the coarsest pegmatite are as much as 6 inches in diameter. Muscovite books as much as 3 inches in diameter occur; they are generally bent or ruled. Biotite books as much as 1 inch in diameter occur in a few places.

The rocks have a white to salt-and-pepper appearance and are fairly fine grained to pegmatitic. They contain various proportions of microcline. Textures are granoblastic to lepidoblastic. The larger muscovite crystals are porphyroclasts. In finer grained rocks having granoblastic textures, microcline replaces oligoclase. Quartz in grains 0.2 to 2 mm in diameter occurs in mosaic-textured aggregates between the larger plagioclase grains. Locally, equant quartz aggregates suggest outlines of a former large grain.

These rocks were emplaced before or during the latest metamorphism. They could be the same age as the light-colored granodiorite and pegmatite, or they could be older. They differ from that unit by having a higher biotite and potassic feldspar content and a somewhat more crosscutting structural habit.

## STRUCTURE AND METAMORPHISM NOT RELATED TO THE BREVARD FAULT ZONE

The gross structure of the Blue Ridge thrust sheet in the Grandfather Mountain area is that of an irregular dome with foliation and layering dipping away from the Grandfather Mountain window (pls. 1, 2 and fig. 32). Foliation and layering in the Cranberry Gneiss near the window are generally subparallel to the Linville Falls fault; dips are gentle in the western part of the Linville Falls quadrangle and steep near the southwestern extension of the window. They steepen somewhat near the subsidiary faults northwest of the window. These faults dip more steeply than the Linville Falls fault and are probably cut by it at depth. Abrupt large-scale flexures in the foliation and layering trend N. 50° W. from the northwest corner of the window and N. 30° E. from the northeast corner. Large gentle folds are found in the Cranberry Gneiss and Beech Granite in the structural saddle between the Grandfather Mountain and Mountain City windows.

In the rocks tectonically overlying the Cranberry Gneiss, dips are gentle and strikes are erratic. The structural and statigraphic relations of these rocks to the Cranberry are complex and are not well understood. In much of the Grandfather Mountain area they are separated from the Cranberry by a metamorphic and a structural discontinuity which may mark a major fault, a possibility that will be discussed in more detail below. The pattern of the amphibolite layers in the mica schist and gneiss (pl. 1) does not indicate structural complexity, but minor folds suggest that the gross structure may consist of recumbent sheared-out isoclines.

The structure and metamorphism of rocks of the Blue Ridge thrust sheet southeast of the Grandfather Mountain window are intimately related to the Brevard fault zone and are discussed in connection with it.

### FOLIATION

Foliation marked by alined micas, tabular quartz-feldspar laminae, and planar arrangement of amphiboles is well developed in almost all the rocks in the Blue Ridge thrust sheet. In most of the mica schist, gneiss, and amphibolite, cataclastic effects are lacking, and foliation apparently formed during synkinematic recrystallization. In most of the plutonic rocks and in some of the pegmatites in the mica schist and amphibolite, on the other hand, the foliation is a cataclastic structure formed by partial or complete granulation and recrystallization; this structure we refer to as cataclastic foliation. Where compositional layering is present, it is parallel with foliation except in the noses of tight minor folds.

### LINEATION

Lineation formed by alined minerals and mineral aggregates and by elongated porphyroclasts and boudins is ubiquitous. It is best developed in the Cran-

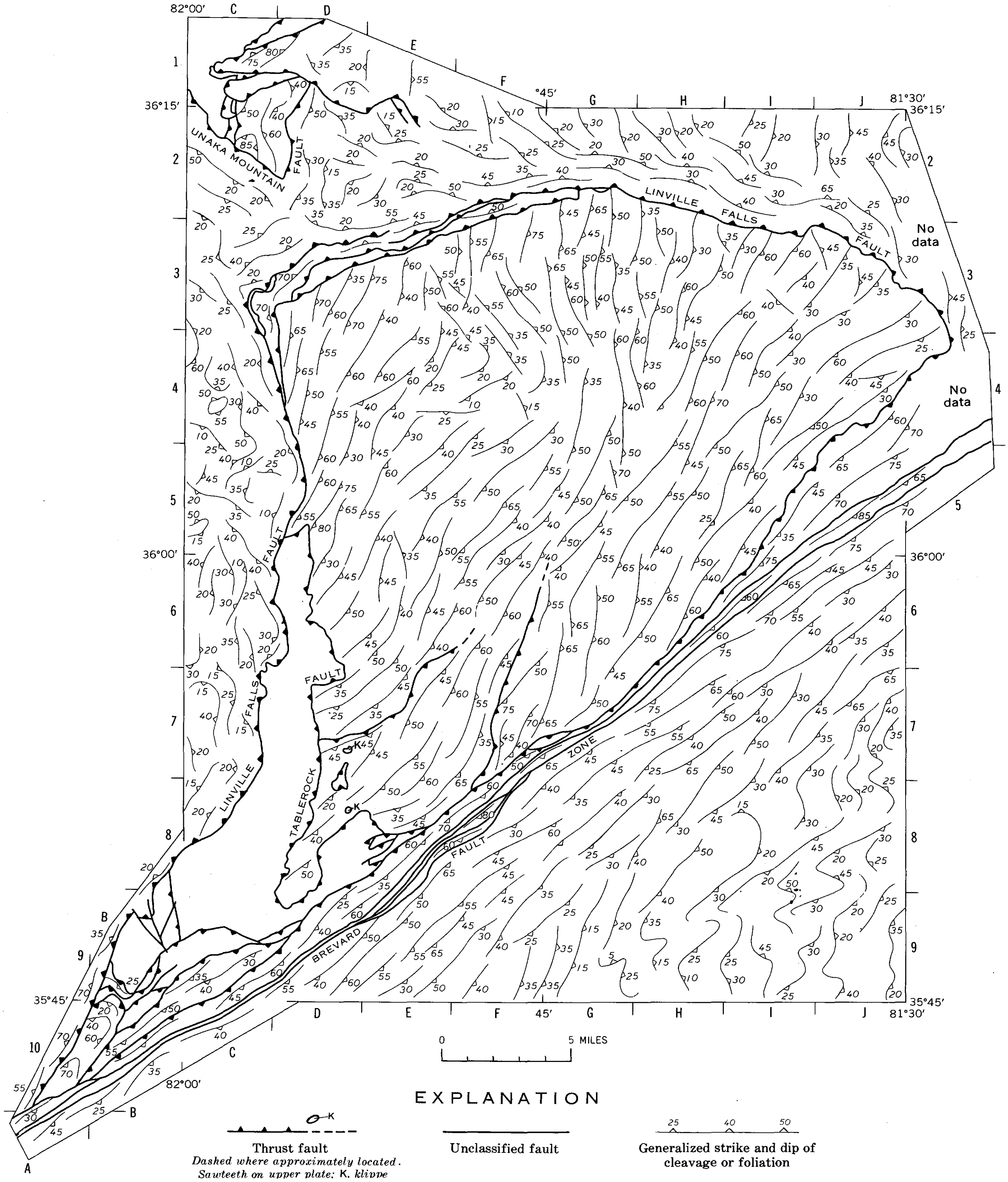



**FIGURE 32.**—Map of the Grandfather Mountain area showing generalized trends of cleavage and foliation. Grid indicated by letters and numbers along margin corresponds to grid on plate 1.

berry Gneiss and Beech Granite, where it is a cataclastic structure formed by drawn-out and incompletely granulated and recrystallized minerals. North and west of the window, the lineation in these rocks trends consistently northwest, but the trend ranges from about N. 60° W. by the northwest corner of the Linville quadrangle to N. 20° W. in the mica schist and gneiss in the west-central Linville Falls quadrangle (fig. 33). In the mica schist and gneiss tectonically above the Cranberry Gneiss, the lineation, like the foliation, formed during synkinematic recrystallization. Lineation in these rocks generally trends northwest, but in a few places both west and north of the window, northeast-trending lineations are found. In one outcrop near the junction of Pine Run with the New River (area I–2, pl. 1), where both lineations were present, the northwest-trending one seems to be the younger.

Regional geologic relations which require northwest transport of the Blue Ridge sheet over the Grandfather Mountain window and Unaka belt indicate that the northwest-trending mineral lineation in the thrust sheet is in the *a* direction. Local evidence in the rocks of the Grandfather Mountain window (discussed below) confirms this interpretation. The northeast-trending lineation may be relict from a deformation older than the thrusting, but until the structures in adjacent areas are more thoroughly studied, these relations will remain undetermined.

#### MINOR FOLDS

Two principal sets of minor folds are found in layered rocks of the Blue Ridge thrust sheet north and west of the Grandfather Mountain window: tight and isoclinal folds having axial planes parallel or nearly parallel with the foliation and layering and axes trending in various directions, and open folds and crinkles having steeply dipping axial planes and northeast-trending axes.

The tight and isoclinal folds are most conspicuous in the markedly layered parts of the Cranberry Gneiss near the contact with the overlying mica schist and gneiss and amphibolite. They are well exposed in roadcuts along Golden Creek (area C–5, pl. 1) and along the Elk River (area J–3, pl. 1). Large folds of this type are exposed just north of the Grandfather Mountain area (Hamilton, in King and Ferguson, 1960, pl. 4*A*). The rare outcrops in which foliation and layering are not parallel are in the noses of these folds (fig. 5). The folds have an amplitude ranging from a few inches to at least tens of feet.

Figure 1*A* on plate 5 shows the orientation of the tight and isoclinal folds west of the Grandfather Mountain window and their geometric relations to the foliation, layering, and mineral lineation. Axial planes of the folds are parallel with the foliation and layering, and their axes form a girdle in the foliation which has a maximum parallel with the mineral lineation.

Figure 1*B* on plate 5 shows that a somewhat similar relationship exists north of the window in the Blowing Rock quadrangle, but there the axes are more evenly distributed along the girdle and do not form a maximum near the mineral lineation.

The open folds and crinkles are most conspicuous in phyllonite and phyllonitic gneiss in the area of outcrop of Cranberry Gneiss. These folds are most numerous in the saddle between the Mountain City and Grandfather Mountain windows. In places, slip cleavage has developed parallel with their axial planes. Very few folds of this type were identified west or northeast of the window. A photograph (Hamilton in King and Ferguson, 1960, pl. 7) shows typical crinkles of this type in a phyllonite zone just north of the Grandfather Mountain area.

The axial planes of the open folds are generally steeply dipping and perpendicular to the mineral lineation; their axes are subhorizontal or northeast plunging and are perpendicular to the mineral lineation (compare pl. 5, fig. 1, *A*–2 and *B*–2 with pl. 5, fig. 1, *C*–1).

Plotting the attitudes of axes and axial planes of open, transitional, and tight folds north of the window in the Blowing Rock quadrangle shows a transition between them (Bryant and Reed, 1970, fig. 6). These relations suggest that open folds may have developed during thrusting and then may have been progressively rotated, tightened, and flattened during northwestward transport of the Blue Ridge thrust sheet.

#### METAMORPHISM

The rocks of the Blue Ridge thrust sheet have been metamorphosed at least twice and perhaps as many as four times. The present distribution of rock of different metamorphic grades is largely due to the most recent metamorphism, which may be of a different age in different areas. Biotite-bearing low-grade rocks form a rim almost surrounding the Grandfather Mountain window and occupying the saddle between the Grandfather Mountain and Mountain City windows (pl. 6*D*). This rim almost coincides with the distribution of the plutonic rocks.

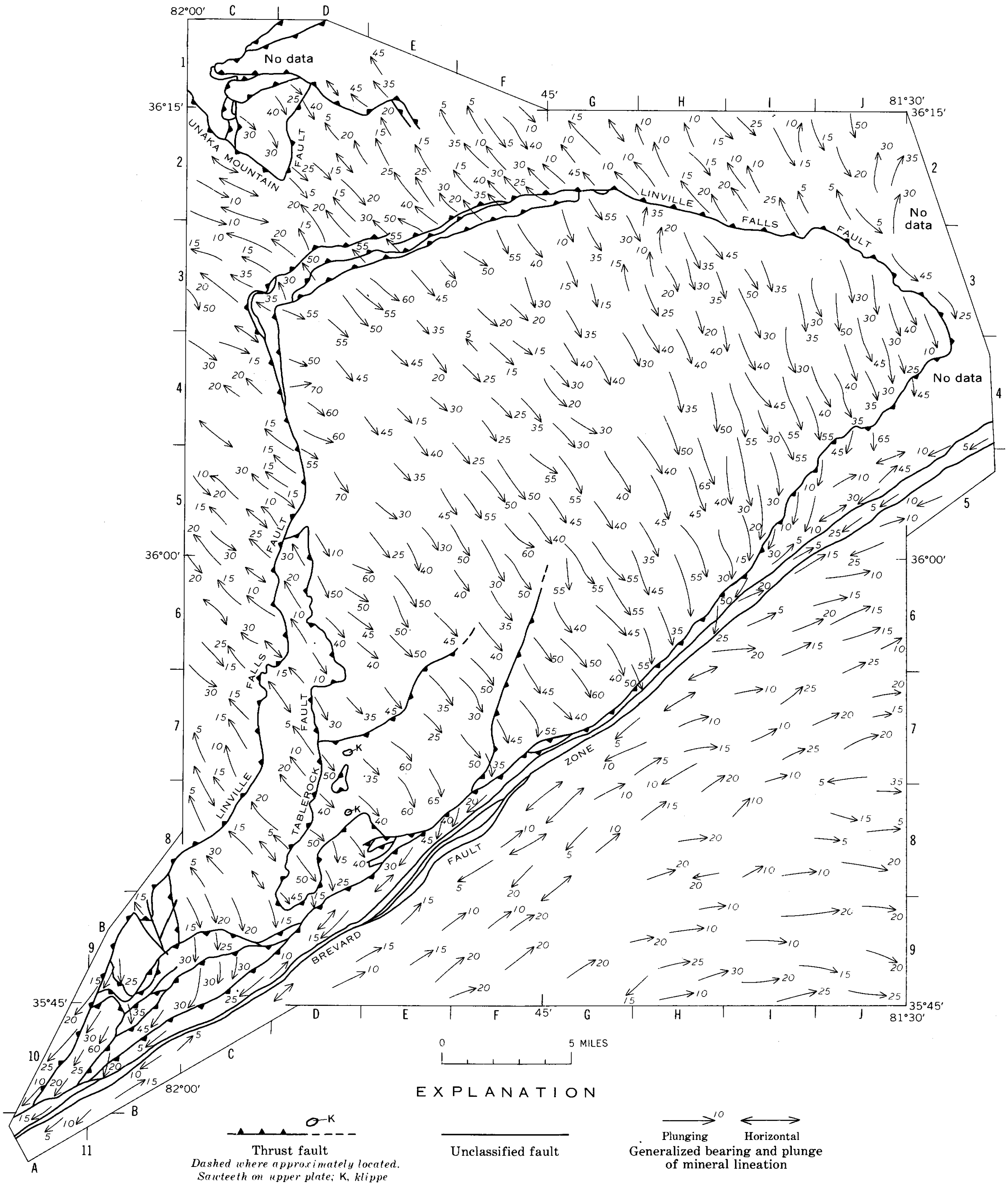



**FIGURE 33.**—Map of the Grandfather Mountain area showing generalized trends of mineral lineation. Grid indicated by letters and numbers in margin corresponds to grid on plate 1. Mineral lineation throughout the area lies on cleavage or foliation planes whose attitudes are shown in figure 32, but because of the generalized nature of the maps, plunges of lineation shown in this figure cannot everywhere be reconciled in detail with attitudes of the planar structures.

West, north, and northeast of the window, outside the rim of low-grade rocks, are medium-grade rocks approximately coinciding with the outcrop area of schist, gneiss, and amphibolite. Southeast of the window and in the intermediate sheet between the Mountain City and Grandfather Mountain windows, the latest metamorphism is of chlorite grade. The metamorphic histories of these rock units and various regional geologic relations and mineral ages discussed below allow us to decipher a history of the rocks and to point out some problems and uncertainties.

The earliest metamorphic event recorded in the rocks of the Blue Ridge thrust sheet is the plutonic metamorphism during which the Cranberry Gneiss formed and sedimentary and volcanic rocks were converted to schist, gneiss, and amphibolite (pl. 6*B*). The hypersthene and monoclinic pyroxene in the Cranberry Gneiss on Little Hump Mountain may be relicts from this episode of metamorphism. If so, the metamorphism probably took place under conditions of the granulite facies, and the Cranberry Gneiss may have once been a charnockite. Jonas (1935) recognized similar rocks in northern and central Virginia, and she believed that the granitic rocks in southwestern Virginia and northwestern North Carolina were the metamorphosed equivalents of the pyroxene-bearing plutonic rocks farther north. The southernmost occurrence of hypersthene in the Blue Ridge belt so far mentioned in the literature is in northern Floyd County, Va. (Dietrich, 1959).

Uranium-lead age determinations on zircon from the Cranberry Gneiss and related rocks in the region indicate that the Cranberry Gneiss formed 1,000 to 1,100 m.y. ago, at about the same time as basement rocks in the Grandfather Mountain window (Davis and others, 1962); one potassium-argon age determination of 1,130 m.y. on hornblende (Thomas, 1963) indicates that the date of metamorphism of the schists, gneisses, and amphibolites was approximately the same. Whether the granitic rocks intrusive into the Cranberry Gneiss were emplaced at that time or somewhat later is uncertain. Isotopic ages of zircon from the Beech Granite in northeastern Tennessee and from the granite just west of the window are all discordant and range from 360 to 820 m.y. (Davis and others, 1962). Unfortunately, no isotopic age determinations are available for zircon from Cranberry Gneiss near those localities. The granites might have been emplaced 800 to 900 m.y. ago, rather than immediately after the metamorphism 1,000 to 1,100 m.y. ago.

If some of the metamorphic rocks of the Blue Ridge thrust sheet are part of the Ashe Formation, as recently suggested by Rankin (1970) and Hadley (1970), these rocks would have been subjected only to the Paleozoic events described below. Field relations described above suggest that at least some of the metamorphic rocks are older, as they are all considered to be in our interpretation.

In Canada, many potassium-argon dates on micas from the Grenville province have a frequency peak between 930 and 940 m.y. but have considerable scatter. On the basis of these ages, a "Grenville Orogeny" is placed in the age range from 880 to 1,000 m.y. (Stockwell, 1964, fig. 2). A compilation of isotopic ages of uraninite, zircon, and cleveite from southern Canada and northern New York State shows a peak at about 1,030 m.y. (Tilton and others, 1960, table 3, samples 1–7). Potassium-argon ages on mica are lower than the ages on the zircon from the same rock (Tilton and others, 1960, table 4, sample 5), and rubidium-strontium ages are generally greater than the potassium-argon ages. These data might be interpreted to show that the "Grenville Orogeny" lasted 150 m.y. or that the orogenic belt was not uplifted and cooled until 150 m.y. after the beginning of the orogeny. We prefer to consider that the "Grenville Orogeny" in the southern Appalachians occurred 1,000 to 1,100 m.y. ago, and we suggest that the apparent younger age of that event in the Grenville province of the Canadian Shield is due to the use of only potassium-argon ages of micas in determining the age of the orogeny.

The Bakersville Gabbro and related mafic rocks intruded the rocks of the Blue Ridge thrust sheet after this metamorphism and before the next, as Kulp and Poldervaart (1956) recognized. It is tempting to correlate the Bakersville with the Linville Metadiabase in the Grandfather Mountain window because of general similarity in lithology and age, but there is no direct evidence for such a correlation.

Potassium-argon ages of micas from several places in the Blue Ridge thrust sheet suggest that there may have been an episode of metamorphism between 450 and 550 m.y. ago (Kulp and Eckelmann, 1961). One of these samples is from about 10 miles west of the Grandfather Mountain area; another is at Ore Knob 10 miles northeast of the area (Thomas, 1963).

Medium-grade dynamothermal metamorphism during the middle Paleozoic (pl. 6*C*) has produced kyanite, staurolite, biotite, muscovite, garnet, and calcic oligoclase or sodic andesine in schist and gneiss; epidote, monoclinic pyroxene, and andesine-labradorite in calc-silicate rocks; and diopside, epidote, hornblende, and calcic oligoclase or sodic andesine in amphibolites. An attempt to construct an ACF diagram for these medium-grade rocks was unsuccessful, probably because of their polymetamorphic character. Potassium-argon ratios in micas from the Spruce Pine district indicate that this metamorphism occurred about 350 m.y. ago. The pegmatite and granodiorite may have been emplaced prior to or during an early stage of this metamorphism.

Whether the Cranberry Gneiss underwent the medium-grade metamorphism during the middle Paleozoic is unknown, but if it did not, the existence of a major fault between it and the structurally overlying rocks would be proved, as the overlying rocks are of medium grade (pl. 6*D*). Ages of micas from basement rocks in the intermediate sheet between the Blue Ridge thrust sheet and the Mountain City window are inconsistent but range from 890 to 660 m.y., showing that those rocks were never subjected to medium-grade metamorphism during the Paleozoic (Long and others, 1959; Davis and others, 1962). Biotite from rocks mapped in the Blue Ridge thrust sheet in northeastern Tennessee by Rodgers (1953a) gave potassium-argon ages in the range from 380 to 695 m.y., and one sample gave a rubidium-strontium age of 715 m.y., showing that the biotite in those rocks was not completely recrystallized during the Paleozoic. No dates are available from the Cranberry Gneiss of the Blue Ridge thrust sheet in the Grandfather Mountain area; the nearest date in Tennessee is from a locality about 5 miles away. Biotite from medium-grade Cranberry Gneiss on Roan Mountain, also about 5 miles west of the Grandfather Mountain area, has a potassium-argon age of 357 m.y., indicating a Paleozoic history for the rock similar to that of the main part of the Spruce Pine district.

If a major fault exists between the rocks of low metamorphic grade and those of medium grade, it may connect with the Gossan Lead and Fries faults of southwestern Virginia (Stose and Stose, 1957, map), as was shown on the 1944 tectonic map of the United States (King and others, 1944). The same fault should lie between Roan Mountain and the belt of iron prospects in northeastern Tennessee (Bayley, 1923). No information is available on the possible position of this fault west of the Spruce Pine district, but the geologic map of the Roan Mountain quadrangle (Keith, 1907a) indicates that it might be found near Hughes Gap on the North Carolina-Tennessee State line and somewhere between Red Hill and Relief in North Carolina.

If the Ashe Formation extends into the northern part of the Grandfather Mountain area, as suggested by Rankin (1970), we think that the map and structural relations along the Cranberry contact necessitate a major fault. A considerable thickness of the Ashe is cut out in a southeasterly direction along the contact in much too sudden a manner to represent a basin margin. Coarse-grained clastic rocks are absent in the metamorphic rocks along this contact.

Most of the plutonic rocks in the Blue Ridge thrust sheet in the Grandfather Mountain area have been incompletely metamorphosed at low grade (pl. 6*A*, *B*). New albite, microcline, biotite, and sericite occur in rocks containing sufficient potassium, whereas chlorite, actinolite, epidote, and albite occur in rocks deficient in that element. Some of the sericite is of the green iron-rich variety. Granitic rocks have been incompletely metamorphosed in most places and contain numerous porphyroclasts of earlier minerals in a matrix of later finer grained minerals produced by cataclastic breakdown and subsequent recrystallization of minerals of the plutonic generation. Most of the rocks are blastomylonitic or phyllonitic gneisses. Locally, all traces of the plutonic minerals have been destroyed to produce blastomylonite and, under certain conditions, phyllonite. Mafic igneous rocks contain rare relict pyroxene, and the arenaceous rocks in the fault slices have relict clastic texture.

The low-grade mineral assemblages and textures are the same as those of the Grandfather Mountain window.

The boundary between the incompletely metamorphosed rocks of low grade and the rocks of medium grade is approximately at the contact between the Cranberry Gneiss and the tectonically overlying mica schist and amphibolite, except at the west margin of the Linville quadrangle, where the boundary is in the Cranberry Gneiss. Near this contact, the medium-grade rocks are partly crushed and contain retrogressive mineral assemblages.

No date for the low-grade metamorphism has been established. The well-developed northwest-trending cataclastic mineral lineation and the cataclastic foliation subparallel with the Linville Falls fault suggest that the metamorphism may have occurred during some phase of thrusting. Thus, it probably occurred later than the metamorphism 350 m.y. ago and earlier than the end of thrusting in the late Paleozoic. (See below for discussion of the age of thrusting.)

The tight and isoclinal folds in the rocks of the Blue Ridge thrust sheet are inferred to have formed during an early stage of thrusting because of the similarity in pattern between them and similar folds in the Table Rock thrust sheet (discussed below). They seem to have formed by tightening, flattening, and passive rotation of earlier more open folds originally formed perpendicular to the direction of transport of the thrust sheet (Bryant and Reed, 1970, b) The crinkles and open folds may have formed at a late stage of thrusting or immediately afterwards. Their orientation in relation to the mineral lineation indicates a response to similarly oriented stress.

Adjacent to the Linville Falls fault southeast of the window in the Blowing Rock quadrangle, a narrow zone of biotite-muscovite schist and gneiss has polymetamorphic textures and has been recrystallized to medium grade. These rocks contain porphyroclasts of muscovite in a fairly fine grained matrix of recrystallized quartz, oligoclase, muscovite, and biotite. Cataclasis and recrystallization probably took place during thrusting. Most of the rocks southeast of the Grandfather Mountain window in the Blue Ridge thrust sheet have been overprinted by structural and metamorphic effects associated with the Brevard fault and described below.

Table 30 (p. 172) summarizes the structural and metamorphic events and their established or inferred dates and sequence.

#### INTERMEDIATE SHEET

In the intermediate sheet, which is mapped as Cranberry Gneiss in the northwestern part of the Linville quadrangle (pl. 1) and which lies tectonically below the Blue Ridge thrust sheet and above the Mountain City window, layering of the plutonic generation is cut by southeast-dipping phyllonite zones. We do not have enough information to deduce the pattern formed by the layering. Much of the rock has not undergone shearing and retrogressive metamorphism, and the pervasive northwest-trending mineral lineation characteristic of the adjacent rocks of the Blue Ridge thrust sheet is poorly developed. Some northeast-trending crinkles are found in the phyllonite zones.

Presumably, the plutonic rocks in the sheet were formed 1,000 to 1,100 m.y. ago, at the same time as those in the Blue Ridge thrust sheet and the Grandfather Mountain window. Paleozoic metamorphism was local and incomplete, and the partial recrystallization was in the chlorite zone. Biotite from a partly altered and somewhat sheared rock on Dark Ridge Creek gave a rubidium-strontium age of 810 m.y. and a potassium-argon age of 660 m.y. (Davis and others, 1962).

## GRANDFATHER MOUNTAIN WINDOW

The oldest rocks in the Grandfather Mountain window are layered migmatitc gneiss and metamorphosed diorite and gabbro that grade into and are cut by granitic rocks of Precambrian age. In the northwestern part of the window, these plutonic rocks are unconformably overlain by the Grandfather Mountain Formation, a thick sequence of interlensing arkose, siltstone, shale, and volcanic rocks of late Precambrian age. The bedded rocks occupy the southeastern limb of a complex synclinorium in which at least two generations of structures are superimposed. The northwestern limb is concealed beneath the Blue Ridge thrust sheet northwest of the window. The basement rocks are intruded by bodies of metadiabase and felsic porphyry which probably correlate with extrusive rocks in the bedded sequence. The granitic rocks were pervasively sheared, retrogressively metamorphosed, and locally converted to phyllonite and blastomylonite at the same time that the overlying sedimentary and volcanic sequence was progressively metamorphosed. Both the granitic rocks and the overlying sequence are apparently autochthonous.

Also exposed in the window are Cambrian and Cambrian(?) quartzite and phyllite of the Chilhowee Group and the overlying Lower Cambrian Shady Dolomite. These rocks form the Tablerock thrust sheet, which has ridden at least 15 miles northwestward over the autochthonous rocks beneath. The Tablerock thrust sheet occupies much of the southwestern part of the window and contains minor structures geometrically related to those in the overlying Blue Ridge thrust sheet. Rocks in the Tablerock thrust sheet have been metamorphosed at the same grade as the underlying autochthonous rocks.

### AUTOCHTHONOUS ROCKS

#### METAGABBRO AND METADIORITE

Small bodies of greenish-gray gneissose to phyllonitic metagabbro and metadiorite crop out in a few places among the plutonic basement rocks in the Grandfather Mountain window. Along Brown Creek and near Upton (areas H–3, G–4, and G–5 pl. 1) several bodies are large enough to map. These bodies are not well exposed, and the relations of the Brown Creek body are obscure enough to make its correlation with the other bodies uncertain. Interior parts of some of these bodies resemble the Linville Metadiabase, but they have gradational boundaries with adjacent granitic rocks and are cut by quartz-feldspar pegmatites. The metagabbro and metadiorite are older than the adjacent granitic rocks, but their age relations to nongranitic layers in layered Wilson Creek Gneiss are not known.

The rocks lack obvious relict igneous textures. They are composed of saussuritized plagioclase, hornblende, quartz, actinolite, and chlorite in proportions that depend on the original composition of the rock and the degree of shearing and recrystallization during the latest metamorphism. Sphene, opaque minerals, and apatite are accessory minerals. A few grains of relict pyroxene occur in hornblende in the Brown Creek body.

### WILSON CREEK GNEISS

The Wilson Creek Gneiss (Bryant, 1962) is a metamorphosed plutonic complex of Precambrian age occupying much of the eastern part of the Grand father Mountain window. Various parts of the Wilson Creek Gneiss were mapped by Keith (1903) as Cranberry Granite, Carolina Gneiss, Beech Granite, metarhyolite, and Unicoi Formation. The Wilson Creek Gneiss consists principally of nonlayered plutonic rock having compositions ranging from diorite to granite and averaging quartz monzonite. Parts of the gneiss are layered, and in areas I–5, J–4, and K–4, the predominantly layered rocks are distinguished on the geologic map (pl. 1). Areas where the gneiss is predominantly dioritic are also separated on the map but have very indefinite contacts (areas I–4, I–3, J–3, J–4). Phyllonite, blastomylonite, and mylonite, mostly derived from Wilson Creek Gneiss are mapped separately in areas I–3 and I–4, and a small area of more uniform quartz monzonite in the Wilson Creek Gneiss near Rose Mountain is delineated in areas E–7, E–8, F–7, and F–8.

The Wilson Creek Gneiss forms large outcrops along Wilson Creek and many other major streams on the Blue Ridge front. Locally, the gneiss is difficult to distinguish from coarse-grained nonbedded arkose of the Grandfather Mountain Formation because both rock types are pervasively sheared and the arkose lacks bedding. However, the gneiss generally contains white quartz-plagioclase-potassic feldspar pegmatites, whereas the arkose contains only segregations of quartz or quartz and pink microcline. Much of the Wilson Creek Gneiss contains biotite, which the arkose lacks.

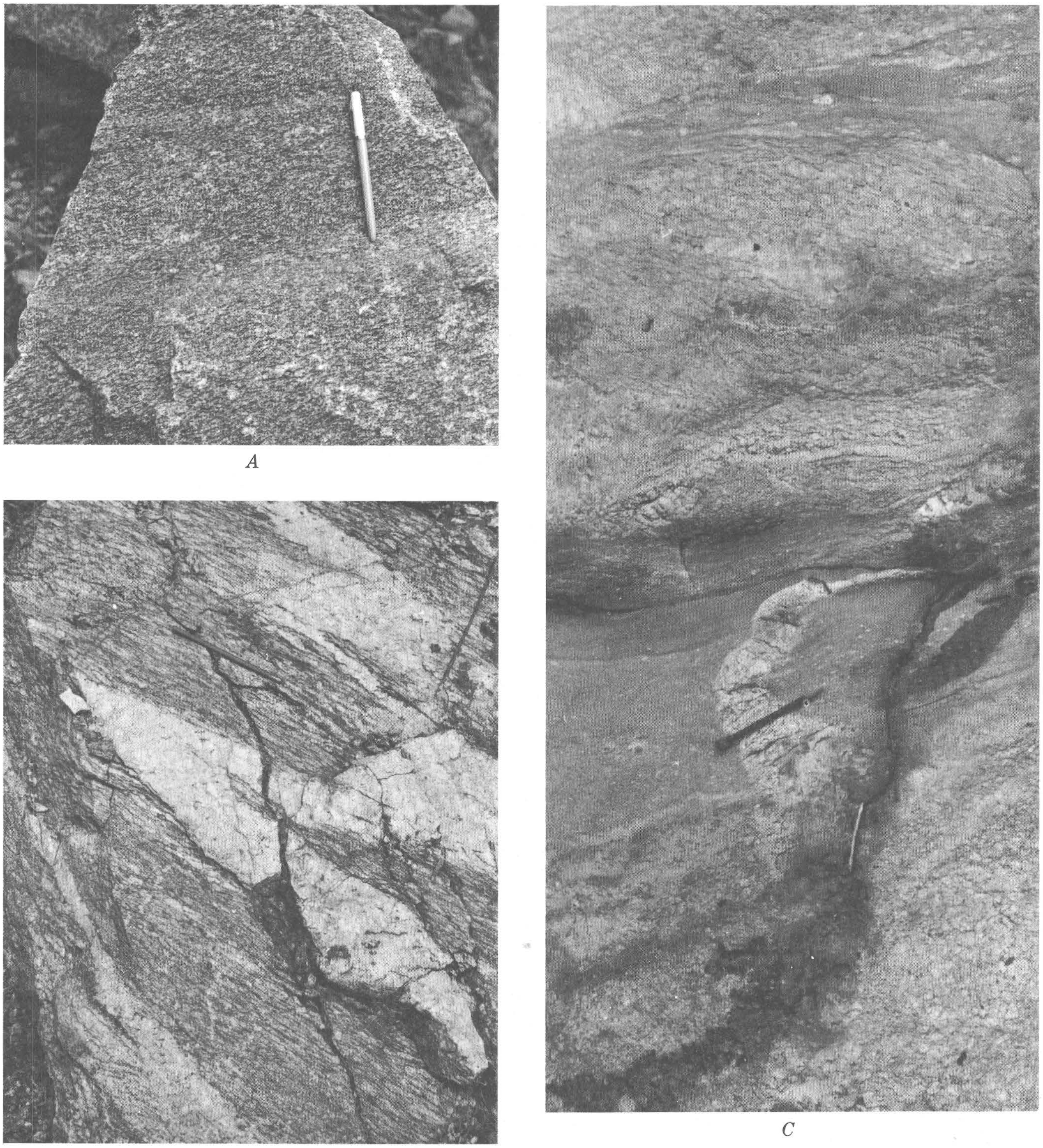
#### MEGASCOPIC FEATURES

The main body of Wilson Creek Gneiss has a variety of aspects because of shearing and recrystallization. It ranges from almost massive, slightly crushed granitic rock to phyllonite and blastomylonite that locally resemble phyllite and volcanic rock. Typically, it is light-gray to greenish-gray medium- to coarse-grained biotite-quartz-feldspar gneiss (fig. 34). The gneiss has a conspicuous cataclastic foliation defined by folia of fine-grained sericite and biotite and a well-developed cataclastic lineation formed by streaks of mica, quartz, or feldspar trains of alined mica, and grooving. Plagioclase-rich gneiss generally his stronger foliation than gneiss rich in potassic feldspar because of decomposition of calcic plagioclase into albite, sericite, and epidote during Paleozoic retrograde metamorphism. The gneiss locally contains layers and pods of biotite schist and biotite amphibolite.

Dikes and pods and irregular bodies of mica-poor quartz-feldspar pegmatite are widespread in the Wilson Creek Gneiss. They range from a few inches to more than 100 feet in thickness and contain feldspar crystals as much as 4 inches long. Some have sharp contacts with the surrounding granitic rock, but other contacts are gradational. The pegmatite consists of quartz, perthite, and plagioclase. It locally contains books of muscovite as much as 1 inch in diameter, but they are rare. Allanite and garnet locally occur as accessory minerals. Generally, the pegmatite seems to be little foliated, especially in exposures perpendicular to the cataclastic mineral lineation, but in many places it has crude foliation despite its lack of mica. Locally, it is an augen gneiss. In some places, the gneiss is cut by thin dikes of fine- to medium-grained light- to medium-gray quartz monzonite in which foliation is parallel to that in the surrounding coarser grained gneiss. These dikes cut some of the pegmatites and are in turn cut by others.

In addition to the pegmatites, the gneiss contains segregation lenses, stringers, and veins of quartz, quartz and chlorite, and rarely, quartz and pink microcline or quartz, calcite, pink microcline, and chlorite. Quartz veinlets are found along joints in some rocks. Veinlets of epidote are found in some of the more plagioclase-rich rocks.

Angular blocks of finely layered amphibole gneiss occur as xenoliths in a coarse pegmatitic phase of the

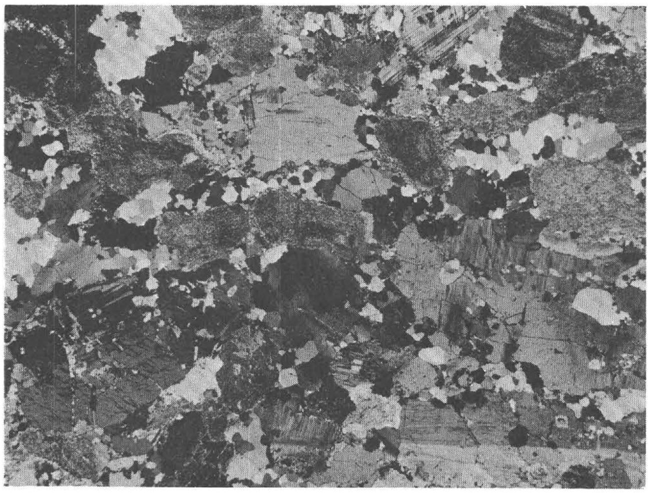


A

B

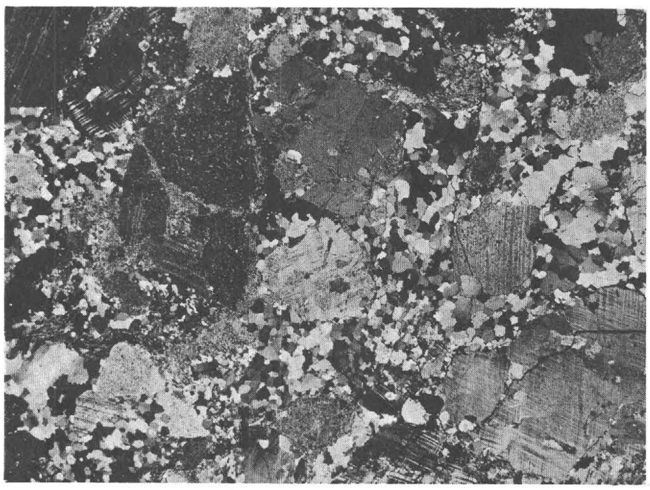
C

FIGURE 34.—Wilson Creek Gneiss. *A*, Typical cataclastic granitic gneiss. From roadcut on east side of North Carolina Highway 181 about 0.5 mile south of bench mark 2868 on Ripshin Ridge (area E–6, pl. 1). *B*, Cataclastic granitic gneiss and pegmatite. From road-metal quarry on east side of Wilson Creek about 0.5 mile northwest of Mortimer (area F–6, pl. 1). This is the location of the sample from which zircon was dated by Davis, Tilton, and Wetherill (1962). *C*, Blastomylonite zone in granitic gneiss and pegmatite. From outcrop along Wilson Creek at 1,760-foot altitude (area F–5, pl. 1). View subperpendicular to cataclastic lineation. Note gradations to blastomylonite containing feldspar porphyroclasts.



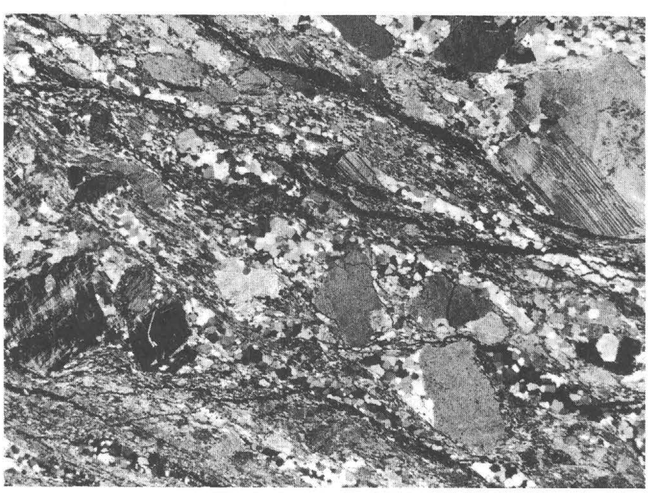
*A*

*D*

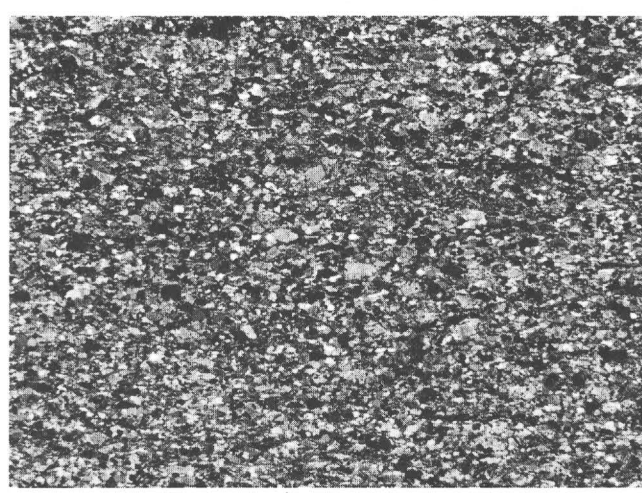


*B*

*E*



*C*



*F*

5 mm

FIGURE 35.—Photomicrographs of Wilson Creek Gneiss. *A*, Recrystallized quartz monzonite mortar gneiss from south fork of Harper Creek 1 mile downstream from Kawana (area F–6, pl. 1). Granoblastic texture with zones of recrystallized quartz mortar. Poikilitic inclusions of quartz and plagioclase in microcline. Microcline bent, broken, and healed by quartz. Larger quartz grains bent and partly broken down into mosaic-textured areas. Plagioclase saussuritized. *B*, Blastomylonitic quartz monzonite gneiss from Harper Creek at the Avery-Caldwell County line (area F–6, pl. 1). Porphyroclasts of plagioclase (pseudomorphosed by albite), microcline, and quartz surrounded by mosaic-textured quartz and recrystallized biotite, sericite, and albite. Section perpendicular to mineral lineation. *C*, Blastomylonitic biotite-epidote-sericite quartz monzonite gneiss from quarry on Bark Camp Ridge (area F–5, pl. 1). Bent and broken porphyroclasts of plagioclase (pseudomorphosed by albite) and microcline in a groundmass of recrystallized quartz, sericite, albite, biotite, and epidote. Section parallel with mineral lineation. *D*, Porphyroclastic mylonite from east end of hill 3715 about 1 mile southeast of Dry Pond Gap (area I–3, pl. 1). Crushed and strung-out porphyroclasts of plagioclase (pseudomorphosed by albite), potassic feldspar, and quartz in a matrix of quartz, feldspar, and recrystallized iron-rich muscovite. Section parallel to mineral lineation. *E*, Porphyroclastic blastomylonite from outcrop along Wilson Creek at an altitude of 1,760 feet (area F–5, pl. 1). From specimen of fine-grained rock shown in figure 34*C*. Small porphyroclasts of saussuritized plagioclase and biotite in a matrix of recrystallized quartz, sericite, biotite, and epidote. Section perpendicular to lineation. *F*. Blastomylonite from an altitude of about 2,600 feet in tributary valley east of Yadkin River just east of Blowing Rock Gneiss contact and north of Watauga-Caldwell County line (area H–4, pl. 1). Quartz, albite, sericite, biotite, and subordinate microcline.

Wilson Creek Gneiss on Upper Creek 0.5 mile above the mouth of Burnthouse Branch (area E–6, pl. 1).

### PETROGRAPHY

The typical Wilson Creek Gneiss (fig. 35*B* and *C*) is composed of porphyroclasts of microcline, plagioclase, biotite, and quartz set in a matrix of fine-grained quartz, plagioclase, sericite, microcline, biotite, chlorite, and epidote. The porphyroclasts are a few millimeters to as much as 1 cm in diameter; matrix grains range from 0.01 to 0.5 mm in diameter, but most are about 0.1 mm. More mafic varieties of gneiss contain porphyroclasts of amphibole. Muscovite porphyroclasts are rare.

The potassic feldspar forms conspicuous porphyroclasts, some as much as 5 cm in diameter. Rock containing such large porphyroclasts resembles the Blowing Rock Gneiss (see below). The common grain size for the potassic feldspar porphyroclasts is 5 mm. They contain inclusions of plagioclase, quartz, and mica. The porphyroclasts are bent and broken, and the fractures are healed by quartz, sericite, albite, biotite, carbonate, or a combination of two or three of these minerals (fig. 35*B* and *D*). Most of the potassic feldspar is microcline, but some is fine-textured string perthite.

Quartz granulates and recrystallizes more easily than the feldspar in many rocks. It forms mosaic-textured areas between feldspar grains even in relatively unsheared rocks (fig. 35*A* and *B*). In places the quartz is concentrated in lenses and stringers which may represent strung-out and recrystallized larger grains. Porphyroclasts, which are as much as 8 mm in diameter, are strongly strained. Many are brecciated or partly granulated and recrystallized into mosaic-textured aggregates or recrystallized grains or mortar.

Plagioclase forms porphyroclasts as much as 8 mm in diameter. They are commonly bent and locally are broken and healed by the minerals of the recrystallized matrix. Albite twinning is widespread, pericline twinning, common, and Carlsbad twinning rare. Almost all the plagioclase grains have been saussuritized, except those in a few rocks in the tectonic slice of Wilson Creek Gneiss above the Tablerock thrust sheet in North Cove (area A–10, pl. 1). There, relict andesine and oligoclase occur. In a few places the altered porphyroclasts have a rim of clear albite. In more sheared rocks the albite, sericite, and epidote derived from breakdown of the plagioclase become separated in varying degrees into clear recrystallized albite, lepidoblastic aggregates of sericite, and discrete grains of epidote.

Brown or reddish-brown biotite forms porphyroclasts as much as 2 mm in diameter. Biotite porphyroclasts are commonly bent and have sagenitic structure. They are generally partly altered to chlorite or lighter colored biotite. Biotite in the matrix is brown, greenish brown, or brownish green and occurs in undeformed flakes about 0.2 mm long. A few flakes are as much as 0.5 mm long.

Most muscovite is fine grained and is derived from breakdown of feldspars, especially plagioclase, but a few bent and partly altered porphyroclasts occur. Limited X-ray data indicate that paragonite is not present in detectable amounts. The new muscovite is mostly synkinematic, but some is postkinematic. It forms flakes generally 0.1 mm long but as much as

0.3 mm long. The new mica in some rocks is green and has a high index of refraction; these properties indicate that the mica is similar to the green iron-rich muscovite so widespread in the Grandfather Mountain Formation. Both the new biotite and muscovite are concentrated in segregations along cleavage planes.

Anhedral grains and aggregates of epidote are widespread in the groundmass of the gneiss. Grain size averages 0.05 to 0.1 mm but reaches as much as 0.5 mm, especially in segregation veinlets. Epidote seems to have been derived mainly from plagioclase. Locally, it has a light-brown absorption, indicating some rare-earth content.

Mafic rocks contain porphyroclasts of hornblende or actinolite as much as 2 mm long. Hornblende commonly has a rim of actinolite. Recrystallized amphibole is actinolite.

Garnet is found locally as an accessory mineral in the Wilson Creek Gneiss. It occurs in the groundmass as subhedral to euhedral crystals that seem to be contemporaneous with the rest of the groundmass minerals, and it also occurs as earlier inclusions in plagioclase porphyroclasts. Some of the garnet has a light yellow absorption.

Chlorite is a major constitutent of a few of the mafic rocks. Most of it is drived from biotite. It is generally green and is mostly the FeMg variety, although MgFe chlorite also occurs.

Other accessory minerals are sphene, ilmenite, leucoxene, magnetite, apatite, zircon, stilpnomelane, carbonate, allanite, and fluorite. Zircon ranges from round to euhedral.

Normal Wilson Creek Gneiss ranges in composition from diorite to granite (fig. 36*A*). Pegmatites range from somewhat altered rocks containing only partly chloritized biotite or saussuritized plagioclase to slightly cataclastic rocks containing a small amount of recrystallized mortar to blastomylonitic gneiss (fig. 37). They range in composition from granite to quartz diorite (fig. 36*B*), but they lack biotite, whereas the Wilson Creek Gneiss typically has a few percent of that mineral. The fine-grained discordant dikes are similar in composition and texture to the pegmatites. Most compositional varieties of the Wilson Creek Gneiss are so irregularly distributed that they cannot be mapped separately.

### LAYERED GNEISS

A conspicuously layered phase of the Wilson Creek Gneiss crops out over an extensive area in the northeastern part of the window (pl. 1). Contacts with typical nonlayered Wilson Creek Gneiss are almost everywhere gradational, and layers are found in the Wilson Creek Gneiss at scattered outcrops as much as 2 miles away from mapped units of layered gneiss. The lens of nonlayered Wilson Creek Gneiss in Chestnut Mountain (area I–5, pl. 1) however, has sharp contacts with the enclosing layered gneiss. The two bodies of layered gneiss mapped in the central Blowing Rock quadrangle (areas I–3 and H–4, pl. 1) are not well exposed. They are phyllonitic and more mafic than the adjacent rock.

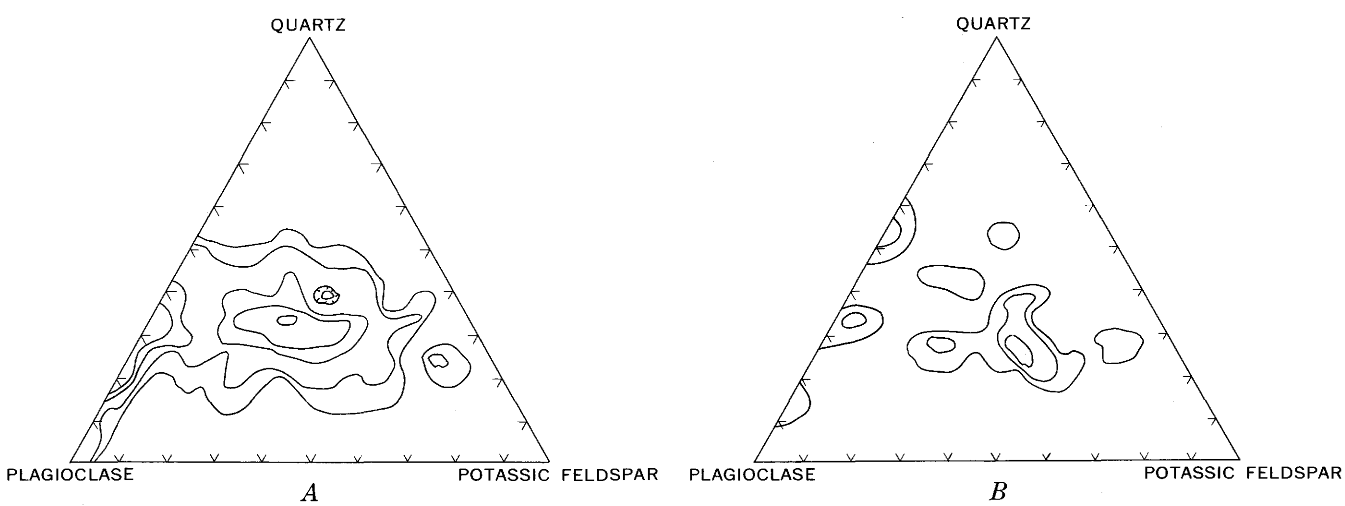



FIGURE 36.—Proportions of quartz, plagioclase, and potassic feldspar in Wilson Creek Gneiss. Epidote and sericite counted as plagioclase. *A*. Typical granitic gneiss. Based on count of 50 random grains in 116 thin sections of rock retaining well-defined plutonic aspect. Contours 1, 2, 4, 6, and 8 percent. *B*, Pegmatite and dikes of granitic rock cutting typical gneiss. Based on count of 50 random grains in each of 19 thin sections. Contours 5, 10, and 15 percent.

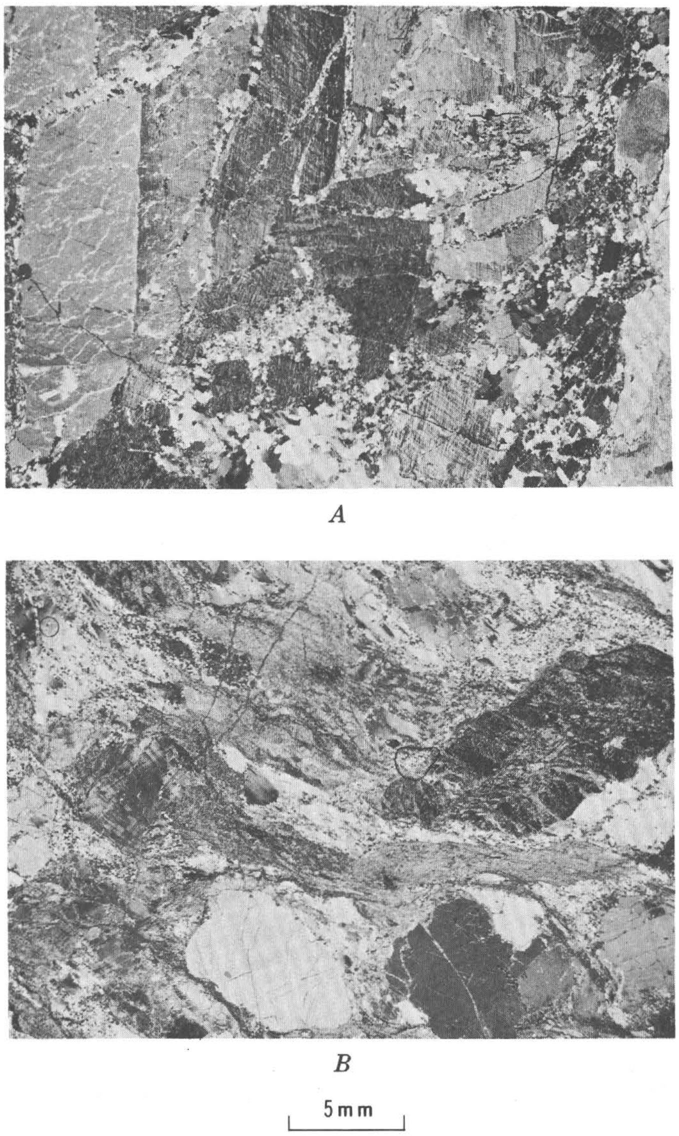



FIGURE 37.—Photomicrographs of pegmatite in Wilson Creek Gneiss. *A*, Sheared pegmatite from discordant dike 8 inches thick in roadcut along North Carolina Highway 181 at the head of Ripshin Branch (area E–6, pl. 1). Large crystals of perthitic microcline, plagioclase (pseudomorphosed by albite), and quartz are bent, broken, and healed by quartz, albite, epidote, sericite, and calcite. Section at 45° to mineral lineation. *B*, Sheared pegmatite from railroad cut east of Woodlawn (area B–10, pl. 1). Porphyroclasts of quartz partly broken down and recrystallized into mosaic-textured aggregates. Porphyroclasts of perthitic microcline bent and broken. Porphyroclasts of plagioclase saussuritized and smeared out. Section perpendicular to mineral lineation.

The layered gneiss consists of interlayered and intergrading plagioclase porphyroclast gneiss, blastomylonitic granitic gneiss, biotite-chlorite schist, muscovite-biotite schist, biotite gneiss, epidote amphibolite, biotite amphibolite, hornblende gneiss, and pegmatite. Layers are one-half of an inch to tens of feet thick. Late segregation lenses are composed of quartz, quartz and pink microcline, quartz and chlorite, and quartz, calcite, biotite, and chlorite.

The layered Wilson Creek Gneiss is similar to much of the Cranberry Gneiss of the Blue Ridge thrust sheet, but the irregular and gradational contact between the layered and the typical Wilson Creek Gneiss indicates that the layered gneiss is inside the window. Cranberry Gneiss may have been adjacent to and southeast of presently exposed layered Wilson Creek Gneiss before thrusting.

The granitic layers are similar to the bulk of the nonlayered Wilson Creek Gneiss, except that directly adjacent to the edge of the window they are more coarsely recrystallized. Most of the biotite schist layers have been completely recrystallized, but in hornblende gneiss and amphibolite, much of the amphibole is porphyroclastic and unaltered or only partly altered to actinolite. Plagioclase is mostly altered to albite even where preserved as porphyroclasts.

### QUARTZ MONZONITE GNEISS

One area of little-foliated coarse-grained light-colored quartz monzonite in the Wilson Creek Gneiss is mapped near Rose Mountain (areas E–7, E–8, F–7, F–8, pl. 1). Other unmapped bodies of similar aspect may occur in the Wilson Creek Gneiss, but if so, they are probably less than a few hundred yards in diameter. The rock is a blastomylonitic gneiss or recrystallized mortar gneiss containing porphyroclasts of microcline, microcline-microperthite, quartz, and saussuritized plagioclase several millimeters in diameter in a recrystallized groundmass of mosaic-textured quartz and minor amounts of brownish-green biotite and sericite. The scarcity of mica accounts for the poor development of foliation. Other accessory minerals are epidote, zircon, fluorite, chlorite, allanite, and opaque minerals.

### PHYLLONITE AND BLASTOMYLONITE

Phyllonite and blastomylonite are found in zones a few inches to thousands of feet wide throughout the Wilson Creek Gneiss, but they are most abundant in the southeastern part of its outcrop area. In the central part of the Grandfather Mountain window west of the belt of Blowing Rock Gneiss, phyllonite generally occurs in well-defined zones concordant with the foliation in the enclosing gneiss, but the zones are discontinuous both laterally and vertically.

East of the outcrop belt of the Blowing Rock Gneiss, the degree of shearing and recrystallization is somewhat greater, and blastomylonites and phyllonites are so widely distributed and intimately mixed with recognizable plutonic rocks that they generally cannot be mapped separately.

Near their margins, the phyllonite zones commonly contain lenses, layers, and pods of less-sheared gneiss and relatively unsheared pegmatite and porphyroclasts of quartz, feldspar, and mica. In many places, the phyllonites lack these relicts and cannot be distinguished from phyllites. The phyllonite is generally medium gray but may be silvery gray, black, or greenish gray. It commonly weathers brown or red. Locally, phyllonite forms the matrix of a breccia composed of gneiss and pegmatite fragments. The phyllonite zones locally contain disseminated graphite or pyrite. The graphitic zones are a few inches to a few feet thick, and individual flakes of graphite are as much as 1 mm in diameter. Some phyllonite zones only a few inches thick are lens shaped when viewed perpendicularly to the lineation. Some zones only 1 to 3 mm thick are as much as 1 foot long parallel with the lineation. The larger discontinuous phyllonite zones may also be elongate parallel with the lineation.

The phyllonites in many places have two lineations at approximately right angles, one formed by elongated and alined mineral grains and the other formed by small crinkles in the foliation.

The phyllonites contain quartz segregation stringers and knots. Epidote knots containing pyrite and sphalerite(?) occur in a few places. Locally, the phyllonites have been enriched by iron in the form of hematite, magnetite, or pyrite, by titanium in the form of ilmenite, and by uranium in the form of torbernite (perhaps originally uraninite).

The phyllonites and blastomylonites contain 30 to 80 percent sericite, 4 to 60 percent quartz, 0 to 30 percent biotite, 0 to 30 percent chlorite, and 0 to 25 percent epidote, depending on the composition of the parent rock. All gradations between gneiss and phyllonite are found. A typical phyllonite contains about 30 percent quartz, 60 percent sericite, and the remaining 10 percent opaque minerals, biotite, chlorite, epidote, and sphene. In some phyllonites, the only mafic mineral is magnetite or ilmenite, but it is finely distributed and gives the rock a dark-gray color. Other accessory minerals are pyrite, calcite, stilpnomelane, zircon, apatite, tourmaline, allanite, pyrrhotite, and porphyroclastic garnet. Quartz, potassic feldspar, plagioclase, biotite, muscovite, or amphibole form porphyroclasts in some phyllonites.

The blastomylonite mapped in areas I–3 and I–4 (pl. 1) was in part mapped by Keith (1903) as metarhyolite. Some of the rocks closely resemble felsic volcanic rocks both in outcrop and thin section. However, along Bailey Camp Creek, gradations in texture between blastomylonitic gneiss and rocks resembling felsic volcanics can be seen. In other places, the blastomylonite contains relict lenses of phyllonitic and blastomylonitic gneiss and pegmatite.

The blastomylonite is a gray to dark-gray and green aphanitic rock locally containing carbonate or quartz-chlorite segregations. It contains porphyroclasts of quartz and feldspar in a matrix of sericite, quartz, and feldspar too fine grained to be accurately identified under the microscope (fig. 35*D*). Some of the rocks contain considerable amounts of chlorite and epidote and were apparently derived from diorite or quartz diorite. A few of the mafic rocks contain lath-shaped plagioclase grains and may therefore have been derived from mafic igneous rock. Locally, the quartz porphyroclasts are subhedral and give the blastomylonite the appearance of a volcanic rock. Some of the rocks mapped as blastomylonite may be volcanic rocks, but if so, they constitute only a minor and unmappable part of the unit. Blastomylonite is found locally in the Wilson Creek Gneiss outside the area mapped as blastomylonite.

### OTHER ROCKS

Light-colored gneiss and rocks somewhat resembling metamorphosed sedimentary rocks occur locally in the Wilson Creek Gneiss. These rocks are light green to white and are intimately associated with typical Wilson Creek Gneiss and phyllonite. Locally, they contain pegmatite and lenses of material resembling plutonic rock. They lack sedimentary structures and seem to grade into typical plutonic gneiss. The origin of these rocks is obscure, but they probably belong to the plutonic complex. Whether they are somewhat silicified blastomylonitic gneisses or some older sedimentary rock is unknown.

These rocks contain 40 to 75 percent quartz and 15 to 50 percent sericite. Some lack feldspar, but others contain as much as 25 percent microcline and 15 percent plagioclase. Rocks of similar aspect but containing more mica and less quartz were called phyllonites. All gradations are found between light-colored quartzitic gneisses and green to gray micaceous phyllonites.

An interesting but rare rock type is a pure granular white quartzite. It forms part of a prominent ledge on the east side of upper Wolfden Branch (area I–3, pl. 1) The outcrop can be seen from Thunder Hill on the Blue Ridge Parkway about 2 miles to the west. The quartzite caps the ledge and is as much as 30 feet thick; it extends about 500 feet along strike. Much of the quartzite has a coarse texture which is partly obscured by shearing. The quartzite locally grades to gneiss. The bulk of the outcrop below the quartzite is light-colored plutonic gneiss containing numerous stringers and pods of similarly sheared quartzite.

A layer of pure quartzite about 30 feet thick crops out on the east side of the valley of Buffalo creek west of Bradshaw school (area I–4, pl 1). The quartzite contains segregation veinlets of quartz along joints. The same layer is 10 feet thick where it crosses Spanish Oak Creek south of the school.

These rocks consist almost entirely of mosaic-textured quartz grains averaging about 0.5 mm in diameter. Rocks in somewhat similar bodies in the Linville Falls quadrangle contain some sericite. These quartzites are probably sheared and recrystallized quartz veins.

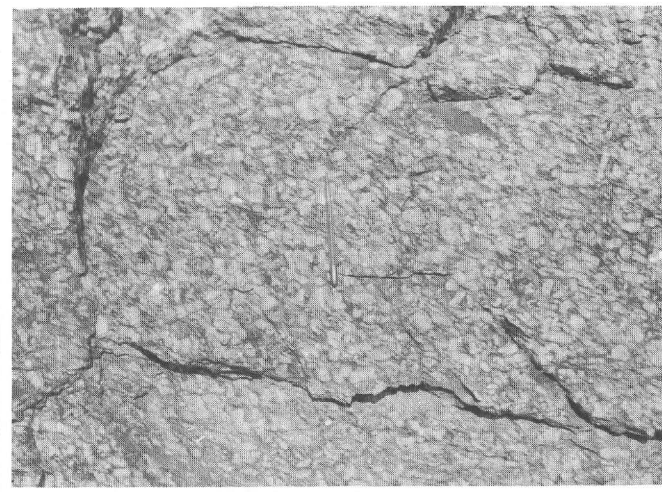
### ORIGIN AND AGE

Intense cataclastic metamorphism obscures the origin of the Wilson Creek Gneiss. The oldest textures preserved are granoblastic, but they are rare. The granitic gneiss has broadly gradational contacts with layered migmatitic schist, gneiss, and amphibolite, but locally the contacts are sharp. In a few places, blocky inclusions of hornblende gneiss occur in the granitic gneiss. Large areas of granitic gneiss lack layers or inclusions yet have various proportions of plagioclase and microcline. Apparently, the Wilson Creek Gneiss is a typical plutonic complex formed by a combination of metasomatic and igneous processes, but the relative importance of these processes cannot be evaluated.

Isotopic ages of zircon from the Wilson Creek Gneiss from the quarry along Wilson Creek in area F–6 (pl. 1) range from 640 to 1,000 m.y., indicating that the Wilson Creek Gneiss is at least 1,000 m.y. old (Davis and others, 1962).

## BLOWING ROCK GNEISS

The Blowing Rock Gneiss (Keith, 1903) occupies a belt 3 to 5 miles wide and 15 miles long in the central part of the Grandfather Mountain window (pl. 1). The rock is very coarse grained strongly foliated blastomylonitic or phyllonitic augen gneiss containing white porphyroclasts of potassic feldspar in a black biotitic matrix (fig. 38). The gneiss is generally quartz monzonite in composition, but granitic and granodioritic varieties are also present. It contains inclusions of finer grained, generally more



*A*

*B*

FIGURE 38.—Blowing Rock Gneiss. *A*, Typical Blowing Rock Gneiss. From roadcut on U.S. Highway 321 on northwest side of south fork of the New River about 0.3 mile west of Brown Creek (area H–3, pl. 1). Note inclusion of fine-grained mafic rock. *B*, Tectonic lenses of pegmatite in phyllonite derived from Blowing Rock Gneiss. Lenses, 1 to 2 feet thick. From railroad cut through ridge north of Green Branch in Tweetsie Amusement Park about 3.2 miles northeast of village of Blowing Rock (area H–3, pl. 1).

mafic rock and dikes and small bodies of light-colored granitic rock and pegmatite. In places it has been converted to phyllonite or blastomylonite.

The contact between the Wilson Creek Gneiss and the Blowing Rock Gneiss is generally gradational over a few tens to a few hundred feet, except where it is extensively modified by shearing. In the northeastern part of the Linville Falls quadrangle (area F–6, pl. 1), phyllonite occurs almost everywhere along the contact, which is steep and parallel with the regional cataclastic foliation. The contact seems to be a fault, but if so, the fault leaves the contact and is lost in the southwestern part of the Blowing Rock quadrangle (area G–5). Near the head of Jackson Camp Creek (area H–4) the contact is also highly sheared, but local irregularities and the lack of offset in the metadiabase to the north indicate that there is no major fault along it.

Gradational zones along contacts generally consist of Wilson Creek Gneiss and Blowing Rock Gneiss in various proportions. In some places, large potassic feldspar crystals characteristic of the Blowing Rock Gneiss appear gradually in the Wilson Creek Gneiss; in other places, fully developed Blowing Rock Gneiss occurs as patches and zones in the Wilson Creek Gneiss. Continuous exposures through gradational contact zones are lacking, but it is generally possible to locate contacts within a few hundred feet.

Relative ages of the Blowing Rock Gneiss and Wilson Creek Gneiss cannot be determined from contact relationships. Isolated pods of Blowing Rock Gneiss in Wilson Creek Gneiss may be either inclusions of older material or zones in the Wilson Creek Gneiss partly converted to Blowing Rock Gneiss. Both gneisses are cut by dikes of light-colored granitic rocks and pegmatite and are believed to be of the same general age.

The Blowing Rock Gneiss is less variable in composition than the Wilson Creek Gneiss. Most noticeable variations are in the content of potassic feldspar and biotite. The rock ranges from dark biotitic types containing scattered white porphyroclasts of potassic feldspar to light-colored augen gneiss containing a rather small amount of biotite.

Foliation in the Blowing Rock Gneiss is defined by folia of fine-grained mica that wrap around the large feldspar porphyroclasts. Lineation is formed by alinement of fine-grained mica streaks, elongate quartz and feldspar (especially plagioclase) grains, and crude alinement of potassic feldspar porphyroclasts.

Layering is rare in the Blowing Rock Gneiss, but it is found locally in area G–6 (pl. 1) and near the contact with the thin unit of layered Wilson Creek Gneiss in area I–3. Some layers consist of granitic gneiss like typical Wilson Creek Gneiss; others are fine-grained biotite gneiss.

Small inclusions of mafic rocks occur locally in the Blowing Rock Gneiss. The mafic rocks are biotite quartz diorite gneiss, epidote-plagioclase-biotite schist, hornblende-biotite schist, amphibolite, and biotite schist. The largest inclusions are at least several tens of feet long. One large inclusion of layered amphibolite is exposed south of the large roadcut on U.S. Highway 321 north of the Blue Ridge Parkway underpass (area H–3, pl. 1); smaller inclusions a few inches to a few feet long are also visible in the same exposure. The inclusions are generally tabular, and some contain scattered porphyroclasts of potassic feldspar. In places, inclusions of mafic gneiss grade to typical Blowing Rock Gneiss, and the matrix of the Blowing Rock Gneiss resembles the included mafic gneiss. Some of the areas mapped as Flattop Schist by Keith (1903) contain these earlier mafic rocks.

Pods, dikes, and lenses of pegmatite in the Blowing Rock Gneiss are a few inches to several feet thick and are crudely foliated. Some have straight contacts and are fracture controlled; others are irregular and discontinuous. They consist principally of quartz, perthite, and plagioclase. The average size of the larger grains is 2 inches, but the maximum is at least 8 inches. Muscovite is less than an inch in diameter and occurs only locally in the pegmatites.

Light-colored equigranular granitic rocks form bodies that range from a few inches in thickness to at least 150 feet in diameter and that cut the Blowing Rock Gneiss. The grain size of the granitic rocks ranges from 1 to 12 mm, and some grade to pegmatite.

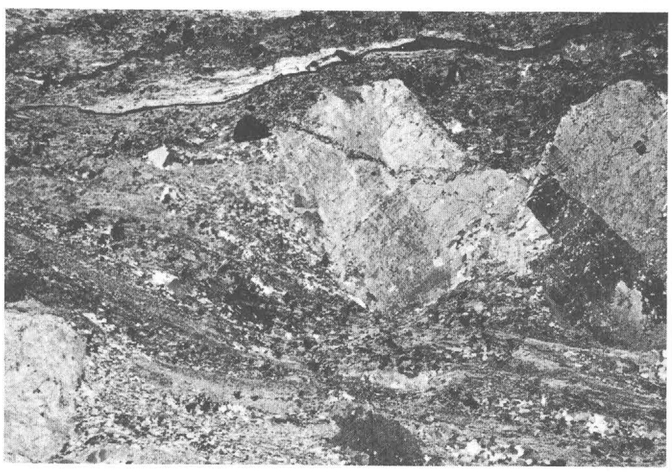
Quartz segregation pods, lenses, and veinlets a few millimeters to a foot thick are widespread in the Blowing Rock Gneiss. Calcite, chlorite, and locally occurring pyrite accompany quartz in many of the segregation veinlets. Veinlets of epidote are less widespread.

Locally, quartz veins as much as 10 feet thick are found; they have a granular texture, suggesting that they have been sheared and recrystallized.

Typical phyllonitic augen gneiss grades into gray or dark-blue-gray phyllonite which occurs in zones 1 to 20 feet thick parallel to the foliation. Zones of

*A*

5 mm



*B*

5 mm

*C*

5 mm

FIGURE 39.—Photomicrographs of Blowing Rock Gneiss. *A*, Porphyroclastic quartz monzonite augen gneiss from roadcut along U.S. Highway 321 just south of contact of Grandfather Mountain formation near the mouth of Goldmine Branch (area H–3, pl. 1). Porphyroclast of potassic feldspar containing patches of plagioclase in a matrix of saussuritized plagioclase, biotite, sericite, calcite, and quartz. A few small porphyroclasts of quartz. *B*, Typical porphyroclastic quartz monzonite augen gneiss from roadmetal quarry along U.S. Highway 321 south of Bailey Camp School (area H–4, pl. 1). Porphyroclasts of potassic feldspar contain patches of plagioclase. Plagioclase porphyroclasts represented by patches of strung-out saussurite. Small biotite porphyroclasts. Matrix composed of recrystallized quartz, sericite, biotite, epidote, and chlorite. Section cut parallel to mineral lineation. Analyzed specimen 1, table 13. *C*, Recrystallized quartz monzonite mortar gneiss from dike 6 feet thick in Blowing Rock Gneiss from roadcut along U.S. Highway 321 south of Brown Creek (area H–3, pl. 1). Perthitic microcline, saussuritized plagioclase, and partly broken down quartz grains in a matrix of recrystallized quartz, albite, and minor amounts of mica.

light-gray or green phyllonite may have been derived from the light-colored granitic rocks. The phyllonites weather red brown. Numerous phyllonite zones are exposed in roadcuts along U.S. Highway 321 north of the village of Blowing Rock (area H–3, pl. 1). In the valley of Matney Branch (northeast part of area H–3, Pl. 1), phyllonite derived from the Blowing Rock Gneiss is in contact with metamorphosed sedimentary and volcanic rocks from which it can be distinguished only with difficulty.

The Blowing Rock Gneiss is studded with porphyroclasts of potassic feldspar averaging ¾ to 1 inch in diameter and gray porphyroclasts of quartz averaging 3 to 5 mm in diameter; locally it contains porphyroclasts of biotite 2 to 3 mm in diameter. The porphyroclasts are set in a black to gray matrix of fine-grained recrystallized biotite, sericite, quartz, albite, and epidote. The grain size of the matrix is about 1 mm, but ranges from 0.01 to 0.5 mm (fig. 39*A* and *B*). Locally, the potassic feldspar porphyroclasts are as much as 3 inches long.

The potassic feldspar is perthitic microcline. It contains inclusions of plagioclase, quartz, and biotite. The microcline porphyroclasts are bent, broken, and healed by quartz, albite, epidote, biotite, and calcite (fig. 39*B*).

Plagioclase occurs as bent, broken, and saussuritized porphyroclasts averaging 0.5 to 3 mm in diameter but locally is as much as 1 cm long. Plagioclase grains included in microcline commonly have saussuritized cores but clear rims. The old altered grains and new recrystallized groundmass grains are both albite.

Quartz is found as strongly strained and drawn-out porphyroclasts and as mosaic-textured aggregates of recrystallized grains.

Biotite occurs in amounts ranging from a trace to 16 percent of the rock. It occurs as bent and partly altered brown porphyroclasts having sagenitic structure and as green or brown new synkinematic flakes 0.1 to 0.2 mm long in the groundmass.

Green FeMg chlorite derived from biotite occurs in some rocks. Locally, it composes as much as 12 percent of the rocks.

Sphene rims opaque minerals, probably ilmenite. It occurs in amounts ranging from 0 to 4 percent. Epidote constitutes 0 to 6 percent of the rocks. Allanite occurs in prisms as much as 1 mm long and as cores of epidote grains. Zircon is in euhedral prisms as much as 0.1 mm long and also as subhedral and anhedral grains. Other accessory minerals are apatite, ilmenite, pyrite, stilpnomelane, and calcite.

The very coarse and uneven grain size of the Blowing Rock Gneiss makes it difficult to obtain accurate modes. Figure 40 shows the proportions of quartz, plagioclase, and potassic feldspar in the Blowing Rock Gneiss and in the inclusions and the dikes cutting it. Because of intense alteration of plagioclase and redistribution of the alteration products in some rocks, the estimated proportion of plagioclase may be too low. (Note especially the mode of the chemically analyzed specimen 1, table 13.) Most rocks are in the quartz monzonite and granite ranges. The Blowing Rock Gneiss was not sampled extensively because petrographic classification based

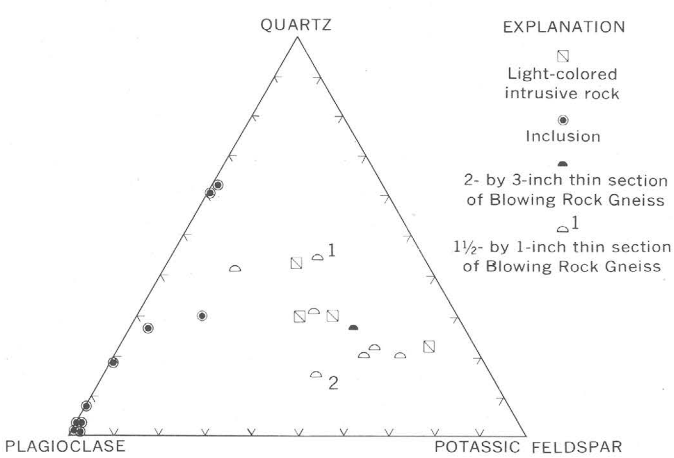



FIGURE 40.—Proportions of quartz, plagioclase, and potassic feldspar in Blowing Rock Gneiss, light-colored intrusive rocks, and inclusions. Based on counts of 50 random grains in thin sections. Numbers refer to anlyzed specimens, table 13.

TABLE 13.—*Chemical analyses of Blowing Rock Gneiss and phyllonite*

[Analysis of sample 2, standard rock analysis by C. L. Parker, U.S. Geol. Survey, 1961. Other analyses, by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis and Hezekiah Smith, U.S. Geol. Survey, 1963. For analysis of sample 1, S determined by induction furnace by I. C. Frost, U.S. Geol. Survey. F and Cl determined by V. C. Smith, U.S. Geol. Survey. Major oxides and CIPW norms are given in weight percent]

| | Gneiss | | Phyllonite | |
|---|---|---|---|---|
| | 1 | 2 | 3 | 4 |
| Field No | GMB-3 | GMB-4 | P1 | P3 |
| Laboratory No | 161252 | H-3264 | 161253 | 161254 |
| **Major oxides** | | | | |
| $SiO_2$ | 61.2 | 63.30 | 73.6 | 63.1 |
| $Al_2O_3$ | 14.2 | 14.87 | 11.6 | 16.8 |
| $Fe_2O_3$ | 1.8 | 1.15 | 1.1 | 2.2 |
| FeO | 4.6 | 3.74 | 2.1 | 3.1 |
| MgO | 1.6 | 1.27 | .79 | 1.1 |
| CaO | 4.7 | 3.14 | 1.7 | 1.8 |
| $Na_2O$ | 3.4 | 3.18 | 1.0 | .19 |
| $K_2O$ | 3.1 | 4.70 | 4.0 | 6.5 |
| $H_2O+$ | 1.2 | 1.20 | 1.4 | 2.2 |
| $H_2O-$ | .05 | .04 | .04 | .06 |
| $TiO_2$ | 1.6 | 1.12 | .72 | 1.6 |
| $P_2O_5$ | .80 | .40 | .26 | .74 |
| MnO | .14 | .09 | .04 | .04 |
| $CO_2$ | .82 | 1.09 | .94 | <.05 |
| F | .22 | .20 | | |
| Cl | .01 | .01 | | |
| BaO | | .16 | | |
| S | .03 | .06 | | |
| Total | 99 | 99.61 | 99 | 99 |
| **CIPW norms** | | | | |
| Q | 20.20 | 20.95 | 50.13 | 33.13 |
| C | 1.03 | 2.72 | 5.34 | 7.95 |
| Or | 18.32 | 27.77 | 23.63 | 38.40 |
| Ab | 28.68 | 26.82 | 8.46 | 1.61 |
| An | 11.55 | 5.03 | .79 | 4.10 |
| Hl | | .02 | | |
| En | 3.98 | 3.16 | 1.97 | 2.74 |
| Fs | 4.52 | 4.11 | 1.83 | 1.31 |
| Mt | 2.61 | 1.67 | 1.60 | 3.19 |
| Il | 3.04 | 2.13 | 1.37 | 3.04 |
| Ap | 1.90 | .95 | .62 | 1.75 |
| Fr | .38 | .37 | | |
| Pr | .06 | .11 | | |
| Cc | 1.86 | 2.48 | 2.14 | |

NOTE.—Minor-element analysis for sample 2 given in table 1.

1. Gray and white cataclastic augen gneiss containing porphyroclasts of white potassic feldspar as much as 3 cm long in a green-gray matrix of fine-grained mica. Porphyroclasts are microcline and perthitic microcline that have been broken and healed by recrystallized quartz and calcite, somewhat bent brown biotite as much as 1 mm long, and saussuritized plagioclase as much as 3 mm in diameter. Matrix is composed of quartz, green biotite, sericite, subordinate FeMg chlorite (replacing biotite), and epidote. Accessory minerals are ilmenite, sphene, apatite, pyrite, allanite, stilpnomelane, and zircon. From small quarry along U.S. Highway 221–321 near spot altitude of 3485, south of Chetola Lake (area H–3, pl. 1).
2. Dark-gray and white cataclastic augen gneiss containing porphyroclasts of white potassic feldspar as much as 2 cm long in a green-gray matrix of fine-grained mica. Porphyroclasts of microcline and perthitic microcline as much as 2 cm long in a matrix of recrystallized albite, sericite, quartz, and green biotite; accessory minerals are sphene, calcite, allanite, zircon, and epidote. Porphyroclasts are filled with inclusions of quartz and plagioclase. From roadcut along U.S. Highway 321 through meander of the South Fork of New River 1.75 miles N. 50° E. of junction between U.S. Highways 321 and 221 in the village of Blowing Rock (area H–3, pl. 1).
3. Porphyroclastic phyllonite derived from Blowing Rock Gneiss. Gray sericitic rock containing numerous quartz and feldspar porphyroclasts. Porphyroclasts of quartz are as much as 7 mm in diameter; perthitic microcline porphyroclasts, 5 mm; and plagioclase porphyroclasts, 2 mm. Matrix is chiefly recrystallized sericite and quartz but contains accessory calcite, pyrite, sphene, zircon, and apatite. From same locality as specimen 2.
4. Phyllonite derived from Blowing Rock Gneiss. Dark-gray phyllitic rock containing a few quartz porphyroclasts. Quartz porphyroclasts as much as 2 mm in diameter in a groundmass of recrystallied sericite, FeMg chlorite, quartz, and accessory sphene, ilmenite, pyrite, apatite, allanite, and zircon. From roadcut along U.S. Highway 221–321, near Middle Fork Church (area H–3, pl. 1).

on thin-section study is probably no more accurate than classification based on field estimates. Chemical analyses (table 13) of two samples show that the Blowing Rock Gneiss is quite similar to the average quartz monzonite (Nockolds, 1954) but that it contains less $SiO_2$, more FeO, and more total iron.

The light-colored intrusive rocks contain little mica and therefore do not have well-developed foliation. In thin section, however, the effects of partial cataclasis and recrystallization are evident. The rocks contain zones of recrystallized mortar and saussuritized plagioclase and are blastomylonitic gneisses or recrystallized mortar gneisses (fig. 39*C*). They are quartz monzonites and granites (fig. 40) that differ from the country rock principally in their lack of mafic minerals and in their smaller and more even grain size.

The inclusions are predominantly fine-grained quartz-plagioclase-biotite gneisses, but range from granodiorite to amphibolite (fig. 40).

Phyllonites in the Blowing Rock Gneiss contain 40 to 80 percent sericite and 0 to 20 percent biotite or chlorite. Quartz porphyroclasts are widespread and feldspar porphyroclasts, less common; some phyllonites lack porphyroclasts. Chemical analyses (table 13) suggest losses in CaO, $Na_2O$, and MgO and gains in $K_2O$ and $H_2O$ during phyllonite formation.

The petrographic data and field evidence do not conclusively establish the origin of the Blowing Rock Gneiss. Less sheared contacts are generally gradational over a few to a few tens of feet, and none are demonstrably intrusive. In some places the contact with Wilson Creek Gneiss is gradational through greater distances. Inclusions of nongranitic rocks are mostly small and resemble inclusions in plutonic igneous rocks; layering is rare. Isotopic ages of zircon from the Blowing Rock Gneiss north of the village of Blowing Rock are nearly concordant and range from 990 to 1,055 m.y., suggesting that the rock formed about 1,055 m.y. ago in the same general metamorphic-plutonic episode as the Wilson Creek Gneiss (Davis and others, 1962).

### BROWN MOUNTAIN GRANITE

The Brown Mountain Granite (Reed, 1964b) is medium to coarse grained, light colored, and homogeneous. It is generally weakly foliated and nonlayered but commonly has a distinct cataclastic lineation. It characteristically crops out in flat exfoliation slabs which make measurement of foliation and lineation difficult.

The contacts of the Brown Mountain Granite are poorly exposed. On the southeast side, the main body is overridden by the Linville Falls and associated faults, and on the west side, the granite apparently overrides volcanic and sedimentary rocks of the Grandfather Mountain Formation. Rocks of the Grandfather Mountain Formation unconformably overlie the granite east of bench mark 1195 on North Carolina Highway 181 area F–7, pl. 1.) and northwest of Adams Mountain (area G–6). Conglomerates in the Grandfather Mountain Formation locally contain pebbles of Brown Mountain Granite. Dikes of granite cut the Blowing Rock and Wilson Creek Gneisses northeast of Adams Mountain (areas G–6 and G–7).

Local zones of silvery medium-gray phyllonite, small bodies of quartz-perthite pegmatite, and small quartz segregation lenses and veinlets occur in the granite. Fluorite is found locally as thin coatings on joint surfaces.

The granite is a recrystallized mortar gneiss or blastomylonitic gneiss composed of porphyroclasts of potassic feldspar as much as 1 cm long and smaller and less abundant porphyroclasts of plagioclase and quartz in a matrix of recrystallized albite, quartz, biotite, sericite, and microcline having grain sizes ranging from 0.05 to 0.5 mm (fig. 41).

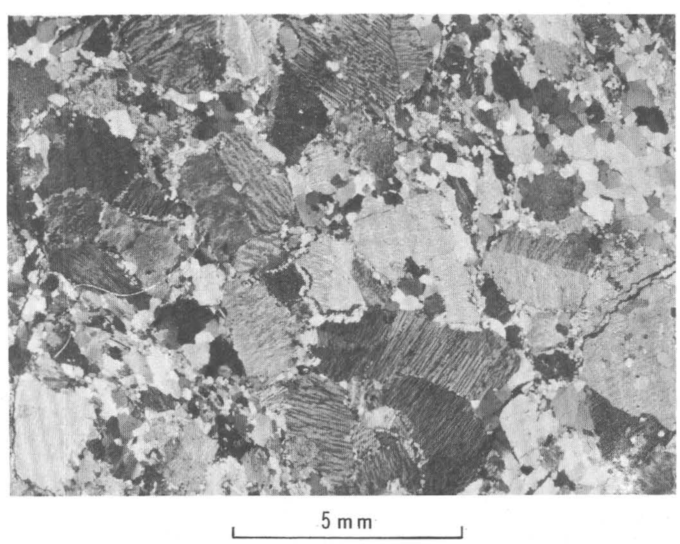



FIGURE 41.—Photomicrograph of typical Brown Mountain Granite from roadcut on east side of Wilson Creek 0.7 mile northeast of Brown Mountain Beach (area G–7, pl. 1). Porphyroclasts of string perthite in groundmass of recrystallized quartz, albite, potassic feldspar, and biotite. Section approximately perpendicular to mineral lineation. Analyzed specimen, table 14.

The potassic feldspar is string and patch perthite and antiperthite. The large porphyroclasts are bent, broken, and healed by quartz, albite, and sericite. Plagioclase porphyroclasts are as much as 6 mm long. They are generally saussuritized and altered to albite. Albite is a major constituent of the groundmass and locally forms rims about 0.05 mm thick on the potassic feldspar grains. Biotite is green or brown and forms irregular clots of recrystallized grains; a few biotite porphyroclasts occur locally. Colorless or light-green sericite forms small flakes in the albite-quartz mosaic. Fluorite occurs as small irregular grains in the mosaic. Accessory minerals are metamict allanite(?), epidote rimming allanite, sphene locally jacketing opaque minerals, zircon, ilmenite, magnetite, and stilpnomelane.

The Brown Mountain Granite is quite uniform in composition (fig. 42). One analyzed specimen (table 14) is siliceous alkali granite low in $K_2O$ and $Al_2O_3$. Recalculation of the chemical analysis into ideal orthoclase and albite places the sample close to the eutectic for the system $NaAlSi_3O_8$-$KalSi_3O_8$-$H_2O$ at about 1,000 kg per cm$^2$ water-vapor pressure (Tuttle and Bowen, 1958, p. 75).

The Brown Mountain Granite probably intruded the Blowing Rock and Wilson Creek Gneisses during a late stage of, or after, the Precambrian metamorphic-plutonic episode. Except for the dikes, no intrusive relations were found, but the map pattern, sharp contacts, lack of layering, uniformity in com-

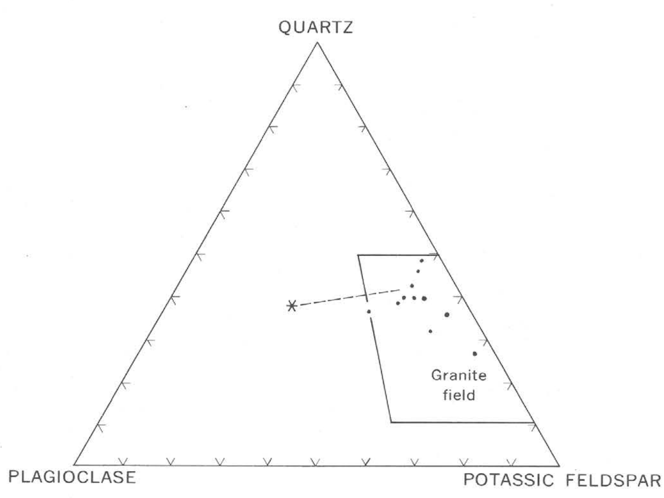



FIGURE 42.—Proportions of quartz, plagioclase, and potassic feldspar (including perthite) in Brown Mountain Granite. Line connects point count on analyzed specimen (table 14) with calculated volume proportion of albite, orthoclase, and quartz. Granite compositional field outlined. Based on count of 50 random grains in each thin section.

TABLE 14.—*Chemical analysis, mode, and norm of Brown Mountain Granite*

[Standard rock analysis by C. L. Parker, Denver, Colo., 1961. Mineral composition determined by point count; P, present but not intersected in counting. Major oxides and CIPW norms given in weight percent; mode, in volume percent]

| Field No. | G | Field No. | G |
|---|---|---|---|
| Laboratory No. | H-3434 | Laboratory No. | H-3434 |
| **Major oxides** | | | |
| $SiO_2$ | 76.85 | $K_2O$ | 4.66 |
| $Al_2O_3$ | 12.06 | $H_2O+$ | .18 |
| $Fe_2O_3$ | .35 | $H_2O-$ | .01 |
| FeO | 1.08 | $TiO_2$ | .10 |
| MgO | .03 | $P_2O_5$ | .00 |
| CaO | .44 | MnO | .02 |
| $Na_2O$ | 3.87 | $CO_2$ | .02 |
| | | Total | 99.67 |
| **Mode** | | | |
| Perthite | 48 | Magnetite | .4 |
| Albite | 9 | Zircon | .2 |
| Quartz | 40 | Fluorite | P |
| Biotite | 2.7 | Muscovite | P |
| | | Points counted | 600 |
| **CIPW norm** | | | |
| Q | 34.94 | En | .07 |
| Or | 27.53 | Fs | 1.57 |
| Ab | 32.73 | Mt | .51 |
| An | 1.78 | Il | .19 |
| Wo | .11 | Cc | .04 |

NOTE.—Minor-element analysis of specimen is given in table 1. Description of analyzed specimen follows:
Coarse-grained, light-colored granite with conspicuous lineation and faint foliation. Crystals of string perthite as much as 7 mm in diameter in a groundmass of quartz, albite, potassic feldspar, and biotite 0.1 to 1 mm in grain size. Most of the groundmass apparently recrystallized during metamorphism. A porphyroclast of biotite 1.5 mm in diameter. From roadcut on east side of Wilson Creek 0.7 mile northeast of Brown Mountain Beach, Lenoir quadrangle (area G-7, pl. 1).

position, and highly perthitic character of the potassic feldspar support this interpretation. The original contacts have been modified during the Paleozoic by shearing and faulting, which produced the present texture and structure of the Brown Mountain Granite.

## GRANDFATHER MOUNTAIN FORMATION

The Grandfather Mountain Formation of late Precambrian age (Bryant, 1962) unconformably overlies the Wilson Creek Gneiss, the Blowing Rock Gneiss, and the Brown Mountain Granite and is tectonically overlain by the Blue Ridge thrust sheet and the Tablerock thrust sheet. The formation is a thick interlayered and intertonguing sequence of metamorphosed arkose, siltstone, and volcanic rocks. It crops out in the northwestern third of the Grandfather Mountain window, where the more resistant rock units control the topographic form of many of the higher mountains on the Blue Ridge, and in several smaller isolated areas on the Blue Ridge front.

The first specific mention of the rocks of the Grandfather Mountain Formation was by Elisha Mitchell (1905, p. 52) in his account of an ascent of Grandfather Mountain in 1828. He observed that the top of the mountain was made of "grau wacke" and the lower part of "clay slate," to both of which he assigned a "transition" age.

Kerr (1875, p. 135–136) assigned the rocks of the Grandfather Mountain Formation to the "Huronian" and described them as follows:

> But the most remarkable part of this belt, both for breadth and peculiar lithology, is found in the region of the Grandfather and Yellow Mounsins [sic], about the headwaters of Toe River, Linville, Elk and Watauga. On upper Linville, and towards the base of the Grandfather and head waters of the Watauga, there are limited beds of argillaceous and hydramica slates and shales; but the prevalent rocks are feldspathic and quartzose slates and grits, sometimes gneisslike, and chloritic and epidotic schists, with epidosites, the latter sometimes enclosing bright red rounded grains of jasper. Along the high spurs of the Grandfather, to the northwest, the Yellow Mountains, are large bodies of greenish epidotic sand stone, feldspathic and quartzose; and along the northward spur of Hanging Rock, hard and gnarled dark gray quartzo-argillaceous slates prevail; but these are also much veined with irregular masses and reticulations of epidote. Passing eastward to the Watauga, from Valley Crucis up the river, the the prevalence of epidotic and chloritic, massive or obscurely bedded rocks, is most striking. Some of the masses are much veined with fine seams of white quartz, while others are amygdaloidal, sprinkled with grains of gypsum and quartz and epidote; while still further east across the Rich Mountains, occur chloritic amygdaloids in which the grains are feldspar, which are much weathered so as to leave the surface of the rock deeply honey-combed. Alternating with these conspicuous and dominant masses along the river, are the slate and gneiss-like grits of Linville, and occasionally silvery, gray greenish and spotted argillaceous and feldspathic slates and shales.
>
> On the Elk River occurs a greenish quartzofeldspathic, thick bedded, compact to friable slate and grit, which gradually passes into a nacreous light-colored, coarse slate-conglomerate—a fine-grained argillaceous quartzite, filled with rounded and flattened pebbles of white and reddish quartz and of hard quartzo-argillaceous slates.

Keith (1903) correlated the arkose units in the Grandfather Mountain Formation with the Unicoi Formation and the siltstone units with the Hampton Formation of the Chilhowee Group of the Valley and Ridge province. He also believed that all the volcanic rocks underlay the arkoses. This naturally led to structural interpretations which differ from ours.

Stose and Stose (1944) correlated the rocks of the Grandfather Mountain Formation with the Chilhowee Group, whereas Rodgers (1953a, p. 22) correlated the volcancic rocks of the formation with those of the Mount Rogers area in southern Virginia, extreme northeastern Tennessee, and northwestern North Carolina.

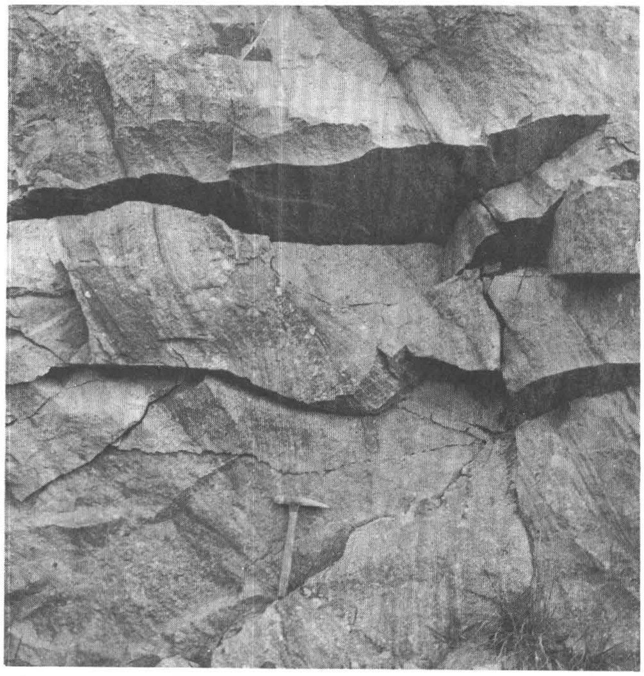
Variations in thickness and facies occur along strike in the Grandfather Mountain Formation. Generally speaking, a variety of volcanic rocks occurs near the base of the formation in the easternmost areas of outcrop, whereas arkose forms the basal deposit farther west. A body of metabasalt, the Montezuma Member, is found in the upper part of the formation. In area D–5 (pl. 1), the upper arkoses and the Montezuma Member pinch out. The thick basal arkose unit partly interfingers with siltstone to the northeast (area G–2), and the second arkose unit above the base pinches out completely.

Stratigraphic sections of the Grandfather Mountain Formation cannot be measured because of incomplete exposures, complex structure, and lack of distinctive marker horizons, but the formation is probably 10,000 to 30,000 feet thick. No major stratigraphic unit is entirely repeated by folding.

All the rocks of the Grandfather Mountain Formation have been metamorphosed, but as original sedimentary and volcanic structures and textures are abundant in the arenaceous sedimentary rocks and the felsic volcanic rocks, sedimentary and igneous rock names are used to describe them. Metamorphic names are generally used for the pelitic rocks and the mafic volcanic rocks which more readily show the effects of metamorphism.

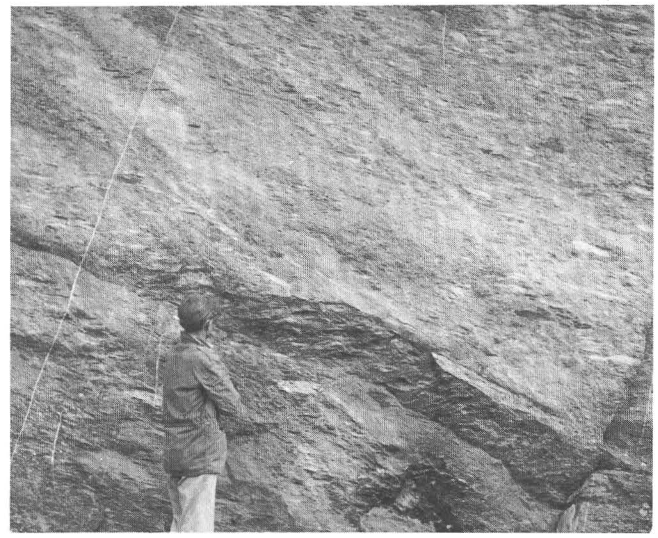
### BASAL CONTACT

The contact between the Grandfather Mountain Formation and the underlying plutonic rocks is exposed in only a few places. In some places its exact position is difficult to determine because of the close resemblance between coarse-grained arkose and phases of the Wilson Creek Gneiss. In many places, however, the basement rocks can readily be distinguished because they contain pegmatite and biotite and are coarser grained than the arkose. Bedding is generally poorly developed in the arkose near the contact.

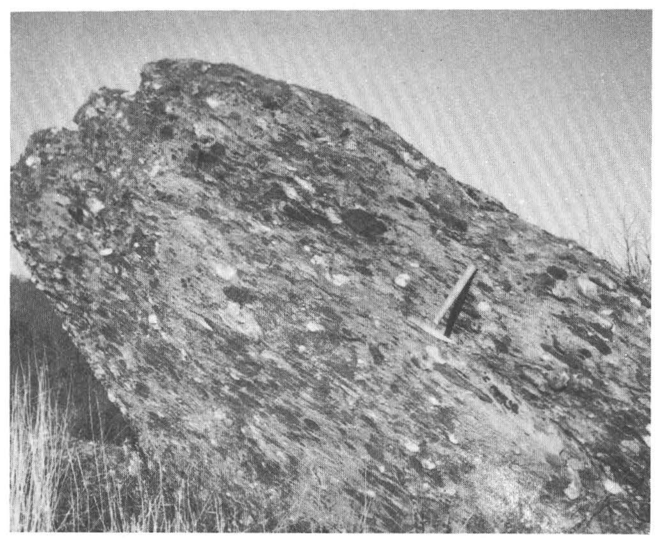


*A*

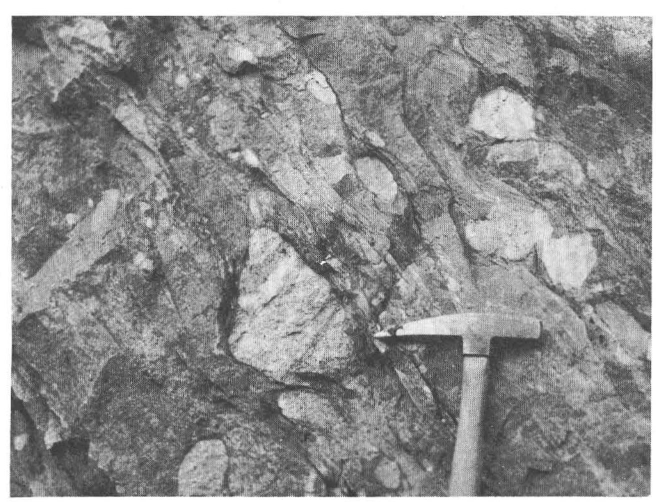
*B*



*C*



*D*



*E*

FIGURE 43.—Arkose of the Grandfather Mountain Formation. *A*, Crossbedding in arkose bed in middle part of exposed Grandfather Mountain Formation. Roadcut along North Carolina Highway 184 northwest of Linville Gap (area E–4, pl. 1). Overturned beds dip east. *B*, Bedding in coarse-grained arkose and conglomerate. Upper part of the lowest continuous arkose unit. Altitude of 3,850 feet on ridge northeast of Shulls Mills (area G–3, pl. 1). *C*, Conglomeratic arkose near top of lowest continuous arkose unit. Lithic fragments flattened along cleavage, which cuts bedding. At 3.850-foot altitude on ridge northeast of Shulls Mills (area G–3, pl. 1). *D*, Conglomerate in highest arkose unit. Quartz pebbles and lithic fragments of sedimentary and volcanic(?) rock elongated parallel to cleagave. Bedding visible. Outcrop has slumped to the left; cleavage originally dipped 80° to the right (east). Near the head of Pigeonroost Creek (northeast part of area E–3, pl. 1). *E*, Conglomerate containing cobbles of plutonic and felsic volcanic rocks in upper part of lowest continuous arkose unit. Roadcut about 0.5 mile southeast of Shulls Mills (area G–3, pl. 1).

An overturned contact between Wilson Creek Gneiss and basal arkose of the Grandfather Mountain Formation is exposed along a Forest Service road (not shown on the base map) northeast of Sassafras Knob (area E–5, pl. 1). Light-green sericite arkose containing scattered clastic grains of quartz and feldspar structurally underlies green to gray cataclastic granitic gneiss that contains porphyroclasts of potassic feldspar and fine-grained sericite and biotite. The gneiss contains small lenses and pods of pegmatite that are lacking in the arkose. Well-developed cleavage dips 25° SE., slightly less than the contact.

The contact between arkose and basement rock is also exposed in an area on Lost Cove Cliffs (area E–5, pl. 1). Arkose on the crest of the cliffs contains beds of quartz and feldspar pebbles that are as much as 2 inches in diameter. Cleavage dips 15° SE. The map pattern indicates that the contact dips gently westward at the top of the cliff, but on the part of the cliffs where it is best exposed, it dips about 35° W. Near the base of the cliffs, it dips 35° SE. to form a fold that is nearly recumbent, although not isoclinal.

The contact between siltstone of the Grandfather Mountain Formation and the Blowing Rock Gneiss is exposed along U.S. Highway 321 near Goldmine Branch (area H–3), where dark-blue-gray biotitic siltstone containing calcareous beds overlies coarse-grained augen gneiss. It is difficult to locate the contact precisely, and several inches of sediment at the base probably contain much material derived from the gneiss.

## CONTACTS BETWEEN ARKOSE AND SILTSTONE UNITS

Contacts between arkose and siltstone units are generally well defined, except on and northeast of Flattop Mountain in the eastern part of the main outcrop area of the Grandfather Mountain Formation (areas G–2, G–3, and H–3, pl. 1). There, units mapped as siltstone contain some interbeds of arkose and mafic volcanic rock, and arkose and siltstone are interbedded for a few tens of feet to perhaps 100 feet near their contacts. Interbedding is also found on the north side of Hodges Mountain (area G–2) and north of the Pack Hill School (area G–4), where arkose and siltstone units interfinger. Gradational contacts between the two lithologies are exposed on the hill east of Newland (area D–4), on Peak Mountain (area E–3), and on Rich Mountain (area G–3).

A sharp contact between arkose and siltstone is well exposed in roadcuts along North Carolina Highway 181 north of Jonas Ridge (area D–6). Coarse-grained, massive, white to light-green conglomeratic arkose grades upward into green, fine-grained, slightly calcareous arkose containing scattered feldspar grains as much as 6 mm in diameter. The arkose is overlain by highly crenulated light-green, dark green, and black phyllite containing light-gray siltstone beds 1 to 2 feet thick.

Soils and float derived from the arkose and siltstone are easily distinguished. The siltstone units form chippy soil containing abundant small rock fragments, whereas the arkose forms granular sandy soil made up of individual grains of quartz and feldspar.

## SEDIMENTARY ROCKS

### ARKOSE

#### MEGASCOPIC FEATURES

Arkose is the predominant rock type of the Grandfather Mountain Formation; it forms large cliffs and slabby outcrops on Grandfather Mountain, Peak Mountain, Sugar Mountain, Hawksbill Mountain, and other peaks and ridges along the Blue Ridge front. Excellent exposures in roadcuts are found at almost any point at which roads cross arkose units.

The arkose units consist of fine- to coarse-grained, thin-bedded to massive, green, light-green, tan, and gray sericitic arkose and feldspathic quartzite containing minor beds of conglomerate, green to gray sericite phyllite, and siltstone. Much of the arkose is poorly bedded, especially in the lowest arkose unit southeast of Grandfather Mountain. Bedding is marked by concentrations of heavy minerals and solution of calcareous zones. Crossbedding (fig. 43*A*) is found locally, but graded bedding is rare. Some of the crossbedding is of the torrential variety. Oscillatory ripple marks are found in a few places. They are well displayed along the road between U.S. Highway 221 and Gragg, where it crosses the west branch of Wilson Creek (area F–4). Sericite phyllite beds one-half of an inch to 6 feet thick occur sparingly in the lower arkose unit but are much more abundant in the upper part of the exposed section. Differences in composition, grain size, and color are more widespread in the higher arkose units. The color of the highest arkose unit is especially variable; siltstones and phyllite beds are gray, green, purple, and maroon. Careful examination of roadcuts along any main highway reveals local bedding in all the arkose units.

Beds and discontinuous lenses of conglomerate (fig. 43*B, C, D, E*) containing pebbles and cobbles of quartz, feldspar, siltstone, phyllite, felsic volcanic rock, arkose, light-colored granoblastic plagioclase-quartz rock, granitic rock, and pegmatite are found throughout the formation. The conglomerate bodies are thickest and most numerous along the crest and on the northwest and north sides of Grandfather Mountain (areas E–4 and F–4, pl. 1) and on Hanging Rock Ridge north of Foscoe (area F–3). Most of the conglomerate lenses are less than 10 feet thick, and the thinnest are pebbly beds only a few inches thick. The thickest and most spectacular conglomerate outcrop is at an altitude of 4,200 feet in Falls Hollow north of the highest peak of Grandfather Mountain (northwest corner of area F–4). There the conglomerate is more than 100 feet thick and contains pebbles, cobbles, and boulders of felsic volcanic rock and muscovite-plagioclase quartzite as much as a foot long. About half a mile to the northeast, boulders of felsic volcanic rock are as much as 2 feet long.

Excellent exposures of conglomerate in the upper part of the lowest arkose unit may be seen along the Blue Ridge Parkway about 0.3 mile southwest of Flat Rock (area E–5). There the conglomerate contains pebbles of quartz and feldspar and lithic fragments of felsic volcanic rock, black or dark-gray argillite or siltstone, gray quartzite, light-green and gray sericite phyllite, light-colored granitic rock, pegmatite, and red jasper. Cleavage is well developed, and pebbles of less competent rocks have been deformed into blade-shaped fragments as much as 10 inches long and only 1 to 2 inches thick, whereas the more competent cobbles are a maximum of 4 inches in diameter. Bedding can be seen in the southern part of the outcrop.

Typically, the conglomerates in the highest exposed arkose unit contain quartz pebbles and fragmenss of gray, maroon, and purple rock and resemble the finer grained interbeds in that unit. Other rock types occur but are much less abundant. The dark rock fragments are flattened in the plane of the cleavage and locally elongated in a northwest-southeast direction parallel to the lineation. These conglomerate beds are generally a few inches to a few feet thick. Rarely do the rock fragments exceed 4 inches in length. Outcrops of such conglomerates (fig. 43*D*) may be seen along North Carolina Highway 184 in the village of Banner Elk and at the head of Pigeonroost Creek (area E–3, pl. 1). An unusually thick, coarse-grained, and diverse conglomerate in the uppermost arkose is exposed along old North Carolina Highway 194 on Mill Timber Creek (area D–5, pl. 1). Several conglomerate beds form a sequence about 40 feet thick. Coarse-grained conglomerate at the bottom contains cobbles as much as 8 inches long. It grades up into interbedded pebble conglomerate and coarse-grained arkose. In addition to the usual quartz and dark-colored argillite pebbles, the conglomerate contains fragments of felsic volcanic rock and quartzite.

An exposure in the lower part of the second arkose unit west of Linville Gap (area E–4, pl. 1) on North Carolina Highway 184 has unusually well-developed sedimentary structures and textures. The arkose (fig. 43*A*) contains quartz and feldspar clasts and lithic fragments of argillaceous rock as much as 2 cm in diameter. Dark beds are rich in heavy minerals. Crossbedding and graded bedding are conspicuous, and the cleavage is poorly developed.

In many places the arkose contains scattered large grains or pebbles of feldspar (especially in the lowest arkose), quartz, and fine-grained dark-colored rock fragments.

The green color of the arkose is due to evenly distributed green mica. Locally, especially in the lower arkose unit, the green mica is concentrated in segre-

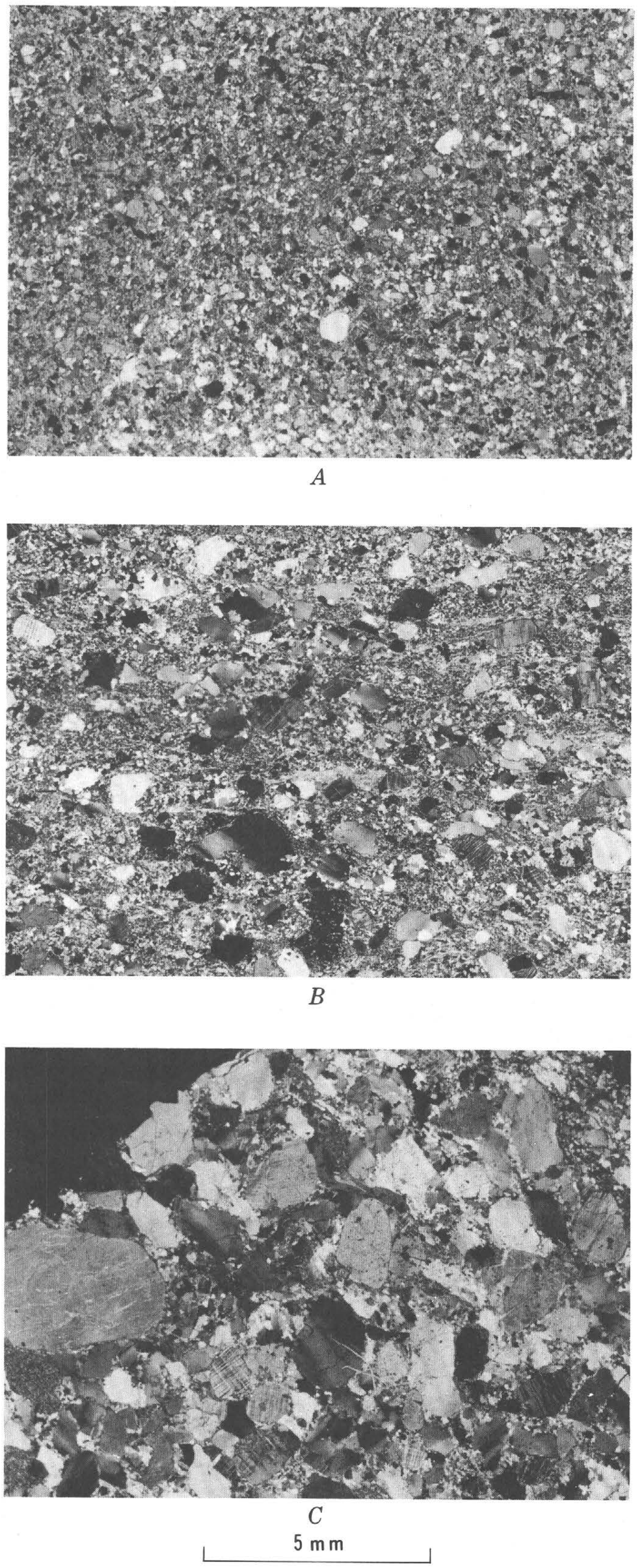



FIGURE 44.—Photomicrographs of arkose in the Grandfather Mountain Formation. *A*, Fine-grained arkose from quarry on east side of Wilson Creek at 3,550-foot altitude (area F–4, pl. 1). Clastic grains of quartz, microcline, and plagioclase (pseudomorphosed by albite) in a matrix of iron-rich muscovite, quartz microcline, and albite. Analyzed specimen 2, table 15. *B*, Fairly coarse grained, poorly sorted feldspathic quartzite from roadcut at 3,250-foot altitude on Rough Ridge (area F–4, pl. 1). Clastic grains of quartz, microcline, and subordinate plagioclase (pseudomorphosed by albite) in a matrix of recrystallized quartz, iron-rich muscovite, and microcline. Analyzed specimen 5, table 15. *C*, Coarse-grained arkose from roadcut on North Carolina Highway 184 northwest of Linville Gap (area E–4, pl. 1). Clastic grains of quartz, microcline, and plagioclase (pseudomorphosed by albite) and fragments of very fine grained quartz-feldspar rock in a small amount of matrix composed of recrystallized quartz and sericite. Some of the larger microcline grains are factured and healed by quartz. From outcrop shown in figure 43*A*.

gation lenses and laminae less than a millimeter thick that are parallel with the cleavage. The micaceous folia are 5 to 10 mm apart and separated by white feldspar-quartz rock.

Calcareous beds and lenses weather brown and in many outcrops are represented only by voids left by solution. On North Carolina Highway 181, 1.2 miles south of the Jonas Ridge School (area D–6, pl. 1), dark-blue-gray arkose contains thin calcareous beds and ellipsoidal light-gray calcareous knots as much as 3 inches long. The knots are surrounded by a halo of bleached arkose.

### TEXTURE AND MINERALOGY

Sedimentary textures are well preserved in much of the arkose (fig. 44). Clastic grains more than 0.2 mm in diameter of the major minerals are easily recognizable, and some smaller clastic grains are identifiable. Some grains of accessory minerals as small as 0.05 mm have the appearance of clasts. Most of the arkose is poorly sorted. Much of it has well-developed cleavage which is generally parallel with bedding southeast of Grandfather Mountain but which cuts across bedding in many places in the northwestern part of the outcrop area.

Typical arkose contains angular to subrounded clastic grains of quartz, potassic feldspar, and plagioclase in a mosaic-textured groundmass of recrystallized quartz and feldspar containing various amounts of well-alined light-green mica. Some of the clasts of quartz and feldspar are several inches in diameter and must have been derived from quartz veins and pegmatite. The larger clasts in typical coarse-grained arkose are generally potassic feldspar about 5 mm in diameter, quartz clasts 2 mm in diameter, and plagioclase 0.8 mm in diameter (fig. 44*B* and *C*). Recrystallized groundmass averages about 0.1 mm in grain size. Rarely, the groundmass is as coarse as 0.2 mm, and in some rocks, especially those in the northwestern part of the window, it is less than 0.05 mm in grain size.

Clastic zircon (locally metamict) is ubiquitous. Clastic grains of apatite are fairly widespread. Sphene is as abundant as zircon, but some of it has recrystallized. Magnetite and ilmenite are at least partly clastic, but some may have recrystallized. Pyrite, which occurs rarely, has recrystallized. Both clastic and metamorphic epidote are widespread. Metamorphic epidote is especially abundant in a few places next to the Linville Metadiabase or the Montezuma Member of Grandfather Mountain Formation, where it has been introduced either during the emplacement of the igneous rock or during metamophism. Calcite is metamorphic. Clastic allanite occurs, and clastic tourmaline is rare.

The clastic grains of potassic feldspar are generally microcline. Some of the larger ones contain inclusions of quartz and plagioclase. These clasts resemble potassic feldspar in the basement rocks in the Grandfather Mountain window. Perthite is much rarer. Some was derived from the plutonic rocks underlying the Grandfather Mountain Formation. Other perthite grains that have a very fine texture and much finely divided opaque material resemble phenocrysts in the felsic volcanic rocks of the Grandfather Mountain Formation. These grains are found in arkose adjacent to felsic flows and in conglomerates containing pebbles of felsic volcanic rock in the Linville and Blowing Rock quadrangles and on the east margin of the southern area of arkose in the Linville Falls quadrangle, about 3 miles west of the largest area of outcrop of the felsic volcanic rocks.

Many of the larger potassic feldspar grains are bent and broken and have been healed by quartz and sericite and rarely by albite and calcite.

In places, quartz clasts are strongly strained and brecciated. Some of the broken grains are healed by mosaics of recrystallized quartz.

All plagioclase clasts have been altered to albite, and plagioclase in the groundmass is also albite. Some of the clasts are bent and broken. A much larger proportion of plagioclase than potassic feldspar is found in the groundmass, and the finer grained arkose generally contains more plagioclase. In eight rocks from the lowest arkose unit, which have a maximum grain size of 0.5 mm, the plagioclase makes up 70 percent of the feldspar; in 17 rocks, which have a maximum grain size of 0.5 to 1 mm, 35 percent of the feldspar is plagioclase.

The green mica is an iron-rich muscovite (Foster and others, 1960). One analyzed sample of this mica has an index of refraction for $\beta$ of 1,6195$\pm$0.0005 and the pleochroism is $\alpha$, colorless; $\beta$, light green; and $\gamma$, light green. Index of refraction for $\beta$ measured on numerous other samples ranges from 1.598 to that of the analyzed sample. Generally, the micas with the higher indices of refraction have the stronger absorption. Colorless muscovite was not studied, and the mica in some rocks may be normal iron-poor muscovite. The range in the index of refraction and absorption suggests that there is probably a range in the iron content of the muscovite in the arkoses.

The mica is generally well alined and, especially in some of the more phyllitic rocks, is bent into small crinkles one to a few millimeters in wavelength. The larger flakes are generally 0.1 mm long; rarely are they as much as 0.5 mm long.

Radiating aggregates of stilpnomelane are found in the arkose in a few places. In one place the stilpnomelane is associated with a heavy-mineral seam.

Green and greenish-brown biotite occurs rarely as a metamorphic mineral of the same grain size as the iron-rich muscovite. Recognizable clastic grains of biotite are even less abundant than the metamorphic biotite. The biotite is locally altered to chlorite.

Phyllite interbeds contain sericite, opaque minerals, and accessory zircon, apatite, epidote, tourmaline, and biotite. Some have a fragmental texture. Fragments of light-green sericite phyllite several millimeters to 2 cm long and lacking opaque minerals occur in a matrix of dark-gray sericite phyllite containing opaque minerals. This rock type is found adjacent to the greenschists and greenstones of the Montezuma Member and also as beds a few to several inches thick in the arkose.

Heavy-mineral beds in the arkose are composed of opaque minerals, sphene, zircon, quartz, feldspar, and mica.

The brown-weathering carbonate beds are composed of sericite, calcite, albite, and quartz. Calcite makes up 10 to 25 percent of the beds. In one place, where calcite occurs in beds and ellipsoidal bodies in dark-blue-gray arkose, carbonates have replaced most of the plagioclase and some of the quartz and make up 50 percent or more of the rock. Epidote is abundant. A halo of bleached rock surrounding the calcareous bodies contains epidote and calcite and lacks the small proportion of chlorite and biotite that gives the color to the country rock. Calcite also occurs as irregular grains and veinlets in the country rock.

### Composition and Classification

In figure 45, estimated modal compositions of arkose are plotted in terms of four major components: quartz, potassic feldspar, plagioclase, and combined mica and chlorite. The content of rock fragments, a component which usually enters any classification of sandstones, is not shown. Rock fragments do not, however, make up a large proportion of most of the samples, and these rocks would not fall into the lithic graywacke or subgraywacke fields of Pettijohn (1957, p. 292) or the lithic wacke or lithic arenite fields of Gilbert (in Williams and others, 1954, p. 292–293).

Most of the arkoses are poorly sorted and contain a few to 50 percent iron-rich muscovite in their matrix. The mica seems to have been derived from clay minerals and iron oxide rather than through breakdown of feldspars, because many small grains of feldspar have not been converted to mica. Most of the arkoses are light colored and sandy looking. Nevertheless, in the classifications of Pettijohn (1957) and Gilbert (in Williams and others, 1954) many of these rocks would be called graywacke. In discussing rocks of similar mica content, although of generally darker color, Hadley and Goldsmith (1963, fig. 14) used the terms muddy arkose and muddy subarkose. We have chosen the classification of Packham (1954) in which sedimentary features are first considered before a name is attached. In his classification, these rocks would fall into the arkose and feldspathic sandstone division of the arkose-quartzose sandstone suite. The mica-rich arkoses probably were derived from muddy arkoses. These light-colored sandstones, which we call arkoses, contrast with beds of dark-gray sandstone found in the siltstone units, which we call graywacke.

The highest exposed arkose (fig. 45*B*) contains less plagioclase and less total feldspar than the lowest arkose (fig. 45*A*). The upper arkose unit contains more numerous siltstone and phyllite interbeds, and the mica content of the unit as a whole is therefore higher than in the lower units.

Rocks in the lowest arkose unit (fig. 45*A*) resemble those of the Newark Series in Connecticut (Krynine, 1950), if it is assumed that muscovite in the arkoses represents original clastic mica and clay. The sandstones of the Newark Series, however, contain considerably less clay and mica than would be indicated by metamorphic mica in arkoses of the Grandfather Mountain Formation. Thus, the Grand-

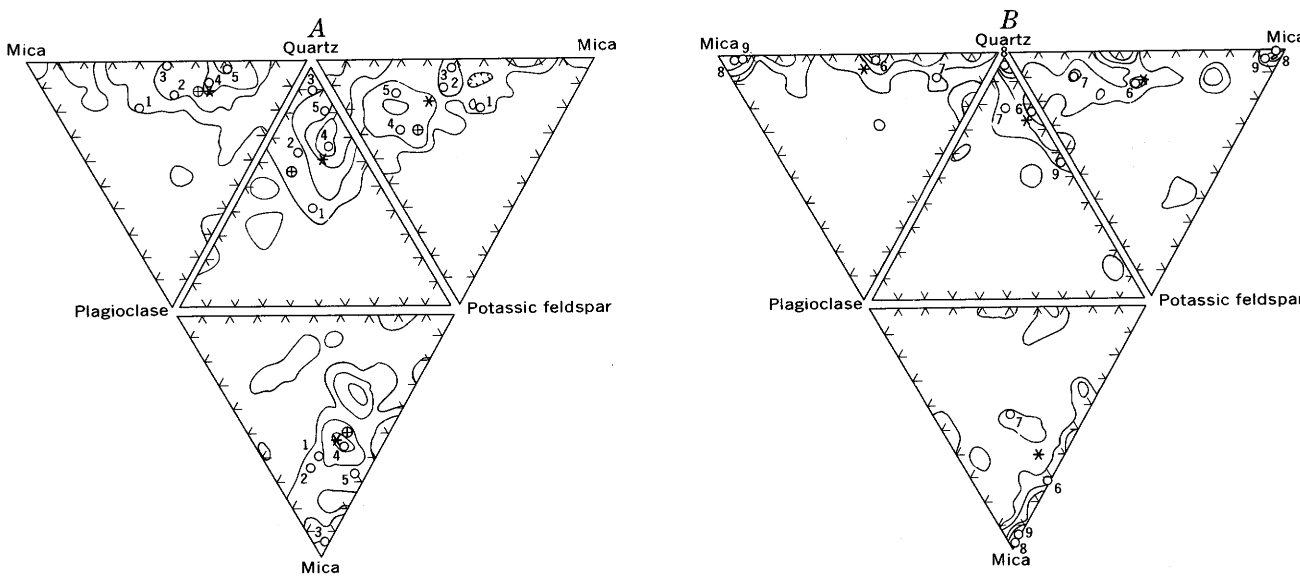



FIGURE 45.—Proportions of quartz, potassic feldspar, plagioclase, and mica in arkose of the Grandfather Mountain Formation. Numbers of analyzed specimens refer to analyses in table 15; *, average. *A*, Lowest arkose unit. ⊕, Weighted average of arkose in the Newark Series as used by Krynine (1950). Based on point counts of 24 thin sections and counts of 50 random grains in each of 55 other thin sections. Contours 1.3, 4, 8, and 12 percent. *B*, Highest exposed arkose unit. Based on counts of 50 random grains in each of 24 thin sections. Contours 4, 8, 17, and 25 percent.

father Mountain arkose is less sorted than arokose of the Newark Series.

Chemical analyses (table 15, Nos. 1–7) of representative samples of arkose and feldspathic quartzite

TABLE 15.—*Chemical analyses, modes, and norms of arkose and related rocks of the Grandfather Mountain Formation*

[Samples 2 and 8, standard rock analyses by C. L. Parker, U.S. Geol. Survey, 1961; sample 4, standard rock analysis by D. F. Powers, U.S. Geol. Survey, 1959; other samples analyzed by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey, 1962. Modes, by point count; P, present but not intersected in counting. Major oxides and CIPW norms given in weight percent; modes, in volume percent. CIPW norms not computable for samples 8 and 9]

| | Lower arkose unit | | | | | Upper arkose unit | | | |
|---|---|---|---|---|---|---|---|---|---|
| | | | | | | Arkose beds | | Phyllite beds | |
| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 |
| Field No. | J-8-4a | GML-8 | RE-102-3 | GML-7 | M-54-3 | P-69-1 | BA-644 | J-20-1 | P-54-3-d |
| Laboratory No. | 160190 | H-3423 | 160191 | F-2590 | 160189 | 160193 | 160192 | H-3427 | 160194 |
| **Major oxides** | | | | | | | | | |
| $SiO_2$ | 67.0 | 74.05 | 74.6 | 76.39 | 78.9 | 75.8 | 84.9 | 46.46 | 46.1 |
| $Al_2O_3$ | 17.0 | 13.04 | 10.6 | 11.18 | 9.1 | 11.3 | 7.8 | 19.77 | 23.3 |
| $Fe_2O_3$ | 2.5 | 2.31 | 2.3 | 2.16 | 3.5 | 1.3 | .63 | 13.25 | 13.2 |
| FeO | .64 | .16 | 1.0 | .63 | .30 | .28 | .17 | 1.15 | 1.0 |
| MgO | .88 | .59 | .53 | .46 | .46 | 1.3 | .42 | 2.77 | 1.0 |
| CaO | .58 | .64 | 1.9 | .40 | 1.3 | .85 | .06 | .00 | .26 |
| $Na_2O$ | 2.4 | 2.42 | .87 | 1.25 | .84 | .77 | 1.6 | .21 | .19 |
| $K_2O$ | 6.6 | 4.70 | 3.8 | 5.29 | 3.3 | 5.0 | 3.0 | 8.97 | 9.5 |
| $H_2O+$ | 1.2 | 1.12 | 1.24 | 1.00 | .75 | 1.2 | .77 | 3.89 | 2.7 |
| $H_2O-$ | .07 | .04 | .04 | .04 | .04 | .07 | .05 | .21 | .12 |
| $TiO_2$ | .66 | .48 | .33 | .52 | .62 | .37 | .09 | 2.67 | 1.6 |
| $P_2O_5$ | .26 | .16 | .07 | .12 | .16 | .10 | .03 | .29 | .39 |
| MnO | .03 | .03 | .04 | .02 | .03 | .03 | .01 | .02 | .06 |
| $CO_2$ | .05 | .01 | 1.3 | .01 | .52 | .43 | <.05 | .01 | <.05 |
| S | -- | -- | -- | -- | -- | -- | -- | -- | -- |
| F | -- | .04 | -- | -- | -- | -- | -- | .06 | -- |
| Cl | -- | .00 | -- | -- | -- | -- | -- | .00 | -- |
| Subtotal | -- | 99.79 | -- | -- | -- | -- | -- | 99.74 | -- |
| Less O | -- | .02 | -- | -- | -- | -- | -- | .02 | -- |
| Total | 100 | 99.77 | 99 | 99.47 | 100 | 99 | 100 | 99.72 | 100 |
| **Modes** | | | | | | | | | |
| Quartz | 25 | 41 | 48 | 50 | 62 | 49 | 65 | 5 | 4 |
| Potassic feldspar | 16 | 8 | 3 | 16 | 11 | 13 | 9 | -- | 3 |
| Plagioclase (albite) | 16 | 13 | .5 | 8 | 2.1 | .8 | 9 | -- | -- |
| Sericite | 41 | 35 | 47 | 25 | 23 | 38 | 15 | 79 | 65 |
| Sphene | 2.2 | 2.0 | .8 | 1.3 | 1.9 | .6 | .4 | 5 | .2 |
| Opaque minerals | -- | -- | -- | -- | .6 | -- | -- | 11 | 28 |
| Calcite | -- | -- | .7 | -- | 1.6 | .6 | -- | -- | -- |
| Epidote | -- | -- | P | -- | .9 | .4 | -- | -- | -- |
| Zircon | P | P | P | P | P | .1 | -- | P | P |
| Apatite | P | .2 | -- | P | P | P | -- | .2 | -- |
| Allanite | -- | P | -- | -- | -- | -- | -- | -- | -- |
| Biotite | -- | -- | -- | -- | -- | P | -- | -- | -- |
| Points counted | 500 | 1,118 | 600 | 1,005 | 1,000 | 706 | 742 | 600 | 600 |
| **CIPW norms** | | | | | | | | | |
| Q | 26.12 | 40.33 | 53.89 | 47.71 | 59.79 | 49.89 | 63.45 | -- | -- |
| C | 5.60 | 3.30 | 4.78 | 2.98 | 3.37 | 4.31 | 1.88 | -- | -- |
| Or | 38.99 | 27.77 | 22.45 | 31.25 | 19.50 | 29.54 | 17.72 | -- | -- |
| Ab | 20.30 | 20.47 | 7.36 | 10.57 | 7.10 | 6.51 | 13.53 | -- | -- |
| An | .86 | 1.83 | .75 | 1.14 | 2.12 | .84 | .10 | -- | -- |
| En | 2.19 | 1.47 | 1.32 | 1.14 | 1.14 | 3.24 | 1.05 | -- | -- |
| Mt | .25 | -- | 2.40 | .59 | -- | -- | .32 | -- | -- |
| Hm | 2.33 | 2.31 | .65 | 1.75 | 3.50 | 1.30 | .41 | -- | -- |
| Il | 1.25 | .40 | .63 | .99 | .67 | .66 | .17 | -- | -- |
| Ru | -- | .27 | -- | -- | .25 | .02 | -- | -- | -- |
| Ap | .62 | .38 | .17 | .28 | .38 | .24 | .07 | -- | -- |
| Fr | -- | .07 | -- | -- | -- | -- | -- | -- | -- |
| Cc | .11 | .02 | 2.96 | .02 | 1.18 | .98 | -- | -- | -- |

NOTE.—Minor-element analyses for samples 2, 4, and 8 given in table 1.

TABLE 15.—*Chemical analyses, modes, and norms of above and related rocks of the Grandfather Mountain Formation*—Continued

1. Light-greenish-gray well-sorted sericite arkose. Clastic grains of quartz, microcline, and plagioclase (altered to albite) in groundmass of recrystallized iron-rich muscovite, quartz, microcline, and albite with a grain size of 0.1 mm. From 3,080-foot altitude on Lost Cove Creek (area E-5, pl. 1).
2. Light-green sericite arkose. Few clastic grains of quartz and microcline as much as 0.5 mm in diameter in groundmass of recrystallied iron-rich muscovite, quartz, microcline, and albite. From quarry on the east side of Wilson Creek at 3,550-foot altitude (area F-4, pl. 1).
3. Green phyllitic quartzite containing lenses of calcite. Clastic grains of quartz and microcline as much as 2 mm in diameter in a matrix of recrystallized quartz 0.02 mm in diameter and light-green iron-rich muscovite 0.01 to 0.2 mm long. A few small quartz-microcline calcite segregation lenses. Mica probably overcounted in mode because of fine grain size. From roadcut 0.8 mile S. 59° E. of Shulls Mills (area G-3, pl. 1).
4. Green arkosic quartzite. Clastic grains of quartz as much as 2 mm in diameter, microcline 1 mm in diameter, and plagioclase (altered to albite) 0.5 mm in diameter in a matrix of recrystallied iron-rich muscovite, quartz, microcline, and albite with a grain size averaging 0.1 mm. From same outcrop as mica analyzed by Foster (Foster and others, 1960). From roadcut on U.S. Highway 221, 1.9 miles N. 62° E. of summit of Grandfather Mountain (area E-4, pl. 1).
5. Light-green feldspathic sericite quartzite containing heavy-mineral seams and a few small rock fragments. Clastic grains of quartz and microcline as much as 2.5 mm in diameter and plagioclase (altered to albite) as much as 1 mm in diameter in a matrix with a grain size of 0.05 to 0.1 mm of recrystallized quartz, iron-rich muscovite, and microcline. A few calcite grains as much as 0.5 mm in diameter occur in the matrix. From roadcut at 3,250-foot altitude on Rough Ridge (area F-4, pl. 1).
6. Light-green feldspathic sericite quartzite. Clastic grains of quartz as much as 1.5 mm in diameter and microcline and plagioclase (altered to albite) as much as 0.5 mm in diameter in a matrix of recrystallized sericite and quartz 0.01 to 0.06 mm in grain size. From roadcut in North Carolina Highway 184 south of Banner Elk near spot altitude of 3645 feet (area E-3, pl. 1).
7. Light-gray well-sorted sericite arkosic quartzite. Clastic grains of quartz and microcline as much as 0.5 mm in diameter and plagioclase (altered to albite) as much as 0.2 mm in diameter with a small proportion of matrix consisting of recrystallized quartz 0.01 to 0.05 mm in diameter and sericite 0.01 to 0.1 mm long. From 4,500-foot altitude on the north ridge of Bald Mountain (area D-3, pl. 1).
8. Light-gray sericite phyllite containing flattened fragments of light-green sericite phyllite as much as 2 cm long. Scattered clastic grains of quartz as much as 1 mm long and phyllite fragments in a matrix of very fine grained iron-rich muscovite and sphene and containing grains of hematite as much as 0.5 mm in diameter. Bed about 4 feet thick in arkose. From a roadcut along North Carolina Highway 181 just east of Montezuma (area D-4, pl. 1).
9. Mottled green, maroon, and gray sericite phyllite. Scattered clastic grains of quartz and microcline as much as 0.35 mm in diameter in a matrix of light-green sericite 0.01 to 0.2 mm long and hematite 0.005 to 0.5 mm in grain size. Hematite overcounted in mode. From roadcut along North Carolina Highway 194 just north of bridge over Blevens Creek crossing (area C-3, pl. 1).

of the Grandfather Mountain Formation resemble those of similar rocks elsewhere. They have a high $Fe_2O_3$:FeO ratio, which is characteristic of unmetamorphosed red arkoses. In the arkoses of the Grandfather Mountain Formation, the ferric iron is mainly in the green muscovite. $Na_2O$ content of the arkoses is a function of the amount of plagioclase; $Al_2O_3$ contents are related to the total feldspar and mica contents. Analyses of typical arkoses defined strictly on a mineralogical basis (Pettijohn, 1957, 1963) have a somewhat higher proportion of alkalis to alumina, probably because they contain a higher proportion of feldspar in relation to mica than do most Grandfather Mountain Formation arkoses.

Figure 46 compares alkali proportions and contents of arkoses of the Grandfather Mountain Formation with averages of sandstone clans according to Middleton (1960) and Pettijohn (1963). Arkoses of the Grandfather Mountain Formation contain more alkalis than the average arkoses, but the alkalis are in about the same proportions. The $K_2O$:$Na_2O$ ratio is substantially greater than 1, a characteristic of taphrogeosynclinal sandstones (Middleton, 1960), most of which are arkosic. Most arkoses of the Grandfather Mountain Formation have about the same ratios of $Al_2O_3$ to $CaO+Na_2O+K_2O$ as average arkoses (fig. 47). Analysis 1, which has a greater content of alumina and alkalis, is a muddy arkose that differs from graywacke in having a high $K_2O$:$Na_2O$ ratio. Analysis 3, which is of a rock poor in feldspar, falls near typical arkoses in this diagram because of its high mica content. Calculated mineral composition suggests that this rock contains about 10 percent feldspar and 30 percent mica and is a muddy feldspathic sandstone.

The mineral composition diagram (fig. 45) indicates that analyses 4 and 5 are most nearly representative of rocks of the lower arkose unit. These anlyses plot quite near the arkose averages on figure 47. Analysis 6 is typical of the rocks in the highest exposed arkose unit; it does not differ in many respects from analysis 4 of the lowest arkose, except for lower contents of iron oxides and $Na_2O$ and higher CaO content.

Analyses 8 and 9 are of sericite phyllite in interbeds in the highest arkose unit. These rocks are chiefly mixtures of iron-rich muscovite, hematite, and quartz. Their origin is problematic. The fragmental texture of sample 8 suggests that it is a tuffaceous rock, but the rock contains clastic grains of quartz. The $K_2O$ and total iron oxide contents are higher than those of most shales. Chemical analyses of rocks believed to be derived from a volcanic-derived saprolite (Reed, 1955) show that they contain less $K_2O$ and $Al_2O_3$ and more total iron oxide and CaO. Perhaps the material in these phyllite beds in the Grandfather Mountain Formation was derived from a residuum containing clay and iron oxide, which was enriched by $K_2O$ during diagenesis or metamorphism, or perhaps it was derived from a mixture of volcanic glass fragments with a minor amount of normal detritus and underwent marked compositional changes during diagenesis and (or) metamorphism.

### SILTSTONE

#### MEGASCOPIC FEATURES

Siltstone units are interbedded and interfinger with arkose units in the Grandfather Mountain Formation. The siltstone commonly underlies valleys be-

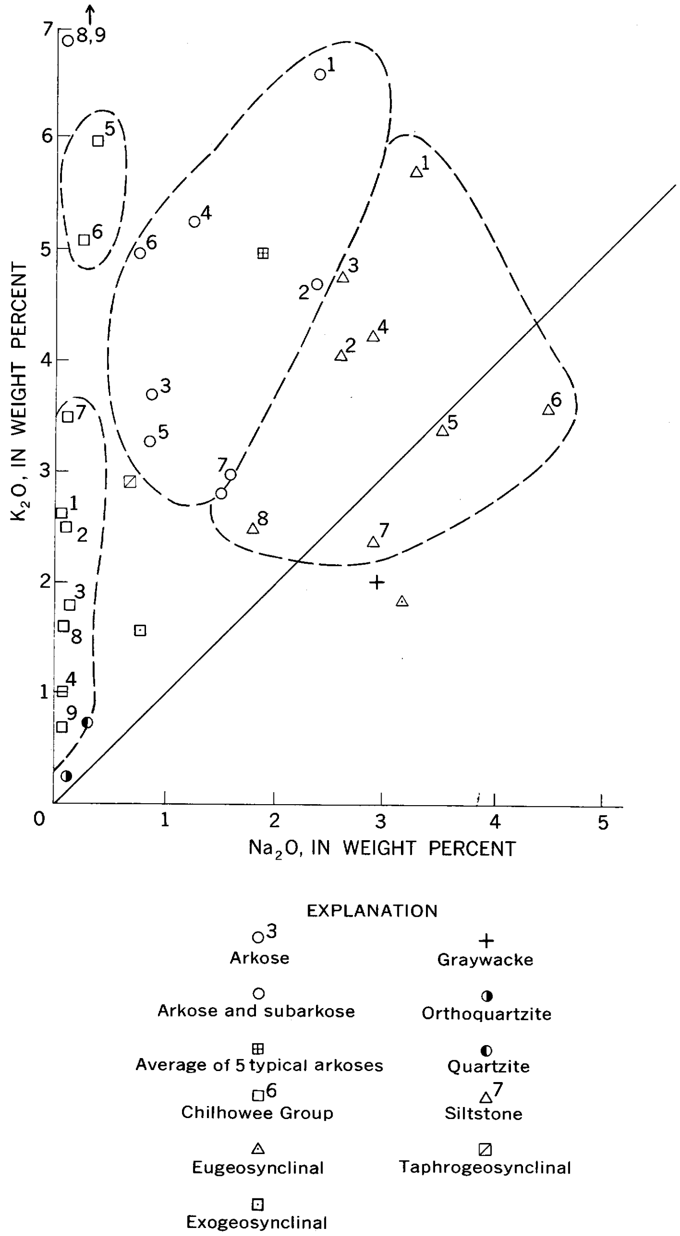



FIGURE 46.—$Na_2O:K_2O$ variation diagram for sedimentary rocks of the Grandfather Mountain Formation and the Chilhowee Group. Numbers of analyzed specimens refer to analyses in tables: arkose (table 15); siltstone (table 16); Chilhowee Group (table 18); mean compositions of graywacke, orthoquartzite, quartzite, and arkose and subarkose (Pettijohn, 1963); averages of taphrogeosynclinal, exogeosynclinal, and eugeosynclinal sandstones (Middleton, 1960); average of five typical arkoses (Pettijohn, 1957).

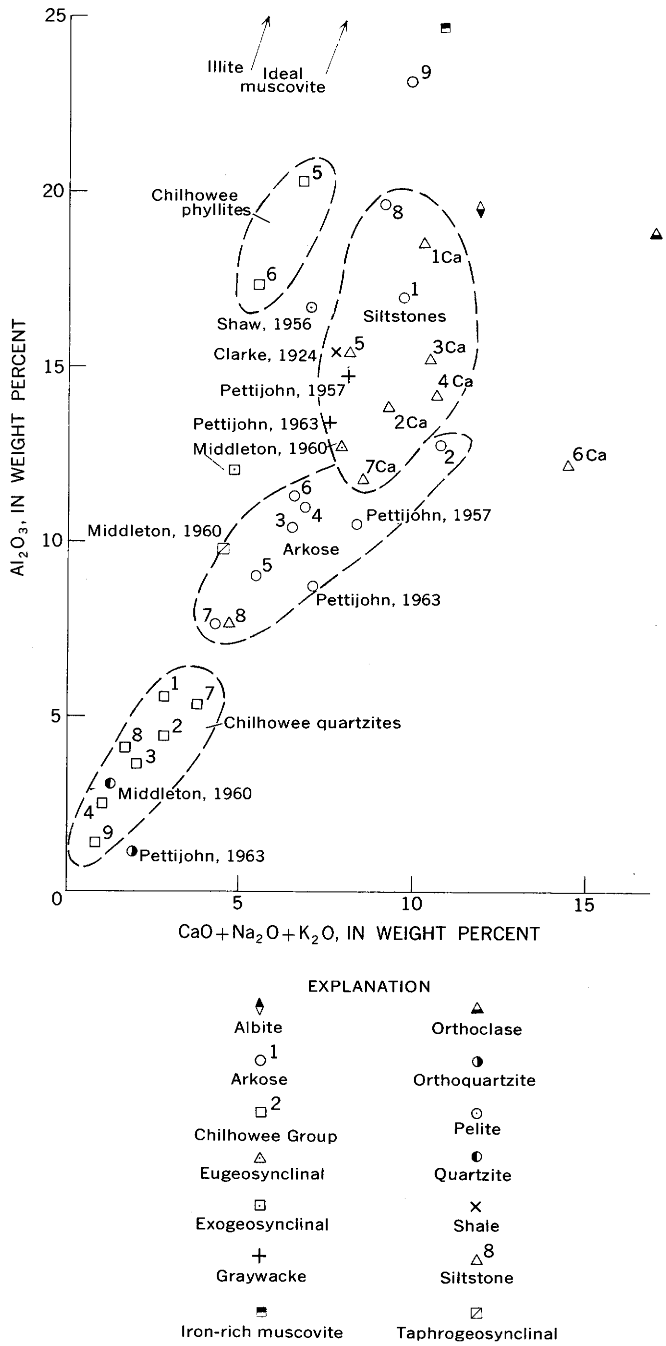



FIGURE 47.—Plot of $Al_2O_3$ against $CaO+Na_2O+K_2O$ in sedimentary rocks of the Grandfather Mountain Formation and the Chilhowee Group. Positions of selected minerals and sedimentary rocks shown. Numbers of analyzed specimens refer to analyses in tables: arkoses of the Grandfather Mountain Formation (table 15); siltstones of the Grandfather Mountain Formation (table 16); rocks of the Chilhowee Group (table 18). Ca indicates calcite content greater than 1 percent.

tween ridges of arkose, such as the one between Linville (area E–4, pl. 1) and Shulls Mills (area G–3).

The siltstone units consist of dark-blue-gray, gray, green-gray, and light-gray, fine-grained, thin-bedded chlorite- and biotite-bearing silstone, phyllite, and graywacke. They locally contain calcareous or dolomitic sandstone layers and lenses and a few lenses of

*A*

*B*

FIGURE 48.—Siltstone in the Grandfather Mountain Formation. *A*, Calcareous phyllite containing numerous quartz-calcite segregations. Roadcut along North Carolina Highway 105, about 1 mile northeast of Foscoe (area F–3, pl. 1). *B*, Laminated argillite and siltstone exposed on a cleavage plane. Roadcut along old North Carolina Highway 195 north of Blevens Creek about 0.1 mile from the Linville Falls Fault (area D–3, pl. 1).

sandy marble. Light-gray to white fine-grained sandstone and phyllite or siltstone in many places alternate in beds 1 mm to 2 inches thick (fig. 48*A*). Very thinly laminated argillite and siltstone beds occur in the uppermost siltstone unit (fig. 48*B*). Arkose beds as much as 20 feet thick are found locally. The sandstones in many places are calcareous and in a few places grade to beds of sandy marble. The marble beds average a foot or two thick but in places are as much as 5 feet thick. All the carbonate-bearing rocks become brown and vuggy upon weathering. Thin sandy beds are lenticular in places, although their shape in some outcrops is plainly due to having been broken and drawn out by movement along cleavage planes that cut across the bedding. Graded bedding is found locally. Graywacke and pebbly graywacke are poorly bedded. They contain clastic grains of plagioclase and quartz and pebbles of mafic volcanic rock, fine-grained quartz-plagioclase rock, and dark sericite phyllite. Crossbedding, which is very widespread in the light-colored arkose, is absent in the gray biotitic and chloritic graywackes.

Many of the calcareous phyllites (fig. 48*A*) contain segregations of quartz and calcite which locally also contain chlorite and rarely, purple fluorite. The calcite is pink in many places.

The lowest siltstone unit in the Blowing Rock quadrangle (areas G–3 and H–3, pl. 1) is siltier than the others and contains interbeds of volcanic material and numerous sandstone beds and lenses. On Flattop Mountain and east of the New River (area H–3), sandstone and volcanic beds are particularly abundant, and the map units, which are quite valid for the rest of the outcrop area of the Grandfather Mountain Formation, become less useful. The middle siltstone, which is continuous from Long Arm Mountain (area D–6) to Hodges Gap (area G–2), is the most calcareous of the siltstone units. Pebbly graywacke is abundant in the uppermost siltstone unit next to the Linville Falls fault in the northwestern part of the Grandfather Mountain window.

### Texture and Mineralogy

The rocks of the siltstone units are generally completely recrystallized because of their original small grain size (fig. 49*A* and *B*). The coarser grained graywackes exhibit relict sedimentary textures (fig. 49*C*) like those in the arkose units; grains larger than 0.1–0.2 mm retain their clastic outlines, and rock fragments, their original textures (fig. 49*D*). Some of the siltstone, especially in the uppermost unit in the northwestern part of the window, seems to have recrystallized less completely; rather fine grained quartz and plagioclase has a fragmental rather than a mosaic texture (fig. 49*C*).

Quartz and plagioclase generally occur in recrystallized grains 0.01 to 0.15 mm in diameter. Clastic grains as much as 3 mm in diameter are found in the coarser or more poorly sorted rocks. Both clastic and recrystallized plagioclase are now albite.

Sericite is the most abundant micaceous mineral. It is 0.01 to 0.2 mm long and well alined. In places it has been deformed by later slip cleavage. A few 0.3- to 0.4-mm grains of muscovite seem to be clastic. No paragonite was detected in several samples of sericite phyllite which were examined by X-ray diffractometer by E. J. Young, U.S. Geological Survey. Probably most of the sericite is muscovite.

Biotite is mostly green and greenish brown, but in the lowest siltstone unit, brown biotite is dominant. Brown biotite is rare in the other siltstone units. Biotite is 0.01 to 0.4 mm long and not as well alined as sericite.

Light-green FeMg chlorite is intergrown with sericite and biotite.

Calcite is a major constituent in some rocks, especially in the lowerst stilstone unit in the Linville quadrangle. It is in grains 0.02 to 0.8 mm in diameter.

Abundant accessory minerals are zircon, sphene, tourmaline, apatite, epidote, ilmenite, magnetite, and pyrite. Opaque minerals are the major constituent of some rocks of the siltstone units, especially in the Blowing Rock quadrangle. Less abundant minerals are allanite and stilpnomelane.

Clastic grains of microcline, hornblende, and epidote are found rarely.

### Composition and Classification

Estimated modal compositions of rocks from the three major siltstone units in the Grandfather Mountain Formation are plotted in figure 50. Chemical analyses (table 16) indicate that modes of these fine-grained rocks are unreliable, but the diagram can be used to compare the units and place the analyzed specimens in relation to the compositional range within each unit. Little information is available from the literature on chemical and mineral compositions of siltstone compared with sandstone. In the Grandfather Mountain Formation, the siltstones lack potassic feldspar and are much richer in micaceous minerals than the arkose. Biotite and chlorite are

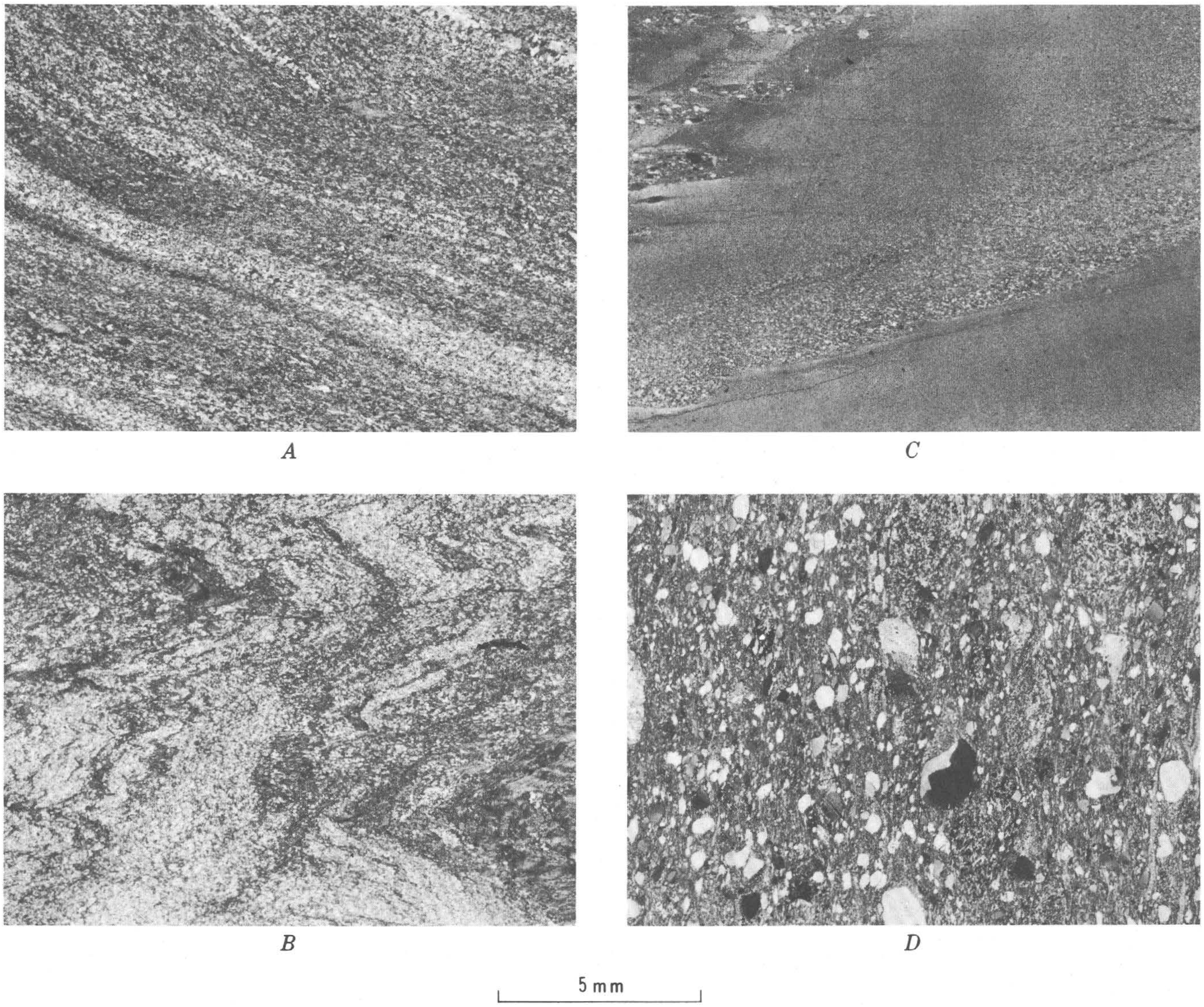



FIGURE 49.—Photomicrographs of siltstone in the Grandfather Mountain Formation. *A*, Chlorite- and biotite-bearing sericite phyllite from roadcut along North Carolina Highway 183 about 0.9 mile west of Cranberry Knob (area D–6, pl. 1). Thin laminae rich in quartz and mica probably represent original silty and shaly beds. Analyzed specimen 5, table 16. *B*, Chlorite-sericite phyllite and siltstone from uppermost exposed siltstone unit at about 3,000-foot altitude west of first sharp bend in Watauga River (area F–2, pl. 1). Silty and shaly laminae tightly folded; axial planes are parallel with north-south cleavage of that area. *C*, Graded bed cut by cleavage in thinly laminated siltstone and argillite from uppermost exposed siltstone unit. North side of hill 3945 north of Newland (area D–4, pl. 1). Outcrop resembles rock shown in figure 48*B*. *D*, Chlorite- and biotite-bearing sericite subgraywacke from roadcut along North Carolina Highway 194, 0.7 mile south of Miller Gap (area D–4, pl. 1). Clastic grains of quartz and plagioclase (pseudomorphosed by albite) and fragments of fine-grained plagioclase-rich felty-textured igneous rock in a matrix of sericite, quartz, chlorite, and biotite.

present in the siltstone but are almost absent in the arkoses.

The lowest siltstone unit (fig. 50*A*) generally contains more mica than the others. The middle siltstone (fig. 50*B*) contains more plagioclase. About half the samples from each of the siltstone units contain biotite. The lowest siltstone unit contains more chlorite and opaque minerals than the others, probably because of admixtures of sedimentary or pyroclastic materials derived from mafic volcanic rock. Calcite occurs in about half the specimens studied from the middle siltstone, and it constitutes more than 10 percent of the rock in about a sixth of the specimens examined.

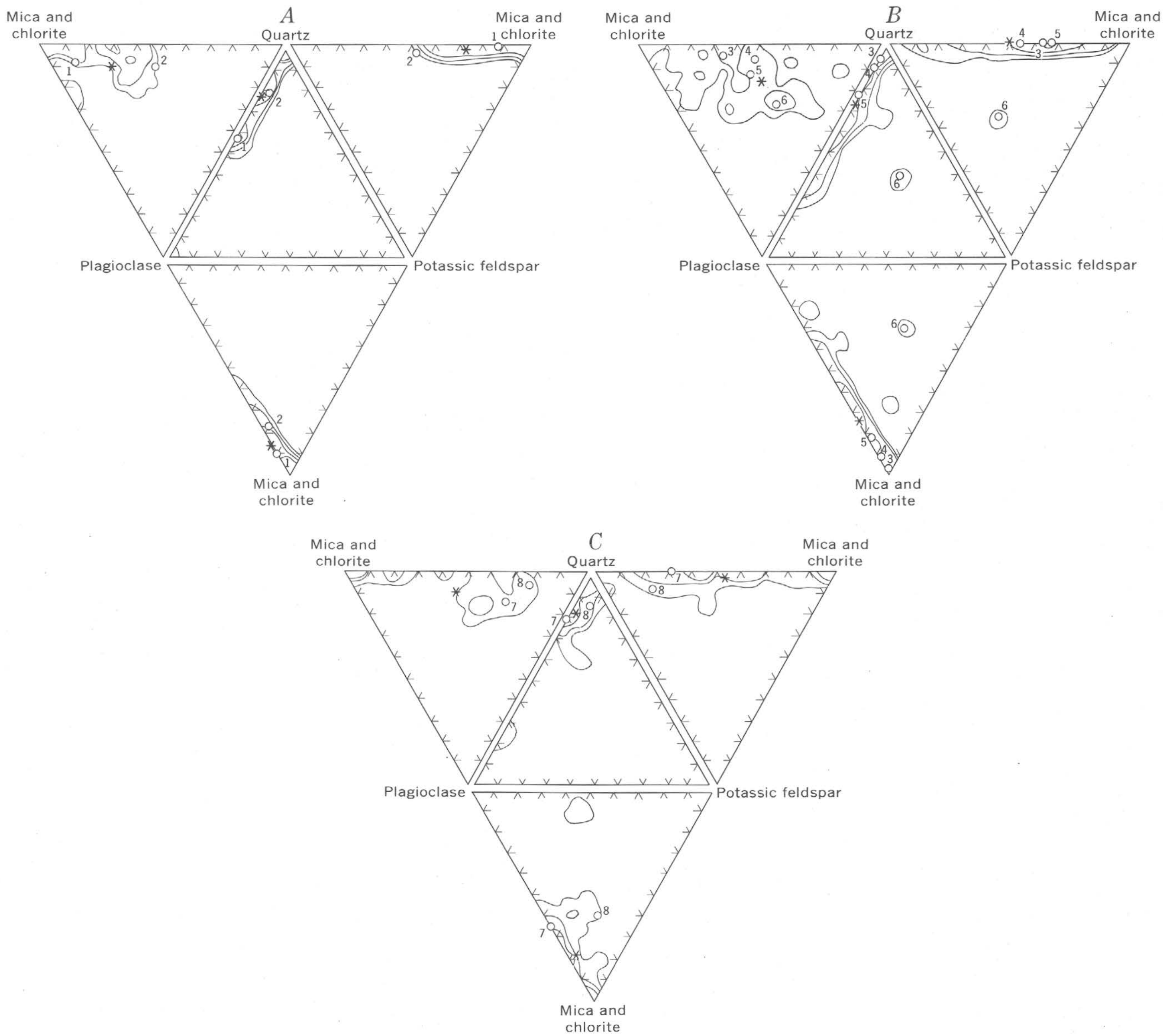



FIGURE 50.—Proportions of quartz, plagioclase, potassic feldspar, and combined mica and chlorite in the siltstones of the Grandfather Mountain Formation. Numbers of analyzed specimens refer to analyses in table 16. *, average. *A*, Lowest siltstone unit. Based on counts of 50 random grains in each of 24 thin sections. Contours 4, 8, 17 and 33 percent. *B*, Middle siltstone unit. Based on counts of 50 random grains in each of 30 thin sections. Contours 3, 7, 13, and 27 percent. *C*, Highest exposed siltstone unit. Based on counts of 50 random grains in each of 19 thin sections. Contours 5, 10, 21, and 42 percent.

The mineral composition diagrams indicate that analysis 1 (table 16) is most typical of the lowest siltstone unit; analysis 5, of the middle unit; and analysis 7, of the uppermost unit. The analyses indicate that the stratigraphically higher siltstone units are generally poorer in $Al_2O_3$ and alkalis than those lower in the section. This is in accord with the mineral composition data, even though X-ray studies and chemical analyses indicate that the amount of albite in most rocks has been underestimated. The rocks of the siltstone units have a lower $K_2O:Na_2O$ ratio than the arkoses (fig. 46); some of the values are less than one. Sandstones intercalated with the siltstones (analyses 6 and 7) would lie in the field of eugeosynclinal rocks of Middleton (1960), most of which are graywackes.

The rocks of the siltstone units generally have a higher ratio of alkalis and CaO to alumina than most shales and argillites (fig. 47). This difference is due to their higher plagioclase content.

TABLE 16.—*Chemical analyses and norms of siltstone and related rocks of the Grandfather Mountain Formation*

[Samples 1, 4, and 6, standard rock analyses by C. L. Parker, U.S. Geol. Survey, 1961; other samples analyzed by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey, 1962; $CO_2$ of sample 3 determined by I. C. Frost, Denver, Colo., 1964. Major oxides and CIPW norms given in weight percent]

| | Lower siltstone unit | | Middle siltstone unit | | | Upper siltstone unit | | |
|---|---|---|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 |
| Field No. | GMB–2 | X–13–2–a | O–58–2 | GML–13 | 2–268 | GML–12 | R–57–2–a | P–37–1–b |
| Laboratory No. | H–3263 | 160188 | 160187 | H–3262 | 160197 | H–3425 | 160195 | 160196 |
| **Major oxides** | | | | | | | | |
| $SiO_2$ | 55.65 | 66.7 | 60.5 | 60.65 | 64.0 | 64.91 | 68.3 | 82.6 |
| $Al_2O_3$ | 18.61 | 14.0 | 15.4 | 14.35 | 15.6 | 12.45 | 11.9 | 7.7 |
| $Fe_2O_3$ | 6.62 | 2.5 | 5.2 | 2.34 | 3.3 | .22 | 3.7 | 1.4 |
| FeO | 2.14 | 2.0 | 1.6 | 3.69 | 3.0 | 1.31 | 2.0 | .69 |
| MgO | 1.94 | 1.5 | 2.5 | 4.51 | 2.4 | .50 | 1.3 | .79 |
| CaO | 1.32 | 2.5 | 3.0 | 2.47 | 1.1 | 6.25 | 3.3 | .28 |
| $Na_2O$ | 3.26 | 2.6 | 2.6 | 3.87 | 3.5 | 4.51 | 2.9 | 1.8 |
| $K_2O$ | 5.74 | 4.1 | 4.8 | 4.26 | 3.4 | 3.69 | 2.4 | 2.5 |
| $H_2O+$ | 1.98 | 1.4 | 2.0 | 1.34 | 1.9 | .41 | 1.4 | .92 |
| $H_2O-$ | .06 | .07 | .11 | .04 | .06 | .07 | .08 | .08 |
| $TiO_2$ | .77 | .68 | .71 | .64 | .92 | .34 | .98 | .36 |
| $P_2O_5$ | .20 | .22 | .19 | .12 | .30 | .22 | .23 | .08 |
| MnO | .04 | .06 | .08 | .12 | .12 | .08 | .08 | .03 |
| $CO_2$ | .84 | 1.7 | 1.74 | .63 | .07 | 4.60 | 1.4 | .13 |
| BaO | .07 | ---- | ---- | .05 | ---- | ---- | ---- | ---- |
| F | .49 | ---- | ---- | .61 | ---- | .08 | ---- | ---- |
| S | .00 | ---- | ---- | .00 | ---- | ---- | ---- | ---- |
| Cl | .00 | ---- | ---- | .00 | ---- | .00 | ---- | ---- |
| Subtotal | 99.73 | ---- | ---- | 99.69 | ---- | 99.64 | ---- | ---- |
| Less O | .21 | ---- | ---- | .26 | ---- | .03 | ---- | ---- |
| Total | 99.52 | 100 | 100 | 99.43 | 100 | 99.61 | 100 | 99 |
| **CIPW norms** | | | | | | | | |
| Q | 12.60 | 33.30 | 22.16 | 11.95 | 25.08 | 23.01 | 37.73 | 61.38 |
| C | 7.04 | 5.21 | 4.96 | 2.22 | 5.05 | 1.04 | 2.33 | 2.02 |
| Or | 33.91 | 24.22 | 28.36 | 25.17 | 20.09 | 21.80 | 14.18 | 14.77 |
| Ab | 27.57 | 21.99 | 21.99 | 32.73 | 29.60 | 38.14 | 24.53 | 15.22 |
| An | ---- | .22 | 2.64 | 3.15 | 3.05 | ---- | 6.02 | .04 |
| En | 3.58 | 3.73 | 6.22 | 11.23 | 5.98 | 1.24 | 3.24 | 1.97 |
| Fs | ---- | .60 | ---- | 4.01 | 1.49 | 1.81 | ---- | ---- |
| Mt | 4.80 | 3.62 | 3.36 | 3.39 | 4.78 | .32 | 3.87 | 1.28 |
| Hm | 3.31 | ---- | 2.88 | ---- | ---- | ---- | 1.03 | .52 |
| Il | 1.46 | 1.30 | 1.35 | 1.22 | 1.75 | .65 | 1.86 | .68 |
| Ap | .47 | .52 | .45 | .28 | .71 | .52 | .54 | .19 |
| Fr | .99 | ---- | ---- | 1.24 | ---- | .14 | ---- | ---- |
| Mg | 1.05 | ---- | ---- | ---- | ---- | .01 | ---- | ---- |
| Cc | .66 | 3.87 | 3.96 | 1.43 | .16 | 10.45 | 3.18 | .30 |

NOTE.—Minor-element analyses for samples 1, 4, and 6 given in table 1.

1. Gray calcareous metasiltstone containing light-green sericite, quartz, and albite with a grain size of 0.02 to 0.06 mm, subordinate magnetite and calcite, and accessory sphene and apatite. From roadcut at sharp outward curve 0.4 mile N. 80° E. from the top of Martin Knob (area G–3, pl. 1).
2. Medium greenish-gray, poorly sorted calcareous sandy biotite-sericite phyllite. Clastic grains of quartz and plagioclase (altered to albite) as much as 1 mm in diameter in a groundmass of recrystallized sericite, quartz, and albite with a grain size of 0.06 mm and subordinate brown biotite and calcite. Accessory sphene, opaque minerals, chlorite, Zircon, and apatite. From roadcut along U.S. Highway 221, 150 feet southeast of spot altitude of 3,803 feet east of Raven Rocks (area G–3, pl. 1).
3. Dark blue-gray fine-grained biotite- and calcite-bearing sericite phyllite. Sericite, quartz, and albite 0.05 mm in grain size and subordinate greenish-brown biotite, calcite, and opaque minerals. Accessory sphene and apatite. Roadcut along U.S. Highway 221 in area E–5 (pl. 1) at first curve south of the village of Linville (area E–4).
4. Dark-gray calcareous biotite-sericite phyllite. Segregations of calcite, quartz, and chlorite contain some purple fluorite. Sericite, green biotite, quartz, and albite 0.03 to 0.16 mm in grain size with subordinate epidote and calcite and accessory sphene. Roadcut along North Carolina Highway 105, 0.4 mile S. 73° W., spot altitude of 3,007 feet in village of Foscoe (area F–3, pl. 1).
5. Dark-gray phyllite, Sericite, quartz, and albite with a grain size of 0.05 to 0.1 mm and subordinate epidote, FeMg chlorite, sphene, opaque minerals, green biotite, and accessory calcite and apatite. A few grains of clastic plagioclase (altered to albite) as much as 1 mm in diameter. Roadcut along North Carolina Highway 183 about 0.7 mile due west of Cranberry Knob (area D–6, pl. 1).
6. Light-gray calcareous biotite arkose. Albite, microcline, and quartz grains are 0.05 to 0.5 mm in diameter; larger grains retain clastic outlines. Subordinate green biotite and calcite. Accessory minerals are sphene, apatite, epidote, allanite, opaque mineral, and zircon. From a 2-foot bed in siltstone in roadcut on North Carolina Highway 105, 0.5 mile southwest of the settlement of Grandfather (area F–3, pl. 1).
7. Dark-greenish-gray pebbly graywacke containing fragments of mafic volcanic rock, sericite phyllite, and granoblastic quartz-plagioclase rock. Contains clastic grains of quartz and plagioclase (altered to albite) as much as 1.5 mm in diameter in a groundmass of quartz, albite, and sericite and subordinate epidote, calcite, sphene, and opaque minerals. Grains in matrix are 0.005 to 0.3 mm in diameter. From roadcut on the north side of a sharp bend of Watauga River at spot altitude of 2,718 feet (area F–2, pl. 1).
8. Gray sandy biotite- and chlorite-bearing metasiltstone. Contains a few fragments of mafic volcanic rock. Clastic grains of quartz as much as 2 mm in diameter and microcline and plagioclase (altered to albite) as much as 1 mm in diameter in a matrix of quartz and sericite 0.01 to 0.05 mm in grain size and accessory greenish-brown biotite, FeMg chlorite, sphene, opaque minerals, calcite, apatite, and tourmaline. From roadcut along North Carolina Highway 194, 0.1 mile northeast of Smoky Gap (southwest corner of area D–3, pl. 1).

Analysis 6 is of a rock from an atypical bed; it has a high ratio of CaO and alkalis to $Al_2O_3$ and contains more calcite and feldspar than the other samples.

## VOLCANIC ROCKS

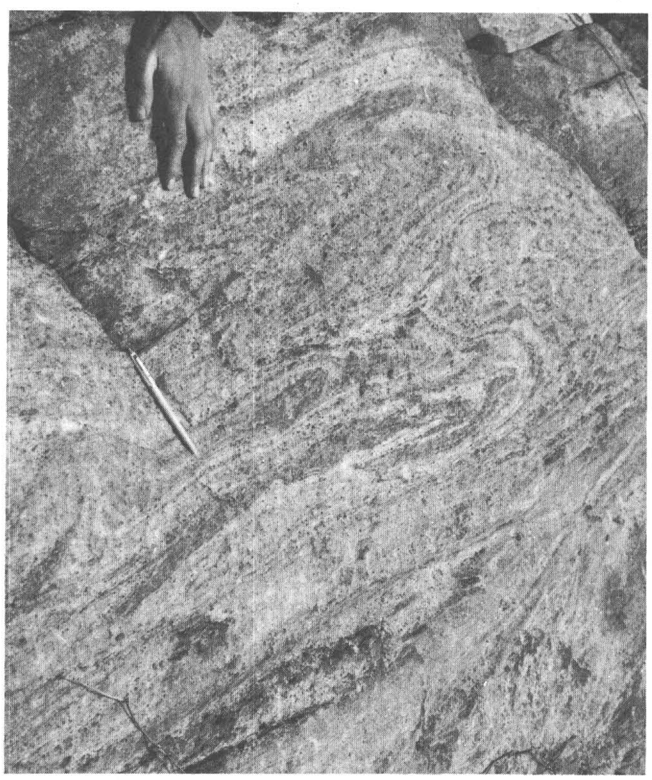
### ROCKS OLDER THAN THE MONTEZUMA MEMBER OF GRANDFATHER MOUNTAIN FORMATION

#### Felsic Volcanic Rocks

Felsic flows, tuffs, and tuffaceous sedimentary rocks occur in the basal part of the Grandfather Mountain Formation, especially in the isolated outcrop areas southeast of the main outcrop belt. It is often difficult to distinguish felsic flows from thick units of tuff or welded tuff even in unmetamorphosed rocks; as all rocks in the Grandfather Mountain Formation are metamorphosed, it is impossible to tell what proportion of the felsic volcanic rocks are of pyroclastic origin. A few dikes and small plugs of felsite cut basement rocks in the Linville Falls and Lenoir quadrangles, but most of them are too small to map.

Some bodies of felsic rocks are homogeneous, but others are heterogeneous, especially near their margins. The body underlying the Montezuma Member of the Grandfather Mountain Formation in the northern part of the window contains interbedded arkose, siltstone, and graywacke adjacent to the Linville Falls fault. A small basalt flow is intercalated with felsic rocks in this body on the lower part of Hodges Mountain (area G–2, pl. 1). Where it is exposed along North Carolina Highway 105 farther to the southwest, the body is homogeneous and coarsely porphyritic.

Along Wilson Creek in the southeast part of area F–6, basalt, arkose, and tuffaceous arkose are interbedded with the felsic volcanic rocks. Faint color banding and differences in grain size may represent bedding, but in some outcrops such differences are obviously flow banding. Adjacent to some of the basalt flows, intricate convolution of color banding (fig. 51*A*) and pseudointrusive relations between the felsic rocks and the basalt may indicate that the underlying felsic material was unconsolidated at the time of extrusion of the basalt. In a few exposures,



*A*

*B*

FIGURE 51.—Felsic volcanic rocks. *A*, Contorted flow banding in felsic volcanic rocks of Grandfather Mountain Formation. East side of Wilson Creek about 0.1 mile upstream from bench mark 1410 (area F–6, pl. 1). *B*, Fragments of basement rock in a matrix of coarse nonbedded volcanic sandstone or tuff. North side of Wilson Creek, 0.15 mile from west edge of the Lenoir quadrangle (area G–6, pl.1).

faint suggestions of crossbedding or local concentrations of lithic fragments suggest that the rocks are water-laid crystal tuffs or tuffaceous sedimentary rocks. Cobbles and boulders of Brown Mountain Granite (fig. 51*B*) are found in nonbedded felsic volcanic rocks at the base of the unit north of Brown Mountain (area G–6, pl. 1).

A few thin felsic flows or beds of tuff occur in the mafic volcanic units at the base of the Grandfather Mountain Formation near Phillips Creek and Gragg Fork (area G–4, pl. 1). A 6-inch bed of felsic volcanic rock at the base of the arkose northeast of Tablerock Mountain (area D–7, pl. 1) is identical with that in the larger extrusive and intrusive bodies and must be a tuff.

The felsic volcanic rocks are light to medium gray and superficially resemble vitreous quartzite, arkose, or mylonitic gneiss. In most outcrops, however, euhedral or partly resorbed phenocrysts of gray to blue-gray quartz 1 to 2 mm in diameter are found. In some exposures, euhedral tan or light pink phenocrysts of potassic feldspar as much as 1 cm long are conspicuous. Some rocks, especially those underlying the Montezuma Member, contain plagioclase phenocrysts. Quartz segregation veinlets are abundant, and felsic volcanic rocks in the southeastern part of the window contain widespread quartz-microcline segregations. Foliation is defined by sericite flakes and elongate lenses and laminae of somewhat coarser grained quartz and feldspar in the fine-grained matrix. It is most conspicuous in the rocks near the Linville Falls fault east of the Brown Mountain Granite.

In thin section the felsic volcanic rocks contain euhedral to anhedral embayed quartz phenocrysts and euhedral phenocrysts of microperthite and microantiperthite in a fine-grained groundmass composed of quartz, feldspar, sericite, and opaque minerals (fig. 52).

Perthite in the phenocrysts has a fine-grained patchy pattern that differs from the coarser and more veinlike pattern of perthite in the plutonic rocks. Both the felsic volcanic rocks immediately below the Montezuma Member and those southwest of Brown Mountain contain plagioclase phenocrysts which have been altered to albite. Some of these plagioclase grains have more complex twinning than that found in plagioclase from the underlying plutonic rocks (fig. 52*C*). Some large feldspar grains seem to be crystal fragments. In some rocks, the fine-grained muscovite is light green and is probably iron rich. The groundmass is so fine grained (0.005 to 0.05 mm) in many specimens (fig. 52*A*) that it is difficult to estimate its composition. The groundmass is generally coarser grained in the rocks north and east of Brown Mountain in the Linville Falls and Lenoir quadrangles than elsewhere (fig. 52*C* and *D*). Quartz veinlets and lenses consist of recrystallized mosaic-textured grains 0.05 to 0.2 mm in diameter. Pyrite is a common accessory mineral. A few rocks contain scattered radiating aggregates of stilpnomelane as much as 0.25 mm in diameter. Zircon occurs as subhedral to euhedral prisms as much as 0.25 mm long in the volcanic rocks; in tuffaceous sedimentary rocks both euhedral and round zircon grains are found. Both green and brown biotite are found, and biotite constitutes as much as 25 percent of the rock. Other accessories are epidote, sphene, allanite, calcite, chlorite, apatite, magnetite, dark-blue-green amphibole, and fluorite.

Chemical analyses (table 17, Nos. 8–16) show that the felsic volcanic flows are rhyolite and quartz latite. Rocks believed to be tuffs or tuffaceous sedimentary rocks are chemically similar to the flow rocks.

#### Mafic Volcanic Rocks

Mafic flows, flow breccias, or tuff breccias occur locally in the lower part of the Grandfather Mountain Formation below the Montezuma Member. All the larger bodies of mafic volcanic rocks contain all these rock types and interfinger with sedimentary rocks at their margins. Beds of tuff and tuffaceous sedimentary rocks, and possibly even very thin flows, are widely distributed in the siltstone units in the Blowing Rock quadrangle and are more numerous near the mapped bodies of mafic volcanic rocks. Some mafic dikes found in the basement rocks may be correlative with the extrusive rocks.

Mafic flows and associated pyroclastic rocks on the southwest slopes of Brown Mountain (area F–7, pl. 1) and near the southeastern contact of the main body of the Grandfather Mountain Formation near Ripshin Ridge (areas E–6 and E–7) are interbedded with lustrous blue phyllite.

The mafic volcanic rocks are dark to light blue gray, gray, green gray to green, and generally contain phenocrysts or crystal fragments of plagioclase that are 5 mm but locally as much as 2 cm long. The large plagioclase grains are commonly partly replaced by calcite. Brown weathered calcite and quartz-calcite segregations are common; chlorite segregations are rare. A few rocks contain segrega-

TABLE 17.—*Chemical analyses and norms of Linville Metadiabase and volcanic rocks of the Grandfather Mountain Formation*

[Samples 1, 2, 3, 6, 8, and 10, standard rock analyses by D. F. Powers, U.S. Geol. Survey, 1959; samples 7 and 16, by C. L. Parker, U.S. Geol. Survey, 1961; other samples analyzed by rapid methods by Paul Elmore, I. H. Barlow, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey, 1962. Sample 4, by W. F. Hillebrand (Clarke, 1900). Major oxides and CIPW norms given in weight percent]

| Map unit (pl. 1) | Linville Meta-diabase | Mafic volcanic rocks of the Montezuma Member | | | Other mafic volcanic rocks | | | Felsic volcanic rocks | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 [1] | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 | 14 | 15 | 16 |
| Field No | GML–3 | GML–1 | GML–2 | ------ | 10–527A | W–84 | GMB–6 | GMB–1 | 24–27 | 34(L) | U–29 | 10–573 | 8V–13 | 8W–85B | RE–94–4 | GMB–5 |
| Laboratory No | F–2418 | F–2416 | F–2417 | ------ | 160149 | F–2419 | H–3426 | F–2591 | 160151 | 14057 | 154236 | 160150 | 160153 | 160152 | 160154 | H–3265 |
| **Major oxides** | | | | | | | | | | | | | | | | |
| $SiO_2$ | 42.91 | 45.26 | 46.09 | 47.85 | 45.6 | 46.08 | 46.15 | 66.68 | 67.7 | 68.87 | 69.0 | 72.8 | 73.1 | 75.2 | 76.2 | 76.79 |
| $Al_2O_3$ | 15.98 | 15.55 | 15.83 | 16.51 | 17.3 | 15.45 | 16.36 | 13.76 | 14.1 | 13.57 | 14.3 | 12.5 | 13.0 | 12.4 | 12.1 | 11.10 |
| $Fe_2O_3$ | 2.78 | 2.45 | 9.28 | 4.16 | 6.3 | 5.08 | 6.35 | 5.11 | 3.6 | 2.45 | 3.5 | 1.0 | 1.7 | 2.2 | 1.6 | 1.15 |
| FeO | 12.82 | 10.45 | 4.95 | 7.43 | 6.0 | 6.90 | 7.79 | .80 | 3.1 | 2.17 | .94 | 2.8 | 1.2 | .95 | .70 | .72 |
| MgO | 5.98 | 6.88 | 6.46 | 6.24 | 6.5 | 3.64 | 2.61 | .67 | .59 | .39 | .56 | .41 | .48 | .44 | .36 | .09 |
| CaO | 7.98 | 10.12 | 6.80 | 7.00 | 6.3 | 8.49 | 5.80 | 1.60 | 1.6 | 1.73 | 1.0 | .87 | .78 | .45 | .39 | .61 |
| $Na_2O$ | 2.55 | 2.42 | 3.41 | 3.20 | 4.0 | 4.90 | 4.79 | 2.47 | 3.2 | 3.54 | 3.5 | 2.2 | 2.1 | 1.1 | 2.2 | 1.62 |
| $K_2O$ | .92 | .54 | .70 | .82 | .09 | .14 | 1.68 | 6.03 | 4.7 | 5.08 | 5.5 | 5.4 | 6.3 | 5.7 | 4.9 | 6.52 |
| $H_2O+$ | 3.97 | 2.41 | 3.42 | 4.00 | 4.0 | 2.86 | 2.84 | 1.04 | .89 | .54 | .56 | .93 | .72 | .91 | .74 | .26 |
| $H_2O-$ | .07 | .09 | .14 | .21 | .06 | .02 | .05 | .03 | .05 | .05 | ------ | .06 | .05 | .04 | .04 | .04 |
| $TiO_2$ | 3.28 | 2.23 | 2.21 | 2.28 | 2.8 | 2.50 | 2.64 | .77 | .62 | .50 | .36 | .35 | .25 | .31 | .19 | .14 |
| $P_2O_5$ | .36 | .33 | .33 | .35 | .50 | .39 | 1.34 | .21 | .22 | .12 | .09 | .04 | .02 | .02 | .01 | .01 |
| MnO | .19 | .20 | .17 | .24 | .24 | .25 | .18 | .05 | .10 | .08 | .10 | .05 | .04 | .02 | .01 | .04 |
| $CO_2$ | .01 | .03 | .01 | ------ | .60 | 3.33 | 1.37 | .41 | <.05 | .61 | .66 | .58 | .41 | <.05 | <.05 | .40 |
| Cl | ------ | ------ | ------ | ------ | ------ | ------ | .01 | ------ | ------ | .01 | ------ | ------ | ------ | ------ | ------ | .01 |
| S | ------ | ------ | ------ | ------ | ------ | ------ | .04 | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | .23 |
| F | ------ | ------ | ------ | ------ | ------ | ------ | .17 | ------ | ------ | .11 | ------ | ------ | ------ | ------ | ------ | .02 |
| BaO | ------ | ------ | ------ | Trace? | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | .02 |
| Subtotal | ------ | ------ | ------ | ------ | ------ | ------ | 100.17 | ------ | ------ | 99.82 | ------ | ------ | ------ | ------ | ------ | 99.77 |
| Less O | ------ | ------ | ------ | ------ | ------ | ------ | .09 | ------ | ------ | .05 | ------ | ------ | ------ | ------ | ------ | .13 |
| Total | 99.80 | 99.96 | 99.80 | 100.29 | 100 | 100.03 | 100.08 | 99.63 | 100 | 99.77 | 100 | 100 | 100 | 100 | 99 | 99.64 |
| **CIPW norms** | | | | | | | | | | | | | | | | |
| Q | ------ | ------ | 1.59 | .51 | 1.13 | 1.45 | 1.23 | 26.54 | 26.58 | 26.37 | 26.69 | 36.83 | 35.37 | 45.44 | 43.32 | 42.20 |
| C | ------ | ------ | ------ | ------ | 1.76 | .46 | 2.81 | 1.72 | 1.37 | 1.10 | 2.52 | 2.90 | 2.31 | 3.65 | 2.49 | 1.28 |
| Or | 5.44 | 3.19 | 4.14 | 4.84 | .53 | .83 | 9.93 | 35.63 | 27.77 | 30.01 | 32.49 | 31.90 | 37.22 | 33.68 | 28.95 | 38.52 |
| Ab | 21.57 | 20.47 | 28.84 | 27.06 | 33.83 | 41.44 | 40.44 | 20.89 | 27.06 | 29.86 | 29.60 | 18.61 | 17.76 | 9.30 | 18.61 | 13.63 |
| An | 29.44 | 29.98 | 25.82 | 28.27 | 24.19 | 18.52 | 10.55 | 3.97 | 6.50 | 3.18 | .20 | .39 | 1.15 | 2.10 | 1.87 | .32 |
| Hl | ------ | ------ | ------ | ------ | ------ | ------ | .02 | ------ | ------ | .02 | ------ | ------ | ------ | ------ | ------ | .02 |
| Wo | 3.22 | 7.46 | 2.37 | 1.86 | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ |
| En | 3.75 | 8.37 | 16.08 | 15.54 | 16.18 | 9.06 | 6.50 | 1.67 | 1.47 | .97 | 1.39 | 1.02 | 1.20 | 1.09 | .90 | .22 |
| Fs | 4.07 | 6.77 | ------ | 6.89 | 1.64 | 4.81 | 4.95 | ------ | 1.88 | 1.28 | ------ | 3.83 | .46 | ------ | ------ | ------ |
| Fo | 7.81 | 6.14 | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ |
| Fa | 9.35 | 5.47 | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ |
| Mt | 4.03 | 3.55 | 10.10 | 6.03 | 9.13 | 7.37 | 9.21 | .51 | 5.22 | 3.55 | 2.31 | 1.45 | 2.46 | 2.23 | 1.74 | 1.22 |
| Hm | ------ | ------ | 2.31 | ------ | ------ | ------ | ------ | 4.76 | ------ | ------ | 1.90 | ------ | ------ | .66 | .40 | .31 |
| Il | 6.23 | 4.24 | 4.20 | 4.33 | 5.32 | 4.75 | 5.01 | 1.46 | 1.18 | .95 | .68 | .66 | .48 | .59 | .36 | .27 |
| Ap | .85 | .78 | .78 | .83 | 1.18 | .92 | 3.17 | .50 | .52 | .28 | .21 | .10 | .05 | .05 | .02 | .02 |
| Fr | ------ | ------ | ------ | ------ | ------ | ------ | .23 | ------ | ------ | .22 | ------ | ------ | ------ | ------ | ------ | .04 |
| Pr | ------ | ------ | ------ | ------ | ------ | ------ | .08 | ------ | ------ | ------ | ------ | ------ | ------ | ------ | ------ | .43 |
| Cc | .02 | .07 | .02 | ------ | 1.36 | 7.57 | 3.12 | .93 | ------ | 1.39 | 1.50 | 1.32 | .93 | ------ | ------ | .91 |

[1] Other elements given: $ZrO_2$, 0.03; $Cr_2O_3$, 0.01; $V_2O_3$, 0.05; CoO, NiO, 0.03; SrO, trace?; $Li_2O$, trace. Complete total 100.41.

NOTE.—Minor-element analyses for samples 1, 2, 3, 6, 7, 8, 10, and 16 given in table 1.

1. Fine-grained gray-green metadiabase containing porphyroclasts of actinolite as much as 4 mm long and ilmenite as much as 2 mm in diameter in a fine-grained matrix of FeMg chlorite, epidote, actinolite, albite, sphene, sericite, ilmenite, and accessory brown biotite and apatite. Roadcut along North Carolina Highway 105 at Linville Gap (area E–4, pl. 1).
2. Greenish-gray greenschist containing actinolite and epidote as much as 0.5 mm in diameter in a matrix composed of FeMg chlorite, albite, sphene, ilmenite, and accessory carbonate and quartz, and relict pigeonite. From abandoned quarry on north side of Bald Mountain (area D–3, pl. 1).
3. Fine-grained gray-green greenschist containing hematite and maghemite grains as much as 5 mm in diameter in a matrix composed of epidote, actinolite, FeMg chlorite, albite, and accessory sericite, apatite, and quartz. Outcrop contains abundant epidote-quartz segregations. Roadcut along North Carolina Highway 184, 0.1 mile south of spot altitude of 3,645 feet, south of Banner Elk (area E–3, pl. 1).
4. Analysis from Clarke (1900, p. 53). "Epidote-chlorite schist one-half mile northeast of Montezuma—contains epidote and feldspar with less chlorite, hornblende, and magnetite." Locality in area D–4 (pl. 1).
5. Fine-grained dark-green faintly schistose greenstone from body 50 to 100 feet thick interlayered with lustrous gray and blue phyllite. Felted mass of MgFe chlorite, albite, epidote, sphene, quartz, and accessory opaque mineral, carbonate, and apatite containing a few plagioclase phenocrysts (altered to albite) as much as 0.8 mm long. From tributary to Steels Creek 0.4 mile northwest of the confluence of Buck Creek and Steels Creek (area E–7, pl. 1).
6. Fine-grained light-greenish-gray schist having relict pilotaxitic texture and containing albite (mainly pseudomorphous after plagioclase laths 0.05 to 0.15 mm long), sericite, epidote, sphene, magnetite, MgFe chlorite (almost isotropic), and carbonate. Rock locally contains amygdules filled with epidote, quartz, and chlorite. From northeast side of Wilson Creek about 0.1 mile north of bridge over Wilson Creek near mouth of Craig Creek (area F–6, pl. 1).
7. Greenish-gray porphyry containing laths and stubby prisms of plagioclase as much as 1 cm long. Plagioclase phenocrysts (altered to albite) in fine-grained groundmass composed of sericite, FeMg chlorite (almost isotropic), magnetite, sphene, epidote, and accessory biotite and apatite. From roadcut along the Blue Ridge Parkway 0.95 mile N. 52° E. of Raven Rocks (area G–3, pl. 1).
8. Greenish-gray metavolcanic rock containing phenocrysts of potassic feldspar and plagioclase as much as 2.5 cm long and quartz as much as 2 mm in diameter. Phenocrysts of perthite, plagioclase (altered to albite), and quartz in a very fine grained groundmass of quartz, potassic feldspar, albite, sericite, and sphene. Accessory minerals are epidote, biotite, opaque mineral, and carbonate. Roadcut along North Carolina Highway 105 about 0.2 mile south of bridge across Watauga River (southwest corner of area G–2, pl. 1).
9. Medium-grained greenish-gray volcanic sedimentary rock or tuff. Contains perthite grains as much as 6 mm in diameter and plagioclase (pseudomorphosed by albite) as much as 3 mm in diameter in mosaic-textured groundmass of recrystallized quartz, albite, and potassic feldspar 0.005 to 0.25 mm in grain size. Matrix contains irregular clots and folia of biotite and epidote. Accessories include opaque minerals, carbonate, sericite, and chlorite. Outcrop in area G–6 (pl. 1) on east side of Wilson Creek, 1.3 miles N. 64° W. of summit of Adams Mountain (area G–7, pl. 1).
10. Fine-grained gray felsic volcanic rock containing phenocrysts of perthite as much as 5 mm in diameter and plagioclase (altered to albite) as much as 2 mm in diameter in a matrix composed of fine-grained microcline, albite, quartz, sericite, biotite, and magnetite. Accessory minerals are sphene, epidote, calcite, and zircon. Rock is strongly sheared and contains quartz-calcite segregations. Cut along road on south side of Johns River 0.17 mile northwest of mouth of Crawley Branch (on boundary between areas G–6 and G–7, pl. 1).
11. Light-gray to white felsite containing phenocrysts of white feldspar 1 mm long and gray quartz blebs 5 mm in diameter. Rock is finely laminated and strongly foliated. Phenocrysts are plagioclase and perthite; groundmass is a mosaic of quartz and feldspar 0.01 to 0.05 mm in grain size containing discontinuous skeins and streaks of sericite. Accessory minerals are magnetite, sphene, calcite, biotite, and epidote. Roadcut on east side of North Carolina Highway 181, 0.18 mile south of bench mark 1195 (north edge of area F–8, pl. 1).
12. Fine-grained dark-blue-gray faintly schistose felsite containing a few feldspar phenocrysts, some as much as 5 mm in diameter. From a dike 15 to 20 feet thick cutting Wilson Creek Gneiss. Phenocrysts of quartz and perthite in a groundmass of mosaic-textured quartz and feldspar 0.01 to 0.05 mm in grain size. Groundmass contains scattered flakes of sericite and biotite and grains of epidote. Accessory minerals are allanite, carbonate, and fluorite. From outcrop in area E–8 (pl. 1) east of Simpson Creek, 1.4 miles west of bench mark 1195 on North Carolina Highway 181 (north edge of area F–8, pl. 1).
13. Medium-grained dark-blue-gray felsite or tuffaceous sedimentary rock containing blue quartz grains (resorbed phenocrysts?) as much as 3 mm in diameter and a few feldspar grains (phenocrysts?) as much as 5 mm in diameter. The large quartz and perthite grains are set in a fine-grained matrix composed of quartz, feldspar, and sericite with a grain size of 0.02 to 0.01 mm. Accessory minerals are allanite, epidote,

TABLE 17.—*Chemical analyses, and norms of Linville Metadiabase and volcanic rocks of the Grandfather Mountain Formation* —Continued

biotite, zircon, carbonate, and opaque minerals. From cut along Wilson Creek road (boundary between areas F–6 and G–6, pl. 1).

14. Medium-grained gray-green rock, probably a lithic crystal tuff or tuffaceous sedimentary rock containing light-pink feldspar grains as much as 2 mm long. Quartz, potassic feldspar, plagioclase (altered to albite), and fragments of fine-grained felsic volcanic(?) rock as much as 1 mm in diameter in a matrix of light-green sericite, quartz, feldspar, and sphene. East end of bridge across Wilson Creek near mouth of Craig Creek (area F–6, pl. 1).

15. Dark-greenish-gray schistose rock, probably lithic crystal tuff or tuffaceous sedimentary rock containing aphanitic clastic fragments as much as 5 mm in diameter. Grains of quartz and potassic feldspar as much as 1.5 mm long and very fine grained fragments of felsic volcanic rock in a fine-grained matrix of recrystallized quartz, feldspar, light-green sericite, and accessory opaque minerals, fluorite, carbonate, and sphene. Same locality as 14.

16. Blue-gray felsite containing quartz and perthite phenocrysts 0.5 to 2.5 mm in diameter in a very fine grained matrix of quartz, sericite, feldspar, and opaque minerals. Accessory minerals are pyrite, allanite, sphene, zircon, and carbonate. From roadcut along Blue Ridge Parkway opposite overlook, 1.25 miles N. 15° W. from junction of U.S. Highways 321 and 221 in the village of Blowing Rock (area H–3, pl. 1).

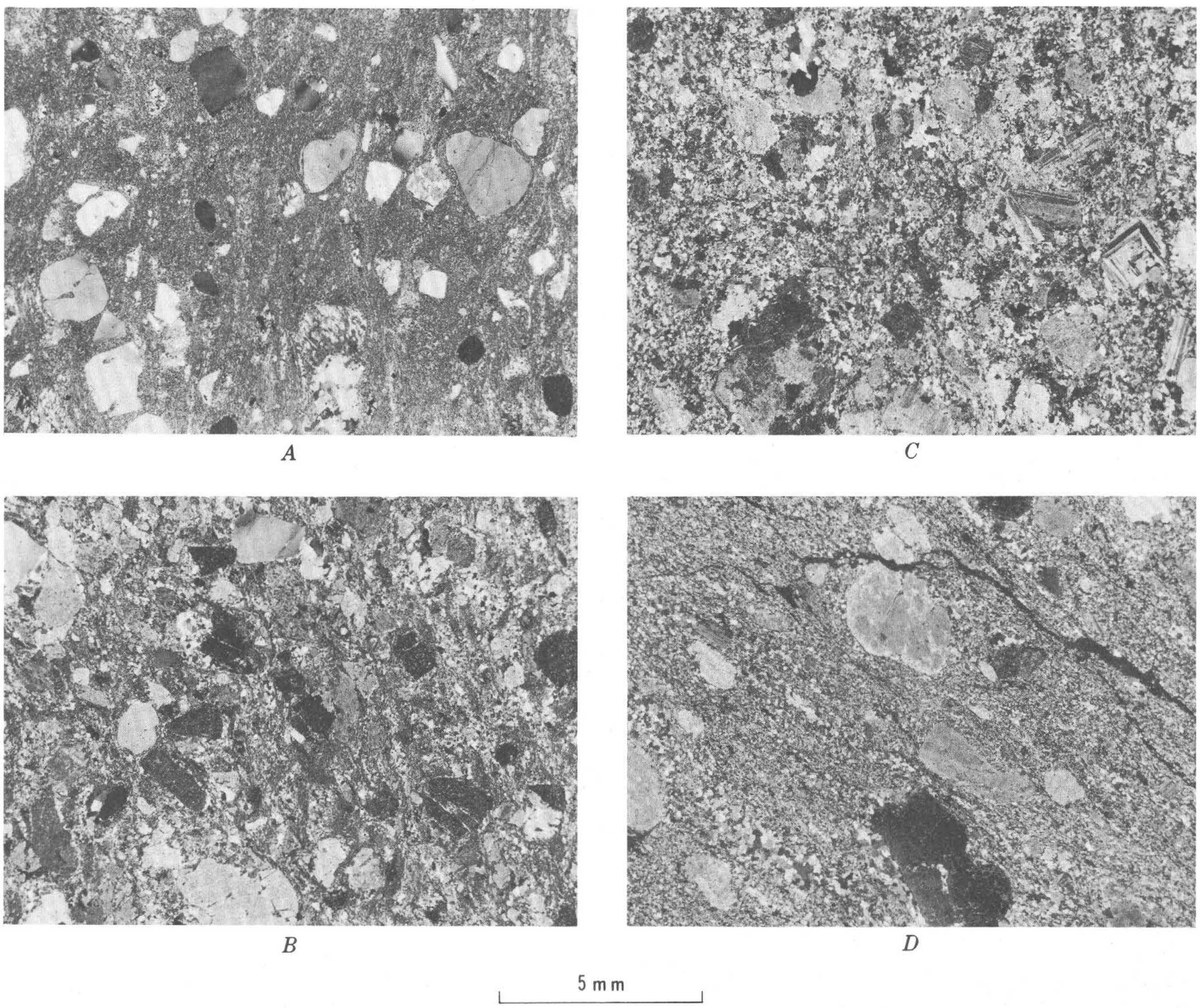



FIGURE 52.—Photomicrographs of felsic volcanic rocks. *A*, Felsic volcanic rock from roadcut along Blue Ridge Parkway about 0.8 mile north of Cone Lake (area H–3, pl. 1). Embayed phenocrysts of quartz and patch perthite in groundmass of very fine grained quartz, feldspar, sericite, and opaque minerals. Analyzed specimen 16, table 17. *B*, Felsic volcanic rock from roadcut along Wilson Creek at east edge of Linville Falls quadrangle (area F–6, pl. 1). Phenocrysts of quartz, fine-textured patch perthite and, rarely, plagioclase (altered to albite) in groundmass of sericite quartz, and feldspar. Analyzed sepcimen 13, table 17. *C*, Volcanic sandstone or tuff from outcrop on north side of Wilson Creek 0.15 mile from west edge of Lenoir quadrangle (area G–6, pl. 1). Phenocrysts of fine-textured perthite and complexly twinned plagioclase (pseudomorphosed by albite) in a groundmass of quartz, albite, potassic feldspar, biotite, and epidote. See figure 51*B* for photograph of outcrop. Analyzed specimen 9, table 17. *D*, Felsic volcanic rock from south side of Johns River in area G–6, pl. 1, 1.2 miles northwest of Collettsville (area H–7, pl. 1). Grains of fine-textured perthite and plagioclase (pseudomorphosed by albite) in a matrix of microcline, albite, quartz, sericite, biotite, and magnetite. Analyzed specimen 10, table 17.

tions of epidote, but they are neither as large nor as ubiquitous as those in the Montezuma Member. Amygdules are as much as 5 cm in diameter and commonly contain quartz and epidote and rarely calcite, albite, or chlorite (fig. 53). Some rocks contain enough magnetite to deflect a compass needle.

Some of the rocks have relict felty texture formed by laths of plagioclase (now albite) generally 0.02 to 0.5 mm long, and in some rocks, gradations in size between laths in the groundmass and lathshaped phenocrysts are found. Other rocks contain stubby euhedral phenocrysts (fig. 54) or anhedral crystal fragments of plagioclase which are bent, broken, and healed by metamorphic minerals and have been altered to calcite and albite. A few have rounded outlines suggesting partial resorption. All gradations exist between rocks having a very marked relict igneous texture and those lacking it. Where plagioclase has recrystallized, it has a mosaic texture and a grain size of 0.01 to 0.2 mm.

Flakes of nearly isotropic FeMg and MgFe chlorite are 0.01 to 0.3 mm long. Sericite is of similar size.


FIGURE 54.—Photomicrograph of mafic porphyry from outcrop along Blue Ridge Parkway in area G–3, plate 1 about 0.6 mile northwest of Cone Lake (area H–3, pl. 1). Laths of plagioclase (altered to albite) in matrix of sericite, albite, chlorite, magnetite, and sphene. Analyzed specimen 7, table 17.

FIGURE 53.—Amygdaloidal greenstone. Epidote-filled amygdules in mafic flow intercalated in felsic volcanic rocks of the Grandfather Mountain Formation on east side of Wilson Creek about 0.1 mile north of bench mark 1410 (area F–6, pl. 1). Note epidote segregation in upper left part of photograph.

Sphene is found in tiny grains 0.001 to 0.1 mm in diameter, which commonly form aggregates. Ilmenite and magnetite grains are 0.002 to 0.3 mm in diameter. Brown biotite flakes 0.05 to 0.2 mm long are intergrown with chlorite. Epidote, in part pistacite, is most abundant in amygdaloidal zones, where it is found in the groundmass as well as in amygdules. Quartz occurs with epidote in amygdules and amygdaloidal zones but is most abundant in tuffaceous sedimentary rocks, where it occurs as clastic grains as much as 2 mm in diameter and in mosaics of recrystallized grains 0.05 to 0.2 mm in diameter.

Some phyllites mapped with the mafic volcanic rocks lack definite volcanic texture and contain little albite or chlorite. They are composed of chlorite, sphene, ilmenite, sericite, and opaque minerals. They were probably fine-grained tuffs or tuffaceous sedimentary rocks.

In the belt of mafic volcanic rocks that extends across Ripshin Ridge from Buck Creek to Upper Creek (areas E–6 and E–5, fig. 4), phyllonites derived from the adjacent Wilson Creek Gneiss and phyllites intercalated with sedimentary rocks and mafic volcanic rocks are intimately associated. Locally, the arkose overlying the volcanic rocks contains phyllite fragments. In some outcrops in this area it is impossible to distinguish phyllites from phyllonites, but commonly the phyllonites can be identified because they grade into gneiss and contain lenses of less-sheared gneiss and pegmatite. Both the phyllites and phyllonites are rich in sericite and contain various quantities of quartz, magnetite, sphene, chlorite, epidote, and plagioclase.

The mafic volcanic rocks in the lower part of the Grandfather Mountain Formation differ from the Montezuma Member in their porphyritic character, partly tuffaceous nature, lack of amphibole, and greater content of albite.

Chemical analyses (table 17, Nos. 5–7) show that these rocks have $SiO_2$ contents typical of basalts, but their rather high $Na_2O$ and low CaO contents suggest that they have been albitized. Concentration of CaO and some $SiO_2$ in segregations and amygdules may have modified the chemistry of the rest of the rock. Their $Na_2O$ contents are greater than those of the Montezuma Member, which indicates that at some stage in their development they were more intensely albitized. MgO contents are lower than those of the Montezuma Member, which probably accounts for the lack of amphibole, and the total iron oxide contents are also slightly lower. $TiO_2$ contents are similar to those of the Montezuma.

### MONTEZUMA MEMBER

The Montezuma Member of the Grandfather Mountain Formation is composed of metabasalt and is found at a higher stratigraphic horizon than the other volcanic rocks of the Grandfather Mountain Formation. Keith (1903) named the unit the Montezuma Schist from exposures near the village of Montezuma in the Linville quadrangle (south edge of area D–4, pl. 1). He believed that it was older than the associated sedimentary rocks. Bryant (1962) found that the metabasalt is interbedded with the surrounding arkoses and renamed it the Montezuma Member of the Grandfather Mountain Formation. It crops out in a continuous belt 15 miles long, mainly in the Linville quadrangle, and is well exposed along the Watauga River and on most hillsides along the outcrop belt. It is cut off by the Linville Falls fault on the northeast and pinches out to the south. Metabasalt also crops out in small areas west of the main outcrop belt; whether these areas are part of the main body brought up in anticlines or separate flows at a stratigraphically higher horizon is generally uncertain.

The Montezuma Member is commonly composed of greenstone and locally, of greenschist. Much of the rock is intermediate between greenstone and greenschist and has a moderately strong cleavage and only an indistinct igneous texture, except for prominent amygdaloidal zones. For convenience, it will be called greenstone. The greenstones are green, blue-green, or blue-gray fine-grained rocks containing epidote and quartz-epidote knots, lenses, and veinlets. Amygdules containing epidote, albite, and quartz occur in zones of massive dark-blue-gray rock. Locally, chlorite and calcite occur in amygdules. Many outcrops contain enough magnetite to deflect a compass needle. Neither pillow structure nor columnar jointing was identified.

Amygdules average 2 to 5 mm in diameter and reach a maximum length of 1.5 cm. They generally stand out on weathered surfaces. Epidote is the most abundant filling, followed, in order of abundance, by albite, quartz, chlorite, and calcite. Chlorite-filled amygdules are commonly flattened and elongated. Some amygdules have epidote rims and quartz cores (fig. 55); others have three zones: albite on the rim, epidote, and quartz in the core. Still others have chlorite cores and epidote rims. Epidote and albite predominate in different amygdules in the same rock.

Yellowish-green to apple-green epidote and quartz-epidote segregations range from irregular stringers and veinlets to lenses as much as 3 feet long. Locally, they contain calcite and fibrous actinolite. The segregations make up about 10 percent of

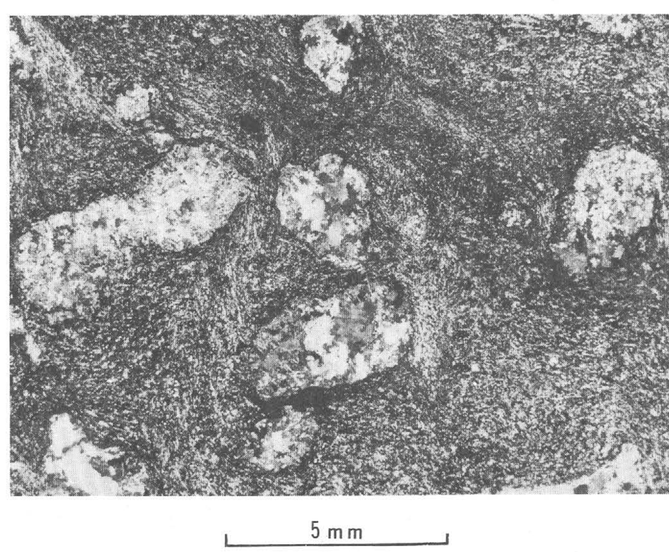



FIGURE 55.—Photomicrograph of amygdaloid in Montezuma Member of Grandfather Mountain Formation exposed in roadcut along Watauga River about 0.1 mile west of east edge of the Linville quadrangle (area F–2, pl. 1). Amygdules have epidote rims and quartz cores. Matrix consists of plagioclase (partly in tiny lath-shaped grains altered to albite), actinolite, sphene, opaque mineral, and chlorite.

the rock in most outcrops; in some they constitute as much as 20 percent. A few greenstone outcrops lack segregations. Quartz or quartz and chlorite locally occur in veinlets within epidote segregations. Rare veinlets of asbestiform actinolite as much as 1 inch thick are found.

Boundaries between flows are commonly marked by a few inches to a few feet of fragmental phyllite. The phyllite is rich in sphene and opaque minerals and may be a tuff or tuffaceous sedimentary rock. It consists of flattened fragments of light-green sericite phyllite 5 mm to 3 cm long in a matrix of dark-gray sericite phyllite. In a few places, similar rocks occur as thin beds in arkose within a few hundred feet stratigraphically of the greenstone. Arkose or siltstone overlying greenstone have been locally epidotized.

In places, tops of flows contain fragments of dark-green-gray rock 1 to 6 inches long. The rock in the fragments is rich in opaque minerals, sphene, and chlorite. It lacks albite and locally contains amygdules. The fragments are in a matrix of lighter gray rock that contains albite and lesser amounts of epidote, actionolite, chlorite, and opaque minerals.

Albite in the greenstone generally forms small anhedral grains 0.01 to 0.05 mm in diameter, but in some rocks it occurs as relict subhedral laths, generally less than 0.5 mm long but locally as much as 1.5 mm long; relicts of more calcic plagioclase were not found.

Light-green actinolite is generally well alined and is locally somewhat bent. It ranges from 0.02 to 1.5 mm long and averages about 0.1 mm long. In places, the larger crystals are randomly oriented postkinematic porphyroblasts. Hornblende rimmed by actinolite occurs in one rock. Green FeMg and MgFe chlorite forms flakes 0.01 to 0.06 mm long.

Epidote is pistacite which occurs in anhedral to subhedral grains 0.008 to 0.4 mm in diameter. In a few rocks, epidote grains are red in hand specimens and have a light pink color in thin section, perhaps because of admixtures of piedmontite.

Opaque minerals are magetite, maghemite, ilmenite, and hematite 0.005 to 0.3 mm in diameter. Long aggregates of opaque minerals are generally nonmagnetic and probably ilmenite. Sphene and leucoxene occur as anhedral grains and aggregates and as rims on opaque minerals. Anhedral relicts of pigeonite were found in only one sample.

Although the proprotions of albite and mafic minerals in single thin sections of the Montezuma Member vary widely, the unit is rather homogeneous overall. Local variations are probably due to the widespread segregations and amygdule fillings. Chemical analyses (table 17, Nos. 2, 3, 4) of greenstones of the Montezuma Member show that they are of basaltic composition. Sample 2, which has low $Na_2O$ and high CaO contents, was free of epidote segregations and may be the most representative analysis of the metabasalt. The other analyses suggest slightly albitized basalt.

The Montezuma Member does not easily fit into any of the categories of basalt. It has the high iron and titanium contents characteristic of tholeiite, but its alumina and (except for analysis 2) soda contents are higher than those of tholeiite. Its alumina content lies between that of tholeiite and high-alumina basalts. Compared with rocks of quite similar aspect from the Catoctin Formation of central Virginia, which may have been derived from tholeiitic basalt (Reed, 1964c), the Montezuma Member contains more alumina and less soda but is quite similar in other respects.

### ORIGIN AND ENVIRONMENT OF DEPOSITION

Abrupt changes in thickness and lithology along strike indicate that the Grandfather Mountain Formation was deposited in a rapidly subsiding basin. Differences between rocks exposed in the southeast part of the window near Brown Mountain and those in the main outcrop belt to the northwest suggest that similar stratigraphic variations occur across strike. Bedding and fold attitudes, many of which are not shown on the maps, but which are summarized below, suggest that the exposed thickness of the formation is about 7,000 feet along the northern edge of the window in the Blowing Rock quadrangle, at least 20,000 feet in the center of the Linville quadrangle, and 9,000 feet in the southern part of the Linville quadrangle (fig. 56). There is no information on thickness and facies variations in a northwest-southeast direction in the main outcrop area.

The conglomerate-filled channels, crossbedding, and ripple marks in the arkose units and the poor sorting of the arenaceous rocks indicate that they were deposited in alluvial fans or deltas adjacent to a region of high relief. Whether the water in the basin was marine or fresh is not known. The mineralogy of the arkoses indicates that they could have been derived in a large part from the underlying basement rocks. Volcanic material in the conglomerates might have come from part of the Grandfather Mountain Formation itself.

The laminated siltstone was probably deposited in shallow water farther from the source of detritus.

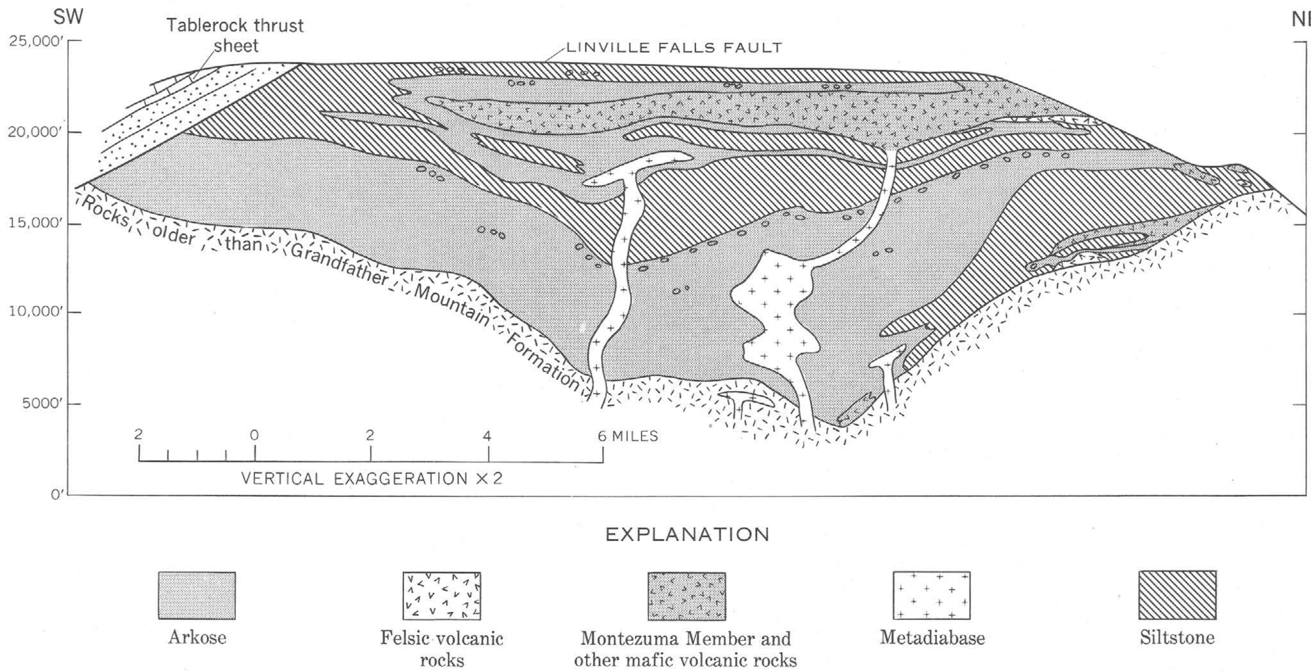



**FIGURE 56.—Diagrammatic stratigraphic section of the Grandfather Mountain Formation constructed on the assumption that rock units cropping out can be projected to a vertical plane parallel with the trend of the units. Section extends from the north-central part of the Linville Falls quadrangle to north-central part of the Blowing Rock quadrangle. Circles indicate occurrences of conglomerate.**

Remoteness from source of clastic material allowed local deposition of admixed carbonate. Compositional differences between siltstones and arkoses may be due to mechanical sorting; the coarse-grained quartz and potassic feldspar remained on the alluvial fans or deltas, while clays or micas and the finer grained quartz and plagioclase were washed farther into the basin. Potassic feldspar in the basement rocks is in large grains, whereas much of the plagioclase and some of the quartz is finer grained.

Occasionally, coarser grained unsorted material was transported beyond the delta and was deposited as graywacke and conglomeratic graywacke. The uppermost siltstone unit, which contains more graywacke-type rocks than the other silstone units, may have been deposited in deeper water.

The positions of the shorelines and deltas and thus the sites of deposition of the two principal lithologies shifted from time to time. Except for extrusion of the basalts of the Montezuma Member, volcanism occurred principally in the area east of the main outcrop belt of the Grandfather Mountain Formation and at the beginning of the filling of the part of the basin presently exposed.

## AGE AND CORRELATION

Stratigraphic relations between the Grandfather Mountain Formation and rocks of Cambrian(?) age cannot be determined in the Grandfather Mountain window because the two sequences are in thrust contact. The Grandfather Mountain Formation, however, differs from the Chilhowee Group in the lenticular stratigraphy, greater and more variable total thickness, greater content of volcanic rock, lack of distinctive marker horizons, absence of orthoquartzite beds, abundance of feldspar, and presence of considerable clastic plagioclase. The differences in mineralogy produce the marked contrast in the chemistry illustrated by figures 46 and 47. Heavy-mineral suites from sandstones of the Grandfather Mountain Formation contain abundant zircon and very little tourmaline, whereas those from Chilhowee sandstones in the Tablerock thrust sheet and in the Unaka belt contain subequal amounts of tourmaline and zircon. Thus, it is unlikely that the Grandfather Mountain Formation is correlative with the Chilhowee Group. Nor does the Grandfather Mountain Formation resemble any known post-Chilhowee rocks in the southern or central Appalachians, and nowhere in the region are post-Chilhowee rocks in stratigraphic contact with the Precambrian metamorphic and plutonic rocks. The Grandfather Mountain Formation is therefore probably older than the Chilhowee Group and younger than the rocks (1,000–1,100 m.y. old) on which it rests; it is believed to be of late Precambrian age.

The Grandfather Mountain Formation contains the southernmost volcanic rocks of late Precambrian age known in the Blue Ridge. The Mount Rogers Formation, 25 miles north of the Grandfather Mountain window (pl. 3) contains the nearest similar rocks. They rest on basement rocks and at least part of them are in the Shady Valley thrust sheet. Available descriptions of the Mount Rogers Formation (Jonas and Stose, 1939; Stose and Stose, 1944, 1949, 1957; King and Ferguson, 1960) indicate that it is composed predominantly of felsic volcanic rock, shale, conglomerate, and arkose. Mafic volcanic rock is a minor constituent. Some of the sedimentary rocks of the Mount Rogers Formation bear a remarkable resemblance to those in the Grandfather Mountain Formation. Green arkoses, conglomerates, and some gray finely laminated slaty rocks in the Mount Rogers Formation resemble some rocks of the Grandfather Mountain Formation (D. W. Rankin, written commun., 1965). The Grandfather Mountain Formation, however, lacks the agglomerates, extremely coarse conglomerates, and red slates found in the Mount Rogers Formation. The site of deposition of the part of the Mount Rogers Formation in the Shady Valley thrust sheet must have been much closer to the site of deposition of the Grandfather Mountain Formation before the northwestward movement of the thrust sheet. In view of the abrupt stratigraphic variations in the Grandfather Mountain Formation, it would be surprising if there were any close correlation between its stratigraphy and that of the Mount Rogers Formation.

Rocks of late Precambrian age are found in the Blue Ridge thrust sheet near Mount Mitchell, 30 miles southwest of the Grandfather Mountain Formation (fig. 12). They have been neither mapped nor studied, but they resemble rocks of the Great Smoky Group in the Blue Ridge thrust sheet 80 miles to the west-southwest (J. B. Hadley, oral commun., 1959). The sedimentary rocks of the Grandfather Mountain Formation more nearly resemble rocks of the Snowbird Group of the Ocoee Series, which are characterized by current bedding, than they do those of the Great Smoky Group, which are characterized by graded bedding (King and others, 1958; J. B. Hadley, oral commun., 1959).

Rocks of the Snowbird Group (Hadley and Goldsmith, 1963, table 7; Hamilton, 1961, table 2), however, differ somewhat from those of the Grandfather Mountain Formation in mineralogy and chemical composition. The sandstones in the Snowbird Group are richer in clastic plagioclase and total feldspar and poorer in mica than the Grandfather Mountain arkoses and therefore have lower $K_2O$:$Na_2O$ ratios. The Snowbird siltstones are very rich in mica and consequently in $K_2O$ and $Al_2O_3$; they chemically resemble shale. Thus, the proportions of CaO, $K_2O$, and $Na_2O$ are reversed between the sandstones and siltstones of the Grandfather Mountain Formation and those of the Snowbird Group.

Relationship between the rocks of the Grandfather Mountain Formation and those of the Ocoee Series cannot be definitely established. Whether the Grandfather Mountain Formation and perhaps the Mount Rogers Formation are a different facies of the Snowbird Group deposited in a different part of the Ocoee basin or whether they were deposited in separate basins and perhaps at somewhat different times is unknown.

Hadley (1970) summarizes the characteristics of the Snowbird Group, Mount Rogers Formation, and Grandfather Mountain Formation and reaches similar conclusions.

Recent isotopic dates on zircon from felsic volcanic rock of the Grandfather Mountain Formation, the Mount Rogers Formation and the Catoctin Formation in southern Pennsylvania indicate an original age of 850 m.y. and prove the approximate synchroneity and Precambrian age of these units (Rankin and others, 1969). These authors note that published zircon ages of the Beech Granite and aegirine-augite granite in the Blue Ridge thrust sheet fall close to the discordia curve for these rhyolites, which suggests that they are correlative.

#### LINVILLE METADIABASE

The Linville Metadiabase intrudes basement rocks and rocks of the Grandfather Mountain Formation stratigraphically below the Montezuma Member. Keith (1903) named the Linville Metadiabase and recognized its relation to the Montezuma Member of the Grandfather Mountain Formation, although he thought that part of it constituted the lower part of a flow. The map relations shown by him were complicated by his belief that the arkoses unconformably overlay the mafic igneous rocks.

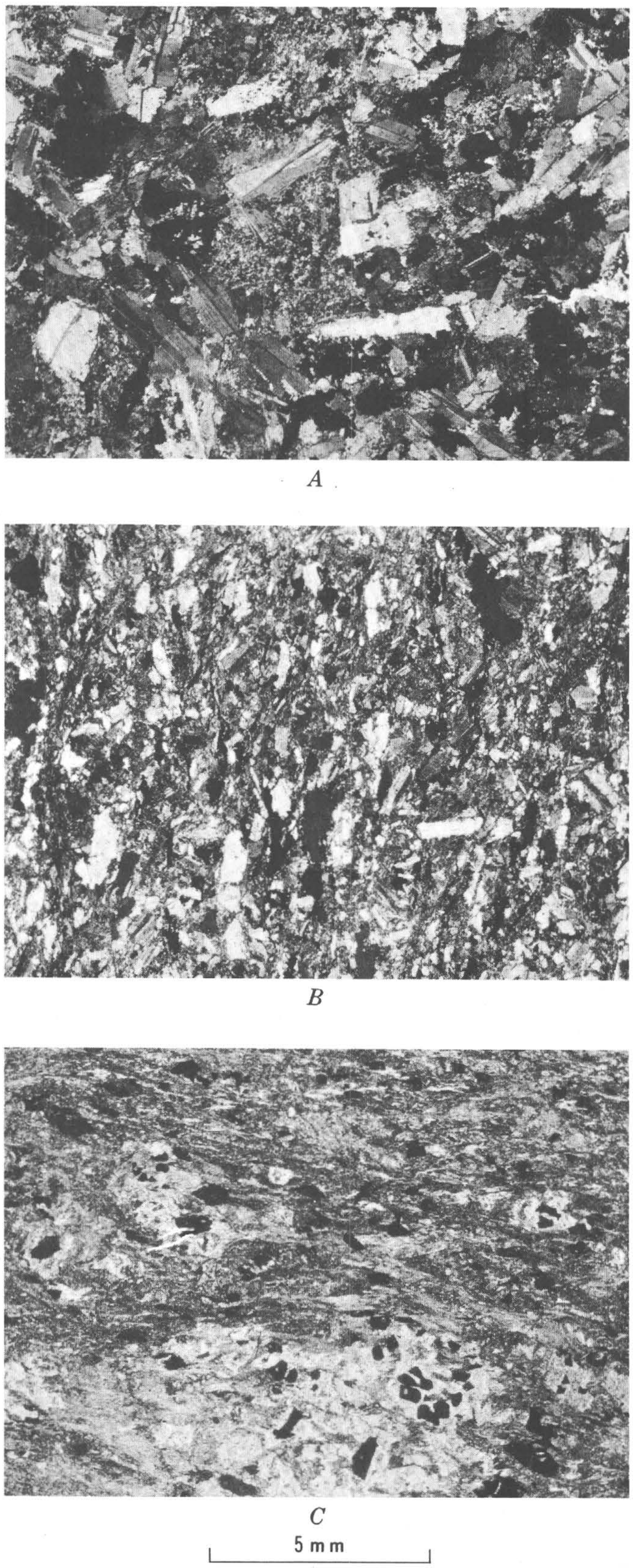



FIGURE 57.—Photomicrographs of Linville Metadiabase. *A*, Altered and slightly sheared coarse-grained diabase from north side of Boone Fork about 0.25 mile east of west edge of Blowing Rock quadrangle (area G–3, pl. 1). Plagioclase laths (altered to albite) in a matrix of recrystallized albite, actinolite, sphene, chlorite, quartz, epidote, and magnetite. *B*, Foliated medium-grained metadiabase from roadcut along North Carolina Highway 183 on south side of Camp Creek 1.3 miles N. 75° W. of summit of Cranberry Knob (area D–4, pl. 1). Laths of plagioclase (altered to albite) in a matrix of chlorite, epidote, actinolite, sphene, stilpnomelane, and opaque minerals. *C*, Schistose metadiabase from roadcut along North Carolina Highway 105 at Linville Gap (area E–4, pl. 1). Broken amphibole porphyroclast (now actinolite) in a fine-grained matrix of chlorite, epidote, actinolite, albite, sphene, sericite, and ilmenite. Analyzed specimen 1, table 17.

Most of the mappable intrusives in the basement rocks are within a mile of the contact with the Grandfather Mountain Formation. Thin dikes of metadiabase or metadiorite are somewhat more widespread in the basement rock than in the Grandfather Mountain Formation. Intrusive bodies in the Grandfather Mountain Formation range from concordant to crosscutting and are as much as 4 miles long. Concordant bodies are generally found in the siltstone units or in thin phyllite beds in the arkose units. Arkose adjacent to metadiabase has been epidotized in a few places. At the margins of the largest metadiabase intrusive are small bodies of porphyroclastic granite to granodiorite; only one was large enough to map.

The Linville Metadiabase is generally poorly exposed. It forms a reddish-brown clayey soil, and small bodies of metadiabase in the arkose and basement rocks in the steep country southeast of the crest of the Blue Ridge were invariably cleared for farms because the soil is so much less stony and sandy than on the adjacent rock. Natural outcrops are abundant in some of the large bodies on Grandfather Mountain, especially the one at the head of Anthony Creek. Metadiabase in the larger bodies typically weathers into rounded boulders.

The metadiabase is dark green, green gray, or blue gray. It contains abundant amphibole megacrysts that range from 0.5 to 10 mm and average about 5 mm in diameter and less abundant laths of plagioclase 2 to 15 mm long, averaging 2 to 3 mm. Segregation knots and lenses of epidote and quartz are much less widespread than in the Montezuma Member. They are found in about 5 percent of the metadiabase outcrops as compared with being found in more than 50 percent of metabasalt outcrops. Rarely, the rock contains veinlets and fractures filled with quartz or albite. Metadiabase ranges from completely altered but unsheared rock with ophitic or diabasic texture to blastomylonitic greenschist (fig. 57). Much of the metadiabase is a blastomylonitic gneiss is which the large crystals are porphyroclasts.

Amphibole is generally light-green actinolite. It occurs as porphyroclasts (fig. 57*C*) and as needles alined on the foliation planes. Relicts of dark-green to reddish-brown hornblende are found in the cores of some of the porphyroclasts. Both the dark core and the lighter colored rims have the same extinction angle. Relicts of pinkish-tan pyroxene are less common. Tiny sphene inclusions are abundant in the hornblende porphyroclasts.

Plagioclase has been completely altered to albite. In less sheared rocks it retains its lath shape (fig. 57*A* and *B*) but is generally filled with inclusions of epidote and sphene; in more sheared rocks it has been recrystallized to clear grains 0.05 to 0.2 mm in diameter.

Green FeMg chlorite is 0.05 to 0.2 mm long and locally is intergrown with biotite or amphibole. Elsewhere it is interstitial to the other minerals. Epidote generally occurs in aggregates, grain size ranging from 0.01 to 0.8 mm. In some specimens, the epidote is pistacite. Sphene mantles opaque minerals and occurs in stringlike aggregates as much as 0.7 mm long. Opaque minerals are magnetite, maghemite, and ilmenite and are as much as 0.3 mm in diameter. Brown stilpnomelane is found both as alined flakes and in rosettes. Individual grains reach a length of 0.5 mm. Brown to dark-green biotite is intergrown with chlorite and locally replaces the amphibole porphyroclasts.

Granite associated with the metadiabase near Dave Coffee Branch (area G–4, pl. 1) is gray and contains perthite porphyroclasts as much as 1 cm in diameter, mosaic-textured quartz, greenish-brown biotite, saussuritized plagioclase, and granophyric intergrowths. Rock in some small bodies is a granodiorite that contains plagioclase porphyroclasts.

Chemical analysis of the Linville Metadiabase shows that its composition is similar to metabasalts of the Montezuma Member of the Grandfather Mountain Formation. It contains less silica and more iron and titanium than mafic intrusive rocks in the Ocoee Series in the Great Smoky Mountains (Hadley and Goldsmith, 1963, table 12).

## ROCKS OF THE TABLEROCK THRUST SHEET

Rocks of the Chilhowee Group of Cambrian and Cambrian (?) age and the Shady Dolomite of early Cambrian age compose the Tablerock thrust sheet, a thin but extensive thrust sheet that lies between autochthonous rocks of the Grandfather Mountain window and the overriding Blue Ridge thrust sheet. The Tablerock thrust sheet occupies much of the southwestern part of the Grandfather Mountain window (pl. 1), a prominent klippe caps Tablerock Mountain (area D–7 pl. 1) for which the thrust sheet and the fault at its base are named.

The rocks in what is now recognized as the Tablerock thrust sheet were described by Maclure (1817, p. 43), who said, after describing rocks of low metamorphic grade in the James River, Virginia area:

> a similar formation about 15 miles long and 2 to 3 miles wide occurs on the north fork of the Catawba River running along Linnville [sic] and John's Mountain near to the Blue Ridge * * *

Kerr (1875, p. 135), after mentioning quartzites on Linville Mountain, said:

> The dip is very irregular and confused but seems to be predominantly westward. Several beds of compact light-colored and gray limestone crop out along the western base of the mountain in the valley of the North Fork, and almost to the head of it, the upper beds being on the west side.

### CHILHOWEE GROUP

Quartzite, feldspathic quartzite, and phyllite, which make up most of the Tablerock thrust sheet, are part of the Chilhowee Group of Cambrian and Cambrian(?) age. The correlation is based upon litholigc similarities and similarity of stratigraphic sequence with rocks of the Chilhowee Group of northeastern Tennessee (King and Ferguson, 1960). The rocks of the Chilhowee Group in the Grandfather Mountain window are unfossiliferous, except for a few occurrences of the worm tube, *Scolithus.* Fossils are very scarce in the Chilhowee Group elsewhere (Keith, 1903; King, 1949; King and Ferguson, 1960).

Complexity of folding and discontinuity of exposures preclude an accurate estimate of the stratigraphic thickness of the Chilhowee Group in the Tablerock thrust sheet, but estimates based on structural sections indicate that at least 4,000 feet of Chilhowee beds is present.

Rocks of the Chilhowee Group have been subdivided into two quartzite units separated by a persistent blue phyllite unit, which constitutes the only mappable marker horizon in the sequence. Because of the isolated tectonic position of the rocks in the Tablerock thrust sheet, correlations with specific formations in the Chilhowee Group of the Unaka belt have not been attempted.

### LOWER QUARTZITE UNIT

The lower unit of the Chilhowee Group in the Tablerock thrust sheet is a sequence of quartzite and feldspathic quartzite containing interbedded green sericite phyllite. The thickness of the unit in the Linville Falls quadrangle ranges from about 800 feet along the Linville River south of Shortoff Mountain (area D–9, pl. 1) to at least 2,200 feet on the east side of Linville Mountain opposite the Chimneys (area D–6). The cliffs on the west side of the Linville Gorge and on Shortoff Mountain, Dobson Knob, the Chimneys, and Tablerock Mountain are composed of the lower few hundred feet of this unit (fig. 3). The higher part of the sequence generally forms rather rounded slopes without prominent cliffs.

The unit consists predominantly of medium- and fine-grained white, gray, or light green quartzite and feldspathic quartzite but contains numerous thin interbeds of green sericite phyllite. The quartzite is generally thin bedded, but massive beds as much as 30 feet thick of medium- and coarse-grained quartzite are common. Much of the quartzite is crossbedded and has dark-blue heavy-mineral streaks parallel to bedding and crossbedding. Angular clasts of pink feldspar as much as 5 mm in diameter are widespread in some beds. Beds of quartz-pebble conglomerate 6 inches to 5 feet thick occur in a few places near the base of the sequence. A few 20-foot beds of vitreous white or gray quartzite occur near the top of the unit.

The quartzite (fig. 58) consists of recrystallized quartz grains 0.05 to 0.2 mm in diameter in a mosaic that encloses detrital grains of strained quartz and microcline and, very rarely, microperthite. The finer grained beds are entirely mosaic textured and lack detrital grains, except for heavy minerals. Microcline content ranges from 0 to 25 percent, and plagioclase is rare. Scattered flakes and discontinuous folia of fine-grained iron-rich muscovite with weak green absorption define a cleavage parallel with bedding. Accessory minerals, chiefly tourmaline, zircon, sphene, ilmenite, and magnetite, occur as scattered grains or are concentrated in thin laminae parallel with bedding or crossbedding. Opaque minerals generally predominate in the heavy-mineral seams. The phyllites consist mainly of fine-grained iron-rich muscovite, but they contain recrystallized quartz and small amounts of FeMg chlorite and brown biotite.

Most of the rocks of the unit are quartzites or muddy quartzites (fig. 59*A*) and have a high content of $SiO_2$ and rather low alkali and $Al_2O_3$ contents (table 18). No analyses of the green phyllite are available, but it probably has a higher $Fe_2O_3$:FeO ratio than the blue phyllite described below.

### PHYLLITE UNIT

The lower quartzite unit is overlain by a thin unit composed of dark phyllite containing interbedded fine-grained gray or white quartzite. The phyllite unit ranges in thickness from a few feet to as much as 400 feet but is generally less than 150 feet thick.

The phyllite is a lustrous, finely laminated, dark-blue blue-gray, or blue rock consisting of folia of fine-grained sericite and thin lenses and laminae of granoblastic quartz, parallel to a strong bedding foliation (fig. 58*D*). The foliation is commonly cut by slip cleavage which produces minor crenulations on the foliation surfaces. The rock contains minor amounts of FeMg chlorite, biotite, magnetite, and ilmenite and scattered grains of zircon and green tourmaline.

Interbeds 0.5 to 6 inches thick of fine-grained light-gray or blue-gray sugary quartzite are common in the phyllite; local layers or blue or white vitreous quartzite are 2 to 20 feet thick, especially where the phyllite is unusually thick.

Thin beds of similar phyllite are interlayered with quartzites in the upper part of the lower quartzite unit and throughout the overlying quartzite unit.

The phyllites (table 18, analysis 5 and 6) have high contents of $K_2O$, $Al_2O_3$, and iron oxides and low contents of $Na_2O$ and CaO; thus they are probably derived from typical clay shales.

### UPPER QUARTZITE UNIT

The upper unit of the Chilhowee Group is a sequence of thin- to thick-bedded medium- to fine-grained white, greenish-gray, or bluish-gray quartzite and felspathic quartzite, ranging in thickness from 1,300 to perhaps 2,500 feet. The unit underlies most of the dip slope on the west side of Linville Mountain. Massive beds of fine-grained vitreous quartzite are more common than in the lower unit, and phyllite is less common. Phyllites of the upper unit resemble the blue phyllite of the middle unit rather than the green phyllites of the lower quartzite unit.

Small-scale crossbedding is common in the quartzites. Conglomerates are absent. Near the summit of Bald Knob and in a few places on the slopes of Linville Mountain, some quartzite beds contain slightly deformed *Scolithus* tubes similar to those common in rocks of the Chilhowee Group elsewhere (King, 1949; King and Ferguson, 1960).

The contact of the upper quartzite unit with the overlying Shady Dolomite is exposed along the North

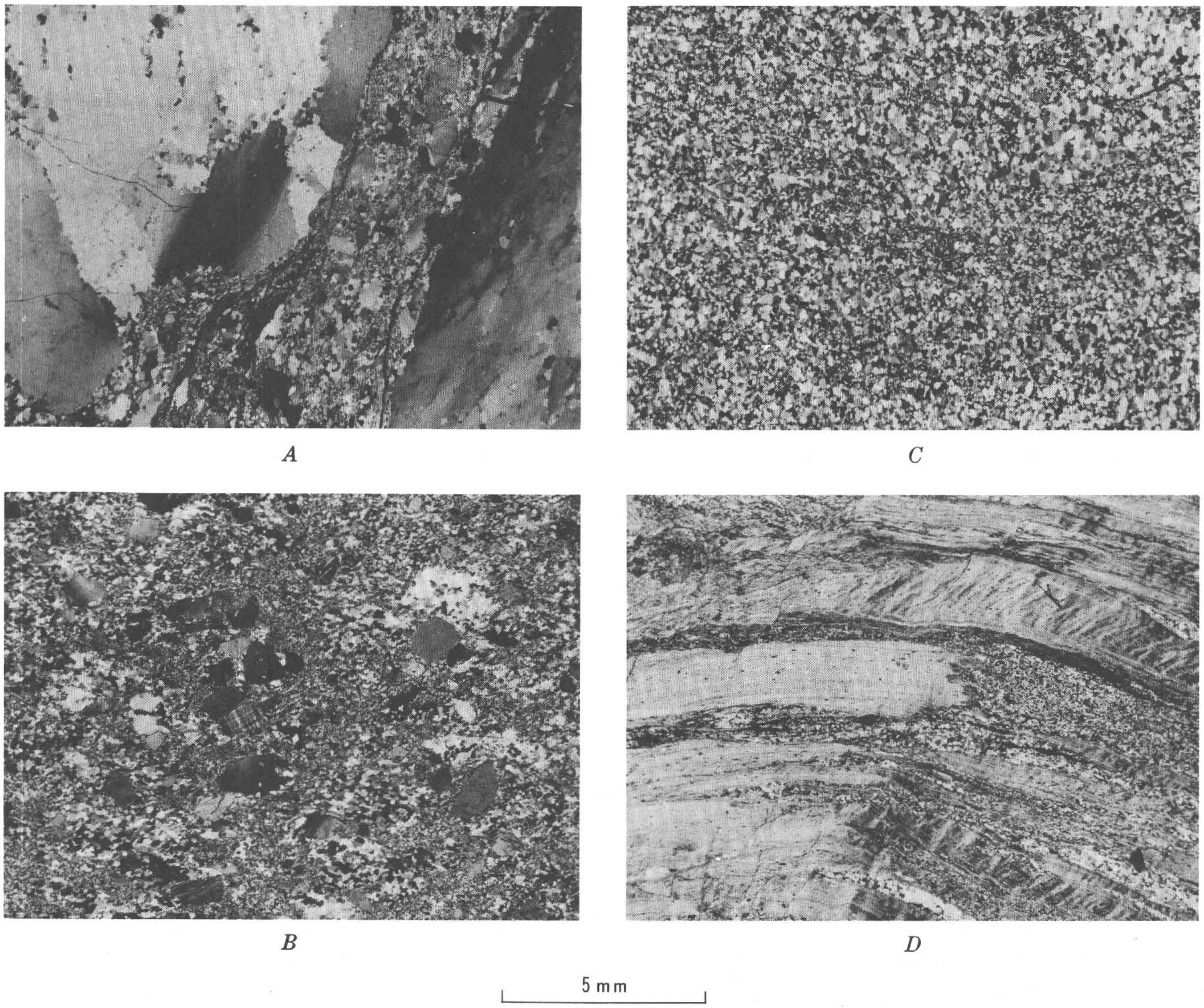



FIGURE 58.—Photomicrographs of rocks of the Chilhowee Group. *A*, Quartz-pebble conglomerate from lower quartzite unit near base of Tablerock thrust sheet east of road intersection at altitude of 2,880 feet on Linville Mountain (area D–8, pl. 1). Quartz pebbles, partly broken; mortar along fractures, recrystallized. Quartz and microcline clasts in matrix of recrystallized quartz and sericite. *B*, Poorly sorted feldspathic quartzite from lower quartzite unit at bottom of Linville Gorge in area D–7, 1.4 miles S. 30° E. of Linville Falls (area D–6, pl. 1). Clastic grains of quartz and microcline in a groundmass of quartz, iron-rich muscovite, and subordinate microcline. Analyzed specimen 2, table 18. *C*, Fine-grained well-sorted quartzite from upper quartzite unit at altitude of 1,760 feet in Stillhouse Branch (area C–8, pl. 1). Mosaic of recrystallized quartz and minor amount of sericite. *D*, Crenulated phyllite from phyllite unit exposed in roadcut on North Carolina Highway 183 about 0.8 mile northeast of bridge across Linville River (area D–6, pl. 1). Sericite and chlorite and fine-grained mosaic-textured lenses and laminae of quartz, crinkled and cut by slip cleavage. Analyzed specimen 6, table 18.

Fork of the Catawba River about 1 mile north of Linville Caverns (area C–7, pl. 1). Massive white quartzite passes up into 15 to 20 feet of thin-bedded quartzite and green sericite phyllite which is overlain by a 6-inch to 1-foot bed containing 2- to 5-mm well-rounded granules of quartz in a calcareous matrix. This bed is directly overlain by the Shady Dolomite (fig. 60). These upper beds probably correspond to the Helenmode Member of the Erwin Formation (John Rodgers, oral commun, 1959).

The quartzites and feldspathic quartzites typically consist of mosaic-textured quartz (fig. 58*C*) and locally occurring microcline 0.05 to 0.1 mm in diameter, scattered flakes and partings of sericite or green

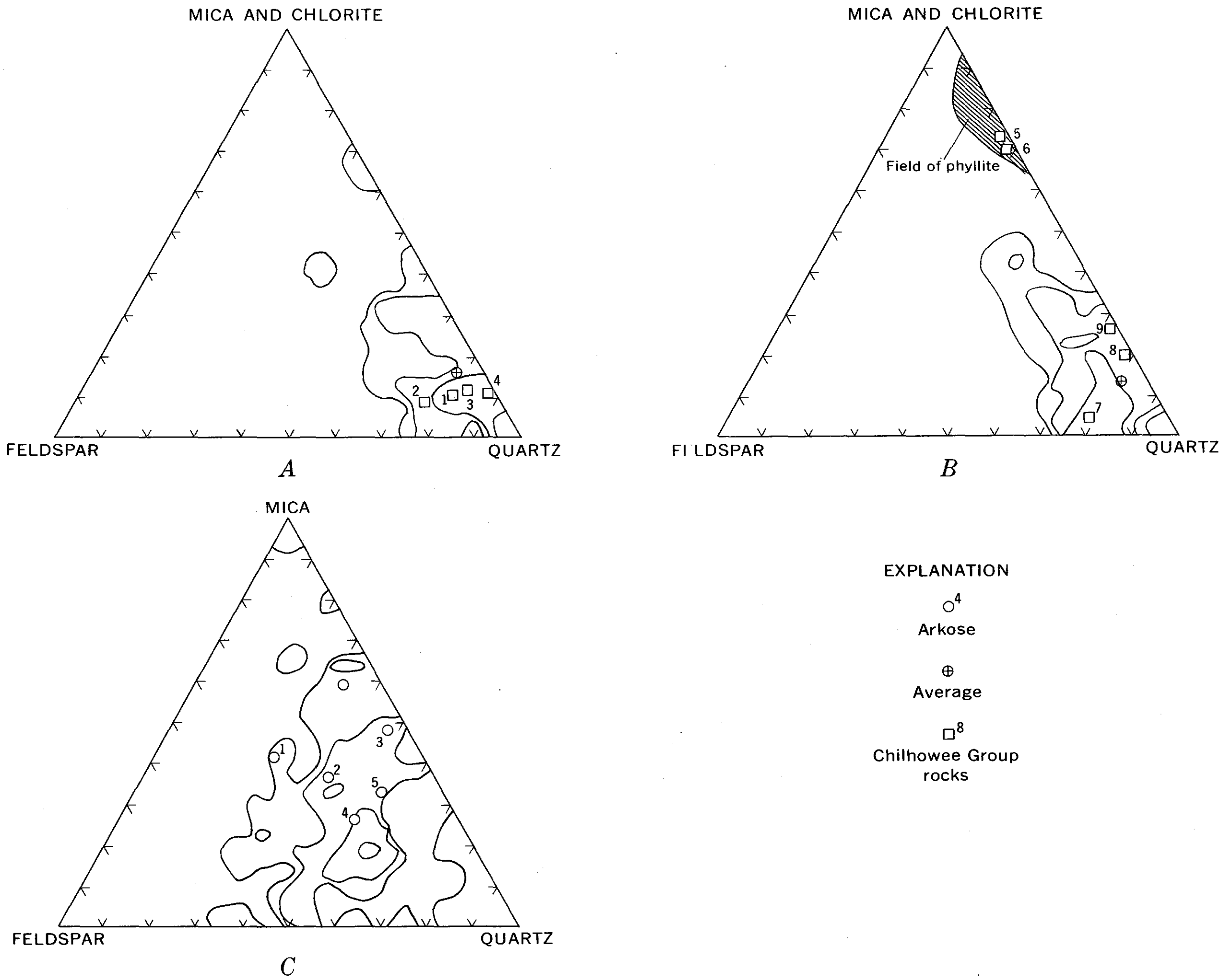



FIGURE 59.—Proportions of quartz, feldspar, and combined mica and chlorite in rocks of the Chilhowee Group in the Tablerock thrust sheet and the lowest arkose of the Granfather Mountain Formation. *A*, Lower quartzite unit of the Chilhowee Group. Numbers of analyzed specimens refer to analyses in table 18. Based on point counts of analyzed specimens and counts of 50 random grains in each of 44 other thin sections. Contours 2, 4, 12, and 25 percent. *B*, Upper quartzite and phyllite units of the Chilhowee Group. Number indicates analyzed specimen in table 18. Based on point counts of analyzed specimens and counts of 50 random grains in each of 40 other thin sections. Contours 2, 5, 14 and 28 percent. *C*, Lowest arkose unit of the Grandfather Mountain Formation. Number indicates analyzed specimens in table 15. Based on point counts of 24 thin sections and counts of 50 random grains in each of 55 other thin sections. Contours 1, 3, 6, and 13 percent.

iron-rich muscovite mica, and some locally occurring brown biotite. Widespread detrital grains of microcline and quartz and scattered detrital grains of perthite are 0.5 to 2 mm in diameter. A few beds contain feldspar clasts as large as 1 cm and quartz granules as large as 3 mm in diameter. Feldspar content is 0 to 25 percent; plagioclase is almost totally absent. Magnetite, ilmenite, sphene, zircon, and tourmaline are the chief accessory detrital minerals and are commonly concentrated in streaks parallel to bedding and crossbedding. Clastic rutile and metamorphic chlorite and stilpnomelane are present locally.

The overall composition of the upper quartzite (fig. 59*B*) is quite similar to that of the lower quartzite (fig. 59*A*); both are somewhat muddy and feldspathic orthoquartzites.

The main difference between the upper and lower quartzite units is that the upper unit is somewhat better sorted, lacks pebble beds, and has gray rather than green phyllites. Individual outcrops of the units are generally similar, and if the phyllite between

TABLE 18.—*Chemical analyses, modes, and norms of rocks of the Chilhowee Group in the Tablerock thrust sheet*

[Samples analyzed by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey, 1962. Modes determined by point counts of 600 grains; P, present but not intersected in counting. Major oxides and CIPW norms given in weight percent; modes, in volume percent. CIPW norms for samples 1, 5, 6, and 7 were not computable]

| | Lower quartzite unit | | | | Phyllite unit | | | Upper quartzite unit | |
|---|---|---|---|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 |
| Field No. | RE–20–1b | 2–249 | RE–90–1 | RE–93–3a | 1–93 | 2–197 | RE–91–1 | 2–195 | 2–196A |
| Laboratory No. | 160173 | 160176 | 160174 | 160177 | 160172 | 160175 | 160171 | 160170 | 160169 |
| **Major oxides** | | | | | | | | | |
| $SiO_2$ | 86.7 | 90.1 | 90.7 | 94.0 | 61.9 | 66.4 | 87.3 | 90.3 | 94.4 |
| $Al_2O_3$ | 5.6 | 4.5 | 3.7 | 2.6 | 20.3 | 17.5 | 5.4 | 4.1 | 1.4 |
| $Fe_2O_3$ | 1.3 | .46 | .92 | .45 | 2.0 | 1.6 | .64 | .74 | 2.5 |
| FeO | .31 | .39 | .69 | .33 | 3.5 | 3.9 | .60 | .60 | .27 |
| MgO | .16 | .21 | .15 | .00 | 1.0 | 1.1 | .08 | .27 | .07 |
| CaO | .05 | .12 | .05 | .05 | .04 | .14 | .06 | .05 | .07 |
| $Na_2O$ | .09 | .13 | .12 | .09 | .38 | .26 | .13 | .08 | .07 |
| $K_2O$ | 2.6 | 2.5 | 1.8 | 1.0 | 6.0 | 5.1 | 3.5 | 1.6 | .70 |
| $H_2O+$ | .78 | .70 | .56 | .58 | 3.2 | 2.9 | .66 | .89 | .58 |
| $H_2O-$ | .12 | .07 | .06 | .06 | .23 | .18 | .09 | .11 | .11 |
| $TiO_2$ | 1.2 | .06 | 1.0 | .67 | .88 | .81 | .90 | .66 | .30 |
| $P_2O_5$ | .04 | .02 | .02 | .01 | .09 | .17 | .05 | .01 | .01 |
| MnO | .00 | .00 | .00 | .00 | .02 | .03 | .02 | .00 | .00 |
| $CO_2$ | <.05 | <.05 | <.05 | <.05 | <.05 | <.05 | <.05 | <.05 | <.05 |
| Total | 99 | 99 | 100 | 100 | 100 | 100 | 99 | 99 | 100 |
| **Modes** | | | | | | | | | |
| Quartz | 77 | 80 | 82 | 87 | 26 | 29 | 75 | 77 | 92 |
| Microcline | 10 | 11 | 5 | .3 | | | 15 | 1.7 | 2.6 |
| Sericite | 11 | 9 | 12 | 11 | 69 | 60 | 5 | 20 | 4 |
| Opaque | 2.2 | P | 1.0 | 1.7 | 4 | 2.3 | 4 | 1.3 | .4 |
| Chlorite | | | | | 2.3 | 9 | | | |
| Tourmaline | P | P | P | | P | | P | P | P |
| Zircon | P | P | P | .2 | | | .2 | P | .2 |
| Sphene | | | P | .2 | | | | | |
| Biotite | | | | | | .2 | | P | |
| Plagioclase | | | | | | | P | P | .2 |
| **CIPW norms** | | | | | | | | | |
| Q | | 79.16 | 82.84 | 89.57 | | | | 83.23 | 91.09 |
| C | | 1.41 | 1.51 | 1.30 | | | | 2.17 | .42 |
| Or | | 14.77 | 10.64 | 5.91 | | | | 9.45 | 4.14 |
| Ab | | 1.10 | 1.01 | .76 | | | | .68 | .59 |
| An | | .46 | .12 | .18 | | | | .18 | .28 |
| En | | .52 | .37 | | | | | .67 | .17 |
| Fs | | .24 | | | | | | | |
| Mt | | .67 | | | | | | .02 | +.00 |
| Hm | | | .92 | .45 | | | | .73 | 2.50 |
| Il | | .11 | 1.46 | .70 | | | | 1.25 | .57 |
| Ru | | | .23 | .30 | | | | | |
| Ap | | .05 | .05 | .02 | | | | .02 | .02 |

1. Poorly sorted light-greenish-gray feldspathic quartzite containing numerous heavy-mineral partings. Very elongate clastic grains of quartz as much as 7 mm long and clastic grains of microcline as much as 5 mm long and 2 mm in diameter in a matrix composed of grains of recrystallized quartz, iron-rich muscovite, and microcline 0.05 to 0.1 mm in diameter. From roadcut on Blue Ridge Parkway 1.2 miles N. 70° E. of bridge over Linville River (area D–6, pl. 1).
2. Poorly sorted light-greenish-gray feldspathic quartzite. Clastic grains of quartz, microcline, and perthitic microcline as much as 1.5 mm in diameter in a groundmass composed of grains of recrystallized quartz, iron-rich muscovite, and subordinate microcline 0.05 to 0.1 mm in diameter. From outcrop along Linville River (area D–7, pl. 1) 1.4 miles S. 30° E. of Linville Falls (area D–6).
3. Light-gray finely laminated feldspathic quartzite with heavy mineral seams and crossbedding. Clastic grains of quartz and microcline as much as 3 mm in diameter in a matrix composed of grains of recrystallized quartz, iron-rich muscovite, and microcline 0.05 to 0.1 mm in diameter. From Wisemans View (area D–7, pl. 1).
4. Fine-grained sugary white to light-green quartzite with thin heavy-mineral seams parallel to bedding. Scattered clastic grains of quartz as much as 1 mm in diameter in a mosaic-textured matrix of quartz (average grain size 0.1 mm) and iron-rich muscovite. From the summit of Tablerock (area D–7, pl. 1).
5. Blue-green phyllite with conspicuous lineation formed by intersection of slip cleavage and bedding foliation. Lepidoblastic sericite and FeMg chlorite as much as 0.3 mm long and laminae of mosaic-textured quartz (average grain size 0.05 to 0.1 mm). From roadcut in Kistler Memorial Highway S. 80° W. from south end of 4,000-foot contour on Laurel Knob (area D–7, pl. 1).
6. Blue-black highly crenulated phyllite with quartz stringers (included in analysis). Iron-rich muscovite and FeMg chlorite layers 1 to 3 mm thick and lenses and laminae of mosaic-textured quartz (grain size 0.05 to 0.1 mm) are crenulated and offset by slip cleavage. Roadcut along North Carolina Highway 183, 0.85 mile N. 25° E. of bridge across Linville River (area D–6, pl. 1).
7. Blue-gray very fine grained, finely laminated feldspathic quartzite interbedded with dark-gray sericite phyllite. Composed of mosaic-textured quartz and microcline 0.05 to 0.1 mm in grain size and lepidoblastic sericite. Finely disseminated opaque material produces dark color. Roadcut along Kistler Memorial Highway at summit of Dogback Mountain (area D–8, pl. 1).
8. Fine-grained white to light-green crossbedded quartzite with conspicuous heavy-mineral laminae at 1- to 2-cm intervals. Clastic grains of quartz as much as 3 mm in length and of microcline as much as 1 mm in diameter in a matrix of recrystallized quartz and iron-rich muscovite 0.02 to 0.1 mm in grain size. Cut along North Carolina highway 183, 0.55 mile north of bridge across Linville River (area D–6, pl. 1).
9. Fine-grained white finely laminated quartzite with heavy-mineral partings. A few clastic grains of quartz as much as 1.5 mm in diameter and of microcline as much as 0.5 mm in diameter in a matrix composed of mosaic-textured quartz 0.05 to 0.1 mm in grain size and subordinate iron-rich muscovite and microcline. From cut along North Carolina Highway 183, 0.65 mile north of bridge across Linville River (area D–6, pl. 1).

FIGURE 60.—Contact between Shady Dolomite and quartzite of the Chilhowee Group in the Tablerock thrust sheet 1 mile north of Linville Caverns on the North Fork of the Catawba River (area C–7, pl. 1). River flows about at contact of massive quartzite and thin-bedded quartzite of the Helenmode Member (mapped with upper quartzite unit.) Lower contact of Shady Dolomite by man's right foot.

them were absent, they could not be mapped separately. A few blue-gray beds in each unit contain as much as 10 percent recrystallized opaque minerals and may be equivalent to the beds of ferruginous quartzites described by King and Ferguson (1960).

### CORRELATION AND REGIONAL RELATIONSHIPS

The Chilhowee Group is a sequence of clastic rocks, chiefly quartzite, arkosic quartzite, shale, and argillite, which constitute the earliest Paleozoic deposits in the Appalachian miogeosyncline. These rocks and their equivalents crop out in a nearly continuous belt along the northwest flank of the Blue Ridge from central Alabama to Pennsylvania. Locally, the Chilhowee Group rests nonconformably on plutonic basement rocks of early Precambrian age, but in many places the basal beds of the Chilhowee lie on sedimentary and volcanic rocks of late Precambrian age with little or no evidence of an important stratigraphic break. The group is conformably overlain by carbonate rocks of well-established Early Cambrian age. In a few places, the uppermost beds of the Chilhowee contain diagnostic Early Cambrian fossils (Butts, 1940b; Laurence and Palmer, 1963); the remainder of the sequence is unfossiliferous, except for the worm tube, *Scolithus*, and is classed as Lower Cambrian (?) because of its stratigraphic position between upper Precambrian and Lower Cambrian rocks.

In the Unaka belt in northeastern Tennessee, northwest of the Grandfather Mountain window, the Chilhowee Group is divided into three formations: the Unicoi (at the base), the Hampton, and the Erwin (at the top). As originally used (Keith, 1903, 1907a, b), these names were applied to units thought to have rather uniform lithology, but later more detailed work has shown that no such uniformity exists. The formations were therefore redefined by King and others (1944, p. 28), and their boundaries were placed at widely traceable marker beds without regard to the lithologic character of the intervening strata. King and Ferguson (1960, p. 33) noted that:

> The units as thus defined are differentiated by tracing individual beds or groups of beds, and by comparing and correlating measured sections from one area to another.

In northeastern Tennessee, the Chilhowee Group occurs in a number of northwestward-traveled thrust sheets. Conspicuous changes in thickness and lithology of formations take place within distances of a few miles along strike in some of the thrust sheets, but by far the most abrupt changes occur between adjoining sheets because of telescoping of sequences that were originally deposited many miles apart. In general, thinner and sandier Chilhowee sections are found in the Doe River inner window of the Mountain City window (pl. 3), whereas thicker and more shaly sections occur in the Shady Valley and Buffalo Mountain thrust sheets which originally lay farther southeast. Basalt flows are widespread in the middle part of the Unicoi Formation in the higher and presumably farther traveled thrust sheets and are found in a few places in the Doe River inner window. The lower part of the Unicoi Formation attains its maximum measured thickness in northeast Tennessee along the southeast side of the Shady Valley thrust sheet (King and Ferguson, 1960, p. 35), where it may include rocks of late Precambrian age, perhaps correlative with the upper part of the Grandfather Mountain Formation.

The comformable contact between the Shady Dolomite and the Chilhowee Group in the Tablerock thrust sheet shows that at least part of the Erwin Formation is present in the thrust sheet, but whether or not the lower strata are correlative with older formations in northeast Tennessee cannot definitely be established. Lack of volcanic rocks and of appreciable thicknesses of shale and the presence of clean quartzites throughout the section suggest that all Chilhowee rocks in the Tablerock thrust sheet may belong to the Erwin Formation. King and Ferguson (1960) reported no conglomerate or feldspathic quartzite in the Erwin in northeasternmost Tennessee, but scattered quartz pebbles occur in the formation in southwestern Virginia (Stose and Stose, 1957) and on Embreeville Mountain in the Buffalo Mountain thrust sheet (Rodgers, 1948). D. W. Rankin (written commun., 1965) found quartz pebble conglomerate and feldspathic quartzite in the Erwin near the Virginia-Tennessee State line, and Oriel (1950) reported that some specimens of quartzite from the Hot Springs window contain as much as 20 percent microcline. Thus, the presence of arkosic quartzite and quartz pebble conglomerate in the Chilhowee section in the Tablerock thrust sheet does not preclude correlation with the Erwin. The lack of any appreciable compositional differences between the upper and lower quartzite units is also in accord with their assignment to a single formation.

The Chilhowee section in the Tablerock thrust sheet most nearly resembles the thick sections of Erwin described by King and Ferguson (1960, pl. 9) in the Doe River inner window of the Mountain City window. The minimum thickness of Chilhowee beds in the thrust sheet, however, is nearly twice as great as the thickest section of Erwin they describe. Our synthesis of the structure (see p. 178–179 and pl. 4) suggests that the Tablerock thrust sheet probably originated southeast of the Buffalo Mountain and Shady Valley thrust sheets, both of which must have come from southeast of the Mountain City window. This structural interpretation, however, is not in agreement with the orderly southeastward change to thicker and more shaly sections in the Chilhowee Group inferred by King and Ferguson (1960, p. 81–82), if all the Chilhowee Group in the Tablerocl-thrust sheet is assigned to the Erwin Formation.

### SHADY DOLOMITE

Shady Dolomite, which overlies sandstones of the Chilhowee Group, is exposed beneath the Linville Falls fault in several small areas along the west edge of the Grandfather Mountain window. It may also be present locally beneath the alluvial deposits along the North Fork of the Catawba River.

The formation consists of fine-grained white, light-gray, or buff-gray crystalline dolomite, generally massive or vaguely mottled, but locally thin bedded or ribboned. Partings and thin beds of light-gray and greenish-gray phyllite are widespread. In some areas the dolomite is silicified to a fine-grained white, sugary or porcelaneous rock resembling quartzite. Near Linville Caverns, partially silicified dolomite contains small veinlets and irregular replacements of honey-yellow to black sphalerite, accompanied by some chalcopyrite and pyrite.

The rock is composed of a granular mosaic of equidimensional and somewhat sutured grains of carbonate generally 0.02 ot 0.2 mm in diameter, although locally as much as 0.5 mm in diameter. Quartz grains participate in the mosaic texture and constitute a trace to 15 percent in the rocks examined.

Bedding is generally not visible in small natural outcrops, but it is conspicuous in large exposures, especially in quarries. The small outcrop area of the formation and the complexity of the structure make it impossible to estimate accurately the stratigraphic thickness of the formation exposed, but it seems that at least 800 feet of dolomite must be present in the hills near the Woodlawn quarrry (area B–9, pl. 1). There, according to Watson and Laney (1906, p. 203),

> * * * several holes were put down [by the State Geological Survey] in such a manner as to include a thicknꞏss of nearly a thousand feet of the stone.

Partial analyses of Shady Dolomite from the Tablerock thrust sheet (Hunter and Gildersleeve, 1946, p. 27–28) show a MgO content ranging from 10 to 21 percent; CaO content, 27 to 31 percent; ignition loss, 41 to 46 percent; Fe, 0.25 to 0.6 percent; and acid insoluble residue, 0.7 to 8.6 percent. Light-colored dolomite has less Fe and MgO than dark-colored dolomite. Complete analyses show the following ranges: $SiO_2$, 0.60 to 5.96 percent; $Al_2O_3$, 0.60 to 1.76 percent; $Fe_2O_3$, 0.49 to 0.73 percent; CaO, 29.13 to 30.93 percent; MgO, 19.56 to 21.22 percent; $K_2O$, 0.26 to 0.41 percent; $P_2O_5$, 0.01 to 0.02 percent; $CO_2$, 40.07 to 47.10 percent (Loughlin and others, 1921). No description of the analyzed rocks is given.

No fossils have been discovered in the dolomite in the Grandfather Mountain window. The Shady is also unfossiliferous in northeastern Tennessee (King and Ferguson, 1960), but a few Early Cambrian fossils have been collected from the formation in southwestern Virginia (Resser, 1938, p. 24–25; Butts, 1940b, p. 54–56). The correlation of the dolomite in the Tablerock thrust sheet with the Shady Dolomite is based on its close lithologic similarities to the Shady in Tennessee (Keith, 1905; John Rodgers, oral commun., 1959; R. A. Laurence, oral commun., 1959) and its stratigraphic position overlying *Scolithus*-bearing quartzite similar to that of the Erwin Formation, with transition beds resembling the Helenmode Member of the Erwin Formation.

## ALLOCHTHONOUS ROCKS OF UNCERTAIN CORRELATION

The small body of quartzite and greenschist along the Linville Falls fault west of Hodges Gap (area G–2, pl. 1) is probably a fault slice emplaced early in the structural history of the area, but its structural relations are not entirely clear. Rocks in the slice have east-dipping cleavage parallel to that in adjacent rocks in the window, but the contact between the quartzite and the adjacent Grandfather Mountain Formation is not parallel to contacts within the Grandfather Mountain Formation.

The quartzite makes up the north side of a steep knob and is well exposed in a roadcut along North Carolina Highway 105 and in a nearby road-metal quarry. The gross pattern of the hill suggests that the quartzite strikes east and dips north, in contrast to the northeast strike of the adjacent Grandfather Mountain Formation. The rock is greenish gray and fairly coarse grained. It contains a few purple-hued beds and lenses, but in most places bedding is obscure. Quartz veinlets are abundant. East of the steep knob the rock has more feldspar and is not quite as resistant.

The quartzite contains strongly strained to brecciated clasts of quartz as much as 5 mm in diameter in a matrix of recrystallized quartz and iron-rich muscovite 0.02 to 0.08 mm in diameter. Zircon, sphene, opaque minerals, and allanite are accessory minerals. A few fragments of volcanic rock are present. The more feldspathic rocks contain clastic microcline.

The greenschist mapped in the quartzite east of the steep knob seems to be exposed in a gentle anticline in the quartzite; quartzite underlying the greenschist is exposed in a roadcut on North Carolina Highway 105.

The quartz-rich composition and lack of plagioclase in the quartzite suggests correlation with rocks of the Chilhowee Group. The scarcity of bedding and lack of clastic tourmaline, however, argue against that correlation. The presence of the greenschist suggests that rocks in the slice may be correlative with either the Unicoi Formation of the Chilhowee Group or the Grandfather Mountain Formation.

## STRUCTURE

### FAULTS

#### TABLEROCK FAULT

The Tablerock fault, named for its exposures in the klippe on Tablerock Mountain (area D–8, pl. 1), separates Lower Cambrian(?) and Lower Cambrian rocks of the Tablerock thrust sheet from autochthonous Precambrian rocks of the Grandfather Mountain window. The fault is parallel with bedding and cleavage in the overriding block, and in most places it truncates cleavage and bedding trends in the underblock. Where Wilson Creek Gneiss is adjacent to the fault, the fault is marked by a zone of lustrous phyllonite and cataclastic gneiss ranging from a few inches to more than 50 feet thick. Cleavage in the phyllonite adjacent to the fault is parallel with the fault plane and concordant with the structures in the overriding block; farther beneath the fault the cleavage curves to become parallel with the regional cataclastic foliation in the underblock. These relationships may be seen on the west side of the Linville River valley opposite Tablerock Mountain. In a few places, thin slices of quartzite derived from the overblock and of arkose derived from the underblock are intercalated with phyllonite and sheared gneiss. Between the Linville River and Longarm Mountain (area D–6, pl. 1), arkose of the Grandfather Mountain Formation forms the underblock, and the fault is marked by finely laminated siltstone and calcareous phyllite, probably a slice derived from the siltstone member of the Grandfather Mountain Formation. Near Crossnore (area D–5, pl. 1) a slice of basement rock is found along the fault. There the fault is folded into a syncline and trancated by the Linville Falls fault.

In the southern part of the Grandfather Mountain window, the Tablerock fault is warped into a gentle anticline, the Bald Knob anticline. The southeastern limb dips 25° to 35° near the Linville Falls fault and the Brevard fault zone. The Tablerock fault is cut by subsidiary faults of the Linville Falls fault east of Shortoff Mountain (area E–8, pl. 1) and near Bald Knob (area C–9, pl. 1). In the southwestern part of the window, the Tablerock thrust sheet is completely overridden by higher slices of Wilson Creek Gneiss and Chilhowee quartzite.

The Tablerock fault carried younger rocks over older. The presence of intercalated slices from the overriding and overridden blocks in the phyllonites along the fault and the different structural patterns in the two blocks indicate a fault of considerable magnitude rather than an unconformity along which there has been minor movement. The difference in structural patterns indicates that the Tablerock thrust sheet has traveled at least 12 miles northwestward over the autochthonous rocks of the window. The Tablerock thrust sheet may be considered a sizable subsidiary beneath the larger and thicker Blue Ridge sheet. It will be shown below that the minor structures in these two thrust sheets are similar.

### OTHER FAULTS

Subsidiaries of the Linville Falls fault are concentrated along the southeastern edge of the window. Small lenses of felsic volcanic rocks in the Grandfather Mountain Formation occur along these subsidiary faults in the northwestern part of the Lenoir quadrangle.

These faults are parallel with cataclastic foliation in the basement rocks and cleavage in the Grandfather Mountain Formation. Most of them are poorly exposed and are drawn largely on the basis of map relationships, but they are locally marked by zones of phyllonite. In many other areas, the basement rocks are highly sheared, but many of the zones of sheared rocks are not continuous, and they are, therefore, not mapped as faults, although many unrecognized faults may be present.

Apparently, these faults formed prior to emplacement of the Tablerock thrust sheet, for they are parallel to cleavage and cataclastic foliation that are truncated by the Tablerock fault.

In the southwestern extension of the window (pl. 1), where thrust sheets of Chilhowee Group rocks and of Wilson Creek Gneiss have overridden the Tablerock thrust sheet, an exceedingly complex map pattern has been produced. These thrusts are probably of the same general age as the Linville Falls and Tablerock faults and are closely related to them.

The allochthonous rocks west of Hodges Gap (area G–2, pl. 1) apparently occur in a slice along an early low-angle thrust fault, older than the Linville Falls but, possibly related to it.

### CLEAVAGE

Cleavage and cataclastic foliation are conspicuous in the autochthonous rocks in the Grandfather Mountain window. They are formed by planar orientation of mica and chlorite flakes and folia and by quartz and feldspar aggregates. Cleavage in the bedded rocks is generally parallel with the axial planes of minor folds and with cataclastic foliation in plutonic basement rocks. Both structures generally strike north or northeast and dip east or southeast (fig. 32). They are truncated by the Linville Falls fault along the north and west sides of the window and are discordant with the cataclastic foliation in the tectonically overlying Blue Ridge thrust sheet; along the southeast side of the window, discordance is less obvious. Cleavage in the Grandfather Mountain Formation and foliation of phyllonite is locally cut by later slip cleavage (not shown in fig. 32).

Locally, cataclastic foliation on the basement rocks has been warped. The most noticeable fold is in the central part of the Blowing Rock quadrangle (areas I–3 and I–4, fig. 32) where the blastomylonite seems to be folded in a sharp but rather local crinkle.

The rocks of the Tablerock thrust sheet do not have the pervasive cleavage of the autochthonous rocks. Metamorphic foliation is generally parallel to bedding. Fracture cleavage in quartzite and slip cleavage in phyllite is locally developed in the noses of minor folds. Along the southeastern edge of the Tablerock thrust sheet, adjacent to the Linville Falls fault, cleavage marked by alinement of micaceous folia is locally developed.

### LINEATION

Mineral lineation consisting of alined micaceous streaks, elongated clastic grains, porphyroclasts, mineral aggregates, pebbles, and amygdules is found on most cleavage and foliation planes in the autochthonous rocks but is absent on the late slip cleavage which locally cuts older cleavage. The lineation trends rather uniformly northwest and plunges southeast in the autochthonous rock (fig. 33).

In the Tablerock thrust sheet, mineral lineation lies on bedding planes or on foliation planes parallel to bedding. It trends northeast or southwest, parallel to lineation in the autochthonous rocks. Northwest of the Bald Mountain axis, it plunges gently northwest, and southeast of the axis, it plunges gently to moderately southeast or south.

Offset of individual beds in the sedimentary rocks and of pegmatite dikes in the plutonic basement rocks shows that movement in the cleavage planes has been parallel with the mineral lineation and demonstrates that this lineation is in the *a* direction.

## FOLDS

### BASEMENT ROCKS

Small tight to isoclinal folds are well developed in the layered part of the Wilson Creek Gneiss in the Blowing Rock quadrangle. Good examples can be seen in roadcuts along U.S. Highway 321 in area I–5, plate 1, and along the Buffalo Creek road in area J–4, plate 1. The axial planes and axes of about 70 percent of the isoclinal folds and about 40 percent of the nonisoclinal folds parallel the foliation and mineral lineation (fig. 61*A*). The axial planes at an angle to the foliation nevertheless strike and dip in the same quadrant; fold axes at an angle with the lineation trend more north and east than the lineation. All gradations between isoclinal and nonisoclinal folds occur, and there is no apparent difference in age between the two types.

In most outcrops of layered Wilson Creek Gneiss, layering and foliation are parallel. The gross structure of the layered rocks probably consists of tight or isoclinal folds sheared out during the formation of the cataclastic foliation in the window rocks.

The blastomylonite and phyllonite in the plutonic basement rocks locally have small-scale crinkles with gently plunging axes trending N. 70° E. and axial planes of various dips trending about N. 60° E. (fig. 61*B*). Locally, slip cleavage parallel to these axial planes cuts the main cataclastic foliation.

Axes of crinkles and intersections of slip cleavage with foliation (fig. 61*B*) generally trend about N. 70° E., roughly perpendicular to the mineral lineation; they plunge gently northeast. The small maxi-

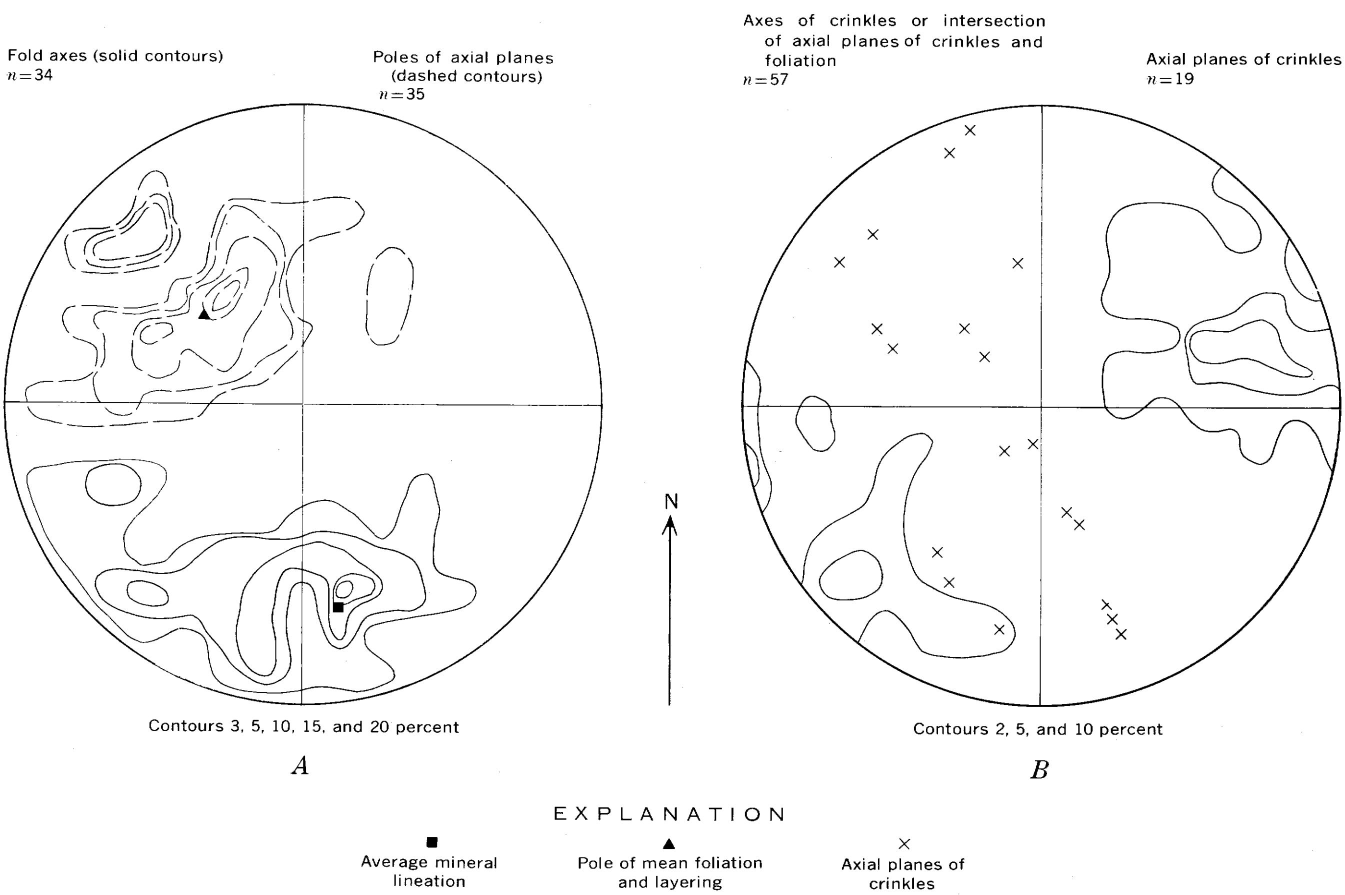



FIGURE 61.—Orientation of minor folds in basement rocks of the Grandfather Mountain window. All diagrams are equal-area projections in the lower hemisphere with planes of projection horizontal and north at top. Contours show percentage of points falling within 1 percent of the area of the diagram. *A*, Tight and isoclinal folds in layered Wilson Creek Gneiss in the Blowing Rock quadrangle. *B*, Axes of crinkles or intersection of axial planes of crinkles and foliation.

mum of southwest-plunging crinkle axes striking N. 50° E. is from measurements in the Linville Falls quadrangle where the mineral lineation averages about N. 40° W. The crinkles may be structures in the *b* direction formed under the same stress conditions but somewhat later than the mineral lineation.

### GRANDFATHER MOUNTAIN FORMATION

#### MAJOR FOLDS

The main outcrop belt of the Grandfather Mountain Formation lies on the southeast limb of a complex synclinorium overturned to the northwest. This structure is cut by both the Linville Falls and Tablerock faults. In the east and central parts of the main outcrop area of the formation, beds are generally overturned; in the northwest part, right-side-up beds are more abundant. The outcrop pattern of rock units is largely controlled by the major synclinorium and its attendant medium- and small-scale folds. No major repetition of rock units is evident, although some minor repetition occurs in the western part of the Linville quadrangle. The northwest limb of the synclinorium is presumably concealed beneath the Blue Ridge thrust sheet northwest of the Grandfather Mountain window.

The nearly horizontal northeast trending axis of the major synclinorium is warped around a northwest-trending axis that passes through the northwest corner of the window (area D–3, pl. 1). Other large- and medium-scale folds are locally superimposed on the earlier synclinorium.

The open syncline plunging gently south under the north end of the Tablerock thrust sheet (area D–5, pl. 1) seems to be a later structure superimposed on the major synclinorium; the two south-plunging folds in area D–4 are moderately sharp and seem to have the same pattern as early minor folds in that area. Other medium-scale folds may influence the outcrop pattern of the arkose unit above the Montezuma Member in the northwest part of the window, but scarcity of bedding and of criteria for tops of beds makes it uncertain whether this outcrop pattern is primarily due to intertonguing of rock units or to folding. At least one body of greenstone seems to be involved in northeast-plunging folds. The long tongue of arkose projecting southward into basement rocks in areas G–4 and G–5 lies in a tight north-plunging syncline apparently superimposed on the earlier northeast-trending folds.

#### MINOR FOLDS

Minor folds are ubiquitous and cause an unknown but probably appreciable amount of small-scale repetition of rock units. The oldest and most widespread folds are asymmetric or overturned to the north or northwest. They are tight or isoclinal in the southeastern part of the main outcrop area of the formation (fig. 62*A* and *B*) but are more open in the northwest part of the area (fig. 62*C*). Siltstone and phyllite, especially where confined between arkose beds, are generally more tightly and complexly folded than the more competent arkose. Such disharmonic folding precludes accurate estimates of stratigraphic thickness of the incompetent rocks.

In the northeastern part of the main outcrop area of the Grandfather Mountain Formation, small-scale open folds are superimposed on the older structures. The open folds deform the cleavage that parallels the axial planes of the earlier folds and have steep to vertical axial planes that strike northeast. Slip cleavage parallel with the axial planes of the open folds offsets the older pervasive cleavage (fig. 63).

#### AGE RELATIONS BETWEEN FOLDS AND OTHER STRUCTURES

Attitudes of bedding in various parts of the main outcrop belt of the Grandfather Mountain Formation are shown by statistical diagrams on plate 7. The orientations of bedding, cleavage, lineation, minor folds, and other structures in the northern and southern parts of the main outcrop belt are summarized in figure 2, plate 5. The bedding diagrams for the individual sectors (pl. 7) clearly reflect the difference in style of the early small-scale folds between the southeastern and northwestern parts of the area. In the southeastern part of the area, where the folds are tight or isoclinal, bedding poles cluster around diffuse maxima or rudely developed partial girdles; in the northwestern part of the outcrop area, folds are more open, and the bedding girdles are better defined and more complete.

The axes of most of the early minor folds (pl. 5, fig. 2, diagrams *A*–4 and *B*–4) are coincident with axes of the bedding girdles. The fold and girdle axes, however, are not everywhere consistent with the gross outcrop pattern of the major lithologic units. In section IV (pl. 7) for example, the girdle axis and fold axes plunge 45° NE., whereas the contacts trend uniformly northeast. In this sector, at least, the minor folds that are reflected in the bedding girdles are apparently superimposed on earlier large-scale folds that control the outcrop pattern. Only locally are the small folds reflected by minor irregularities in the map pattern. Similar relations are found in

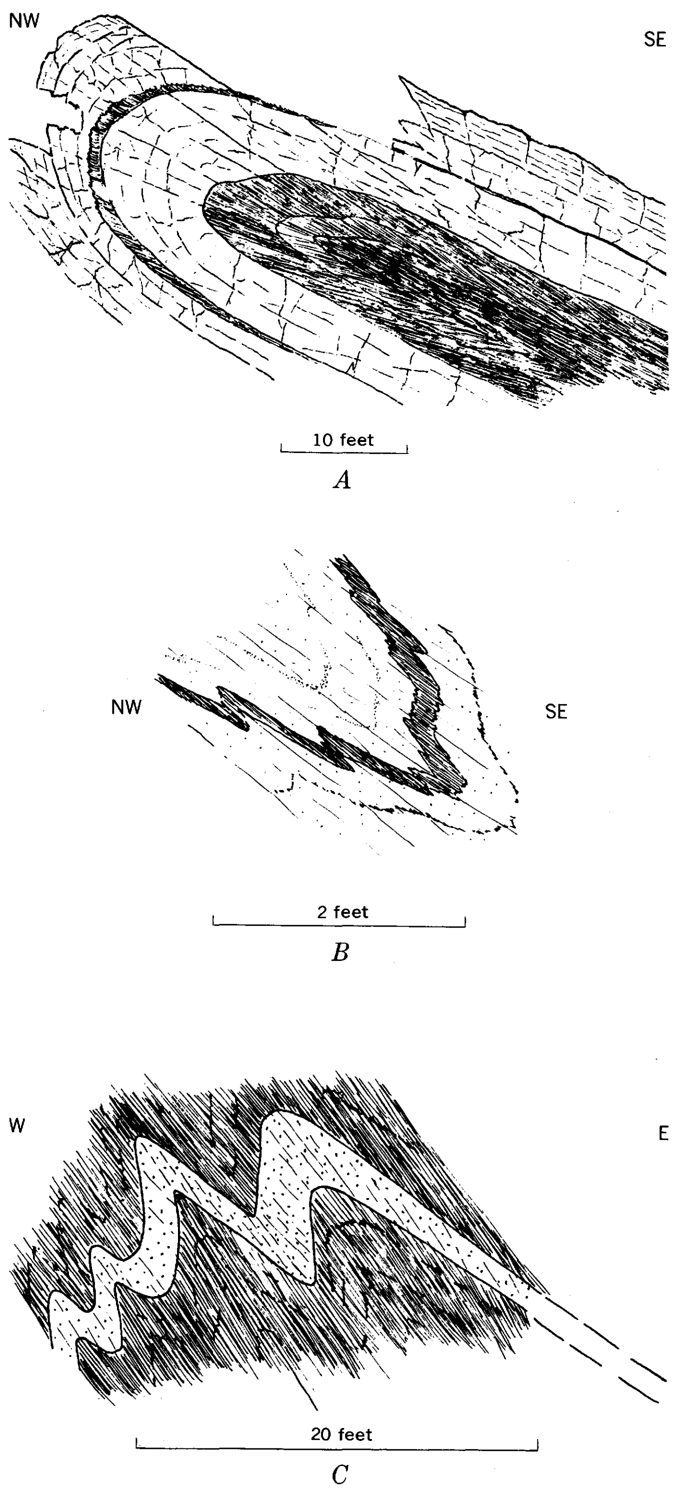



FIGURE 62.—Typical early folds in the Grandfather Mountain Formation. For location of sectors see plate 7. *A*, Isoclinal fold in arkose in sector VII. *B*, Overturned asymmetric fold in arkose and phyllite on boundary between sectors IV and VII. Fold axis trends N. 45° E. and plunges 5° NE. *C*, Asymmetric folds in siltstone and arkose in the northwest part of sector XIII. Fold axes trend N. 5° E. and plunge 5° N.

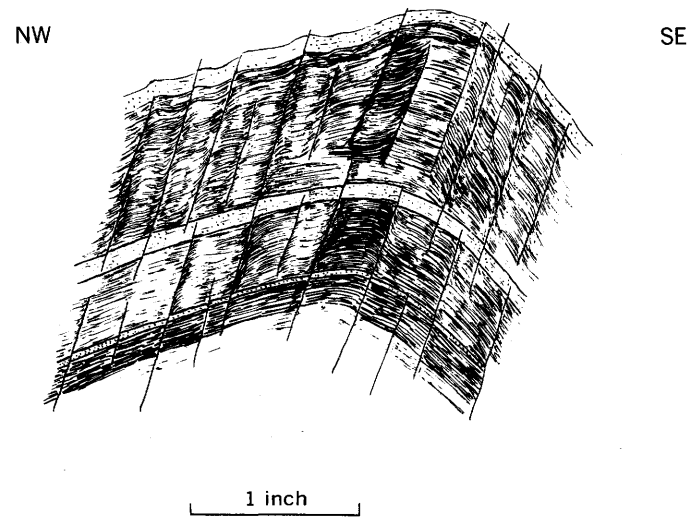



FIGURE 63.—Late slip fold in siltstone of the Grandfather Mountain Formation in sector IV (pl. 7). Slip cleavage parallel to axial plane of this fold cuts earlier cleavage parallel to bedding in the limbs of earlier isoclinal folds. Fold axis trends N. 45° E. and plunges 60° NE.

sectors I, III, and V, but elsewhere the discordance between gross map pattern and girdle and fold axes is not evident, and it is therefore uncertain whether the early minor folds are contemporaneous with or superimposed on the large early folds.

The conspicuous regional cleavage, $S_2$ (pl. 5, fig. 2, diagrams *A*–2 and *B*–2), is parallel to the axial planes of the minor folds (pl. 5, diagrams *A*–4 and *B*–4) and was presumably developed at the same time. Where the trend of the minor folds diverges from that of the large early folds, the cleavage is clearly superimposed on the older structures. Nowhere have the two cleavages been found in the same outcrop, but the abrupt change (fig. 32) from northeast-striking cleavage parallel to the trend of lithologic units in sectors V, VI, and VII to cleavage striking north or northwest at high angles to the contacts in sectors II, III, and IV strongly suggests that the cleavages and minor folds are of two distinct generations. Where the minor folds parallel the larger folds, it is not possible to distinguish the minor folds and the cleavage that date from the earliest episode of folding from the later structures.

The pervasive mineral lineation ($L_1$) lies on the $S_2$ cleavage planes and maintains virtually the same orientation throughout the Grandfather Mountain Formation (fig. 33). In the northern part of the outcrop area the lineation is approximately normal to the maximum concentration of axes of the early minor folds (pl. 5, fig. 2, diagram *A*–3), but in the southern part of the area it is oblique to the axes of the minor folds. These relations suggest that it was formed concurrently with the second generation of cleavage and minor folds in the northern part of the area, whereas in the southern part it was formed by renewed movement along cleavage planes of the first generation. The striking parallelism in trend between the mineral lineation in the Grandfather Mountain Formation and the cataclastic mineral lineation in nearby parts of the Blue Ridge thrust sheet indicates that both structures were formed during northwestward transport of the thrust sheet.

The small open folds (pl. 5, fig. 2, diagram *A*–5) are superimposed on folds of the second generation in the northern part of the area. Nowhere do they affect the outcrop pattern of mappable lithologic units, nor are they reflected to any recognizable degree in the bedding girdles. The late slip cleavage parallel to their axial planes (pl. 5, fig. 2, diagram *A*–6) is also parallel to a strongly developed regional joint system (pl. 5, fig. 2, diagram *A*–7) which is also recognizable in the southern part of the area (pl. 5, fig. 2, diagram *B*–5) where the open folds and slip cleavage are absent. The occurrence of similar open folds and joints in adjacent parts of the Blue Ridge thrust sheet (pl. 5, fig. 1, diagram *C*–1) suggests that these structures formed subsequent to or in the closing phases of thrusting.

The following sequence of structural events is inferred from the relations between folds and other structures in the Grandfather Mountain Formation:

1. Initial folding—formation of medium- and large-scale folds that control the outcrop pattern of lithologic units and of associated minor folds and cleavage of the first generation.
2. Second folding—formation of minor folds and cleavage of the second generation; obliteration of earlier minor folds and cleavage in the northern part of the main outcrop area. Concurrent with or closely followed by 3.
3. Reactivation of all preexisting cleavages and formation of mineral lineation during northwestward movement of the Blue Ridge thrust sheet.
4. Formation of small open folds, crenulations, slip cleavage, and predominant regional joint system.

## TABLEROCK THRUST SHEET

### MAJOR FOLDS

Most irregularities in outcrop pattern of the stratigraphic units in the Tablerock thrust sheet (pl. 1) are due to intersection of gently dipping units with the rugged topography. Except for the Bald Knob anticline, the shallow syncline in the northern end of thrust sheet, and the folds in the southwestern part of the thrust sheet (pl. 1), no folds large enough to affect the gross outcrop pattern of the lithologic units have been detected. Locally, however, folds are mappable at a 1:48,000 scale. The overturned rocks in the southwestern extension of the window are interpreted to be a part of an overturned limb of an isoclinal fold in the rocks of the Tablerock thrust sheet that has subsequently been cut by thrust faults and overridden by basement rock (fig. 64). In the fence diagram (pl. 2), the overturn has been extended well to the north for purposes of illustration; on the surface, the northernmost extensive area of overturned rocks is on the west flank of Dobson Knob (area C–9, pl. 1).

### MINOR FOLDS

Smaller scale folds, visible in single outcrops or inferred from closely spaced groups of outcrops, are common. The minor folds are of two types and are apparently of at least two generations.

Minor folds of the first type (fig. 65) are tightly appressed or isoclinal and have horizontal or gently plunging axes with erratic trends. Many of these folds have flat or gently dipping axial planes, but a few axial planes dip as much as 50°. Most of the folds are overturned to the west, northwest, or north; a few are overturned to the northeast. Upright limbs are many times longer than steeply dipping or overturned limbs.

Beds are thickened in fold noses and thinned on the limbs; locally incompetent phyllite layers are pinched out. Thin quartzite layers in phyllite are boudinaged and locally remain only as discontinuous lenses in fold noses. Minor folds in phyllitic layers are commonly disharmonic with respect to more competent quartzite layers. Locally, slip cleavage in phyllite and rude fracture cleavage in quartzite are developed parallel to the axial planes of these folds.

Minor folds of the second type are open symmetrical or slightly asymmetrical folds in quartzite and associated crenulations in phyllite (fig. 66). Most of these folds have nearly horizontal northeast-trending axes, and where the folds are asymmetric, axial planes dip southeast.

The wavelength of the open folds ranges from a few inches to several tens of feet but is generally less than 5 feet. Crenulations in phyllite range from about 5 mm to less than 1 mm in amplitude and are

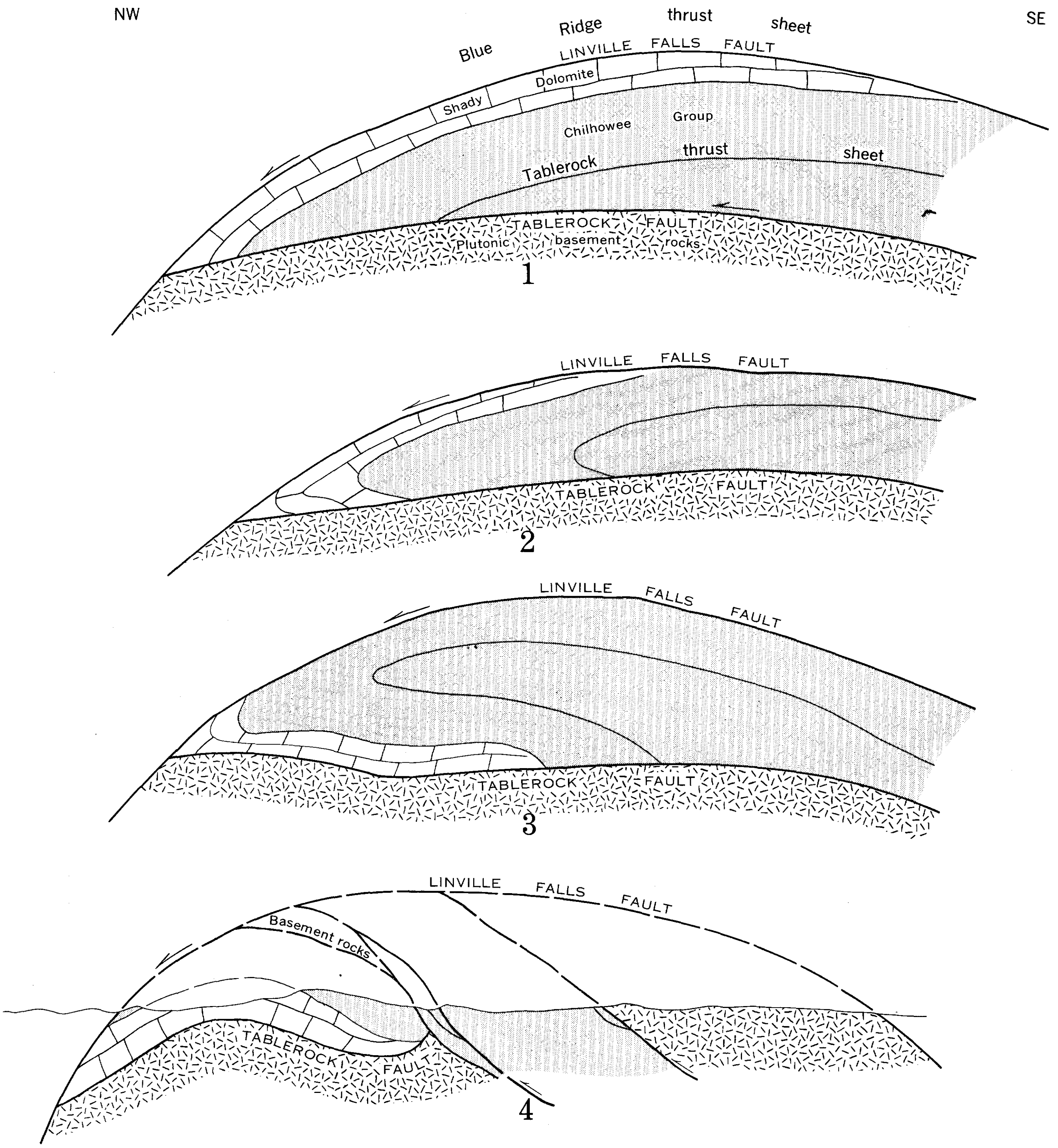



FIGURE 64.—Possible mode of development of antiform in Shady Dolomite near Woodlawn (area B–9, pl. 1). 1, Tablerock thrust sheet moves as tectonic slice during concurrent movement along Linville Falls and Tablerock faults. 2, Development of isoclinal fold in Tablerock thrust sheet as movement along Tablerock fault ceases. 3, Further overturning of fold during continued movement of Blue Ridge thrust sheet along Linville Falls fault. 4, Tablerock thrust sheet overridden by numerous other fault slices during latest stages of movement along Linville Falls fault.

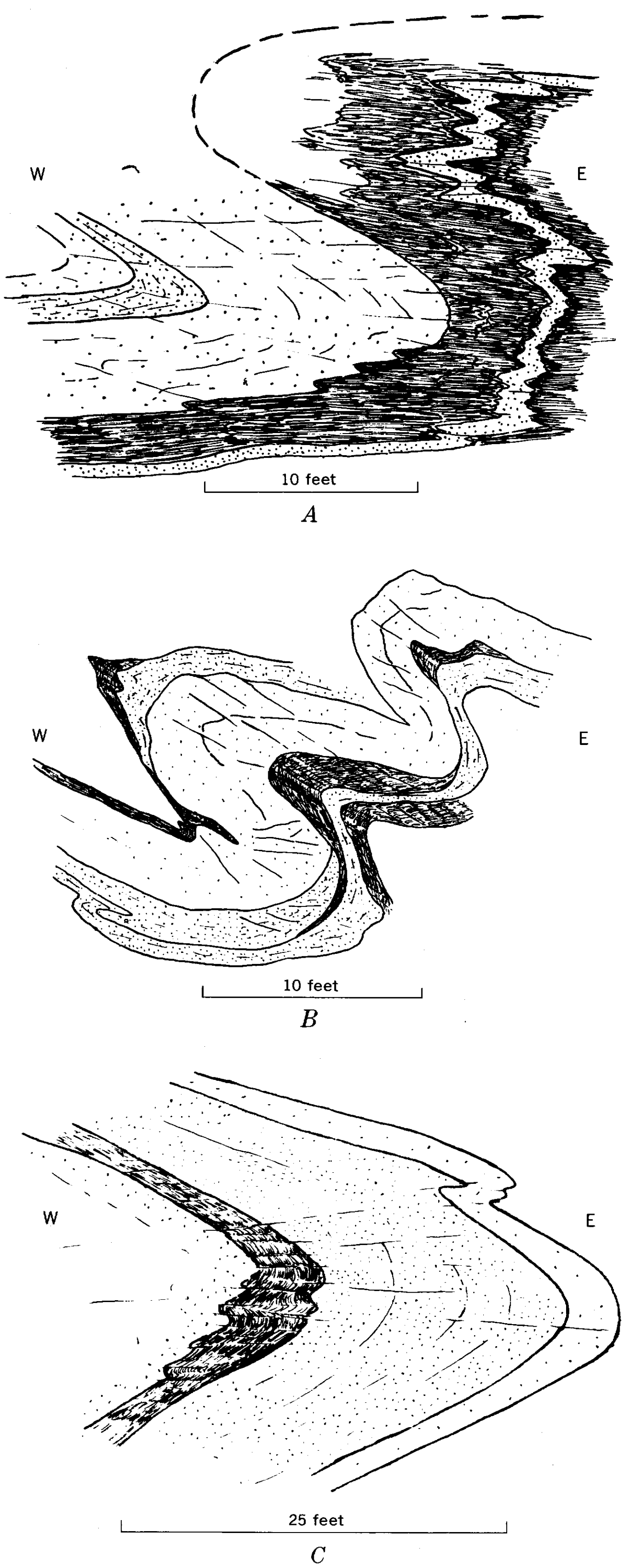



FIGURE 65.—Tightly appressed and isoclinal folds in rocks of the Chilhowee Group in the Tablerock thrust sheet. For location of sectors, see plate 7. *A*, Recumbent folds in quartzite and phyllite in northeastern part of sector E. Fold axes trend N. 5° W. and plunge 5° N. *B*, Overturned asymmetric folds in quartzite and phyllite in southeastern part of sector F. Fold axes are horizontal and trend N. 15° W. Axes of crenulations on the limbs of these folds trend N. 40° E. and plunge 10° NE. *C*, Nose of recumbent anticline in quartzite and phyllite in the northeastern part of sector E. Fold axis is horizontal and trends N. 25° E.

parellel to axes of larger folds of this type in adjacent quartzite beds.

Individual layers maintain rather uniform thicknesses in the open folds, and boudinage effects are less conspicuous than in the earlier folds. Axial-plane cleavage commonly is found in noses of asymmetric folds but is generally absent in the more open symmetrical folds.

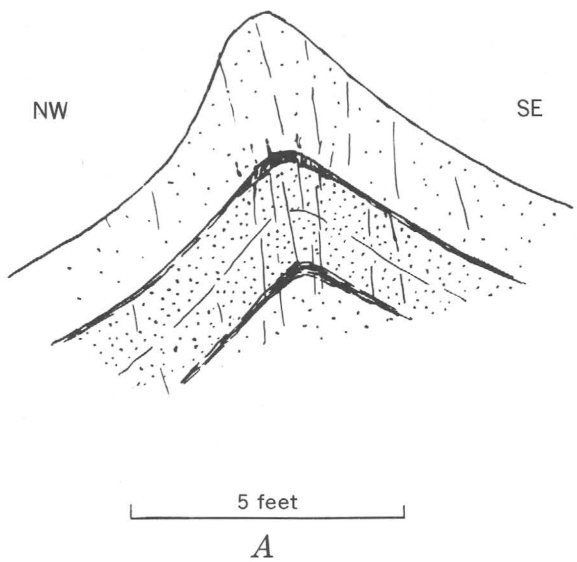
Locally, open folds and small crenulations of the second type appear on the limbs of isoclinal folds of the first type and have axes transverse to axes of the earlier folds. Where folds of the second set are asymmetric or overturned, they become indistinguishable from folds of the first type, and some observed folds may have been assigned to the wrong set.

Southeast of the axis of the Bald Knob anticline (pl. 7), open folds and crenulations of the second set are less common. Cleavage dips moderately southeast, and field relations suggest that much of the cleavage in this area is related to shearing along the Linville Falls fault rather than to minor folds. Most of the observed minor crenulation axes are parallel to the intersection of cleavage and bedding.

The most strongly developed joints in the rocks of the Tablerock thrust sheet are nearly vertical and strike northeast. These joints are not mineralized and cut all other structures. Joints in the more massive quartzite beds are spaced tens of feet apart, but in phyllite and sericite quartzite beds they are more closely spaced and locally resemble fracture cleavage. No lineations have been observed on the joint surfaces.

### STRUCTURAL GEOMETRY

The geometry of the various structures in the Tablerock thrust sheet in the Linville Falls and Linville quadrangles is summarized in the contour diagrams (pl. 5, fig. 3). Structures in the part of the thrust sheet southwest of the Linville Falls quadrangle were not studied in detail. Separate sets of



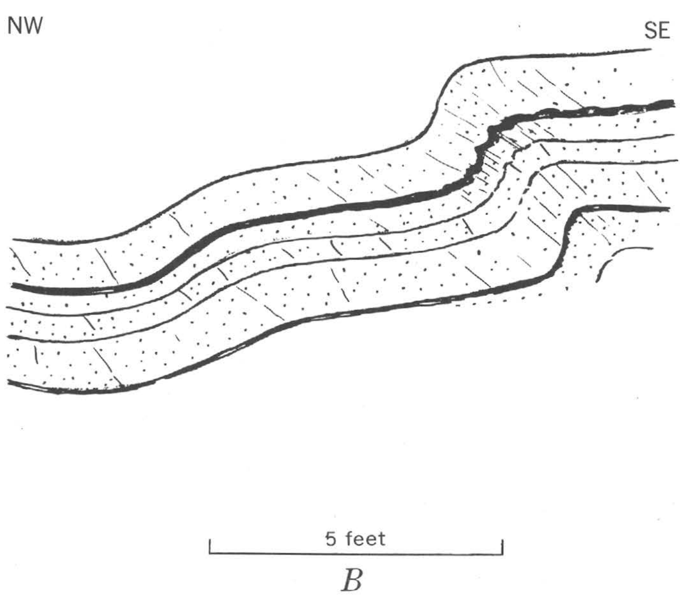




FIGURE 66.—Open folds in rocks of the Chilhowee Group in the Tablerock thrust sheet. For location of sectors, see plate 7. *A*, Slightly asymmetric open fold in quartzite in southwestern part of sector B. Note development of cleavage parallel to axial plane. Fold axis trends N. 50° E. and plunges 10° SW. *B*, Open folds in quartzite near Linville Falls in southwestern part of sector A. Fold axes are horizontal and trend N. 45° E.

diagrams have been prepared for the parts of the thrust sheet northwest and southeast of the axis of the Bald Knob anticline. Diagrams of bedding poles for small subdivisions of the area are shown on plate 7.

The bedding diagrams (pl. 7 and pl. 5, fig. 3, diagrams *A*–1 and *B*–1) show that most of the bedding attitudes are consistent with the generally simple structure of the thrust sheet as inferred from mapping. In most of the diagrams, bedding poles cluster around single high maxima and indicate gentle dips away from the axis of the Bald Knob anticline. Only locally are indistinct bedding girdles evident, and even these largely disappear on the summary diagrams (pl. 5, fig. 3, diagrams *A*–1 and *B*–1), showing that the minor folds have little influence on the statistical distribution of bedding poles.

Axes of small tightly appressed or isoclinal folds (pl. 5, fig. 3, diagrams *A*–2 and *B*–2) form a nearly complete horizontal girdle having low maxima in the east-west, north-south, and northeast-southwest positions. Most axial planes are nearly horizontal or have low dips to the south, southeast, or northeast.

Orientations of axes of open folds and crenulations and of poles of cleavage planes are summarized in diagrams *A*–3 and *B*–3, pl. 5, fig. 3. Fold and crenulation axes northwest of the Bald Mountain anticline have a remarkably constant orientation and cluster around a single high maximum at S. 40° W., plunging 10° SW. Cleavage poles are normal to the fold-axis maximum. Most cleavage planes dip gently or moderately southeast and are apparently related to folds of the second type, but some of the gently dipping or horizontal cleavage planes are probably related to the earlier folds.

Mineral lineations have very consistent orientation patterns (pl. 5, fig. 3, diagrams *A*–4 and *B*–4): Northwest of the Bald Mountain anticline, the average trend is about N. 40° W. and horizontal; southeast of the axis, the most frequent trend is S. 25° E., plunging 30° SE., but the diagram also shows a small submaximum with a more southerly trend.

Diagram *C* (pl. 5, fig. 3) shows the orientation of joint poles. No differences were noted between the distribution of poles on the two sides of the Bald Mountain anticline; therefore, the data are summarized on a single diagram. Most joints are nearly vertical and strike about N. 70° E.; a small proportion strike northwest.

### RELATION BETWEEN STRUCTURES IN THE TABLEROCK THRUST SHEET AND THRUSTING

The character of the mineral lineation, the close correspondence in trend between the lineation in the Tablerock thrust sheet and the similar lineation in the overridden rocks (compare diagrams *A*–4 and *B*–4, pl. 5, fig. 3, with diagrams *A*–3 and *B*–3, pl. 5, fig. 2), and the occurrence of similar lineations in blastomylonite along the Linville Falls fault and in retrogressively metamorphosed rocks of the Blue Ridge thrust sheet above the fault all suggest that the lineation is a cataclastic *a* lineation formed during movement along the major thrust faults.

The consistent orientation of the axes of minor open folds and crenulations normal to the lineation (fig. 67) suggests that these structures were formed during the same episode of movement. The direction of asymmetry of these folds is consistent with northwestward movement of the overriding Blue Ridge thrust sheet in the direction of the mineral lineation. The folds may have been a somewhat later response to movement of the thrust sheet.

The distribution of the earlier fold axes (pl. 5, fig. 3, diagram *A*–2) has nearly perfect orthorhombic symmetry, the mineral lineation and the axes of the later folds being along the symmetry planes. The reason for this interesting distribution is not clear, but the pattern suggests that tight folds were not formed by regional folding prior to and independent of the thrusting. They may have formed in diverse orientations as a result of local stress variations during an early stage of thrusting, but the present pattern is probably largely due to rotation of the folds by differential forward movement of the thrust sheet in the direction of the *a* lineation. The direction of overturning of the earlier folds is consistent with this interpretation.

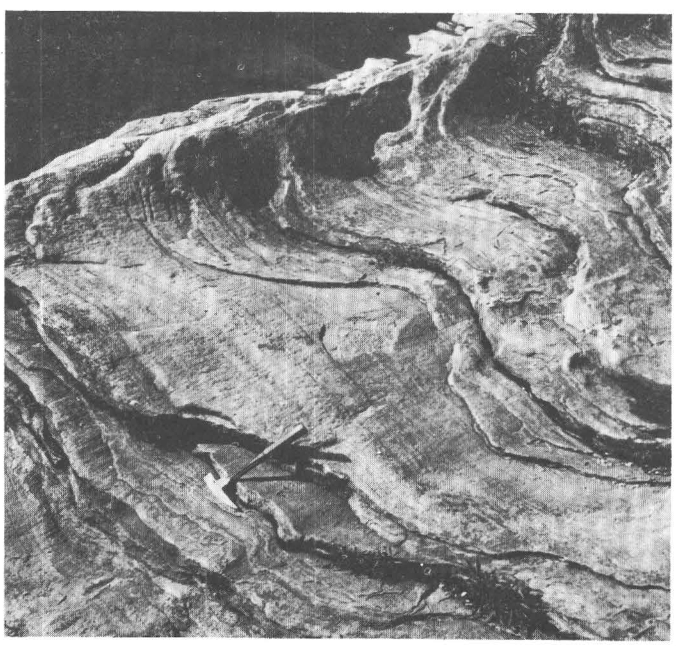


FIGURE 67.—Mineral lineation and axes of open folds in quartzite of the Chilhowee Group in the Tablerock thrust sheet at overlook along National Park Service trail at head of Linville Falls (area D–6, pl. 1). Fold axes trend N. 45° E.; mineral lineation trends N. 45° W.

The earlier folds do not involve the Linville Falls or Tablerock faults, but small open folds and crenulations of the second type are found in phyllonitic and blastomylonitic rocks along the faults. Where the fault planes are exposed, they are not folded. The younger folds therefore probably formed during a late stage of thrusting but before movement had ceased.

The distribution of joint poles (pl. 5, fig. 3, diagram *C*) does not fit in with the symmetry of the other structures in the Tablerock thrust sheet but is identical with the distribution patterns of joint and fracture cleavage poles in the Grandfather Mountain Formation (pl. 5, fig. 2, diagrams A–6, A–7, B–5). It is therefore inferred that the major joint sets formed after the Tablerock thrust sheet reached its present position. Formation of the joints may be related to the gentle arching of the thrust sheet along the Bald Knob anticline.

The structural geometry of the Tablerock thrust sheet more nearly resembles that of the adjacent part of the overlying Blue Ridge thrust sheet than it does that of the autochthonous rocks of the Grandfather Mountain window (compare pl. 5, fig. 1 and pl. 5, fig. 2). This seems to confirm that formation of the minor structures in the Tablerock thrust sheet was closely related to the thrusting process.

## METAMORPHISM

The Precambrian basement rocks in the Grandfather Mountain window (pl. 6) underwent plutonic metamorphism which is dated by uranium-lead isotopic ratios in zircon from the Wilson Creek Gneiss and Blowing Rock Gneiss as 1,000 to 1,100 m.y. ago, approximately equivalent to the time of metamorphism of the Grenville Series in the Adirondacks (Davis and others, 1962). The age and previous history of the pregranitic rocks is unknown. It is also uncertain whether the Brown Mountain Granite was emplaced during this plutonic event or at some later time before the deposition of the Grandfather Mountain Formation.

After deposition of the upper Precambrian rocks, all the rocks of the Grandfather Mountain window were sheared and metamorphosed to a low and rather uniform grade. Progressive metamorphism of the upper Precambrian rocks and concurrent retrogressive metamorphism of the plutonic basement produced mineral assemblages characteristic of the quartz-albite-epidote-biotite subfacies of the greenschist facies of Fyfe, Turner, and Verhoogen (1958). The more mafic Precambrian rocks contain porphyroclastic hornblende and recrystallized actinolite and chlorite. Relict oligoclase or andesine is found in a few of the rocks.

Some basement rocks were only incipiently sheared and recrystallized, whereas others were entirely converted to blastomylonite or phyllonite.

Original textures in sedimentary and volcanic rock were not completely destroyed, except in some of the finer grained rocks. However, all plagioclase was altered to albite, and almost all pyroxene, to chlorite and actinolite. The fine-grained mafic volcanic rocks, such as those of the Montezuma Member, were almost completely reconstituted to albite-epidote-chlorite-actinolite greenschists. Coarser grained volcanic rocks contain relict phenocrysts of potassic feldspar, plagioclase altered to albite, quartz, and, rarely, pyroxene. The Linville Metadiabase was metamorphosed to rocks ranging from strongly sheared and completely reconstituted greenschist to altered but relatively unsheared diabase.

The recrystallized minerals are locally coarser grained near the southeast side of the window. In the Blowing Rock quadrangle, recrystallized Wilson Creek Gneiss directly adjacent to the Linville Falls fault contains recrystallized oligoclase ($An_{15}$).

This metamorphism was apparently contemporaneous with formation of cataclastic foliation in the basement rocks, the two older cleavages in the Grandfather Mountain Formation, and the cleavage parallel with bedding in the Tablerock thrust sheet. A single rubidium-strontium date on biotite from the Wilson Creek Gneiss (Davis and others, 1962) suggests that this metamorphism occurred about 350 m.y. ago, near the end of the middle Paleozoic.

Rankin, Stern, Reed and Newell (1969) interpreted discordant zircon ages from felsic volcanic rocks of the Grandfather Mountain Formation as being due to episodic lead loss at 240 m.y. during late Paleozoic thrusting. The exact time of this lead loss in relation to the sequence of minor structures in the rocks of the window is not known.

The inferred ages and relationships of geologic events in the Grandfather Mountain window are summarized in table 29.

Green iron-rich muscovite occurs throughout the Grandfather Mountain window in metamorphosed basement rock, sandstone, and felsic volcanic rock. Its occurrence in the arkoses of the Grandfather Mountain Formation could be due to an originally high content of ferric iron, and the arkoses could have been red beds whose color was changed during metamorphism. The environment of deposition is probably not as critical as metamorphic conditions in controlling oxidation or reduction of iron during regional metamorphism, however, as the mica is also found in metamorphosed quartzite of the Chilhowee Group, felsic volcanic rock, and basement rock as well as in the arkoses.

The widespread occurrence of the iron-rich muscovite is probably related to bulk composition of the rocks and to metamorphic conditions. Ernst (1963) suggested that a high ratio of fluid pressure to total pressure and low temperature favors the formation of phengite in glaucophane schist terranes of Japan, California, and the Alps. Lambert (1959) analyzed muscovites from psammatic and pelitic rocks of various metamorphic grades in the Moine Schist and found that rocks of low metamorphic grade contain muscovite richer in $Fe_2O_3$ than those of higher grade. He attributes the high ferric iron content of the muscovite to metamorphism under unusually strong oxidizing conditions. The presence of the iron-rich muscovitic mica in the rocks of the Grandfather Mountain area may indicate that similar conditions prevailed.

Bryant (1967) discussed the occurrence of the iron-rich muscovite in more detail and concluded that rocks of the Grandfather Mountain window were open to oxygen during metamorphism, which took place under conditions of high shearing stress, low temperature, and high $P_{H_2O}$.

## INNER PIEDMONT BELT

The rocks southeast of the Brevard fault zone form a structurally complex metamorphic terrane consisting principally of layered biotite and biotite-amphibole gneiss, mica and sillimanite schist, amphibolite, and large concordant bodies of cataclastic augen gneiss. These rocks are invaded by concordant or semiconcordant plutons of granitic rocks, principally quartz diorite and granodiorite, commonly flanked by zones of migmatites apparently derived from the enclosing rocks. The gneisses and schists in the southeastern part of the area are of uniform high metamorphic grade, but in a belt 4 to 5 miles wide (pl. 6*D*) immediately southeast of the Brevard fault zone, rocks of the Inner Piedmont belt have been intensely sheared and retrogressively metamorphosed under medium- and low-grade conditions and have a structural pattern differing from that of the high-grade rocks.

Deep colluvial cover, scarcity of float, and lack of continuous exposures seriously handicap detailed geologic mapping in the Inner Piedmont belt. These handicaps, plus the structural complexity and lack of distinctive marker horizons, have obscured the stratigraphic sequence of the layered rocks. These rocks are therefore grouped into lithologic units for the purposes of mapping and description; no implication is made as to their original stratigraphic relationships. The intrusive rocks and migmatites are described in order of their inferred ages. Contacts between rock units in the Inner Piedmont are commonly gradational and are generally poorly exposed. Many of the contacts shown on the geologic map are therefore diagrammatic at best.

## ROCK UNITS

### LAYERED ROCKS

#### BIOTITE GNEISS

Biotite gneiss interleaved with biotite and biotite-muscovite schist and biotite-hornblende gneiss and containing pods and layers of amphibolite, quartzite, quartz schist, and calc-silicate rocks constitutes the bulk of the layered rocks in the Inner Piedmont belt. The typical gneiss is a fine-grained well-layered light-, medium-, or dark-gray rock consisting of various proportions of quartz, plagioclase, and biotite and subordinate amounts of muscovite, epidote, potassic feldspar, garnet, and chlorite. Common accessory minerals are zircon, apatite, allanite, magnetite, ilmenite, and pyrite. Limonite, carbonate minerals, and clay minerals are common products of incipient weathering.

Layers range from fractions of an inch to several feet in thickness. Differences in color and texture between layers are due chiefly to variations in the proportions of quartz, feldspar, and biotite. Contacts between layers are sharp (fig. 68). In exceptionally good outcrops, some individual layers and groups of layers can be traced for several hundred feet along strike without significant variations in thickness, but more competent layers are commonly boudinaged and many layers lens out, apparently because of shearing parallel to the layering planes.

Foliation is defined by parallel orientation of micaceous minerals and by quartz-feldspar folia. It is generally parallel to the layering but locally transects layering in fold noses, where it lies parallel to axial planes. With the possible exception of the layering, no sedimentary textures or structures have been observed in the gneisses.

Where they have not been affected by shearing and retrogressive metamorphism related to the Brevard fault zone, the biotite gneisses are of high metamorphic grade and are granoblastic or rudely foliated (fig. 69*A*). Quartz and plagioclase form an inequigranular mosaic of sutured grains generally ranging from 0.1 to 1.0 mm in diameter and arranged with their long dimensions parallel to the foliation. Plagioclase also occurs as scattered, faintly zoned porphyroblasts as much as 1 cm in diameter. Untwinned potassic feldspar forms intergranular films and small irregular grains interstitial to quartz and plagioclase in the mosaic, and microcline occurs as porphyroblasts as much as 1 cm in diameter, having reaction rims of myrmekite at contacts with plagioclase. Biotite occurs in stubby subhedral to anhedral flakes 0.25 to 2.0 mm long. The smaller flakes are commonly randomly oriented, but the larger flakes are rudely alined with the foliation, suggesting that the mineral is synkinematic or postkinematic. The biotite is most commonly pleochroic in shades from light yellow to deep red brown, but in some specimens it is olive green or deep smoky brown. Muscovite occurs as rudely alined synkinematic flakes comparable in size with the largest biotite flakes and as a secondary mineral forming fringes on the ends of biotite books, skeletal aggregates replacing feldspar, and sericite flakes in feldspar. Colorless or light-yellow nonpleochroic epidote forms subhedral prismatic grains, small granules, and anhedral skeletal aggregates, generally less than 0.5 mm in diameter. Many epidote grains and aggregates enclose cores of orange-red or reddish-brown metamict allanite. Garnet forms irregular skeletal grains and subhedral to euhedral porphyroblasts as much as 1.5 mm in diameter, which show no evidence of rotation. The garnet is colorless in thin section and deep wine red in hand specimen. Zircon, apatite, sphene, and magnetite are ubiquitous minor accessory minerals. Sillimanite occurs as needles and fibrous aggregates in biotite in a few specimens from the area of sillimanite-grade rocks.

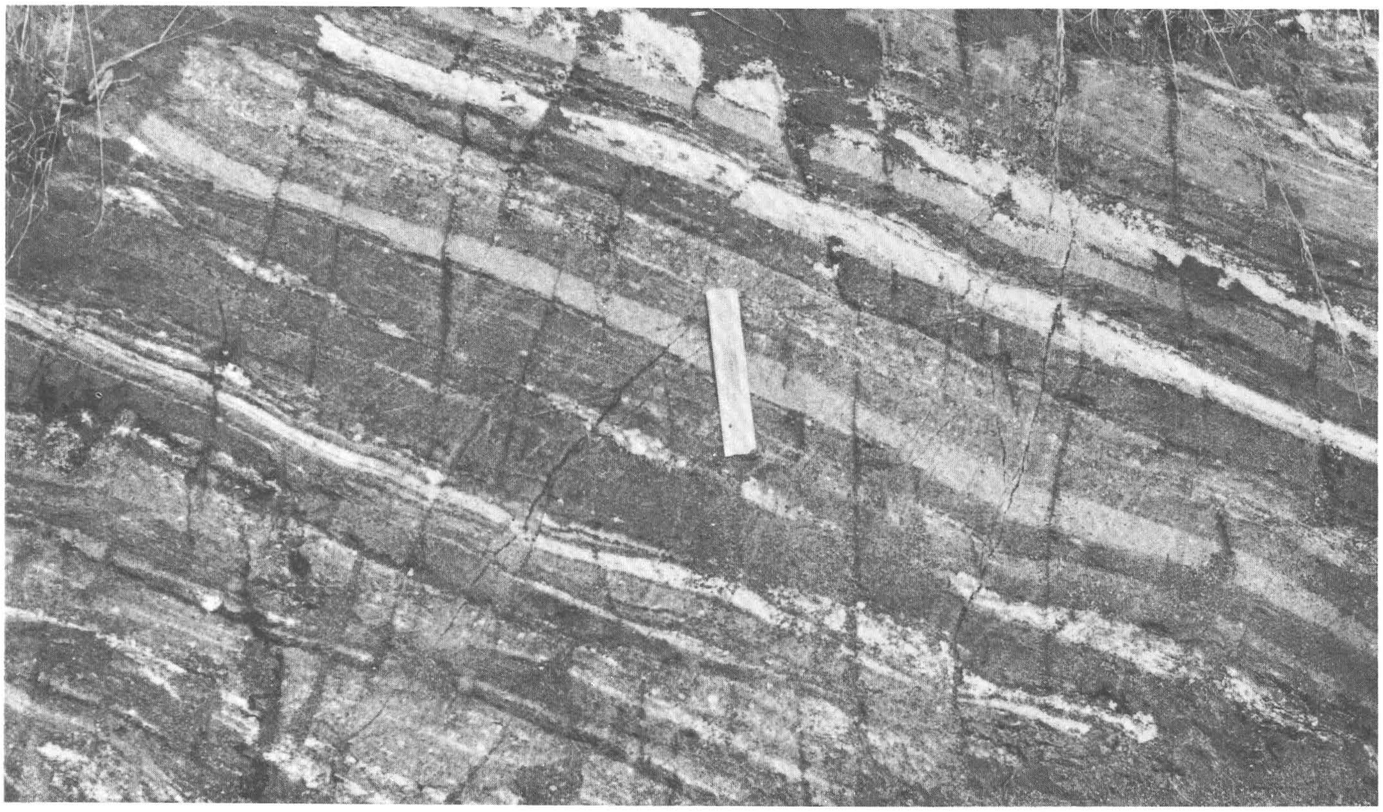


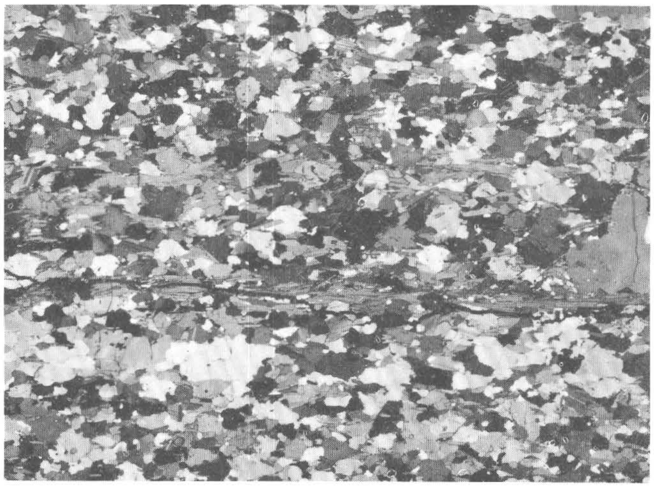
FIGURE 68.—Interlayered fine-grained biotite and hornblende-biotite gneiss containing small concordant pods of biotite pegmatite. Several layers contain scattered porphyroclasts of potassic feldspar. Saprolite exposure in roadcut on northeast side of county road 0.3 mile northwest of Abingdon (area I–7, pl. 1). Scale is about 7 inches long.

Plagioclase in the gneiss is generally about $An_{30}$, but ranges from $An_{25}$ to $An_{45}$. The smaller grains have about the same composition as the porphyroblasts. In some specimens, plagioclase grains have thin albitic rims. Plagioclase compositions in adjacent layers differ by as much as 10 percent An, showing that the variations in the composition of plagioclase are controlled more by original bulk composition of the rocks than by the metamorphic grade.

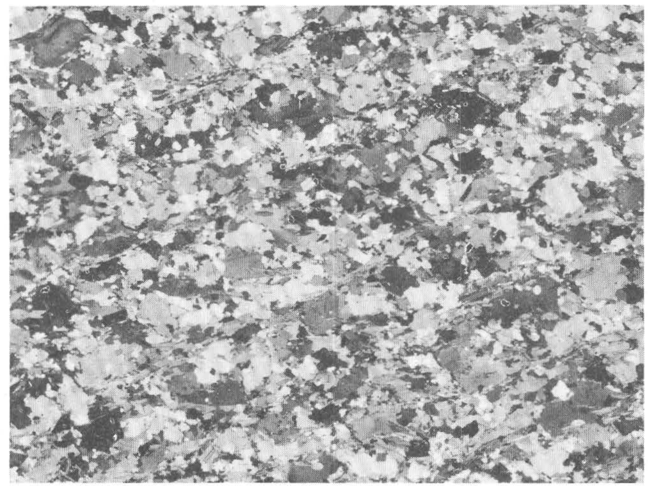
Stable mineral assemblages in the biotite gneisses that have not been affected by shearing and retrogressive metamorphism associated with the Brevard fault zone are summarized in table 19. In addition to the minerals listed in the table, zircon, apatite, sphene, and magnetite are stable accessory minerals in all assemblages.

In the belt of shearing and retrogressive metamorphism associated with the Brevard fault zone, the gneisses have cataclastic and polymetamorphic textures which become more conspicuous as the fault zone is approached (fig. 69, *B–F*). Porphyroclasts of plagioclase, potassic feldspar, muscovite, and biotite are set in a fine-grained recrystallized matrix (fig. 69 *D–F*). Where recrystallization has taken place under medium-grade conditions, the matrix consists principally of quartz, plagioclase, and potassic feldspar in an inequigranular mosaic of sutured grains 0.05 to 0.3 mm in diameter and stubby flakes and anhedral grains of synkinematic and postkinematic biotite 0.1 to 0.5 mm long randomly oriented or rudely alined parallel to the foliation. Biotite in the recrystallized matrix is generally pleochroic in shades of gray, green, or dark brown, although some is reddish brown, similar to that in the high-grade gneiss. Synkinematic muscovite forms thin flakes alined with the foliation, and epidote forms skeletal prismatic grains and irregular aggregates as much as 0.5 mm in diameter. Most of the epidote grains seem to be postkinematic, but some contain cores of orange-red or reddish-brown metamict allanite similar to those in epidote grains in the unrecrystallized high-grade gneiss. Scattered small skeletal grains of garnet also seem to be postkinematic.

Plagioclase porphyroclasts resemble the plagioclase porphyroblasts in the unrecrystallized gneiss, but they are generally small and have bent and

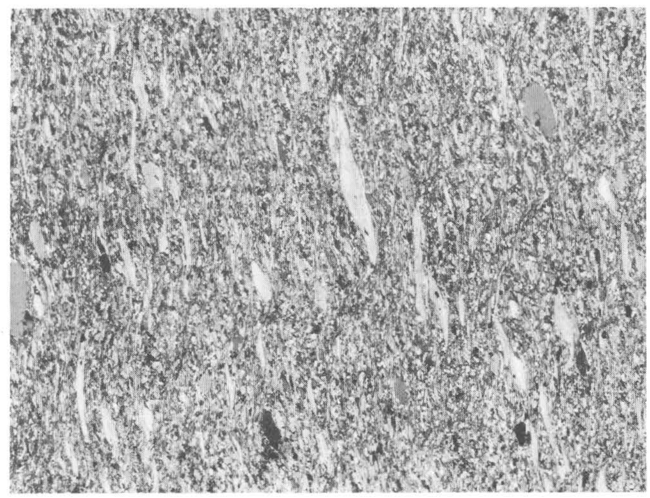


*A*

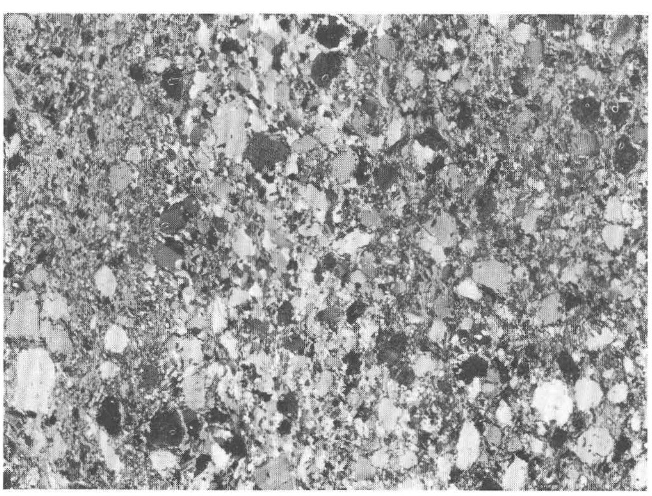


*B*

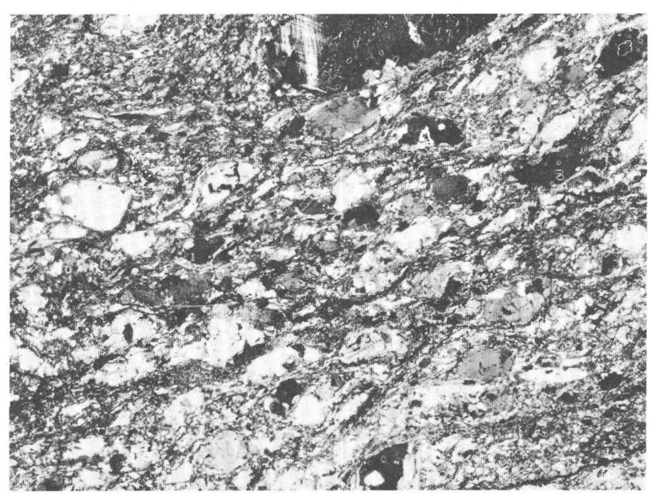
*C*



*D*



*E*



*F*

5mm

FIGURE 69.—Photomicrographs of biotite gneiss. *A*, Plagioclase-biotite-quartz gneiss from quarry beside U.S. Highway 321, 0.3 mile southeast of Caldwell County courthouse in Lenoir (area J–7, pl. 1). Dark layer in fine-grained biotite gneiss. Polymetamorphic features lacking. Plagioclase is sodic andesine. *B*, Biotite-plagioclase-quartz gneiss from roadcut 0.6 mile N. 41° W. of Abingdon (area I–7, pl. 1). Incipient development of mortar, new well recrystallized. Plagioclase is sodic andesine. Analyzed speciment 6, table 20. *C*, Biotite-muscovite–plagioclase-quartz gneiss from roadcut 0.25 mile east-northeast of Fleming Chapel (area H–7, pl. 1). Recrystallized polymetamorphic texture. Some larger muscovite flakes are porphyroclasts. Plagioclase is calcic oligoclase. *D*, Porphyroclastic muscovite-plagioclase-quartz-biotite gneiss from roadcut on east side of road 0.9 mile N. 50° E. of Arneys Store (area G–8, pl. 1). Analyzed specimen 3, table 20. *E*, Quartz-plagioclase-biotite gneiss 1.5 miles S. 57° E. of village of Happy Valley (area J–6, pl. 1). Porphyroclasts of plagioclase in a mosaic of anhedral recrystallized quartz, sodic andesine, and well-alined biotite. Analyzed specimen 2, table 20. *F*, Blastomylonitic microcline-muscovite-quartz-biotite-plagioclase gneiss from roadcut 0.4 mile east of Adako (area G–7, pl. 1). Porphyroclasts of perthitic microcline, calcic oligoclase, muscovite, biotite, and quartz in a matrix of recrystallized quartz, albite, biotite, muscovite, and epidote.

TABLE 19.—*Stable-mineral assemblages in biotite gneiss of high metamorphic grade in the Inner Piedmont belt*

[Stable accessory minerals omitted; ×, present]

| Assemblage | Quartz | Plagioclase (An content) | Biotite (color parallel to Z) | Potassic Feldspar | Muscovite | Epidote | Garnet | Sillimanite |
|---|---|---|---|---|---|---|---|---|
| 1 | × | 25 | Black | × | × | × | | |
| | × | 27 | Olive green | × | × | × | | |
| | × | 30 | Reddish brown | × | × | × | | |
| | × | 30 | Grayish brown | × | × | × | | |
| 2 | × | 27 | Grayish brown | × | | × | × | |
| | × | 25 | Greenish gray | × | | × | × | |
| 3 | × | 33 | Reddish brown | × | | × | | |
| | × | 24 | Dark brown | × | | × | | |
| 4 | × | 35 | Reddish brown | | × | × | × | |
| 5 | × | 35 | Black | | × | × | | |
| | × | 29 | Reddish brown | | × | × | | |
| 6 | × | 36 | do | | × | | × | × |
| 7 [1] | × | 30 | do | | × | | × | |
| | × | 37 | do | | × | | × | |
| | × | 35 | Dark brown | | × | | × | |
| 8 | × | 35 | Reddish brown | | | × | × | |
| | × | 35 | Dark brown | | | × | × | |
| 9 | × | 35 | Olive green | | | × | | |

[1] Typical mineral assemblage of mica schist, but also found in biotite gneiss.

broken twin lamellae and ovid outlines (fig. 69*E* and *F*). They are calcic oligoclase or sodic andesine similar in composition to the recrystallized plagioclase in the mosaic, but some have thin rims of albite. Porphyroclasts of microcline or microperthite and aggregates of angular grains apparently derived from crushed porphyroclasts are common in some of the gneiss (fig. 69*F*). The potassic feldspar porphyroclasts are generally less than 1 mm long and are commonly surrounded by a jacket of quartz and plagioclase formed by recrystallization of myrmekite. Muscovite forms conspicuous porphyroclasts as much as 2.5 mm long alined with the foliation. Some of these have warped or folded cleavages and many have cross sections shaped like a much flattened parallelogram, similar to the head of a double-bitted axe viewed parallel with the handle (fig. 69*D*). Porphyroclasts of biotite are smaller and less common than those of muscovite. They are generally dark brown and have warped cleavages and sagenitic webs of opaque minerals.

Adjacent to the Brevard fault zone, and in local zones throughout the belt of shearing and retrogression associated with it, the biotite gneiss has been recrystallized under low-grade conditions. Biotite in the recrystallized matrix has been partly or entirely altered to chlorite, plagioclase is partly altered to albite, and garnet is partly replaced by chlorite. Sericite is abundant, and tourmaline is a common accessory mineral.

Chemical analyses, modes, norms, and descriptions of selected specimens of biotite gneiss are given in table 20, together with average modes of rocks mapped as biotite gneiss in the zone of high-grade regional metamorphism and in the belt of shearing and retrogressive metamorphism associated with the Brevard fault zone. The principal modal variations in all rocks mapped as biotite gneiss are shown in figures 70 and 71.

TABLE 20.—*Chemical analyses, modes, and norms of biotite gneiss of the Inner Piedmont belt*

[Analyses of samples 2 and 7 determined by standard chemical methods by D. F. Powers, U.S. Geol. Survey. Other analyses, by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey. Modes of analysed samples, by point counts of 600 grains; P, present but not intersected in counting; Tr, trace; Nd, not determined. Major oxides and CIPW norms given in weight percent; modes, in volume percent]

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 [1] | 11 [2] | 12 [3] |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Field No. | 30–134502 | BR–71A | A–85B | 18–1680 | 33–2196B | 32–2009 | BR–71B | 30–1395D1 | 29–1148 | | | |
| Laboratory No. | 160163 | I–4058 | 16015B | 160157 | 160165 | 160164 | I–4059 | 160162 | 160160 | | | |
| **Major oxides** | | | | | | | | | | | | |
| $SiO_2$ | 56.9 | 60.81 | 66.1 | 69.0 | 69.3 | 69.9 | 71.12 | 73.8 | 79.5 | | | |
| $Al_2O_3$ | 17.6 | 17.53 | 15.8 | 15.2 | 14.0 | 14.4 | 15.24 | 15.0 | 11.4 | | | |
| $Fe_2O_3$ | .68 | 1.62 | 1.5 | 1.2 | 3.0 | 1.1 | .97 | .10 | .71 | | | |
| FeO | 7.5 | 4.22 | 4.5 | 3.5 | 2.2 | 3.0 | 1.42 | 1.2 | .27 | | | |
| MgO | 2.2 | 2.42 | 1.7 | 1.4 | 1.5 | 1.4 | .69 | .50 | .17 | | | |
| CaO | 5.0 | 4.22 | 1.1 | 1.6 | 2.8 | 2.8 | 2.35 | 2.8 | .18 | | | |
| $Na_2O$ | 3.4 | 3.84 | 2.7 | 2.6 | 3.6 | 3.6 | 3.99 | 4.7 | 3.8 | | | |
| $K_2O$ | 2.2 | 2.63 | 3.6 | 3.0 | 1.5 | 1.8 | 2.63 | .75 | 2.3 | | | |
| $H_2O+$ | .85 | .97 | 1.3 | 1.4 | .86 | .91 | .55 | .56 | .85 | | | |
| $H_2O-$ | .04 | .07 | .06 | .14 | .12 | .10 | .09 | .03 | .06 | | | |
| $TiO_2$ | 2.1 | .88 | .93 | .78 | .68 | .56 | .34 | .20 | .13 | | | |
| $P_2O_5$ | .94 | .34 | .15 | .18 | .25 | .28 | .12 | .05 | .01 | | | |
| MnO | .22 | .10 | .11 | .05 | .08 | .07 | .05 | .00 | .00 | | | |
| $CO_2$ | <.05 | .02 | <.05 | <.05 | <.05 | <.05 | .01 | .15 | <.05 | | | |
| Cl | Nd | .02 | Nd | Nd | Nd | Nd | .01 | Nd | Nd | | | |
| F | Nd | .09 | Nd | Nd | Nd | Nd | .03 | Nd | Nd | | | |
| Total | 100 | 99.78 | 100 | 100 | 100 | 100 | 99.61 | 99 | 99 | | | |
| **Modes** | | | | | | | | | | | | |
| Quartz | 23 | 25 | 32 | 38 | 42 | 38 | 32 | 37 | 51 | 35.2 | 31.6 | 32.8 |
| Plagioclase | 45 | 39 | 15 | 26 | 41 | 41 | 50 | 54 | 32 | 32.5 | 29.6 | 30.6 |
| Potassic feldspar | | .2 | .5 | 1.7 | 1.0 | .3 | 5 | 1.5 | 7 | 7.2 | 6.3 | 6.6 |
| Biotite | 25 | 28 | 36 | 20 | 12 | 18 | 8 | 5 | 1.0 | 14.3 | 17.3 | 16.3 |
| Muscovite | | 5 | 15 | 13 | 1.0 | 1.2 | 4 | 1.8 | 9 | 4.0 | 9.7 | 7.8 |
| Epidote | .3 | 2.8 | .8 | .8 | 1.2 | .8 | .5 | .2 | | 2.9 | 2.8 | 2.8 |
| Apatite | 1.2 | .2 | .2 | .5 | .2 | .3 | P | .3 | | Tr | Tr | Tr |
| Garnet | 3 | | | P | | | | | | .8 | .3 | .4 |
| Chlorite | | .2 | | | | | .7 | | | Tr | Tr | Tr |
| Zircon | | P | .2 | P | | P | | | .2 | Tr | Tr | Tr |
| Carbonate | | | | | | | | .2 | | Tr | Tr | Tr |
| Limonite | | | | .8 | | | | | 1.0 | Tr | Tr | Tr |
| Opaque minerals | 2.8 | .5 | .3 | .2 | 1.0 | | .2 | P | | Tr | Tr | Tr |
| **CIPW norms** | | | | | | | | | | | | |
| Q | 12.76 | 14.79 | 29.58 | 35.47 | 34.84 | 33.06 | 31.69 | 36.60 | 48.00 | | | |
| C | 2.79 | 1.79 | 5.82 | 5.20 | 1.97 | 2.11 | 1.95 | 1.84 | 2.36 | | | |
| Or | 13.00 | 15.54 | 21.27 | 17.72 | 8.86 | 10.64 | 15.54 | 4.43 | 13.59 | | | |
| Ab | 28.73 | 32.33 | 22.83 | 21.99 | 30.45 | 30.45 | 33.67 | 39.75 | 32.14 | | | |
| An | 18.66 | 18.04 | 4.48 | 6.76 | 12.25 | 12.06 | 10.63 | 12.61 | .83 | | | |
| Hl | | .03 | | | | | .02 | | | | | |
| En | 5.48 | 6.02 | 4.23 | 3.48 | 3.73 | 3.48 | 1.72 | 1.24 | .42 | | | |
| Fs | 10.15 | 5.14 | 5.69 | 4.24 | .59 | 3.80 | 1.34 | 1.79 | | | | |
| Mt | .99 | 2.35 | 2.18 | 1.74 | 4.35 | 1.60 | 1.41 | .14 | .49 | | | |
| Hm | | | | | | | | | .37 | | | |
| Il | 3.99 | 1.67 | 1.77 | 1.48 | 1.29 | 1.06 | .65 | .38 | .25 | | | |
| Ap | 2.23 | .80 | .36 | .43 | .59 | .66 | .28 | .12 | .02 | | | |
| Fr | | .15 | | | | | .05 | | | | | |
| Cc | | .04 | | | | | | .02 | .34 | | | |

[1] Total of 50 random grains counted in each of 44 thin sections of biotite gneisses and related rocks not showing conspicuous polymetamorphic textures.
[2] Total of 50 random grains counted in each of 91 thin sections of biotite gneiss and related rocks showing conspicuous polymetamorphic textures.
[3] Weighted average of columns 10 and 11.
NOTE.—Minor—element analyses for samples 2 and 7 given in table 1.

1. Quartz-biotite-plagioclase gneiss. Fine-grained dark-gray biotite gneiss in 2-cm layer interleaved with biotite-amphibole gneiss and quartz-plagioclase gneiss. Rock is composed of irregular interlocking grains of quartz and plagiocvlase ($An_{35}$) 0.1 to 0.5 mm across and ragged undeformed flakes of biotite as much as 0.5 mm long. Biotite is strongly alined parallel to layering. Garnet occurs in skeletal porphyroblasts as much as 5 mm in diameter. North wall of Causby quarry, on east side of Hunting Creek, 0.9 mile S. 10° W. of confluence of the Catawba River and Johns River (area I–9, pl. 1).
2. Quartz-biotite-plagioclase gneiss. Fine-grained dark-gray gneiss interlayered with quartz-feldspar gneiss and amphibolite. Rock has polymetamorphic texture and consists of ragged porphyroclasts of plagioclase as much as 6 mm in diameter, in a mosaic of anhedral recrystallized quartz and plagioclase ($An_{35}$) 0.01 to 0.5 mm in diameter and well-alined synkinematic or postkinematic biotite 0.05 to 5 mm long. Epidote occurs in irregular prismatic grains 0.1 to 1 mm long, many with allanite cores. 1.5 miles S. 57° E. of village of Happy Valley (area J–6, pl. 1).
3. Muscovite-plagioclase-quartz-biotite gneiss. Layered fine-grained dark-gray gneiss containing thin layers of white quartz-feldspar gneiss. Light-colored layers not included in analysis. Gneiss contains scattered porphyroclasts of potassic feldspar as much as 1 cm long; flakes of muscovite as much as 3 mm across lie on the foliation planes. Texture is polymetamorphic. The rock consists of a mosaic of anhedral recrystallized quartz and plagioclase ($An_{25}$) 0.02 to 0 2 mm in diameter and irregular to tabular synkinematic and postkinematic biotite 0.05 to 0.25 mm long. Porphyroclasts of untwinned plagioclase 0.25 to 0.5 mm across are set in the mosaic. Muscovite occurs as small synkinematic flakes and as rhomboid porphyroclasts 0.5 to 3 mm long. Potassic feldspar forms small irregular grains and intergranular films in the recrystallized mosaic; no porphyoclasts of potassic feldspar occur in the analyzed specimen. Epidote occurs as small skeletal crystals, some containing clinozoisite cores. Roadcut on east side of road east of Johns River, 0.9 mile N. 50° E. of Arneys Store (area G–8, pl. 1).
4. Muscovite-biotite-plagioclase-quartz gneiss. Fine-grained dark-gray schistose gneiss containing plagioclase and potassic feldspar porphyroclasts 0.5 to 2.5 cm long. Texture is polymetamorphic. Rock consists of mosaic of granoblastic quartz and plagioclase ($An_{30}$) 0.05 to 0.1 mm in diameter and rudely alined flakes of synkinematic and postkinematic biotite 0.05 to 0.2 mm long. Epidote forms irregular poikolitic grains, some containing allanite cores. Small outcrop in tributary of Canoe Creek 0.4 mile S. 10° W. of Tablerock Church on North Carolina Highway 181 (area F–9, pl. 1).

TABLE 20.—*Chemical analyses, modes, and norms of biotite gneiss of Inner Piedmont belt*—Continued.

5. Biotite-plagioclase-quartz gneiss. Dark-gray fine-grained layered gneiss. Texture is polymetamorphic. Rock is composed of mosaic of anhedral recrystallized quartz and plegioclase ($An_{30}$) 0.1 to 0.5 mm in diameter and irregular to tabular flakes of synkinematic and postkinematic biotite 0.1 to 1 mm long. A few porphyroclasts of twinned plagioclase as much as 1 mm long are set in the mosaic. Epidote occurs as irregular to prismatic grains as much as 0.25 mm long, many containing cores of allanite. From outcrop 250 feet northeast of summit of Peaked Top (area I–7, pl. 1).
6. Biotite-quartz-plagioclase gneiss. Medium-grained dark-gray gneiss containing thin layers of biotite schist and small pods of pegmatite (only the gneiss is included in the analyzed specimen). Texture not polymetamorphic. Rock consists of mosaic of anhedral inequigranular quartz 0.05 to 0.5 mm in diameter, zoned plagioclase ($An_{30-37}$) as much as 1 mm in diameter, and flakes of synkinematic biotite 0.01 to 1 mm long. Muscovite is in scattered synkinematic and postkinematic flakes, and epidote forms subhedral prisms about 0.1 mm long, commonly containing allanite cores. Roadcut on road between Abingdon and Collettsville, 0.6 mile N. 41° W. of Abingdon (area I–7, pl. 1).
7. Potassic feldspar-biotite-quartz-plagioclase gneiss. Light-gray medium-grained gneiss interlayered with dark-gray biotite gneiss (sample 2). Texture is polymetamorphic. Rock consists of mosaic of irregular grains of recrystallized quartz and plagioclase 0.02 to 0.5 mm in diameter and irregular flakes of synkinematic and postkinematic biotite 0.05 to 0.25 mm long. Abundant porphyroclasts of plagioclase ($An_{30}$) 0.25 to 1 mm in diameter, some sericitized and partly albitized. Muscovite occurs as scattered synkinematic flakes. Potassic feldspar forms small irregular grains and intergranular films in the quartz-plagioclase mosaic. Same locality as sample 2.
8. Biotite-quartz-plagioclase gneiss. Fine-grained light-gray gneiss in 3-cm layer interlayered with dark-gray biotite gneiss (sample 1) and biotite amphibole gneiss. Texture is not polymetamorphic. Rock consists of anhedral grains of plagioclase ($An_{24}$) averaging 0.5 mm in diameter, lenticular grains of quartz as much as 1 mm long elongated parallel to foliation, and scattered rudely alined stubby flakes of biotite 0.1 to 0.25 mm long. Muscovite occurs as sericite flakes and skeletal grains replacing plagioclase. Potassic feldspar forms intergranular films and perthitic blebs in plagioclase. Some plagioclase slightly albitized. Same locality as sample 1.
9. Potassic feldspar-muscovite-plagioclase gneiss. Light-gray medium- to fine-grained gneiss, well foliated but nonlayered. Texture is not conspicuously polymetamorphic. Rock consists of mosaic of recrystallized granoblastic quartz 0.02 to 0.5 mm in diameter, clear twinned plagioclase ($An_{55}$) 0.5 to 1.0 mm in diameter, and strongly alined flakes of synkinematic muscovite as much as 0.7 mm long. Potassic feldspar occurs in small irregular grains interstitial to quartz and plagioclase. Outcrop 0.6 mile S. 55° E. of Arneys Store (area G–8, pl. 1).

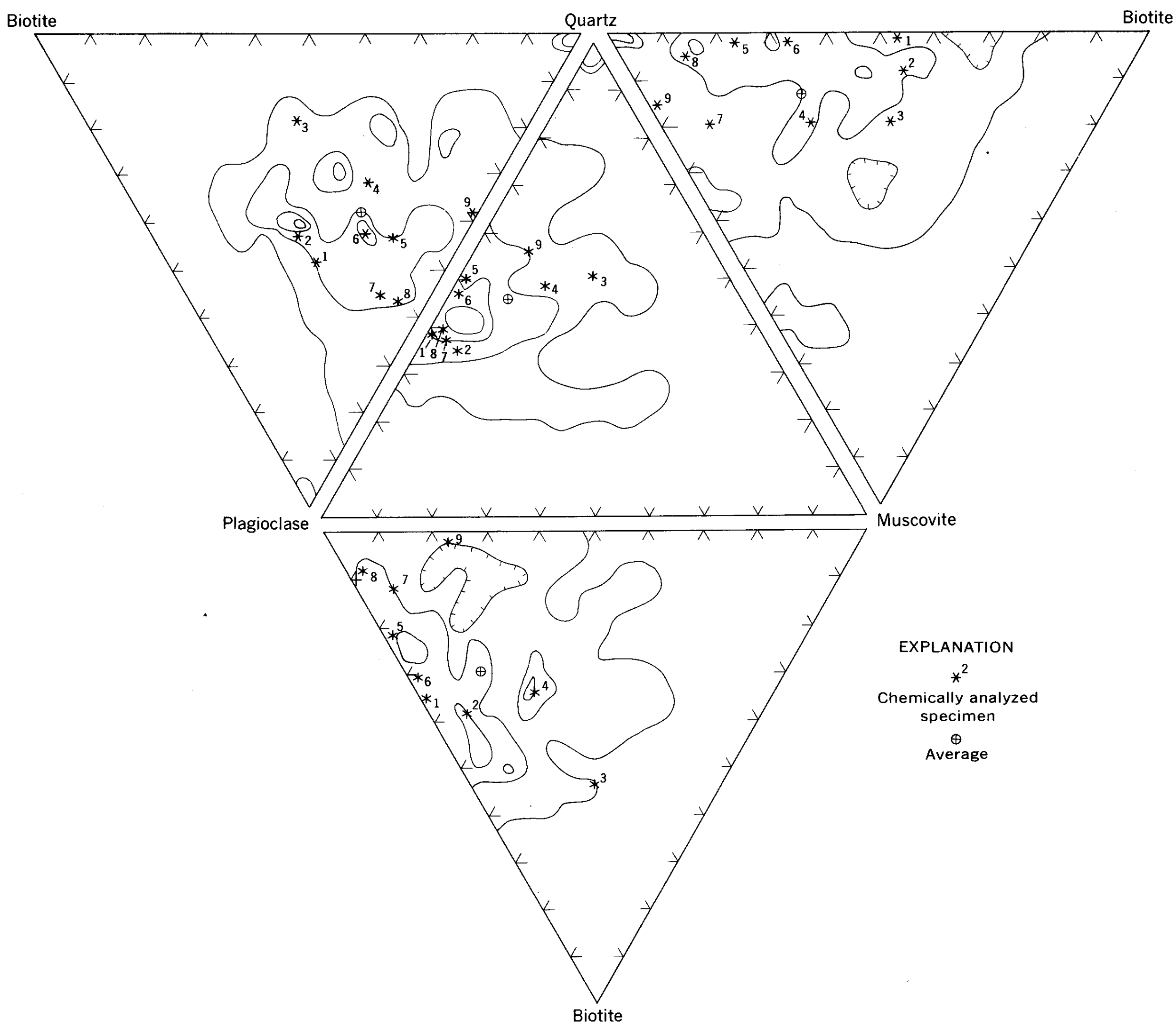



FIGURE 70.—Proportions of quartz, plagioclase, muscovite, and biotite in biotite gneisses of the Inner Piedmont belt. Based on counts of 50 random grains in each of 135 thin sections. Contours 0.75, 3, 9, and 12 percent. Numbers of analyzed specimens refer to analyses in table 20.

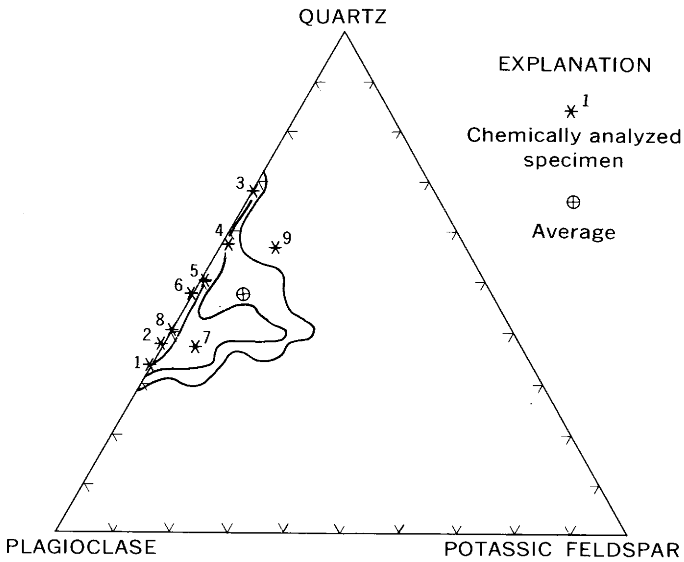



FIGURE 71.—Proportions of quartz, plagioclase, and potassic feldspar in biotite gneisses of the Inner Piedmont belt. Based on counts of 50 random grains in each of 135 thin sections. Contours 1.6, 3.2, and 4.7 percent. Numbers of analyzed specimens refer to analyses in table 20.

MICA SCHIST

Mica schist is interlayered with biotite gneiss in all proportions and locally predominates over gneiss in large enough areas to be distinguished on the geologic maps. The schist forms layers ranging from a few inches to several tens of feet in thickness and is conformable with the layering in the enclosing gneiss. The mica schist is a strongly foliated medium- to coarse-grained light-gray or greenish-gray rock composed of various proportions of biotite, muscovite, quartz, and plagioclase, and commonly containing porphyroblasts or porphyroclasts of garnet and plagioclase.

Where the schist has not been subjected to shearing and retrogression along the Brevard fault zone, it consists of folia of strongly alined synkinematic flakes of muscovite and biotite 1 to 5 mm long alternating with folia composed of irregular granoblastic grains of quartz and clear-twinned plagioclase (sodic andesine) 0.25 to 5 mm in diameter (fig. 72*A*). Garnet forms subhedral to euhedral poikilitic grains, generally less than 5 mm in diameter but locally as much as 3 inches in diameter (fig. 72*A*). Zircon, apatite, and magnetite are the most common minor accessory minerals, and rutile occurs in a few specimens. Some of the schist contains porphyroblasts of plagioclase as much as 1 cm in diameter.

The only stable mineral assemblage found in the mica schist unaffected by shearing along the Brevard zone is quartz-plagioclase-biotite-muscovite-garnet.

In the belt of shearing and recrystallization along the Brevard fault zone, the schist has conspicuously polymetamorphic textures (fig. 72*B–D*). Warped and bent books of muscovite and biotite as much as 5 mm long and grains of plagioclase as much as 5 mm in diameter having warped and broken twin lamellae form porphyroclasts set in a finely recrystallized granoblastic matrix of recrystallized quartz and plagioclase grains 0.01 to 0.1 mm in diameter. Coarser grained recrystallized quartz also forms segregation folia parallel to the foliation. Garnets contain cores having helicitic structures, commonly jacketed by clear synkinematic or postkinematic overgrowths. Scattered grains of kyanite and staurolite as much as 5 mm in diameter are fairly widespread in the polymetamorphic schist. Locally, staurolite is associated with an isotropic mineral with very light green absorption and high relief, which might be a spinel. Whether the staurolite and kyanite grains are porphyroblastic or porphryoclastic near the southeast margin of the zone of shearing and recrystallization (pl. 6*D*) is not certain, but to the northwest they are clearly porphyroclastic. Plagioclase in the polymetamorphic schist is generally more sodic than that in the schist to the southeast, and colorless epidote is present as scattered irregular grains having moderate to low birefringence. Adjacent to the Brevard fault zone, and in irregularly distributed local areas throughout the belt of shearing and metamorphism associated with it, the latest recrystallization occurred at low metamorphic grade. Plagioclase is partially or completely replaced by albite or sericite, biotite and garnet are altered to chlorite, and staurolite and kyanite are jacketed by sericite. In these areas, undeformed needles of green tourmaline are very common in the schist.

The variation in proportions of the principal minerals in all rocks mapped as mica schist is shown in the modal composition diagrams (fig. 73).

Chemical analyses, modes and norms of two specimens of schist and the average mode of all the mica schists examined are given in table 21.

HORNBLENDE GNEISS AND AMPHIBOLITE

Fine- and medium-grained medium- to dark-gray biotite-hornblende gneiss, hornblende gneiss, and amphibolite form layers and lenses intercalated with biotite gneiss. These rocks are ubiquitous, but nowhere do they predominate over biotite gneiss in a sufficiently large area to be distinguished on the map, and they are therefore mapped with the biotite gneiss. Areas where amphibolite is particularly

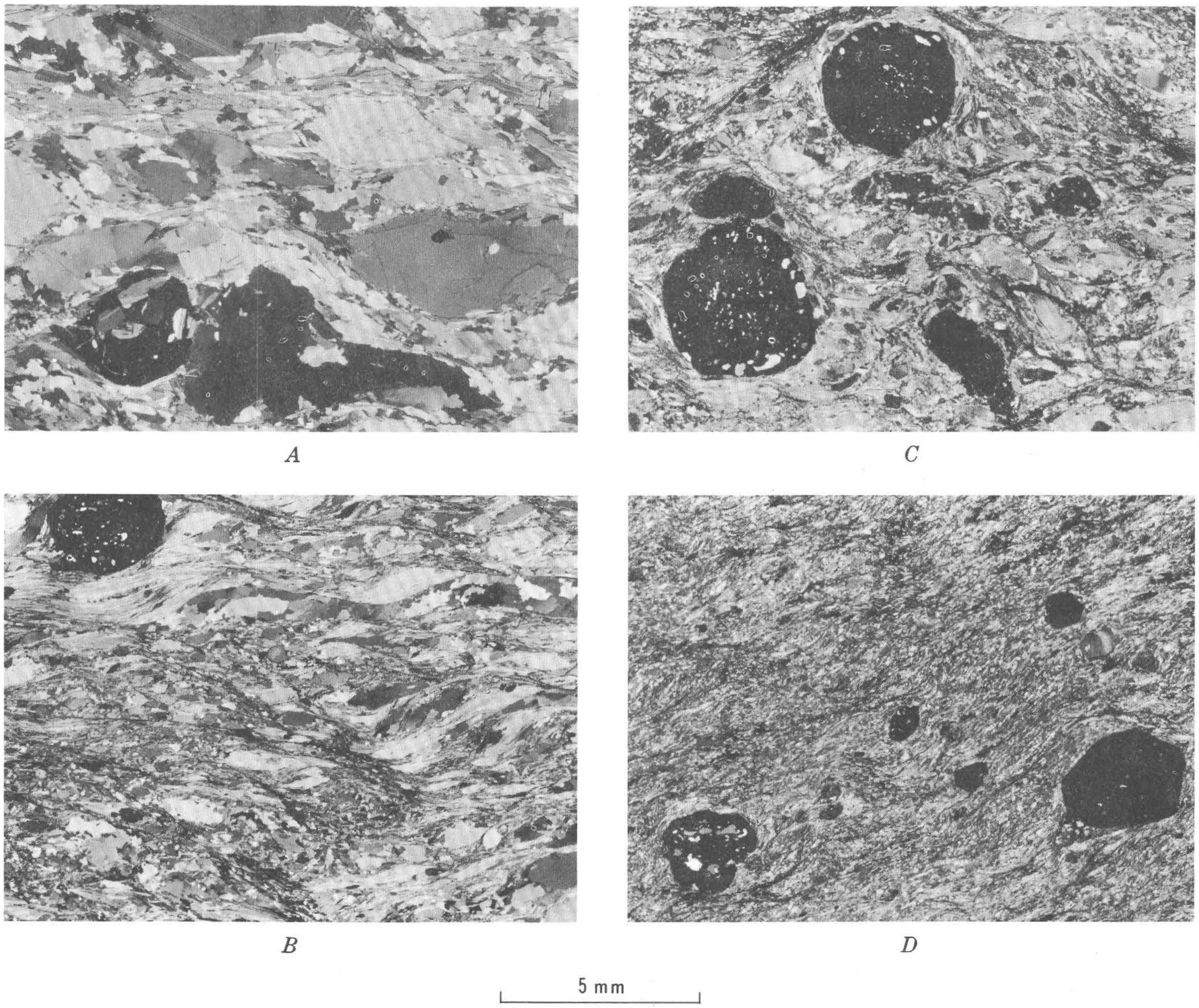



FIGURE 72.—Photomicrographs of mica schist. *A*, Coarse-grained garnet-bearing muscovite-biotite-plagioclase-quartz schist from outcrop along Lower Creek 1.6 miles southeast of Chesterfield (area H–9, pl. 1). Large grains of sodic andesine, quartz, and garnet in a matrix of quartz, sodic andesine, and synkinematic micas. *B*, Garnet-bearing plagioclase-quartz-muscovite-biotite schist from roadcut on new road from Abingdon to Collettsville, 1.0 mile N. 39° W. of Abingdon (area I–7, pl. 1). Porphyroclasts of calcic oligoclase and of bent muscovite in a matrix of recrystallized quartz, biotite, muscovite, and calcic oligoclase. Garnets with sieve texture. Quartz segregation stringer. Analyzed specimen 2, table 20. *C*, Staurolite-bearing kyanite-chlorite-biotite-garnet-quartz-muscovite schist from roadcut along North Carolina Highway 126, 0.9 mile east of bridge over Linville River (area E–9, pl. 1). Porphyroclasts of sieve-textured garnet, staurolite, kyanite, and bent muscovite in a matrix of recrystallized quartz, muscovite, and biotite. Kyanite partly altered to sericite, and biotite partly altered to chlorite. *D*, Kyanite- and staurolite-bearing garnet-chlorite-quartz-plagioclase-biotite-muscovite schist from roadcut on North Carolina Highway 181, 1.1 miles southeast of Smyrna Church (area F–8, pl. 1). Porphyroclasts of staurolite and sieve-textured garnet in a matrix of recrystallized quartz, calcic oligoclase, biotite, and muscovite. Local alteration of staurolite to sericite and of garnet to chlorite.

abundant are indicated by an overprint on plate 1. Layers of hornblende-bearing rocks range from a few inches to several feet in thickness, and contacts with the biotite gneiss are sharp. Where the biotite gneiss has been sheared, hornblende gneiss and amphibolite layers are commonly pulled apart into disconnected pods and lenses, some of which have been rotated, so that layering within them stands at high angles to layering in the enclosing rocks.

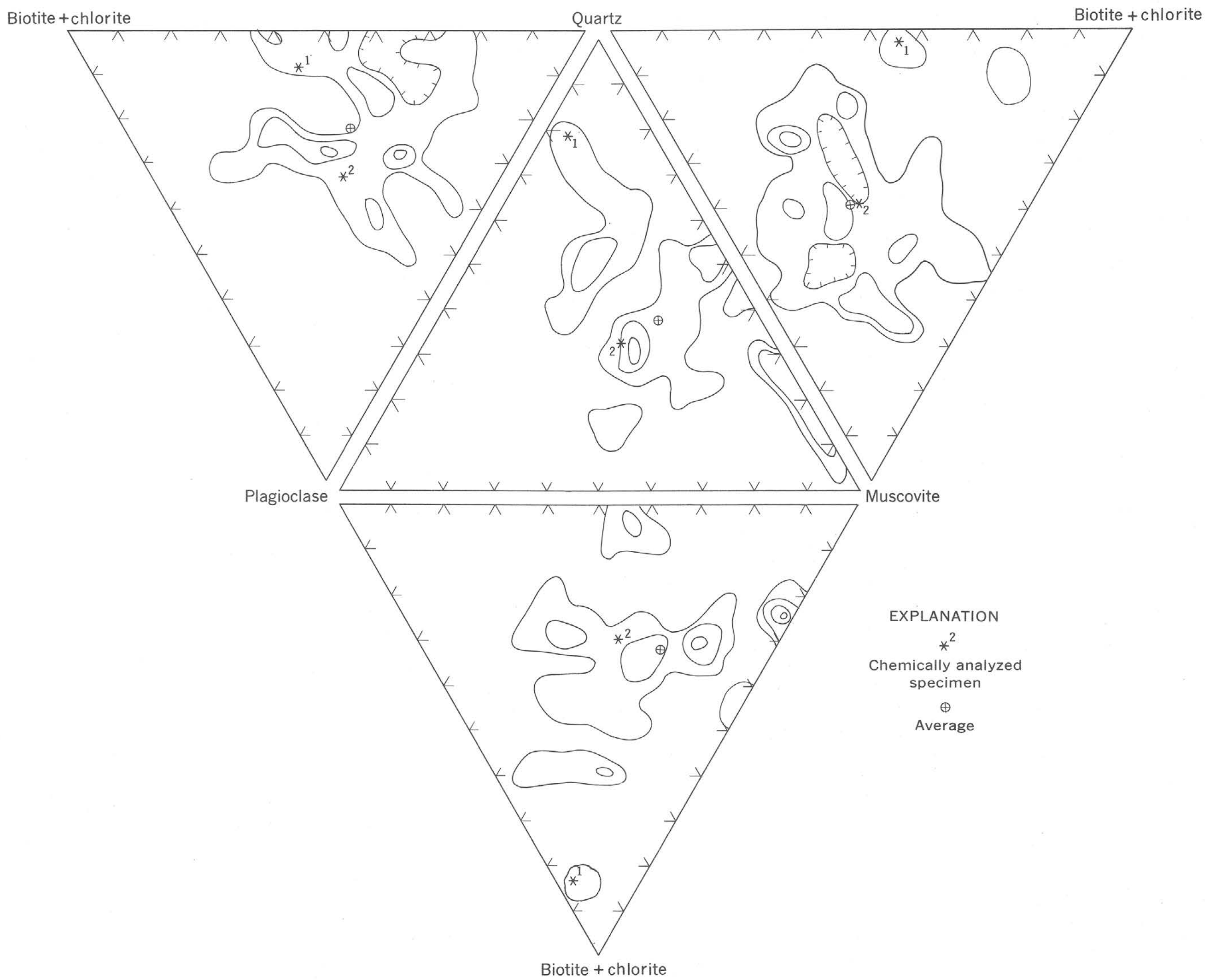



**FIGURE 73.—Proportions of quartz, plagioclase, muscovite, and biotite plus chlorite in mica schist of the Inner Piedmont belt. Based on counts of 50 random grains in each of 35 thin sections. Contours 3, 6, 12, and 17 percent. Numbers of analyzed specimens refer to analyses in table 21.**

The hornblende rocks are granoblastic or rudely foliated or lineated, depending on the relative proportions of biotite and hornblende. Commonly the hornblende-bearing gneisses and amphibolites do not have recognizable polymetamorphic textures, even in the belt of shearing and retrogression along the Brevard fault zone where polymetamorphic textures are conspicuous in the enclosing rocks. Apparently hornblende-rich rocks were resistant to physical and chemical breakdown under the metamorphic conditions prevailing in that belt.

The biotite-hornblende gneisses consist of various proportions of quartz, plagioclase, bornblende, and biotite. Anhedral grains of quartz and plagioclase 0.25 to 2.0 mm in diameter form an inequigranular mosaic interstitial to randomly arrayed or rudely alined stubby flakes of biotite as much as 2.0 mm long and irregular to subhedral poikilitic grains of hornblende 0.25 to 3.0 mm in diameter. Much of the plagioclase is faintly zoned; it is generally calcic oligoclase or andesine. In some specimens, plagioclase forms porphyroblasts as much as 3.0 mm in diameter. Epidote occurs as prismatic grains and irregular skeletal aggregates. It is generally colorless, but some is light yellow and fainly pleochroic. Some grains contain cores of allanite; others contain cores of clinozoisite. Apatite, sphene, and magnetite are

TABLE 21.—*Chemical analyses, modes, and norms of schists in the Inner Piedmont belt*

[Chemical analyses determined by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey. Modes, by point counts of 600 grains; P, present but not intersected in counting; Tr, trace. Brackets indicate chlorite replacing biotite and garnet. Major oxides and CIPW norms given in weight percent; modes, in volume percent.]

| | 1 | 2 | 3 [1] | 4 [2] |
|---|---|---|---|---|
| Field No | 18–1652 | 32–2012A | | |
| Laboratory No | 160141 | 160142 | | |
| **Major oxides** | | | | |
| $SiO_2$ | 58.9 | 59.0 | | |
| $Al_2O_3$ | 13.0 | 17.9 | | |
| $Fe_2O_3$ | 4.9 | 1.9 | | |
| FeO | 4.4 | 6.9 | | |
| MgO | 6.7 | 2.4 | | |
| CaO | .40 | 2.1 | | |
| $Na_2O$ | .58 | 2.8 | | |
| $K_2O$ | 3.6 | 3.0 | | |
| $H_2O+$ | 1.9 | 1.2 | | |
| $H_2O-$ | .09 | .04 | | |
| $TiO_2$ | 1.2 | 1.1 | | |
| $P_2O_5$ | .25 | .48 | | |
| MnO | .40 | .11 | | |
| $CO_2$ | .05 | .05 | | |
| Aqua regia soluble sulfur as S | 2.3 | | | |
| Total | 99 | 99 | | |
| **Modes** | | | | |
| Quartz | 36 | 23 | 26.9 | 69.8 |
| Plagioclase | 7 | 21 | 13.9 | 2.8 |
| Potassic feldspar | | | .4 | |
| Biotite | 46 | 20 | 20.9 | 4.2 |
| Muscovite | 2.3 | 26 | 29.7 | 18.2 |
| Aluminum silicates | | | .4 | 1.1 |
| Staurolite | 2.2 | | | |
| Epidote | | 1.0 | .5 | |
| Garnet | 2.7 | 4 | 3.3 | 1.2 |
| Chlorite | .8 | 1.8 | [5.2·] | |
| Zircon | P | P | Tr | Tr |
| Apatite | .1 | 1.0 | Tr | Tr |
| Tourmaline | | P | Tr | |
| Rutile | .3 | | Tr | |
| Opaque minerals | 2.7 | 2.8 | 3.0 | 1.7 |
| **CIPW norms** | | | | |
| Q | 31.76 | 20.34 | | |
| C | 8.14 | 7.50 | | |
| Or | 21.27 | 17.72 | | |
| Ab | 4.90 | 23.68 | | |
| An | .04 | 6.96 | | |
| En | 16.68 | 5.98 | | |
| Fs | | 9.49 | | |
| Mt | 3.70 | 2.76 | | |
| Hm | 2.34 | | | |
| Il | 2.28 | 2.09 | | |
| Ap | .59 | 1.14 | | |
| Pr | 4.30 | | | |
| Cc | .11 | .11 | | |

[1] Average of 50 random grains counted in each of 35 thin sections of mica schists.
[2] Average of 50 random grains counted in each of 13 thin sections of quartz schists.

1. Quartz-biotite schist. Medium-grained dark-gray biotite schist. Rock consists of well-alined flakes of reddish-brown synkinematic biotite 1 to 2 mm long, interleaved with folia of granoblastic quartz 0.05 to 1 mm in diameter. Muscovite occurs as scattered flakes alined with biotite, and plagioclase ($An_{30}$), as small clear grains in the quartz mosaic. Garnet and staurolite form large synkinematic or postkinematic poikilitic grains. Opaque grains are chiefly pyrite. Outcrop 0.8 mile N. 47° W. of school at Oak Hill (area F–9, pl. 1).

TABLE 21.—*Chemical analyses, modes, and norms of schists in the Inner Piedmont belt*—Continued

2. Biotite-plagioclase-quartz-muscovite schist. Lustrous fine-grained biotite-muscovite schist containing porphyroclasts of bent muscovite interlayered with dark-gray biotite and biotite-amphibole gneiss. Texture is polymetamorphic. The rock consists of a mosaic of recrystallized granoplastic quartz and plagioclase 0.05 to 0.2 mm in diameter and of irregular alined synkinematic and postkinematic flakes of reddish-brown biotite 0.05 to 0.5 mm long; the rock also contains porphyroclasts of plagioclase ($An_{23-28}$) 0.25 to 2 mm in diameter and of muscovite as much as 5 mm long. Garnet occurs in subhedral porphyroblasts as much as 5 mm in diameter, containing poikilitic inclusions of quartz and plagioclase. Roadcut on new road from Abingdon to Collettsville 1.0 mile N. 39° W. of Abingdon (area I–7, pl. 1).

common stable accessory minerals, and zircon occurs in some specimens. Chlorite, sericite, and carbonate are products of incipient alteration in many specimens.

Hornblende gneiss and amphibolite are composed principally of various proportions of plagioclase and hornblende and commonly contain smaller amounts of quartz and epidote. We refer to rocks containing less than 50 percent hornblende as hornblende gneisses and to those containing more than 50 percent as amphibolites.

These rocks consist of mosaics of rudely alined irregular poikilitic grains of hornblende 0.2 to 2.0 mm long, anhedral grains of plagioclase 0.2 to 0.5 mm in diameter, and smaller irregular grains of quartz that are interstitial to plagioclase and that also occur as poikilitic inclusions in hornblende. Plagioclase is generally andesine and is well twinned and commonly faintly zoned. Hornblende is pleochroic in shades of green and blue green. Some amphibolites contain scattered anhedral grains of colorless or light-green diopside, and many of them contain scattered small flakes of reddish-brown biotite. Colorless or pale-yellow epidote forms skeletal grains and irregular aggregates, generally less than 0.2 mm in diameter. A few specimens contain euhedral to subhedral grains of garnet 0.05 to 0.2 mm in diameter, concentrated along foliation planes. Sphene is a ubiquitous accessory mineral; it forms anhedral to subhedral grains as much as 0.5 mm in diameter and aggregates of smaller grains, generally concentrated along foliation planes. Apatite, zircon, and magnetite are other common accessories.

Although polymetamorphic textures are not conspicuous, hornblende gneiss and amphibolite in the zone of shearing and recrystallization associated with the Brevard fault zone are generally finer grained and more schistose and contain more epidote and a more sodic plagioclase than similar rocks to the southeast. Hornblende is partly altered to chlorite or to actinolite.

Stable mineral assemblages found in hornblende-bearing rocks are summarized in table 22. Compositions, modes, and norms of selected specimens of am-

TABLE 22.—*Stable-mineral assemblages in hornblende gneisses and amphibolites in the Inner Piedmont belt*

[Stable accessory minerals omitted; ×, present. Assemblages 1 and 2 are in biotite-hornblende gneiss; assemblages 3 through 9, in hornblende gneiss and amphibolite]

| Assemblage | Quartz | Plagioclase (An content) | Hornblende (color parallel to Z) | Biotite (color parallel to Z) | Epidote | Garnet | Diopside |
|---|---|---|---|---|---|---|---|
| 1 | × | 30 to 35 | Blue green | Dark brown | × | | |
| | × | 45 | Light gray green | Reddish brown | × | | |
| 2 | × | 38 | Gray green | Reddish brown | | | |
| | × | 45 | do | Reddish brown | | | |
| | × | 30 | Dark gray green | Dark chestnut brown | | | |
| | × | 90 | Light gray | Red brown | | | |
| 3 | | 35 to 45 | Blue green | Medium brown | | | |
| | | 50 to 60 | Light gray green | Red brown | | | |
| 4 | × | 40 | Blue green | | × | | |
| 5 | | 40 to 45 | Light gray green | | × | | |
| 6 | × | 35 | Gray green | | | × | |
| 7 | × | 40 to 45 | do | | | × | × |
| 8 | × | 45 to 55 | Light gray green | | × | | × |
| | × | 80 to 85 | do | | × | | × |
| 9 | × | 90 | Light gray | Red brown | | | |

TABLE 23.—*Chemical analyses, modes, and norms of biotite-hornblende gneiss and amphibolite in the Inner Piedmont belt*

[Chemical analyses determined by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey. Modes of analyzed samples determined by point count of 600 points; P, present but not intersected in counting; Tr, trace. Major oxides and CIPW norms given in weight percent; modes, in volume percent]

| | Amphibolites | | | Biotite-hornblende gneiss | | | | | Averages | | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 [1] | 10 [2] | 11 [3] |
| Field No. | 28–1111D | 29–1343C | 29–1218 | 28–1126 | 30–1379A | 30–1669A | 30–1395A | 33–2146A | | | |
| Laboratory No. | 160138 | 160140 | 160139 | 160159 | 161250 | 161251 | 160161 | 160156 | | | |
| **Major oxides** | | | | | | | | | | | |
| $SiO_2$ | 44.2 | 52.6 | 53.1 | 38.1 | 52.6 | 52.9 | 57.0 | 60.9 | | | |
| $Al_2O_3$ | 17.2 | 15.1 | 12.5 | 15.4 | 18.1 | 16.4 | 16.6 | 15.3 | | | |
| $Fe_2O_3$ | 2.3 | 1.2 | 2.5 | 5.1 | .90 | 1.8 | 1.0 | 2.6 | | | |
| FeO | 11.2 | 8.0 | 5.8 | 10.6 | 7.0 | 8.4 | 5.9 | 4.6 | | | |
| MgO | 7.4 | 7.8 | 10.8 | 9.8 | 5.1 | 3.6 | 5.3 | 3.6 | | | |
| CaO | 12.0 | 11.0 | 11.9 | 10.5 | 8.3 | 7.6 | 7.2 | 6.7 | | | |
| $Na_2O$ | 1.7 | .87 | .98 | .94 | 3.3 | 3.0 | 3.0 | 3.2 | | | |
| $K_2O$ | .89 | .90 | .37 | 1.9 | 1.6 | 2.2 | 1.6 | 1.3 | | | |
| $H_2O+$ | .71 | .75 | .65 | 1.4 | 1.0 | 1.2 | .87 | .92 | | | |
| $H_2O-$ | .04 | .04 | .11 | .22 | .07 | .10 | .05 | .04 | | | |
| $TiO_2$ | 1.0 | .71 | .24 | 2.2 | 1.0 | 2.5 | .84 | .76 | | | |
| $P_2O_5$ | .19 | .13 | .04 | 2.3 | .46 | .37 | .18 | .22 | | | |
| MnO | .23 | .19 | .16 | .14 | .16 | .18 | .11 | .13 | | | |
| $CO_2$ | .05 | .05 | .05 | .05 | .05 | .05 | .12 | .05 | | | |
| Total | 99 | 99 | 99 | 99 | 100 | 100 | 100 | 100 | | | |
| **Modes** | | | | | | | | | | | |
| Quartz | | 10 | 10 | | 4 | 14 | 13 | 19 | 3.1 | 5.8 | 19.1 |
| Plagioclase | 20 | 25 | 13 | 21 | 35 | 36 | 42 | 40 | 23.8 | 12.2 | 38.5 |
| Biotite | .5 | 5 | | 18 | 20 | 25 | 21 | 14 | .7 | 1.0 | 15.3 |
| Muscovite | .7 | | | | | | .2 | 4 | .1 | 1.2 | Tr |
| Hornblende | 77 | 60 | 73 | 36 | 34 | 22 | 22 | 18 | 64.9 | 55.8 | 22.2 |
| Pyroxene | | | .8 | 18 | | | | | .5 | Tr | |
| Epidote | .5 | | 2.2 | .5 | | | | 2.5 | 2.2 | 19.8 | 1.1 |
| Apatite | .1 | .3 | .3 | | | .3 | .3 | .7 | | | |
| Chlorite | .1 | | | | .1 | | .8 | 1.5 | 1.5 | .8 | |
| Zircon | | P | P | | | | | | Tr | Tr | Tr |
| Carbonate | | | | | | | .3 | | Tr | | |
| Sphene | | P | | | 8.3(?) | 4 | 1.1 | .5 | .9 | 1.0 | 1.2 |
| Opaque minerals | 1.5 | .6 | | 4 | .1 | | .7 | .2 | .9 | Tr | Tr |
| **CIPW norms** | | | | | | | | | | | |
| Q | | 7.61 | 7.24 | | .18 | 4.16 | 7.74 | 17.68 | | | |
| Or | 5.26 | 5.32 | 2.19 | 11.22 | 9.45 | 13.00 | 9.45 | 7.68 | | | |
| Ab | 11.54 | 7.36 | 8.29 | 7.92 | 27.91 | 25.37 | 25.15 | 27.06 | | | |
| An | 36.67 | 34.64 | 28.62 | 32.19 | 29.86 | 24.79 | 27.24 | 23.55 | | | |
| Ne | 1.54 | | | .01 | | | | | | | |
| Wo | 8.89 | 7.83 | 12.45 | 1.90 | 3.34 | 4.25 | 2.90 | 3.31 | | | |
| En | 4.46 | 19.42 | 26.89 | 1.20 | 12.70 | 8.96 | 14.30 | 8.96 | | | |
| Fs | 4.23 | 12.88 | 8.48 | .58 | 10.76 | 10.14 | 8.85 | 5.28 | | | |
| Fo | 9.78 | | | 16.26 | | | | | | | |
| Fa | 10.20 | | | 8.72 | | | | | | | |
| Mt | 3.34 | 1.74 | 3.62 | 7.40 | 1.30 | 2.61 | 1.39 | 3.77 | | | |
| Il | 1.90 | 1.35 | .46 | 4.18 | 1.90 | 4.75 | 1.60 | 1.44 | | | |
| Ap | .45 | .31 | .10 | 5.45 | 1.09 | .88 | .34 | .52 | | | |
| Cc | .11 | .11 | .11 | .11 | .11 | .11 | .30 | .11 | | | |

TABLE 23.—*Chemical analyses, modes, and norms of biotite-hornblende gneiss and amphibolite in the Inner Piedmont belt*—Continued

[1] Averages of 50 random grains counted in each of 16 thin sections of amphibolites not affected by shearing along the Brevard fault zone.
[2] Average of 50 random grains counted in each of nine thin sections of amphibolites from zone of rocks affected by shearing along the Brevard fault zone.
[3] Average of 50 random grains counted in each of 23 thin sections of biotite-hornblende gneisses.

1. Plagioclase amphibolite. Medium-grained dark-gray amphibolite in folded layer in medium-grained migmatitic biotite gneiss containing granitic layers and streaks. Amphibolite is weakly foliated but has conspicuous lineation of alined hornblende prisms. Rock is unsheared; it consists of subhedral to anhedral grains of green hornblende 0.5 to 1.5 mm long and anhedral grains of twinned and zoned plagioclase ($An_{50-60}$) 0.25 to 1 mm long, some with warped twin lamellae. Biotite is reddish brown and occurs in irregular flakes as much as 1 mm long. North wall of abandoned quarry on east side of Hunting Creek 0.9 mile S. 12° W. of confluence of Johns River and the Catawba River (area H–9, pl. 1).
2. Quartz-plagioclase amphibolite. Medium-grained dark-gray amphibolite in pod 3 feet long in streaky biotite quartz monzonite containing thin layers of biotite schist and gneiss. Moderately strong foliation and lineation in amphibolite is due to alinement of hornblende; 1- to 2-mm flakes of biotite are conspicuous on foliation surfaces. Rock consists of rudely alined irregular poikilitic grains of green hornblende 0.25 to 1 mm long and irregular well-twinned and faintly zoned plagioclase (about $An_{90}$), both containing small blebs of quartz. Quartz also occurs as small irregular interstitial grains less than 0.25 mm in diameter. Thin undeformed flakes of reddish-brown biotite are alined with the foliation and seem to have crystallized with the amphibole and plagioclase. Roadcut on north side of U.S. Highway 70 (business route), 100 feet east of the junction of the U.S. Highway 64–70 bypass around Morganton (area H–9, pl. 1).
3. Quartz-plagioclase amphibolite. Medium-grained amphibolite in 2-foot-long pod in saprolitized layered biotite gneiss containing abundant pods and layers of medium- or coarse-grained biotite quartz diorite. Amphibolite is well foliated and weakly lineated because of alinement of hornblende. Rock is unsheared. It consists of well-alined poikilitic prismatic grains of green hornblende as much as 3 mm long, irregular grains of strained quartz, and twinned and zoned plagioclase ($An_{70-75}$) as much as 2 mm in diameter that is interstitial to hornblende and occurs as inclusions in hornblende. A few irregular grains of a colorless pyroxene, probably diopside, are intergrown with hornblende and seem to have crystallized simultaneously with it. Roadcut in area H–8 (pl. 1), 1.4 miles N. 40° W. of the school at Chesterfield (area H–9, pl. 4).
4. Pyroxene-biotite-hornblende gneiss. Dark-gray medium-grained nonlayered rudely foliated gneiss from isolated homogeneous outcrop. Probably a layer in layered biotite gneiss. Rock is unsheared and is composed of clear irregular grains of twinned plagioclase ($An_{82}$) 0.25 to 1.5 mm in diameter, irregular poikilitic grains of green hornblende 0.5 to 1.5 mm in diameter, stubby flakes of reddish-brown biotite 0.25 to 0.75 mm long, and irregular equidimensional grains of diopside 0.25 to 2 mm in diameter. Diopside is jacketed and partially replaced by hornblende. Magnetite occurs as small irregular grains, chiefly in hornblende; apatite forms subhedral prisms and round grains included in amphibole and biotite. Small outcrop north of county road, 0.3 mile N. 64° W. of bench mark 1013 at north end of bridge on which North Carolina Highway 18 crosses Johns River (area H–9, pl. 1).
5. Plagioclase-biotite-hornblende gneiss. Medium-grained dark-gray biotite-amphibole gneiss in 10-foot layer in migmatitic biotite gneiss. Rock is unsheared and consists of interlocking irregular grains of twinned and faintly zoned plagioclase ($An_{35-40}$) and rudely alined poikilitic grayish-green hornblende, both 0.25 to 1 mm in diameter. Biotite is reddish brown and occurs in irregular undeformed flakes 0.25 to 2.5 mm long, alined with the foliation. Sphene occurs in round or wedge-shaped grains 0.05 to 0.2 mm across, forming clusters as much as 1 mm across. More sphene is recorded in the modal analysis than is indicated by the $TiO_2$ in the chemical analysis. Roadcut on west side of road north of Huffman Bridge over Rhodiss Lake, 0.4 mile N. 20° E. of south end of bridge (west edge of area I–9, pl. 1).
6. Quartz-hornblende-biotite-plagioclase gneiss. Medium-grained dark-gray well-foliated gneiss containing abundant pods as much as 5 feet long of light-colored granitic rock. Rock is unsheared. It consists of well-alined irregular stubby grains of undeformed brown biotite and anhedral to subhedral green hornblende 0.25 to 1.0 mm long, anhedral grains of quartz 0.25 to 1.0 mm in diameter, and twinned and faintly zoned plagioclase ($An_{35}$) 0.25 to 1.5 mm in diameter. Sphene occurs in anhedral or wedge-shaped grains, some in clusters as much as 0.5 mm in diameter. Roadcut on north side of road, just north of Smoky Creek, 1.5 miles N. 52° E. of south end of Huffman Bridge (area I–9, pl. 1).
7. Quartz-biotite-hornblende-plagioclase gneiss. Dark-greenish-gray coarse-grained well-foliated nonlayered gneiss. Rock is unsheared. It consists of well-alined flakes of brown biotite as much as 2 mm long, irregular small prisms and sieve-textured grains of gray-green hornblende as much as 1 mm long, and folia of irregular granoblastic quartz and twinned and faintly zoned plagioclase ($An_{45}$), averaging 0.25 to 0.5 mm in diameter. South end of east wall of Causby quarry, 0.9 mile S. 10° W. of confluence of Catawba River and Johns River (area H–9, pl. 1).
8. Biotite-quartz-hornblende-plagioclase gneiss. Well-foliated medium- to coarse-grained gneiss forming streak in migmatite biotite quartz diorite (samples 4 and 7, table 28). Rock is unsheared. It consists of large, very irregular grains of faintly zoned plagioclase ($An_{35}$) as much as 6 mm in diameter, many having albite fringes and some being partly sericitized and saussuritized. Quartz forms irregular grains 0.1 to 1 mm in diameter interstitial to plagioclase. Brown biotite occurs as irregular poorly alined flakes as much as 3 mm long. Some partly chloritized green hornblende forms irregular poikilitic grains 3 mm in diameter containing blebs of quartz and plagioclase. Small irregular grains of colorless epidote are included in biotite and partially replace plagioclase. Roadcut on east side of North Carolina Highway 18 bypass west of Lenoir, 1.1 miles S. 68° W. of Caldwell County courthouse in Lenoir (area J–7, pl. 1).

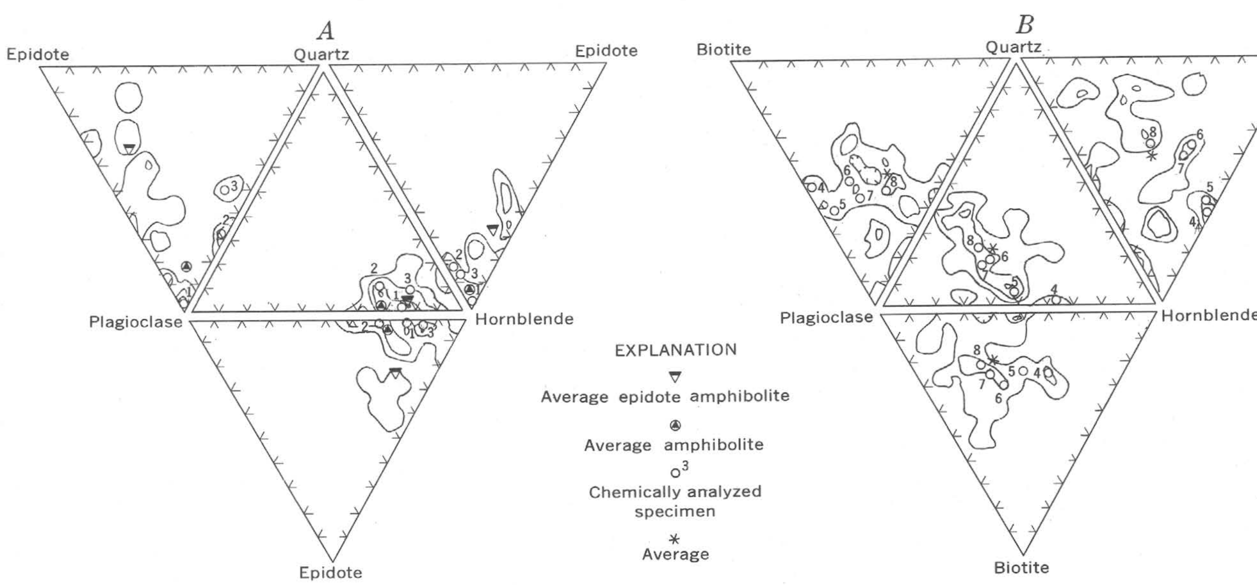



FIGURE 74.—Model composition of hornblendic rocks of the Inner Piedmont belt. Numbers of analyzed specimens refer to analyses in table 23. *A*, Proportions of quartz, plagioclase, hornblende, and epidote in amphibolites. Based on counts of 50 random grains in each of 25 thin sections. Contours 4, 8, 17, and 33 percent. *B*, Proportions of quartz, plagioclase, hornblende and biotite in biotite-hornblende gneisses. Based on point counts of analyzed specimens and counts of 50 random grains in each of 22 other thin sections. Contours 4, 7, and 15 percent.

phibolite and biotite-hornblende gneiss and average modal compositions of these rocks are given in table 23. Modal composition diagrams showing range in proportions of the principal minerals in the biotite-hornblende gneiss and in amphibolite are given in figure 74.

### SILLIMANITE SCHIST

Sillimanite schist interlayered with sillimanite-bearing quartzite and containing a few pods and layers of calc-silicate rocks forms an extensive map unit in the southeastern part of the Lenoir quadrangle. Similar sillimanite schist occurs in layers and lenses interleaved with biotite gneiss northwest of the main body. The sillimanite schist is a strongly foliated lustrous rock composed principally of sillimanite, muscovite, biotite, and quartz. It is light to medium gray and commonly has a faint purple cast on slightly weathered surfaces. Aggregates of fine sillimanite needles give the foliation surfaces a silky luster. Porphyroblasts of muscovite, biotite, and garnet are conspicuous in most hand specimens, and lenses, knots, and thin folia of quartz and feldspar are ubiquitous. In many outcrops, foliation in the schist is corrugated by chevron folds, having amplitudes of as much as an inch. Locally, sillimanite needles are rudely alined with the axes of these folds, but elsewhere sillimanite is randomly arrayed on the foliation surfaces and is wrapped around the fold axes. Much of the sillimanite schist is partially or completely altered to lustrous silvery chlorite-sericite schist or phyllite in which sericite pseudomorphs of sillimanite are recognizable. In these rocks, prisms of black tourmaline as much as 1 cm long are abundant.

The unaltered sillimanite schist consists of a deformed mosaic of intergrown dark-brown to black biotite flakes as much as 1.0 mm long and sillimanite needles and fibrolite aggregates, some of which partially replace biotite. The biotite and sillimanite are bent around noses of small chevron folds visible in hand specimen. Undeformed flakes of reddish-brown biotite as much as 1.0 mm in diameter and skeletal crystals of muscovite as much as 5 mm long form scattered porphyroblasts in the older deformed biotite-sillimanite mosaic (fig. 75*A*). Skeletal grains and aggregates of garnet, some as much as 3 cm in diameter, and a few large undeformed prisms of sillimanite also seem to be postkinetic. Both muscovite and garnet contain unrotated helicitic trains of fine sillimanite needles which have the pattern of the chevron folds. Grains of anhedral quartz 0.25 to 1.0 mm in diameter form segregation folia interleaved with the old biotite and sillimanite. These folia are wrapped around the noses of the chevron folds, but it is difficult to tell whether they formed during or after the folding. The quartz folia commonly contain scattered flakes of postkinematic biotite and a few

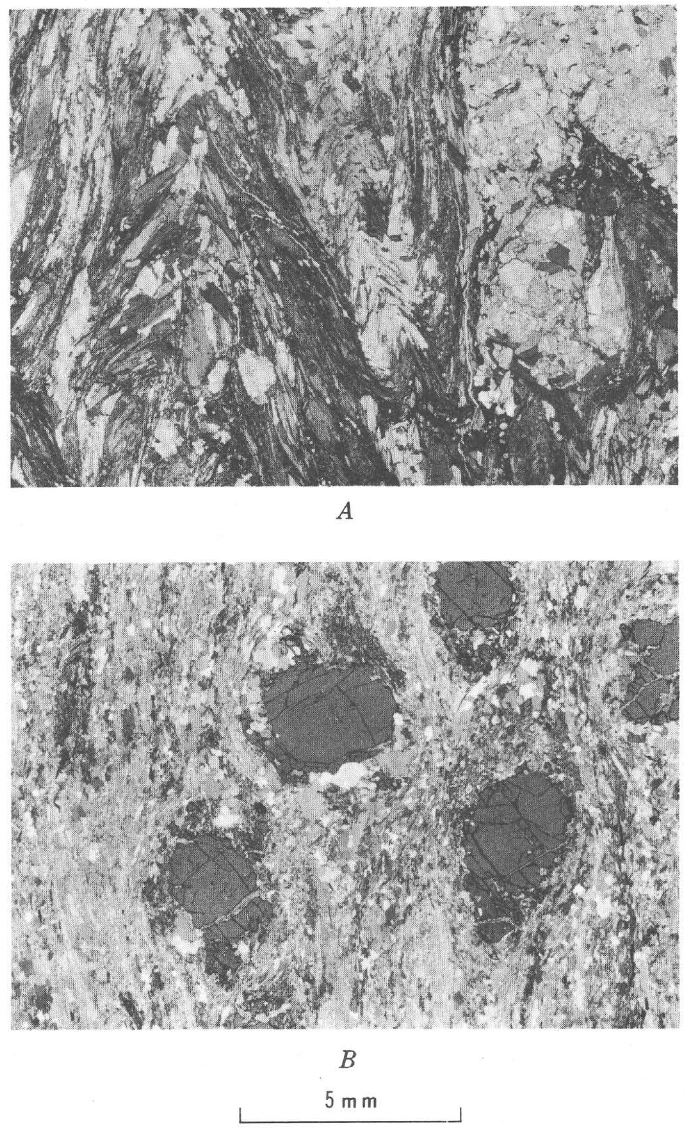



FIGURE 75.—Photomicrographs of sillimanite schist. *A*, Quartz-plagioclase-muscovite-sillimantie-biotite schist from south side of Rhodiss Lake about 1.6 miles west of Castle Bridge (area J–9, pl. 1). Sillimanite needles and fibrolite bundles partly parallel plane of section and bent around folds and are partly perpendicular to plane of section and parallel fold axes. Biotite in polygonal arcs. Muscovite in polygonal arcs and as postkinematic porphyroblasts. Quartz and plagioclase ($An_{30}$) concentrated in segregation. Polarizers at 45°. *B*, Altered sillimanite schist. Garnet-quartz-chlorite-sericite schist from roadcut 0.6 mile northwest of center of Drexel (area I–9, pl. 1). Sericite pseudomorphous after sillimanite, and chlorite, after biotite. Some sillimanite and chlorite included in garnet rims and in quartz. Polarizers at 45°.

anhedral grains of plagioclase ($An_{30}$). The larger quartz-feldspar pods and knots in the schist contain as much as 25 percent plagioclase of the same composition. Principal stable accessory minerals in the sillimanite schist are zircon and magnetite. Rutile occurs in a few specimens. Leucoxene forms fringes on biotite; garnet is commonly partly altered to limonite, and plagioclase, to clay minerals.

Where the sillimanite schist has been altered, sillimanite is replaced by pseudomorphs or structureless felted aggregates of sericite, biotite is replaced by chlorite, and garnet is partly or completely chloritized (fig. 75*B*). In spite of the alteration, the microscopic structure and texture of the original sillimanite schist is generally well preserved. A few needles of sillimanite are preserved as inclusions in muscovite porphyoblasts and in the cores of garnets. Subhedral to euhedral prisms of porphyroblastic tourmaline are abundant in some of the altered rocks. Tourmaline is black in hand specimen and colorless to yellowish brown in thin section. Many grains have light-green rims.

A chemical analysis of one specimen of altered sillimanite schist and average modal compositions of altered and unaltered sillimanite schist are given in table 24.

TABLE 24.—*Chemical analysis and modes of sillimanite schist*

[Chemical analyses determined by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey. Mode of analyzed sample determined by count of 600 points; P, present but not intersected in counting; Tr, trace. Major oxides given in weight percent; modes, in volume percent]

| | 1 | 2[1] | 3[2] |
|---|---|---|---|
| Field No. | 30-1480 | | |
| Laboratory No. | 160143 | | |
| **Major oxides** | | | |
| $SiO_2$ | 40.0 | | |
| $Al_2O_3$ | 32.1 | | |
| $Fe_2O_3$ | 2.7 | | |
| FeO | 8.2 | | |
| MgO | 2.3 | | |
| CaO | .20 | | |
| $Na_2O$ | 1.0 | | |
| $K_2O$ | 5.7 | | |
| $H_2O+$ | 4.2 | | |
| $H_2O-$ | .15 | | |
| $TiO_2$ | 1.7 | | |
| $P_2O_5$ | .21 | | |
| MnO | .24 | | |
| $CO_2$ | .05 | | |
| Total | 99 | | |
| **Modes** | | | |
| Quartz | 8 | 29.0 | 2.4 |
| Plagioclase | .1 | .5 | 2.0 |
| Biotite | | | 31.1 |
| Muscovite | 61 | 52.0 | 16.2 |
| Sillimanite | 2.2 | .5 | 40.4 |
| Garnet | 4 | 2.5 | 2.0 |
| Chlorite | 18 | 9.7 | |
| Limonite and leucoxene | 1.7 | 2.5 | 4.6 |
| Zircon | P | Tr | Tr |
| Tourmaline | | .5 | Tr |
| Apatite | P | Tr | |
| Rutile | | | Tr |
| Sphene | | | |
| Opaque minerals | 6 | 2.8 | 1.2 |

[1] Average of 50 random grains counted in each of eight thin sections of hydrothermally altered sillimanite schists.
[2] Average of 50 random grains counted in each of five thin sections of unaltered sillimanite schists.

1. Sillimanite-bearing chloride-muscovite schist. Lustrous light-gray fine-grained schist containing abundant small pegmatite pods and quartz segregations. Sillimanite needles and sericite pseudomorphs after sillimanite are randomly oriented on foliation planes. Rock consists of alined felted aggregates of sericite probably replacing sillimanite bundles, scattered porphyroblasts of garnet partly replaced by chlorite, and aggregates of anhedral quartz in segregation folia 1 to 2 mm in diameter. Small plates and irregular grains of opaque minerals are disseminated in sericite aggregates. Outcrop on south side of road 0.25 mile east of Hoyle Creek, 1.7 mile S. 46° W. of south end of Castle Bridge (area J–9, pl. 1).

### OTHER INTERLAYERED ROCKS

Layers and pods of impure quartzite, quartz schist, calc-silicate rocks, anthophyllite gneiss, and marble are locally intercalated with the layered gneisses and mica schist. These rocks constitute a very minor proportion of the layered rocks, but they are important as indicators of the origin of the rocks enclosing them. Stable mineral assemblages in these rocks are listed in table 25.

#### QUARTZITE AND QUARTZ SCHIST

Feldspathic quartzite (Used here as a general term for quartz-rich gneiss or quartz granofels.) and quartz schist occur as layers and pods interleaved with biotite gneiss, mica schist, sillimanite schist, hornblende gneiss, and amphibolite. The quartzite occurs as isolated pods and as continuous layers 1 inch 2 feet thick and is most commonly interlayered with sillimanite schist or biotite gneiss. Quartz schist forms layers as much as 20 feet thick and is generally interlayered with mica schist.

The quartzite is a medium- to dark-gray sugary to vitreous fine- to medium-grained rock, commonly thinly layered, but only rudely foliated. It consists of 60 to 80 percent anhedral quartz grains 0.5 to 1.0 mm in diameter, 0 to 20 percent plagioclase, and subordinate amounts of epidote, garnet, hornblende, and biotite. Sphene, zircon, rutile, and opaque minerals are common accessories. The quartz is commonly deeply sutured and has a rude dimensional orientation parallel to the foliation, which is commonly defined by folia of coarser and finer grained quartz mosaic. Plagioclase occurs in scattered untwinned grains in the quartz mosaic. It is generally sodic andesine. Colorless epidote having low to moderate bi-

**TABLE 25.**—*Stable-mineral assemblages in quartzite, quartz schist, calc-silicate rocks, and anthophyllite gneiss in the Inner Piedmont belt not affected by shearing along the Brevard fault zone*

[Stable accessory minerals omitted; ×, present]

| | Assemblage | Quartz | Plagioclase (An content) | Hornblende (color parallel to Z) | Biotite (color parallel to Z) | Muscovite | Epidote | Garnet | Diopside | Sillimanite | Anthophyllite |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Quartzite | 1 | × | | Blue green | | | × | × | | | |
| | 2 | × | | | | | × | | | | |
| | 3 | × | 36 | Blue green | Yellow brown | | × | | | | |
| | 4 | × | 88 | Light gray green | Red brown | | | × | | | |
| | 5 | × | | | do | × | | | | × | |
| | 6 | × | 38 | | do | × | | | | | |
| Quartz schist | 7 | × | | | | × | | | | | |
| | 8 | × | | | Light brown | × | | | | | |
| | 9 | × | | | Black | × | | × | | | |
| | 10 | × | | | Red brown | × | | × | | × | |
| Calc-silicate rocks | 11 | × | 80 | Greenish gray | do | | | × | × | | |
| | 12 | × | 80 | do | do | | × | × | | | |
| | 13 | × | 37 | Dark green | | | × | × | | | |
| | 14 | | × | Dark blue green | | | × | | × | | |
| | 15 | | 5 | | | | × | | | | |
| Anthophyllite gneiss | 16 | × | 35 | | | | | | | | × |

refringence occurs in scattered irregular skeletal grains 0.05 to 0.5 mm in diameter, commonly concentrated in thin layers. Garnet forms anhedral to subhedral poikilitic grains 0.1 to 0.5 mm in diameter, commonly forming aggregates parallel to layering. Anhedral grains of hornblende and biotite interstitial to quartz and plagioclase are irregularly distributed and randomly oriented. Quartzite layers in sillimanite schist commonly contain fibrolite aggregates alined with the foliation and scattered sillimanite needles in quartz.

The quartz schist is a medium- to coarse-grained light-gray to light-pink strongly foliated lustrous rock; it is most commonly intercalated with mica schist, but it is also interlayered in biotite and biotite-hornblende gneiss. The layers range from a few inches to as much as 30 feet in thickness; locally, quartz schist makes up a considerable proportion of the rocks mapped as mica schist. The rock consists of 60 to 80 percent quartz, 5 to 25 percent muscovite, 0 to 10 percent biotite, and generally contains small amounts of garnet and plagioclase. Sillimanite and staurolite occur in some specimens, and zircon, apatite, pyrite, and magnetite are common accessory minerals. An average modal composition of quartz schist is given in column 4, table 21. As muscovite content decreases, the quartz schist is compositionally gradational into quartzite; as biotite and plagioclase content increases, the quartz schist is gradational into mica schist.

Quartz forms an inequigranular mosaic of sutured elongate anhedral grains 0.1 to 3.0 mm long and segregation laminae of larger grains alined with the foliation. A few anhedral grains of calcic oligoclase or sodic andesine are scattered in the quartz mosaic. Muscovite occurs in synkinematic and postkinematic flakes 0.25 to 5 mm long, forming redundant folia or scattered through the quartz mosaic. Some muscovite books contain abundant poikilitic inclusions of quartz. Biotite forms scattered synkinematic flakes, as much as 3 mm long, generally smaller than muscovite and interleaved with it. Garnet forms subhedral to anhedral porphyroblasts, as much as 3 mm in diameter, many of them containing abundant poikilitic inclusions of quartz and magnetite. Sillimanite forms fibrolite aggregates in muscovite and biotite and scattered rudely alined needles in quartz.

In polymetamorphic quartz schist in the belt of shearing and metamorphism along the Brevard fault zone, muscovite forms wedge-shaped porphyroclasts, biotite is largely recrystallized to small anhedral grains interstitial to quartz, and quartz is intensely strained and partially recrystallized. Garnets commonly have conspicuous snowball strcture. In a few places in the belt of polymetamorphic rocks, the quartz schist contains scattered porphyroblasts or porphyroclasts of poikilitic staurolite.

### CALC-SILICATE ROCKS

Fine- to medium-grained medium-gray, greenish-gray, or yellowish-green calc-silicate granofelses form layers or pods 1 inch to 2 feet thick in the gneisses and schists. They are most commonly interleaved with hornblende gneiss and amphibolite but also occur in biotite gneiss and sillimanite schist. They consist of various proportions of quartz, plagioclase, epidote, hornblende, pyroxene, and garnet. Some calc-silicate rocks contain as much as 50 percent quartz and are compositionally gradational with quartzites. Others lack quartz and are composed entirely of plagioclase, pyroxene and epidote, or plagioclase and epidote. Contacts of the layers are sharp, and their composition, internal layering, and continuity suggest that they were originally sedimentary beds, rather than metamorphic segregations.

Textures are generally granular. Quartz and plagioclase form mosaics of anhedral sutured grains 0.05 to 0.5 mm in diameter or occur as scattered grains interstitial to epidote, pyroxene, and garnet. Plagioclase ranges from sodic andesine in some rocks to bytownite in others. Epidote forms irregular to subhedral skeletal grains as much as 1 mm in diameter. It is generally colorless and has moderate to low birefringence and anomalous interference colors. In some of the calc-silicate rocks, however, it is pleochroic in shades of yellow and has high birefringence. Pyroxene is colorless to light-apple-green diopside that forms anhedral poikilitic grains as much as 2 mm in diameter. Hornblende occurs as poikilitic grains generally less than 0.5 mm in diameter. In some rocks it seems to be contemporaneous with diopside, but in others it forms partial jackets around diopside. Garnet forms subhedral to anhedral skeletal grains generally less than 1 mm in diameter. Sphene, apatite, and magnetite are common accessory minerals, and zircon occurs in a few specimens. Near the Brevard fault zone, plagioclase in some rocks is entirely altered to albite and clinozoisite, and hornblende, garnet, and pyroxene are partly altered to chlorite.

### ANTHOPHYLLITE GNEISS

In outcrops on the east side of Canoe Creek, 1.2 miles S. 54°W. of Oak Hill School (area F–9, pl. 1), anthophyllite gneiss is interlayered with biotite gneiss, hornblende gneiss, and amphibolite. The anthophyllite gneiss is a medium-grained dark-gray strongly foliated rock having sheaves of gray anthophyllite randomly arrayed on the foliation planes. The layers are a few inches to several feet thick. The rock consists of 5 to 10 percent quartz, 40 to 50 percent andesine, and 20 to 30 percent anthophyllite and contains a few flakes of biotite and scattered grains of magnetite. Quartz and plagioclase occur in anhedral grains 0.1 to 0.5 mm in diameter. Anthophyllite prisms as much as 1 cm long lie in the plane of foliation but randomly oriented within it. Anthophyllite is partly altered to a pale-green monoclinic amphibole and locally, to chlorite. Apatite is a minor accessory mineral.

### MARBLE

Layers and pods of impure calcite marble are intercalated with sheared and retrogressively metamorphosed gneiss, schist, and amphibolite in a few places adjacent to the Brevard fault zone. These localities are in: (1) roadcuts on the south side of the county road south of the Catawba River, 2.1 to 2.3 miles S. 52°W. of the junction of U.S. Highways 70 and 221 northwest of Marion (area A–10, pl. 1), (2) roaducts on the east side of the county road east of the Catawba River, 1.0 mile S. 39°W. of the same highway junction (area A–10), (3) a roadcut on the north side of the county road north of the head of Lake James, 0.3 mile N. 58° W. of the center of the Clinchfield Railroad trestle over the head of the lake (area C–10), (4) outcrops on the northwest shore of Lake James, 2.1 miles N. 39° E. of the same trestle (area C–10), (5) roadcuts in area D–9 on the north side of the county road north of Lake James, 1.1 miles S. 84° E. of the U.S. Geological Survey gaging station at the mouth of the Linville River (area E–9, pl. 1), and (6) outcrops southeast of the Johns River, 0.5 mile S. 83° E. of Collettsville (area H–7, pl. 1). The first three of these occurrences of marble have been described by Conrad (1960), who described a similar occurrence about 5 miles southwest of Marion.

The marble is a fine- to medium-grained medium-gray rudely foliated rock, commonly having a light-blue or purple cast on fresh surfaces. Porphyroclasts of muscovite and flakes of red-brown biotite as much as 5 mm in diameter are conspicuous in hand specimens, and the marble contains thin micaceous partings and quartz-feldspar folia, many of them highly contorted. The marlbe occurs in pods and layers conformable with the layering in the enclosing schist and gneiss. Contacts with the enclosing rocks are commonly gradational but are locally sharp. The exposed marble layers range from a few inches to at least 30 feet thick, and Conrad (1960) reported that a diamond-drill hole near the outcrop 1.0 mile southwest of the junction of U.S. Highways 70 and 221 penetrated nearly 130 feet of marble. The marble is interlayered with biotite gneiss, quartz-feldspar gneiss, biotite-muscovite-chlorite schist, pegmatite, and amphibolite, some of which contains diopside. The enclosing rocks have well-developed cataclastic textures and structures related to shearing and retrogressive metamorphism along the Brevard fault zone, but these features are not everywhere conspicuous in the marble itself.

The marble has a granular to rudely foliated texture. Inequigranular grains of twinned calcite 0.25 to 1.0 mm in diameter form 25 to 90 percent of the rock. Quartz and plagioclase occur in scattered anhedral grains interstitial to calcite and in thin layers and folia parallel to the foliation. Muscovite occurs as wedge-shaped deformed porphyroclasts, as much as 1 mm long, and as small synkinematic or postkinematic flakes 0.1 to 0.2 mm long. Reddish-brown biotite forms synkinetimatic and postkinematic flakes and anhedral grains, as much as 0.25 mm long, generally concentrated in the quartz-plagioclase layers. Scattered granules of clinozoisite are common in some specimens, and apatite, sphene, and magnetite are common accessory minerals. The marble near Collettsville contains porphyroclasts of microcline as much as 2 mm in diameter, fringed with myrmekite, and abundant chlorite derived from alteration of biotite.

Plagioclase is clear and well twinned, and some is faintly zoned. Its composition is about $An_{35}$. The carbonate has the optical properties of calcite, but a partial chemical analysis of the marble near Marion quoted by Conrad (1960) indicates that 11 percent of weight of the total carbonate in the rock is $MgCO_3$. If no disseminated dolomite has been overlooked in the grain mounts, the carbonate must be a highly magnesian calcite.

In all the known outcrops, the marble is associated with polymetamorphic rocks within a few hundred yards of the Brevard fault zone, and it is not clear which minerals crystallized during metamorphism prior to faulting and which crystallized during shearing and metamorphism during movement along the fault zone. Mineral assemblages in the marble are therefore not listed in table 25.

### ORIGIN

Shearing and metamorphism have completely obliterated any textures and structures that might have indicated the nature of the original materials from which the layered rocks of the Inner Piedmont were derived. The layered character of the sequence suggests that the rocks were derived from a stratified sequence, but deformation has obscured or obliterated original stratigraphic relationships, and even the contacts between individual layers are probably largely tectonic. There is no evidence of any important modification of the bulk composition of the layered rocks during metamorphism, and the chemistry of the rocks, therefore, furnished the only clues, however tenuous, to the protoliths from which they were derived.

Some of the rocks in the Inner Piedmont are clearly of sedimentary origin. These include the quartzite, quartz schist, marble, sillimanite schist, and some of the amphibolites and the lime-silicate rocks containing diopside and bytownite. The origin of the remainder of the gneisses, schists, and amphibolites that make up the bulk of the layered rocks of the Inner Piedmont is less obvious.

Terranes similar to the layered rocks of the Inner Piedmont are commonly believed to have been produced by metamorphism of thick eugeosynclinal accumulations of graywackes and volcanic rocks. To test this hypothesis, chemical analyses of gneisses and schists from the Inner Piedmont are compared with analyses of rocks from the Aleutian Islands (Hamilton, 1963, p. 70–71), the Carolina slate belt in the Albermarle-Denton area, North Carolina (A. A. Stromquist, written comm., 1960), and with analyses of graywackes quoted by Pettijohn (1963, table 6).

Hamilton (1963, p. 70–71) has briefly described the geology of the Aleutian Islands and has summarized a large number of recent rock analyses. The Aleutians are a chain of volcanic islands formed of lavas and tuffs, of sedimentary rocks derived directly from them, and of dikes, sills, and plutons intrusive into them. Two chief groups of rocks are exposed: older eugeosynclinal submarine lavas and tuffs, generally altered and partly reworked by water, and younger unaltered subaerial flows and tuffs unconformably overlying the eugeosynclinal rocks and forming great stratovolcanoes, some of which are still active. Where exposed on the islands, the submarine volcanic rocks have yielded fossils of various ages from Pennsylvanian to Tertiary. The subaerial volcanic rocks range in age from later Tertiary to Recent and are predominantly andesite with lesser dacite and basalt and rare rhyodacite and quartz latite. Compositionally, the submarine volcanic rocks are chiefly andesite, basalt, dacite, and their albitized equivalents, keratophyre, spilite, and quartz keratophyre. Rhyodacite and quartz latite are present but are uncommon.

The Carolina slate belt is a group of volcanic and associated sedimentary rocks of low metamorphic grade that extends from southern Virginia southward across the Carolinas into northern Georgia (King, 1955, p. 343). The rocks of the slate belt form a sequence of alternating rhyolitic to basaltic volcanic rocks, interbedded with argillite, siltstone, sandstone, and their tuffaceous equivalents, and intruded by dikes, stocks, sills, and batholiths of granitic and gabbroic rocks. Some of the volcanic and sedimentary rocks are probably submarine, but others are clearly subaerial (Stromquist and Conley, 1959; Sundelius, 1963; Conley and Bain, 1965). Some of the rocks in the Albemarle-Denton area have yielded zircon that has been dated as Ordovician (White and

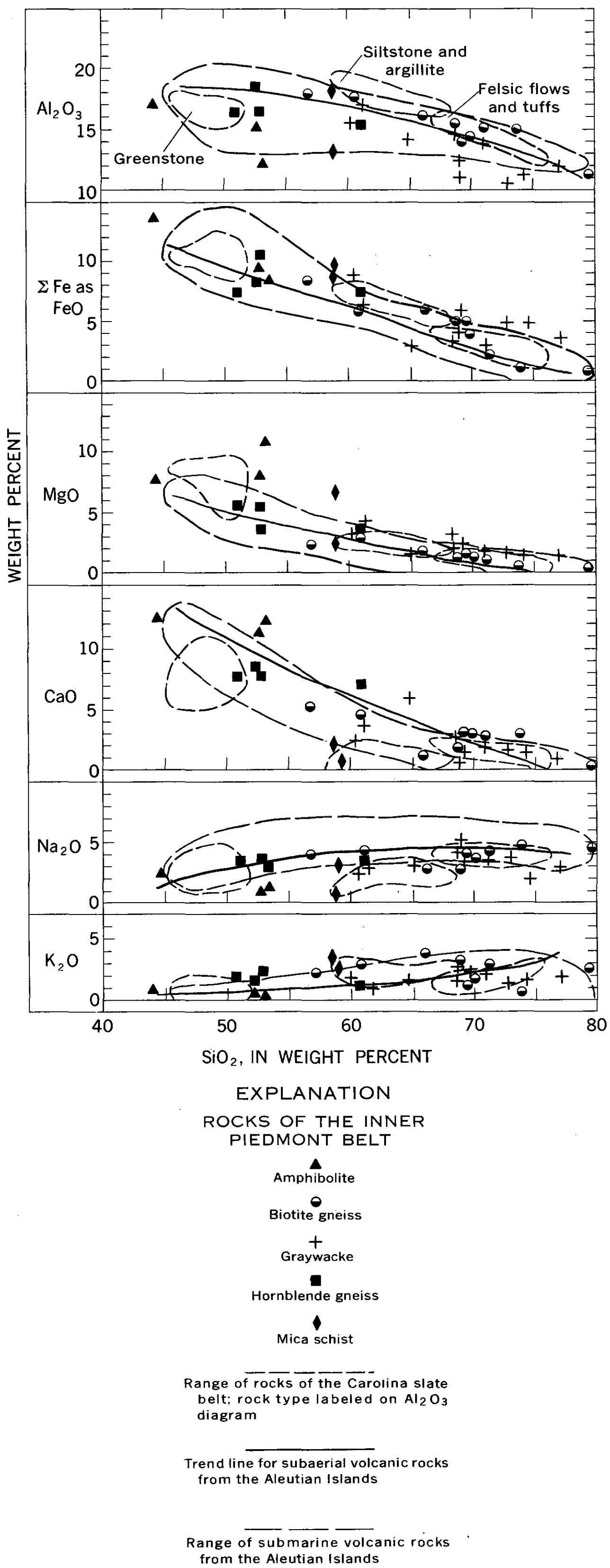



FIGURE 76.—Silica variation diagrams comparing layered gneisses, schists, and amphibolites of the Inner Piedmont belt with volcanic rocks of the Aleutian Islands, rocks of the Carolina slate belt, and graywackes. Data for the Aleutian rocks from Hamilton (1963); for the slate-belt rocks, from A. A. Stromquist, U.S. Geological Survey; and for graywackes, from Pettijohn (1963). Amphibolite and hornblende gneiss analyses from table 23; biotite gneiss, table 20; mica schist, table 21.

others, 1963). Several distinct depositional sequences have been recognized, and rocks of widely different ages may be present. Many chemical analyses of rocks from the slate belt have been published (Butler, 1964), but many are of doubtful reliability and the specimens are inadequately described. A group of unpublished modern analyses of well-described rocks from the Albemarle-Denton area has kindly been made available by A. A. Stromquist, U.S. Geological Survey, and have been selected for comparison.

Analyses of 10 graywackes, many of them associated with tuffs and submarine volcanic rocks in eugeosynclinal environments, are quoted by Pettijohn (1963, table 6) and are used for comparison. These rocks are from a variety of geologic provinces and range in age from Precambrian to Eocene.

The silica variation diagrams (fig. 76) show that most of the biotite and hornblende gneisses of the Inner Piedmont are chemically comparable to the volcanic rocks of the Aleutian Islands and to some of the rocks of the Carolina slate belt, as well as to the graywackes. Their compositions most closely approach those of the subaerial volcanic rocks of the Aleutians, but many of them tend to be slightly richer in $Al_2O_3$ and slightly lower in $Na_2O$. They tend to show a complete range of $SiO_2$ content, having unimodal distribution, from hornblende gneisses resembling andesites in composition to biotite gneisses resembling rhyodacites and having a broad frequency maximum in the range of 65 to 70 percent. In the Aleutian suite, the $SiO_2$ content also has a unimodal distribution, but it has a frequency maximum in the range of 55 to 60 percent, whereas the gneisses of the Inner Piedmont more commonly have $SiO_2$ contents in the range of 65 to 70 percent.

Many of the more silicic gneisses closely resemble the felsic volcanic rocks of the slate belt, but in the slate belt, at least in the Albemarle-Denton area, the $SiO_2$ content has a strong bimodal distribution, and volcanic rocks having $SiO_2$ contents in the range of 55 to 65 percent are absent. Siltstones and argillites in the slate belt commonly have $SiO_2$ contents in the latter range, but the gneisses of the Inner Piedmont having comparable $SiO_2$ contents are distinctly richer in CaO and somewhat richer in $Na_2O$.

Most of the graywackes for which analyses are quoted by Pettijohn (1963, table 6) have $SiO_2$ contents in the range 65 to 75 percent. Inner Piedmont gneisses having $SiO_2$ contents in this range are appreciably richer in $Al_2O_3$, somewhat richer in CaO, and poorer in MgO and total iron.

Most of the layered gneisses have an excess of $Na_2O$ over $K_2O$, a feature that is very characteristic of graywackes (Middleton, 1960; Pettijohn, 1963). Figure 77 shows, however, that both the subaerial volcanic rocks of the Aleutian Islands and the rocks of the Carolina slate belt also have $Na_2O$ : $K_2O$ ratios are greater than 1, so that this property cannot be used to distinguish volcanic rocks from graywackes. The submarine volcanic rocks of the Aleutians have $Na_2O$ :$K_2O$ ratios greater than 1, but the ratios are quite variable because of extensive albitization, and they are therefore not shown on the diagram.

Figure 78 is a plot of the molecular proportions of alumina, lime, and total alkalies in the various rocks. The gneisses of the Inner Piedmont fall in a trend very similar to the trend line for the Aleutian subaerial volcanic rocks, and many plot in positions intermediate between the felsic and mafic volcanic rocks of the slate belt. Some fall in the same general field as the graywackes, but as a group the gneisses seem to plot along a volcanic trend.

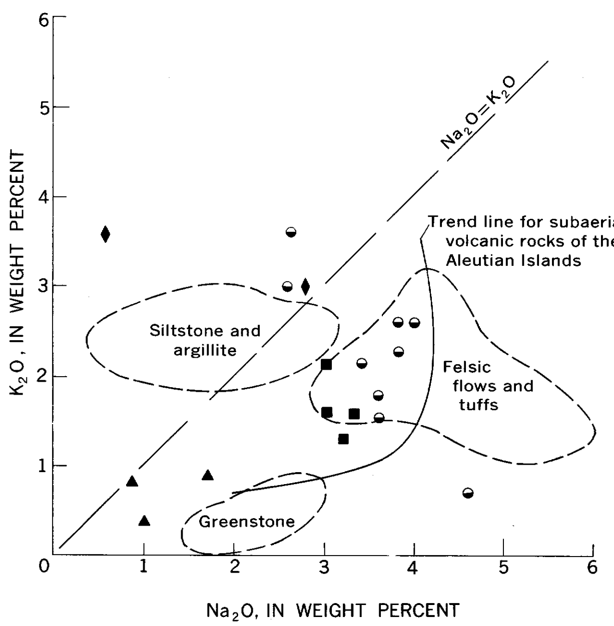



FIGURE 77.—$Na_2O$:$K_2O$ variation diagram comparing layered gneisses, schists, and amphibolites of the Inner Piedmont belt with volcanic rocks of the Aleution Islands and rocks of the Carolina slate belt. Symbols and sources of data are the same as in figure 76.

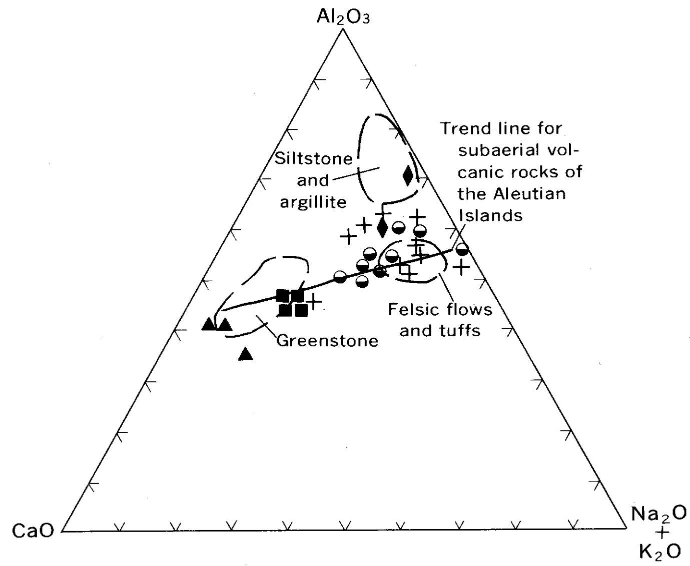



FIGURE 78.—Molecular proportions of $Al_2O_3$, CaO, and total alkalies in layered gneisses, schists, and amphibolites of the Inner Piedmont belt compared with subaerial volcanic rocks of the Aleutian Islands, rocks of the Carolina slate belt, and graywackes. Symbols and sources of data same as in figure 76.

The two analyses of mica schist and the three analyses of amphibolite do not fit trends of the biotite and hornblende gneisses in figures 76, 77, and 78. The schists have higher total Fe and $K_2O$ and lower $Na_2O$ and CaO contents than gneisses of comparable $SiO_2$ content, and one of them has low $Al_2O_3$ and high MgO contents. One of the amphibolites has a lower $SiO_2$ content than any of the volcanic rocks of the Aleutians or the slate belt. The other two amphibolites have higher MgO and CaO and lower $Na_2O$ contents than any of the volcanic rocks of similar $SiO_2$ content. One of them also contains considerably less $Al_2O_3$.

The chemical anlayses suggest that many of the biotite and hornblende gneisses and perhaps some of the amphibolite could have been produced by metamorphism of volcanic rocks similar to those of the Aleutian volcanic arc but that some of the biotite gneisses may also have been derived from graywackes. The mica schists were probably interbedded shales or argillites, and at least some of the amphibolites were shaly dolomites. Also interlayered with the volcanic-graywacke sequence were calcareous and argillaceous sandstone (now metamorphosed to quartzite and quartz schist) and impure dolomitic limestone (now metamorphosed to marble, calc-silicate rocks, or anthophyllite gneiss). The sillimanite schist must originally have been an aluminous shale, perhaps derived from clays produced during chemical weathering of volcanic rocks.

### AGE

The age of the layered rocks of the Inner Piedmont belt in the Grandfather Mountain area is not obvious. The rocks are, however, older than the middle or lower Paleozoic granitic rocks that invade them. Keith (1905) mapped the biotite gneisses and mica schists in the adjacent part of the Mount Mitchell quadrangle as Carolina Gneiss, and the hornblende gneiss and amphibolite as Roan Gneiss, both of which he believed to be of Archean or early Precambrian age. Both terms have now been abandoned. Many of the rocks so mapped in the Blue Ridge are clearly of early Precambrian age, but there is no evidence that the rocks mapped by Keith as Carolina and Roan in the Inner Piedmont belt are of the same age.

Recent work in central Virginia (Bloomer, 1950; Brown, 1958; Espenshade, 1954; Smith and others, 1964) has shown that large parts of the Piedmont are underlain by sedimentary and volcanic rocks of late Precambrian and early Paleozoic age. According to Kesler (1944) and Overstreet and Bell (1960; 1965, p. 100–102, 108–109), many of the rocks in the Inner Piedmont belt in North Carolina and South Carolina may be of late Precambrian or early Paleozoic age, and some of them may be equivalents of less metamorphosed Paleozoic rocks in the Kings Mountain belt and the Carolina slate belt. The layered rocks of the Inner Piedmont belt in the Grandfather Mountain area are therefore probably of late Precambrian or early Paleozoic age, and rocks of both ages may be present.

### METAMORPHISM

The layered rocks of the Inner Piedmont belt are of high metamorphic grade, except where they have been affected by shearing and retrogressive metamorphism associated with the Brevard fault zone or by local shearing and hydrothermal alteration. They contain a complex array of mineral assemblages (tables 18, 21, and 24), which occur in interlayered rocks and which have no conspicuous pattern of distribution that would suggest a regional metamorphic gradient. It is therefore assumed that the high-grade regional metamorphism of the layered rocks took place under conditions that were rather uniform within the area studied.

The diversity and complexity of the composition and mineralogy of the layered rocks and the lack of detailed data on compositions of the coexisting minerals precludes adequate representation of the mineral assemblages on phase diagrams. A schematic ACF diagram (fig. 79), however, illustrates the general paragenetic relationships.

The layered rocks contain mineral assemblages that in general correspond to the cooler part of the sillimanite zone and to the almandine-amphibolite facies of Turner (Fyfe and others, 1958), but the ACF diagram (fig. 79) does not agree closely with those for any of the subfacies described by Turner. The appearance of the apparently stable mineral pairs—sillimanite-muscovite, epidote-diopside, and epidote-plagioclase—suggests pressures and temperatures somewhat lower or $P_{H_2O}$ somewhat higher than those characteristic of Turner's sillimanite-almandine subfacies, but the occurrence of sillimanite instead of kyanite and the absence of staurolite in rocks of suitable composition do not correspond to either the kyanite-muscovite-quartz or staurolite-quartz subfacies.

Shearing and recrystallization of the rocks along the Brevard fault zone have produced polymetamorphic rocks, many of which contain parts of two or more mineral assemblages. The widespread occurrence of recrystallized biotite, muscovite, oligoclase, and epidote and the presence of porphyroclasts of staurolite and kyanite in some of the rocks suggest that the initial recrystallization was under medium-grade conditions, perhaps corresponding to the low-intensity part of the almandine-amphibolite facies of Turner. Later shearing, largely concentrated closer to the Brevard fault zone, has resulted in further retrogression, producing low-grade assemblages containing sericite, chlorite, epidote, and albite, which are characteristic of Turner's greenschist facies.

## HENDERSON GNEISS

The Henderson Gneiss is a biotite quartz monzonite augen gneiss that forms pods and elongate lenses concordant with foliation and layering in the enclosing layered rocks (pl. 1). It is most common in the belt of polymetamorphic rocks southeast of the Brevard fault zone, but it also occurs in a few small bodies among the unsheared rocks farther southeast. The name Henderson Gneiss has been applied to the gneiss in the Grandfather Mountain area because of its strong lithologic similarities to and apparent continuity with rocks mapped as Henderson Granite by Keith (1905, 1907b) in Henderson County, N.C., about 30 miles southeast. The lithologic designation

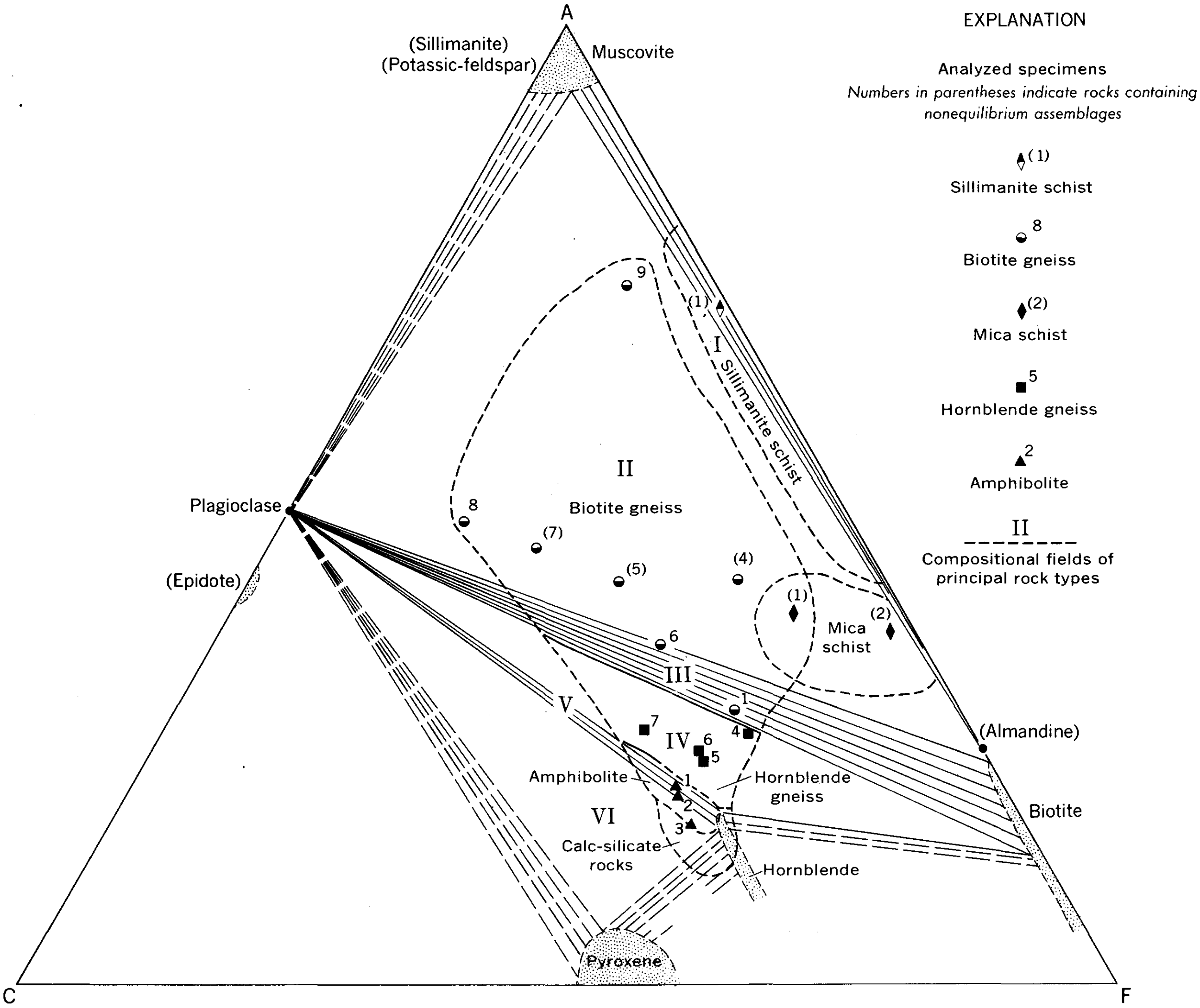



FIGURE 79.—Schematic ACF diagram for layered rocks of the Inner Piedmont belt containing excess $SiO_2$. Positions of fields and tie lines largely schematic, but consistent with analyses of rocks studied and published analyses of minerals from rocks of similar metamorphic grade. Possible additional minerals are in parentheses. Compositional fields of the principal rock types outlined by short dashed lines; quartzite, quartz schist, and marble are omitted. Fields containing observed assemblages are numbered. Quartz is an additional phase in all assemblages; epidote appears in all assemblages; epidote appears in all observed fields but not in all assemblages. Garnet occurs in fields I and II in rocks where the molecular proportion of FeO exceeds that of MgO. Potassic feldspar occurs in fields I and II in rocks in which the molecular ratio of excess $Al_2O_3$:$K_2O$ is less than 3. Sillimanite occurs in field I in rocks containing high proportions of $Al_2O_3$, but the ratios controlling its appearance in rocks of the Grandfather Mountain area are not determined by the data available. Analyses of similar rocks by Shaw (1956) indicate it is confined to rocks in which the molecular ratio $Al_2O_3$:$Na_2O+K_2O+$ CaO exceeds 2. Occurrence of sphene is limited to field III. Other accesory minerals appear in all fields. Sillimanite schist analyses from table 24; biotite gneiss, table 20; mica schist, table 21; hornblende gneiss and amphibolite, table 23.

was changed from granite to gneiss by Reed (1964b) to better describe the lithology of the unit.

The gneiss is a fine- to medium-grained light- to medium-gray rock, containing abundant light-gray to white augen of potassic feldspar 0.5 to 3.0 cm long. The augen typically have single Carlsbad twins and are rimmed by a thin chalky-white jacket of quartz and plagioclase. Where the rock has not been subjected to shearing associated with the Brevard fault zone, it is weakly foliated, and the augen are ovoid in outline and rudely alined or randomly arrayed. Most bodies of Henderson Gneiss, however, are in the belt of polymetamorphic rocks associated with the Brevard fault zone (pl. 6*C*, *D*). There the gneiss is strongly foliated and lineated, and the feldspar augen are reduced to lentil-shaped porphyroclasts, flattened in the foliation and having their long axes alined with the cataclastic lineation. The gneiss is typically nonlayered in the centers of the larger bodies, but it commonly becomes rudely layered near the contacts and passes gradationally into layered gneisses containing scattered potassic feldspar augen. This transition is most conspicuous along strike from the ends of the larger lens-shaped bodies. Layered biotite gneiss near bodies of Henderson Gneiss commonly contains layers a few inches to several feet thick containing abundant augen of potassic feldspar similar to those in the Henderson but generally in a finer grained and darker, more biotitic groundmass.

Where the gneiss is unsheared, the matrix consists of an inequigranular mosaic of anhedral grains of sodic andesine and microcline 0.5 to 2 mm in diameter, strained quartz 0.1 to 2.0 mm in diameter, and rudely alined irregular flakes of reddish-brown biotite as much as 2.0 mm long. Apatite occurs as scattered prisms, irregular aggregates, and commonly as inclusions in biotite; allanite forms small inclusions surrounded by pleochroic halos in biotite. Other accessory minerals are sphene, zircon, and magnetite. Sericite occurs as an alteration product in the feldspars and as fringes on the ends of biotite books. The feldspar augen are porphyroblasts of perthitic microcline, typically containing rectangular inclusions of plagioclase and amoeboid blebs of quartz. The microcline porphyroblasts are partially or entirely rimmed by myrmekite, which invades the potassic feldspar in wartlike indentations. A few specimens also contain porphyroblasts of andesine as much as 1.0 cm in diameter.

Where the gneiss has been sheared and recrystallized near the Brevard fault zone, it consists of an inequigranular granoblastic-lepidoblastic mosaic of quartz, plagioclase (chiefly calcic oligoclase), potassic feldspar, biotite, muscovite, epidote, and small amounts of chlorite, garnet, and green hornblende. The grains in the mosaic are 0.01 to 0.1 mm in diameter. Foliation is defined by segregation lamellae of quartz and feldspar and by diffuse folia of biotite and muscovite (fig. 80*A*). Biotite occurs in small irregular flakes and grains and seems to be largely late synkinematic and partly postkinematic. Muscovite seems to be largely contemporaneous with the biotite, but a few specimens contain large bent porphyroclasts of old muscovite. Epidote occurs as scattered grains and poikilitic grains and aggregates; some

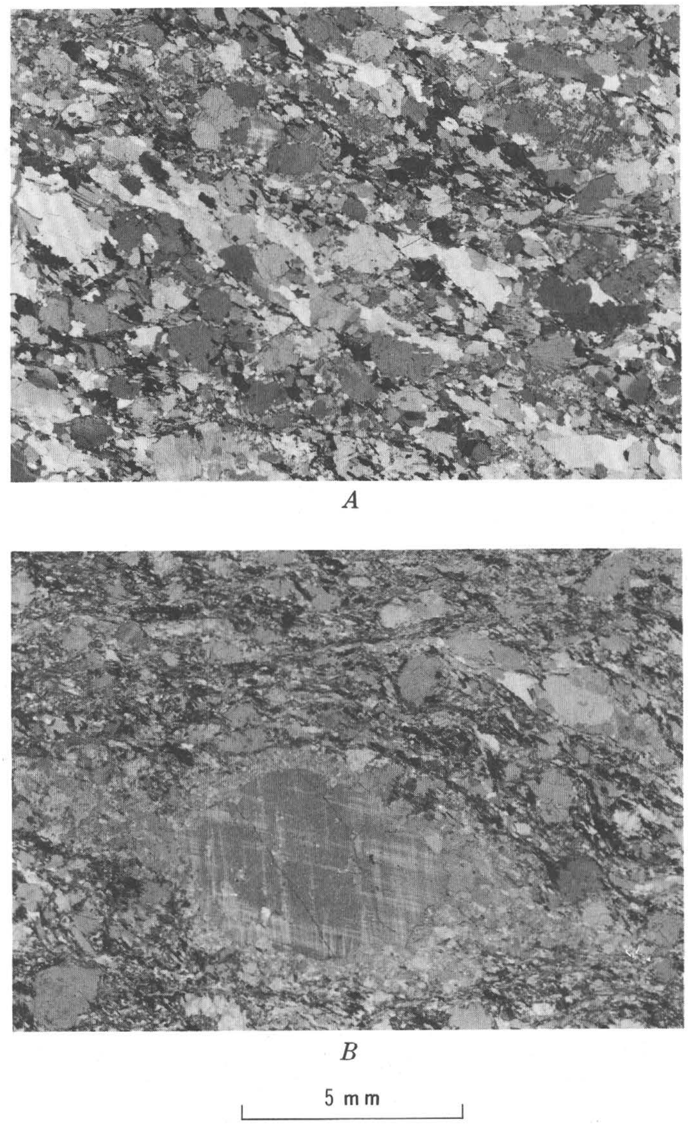



FIGURE 80.—Photomicrographs of Henderson Gneiss. *A*, Biotite quartz monzonite gneiss from east shore of Lake James about 1 mile west-northwest from River Valley Church (area E–9, pl. 1). Microcline and calcic oligoclase relicts of an earlier generation and recrystalized quartz, calcic oligoclase, biotite, and microcline. Some coarse-grained quartz in segregations. From near southeastern margin of belt of rock containing conspicuous polymetamorphic textures. Outcrop has microcline porphyroclasts as long as 1 cm. Polarizers at 45°. *B*, Porphyroclastic epidote-biotite-muscovite quartz monzonite gneiss from West side of Upper Creek, 0.2 mile west of east edge of the Linville Falls quadrangle (area F–8, pl. 1). Large and small porphyroclasts of microcline and smaller porphyroclasts of calcic oligoclase in a matrix of recrystallized quartz, microcline, calcic oligoclase, muscovite, biotite, and epidote. Typical jacket of myrmekite, quartz, and plagioclase on microcline porphyroclast, which tapers off to spindle-shaped extension parallel with lineation. Polarizers at 45°.

TABLE 26.—*Chemical analyses, modes, and norms of Henderson Gneiss and biotite gneiss transitional into Henderson Gneiss*

[Analysis of sample 2 determined by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey. Other analyses, by standard chemical methods by C. L. Parker and D. F. Powers, U.S. Geol. Survey. Nd, not determined. Modes of anlyzed specimens determined by point counts; P, present but not intersected in counting; Tr, trace. Major oxides and CIPW norms given in weight percent; modes, in volume percent]

| | 1 | 2 | 3 | 4 | 5 [1] | 6 [2] |
|---|---|---|---|---|---|---|
| Field No | 28–1045 | A–318 | 15–1061 | 21–2331 | | |
| Laboratory No | I–4061 | 161247 | H–3433 | H–3432 | | |
| **Major oxides** | | | | | | |
| $SiO_2$ | 64.08 | 67.4 | 70.37 | 75.46 | | |
| $Al_2O_3$ | 16.16 | 15.4 | 14.59 | 12.90 | | |
| $Fe_2O_3$ | 1.51 | .72 | 1.30 | .69 | | |
| FeO | 3.28 | 3.8 | 1.48 | 1.17 | | |
| MgO | 1.34 | 1.1 | .61 | .04 | | |
| CaO | 3.17 | 2.8 | 1.58 | .40 | | |
| $Na_2O$ | 3.66 | 3.8 | 3.33 | 3.73 | | |
| $K_2O$ | 4.33 | 2.6 | 4.95 | 4.74 | | |
| $H_2O+$ | .55 | .89 | .72 | .43 | | |
| $H_2O-$ | .05 | .13 | .11 | .12 | | |
| $TiO_2$ | .97 | .82 | .47 | .11 | | |
| $P_2O_5$ | .29 | .38 | .12 | .01 | | |
| MnO | .10 | .10 | .07 | .03 | | |
| $CO_2$ | .03 | .05 | .01 | .02 | | |
| Cl | .01 | Nd | Nd | Nd | | |
| F | .09 | Nd | Nd | Nd | | |
| Less O | .04 | | | | | |
| Total | 99.58 | 100 | 99.71 | 99.85 | | |
| **Modes** | | | | | | |
| Quartz | 20 | 29 | 34 | 37 | 26.8 | 30.1 |
| Plagioclase | 40 | 42 | 24 | 31 | 27.5 | 32.9 |
| Potassic feldspar | 14 | 6 | 19 | 26 | 20.0 | 22.4 |
| Biotite | 20 | 20 | 10 | 3 | 12.3 | 10.1 |
| Muscovite | 1.5 | 2.3 | 9 | 1.3 | 8.6 | 3.5 |
| Amphibole | | | | .3 | | Tr |
| Epidote | 2.7 | .2 | 4 | .2 | 3 | .3 |
| Apatite | .8 | .8 | | | Tr | .1 |
| Zircon | P | P | P | | .1 | Tr |
| Sphene | 1.0 | .2 | .1 | | Tr | Tr |
| Opaque minerals | | Tr | | | .5 | Tr |
| Garnet | | | .3 | .6 | Tr | Tr |
| Points counted | 662 | 600 | 600 | 610 | | |
| **CIPW norms** | | | | | | |
| Q | 17.12 | 26.56 | 27.69 | 34.15 | | |
| C | .68 | 2.27 | 1.20 | .98 | | |
| Or | 25.58 | 15.36 | 29.24 | 28.00 | | |
| Ab | 30.88 | 32.14 | 28.16 | 31.54 | | |
| An | 13.08 | 11.09 | 6.99 | 1.79 | | |
| Hl | .02 | | | | | |
| En | 3.34 | 2.74 | 1.52 | .10 | | |
| Fs | 3.36 | 5.21 | 1.00 | 1.54 | | |
| Mt | 2.19 | 1.04 | 1.88 | 1.00 | | |
| Il | 1.84 | 1.56 | .89 | .21 | | |
| Ap | .69 | .90 | .28 | .02 | | |
| Fr | .16 | | | | | |
| Cc | .07 | .11 | .02 | .04 | | |

[1] Average of 50 random grains counted in each of 13 thin sections of fine-grained biotite gneisses containing conspicuous potassic feldspar augen and believed to be marginal phases of Henderson Gneiss.

[2] Average of 50 random grains counted in each of 15 thin sections of typical Henderson Gneiss.

NOTE.—Minor-element analyses for samples, 1, 3, and 4 given in table 1.

1. Marginal facies of Henderson Gneiss. Strongly foliated and lineated nonlayered fine-grained medium-gray biotite-quartz-plagioclase gneiss containing abundant ovoid augen of pink potassic feldspar as much as 1 cm long, elongated parallel to lineation. Augen are jacketed by a thin mantle of quartz and plagioclase. Texture is conspicuously cataclastic. The augen are porphyroclasts of microcline set in a matrix of recrystallized anhedral quartz and plagioclase ($An_{32}$) 0.05 to 0.25 mm in diameter and stubby flakes of synkinematic and postkinematic brown biotite as much as 0.2 mm long. Plagioclase also occurs in faintly zoned porphyroclasts as much as 3 mm in diameter. Epidote occurs in irregular grains and subhedral prisms less than 0.2 mm long. Muscovite forms small postkinematic flakes. Microcline porphyroclasts are jacketed by myrmekite partly recrystallized to quartz and plagioclase. Roadcut on southeast side of county road southeast of Yadkin River, 1.0 mile S. 86° E. of village of Happy Valley (area J–6, pl. 1).
2. Typical Henderson Gneiss. Well-foliated nonlayered medium-grained biotite-quartz-plagioclase gneiss containing abundant potassic feldspar augen as much as 3 cm long. Augen have single Carlsbad twins and conspicuous thin chalky-white jackets of quartz and plagioclase. Texture is not polymetamorphic. Porphyroblasts of microcline and of plagioclase ($An_{30}$) as much as 0.4 mm in diameter are set in a matrix of inequigranular anhedral quartz 0.1

TABLE 26.—*Chemical analyses, modes, and norms of Henderson Gneiss*—Continued

to 2 mm in diameter and plagioclase ($An_{30}$) as much as 0.5 mm in diameter and randomly oriented irregular undeformed flakes of brown biotite as much as 1 mm long. Muscovite forms stubby primary books and scattered secondary small flakes and aggregates in feldpsar. Microcline porphyroblasts are surrounded and penetrated by myrmekite. Outcrop about 40 feet above south shore of Rhodiss Lake, 1.6 mile S. 75° E. of south end of Huffman Bridge (area I–9, pl. 1).

3. Cataclastic Henderson Gneiss. Strongly foliated and lineated nonlayered fine-grained muscovite-biotite-plagioclase-quartz gneiss containing augen of pink potassic feldspar as much as 1 cm long elongated parallel to foliation. Texture is strongly cataclastic. Rock consists of crushed porphyroclasts of microcline and microperthite set in recrystallized matrix of granoblastic grains of quartz and plagioclase ($An_{15}$) and well-alined flakes of synkinematic and postkinematic greenish-brown biotite and muscovite 0.01 to 0.1 mm long. Epidote forms granules and small sieve-textured grains. One sieve-textured grain of garnet 0.25 mm in diameter appears to be a porphyroblast. Roadcut on east side of dirt road on northeast side of Upper Creek, 0.9 mile S. 89° E. of Fairview Church (area F–8, pl. 1).
4. Henderson Gneiss. Strongly foliated and lineated fine-grained biotite quartz monzonite gneiss containing lentil-shaped potassic feldspar aggregates elongated parallel to lineation. Texture is cataclastic. Rock consists of scattered porphyroclasts of microcline as much as 5 mm in diameter set in an inequigranular mosaic of recrystallized granoblastic quartz, plagioclase ($An_{20}$), and microcline 0.1 to 1.0 mm in diameter and ragged grains of synkinematic and postkinematic olive-green biotite and muscovite 0.05 to 0.5 mm long. Garnet forms sieve-textured porphyroblasts as much as 5 mm in diameter, associated with irregular grains of green hornblende as much as 0.5 mm in diameter. Outcrop on south shore of Lake James, 0.4 mile S. 30° W. of Rock Hill Church (area D–9, pl. 1).

grains have allanite cores. Some specimens contain small skeletal garnets.

The feldspar porphyroclasts are chiefly microcline and microcline-microperthite, but some twinned plagioclase porphyroclasts occur. The potassic feldspar porphyroclasts have ragged outlines and are jacketed by mosaics of quartz, plagioclase, myrmekite, and recrystallized potassic feldspar in which the grains are slightly coarser than in the adjacent groundmass (fig. 80*B*). Commonly, the porphyroclastic grains have been broken and the fractures healed by the mosaic.

The chief accessory minerals in the gneiss are magnetite, sphene, apatite, and light-pink or orange zircon. The zircon occurs in slender euhedral prisms, some as long as 0.5 mm.

Chemical analyses, modes, and norms of typical specimens of Henderson Gneiss and average modes of Henderson Gneiss and of biotite gneisses transitional into Henderson Gneiss are given in table 26. Ranges in proportions of the principal minerals are indicated by modal variation diagrams. (figs. 81 and 82).

The Henderson Gneiss contains a much higher proportion of potassic feldspar than the layered biotite gneisses, but the proportions of the other minerals are about the same. Chemically, the Henderson closely resembles biotite gneiss of similar $SiO_2$ content, but it contains appreciably more $K_2O$ and therefore has a higher $K_2O/Na_2O$ ratio and a higher ratio of total alkalies to alumina.

The gradational contacts between the Henderson Gneiss and the enclosing rocks and the compositional similarity between them suggest either that the Henderson Gneiss may have originated by recrystalli-

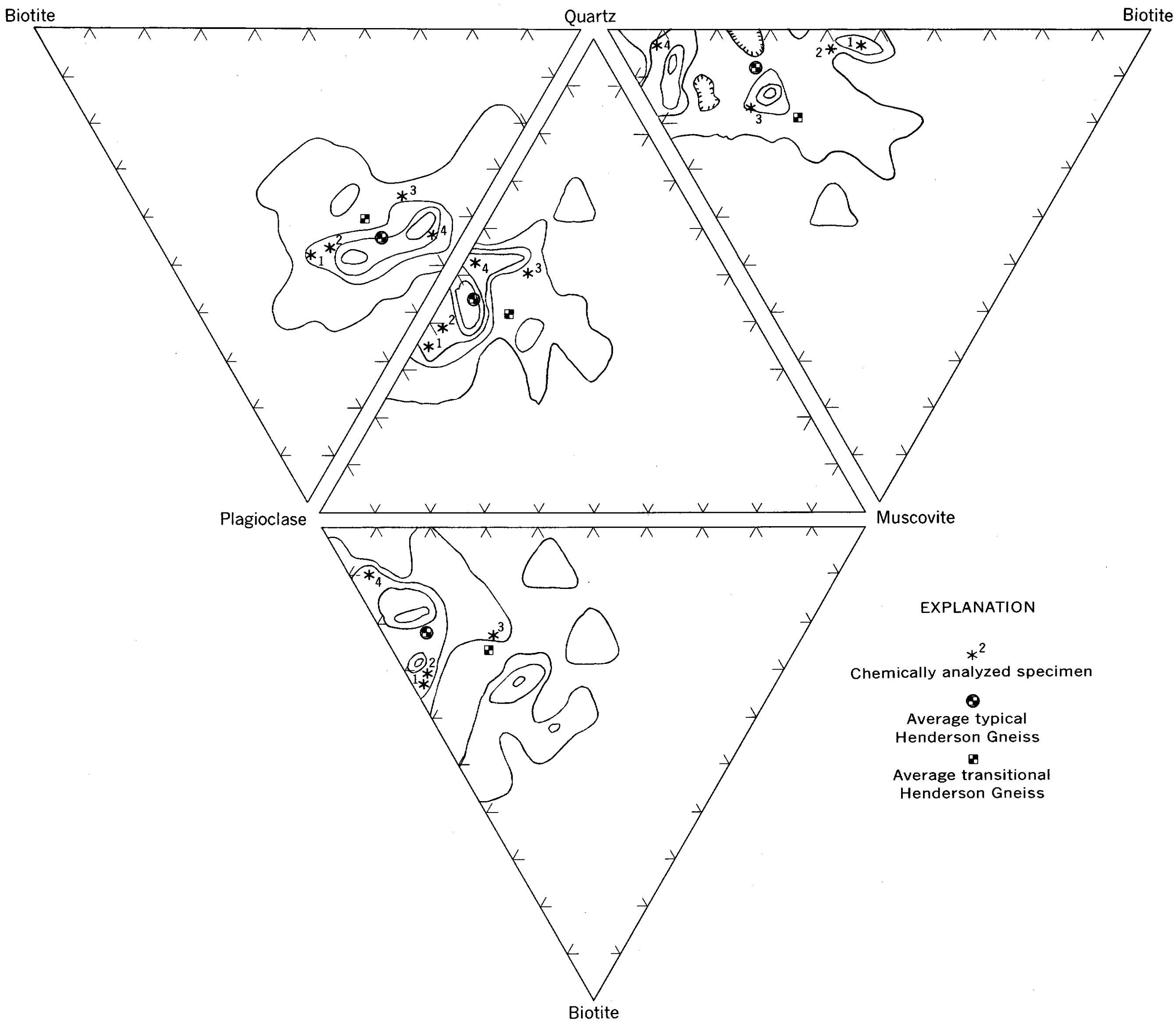



FIGURE 81.—Proportions of quartz, plagioclase, biotite, and muscovite in Henderson Gneiss and biotite gneisses transitional into Henderson Gneiss. Based on point counts of analyzed specimens and counts of 50 random grains in each of 13 thin sections of typical Henderson Gneiss and 11 thin sections of transitional gneisses. Contours 4, 7, 10, 14, and 18 percent. Numbers of analyzed specimens refer to analyses in table 26.

zation of slightly more potassium-rich rocks in the original volcanic and sedimentary sequence or that it was the product of the addition of $K_2O$ to parts of the stratified sequence. The small bodies of unsheared Henderson Gneiss outside the belt of polymetamorphic rocks along the Brevard fault zone are all closely associated with rocks similar to the Toluca Quartz Monzonite. In the Causby quarry, on the east side of Hunting Creek, 0.9 mile S. 10° W. of the confluence of the Catawba River and Johns River (area H–9, pl. 1), biotite and biotite-amphibole gneiss adjacent to a small body of biotite granodiorite has been converted to augen gneiss closely resembling the Henderson Gneiss, a fact suggesting that the Henderson Gneiss may be related to the Paleozoic granitic rocks. The larger and more abundant bodies of Henderson Gneiss in the belt of polymetamorphic rocks along the Brevard fault zone are not closely associated with the granitic rocks, although they contain sheared pegmatite bodies containing heayy-mineral suites similar to those of the granitic rocks. The shape of the Henderson bodies in the belt of polymetamorphic rocks has apparently been greatly modified by shearing; many of these rocks

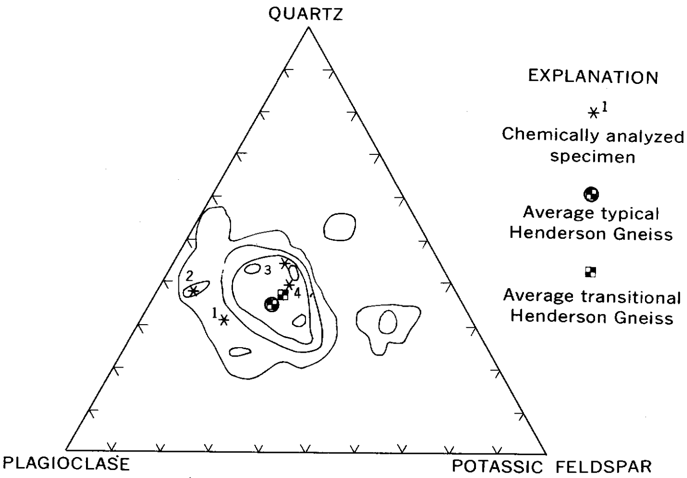



FIGURE 82.—Proportions of quartz, plagioclase, and potassic feldspar in Henderson Gneiss and biotite gneisses transitional into Henderson Gneiss. Based on point counts of analyzed specimens and counts of 50 random grains in each of 13 thin sections of typical Henderson Gneiss and 11 thin sections of transitional gneisses. Contours 4, 7, 10, and 14 percent. Numbers of analyzed specimens refer to analyses in table 26.

and the rocks now associated with them may be exotic and may have been carried for long distances during early movement along the Brevard fault zone. It therefore seems unlikely that the origin of the Henderson Gneiss can be satisfactorily determined from the evidence available in the Grandfather Mountain area. This problem might better be attacked farther southwest, where Keith (1905, 1907b) has mapped large areas of Henderson Gneiss many miles southeast of the Brevard fault zone.

Keith (1905, 1907b) believed that the Henderson Gneiss was of early Precambrian (Archean) age, largely because of its relations to layered gneisses which he also considered to be of early Precambrian age. The enclosing rocks are now believed to be of late Precambrian or early Paleozoic age. Field relations in the Grandfather Mountain area suggest that the Henderson is older than or related to the granitic rocks of early or middle Paleozoic age, and it is therefore believed to be of late Precambrian or early Paleozoic age.

### GRANITIC ROCKS AND MIGMATITE

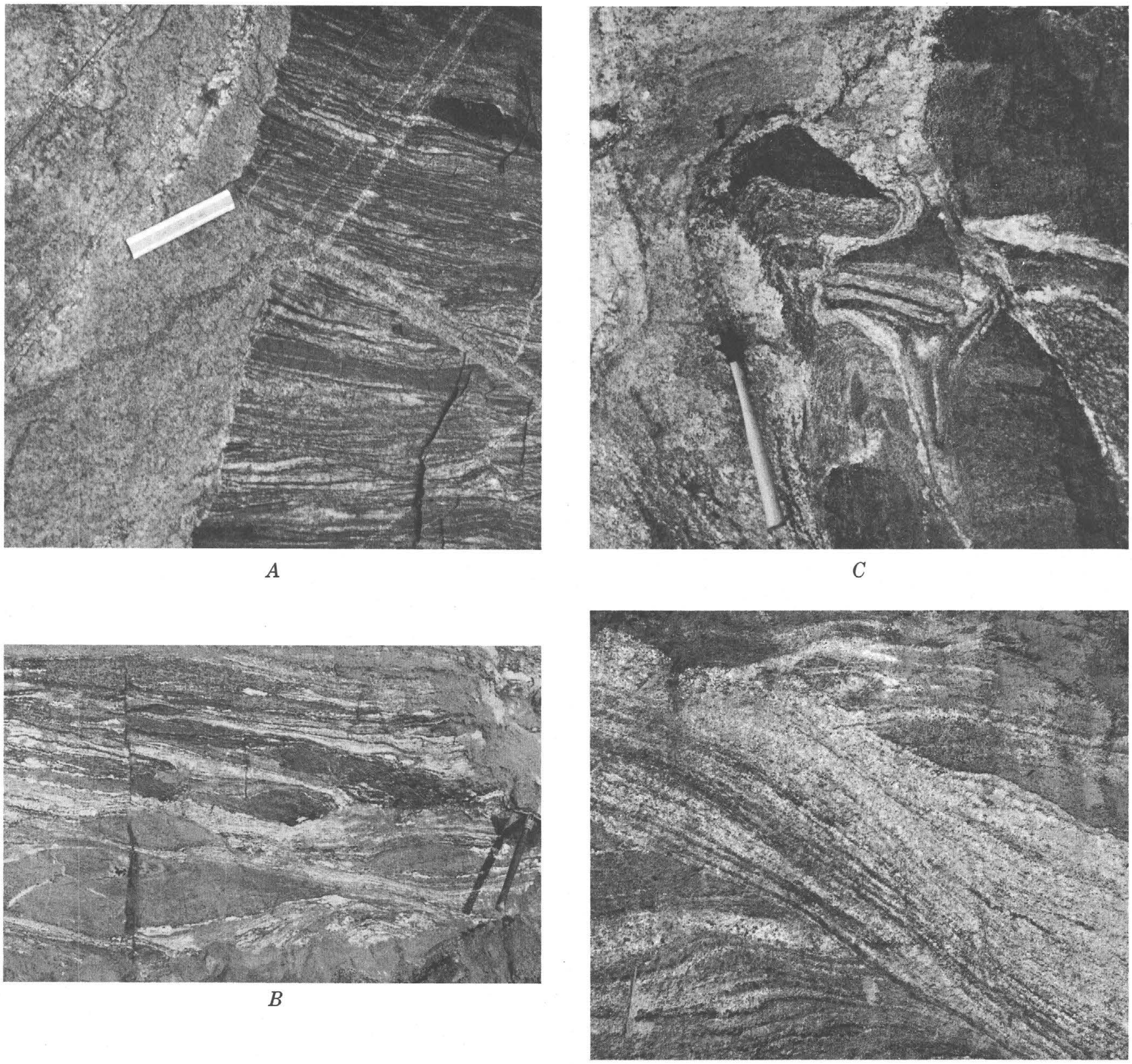
The layered rocks of the Inner Piedmont are invaded by extensive concordant bodies of granitic rocks ranging in composition from quartz diorite to quartz monzonite. The granitic bodies are flanked by diffuse zones of migmatite in the wallrocks, and their emplacement has been accompanied by emplacement of myriad dikes, pods, and knots of pegmatite in the enclosing rocks. Strangely, granitic bodies have not been found in the belt of sheared and retrogressively metamorphosed rocks along the Brevard fault zone, although pegmatites are common in this belt and the large bodies of Henderson Gneiss are confined to it.

The granitic rocks closely resemble the Toluca Quartz Monzonite of the Shelby area (Griffitts and Overstreet, 1952; Overstreet and others, 1963) in their general appearance, distribution, structural relations, and accessory mineral suite, but they differ from the Toluca in the preponderance of quartz diorite and granodiorite over quartz monzonite. The name Toluca Quartz Monzonite is therefore not formally applied to the granitic rocks in the Grandfather Mountain area, although the main belt of granitic rocks between Lenoir and Morganton can be traced southwestward from Morganton and is apparently continuous with rocks identified as Toluca by Overstreet and Griffitts (1955) along U.S. Highway 64 in the southeastern part of McDowell County.

#### MEGASCOPIC FEATURES AND FIELD RELATIONS

The granitic rocks are generally medium to coarse grained and light to medium gray; they are inequigranular and locally contain porphyroblasts of microcline as much as 2 cm long. Many of them are strongly gneissic, but in some of the larger bodies they are massive or very weakly foliated. Foliation is commonly defined by subparallel dark biotitic streaks and by faint parallelism of micas. The foliation is generally rudely parallel to foliation and layering in the enclosing rocks, but in some place cross-cutting dikes of granitic rock have planar flow structures parallel to their walls (fig. 83). Inclusions of amphibolite and hornblende gneiss are common in the more massive granitic rocks and are generally surrounded by thin reaction rims of biotite schist. Inclusions of other rocks are rare.

Most of the larger granitic bodies are concordant or semiconcordant with the structures in the enclosing rocks, but locally the contacts are sharply discordant, and small disordant dikes are common (fig. 83*A* and *D*). Locally, the contacts are sharp, but the larger granitic bodies more commonly pass into the surrounding rocks through broad migmatitic zones composed of strongly gneissic granitic rock interleaved with folia of schist and layered biotite gneiss and containing abundant pods of amphibolite and biotite gneiss (fig. 83*B*). In the Lenoir quadrangle, the boundary between migmatitic granitic rocks and migmatitic wallrocks has been arbitrarily drawn



A

C

B

D

FIGURE 83.—Migmatite and granitic rocks. *A*, Contact of discordant dike of fine-grained muscovite-biotite quartz monzonite and related pegmatite cutting strongly foliated migmatitic biotite gneiss containing a pod of dark amphibolite. Gneiss is feldspathized along joints extending into it from the granitic rock. Fresh rock exposed in abandoned quarry 200 feet southwest of U.S. Highway 321, 0.1 mile north of southern city limits of Lenoir (area J–7, pl. 1). *B*, Pods of amphibolite in migmatite composed of interleaved layered biotite gneiss and streaky amphibole-bearing biotite quartz diorite. Saprolite exposures on North Carolina Highway 18 bypass around Morganton, about 0.1 mile north of intersection with North Carolina Highway 181, on east bank of Catawba River (south edge of area G–9, pl. 1). *C*, Blocks of layered biotite-hornblende gneiss and amphibolite in streaky inequigranular quartz diorite or granodiorite and pegmatite. Saprolite exposure in roadcut on northwest side of North Carolina Highway 18, 0.2 mile southeast of the Burke-Caldwell County Line (area H–8, pl. 1). *D*, Thinly layered fine-grained biotite and biotite-hornblende gneiss cut by dike of streaky biotite quartz diorite having flow structure parallel to its walls. The layered gneiss contains concordant lenses and pods of granitic rock, some of which are cut by the dike and some of which may have been fed by the dike. Saprolite exposure in roadcut on north side of county road, 0.4 mile S. 17° E. of Littlejohn Church (area H–8, pl. 1). Width of view about 1.5 feet.

where the granitic rocks seem to make up about half the total volume.

Muscovite-quartz-microcline-plagioclase pegmatite forms dikes, pods, and knots in the granitic rocks, migmatites, and in the nonmigmatitic rocks throughout the area, but the pegmatite bodies are most common in the migmatite zones adjacent to the granitic bodies. The pegmatite bodies range from a few inches to several tens of feet in thickness, and some are several hundred feet long. Some of them are closely associated with and pass irregularly into granitic rocks, but the relation between the granitic rocks and other pegmatites cannot be conclusively demonstrated, although all the pegmatite bodies sampled contain heavy-mineral suites similar to those of the granitic rocks.

## PETROGRAPHY

The granitic rocks typically consist of an unsheared inequigranular mosaic of anhedral grains of quartz, plagioclase, and microcline in various proportions and of scattered rudely alined or randomly oriented flakes of biotite and muscovite. The plagioclase is most commonly well-twinned and faintly zoned oligoclase or sodic andesine in irregular grains 1 to 3 mm in diameter. It also forms porphyroblasts as much as 1 cm in diameter in a few specimens. Compositions of plagioclase in granitic rocks of various compositions are shown in figure 84. Plagioclase grains in many specimens are rimmed by thin films of albite. Quartz in the mosaic is generally smaller than and interstitial to the plagioclase, but in some specimens it also forms segregation laminae parallel to the foliation. Microcline forms small irregular grains intergrown with and intersitial to quartz and plagioclase in the mosaic, and in some specimens it occurs as porphyroblasts as much as 2 cm long, similar to those in the Henderson Gneiss. Myrmekite reaction rims are common at contacts between microcline and plagioclase and form wartlike projections into the potassic feldspar.

Stubby irregular flakes of brown or reddish-brown biotite 0.1 to 1.0 mm long are randomly oriented or rudely alined. Commonly, they are arranged in streaks defining a rude foliation. Primary muscovite is subordinate to biotite in most rocks; it forms slender flakes interleaved with biotite and commonly seems to have crystallized somewhat later. Secondary muscovite forms fringes on the ends of biotite books and skeletal grains and sericite aggregates replacing feldspars. Colorless epidote with moderate to high birefringence occurs as skeletal aggregates and clusters of anhedral grains, commonly in or adjacent

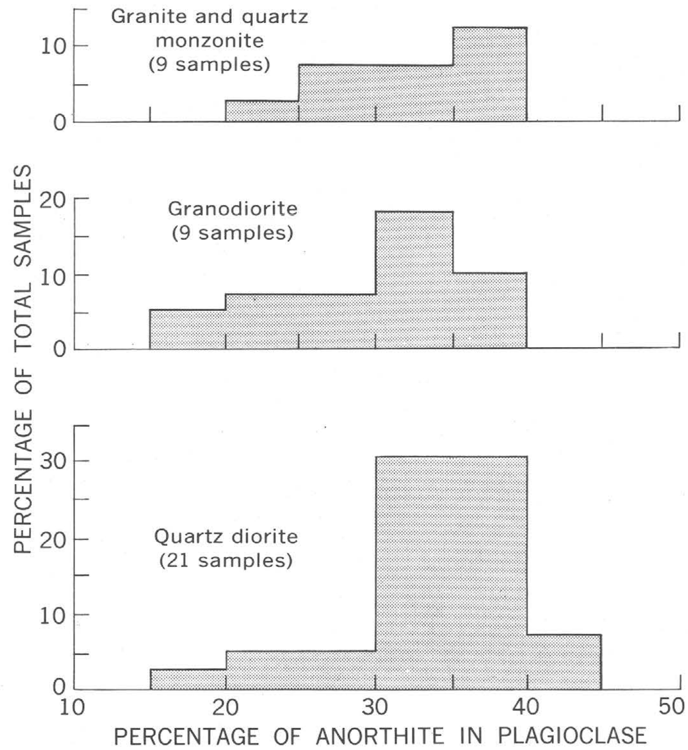



FIGURE 84.—Histograms showing percentage of granitic rocks containing various compositions of plagioclase whose indices of refraction were determined by oil-immersion methods. Some rocks contain plagioclase with 10-percent or greater range of anorthite content and plot in several adjacent bars; the total therefore is more than 100 percent. Based on 39 samples.

to biotite. Some epidote grains have cores of metamict allanite. Garnet forms subhedral to anhedral equidimensional porphyroblasts as much as 0.25 mm in diameter in some specimens. Aggregates of small anhedral grains of sphene commonly lie along biotite cleavages. Magnetite in scattered anhedral grains and aggregates and apatite in small blunt prisms are common minor accessories. A few of the granitic rocks contain scattered skeletal grains of green hornblende partly replaced by aggregates of biotite. Chlorite, carbonates, and clay minerals are common in slightly altered rocks. In addition to epidote, apatite, garnet, and magnetite, heavy-mineral suites panned from saprolitized granitic rocks contain abundant stubby prisms of colorless or light-pink zircon, anhedral orange grains of monazite, scattered euhedral pyramids of light-amber xenotime, and a few prisms of black tourmaline.

The migmatites differ from the other granitic rocks only in their more gneissic textures and in their higher contents of biotite and lower contents of potassic feldspar. All gradations in texture between the layered biotite gneiss and the granitic rock are found. Mineralogically, they are apparently identical with the granitic rocks.

COMPOSITION

Chemical analyses, modes, and norms of selected specimens of granitic rocks are given in table 27. Similar data for the migmatic rocks are given in table 28. A combined modal variation diagram showing ranges in proportions of quartz, plagioclase, biotite, and muscovite in granitic rocks and in migmatitic gneisses related to them is given in figure 85.

The granitic rocks and the migmatites associated with them closely resemble the layered gneisses in their modal composition. They differ from the layered gneisses only in containing slightly higher proportions of potassic feldspar and slightly lower proportions of biotite. A modal variation diagram showing the range in relative proportions of quartz, plagioclase, and potassic feldspar in the granitic rocks and migmatites is given in figure 86.

A silica variation diagram (fig. 87) of the analyses from tables 27 and 28 shows that the similarity in

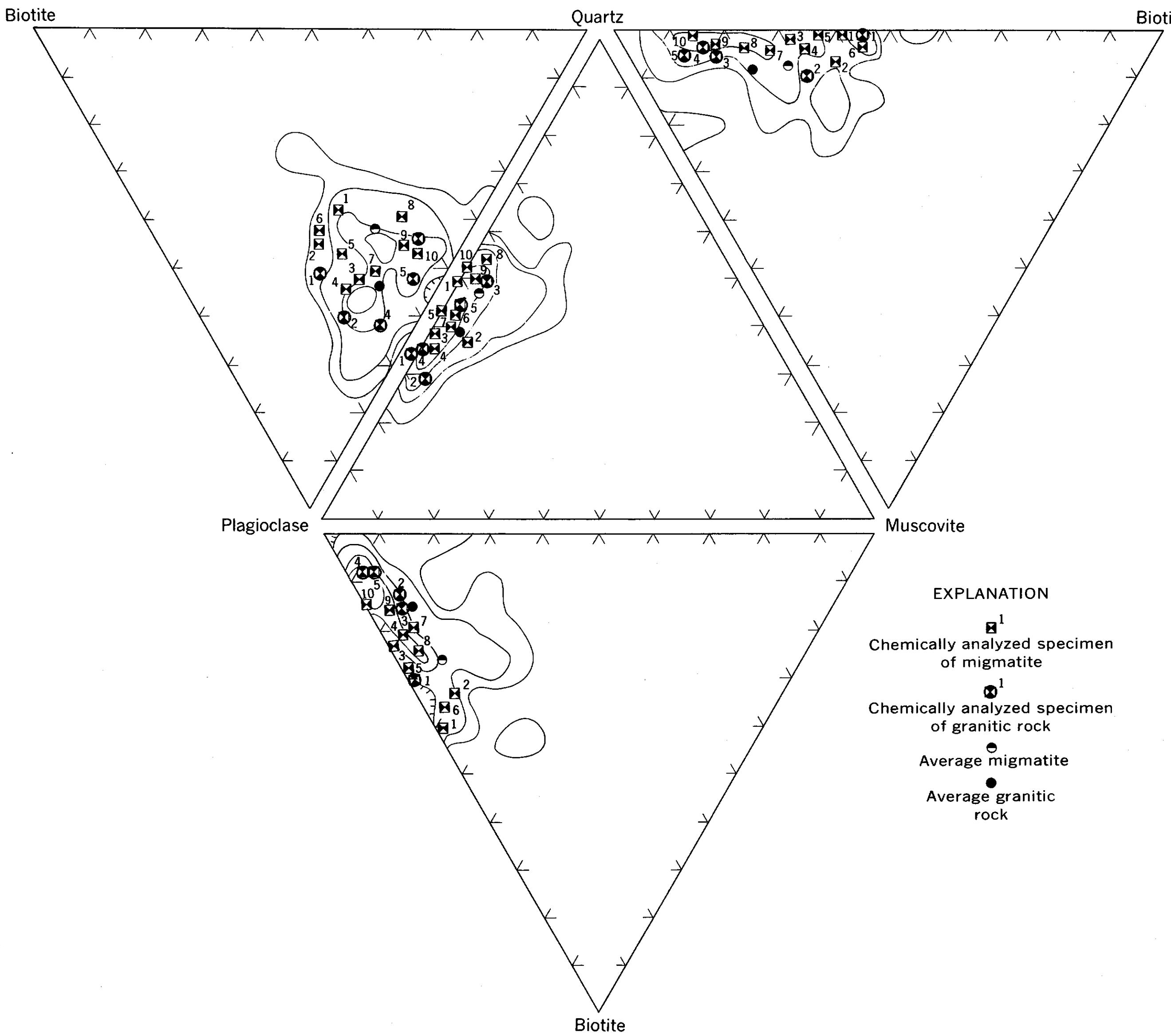



FIGURE 85.—Proportions of quartz, plagioclase, biotite, and muscovite in migmatitic biotite gneisses and granitic rocks related to the Toluca Quartz Monzonite. Based on counts of 50 random grains in each of 24 thin sections of migmatitic gneisses and each of 26 thin sections of granitic rocks. Contours 2, 4, 8, 12, and 16 percent. Numbers of migmatite specimens refer to analyses in table 28; numbers of granitic specimens, to analyses in table 27.

TABLE 27.—*Chemical analyses, modes, and norms of granitic rocks probably related to the Toluca Quartz Monzonite*

[Chemical analyses of samples 1 and 3 determined by standard chemical methods by D. F. Powers and C. L. Parker, U.S. Geol. Survey. Other analyses, by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey. Nd, not determined. Modes of analyzed specimens determined by point counts; P, present but not intersected in counting; Tr, trace. Major oxides and CIPW norms given in weight percent; modes, in volume percent]

| | 1 | 2 | 3 | 4 | 5 | 6 [1] |
|---|---|---|---|---|---|---|
| Field No. | 2583(L) | 30–1395E3 | B | A–607 | 31–1669B | |
| Laboratory No. | I–4060 | 160146 | H–3431 | 161249 | 160147 | |
| **Major oxides** | | | | | | |
| $SiO_2$ | 64.69 | 68.5 | 71.83 | 74.2 | 75.5 | |
| $Al_2O_3$ | 16.53 | 17.7 | 14.35 | 15.3 | 14.4 | |
| $Fe_2O_3$ | .53 | .25 | .55 | .30 | .14 | |
| FeO | 3.89 | .99 | 1.76 | .64 | .62 | |
| MgO | 1.79 | .46 | 1.14 | .42 | .35 | |
| CaO | 4.32 | 1.4 | 2.04 | 3.2 | 1.8 | |
| $Na_2O$ | 2.98 | 3.8 | 3.38 | 4.9 | 3.2 | |
| $K_2O$ | 3.32 | 6.0 | 3.80 | .62 | 4.0 | |
| $H_2O+$ | .64 | .60 | .54 | .70 | .45 | |
| $H_2O-$ | .04 | .03 | .04 | .02 | .04 | |
| $TiO_2$ | .70 | .22 | .19 | .09 | .11 | |
| $P_2O_5$ | .19 | .05 | .08 | .03 | .01 | |
| MnO | .10 | .00 | .06 | .02 | .00 | |
| $CO_2$ | .10 | .07 | .09 | .05 | .05 | |
| F | .07 | Nd | Nd | Nd | Nd | |
| Cl | .02 | Nd | Nd | Nd | Nd | |
| Total | 99.87 | 100 | 99.85 | 100 | 100 | |
| **Modes** | | | | | | |
| Quartz | 24 | 17 | 37 | 32 | 32 | 29.8 |
| Plagioclase | 46 | 40 | 35 | 60 | 39 | 44.8 |
| Potassic feldspar | 8 | 31 | 18 | | 22 | 12.5 |
| Biotite | 20 | 8 | 7 | 6 | 4 | 8.9 |
| Muscovite | .2 | 2.8 | 3 | 1.7 | 2.5 | 3.8 |
| Hornblende | .3 | | | | | Tr |
| Epidote | .8 | .1 | .7 | 1.1 | P | .6 |
| Apatite | .2 | P | | | .2 | Tr |
| Chlorite | | | | .2 | | Tr |
| Zircon | | .1 | | | P | Tr |
| Carbonate | | .2 | .1 | | | Tr |
| Sphene | 1.0 | P | | .1 | .3 | Tr |
| Opaque minerals | | P | | .1 | | Tr |
| Points counted | 600 | 900 | 700 | 600 | 600 | |
| **CIPW norms** | | | | | | |
| Q | 21.24 | 19.54 | 30.88 | 35.72 | 37.00 | |
| C | 1.06 | 2.70 | 1.37 | .94 | 1.68 | |
| Or | 19.62 | 35.45 | 22.45 | 3.66 | 23.63 | |
| Ab | 25.06 | 32.14 | 28.58 | 41.44 | 27.06 | |
| An | 19.10 | 6.18 | 9.03 | 15.36 | 8.55 | |
| Hl | .03 | | | | | |
| En | 4.46 | 1.14 | 2.84 | 1.05 | .87 | |
| Fs | 5.73 | 1.25 | 2.58 | .82 | .84 | |
| Mt | .77 | .36 | .80 | .44 | .20 | |
| Il | 1.33 | .42 | .36 | .17 | .21 | |
| Ap | .45 | .12 | .19 | .07 | .02 | |
| Fr | .13 | | | | | |
| Cc | .23 | .16 | .20 | .11 | .11 | |

[1] Total of 50 random grains counted in each of 26 thin sections of granitic rocks.
NOTE.—Minor-element analyses for samples 1 and 3 given in table 1.

1. Coarse-grained well-foliated streaky biotite granodiorite containing pods of coarse biotite pegmatite. Rock consists of unsheared inequigranular mosaic of anhedral grains of quartz as much as 4 mm long, plagioclase ($An_{30-40}$) 1 to 5 mm long, and rudely alined irregular grains of brown biotite 0.5 to 2 mm long. Microcline forms small blebs in plagioclase and irregular grains interstitial to quartz and plagioclase; it is fringed with myremkite at contacts with plagioclase. Larger plagioclase grains seem to be porphyroblastic. Green hornblende occurs in scattered skeletal grains as much as 1 mm in diameter partly replaced by biotite. Epidote forms small scattered grains containing low birefringent brown cores. Muscovite occurs as very sparse secondary flakes in feldspar. Abandoned quarry on northwest side of McGalliard Creek, 2.0 miles N. 78° E. of Drexel (area I–9, pl. 1).
2. Coarse-grained streaky biotite quartz monzonite granodiorite containing scattered augen of potassic feldspar as much as 2 cm long. Rock consists of an unsheared inequigranular mosaic of anhedral grains of quartz as much as 5 mm long and plagioclase ($An_{20-30}$) as much as 5 mm long, intergranular films and irregular grains of microcline interstitial to quartz and plagioclase, and rudely alined irregular flakes of brown biotite 0.2 to 2 mm long. Plagioclase also forms zoned porphyroblasts as much as 1 cm in diameter. Muscovite occurs as irregular flakes contemporaneous with biotite, as fringes on biotite, and as sericite in feldspars. Epidote forms prismatic grains as much as 0.2 mm long. North wall of Causby quarry, 0.9 mile S. 10° W. of confluence of Catawba River and Johns River (area H–9, pl. 1).
3. Biotite granodiorite. Medium-grained faintly foliated equigranular biotite granodiorite containing scattered porphyroblasts of potassic feldspar 1 to 3 inches long and a few streaks of coarse biotite schist. Rock consists of unsheared mosaic of anhedral grains of quartz and plagioclase ($An_{23-28}$) 2 to 3 mm in diameter and microcline as much as 5 mm in diameter and ragged flakes of dark-brown biotite 0.5 to 1.5 mm long interstitial to and included in quartz and feldspar. Plagioclase is twinned and faintly zoned and commonly passes into myrmekite at contacts with microcline. Muscovite occurs in ragged primary flakes and as secondary sericite in feldspars. Epidote forms irregular poikilitic grains and aggregates, commonly in biotite. Roadcut on east side of road between Oak Hill and Willow Tree school just south of bridge over Canoe Creek, 1.0 mile S. 27° W. of Oak Hill school (area F–9, pl. 1).
4. Medium- to coarse-grained light-colored quartz diorite containing pods of amphibolite. Rock consists of unsheared inequigranular mosaic of anhedral grains of quartz and plagioclase ($An_{30-36}$) as much as 5 mm in diameter and irregular nonalined flakes of partly chloritized brown biotite as much as 2 mm long. Muscovite forms fringes on biotite and felted sericite aggregates in feldspars. Albite forms fringes on plagioclase. Epidote forms irregular skeletal grains, some containing low birefringent brown cores, chiefly in biotite. Roadcut on south side of road, 2.0 miles S. 60° E. of school at Gamewell (area I–8, pl. 1).
5. Medium-grained streaky biotite quartz monzonite in 5-foot-long pod in layered amphibole-biotite gneiss. Rock consists of unsheared mosaic of anhedral grains of quartz 1 to 3 mm in diameter, plagioclase ($A_{30-34}$) 1 to 5 mm in diameter, and microcline 1 to 5 mm in diameter and scattered irregular nonalined grains of brown biotite as much as 1 mm long. Sieve-textured porphyroblasts of microcline contain abundant inclusions of quartz and plagioclase. Myrmekite is common at contacts between microcline and plagioclase. Albite forms fringes on plagioclase. Muscovite forms spongy skeletal grains, fringes on biotite and sericite aggregates in feldspars, and is largely secondary. Roadcut on north side of county road on west side of Smoky Creek, 1.4 miles N. 55° E. of north end of Huffman Bridge (area I–9, pl. 1).

TABLE 21.—*Chemical analyses, modes, and norms of granitic rocks probably related to the Toluca Quartz Monzonite*—Continued

mineral composition is concomitant with a close similarity in chemical composition. There is apparently no demonstrable difference in proportions of the major oxides between migmatitic rocks and nomigmatitic layered biotite gneisses of equivalent $SiO_2$ content. The granitic rocks also closely resemble the nomigmatitic biotite gneisses, but they have a generally slightly higher alumina and slightly lower total iron content, and many of them have a somewhat higher $K_2O$ content than nonmigmatitic gneisses of equivalent $SiO_2$ content. Figure 88 shows that the molecular proportions of $Al_2O_3$, CaO, and $K_2O$ + $Na_2O$ in the granitic rocks and migmatites are nearly identical with those in the layered biotite gneisses.

The $Na_2O:K_2O$ ratio (fig. 89) of most of the migmatites is in the field of biotite gneisses, but a few of the migmatites and most of the granitic rocks have $Na_2O:K_2O$ ratios less than one. The only granitic rock having a $Na_2O:K_2O$ ratio greater than one is a plagioclase-rich quartz diorite lacking potassic feldspar and containing abundant inclusions of amphibolite.

## ORIGIN

The close chemical similarity between the granitic rocks and the enclosing biotite gneisses suggests that the granitic rocks could have been derived from the country rocks with only minor changes in bulk composition. The generally conformable habit of the larger granitic bodies and the gradational contacts from granitic rocks through migmatites into non-

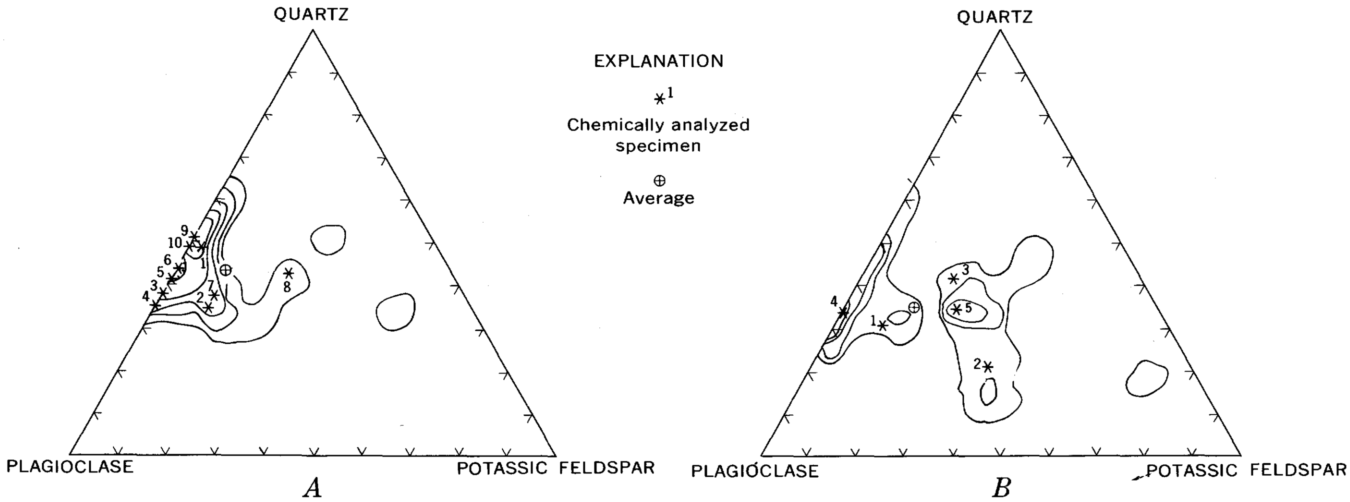



FIGURE 86.—Proportions of quartz, plagioclase, and potassic feldspar in migmatitic biotite gneisses and granitic rocks. *A*, Migmatitic biotite gneisses. Based on point counts of analyzed rocks and counts of 50 random grains in each of 14 other thin sections. Contours 4, 8, 12, 16, and 25 percent. *B*, Granitic rocks. Based on point counts of analyzed rocks and count of 50 random grains in each of 19 other than sections. Contours 4, 8, 12, and 23 percent. Numbers of analyzed specimens refer to analyses in tables 27 and 28.

migmatitic gneisses suggests that a large part of the granitic material may have formed by in situ recrystallization of the biotite gneiss of appropriate composition, perhaps with local addition of $K_2O$. Discordant dikes of pegmatite and granitic rocks are common, however, and the larger granitic bodies locally contain rotated inclusions of layered amphibolite and amphibole gneiss and in some places have sharp discordant contacts, showing that at least some of the granitic material moved. The absence of migmatites and granitic rocks comparable in composition with the hornblende gneisses and amphibolites and the preponderance of these rocks as inclusions is compatible with the hypothesis of origin of the granitic rocks by recrystallization and partial anatexis of the biotite gneisses, the interlayered hornblendic rocks being left as undigested relics.

The general structural conformity between the granitic rocks and migmatites and the enclosing rocks, the similarity in mineralogy and mineral assemblages between the granitic rocks, migmatites, and enclosing rocks, and the absence of metamorphic effects and chilled margins at the contacts of the granitic bodies indicate that the granitic rocks formed by recrystallization and partial anatexis of the country rocks during the climax of the high-grade regional metamorphism. The granitic bodies in this part of the Inner Piedmont belt have all the characteristics of the "plutons of the catazone" described by Buddington (1959).

There is as yet no unanimity of opinion on the correlation between metamorphic facies and the temperature and pressure ranges that they represent. Luth, Jahns, and Tuttle (1964, p. 760) indicated that water-saturated granitic magma could be formed by fusion of rocks of appropriate composition at temperatures of about 625° C under $P_{H_2O}$ of 10 kilobars, corresponding to depths of 35 to 40 km (kilometers), if $P_{H_2O}$ is assumed equal to lithostatic load (Hamilton, 1963, fig. 78), or at temperatures of about 700°C at depths of about 5 km. Recent recalculation of the aluminum silicate triple point based on new experimental data (Newton, 1966) indicated that sillimanite is stable in the temperature range 600° to 700°C at pressures between about 2 and 7 kilobars, corresponding to depths of 7.5 to 26.5 km. These data probably provide a reasonable estimate of the possible range of temperatures and pressures that prevailed during the climax of regional metamorphism and the formation of the granitic rocks in this part of the Inner Piedmont.

## AGE

No stratigraphic evidence or geochronologic data are available that would unequivocally establish the age of the granitic rocks in the Inner Piedmont belt in the Grandfather Mountain area. The Toluca Quartz Monzonite, with which the granitic rocks in the Grandfather Mountain area are believed to be correlative, has been assigned an Ordovician age by Overstreet and Griffiths (1955) on the basis of lead-

TABLE 28.—*Chemical analyses, modes, and norms of migmatitic biotite gneisses*

[Analysis of sample 1 determined by standard chemical methods by D. F. Powers, U.S. Geol. Survey. Other analyses, by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey. Nd, not determined. Modes of analyzed specimens determined by count of 600 points; P, present but not intersected in counting; Tr, trace. Major oxides and CIPW norms given in weight percent; modes, in volume percent]

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 [1] |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Field No. | 2028-L | 30-1395F-1 | 28-1111A | 33-2146B | A-580 | 32-2035B | 33-2146D | 30-1379B | 28-1011C | 29-1211a | |
| Laboratory No. | I-4062 | 161334 | 160144 | 161245 | 161248 | 160155 | 161246 | 160145 | 161243 | 161244 | |
| **Major oxides** | | | | | | | | | | | |
| $SiO_2$ | 63.18 | 66.6 | 66.9 | 68.0 | 69.6 | 69.8 | 70.4 | 73.8 | 74.0 | 75.8 | |
| $Al_2O_3$ | 15.38 | 15.1 | 17.0 | 14.9 | 14.9 | 14.6 | 14.9 | 13.8 | 13.2 | 12.9 | |
| $Fe_2O_3$ | 1.75 | .66 | .90 | 1.3 | 1.2 | .75 | .74 | .35 | .84 | .56 | |
| FeO | 3.96 | 3.4 | 2.7 | 3.4 | 2.8 | 3.3 | 2.6 | 1.4 | 1.6 | 1.4 | |
| MgO | 2.88 | 1.0 | 1.2 | 1.7 | 1.8 | 1.6 | 1.2 | .55 | 1.4 | 1.0 | |
| CaO | 5.08 | 3.6 | 4.2 | 3.3 | 2.9 | 2.6 | 2.6 | 1.2 | 3.0 | 4.0 | |
| $Na_2O$ | 2.67 | 3.0 | 4.0 | 3.4 | 3.6 | 3.1 | 3.3 | 2.9 | 3.8 | 2.8 | |
| $K_2O$ | 2.84 | 3.9 | 1.6 | 2.1 | 2.0 | 2.3 | 3.3 | 4.8 | 1.0 | .78 | |
| $H_2O+$ | .82 | .66 | .80 | .84 | .85 | .85 | .69 | .56 | .48 | .77 | |
| $H_2O-$ | .05 | .05 | .02 | .07 | .06 | .04 | .04 | .03 | .02 | .05 | |
| $TiO_2$ | .56 | .86 | .36 | .20 | .40 | .62 | .48 | .30 | .32 | .29 | |
| $P_2O_5$ | .12 | .53 | .21 | .20 | .09 | .24 | .12 | .12 | .09 | .09 | |
| MnO | .12 | .07 | .05 | .08 | .08 | .06 | .05 | .02 | .06 | .04 | |
| $CO_2$ | .15 | .48 | <.05 | <.05 | <.05 | <.05 | <.05 | <.05 | <.05 | <.05 | |
| Cl | .01 | Nd | Nd | Nd | Nd | Nd | Nd | Nd | Nd | Nd | |
| F | .05 | Nd | Nd | Nd | Nd | Nd | Nd | Nd | Nd | Nd | |
| Total | 99.60 | 100 | 100 | 100 | 100 | 100 | 100 | 100 | 100 | 100 | |
| **Modes** | | | | | | | | | | | |
| Quartz | 33 | 25 | 32 | 29 | 33 | 31 | 32 | 36 | 43 | 45 | 34.7 |
| Plagioclase | 33 | 38 | 51 | 52 | 45 | 41 | 44 | 29 | 42 | 44 | 36.0 |
| Potassic feldspar | 2.2 | 8 | .3 | | .02 | | 9 | 23 | | | 7.2 |
| Biotite | 23 | 22 | 15 | 15 | 19 | 25 | 12 | 10 | 9 | 8 | 14.9 |
| Muscovite | | 4 | 1.3 | 2.3 | .3 | 2.5 | 1.7 | 1.8 | 2.0 | | 4.4 |
| Amphibole | 4 | | | | .7 | | | | 1.0 | 1.0 | .3 |
| Epidote | 4 | .2 | .3 | 1.0 | 1.0 | | 1.0 | | 2.2 | 1.6 | 1.1 |
| Apatite | .2 | 1.0 | .3 | | .3 | .4 | | .4 | | | Tr |
| Chlorite | | | P | | | | | | .3 | | Tr |
| Zircon | | | .1 | | | .1 | | | | .2 | Tr |
| Carbonate | .3 | .8 | .4 | | | .2 | .2 | | | | Tr |
| Sphene | | .5 | | | | | .2 | .1 | .2 | .2 | Tr |
| Opaque minerals | | | | .2 | .3 | .2 | .3 | | | | Tr |
| Garnet | | | | | | | | P | | | Tr |
| **CIPW norms** | | | | | | | | | | | |
| Q | 20.55 | 25.84 | 25.65 | 28.90 | 30.73 | 33.64 | 30.00 | 34.70 | 39.00 | 45.96 | |
| C | | 1.78 | 1.56 | 1.52 | 1.76 | 2.86 | 1.46 | 1.94 | .63 | .40 | |
| Or | 16.78 | 23.04 | 9.45 | 12.41 | 11.82 | 13.59 | 19.50 | 28.36 | 5.91 | 4.61 | |
| Ab | 22.51 | 25.37 | 33.83 | 28.75 | 30.45 | 26.22 | 27.91 | 24.53 | 32.14 | 23.68 | |
| An | 21.64 | 11.36 | 19.46 | 15.06 | 13.80 | 11.33 | 12.11 | 5.17 | 14.29 | 19.25 | |
| Hl | .02 | | | | | | | | | | |
| Wo | .63 | | | | | | | | | | |
| En | 7.17 | 2.49 | 2.99 | 4.23 | 4.48 | 3.98 | 2.99 | 1.37 | 3.48 | 2.49 | |
| Fs | 5.12 | 4.41 | 3.71 | 4.99 | 3.64 | 4.53 | 3.46 | 1.82 | 1.83 | 1.70 | |
| Mt | 2.54 | .96 | 1.30 | 1.88 | 1.74 | 1.09 | 1.07 | .51 | 1.22 | .81 | |
| Il | 1.06 | 1.63 | .68 | .38 | .76 | 1.18 | .91 | .57 | .61 | .55 | |
| Ap | .28 | 1.26 | .50 | .47 | .21 | .57 | .28 | .28 | .21 | .21 | |
| Fr | .09 | | | | | | | | | | |
| Cc | .34 | 1.09 | | | | | | | | | |

[1] Average of 50 random grains counted in each of 24 thin sections of migmatitic biotite gneisses.

NOTE.—Minor-element analyses for sample 1 given in table 1.

1. Hornblende gneiss. Medium-grained well-foliated streaky dark-gray migmatitic biotite gneiss containing light-colored quartz-feldspar folia and veinlets. Rock consists of mosaic of irregular interlocking grains of quartz 0.25 to 1.5 mm in diameter, clear twinned plagioclase ($An_{40}$) 0.25 to 1 mm in diameter, and well-alined undeformed flakes of brown biotite as much as 1 mm long. Microcline forms irregular grains intergrown with quartz and plagioclase and commonly is fringed with myrmekite. Green hornblende occurs as scattered irregular grains as much as 1.5 mm in diameter. Colorless epidote forms irregular grains and aggregates as much as 1 mm long. Roadcut on east side of U.S. Highway 321, 0.4 mile south of bridge over Lower Creek (area J-7, pl. 1).
2. Strongly foliated nonlayered biotite gneiss containing scattered potassic feldspar augen. This rock is apparently transitional between biotite-amphibole gneiss (specimen 7, table 23) and biotite quartz monzonite (specimen 2, table 27). Rock consists of inequigranular mosaic of irregular grains of quartz, plagioclase ($An_{30-35}$), microcline, and very irregular flakes of synkinematic and postkinematic brown biotite. Most quartz and feldspar grains are 0.1 to 0.5 mm, but plagioclase and microcline also form porphyroblasts as much as 5 mm in diameter, and quartz occurs as 4-mm grains in segregation folia. Muscovite occurs as skeletal grains replacing feldspars, and epidote, as small irregular grains in biotite, many with cores of brown allanite. Near northwest corner of Causby quarry, 0.9 mile S. 10° W. of confluence of Catawba River and Johns River (area H-9, pl. 1).
3. Layered medium-grained migmatitic biotite gneiss containing folded layers of amphibolite and layers and streaks of biotite quartz diorite. Neither amphibolite nor quartz diorite are included in analyzed specimen. Rock consists of unsheared mosaic of granoblastic quartz and plagioclase ($An_{35}$) in grains 0.5 to 1 mm in diameter and irregular flakes of undeformed reddish-brown biotite as much as 1 mm long. Plagioclase has thin albite fringes and contains a few irregular blebs of potassic feldspar. Epidote forms irregular to subhedral grains, some with allanite cores 0.01 mm in diameter. Muscovite occurs in small flakes probably derived from alteration of feldspar. North wall of abandoned quarry on east side of Hunting Creek 0.9 mile S. 12° W. of confluence of Johns River and the Catawba River (area H-9, pl. 1).
4. Coarse-grained streaky migmatitic quartz diorite gneiss (cf. specimen 7) interleaved with folia of coarse-grained biotite schist, fine-grained biotite gneiss, and amphibole-biotite gneiss (specimen 8, table 23). Rock consists of inequigranular mosaic of irregular grains of quartz as much as 2 mm in diameter, plagioclase ($An_{35}$) as much as 8 mm in diameter, and randomly oriented flakes of dark-brown biotite 0.25 to 1.5 mm long. Plagioclase is twinned and faintly zoned, and some has fringes of albite. Sericite forms fringes on biotite and aggregates

TABLE 28.—*Chemical analyses, modes, and norms of megmatitic biotite gneisses*—Continued

in feldspar. Epidote occurs as irregular aggregates, chiefly in biotite. Roadcut on east side of North Carolina Highway 18 bypass west of Lenoir, 1.1 miles S. 68° W. of Caldwell County Courthouse (area J–7, pl. 1).

5. Coarse-grained, streaky well-foliated migmatitic quartz diorite gneiss. Rock consists of inequigranular mosaic of grains of quartz as much as 3 mm long, plagioclase ($An_{[illegible]}$) as much as 5 mm in diameter, and irregular alined flakes of olive-green biotite as much as 4 mm long. Plagioclase is twinned and faintly zoned, and some grains have fringes of albite. Some plagioclase twin lamellae and biotite flakes are warped. Potassic feldspar occurs in small irregular grains interstitial to quartz and plagioclase. Scattered poikilitic grains of blue-green hornblende are partly replaced by biotite. Epidote occurs in skeletal aggregates and scattered subhedral grains in biotite. Roadcut on east side of secondary road, 1.2 miles S. 74° E. of school at Gamewell (area I–8, pl. 1).
6. Coarse-grained, poorly foliated streaky migmatitic quartz diorite gneiss containing thin folia of biotite schist and boudinaged dikes of biotite pegmatite. Rock consists of an inequigranular mosaic of irregular grains of quartz 0.1 to 1.5 mm long, and of plagioclase ($An_{35}$) 0.5 to 2 mm long, and flakes of alined undeformed brown biotite 0.1 to 1.0 mm long. Plagioclase porphyroblasts as much as 5 mm in diameter are elongated parallel to foliation. Muscovite occurs in flakes syngenetic with biotite and as sericite from local alteration of plagioclase. Abandoned quarry on north side of Millers Creek, 1.55 miles S. 10° W. of Caldwell County Courthouse (area J–7, pl. 1).
7. Medium-grained, well-foliated streaky migmatitic granodiorite gneiss (cf. specimen 4) interleaved with folia of coarse-grained biotite schist, fine-grained biotite gneiss and amphibole-biotite gneiss (specimen 8, table 23). Rock consists of inequigranular mosaic of irregular grains of quartz as much as 3 mm in diameter, plagioclase ($An_{30\text{-}38}$) 1 to 4 mm in diameter, and stubby rudely alined grains of undeformed brown biotite as much as 3 mm long. Microcline forms irregular grains as much as 5 mm long, elongated parallel to foliation, partly jacketed, and deeply penetrated by myrmekite. It also occurs in small irregular grains interstitial to plagioclase. Muscovite is largely secondary from alteration of plagioclase. Epidote forms rudely prismatic skeletal grains containing brown low birefringent cores. Same locality as specimen 4.
8. Medium-grained, strongly foliated migmatitic quartz monzonite containing layers of fine-grained amphibole-biotite gneiss. Rock consists of inequigranular mosaic of irregular grains of quartz 0.1 to 1.0 mm in diameter, plagioclase ($An_{29}$) 0.25 to 1.0 mm in diameter, and alined stubby flakes and irregular grains of undeformed red-brown biotite 0.1 to 1 mm long. Plagioclase also occurs as porphyroblasts as much as 5 mm in diameter. Myrmekite occurs at contacts between plagioclase and microcline. Muscovite forms fringes on biotite and sericite clusters in feldspars and is apparently all secondary. Roadcut on west side of road north of Huffman Bridge, 0.4 mile N. 20° E. of south end of bridge (area I–9, pl. 1).
9. Medium-grained, layered, and well-foliated quartz diorite gneiss containing streaks of biotite schist and amphibole-biotite gneiss and pods and blocks of amphibolite. Rock consists of an inequigranular mosaic of irregular grains of quartz 0.1 to 1.0 mm in diameter and plagioclase ($An_{34}$) 0.5 to 4.0 mm in diameter, and rudely alined irregular flakes of brown biotite 0.1 to 1.5 mm long. Plagioclase is partly sericitized and contains blotches of albite. Bluish-green hornblende occurs in poikilitic grains as much as 1 mm long. Epidote forms irregular aggregates and small grains, many with brown weakly birefringent cores. East bank of the Catawba River 0.1 mile north of North Carolina Highway 181 (south edge of area G–9, pl. 1).
10. Coarse-grained, streaky migmatitic quartz diorite gneiss containing streaks and layers of amphibole quartz diorite. Rock is composed of inequigranular mosaic of irregular grains of quartz as much as 4 mm in diameter and plagioclase ($An_{37\text{-}45}$) 1 to 4 mm in diameter, and rudely alined irregular flakes of brown biotite as much as 1.5 mm long. Blue-green hornblende occurs in irregular poikilitic grains as much as 1 mm in diameter. Epidote forms irreguar aggregates and skeletal grains, some with light yellow weakly birefringent cores. South side of Warrior Fork, 0.4 mile N. 55° W. of confluence with the Catawba River (area G–9, pl. 1).

alpha age determinations on zircon and monazite that yielded dates of about 400 m.y. More recent isotopic uranium-lead determinations on zircon from the Toluca in its type locality give discordant ages ranging from 405 to 485 m.y. (Davis and others, 1962). Biotite from the same rock gave a rubidium-strontium age of 250 m.y. Davis, Tilton, and Wetherill (1962, p. 1993) interpreted these data as indicating a minimum age of 400 m.y. for the zircon, but pointed out that "it is not possible from this information to reach any conclusions about the time of intrusion of the rock itself." They believed that many of the zircon crystals were inherited from sedimentary rocks. They attributed the biotite age of 250 m.y. to the effect of a later metamorphism.

Even if the age of 400 m.y. were accepted as the true age of the Toluca Quartz Monzonite, recent revisions of the absolute geologic time scale (Holmes, 1959; Kulp, 1961) indicate that the geologic age would be latest Silurian or Early Devonian rather than Ordovician. In view of the uncertainties involved in the interpretation of the available isotopic age data and in the correlation between isotopic and geologic ages, we suggest that the Toluca Quartz Monzonite be assigned to the early or middle Paleozoic rather than specifically to the Ordovician. The correlative granitic rocks in the Inner Piedmont belt in the Grandfather Mountain area are also believed to be of early or middle Paleozoic age. A single unpublished potassium-argon age determination by J.L. Kulp (written commun., 1961) on muscovite from quartz monzonite from a roadcut 0.5 mile N. 5° E. of Willow Tree school (area F–9, pl. 1) is consistent with this age assignment and indicates that the muscovite has not been affected by the metamorphism of 250 m.y. ago that affected the biotite in the Toluca Quartz Monzonite at its type locality.

### QUARTZ MONZONITE GNEISS

On the southeast side of the Brevard fault zone in the southern Blowing Rock and northern Lenoir quadrangles (areas J–5 and I–6, pl. 1) is a long thin body of white fine- to medium-grained well-foliated and lineated quartz monzonite gneiss. Small lenses of quartz monzonite, most of which are too small to map, are strung out along the Brevard fault northeast and southwest of the main body, and some of the material mapped as blastomylonite to the southwest may have been derived from similar rock.

Except for one lens in the Lenoir quadrangle, the bodies of light-colored quartz monzonite gneiss have sharp and very sheared contacts on the northwest side, whereas on the southeast side they are interlayered with gneiss of the Inner Piedmont belt.

The rock ranges from blastomylonitic gneiss to blastomylonite and contains porphyroclasts of feldspar and mica. Alined mineral grains form the lineation. The main body of quartz monzonite gneiss includes only a few layers of other rock types, but to the southeast the white granitic gneiss forms layers in typical gneiss and schist of the Inner Piedmont. In one outcrop just west of Patterson (area I–6, pl. 1) the quartz monzonite gneiss contains septa of

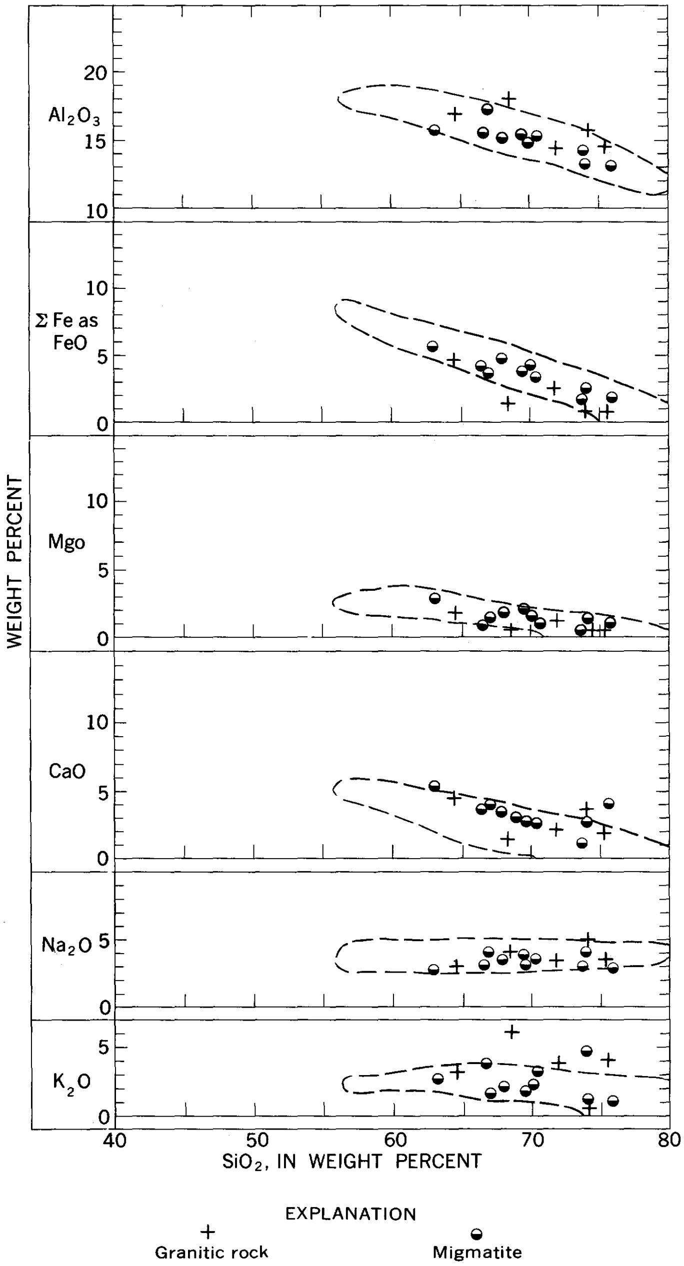



FIGURE 87.—$SiO_2$ variation diagrams comparing granitic rocks and migmatites of the Inner Piedmont belt with layered biotite gneisses. Dashed line encloses field of biotite gneisses from figure 76.

gneiss in which the layering is conformable with the cataclastic foliation of the granitic rock.

The rock consists of porphyroclasts of microcline, plagioclase, quartz, and muscovite about 2 mm in diameter set in a mosaic-textured groundmass of recrystallized quartz, plagioclase, microcline, sericite, biotite, epidote, and chlorite ranging from 0.01 to 0.2 mm in grain size (fig. 90). The plagioclase is sodic oligoclase and albite (about $An_{10}$). Some of the re-

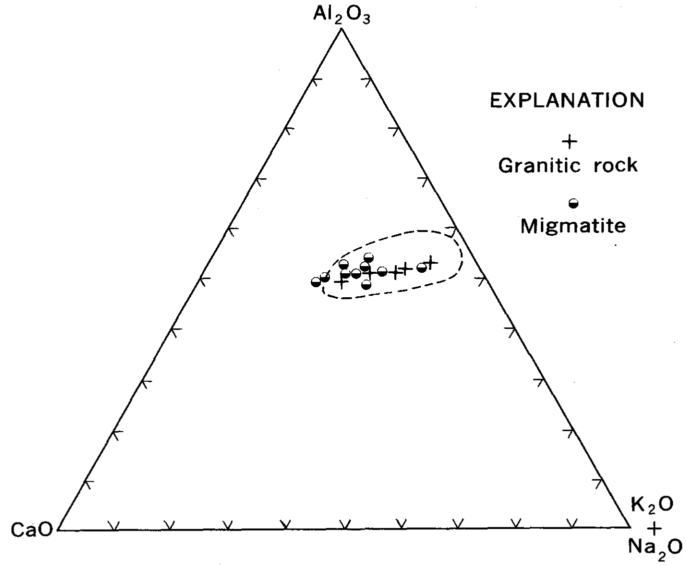



FIGURE 88.—Molecular proportions of $Al_2O_3$, CaO, and total alkalies in granitic rocks and migmatites. Dashed line encloses field of biotite gneisses from figure 78.

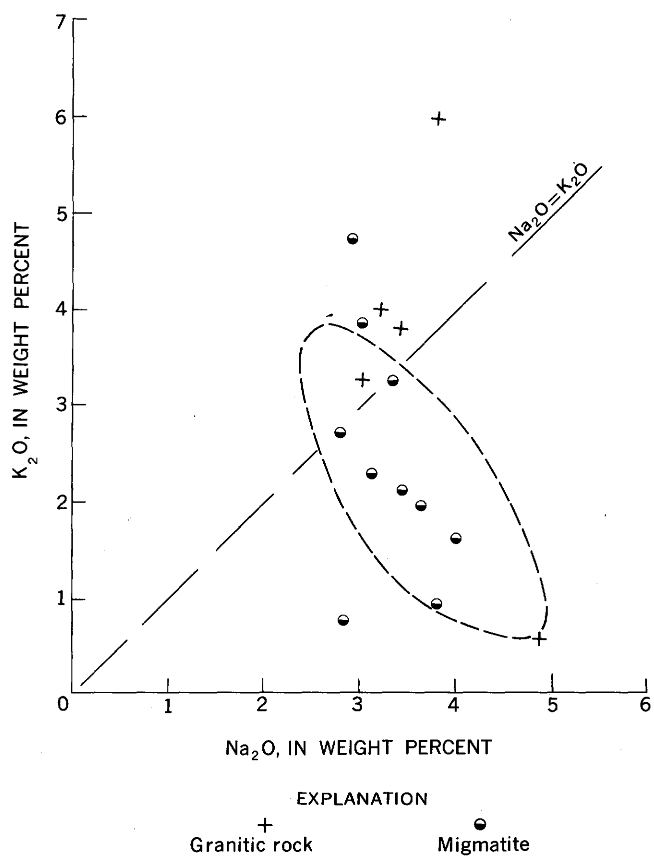



FIGURE 89.—$Na_2O$:$K_2O$ variation diagram comparing granitic rocks and migmatites of the Inner Piedmont belt with layered biotite gneisses. Dashed line encloses field of layered biotite gneisses from figure 77.

FIGURE 90.—Photomicrograph of quartz monzonite blastomylonite from east fork of Church Branch (area J–5, pl. 1). Small porphyroclasts of plagioclase (pseudomorphosed by albite) and microcline in a groundmass of recrystallized quartz, albite, microcline, muscovite, and subordinate biotite and epidote. Section cut parallel to mineral lineation.

crystallized biotite and epidote is as much as 1 mm long. Accessory minerals are garnet, zircon, apatite, and opaque minerals.

The age of this granitic rock has not been established but it is younger than the country rock and older than the faulting and retrogressive metamorphism. It, therefore, may be anywhere from late Precambrian to middle Paleozoic in age.

### ULTRAMAFIC ROCKS

Serpentine schist and altered peridotite and dunite are intercalated with layered rocks of the Inner Piedmont belt in a few places. The bodies of ultramafic rocks range from pods a few inches long to a lens 1,000 feet across and 4,000 feet long (area F–8, pl. 1). A series of disconnected ultramafic lenses as much as 200 feet thick extends along strike for nearly 2 miles between the Yadkin River and the head of Celia Creek (areas G–7 and H–7, pl. 1). The ultramafic bodies have been found only in the belt of polymetamorphic rocks southeast of the Brevard fault zone, and all occur along a well-defined trend parallel to the fault zone. Contacts of the ultramafic bodies are everywhere conformable with layering and foliation in the enclosing rocks, and most of the bodies are elongated parallel to the regional strike of the enclosing rocks. The ultramafic rocks commonly form resistant float and fresh natural outcrops, unlike most other rocks in the Piedmont, but contacts of most of the bodies are poorly exposed.

The ultramafic rocks are chiefly lustrous, fine- to medium-grained, dark-gray-green magnesian schists, commonly having 0.5 to 1 cm knots of light-green to gray amphibole on weathered surfaces. Schistosity is generally well developed, but in the centers of some of the larger bodies the rocks are nonfoliated and have textures resembling those of unaltered dunite. The rocks are principally composed of tabular or felted aggregates of prochlorite(?) interleaved with magnetite and small flakes and aggregates of talc. Colorless to light-green amphibole occurs in ragged prisms and in poikilitic grains and tiny needles interwoven with chlorite. Relict poikilitic grains of clinopyroxene (diopside?) occur in many specimens, and a few also contain relict grains of olivine partially replaced by antigorite(?). In the nonfoliated rocks having textures relict from dunite, the olivine grains are entirely pseudomorphosed by antigorite or prochlorite set in a felted matrix of prochlorite, antigorite, and magnetite. Some of these rocks are strongly magnetic in hand specimen. Veins of fibrous anthophyllite as much as 1 foot thick cut the unsheared ultramafic rocks in several of the larger bodies. One of these, 1.0 mile N. 86 W. of Conways Chapel (area G–7, pl. 1) has been mined for asbestos and is described by Conrad, Wilson, Allen, and Wright (1963).

There is no direct evidence on the age of the ultramafic rocks in the Inner Piedmont belt. Reed (1964b) reported the occurrence of what he believed to be a biotized and feldspathized ultramafic rock adjacent to Henderson Gneiss in an outcrop on the south shore of Lake James, 1.5 miles southeast of the mouth of the Linville River (area E–9, pl. 1), and suggested that the ultramafic body was older than the Henderson Gneiss. On the other hand, the fact that the ultramafic bodies are apparently confined to the belt of sheared polymetamorphic rocks along the

Brevard fault zone and are alined with the fault zone may indicate a genetic relation. If so, the ultramafic rocks are either exotic lenses or are younger than the granitic rocks of early or middle Paleozoic age cut by the fault zone. Until the age relations between the ultramafic rocks and the granitic rocks are clearly established, the age of the ultramafic rocks is entirely undetermined.

## STRUCTURE OF ROCKS OF THE INNER PIEDMONT NOT SHOWING METAMORPHIC EFFECTS RELATED TO THE BREVARD FAULT ZONE

Description and interpretation of the complex structure of the rocks of the Inner Piedmont belt is impeded by the lack of recognizable stratigraphic units, obliteration of original textures and structures, and the difficulty of detailed geologic mapping. Deep weathering and extensive soil cover make it impossible to recognize structures intermediate in scale between those observed in single small outcrops and those inferred from the necessarily crude geologic maps. Therefore, statistical orientation diagrams are necessary to describe the structural geometry and decipher the history of these rocks. Although it seriously handicaps geologic mapping, the deep weathering is not entirely disadvantageous, for in undisturbed saprolite, bedrock structures are almost perfectly preserved and may be easily dissected with a pocketknife or entrenching tool for measurement and three-dimensional study.

The general trends of structures in the Inner Piedmont belt are shown on the foliation map (fig. 32), the lineation map (fig. 33), and on the geologic map (pl. 1). These maps show an obvious contrast in map pattern and structural fabric between the high-grade metamorphic rocks in the southeastern part of the area and the polymetamorphic rocks adjacent to the Brevard fault zone. Structures of the polymetamorphic rocks are related to the Brevard zone and are discussed below.

### LAYERING AND FOLIATION

Although no undisputed sedimentary features are preserved in the Inner Piedmont rocks, the strongly layered character of many of the gneisses and schists indicates that many of them were originally stratified rocks. The continuity of individual layers and groups of layers, the compositional contrasts between adjacent layers, and the variety of lithologies represented strongly suggest that the layering is inherited from bedding. Contacts between layers are largely tectonic, however, and flowage parallel to layering is indicated by attenuation of ductile mica-rich layers and boudinage of brittle amphibolite and quartzite layers to form detached pods and blocks, some of which are rotated with respect to the more ductile enclosing rocks. The arrangement of the various individual lithologies in the layered rocks probably has little or no relation to their original stratigraphic sequence.

Foliation due to alinement of micaceous and prismatic minerals and quartzo-feldspathic folia parallel to layering is conspicuous in most rocks. It is best developed in schists and the more micaceous gneisses and is least apparent in quartzo-feldspathic gneiss.

Foliation due to parallel arrangement of biotite flakes and streaks and quartzo-feldspathic folia is well developed in migmatite and in many of the granitic rocks. Foliation in most of the migmatites and granitic rocks is parallel to the contacts of the rock bodies and to layering and foliation in the enclosing rocks. Some discordant dikes and a few larger granitic bodies have streaky planar flow structure parallel to their walls and at high angles to foliation and layering in the wallrocks (fig. 83*D*).

### FOLDS AND LINEATION

Many outcrops of sillimanite schist have small crenulations in foliation which range in amplitude from fractions of an inch to several inches. Faint slip cleavage is developed parallel to the axial planes in many places (fig. 91*A* and *B*). Folds in competent layers of quartzite and gneiss interlayered with sillimanite schist have amplitudes of a few inches to several feet. They are sharp but generally not isoclinal and have only minor thickening in the noses. Their axes and axial planes are parallel to those of crenulations in the enclosing sillimanite schist. Folds are rarely observed in the rocks between the sillimanite schist and the main belt of granitic rocks, but where they are present, they are generally more tightly appressed and have more pronounced thickening of ductile layers in their noses than do folds in the sillimanite schist. They closely resemble flexural flow folds described by Donath and Parker (1964).

### STRUCTURAL GEOMETRY

Figure 4 on plate 5 shows the statistical distribution of measurements of layering, foliation, lineation, fold axes, and axial planes in rocks of the Inner Piedmont belt in the area where no metamorphic effects related to the Brevard fault zone were recognized (pl. 6*D*). This area is subdivided into three zones: the main belt of granitic rocks and migmatite

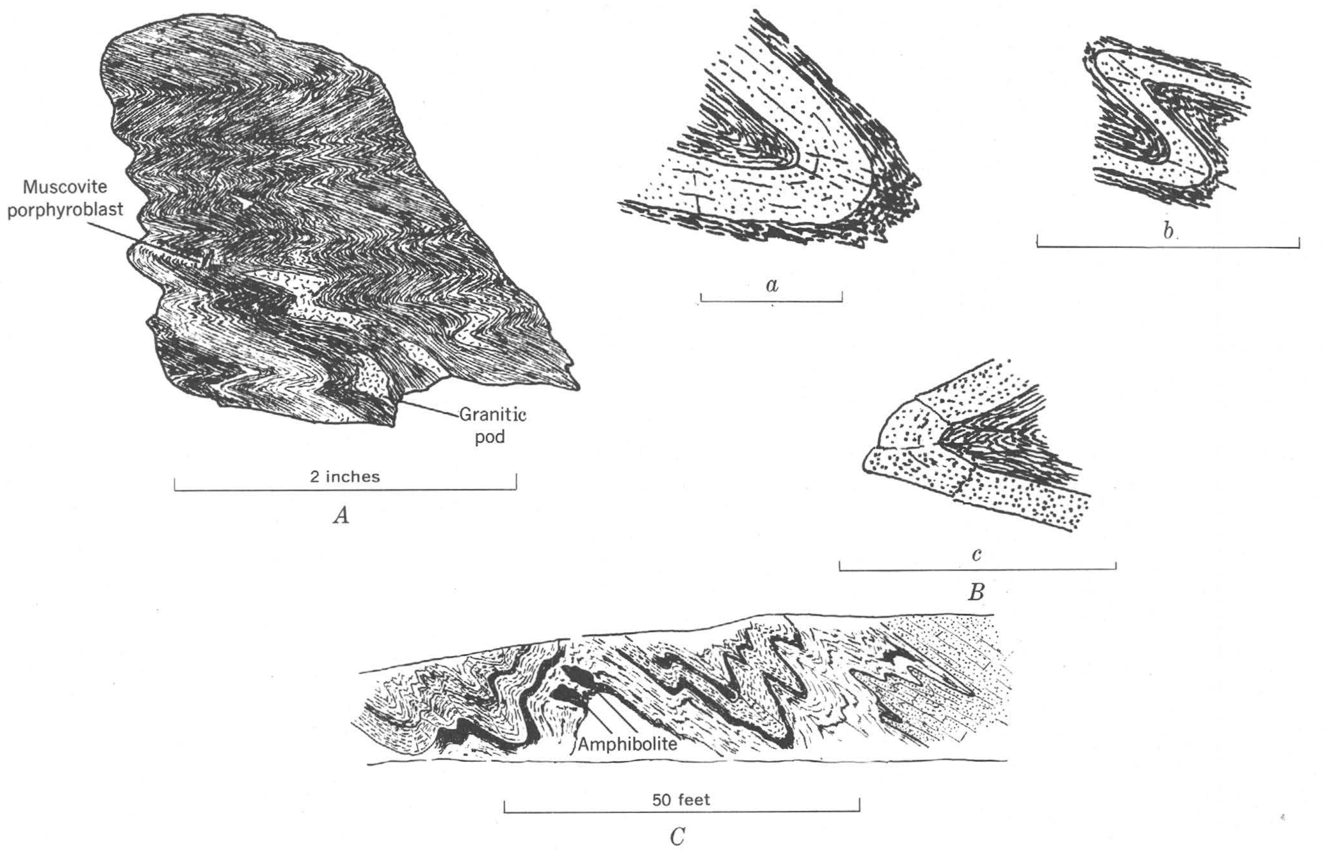



FIGURE 91.—Folds in Inner Piedmont rocks. *A*, Crenulations in sillimanite schist. View on AC plane with north to the left, Small granitic pods are conformable with folded foliation. Muscovite porphyroblasts are undeformed and cut across fold noses. Drawn from a hand specimen. South side of Rhodhiss Lake, 1.6 miles S. 82° W. of south end of Castle Bridge (area J–9, pl. 1). *B*, Folds in quartzite layers in sillimanite schist. Scale below each fold represents 1 foot. All folds viewed on AC plane with north to the left. *a*. Roadcut through Cajah Mountain in area J–8, 1.3 miles N. 6° E. of school at Baton (area J–9, pl. 1). *b*. Same roadcut about 60 feet south of 1. *c*. South short of Rhodhiss Lake, 1.3 miles S. 80° E. of south end of Castle Bridge (area J–9, pl. 1). *C*, Folds in interlayered biotite gneiss, mica schist, and quartz schist (stippled at right side of drawing). Note isolated rotated blocks of amphibolite. Direction of view is northeast, approximately parallel to fold axes. Roadcut in saprolite on northeast side of North Carolina Highway 126, 0.6 mile S. 34° E. of Grandview Church (area F–9, pl. 1).

which extends continuously northeastward from just west of Morganton through Lenior (pl. 5, fig 4, col. *A*); the area to the southeast which is mapped as predominantly layered gneiss, but which also contains isolated boidoes of mica schist, migmatite, and granitic rocks (col. *B*); and the sillimanite schist and interlayered rocks in the southeastern corner of the Lenoir quadrangle (col. *C*). The structures in these belts are discussed in order of decreasing distance from the Brevard zone.

Separate plots show that, within any given area, layering and foliation in the layered rocks and foliation (other than obvious flow structures) in nonlayered rocks have orientation patterns that are statistically indistinguishable, and these structures are therefore plotted together in the upper row of diagrams.

Layering and foliation in both groups of rocks southeast of the main belt of granitic rocks strike almost north-south and dip gently or moderately east. Map patterns, particularly that of the northeastern contact of the sillimanite schist, suggest major open folds whose axes plunge gently east. Poles to layering and foliation cluster around single broad maxima (diagrams *B*–1 and *C*–1, fig. 4, pl. 5), with only the vaguest suggestion of girdles around east-plunging axes. Mineral lineation, marked by alinement of sillimanite needles and fibers in the sillimanite schist and by hornblende needles and long dimensions of mica flakes and clusters and quartz-feldspar aggregates in other rocks, plunges eastward. Lineation projections (diagrams *B*–2 and *C*–2, fig. 4, pl. 5) are arrayed in partial girldes along the trace of the foliation and layering maxima and culminate in point maxima down the dip of the foliation and layering.

Crenulations and minor folds observed in individual outcrops have axes plunging gently east, parallel to the mineral lineation; axial planes of most of the minor folds dip gently or moderately north or northeast, generally at angles slightly steeper than the foliation and layering (diagrams *B*–3 and *C*–3, fig. 4, pl. 5).

Although metamorphic effects related to the Brevard fault zone are not recognized in the main belt of granitic rocks, the structural geometry of the rocks in this belt (col. *A*, fig. 4, pl. 5) is more like that of the polymetamorphic rocks to the northwest than that of the rocks to the southeast. Foliation and layering strike northeast and dip consistently southeast; poles cluster symmetrically around a well-defined maximum (diagram *A*–1, fig. 4, pl. 5). Mineral lineation plunges northeast and forms a single high maximum in the projection (diagram *A*–2, fig, 4, pl. 5). Mineral lineation is absent in most of the granitic rocks and is very rare in the migmatites. No major folds are apparent in the map pattern, but minor folds in layered gneiss and migmatite resemble those in the rocks to the southeast. Their axes plunge northeast, approximately parallel to the mineral lineation, and their axial planes dip southeast, parallel to layering and foliation (diagram *A*–3, fig. 4, pl. 5). Two maxima appear in the fold axis orientations, one plunging very gently N. 45° E. and the second plunging more steeply northeast and lying on a partial small-circle girdle that extends to near the position of the fold-axis maxima in the rocks to the southeast (diagrams *B*–3 and *C*–3, fig. 4, pl. 5). This suggests that the second group of folds may have been rotated into their present orientation from an original orientation parallel to folds in the rocks to the southeast.

### AGE RELATIONS BETWEEN FOLDING, FORMATION OF MINERAL LINEATION, AND EMPLACEMENT OF GRANITIC ROCKS

In the sillimanite schist, felted aggregates of randomly arrayed sillimanite fibers are bent around noses of crenulations and minor folds, whereas larger sillimanite needles are arranged parallel to fold and crenulation axes and form the mineral lineation. Apparently, sillimanite began crystallizing prior to folding and continued to grow parallel to fold axes during folding. Folding in the sillimanite schist is therefore inferred to have occurred during high-grade regional metamorphism, somewhat later than the beginning of crystallization of sillimanite. In other rocks, the age relations between high-grade regional metamorphism and folding are not as clear. However, the similarity in structural geometry between the sillimanite schist and the rocks immediately to the northwest and the parallelism between fold axes and mineral lineation formed by alinement of hornblende prisms, mica flakes, and quartz-feldspar aggregates in layered gneiss, mica schist, and amphibolite suggest that folding was contemporaneous with high-grade regional metamorphism and that the lineation is due to preferential growth of minerals parallel to fold axes.

Most mappable granitic bodies are concordant or semiconcordant in gross aspect; foliation in the granitic rocks and flanking migmatite generally is parallel to layering and foliation in the enclosing rocks. Minor folds commonly involve foliation in migmatite, and some dikes and pods of granitic rocks and pegmatite are apparently folded along with their wallrocks. In many places, however, dikes of granitic rocks are sharply discordant with folded layering and foliation in the enclosing rocks (fig. 92*A*) and contain rotated inclusions, some of which have complex folds that must have formed before emplacement of the enclosing granitic rock (figs. 92*B*, 93).

These relations indicate that most of the granitic rocks were emplaced during folding and high-grade regional metamorphism. Some granitic material, however, remained mobile long enough to be intruded as discordant dikes and local crosscutting bodies after the peak of regional metamorphism and formation of the minor folds.

## BREVARD FAULT ZONE

In the Grandfather Mountain area, the Brevard fault zone comprises a number of subparallel anastomosing faults marked by belts of mylonite, blastomylonite, and phyllonite intercalated with tectonic lenses and slices of phyllonitic schist and gneiss, some of which are apparently derived from adjacent rocks and some of which are probably exotic. On the geologic map (pl. 1), rocks of the Brevard zone are subdivided into blastomylonite and related siliceous mica-poor rocks and mica-rich phyllonitic schist and gneiss.

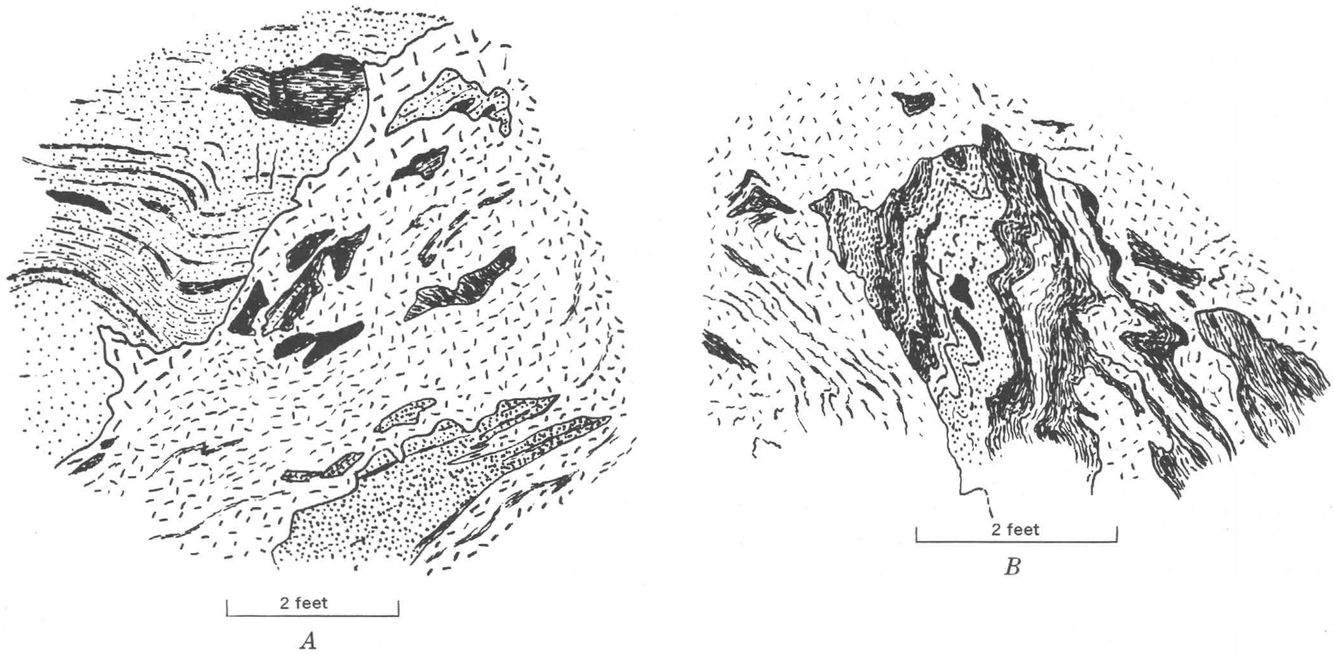



FIGURE 92.—Field relations of granitic rocks. *A*, Dike of leucocratic granitic rock and pegmatite cutting layered biotite gneiss, biotite-hornblende gneiss, and fine-grained biotite schist and containing rotated and partly digested inclusions of the wallrocks. Biotite pegmatite forms a poorly defined zone along the contact. Drawn from a photograph. Roadcut in saprolite along farm road, 2.2 miles S. 77° E. of school at Gamewell (area I–8, pl. 1). *B*, Inclusion of complexity folded interlayered biotite and biotite-hornblende gneiss and amphibolite in leucocratic granitic rock and migmatite. Drawn from a photograph. Roadcut in saprolite in area J–6, 300 feet south of Zacks Fork Creek, 3.25 miles N. 36° E. of courthouse in Lenoir (area J–7, pl. 1).

FIGURE 93.—Fold in layered biotite and hornblende-biotite gneiss cut by dike of light-colored granitic rock containing blocky inclusions of layered gneiss. Diagonal dark streaks are iron-stained zones along joints. Saprolite exposure in borrow pit in area H–8 on west side of road, 1.7 miles N. 16° E. of school at Chesterfield (area H–9, pl. 1). Direction of view is east, approximately parallel with fold axis.

Rocks in the Blue Ridge thrust sheet and Inner Piedmont belt adjacent to the Brevard fault zone have structures and textures indicating that they have been pervasively sheared and retrogressively metamorphosed during movement along faults in the Brevard zone. Unlike most of the rocks in the Brevard zone itself, however, these polymetamorphic rocks can be identified with appropriate map units in the flanking tectonic blocks. The polymetamorphic rocks are therefore mapped and described with the rocks of the Blue Ridge thrust sheet, Inner Piedmont belt, and Grandfather Mountain window, but their structure is discussed in conjunction with the structure of the rocks in the Brevard fault zone. The slice of layered gneiss in the Brevard zone in the Linville Falls quadrangle (areas E–8 and F–8, pl. 1) is so like polymetamorphic layered gneiss adjacent to the zone in the Inner Piedmont that it is not described separately, although it is designated by a different pattern and symbol (gn) on plate 1.

## ROCK UNITS

### BLASTOMYLONITE

Blastomylonite interleaved with mylonite, phyllonite, and phyllonitic schist and gneiss forms a narrow

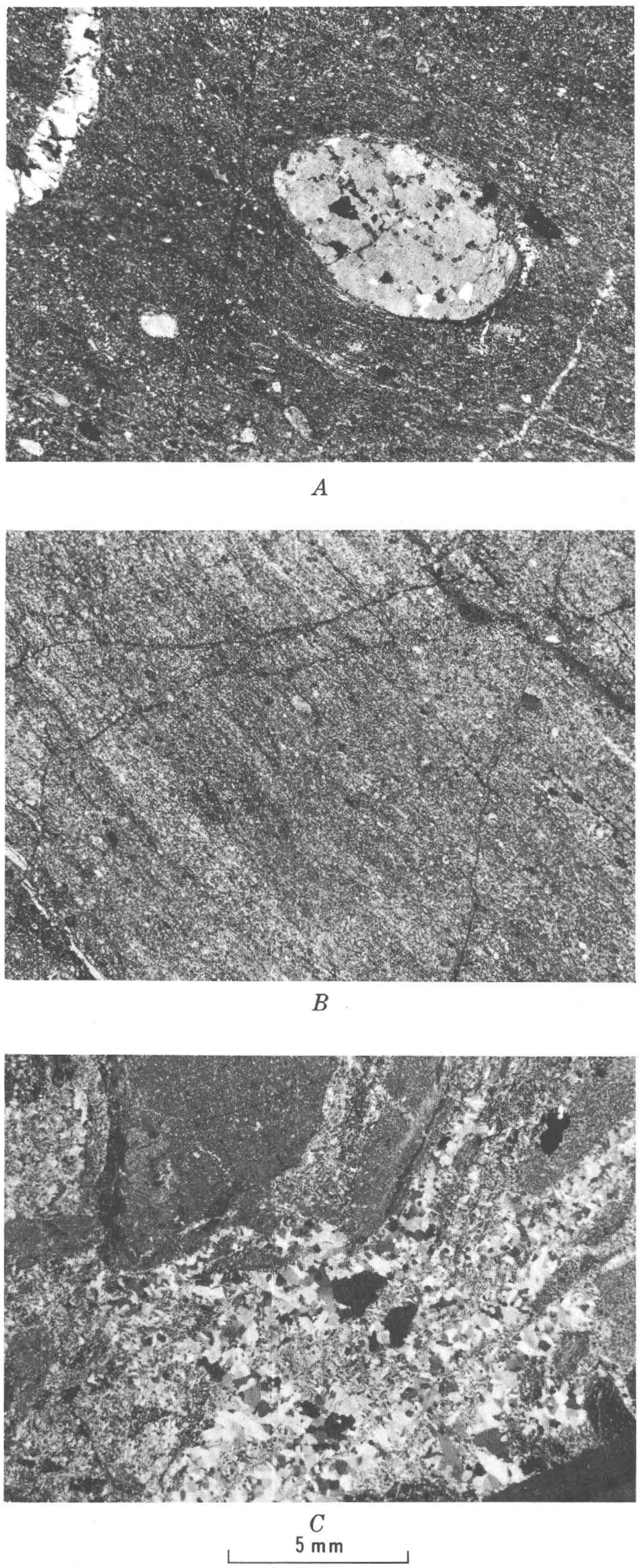



FIGURE 94.—Photomicrographs of blastomylonite. *A*, Porphyroclastic blastomylonite probably derived from Henderson Gneiss, outcrop along Roses Creek about 1 mile southeast of the Linville Falls fault (area F–8, pl. 1). A porphyroclast of perthitic microcline in a matrix of recrystallized biotite, epidote, sericite, quartz, and feldspar. Late quartz veinlet. Section perpendicular to mineral lineation. *B*, Blastomylonite from same outcrop as *A*. Epidote, sericite, microcline, albite, and quartz. *C*, Mylonite breccia from outcrop on point on north shore of Lake James, south of mouth of Dales Creek (area C–9, pl. 1). Fragments of mylonite in a matrix of recrystallized introduced(?) quartz.

but continuous belt in the Brevard fault zone in the Linville Falls quadrangle and southwestward beyond the southwestern end of the Grandfather Mountain window. Blastomylonite also occurs in small lenses elsewhere along the Brevard fault zone and as thin layers and zones parallel to foliation in schist and gneiss on both sides, but these bodies are too small to distinguish on the geologic map (pl. 1).

The blastomylonite is a very fine grained or aphanitic flinty rock, generally gray, greenish gray, buff, or pink, and commonly it has a dull waxy luster. Some types superficially resemble quartzite or altered volcanic rock. Porphyroclasts of microcline, patch perthite, and plagioclase 0.5 to 1 cm in diameter are common and, in some places, are larger (fig. 94*A*). The rock is composed of fine-grained recrystallized quartz, albite, and untwinned potassic feldspar and scattered flakes and wavy folia of sericite, biotite, and chlorite (fig. 94*B*). Foliation is defined by lentils of coarser and finer grained mosaic and quartz segregation laminae, as well as by micaceous folia. Porphyroclasts of muscovite are common. Some of the rock is a microbreccia in which angular fragments of microcyrstalline material containing wavy bands of pseudotachylyte are set in a recrystallized matrix. The microcrystalline fragments and the coarser granoblastic mosaic (fig. 94*C*) are cut by anastomosing quartz-feldspar mortar and microbreccia. Blastomylonite derived from mafic rocks contains abundant chlorite and epidote, and some of it has small porphyroclasts of hornblende and sphene. Locally, the blastomylonite is silicified and consists almost entirely of secondary quartz. The blastomylonite weathers to a structureless buff or cream-colorder clayey saprolite, locally containing irregular zones of chippy-weathering silicified blastomylonite.

Foliation is commonly well developed in the blastomylonite and is generally parallel to foliation in the encosing rocks and to foliation in interleaved phyllonitic rocks. Contacts between blastomylonite and adjacent rocks are gradational through distances ranging from a few inches to hundreds of feet.

### PHYLLONITIC SCHIST AND GNEISS

Most of the Brevard fault zone in the eastern half of the Grandfather Mountain area consists of a narrow belt of phyllonitic schist interlayered with phyllonitic and blastomylonitic gneiss and phyllonite. The belt extends southwestward to Lake James and northeastward out of the map area.

The phyllonitic schist is a lustrous fine-grained gray, gray-green, or blue-green chlorite-sericite schist containing conspicuous bent porphyroclasts of muscovite as much as 2 cm in diameter, partly replaced by sericite, and commonly containing porphyroclasts of garnet 0.5 to 2 cm in diameter, partially or entirely altered to chlorite. The muscovite porphyroclasts and the sericite aggregates replacing them give much of the schist a curly, scaly, lensy, or wavy appearance; in soils derived from weathering of the schist they form chips resembling buttons or large fish scales. With decrease in the size and abundance of muscovite porphyroclasts, the phyllonitic schist passes into lustrous gray or greenish-gray chlorite-sericite phyllonite.

The phyllonitic schist commonly contains layers of light- to medium-gray phyllonitic or blastomylonitic gneiss; locally, a few layers of amphibolite as much as 20 feet thick are interlayered with the schist and gneiss. In a few places, the gneiss layers are clearly beds, and locally they have graded bedding. Small stringers and segregation lenses of finely granular quartz are abundant in the schist and gneiss; the texture suggests that the quartz has been sheared and recrystallized.

The phyllonitic schist resembles some of the polymetamorphic schists in both the Blue Ridge thrust sheet and the Inner Piedmont belt, but it is richer in sericite, chlorite, plagioclase, and garnet. Pegmatite has not been found in phyllonitic schist or interlayered rocks in the Brevard zone, although small pegmatite bodies are ubiquitous in the schist and gneiss in adjacent parts of the Inner Piedmont and the Blue Ridge thrust sheet.

The phyllonitic schist differs from the interlayered gneiss chiefly in the proportion of micaceous minerals. In both rocks, recrystallized quartz and scattered grains of plagioclase and epidote form a mosaic of grains 0.01 to 0.2 mm in diameter. Plagioclase is generally albite, but in a few rocks, the plagioclase has not been completely decalcified. Synkinematic and postkinematic flakes of sericite and chlorite 0.05 to 0.2 mm long are scattered through the quartz mosaic and form wisps and laminae which define the foliation. A few rocks contain synkinematic brown biotite, but in most, biotite is entirely replaced by chlorite.

Muscovite porphyroclasts are very abundant in the phyllonitic schist, and a few occur in the interlayered gneiss. In cross section they have the outline of flattened parallelograms with the long axes 0.5 to 1 cm long and short axes 1 to 2 mm long. The long axes are alined with the foliation. Cleavage in some of the porphyroclasts is undeformed, but in many it is gently warped, and in a few it is sharply bent (fig. 95). Muscovite porphyroclasts are partly sericitized in many rocks, and in some, they are entirely replaced by sericite aggregates, which, with more shearing, are distributed so that no trace of the large muscovite remains.

Garnet crystals range from skeletal to euhedral, and many contain rotated helicitic trains of magnetite and quartz inclusions. Muscovite porphyroclasts are bent around garnets, and in some rocks the garnets seem to be rounded. Most of the garnets are at least partly replaced by chlorite, and in some rocks they are entirely pseudomorphosed or remain only as scattered relicts in chlorite aggregates. In many rocks it is not clear whether the garnets are porphyroblasts or porphyroclasts, but locally they are clearly porphyroclastic, and their alteration to chlorite suggests that they are all porphyroclastic, at least in relation to the latest recrystallization.

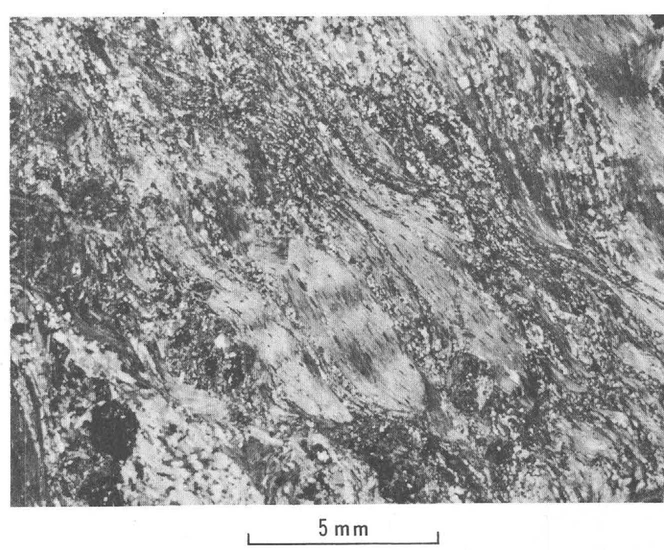



FIGURE 95.—Photomicrograph of typical phyllonitic mica schist from roadcut on U.S. Highway 321 at south edge of Blowing Rock quadrangle (south edge of area J–5, pl. 1). Bent porphyroclasts of white mica and partly chloritized garnet in a matrix of quartz and white mica.

Magnetite is the principal opaque mineral and occurs as scattered small grains and aggregates which constitute several percent of some rocks. Small synkinematic and postkinematic needles of tourmaline are widespread; the tourmaline is black in hand specimen and green or dark blue green in thin section. Other typical accessory minerals are zircon, apatite, and sphene. Locally, the schist contains scattered relicts of staurolite 0.2 to 0.5 mm in diameter embedded in sericite aggregates that were probably derived from staurolite porphyroclasts.

A chemical analysis of a typical specimen of phyllonitic schist (table 29) shows $Na_2O$, $K_2O$, and $Al_2O_3$ contents compatible with the amounts of chlorite and micas in the mode, assuming that the chlorite is aluminous and that some of the white mica is paragonite. X-ray study by A. J. Gude, 3d, of two other similar specimens of phyllonitic schist indicates that no albite is present and that about half of the white mica identified as sericite is paragonite. No other rocks in the Grandfather Mountain area are known to contain significant amounts of paragonite.

TABLE 29.—*Chemical analysis, mode, and norm of typical phyllonitic mica schist*

[Analysis determined by rapid methods by Paul Elmore, Samuel Botts, Gillison Chloe, Lowell Artis, and Hezekiah Smith, U.S. Geol. Survey, 1962. Mode, by count of 600 points; P, present but not intersected in counting. Major oxides and CIPW norms given in weight percent; modes, in volume percent]

| Field No. | 25-237 | Field No. | 25-237 |
|---|---|---|---|
| Laboratory No. | 160198 | Laboratory No. | 160198 |
| **Major oxides** | | | |
| $SiO_2$ | 53.7 | $K_2O$ | 2.4 |
| $Al_2O_3$ | 23.0 | $H_2O+$ | 3.3 |
| $Fe_2O_3$ | 5.8 | $H_2O-$ | .09 |
| FeO | 5.0 | $TiO_2$ | 1.3 |
| MgO | 2.4 | $P_2O_5$ | .34 |
| CaO | .71 | MnO | .25 |
| $Na_2O$ | 1.9 | $CO_2$ | <.05 |
| | | Total | 100 |
| **Mode** | | | |
| Muscovite and paragonite | 50 | Epidote | .5 |
| Quartz | 24 | Staurolite | .2 |
| Chlorite | 18 | Limonite | .3 |
| Opaque mineral | 5 | Tourmaline | .2 |
| Garnet | 1.5 | Sphene | .3 |
| | | Apatite | P |
| **CIPW norm** | | | |
| Q | 28.10 | En | 5.98 |
| C | 16.80 | Fs | 2.71 |
| Or | 14.18 | Mt | 8.41 |
| Ab | 16.07 | Il | 2.47 |
| An | 1.30 | Ap | .80 |

NOTE.—Description of analyzed specimen follows:

Lustrous green chlorite schist with silvery muscovite aggregates forming knots as much as 1 inch in diameter. Porphyroclasts of white mica as much as 1 cm in diameter are bent, folded, and partly granulated. Laminae formed by fine-grained white mica alternating with mosaic-textured quartz (grain size averaging 0.1 mm but reaching as much as 0.5 mm). MgFe chlorite is derived from alteration of garnet and possibly of biotite. Subhedral to euhedral garnet porphyroclasts 0.5 to 1 mm in diameter partly altered to chlorite; some have rotated helicitic structure. Anhedral to euhedral magnetite 0.1 to 0.7 mm in diameter and finer grained opaque mineral included in mica. Staurolite in porphyroclasts as much as 0.3 mm long. From southwest side of Coffey Creek 0.7 mile northwest of North Carolina Highway 90 (area H-6, pl. 1).

The absence of pegmatite in the phyllonitic schist and gneiss in the Brevard fault zone and the mineralogical differences between these rocks and polymetamorphic schists and gneisses in adjacent parts of the Blue Ridge thrust sheet and Inner Piedmont belt suggest that the schist and gneiss in the fault zone form an exotic tectonic slice derived from outside the Grandfather Mountain area. In texture and mineralogy the rocks resemble parts of the Candler Formation in the James River synclinorium in Virginia (Redden, 1963), but further detailed mapping along the Brevard fault zone northeast and southwest of the Grandfather Mountain area will be necessary to establish the source of the slice and the age and correlation of the rocks in it.

## STRUCTURE OF ROCKS OF THE BREVARD FAULT ZONE AND FLANKING POLYMETAMORPHIC ROCKS

The geologic contrasts between the Blue Ridge belt northwest of the Brevard fault zone and the Inner Piedmont belt southeast of the zone are obvious on the generalized geologic map (pl. 1) and on the structure maps (figs. 32 and 33). The abrupt lithologic change between rocks of the Blue Ridge belt and rocks of the Inner Piedmont belt is marked by the phyllonitic and mylonitic rocks of the Brevard. The changes in structural geometry are less abrupt and take place in a broader zone, mostly in the flanking rocks having metamorphic effects related to the Brevard fault zone.

The orientation of lineation (fig. 33) in the Blue Ridge thrust sheet and Grandfather Mountain window swings clockwise and becomes nearly horizontal and trends northeast in the fault zone and in the polymetamorphic rocks of the Inner Piedmont belt to the southeast. Southeast of the belt of polymetamorphic rocks, the lineation changes abruptly to the east-west trend characteristic of the high-grade metamorphic rocks of the Inner Piedmont.

Statistical diagrams (fig. 5, pl. 5) summarize the structural geometry of rocks of the Brevard fault zone and of the adjacent rocks affected by shearing and retrogressive metamorphism related to it.

Autochthonous basement rocks in the Grandfather Mountain window all have a uniform structural pattern. Cataclastic foliation formed during low-grade regional metamorphism strikes northeast and generally dips 40° to 60° SE. (fig. 32). Cataclastic lineation plunges southeast, nearly down the dip of the foliation planes (fig. 33). Diagrams *A*–1 and *A*–2 (fig. 5, pl. 5) show the orientation of foliation and lineation in the autochthonous basement rocks in an arbitrary belt 2 miles wide along the southeast edge of the window. Poles to foliation cluster around a single high maximum corresponding to a stike of N. 55° E. and dip of 45° SE. Projections of lineations form a point maximum plunging 45° S. 25° E. This pattern is consistent with the structural pattern in the autochthonous basement rocks elsewhere in the Grandfather Mountain window and is unrelated to the Brevard zone.

Most of the basement rocks are nonlayered and consequently no folds can be seen, but the layered phase of the Wilson Creek Gneiss near the Linville Falls fault in the eastern part of the window (pl. 1) has tight and isoclinal folds with axial planes parallel to cataclastic foliation. Axes of most of these folds plunge southeast, parallel to cataclastic mineral lineation (diagram *A*–3, fig. 5, pl. 5), but some axes plunge gently south or southwest, forming a partial girdle in the orientation diagram.

Structures in the Tablerock thrust sheet and in the higher thrust slices in the southwestern part of the Grandfather Mountain window are more closely related to those of the overriding Blue Ridge thrust sheet than to those in the autochthonous rocks in the window. The geometry of structures in the Tablerock thrust sheet and higher thrust sheets in the window in an arbitrary belt 2 miles along the southeastern edge of the window is summarized in the diagrams in column *B* of figure 5, pl. 5. Bedding is nearly parallel to foliation in rocks of the Chilhowee Group, and bedding and foliation poles are therefore both plotted in diagram *B*–1, figure 5, plate 5. They are scattered around a single point maximum, indicating a more easterly strike and gentler dip than foliation in the autochthonous basement rocks along the southeastern edge of the window. Mineral lineation generally plunges gently to moderately south and forms a partial girdle in the diagram (*B*–2, fig. 5, pl. 5), in a manner that corresponds to the conspicuous swing from southeast plunges parallel to the regional cataclastic lineation in the Grandfather Mountain window to south and southwest plunges near the Brevard zone (fig. 33). The change in direction of mineral lineation in the Tablerock thrust sheet and higher thrust sheets occurs within 2 miles of the southeastern boundary of the window, but it is not apparent in autochthonous basement rocks in the window at corresponding distances from the window boundary and the Brevard fault zone.

Axes of crenulations and isoclinal and subisoclinal folds in rocks of the Tablerock thrust sheet are parallel or subparallel to mineral lineation and have a similar distribution. Axial planes of the folds are nearly parallel to foliation (diagram *B*–3, fig. 5, pl. 5). The orientation pattern of crenulation axes and of fold axes and axial planes closely resemble that in the layered Wilson Creek Gneiss in the northeastern part of the window (diagram *A*–3, fig. 5, pl. 5).

Diagrams in column *C* of figure 5, plate 5 show orientation of structures in rocks of the Blue Ridge thrust sheet in the narrow belt between the southeastern side of the Grandfather Mountain window and the Brevard fault zone. Most of these rocks were originally of medium or high metamorphic grade, but most of them have been sheared and retrogressively metamorphosed under low-grade conditions during movement along the Linville Falls fault and the Brevard fault zone. Some of the medium-grade rocks in the northeastern part of the belt (pl. 6) have been remetamorphosed at medium grade, but some are apparently not polymetamorphic. Layering and foliation in the rocks of this belt are nearly everywhere parallel; the poles are distributed in a single maximum reflecting strikes parallel to the Brevard zone and dips averaging about 50° SE. Cataclastic mineral lineations formed during shearing and retrogressive metamorphism are arrayed in a partial great-circle girdle (diagram *C*–2, fig. 5, pl. 5) corresponding to the clockwise swing from moderate southward plunges near the Linville Falls fault to subhorizontal northeast trends near the Brevard fault zone (fig. 33). Axes of small isoclinal and subisoclinal folds (diagram *C*–3, fig. 5, pl. 5) plunge gently northeast or southwest; axial planes are subparallel to layering and foliation, but many are more steeply dipping. Medium-grade rocks in the northeastern part of the belt have many sharp crinkles with subhorizontal axes (figs. 6*A*, 96) trending northeast parallel to minor folds elsewhere in the belt. Axial planes of the crinkles are marked by a steeply dipping slip cleavage in the nonpolymetamorphic rocks which is parallel to cataclastic foliation in the polymetamorphic medium-grade rocks to the southeast.

Foliation and layering of phyllonitic and blastomylonitic rocks in the Brevard fault zone strike northeastward, parallel to the trend of the zone, and generally dip steeply southeastward, but vertical or even northwestward dips are found locally. Poles of

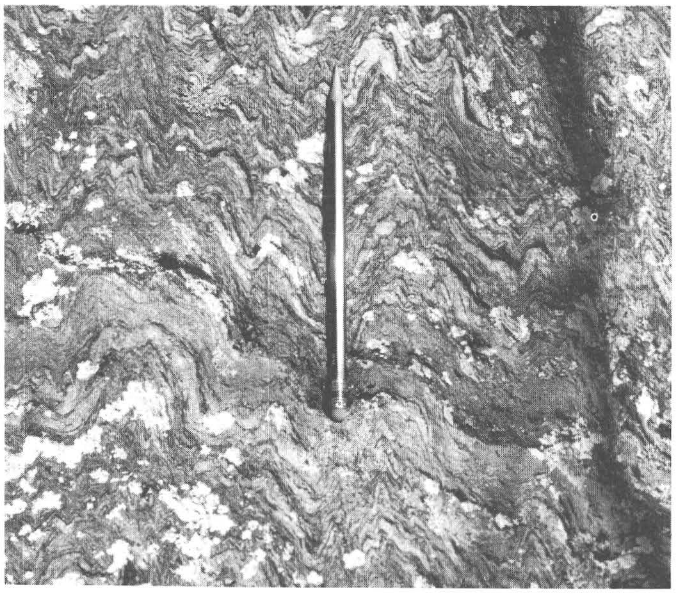


FIGURE 96.—Crinkles with steep axial planes and northeast-trending axes in biotite-muscovite schist and gneiss of the Blue ridge thrust sheet southeast of the Grandfather Mountain window. At altitude of 1,710 feet near head of Cove Branch (area J–5, pl. 1).

foliation and layering (diagram *D*–1, fig. 5, pl. 5) cluster around a single broad maximum similar to the foliation and layering maximum in the Blue Ridge thrust sheet northwest of the fault zone.

Mineral lineation is conspicuous. It is marked by alinement of recrystallized mineral grains and aggregates, elongation of porphyroclasts, and streaking and grooving on foliation planes and is indistinguishable from lineation in low-grade polymetamorphic rocks northwest of the fault zone. Most lineations are subhorizontal and trend northeast, parallel to lineation in polymetamorphic rocks southeast of the zone, but many plunge southwest or south, forming a poorly defined partial great-circle girdle in the statistical diagram (diagram *D*–2, fig. 5, pl. 5).

No large folds are apparent in the rocks of the Brevard zone, but minor folds and crenulations are common, especially in phyllonitic schist and layered gneiss in tectonic slices within the zone. The minor folds in the gneissic rocks are isoclinal or subisoclinal and have steeply dipping axial planes; they commonly have slip cleavage parallel to axial planes. Layers maintain approximatey uniform thicknesses parallel to axial planes. Fold axes show wide variation in attitude within single outcrops, and some individual axes are visibly curved.

Many of the phyllonitic rocks have steeply dipping slip cleavage planes which offset and crenulate the older foliation. Crenulations range in amplitude and wavelengths from a fraction of an inch to several inches. Their axes are approximately parallel to the other fold axes, and they show similar variation in attitude in single outcrops.

The attitudes of the minor folds and crenulations (diagram *D*–3, fig. 5, pl. 5) have a distribution pattern similar to that of the mineral lineation. The shape of the folds and the distribution of their axes in the spherical projection in a great circle rather than in a single maximum indicate that they are passive flow folds (Donath and Parker, 1964).

The geometry of the structures in the belt of rocks in the Inner Piedmont having polymetamorphic features related to the Brevard fault zone is summarized in the diagrams in columns *E* and *F* of figure 5, pl. 5. The diagrams in column *E* show structures in rocks thoroughly sheared and completely recrystallized under medium- or low-grade conditions adjacent to the fault zone. The diagrams in column *F* show orientation of structures in the rocks partially sheared and recrystallized under medium-grade conditions along the southeast side of the belt of polymetamorphic rocks.

Foliation and layering strike northeast and dip southeast at moderate to steep angles which increase as the fault zone is approached; poles cluster symmetrically around high point maxima (diagrams *E*–1 and *F*–1) which are more sharply defined than those in the nonpolymetamorphic rocks to the southeast.

Mineral lineation is much more conspicuous in the polymetamorphic rocks than in the rocks to the southeast. It is marked by streaking of flakes and clusters of new micas and recrystallized quartz-feldspar aggregates on foliation planes and by alinement of long dimensions of feldspar porphyroclasts. The lineation lies nearly parallel to the strike of the foliation planes. Projections of lineation in the statistical diagrams (*E*–2 and *F*–2) show nearly symmetrical concentrations around high point maxima having slight northeast plunges.

No major folds are suggested by the map patterns; bodies of mica schist and Henderson Gneiss form long strips and concordant lenses that resemble megaboudins in the layered gneiss country rock. Minor folds (fig. 97) are subisoclinal and have axes trending northeastward, parallel to mineral lineation, and axial planes that generally dip steeply to moderately southeast. Near the Brevard fault zone, axial planes are steep or vertical, and the folds are more tightly appressed. Many layers maintain rather uniform thicknesses in these folds, although most

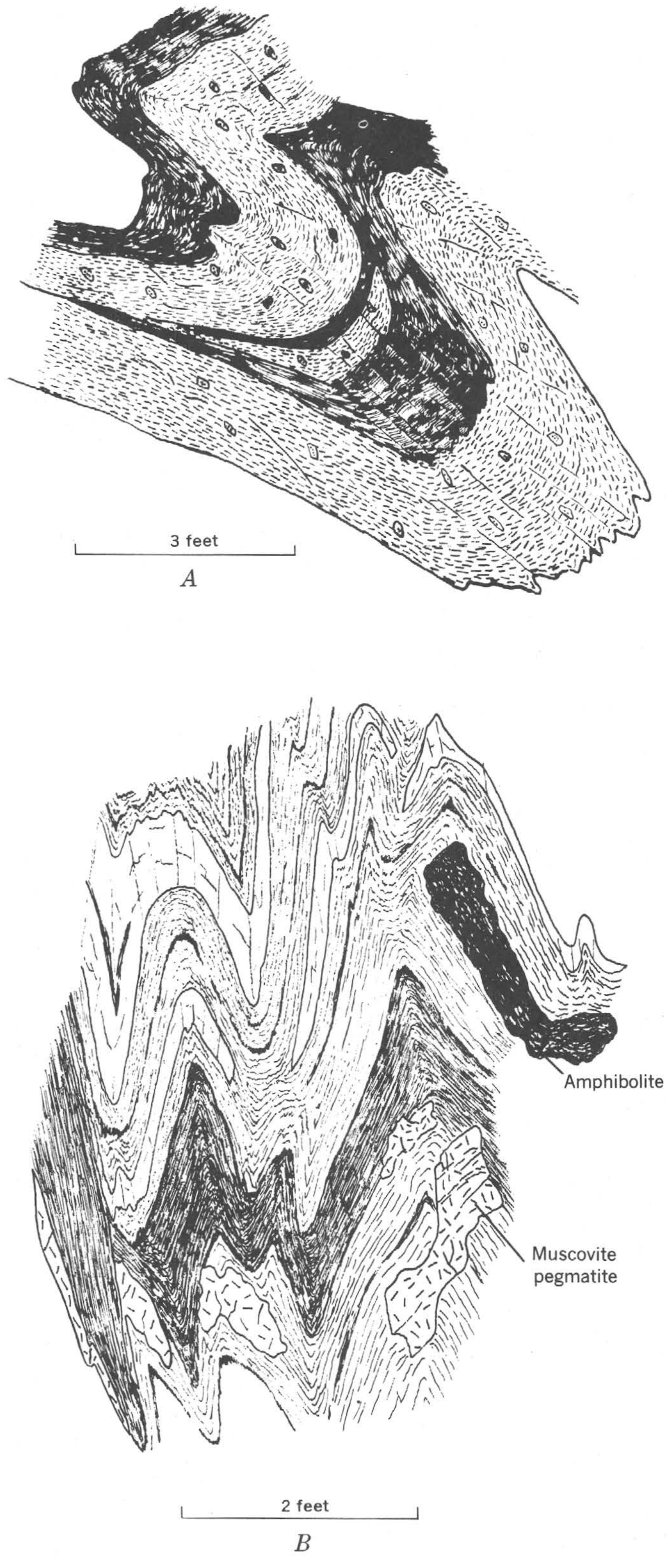



FIGURE 97.—Folds in polymetamorphic rocks of the Inner Piedmont belt near the Brevard fault zone. *A*, Folds in interlayered fine-grained dark-gray biotite gneiss and medium-grained Henderson Gneiss containing potassic feldspar augen elongated parallel to the fold axes. Direction of view is northeast, approximately parallel to fold axes. Note incipient fracture cleavage parallel to axial planes. Saprolite exposure on shore of Lake James, 0.85 mile N. 88° W. of Linville Church (area E–9, pl. 1). *B*, Folds in interlayered light- and dark-gray biotite gneiss and white quartzo-feldspathic gneiss containing boudins of amphibolite and pods of cataclastic muscovite pegmatite. Direction of view is northeast, parallel to fold axes. Roadcut in saprolite on northeast side of road 1.2 miles S. 55° E. of Fairview Church (area F–8, pl. 1).

layers are conspicuously thickened in the apices, but have approximately uniform thickness measured parallel to the trace of the axial plane. In some layers, slip cleavage is locally developed parallel to axial planes. Boudins and blocks of pegmatite, amphibolite, and other less ductile rocks show no consistent geometric relation to fold noses and must have been produced by shearing parallel to layering prior to formation of the folds.

### RELATION BETWEEN STRUCTURES IN THE BREVARD FAULT ZONE AND STRUCTURES IN ROCKS TO THE NORTHWEST

The northwest-trending cataclastic mineral lineation in the rocks of the Grandfather Mountain window and in the low-grade polymetamorphic rocks near the sole of the Blue Ridge thrust sheet (fig. 33) is in the *a* direction, parallel to tectonic transport of the Blue Ridge and Tablerock thrust sheets. Similar cataclastic lineation in the Brevard fault zone and in polymetamorphic rocks to the southeast is subhorizontal and trends northeast, parallel to the strike of the zone. It is believed to be an *a* lineation formed during strike-slip movement along the Brevard fault zone (Reed and Bryant, 1964b).

Previously we interpreted the swing of the cataclastic lineation in the rocks of the Blue Ridge belt near the Brevard zone as the result of drag along the Brevard and concluded that movement along the Brevard was therefore right lateral and younger than thrusting (Reed and Bryant, 1964b). However, the statistical orientation diagrams (fig. 5, pl. 5) show that the geometry of the structures in the thrust sheets northwest of the Brevard is similar to that in the Brevard zone itself and that the change in direction of the cataclastic lineation is not as apparent in autochthonous basement rocks as in the thrust sheets in the Grandfather Mountain window near the Brevard zone. Rotation of the cataclastic lineation by drag would require concomitant rotation of planes containing the lineation, unless, by coincidence, the rotation occurred around an axis exactly perpendicular to the planes. The planes containing the lineation strike consistently northeast and dip southeast and show no evidence of rotation by drag.

In the orientation diagrams (fig. 5, pl. 5), poles to foliation and layering in the Brevard zone (diagram *D*–1) and in the Blue Ridge thrust sheet between the Brevard and the Grandfather Mountain window (diagram *C*–1) form single high maximums in about the same position as poles to foliation in autochthonous basment rocks in adjoining parts of the window (diagram *A*–1). In the Tablerock thrust sheet, the cataclastic mineral lineation lies on bedding planes and foliation planes parallel to bedding; poles of these planes cluster around a point maximum in a different position than the foliation and layering maximums in the adjoining rocks (diagram *B*–1). Mineral lineation in the Tablerock thrust sheet is involved in the swing in regional lineation and forms a partial great-circle girdle in the diagram (diagram *B*–2) similar to the lineation girdles in the Blue Ridge thrust sheet (diagram *C*–2) and the Brevard zone (diagram *D*–2). If the swing in lineation in diagram *B*–2 were due to rotation around an axis normal to bedding and foliation in the Tablerock thrust sheet (diagram *B*–1), it should have rotated foliation and layering planes in the Blue Ridge thrust sheet and produced a small-circle girdle in the foliation (diagram *C*–1). Conversely, if the lineation swing in the Blue Ridge thrust sheet (diagram *C*–2) were due to rotation around the corresponding layering and foliation maximum (diagram *C*–1), it should have produced a small-circle girdle in the bedding and foliation diagram for the Tablerock thrust sheet (diagram *B*–1). As neither diagram *B*–1 nor *C*–1 shows any indication of such rotation, it seems unlikely that the lineation swing is due to simple rotational drag.

There is no indication that the northeast-trending cataclastic lineation in the Brevard zone has overprinted an older northwest-trending lineation in the rocks to the northwest. All the lineations are petrographically similar, and nowhere are diversely oriented lineations superimposed. The smooth swing in lineation trends and the resulting continuity in the girdles in diagrams *B*–2, *C*–2, and *D*–2 could hardly have resulted from overprinting an older lineation by a younger one. The similarity in lineation pattern between the Brevard zone (diagram *D*–2) and the adjacent Blue Ridge thrust sheet (diagram *C*–2) is particularly striking. This similarity could not be the result of rotation of older lineation or of the eradication of older lineation and formation of new lineation during movement along the Brevard, for in the Brevard zone itself the old lineation should have been either entirely rotated or completely destroyed by the intense shearing and almost complete recrystallization that took place, as shown by the blastomylonitic and phyllonitic rocks in the Brevard zone.

The northeast-trending cataclastic *a* lineation along the Brevard zone is, therefore, probably of the same age as northwest-trending cataclastic *a* lineation in rocks of the Blue Ridge thrust sheet and Grandfather Mountain window. The swing in lineation trends in the thrust sheets northwest of the Brevard must be due to change in direction of tectonic transport, from nearly horizontal and northeast along the southeast side of the Brevard to northward and upward near the northwest side of the zone to northwestward in the Tablerock and Blue Ridge thrust sheets farther from the Brevard zone. Thus, the lineation pattern indicates that strike-slip movement along the Brevard fault zone was contemporaneous with northwestward thrusting of rocks of the Blue Ridge.

Axes of folds and crenulations in the Brevard zone and in the flanking rocks are generally parallel or subparallel to the cataclastic lineation. The close similarity between the orientation pattenrs of the fold axes (diagrams *A*–3 through *F*–3, fig. 5, pl. 5) and of the lineation (diagrams *A*–2 through *F*–2) indicates that most of the fold axes are in the *a* direction and that they formed at the same general time as the lineation. Many folds in the Blue Ridge thrust sheet north and west of the Grandfather Mountain window also have axes parallel to cataclastic mineral lineation (fig. 1, pl. 5, diagram *A*–4). The consistent parallelism between fold axes and cataclastic lineation in the zone between northwest-trending lineation in the Blue Ridge and Tablerock thrust sheets and northeast-trending lineations in the polymetamorphic rocks southeast of the Brevard zone is consistent with the conclusion that the two lineations are contemporaneous.

### RELATION BETWEEN STRUCTURES IN POLYMETAMORPHIC ROCKS SOUTHEAST OF THE BREVARD ZONE AND STRUCTURES IN OTHER ROCKS OF THE INNER PIEDMONT

Shearing and recrystallization during retrogressive metamorphism have produced a structural pattern in the polymetamorphic rocks of the Inner Piedmont adjacent to the Brevard fault zone that is quite different from that of the unaffected rocks farther southeast. Although the geometry and symmetry of the two groups of structures is similar, those in the polymetamorphic rocks are apparently related to movement along the Brevard fault, whereas those in the rocks to the southeast date from the earlier high-grade regional metamorphism.

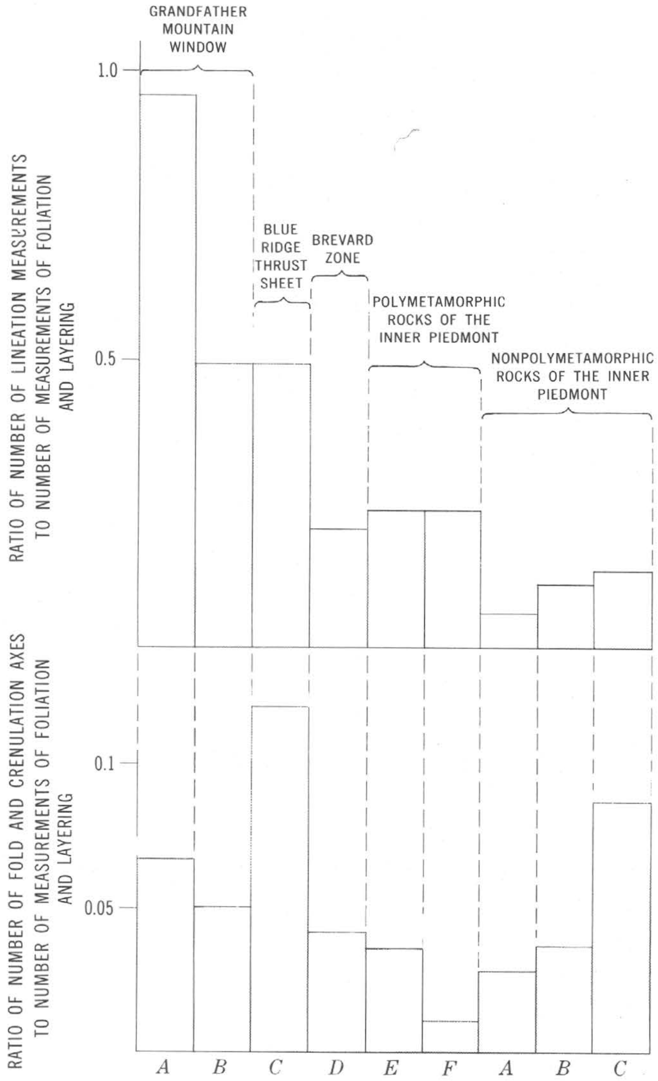



FIGURE 98.—Graphs showing ratios of numbers of measurements of fold and crenulation axes and mineral lineation to numbers of measurements of layering and foliation in rocks of the Brevard fault zone and adjacent parts of the Grandfather Mountain window, Blue Ridge thrust sheet, and Inner Piedmont belt. Letters under first six graphs, *A–F*, correspond to areas for which orientation diagrams are given in figure 5, plate 5; letters under last three graphs, *A–C*, to figure 4, plate 5.

As attitudes of layering or foliation can be measured in almost every outcrop of bedrock or saprolite, the degree of development of mineral lineation and the relative abundance of minor folds can be estimated by comparing the number of attitudes of mineral lineation or fold axes measured in a given area with the number of measurements of foliation and layering in the same area. Figure 98 shows relative degree of lineation development and abundance of minor folds in areas corresponding to the columns of figures 4 and 5, plate 5. Mineral lineation is most strongly developed northwest of the Brevard zone, but it is far more conspicuous in the Brevard zone and in the polymetamorphic rocks of the Inner Piedmont adjacent to it than in the nonpolymetamorphic rocks farther southeast. Clearly, the northeast-trending mineral lineation in the rocks near the Brevard zone cannot be explained by simple rotational drag of older east-trending mineral lineation of the Piedmont rocks. The lineation map (fig. 33) suggests that locally some rotation of the old mineral lineation in the Inner Piedmont rocks may have occurred, but most, and probably all, of the northeast-trending mineral lineation in the polymetamorphic rocks is younger. Nowhere, however, have both lineations been recognized in the same outcrop.

Minor folds are most abundant, or at least most easily measured, in the sillimanite schist of the Inner Piedmont, and they are considerably less conspicuous in the other nonpolymetamorphic Piedmont rocks. They are least abundant along the southeast edge of the belt of polymetamorphic rocks and become more common as the Brevard zone is approached. This distribution suggests that the east-trending folds in the Piedmont rocks were obliterated during shearing and retrogressive metamorphism along the Brevard zone and that continued movement in the polymetamorphic rocks produced new folds, with northeast-trending axes, which increase in abundance and intensity closer to the fault zone. Only locally (diagram *A*–3, fig. 4, pl. 5) is there evidence that some of the older east-trending folds are rotated, but these rotated folds were obliterated in the more strongly sheared rocks to the northwest.

The inferred age relation between the two sets of folds and the lineations associated with them is supported by their relations to the granitic rocks and pegmatites. Discordant granitic and pegmatitic dikes cut across east-plunging folds in the nonpolymetamorphic rocks in the Inner Piedmont, but similar pegmatites were sheared and boudinaged before formation of northeast-trending folds in the polymetamorphic rocks.

## DIABASE

A remarkably straight and continuous dike of unmetamorphosed olivine diabase a few feet to several hundred feet thick enters the Grandfather Mountain area near Drexel (area I–9, pl. 1) and extends northwestward for more than 15 miles. It passes without deflection from the Inner Piedmont belt across the Brevard fault zone and the Linville Falls fault and

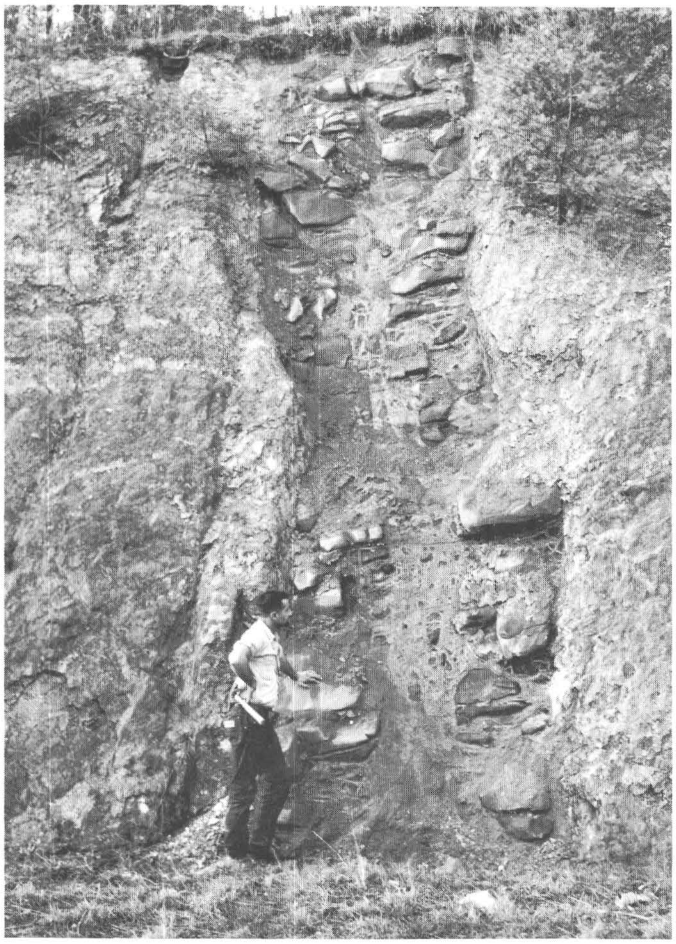


FIGURE 99.—Dike of unmetamorphosed diabase of Late Triassic(?) age in sheared and retrogressively metamorphosed gneiss of the Inner Piedmont in the Brevard fault zone. Note spheroidal weathering. Roadcut in saprolite on east side of Wilson Creek southwest of Adako (area G–7, pl. 1).

into the basement rocks in the Grandfather Mountain window. Smaller disconnected diabase dikes locally parallel the main dike, and a few are found along the same trend in the basement rocks at several places along Wilson Creek as far northwest as the mouth of Harper Creek (area F–6, pl. 1). Keith and Sterrett (1954) mapped a large and nearly continuous diabase dike along the same trend for at least another 15 miles southeast of the Grandfather Mountain area. Diabase has been found nowhere else in the Grandfather Mountain window or in the Inner Piedmont part of the area, but a few dikes and a sill too small to map have been found in rocks of the Blue Ridge thrust sheet near the mouth of Fall Branch (area C–5).

The diabase is a fine- to medium-grained dark-blue-gray rock which weathers to shades of ochre on exposed surfaces. It is more resistant to weathering than most of the enclosing rocks and forms abundant round boulders and cobbles in the soil and a few natural outcrops along streams (fig. 99).

The diabase has ophitic textures in which slender laths of twinned and strongly zoned plagioclase (generally bytownite) are enclosed in poikilitic grains of augite and olivine or intergranular textures in which olivine and augite grains are interstitial to the plagioclase laths (fig. 100). In a few specimens, augite and olivine also form anhedral to subhedral phenocrysts. The rock consists of 40 to 60 percent plagioclase, 20 to 35 percent augite, 1 to 5 percent magnetite, 0 to 25 percent olivine, and locally contains a few grains of basaltic hornblende and biotite.

In the centers of the larger dikes, the diabase is medium grained and generally has ophitic textures in which the plagioclase laths are 1 to 2 mm long and the poikilitic augite and olivine grains are 1 to 2 mm in diameter (fig. 100). In the small dikes and in chilled margins within 1 or 2 feet of the contacts of the larger bodies, the rock is finer grained and textures are more commonly intergranular. Plagioclase laths are 0.2 to 0.5 mm long, and intergranular grains of augite and olivine are 0.1 to 0.2 mm in diameter.

The unmetamorphosed character of the diabase, and the undeflected northwest trend of the principal dike athwart all other structures clearly shows that

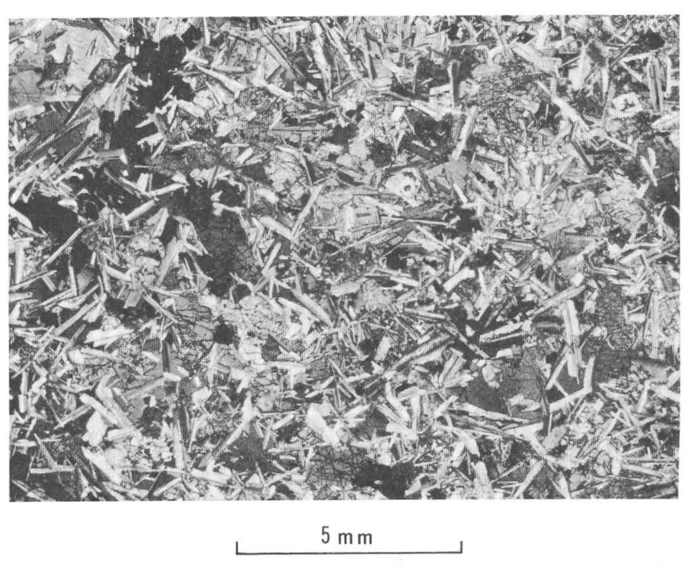



FIGURE 100.—Photomicrograph of diabase. Unmetamorphosed diabase from roadcut on North Carolina Highway 18 (area H–8, pl. 1) 1 mile northeast of Chesterfield. Ophitic texture. Laths of bytownite enclosed in grains of augite and olivine. From interior of dike at a place where the dike is about 100 feet thick.

the diabase was emplaced after all the metamorphisms and major deformations of the enclosing rocks. The sharp chilled contacts of the dikes show that the diabase was intruded into relatively cool wallrocks; lack of offset of layers in the country rocks across the dikes shows that no faulting was involved during their emplacement.

The trend of the principal diabase dike in the Grandfather Mountain area is parallel with similar dikes elsewhere in the Piedmont in North Carolina, South Carolina, and Georgia (King, 1961; Lester and Allen, 1950). These dikes are generally considered to be of Late Triassic age because of their relations to rocks of the Upper Triassic Newark Group, but they may be as young as Early Cretaceous (Reinemund, 1955). This dike is clearly part of the same dike swarm, and the diabase in the Grandfather Mountain area is therefore probably of Late Triassic age but is possibly younger.

# MAJOR FAULTS BOUNDING TECTONIC UNITS

## THRUST FAULTS

### LINVILLE FALLS FAULT

The Linville Falls fault forms the boundary of the Grandfather Mountain window. Along it Precambrian crystalline rocks of the Blue Ridge thrust sheet are carried over autochthonous Precambrian plutonic rocks and metamorphosed upper Precambrian sedimentary and volcanic rocks and over Cambrian and Cambrian(?) rocks in the Tablerock thrust sheet. Branches of the Linville Falls fault cut the Tablerock fault east of Shortoff Mountain (area E–8, pl. 1). The imbricate faults in areas B–9 and C–9 and the faults marked by slices of Chilhowee quartzite in the Blue Ridge thrust sheet on the northwest side of the window are somewhat older subsidiary faults of the Linville Falls fault, as the map pattern shows that they are cut by it. The Tablerock fault may also be an older subsidiary thrust.

On the southeast side of the Grandfather Mountain window, the Linville Falls fault dips southeast, parallel to foliation in adjacent parts of the window and the Blue Ridge thrust sheet. On the north and west sides of the window, there is marked structural discordance between the fault and the cleavage in the autochthonous window rocks (fig. 32); locally, foliation in the nearby window rocks is dragged in a sinistral sense into parallelism with the fault. The fault plane is parallel to cataclastic foliation in the overriding rocks.

Where the Linville Falls fault overlies the Tablerock thrust sheet, no discordance is apparent between foliation in the Blue Ridge thrust sheet above the fault and cleavage and bedding in the Tablerock thrust sheet beneath it, although mapping shows discordance in the gross structure.

The Linville Falls fault is best exposed on the west side of the Linville River 100 yards upstream from the end of the National Park Service trail to the head of Linville Falls (area D–6, pl. 1). There, rudely layered Cranberry Gneiss overlies green sericite quartzite of the Tablerock thrust sheet (fig. 101). The contact, which dips gently west, is marked by 6 to 18 inches of white to green finely laminated blastomylonite (fig. 102) which is parallel to bedding in the quartzite and to foliation in the gneiss. Several other thin blastomylonite layers occur in the quartzite 100 to 200 feet south of the exposure of the main fault.

The fault plane separating Cranberry Gneiss above from Shady Dolomite beneath is exposed near the base of a prominent cliff on the west side of the

FIGURE 101.—Linville Falls fault at its type locality. Exposure on west side of Linville River about 100 yards upstream from end of National Park Service trail to head of Linville Falls (area D–6, pl. 1). Massive rock is coarse-grained blastomylonitic quartz monzonite gneiss of the Blue Ridge thrust sheet. It overlies less resistant zone of blastomylonite, a layer of which interfingers with the quartz monzonite. Trash in foreground lies on outcrop of quartzite of the Chilhowee Group in the Tablerock thrust sheet.


FIGURE 102.—Photomicrograph of blastomylonite from Linville Falls fault at Linville Falls (fig. 101). Partly recrystallized mylonite containing recrystallized quartz and iron-rich muscovite and some unrecrystallized quartz and feldspar. Crinkles trend northeast, parallel with late folds in nearby rocks of the Tablerock thrust sheet (fig. 67). Section cut parallel to mineral lineation.

valley of the north fork of the Catawba River about 0.3 mile north of Linville Caverns (area C–7, pl. 1). Near the fault, the gneiss has been reduced to a dark-green siliceous blastomylonite. The fault is marked by a 6-inch quartz vein. The Shady Dolomite is shattered and silicified in a 1- to 2-foot zone below the fault. The fault plane is parallel to layering and foliation in the gneiss; bedding is not apparent in the dolomite.

The fault is also well exposed on the north side of the window in a roadcut along North Carolina Highway 194 about half a mile east of Bowers Gap (area E–3, pl. 1; fig. 103). There, blastomylonitic Cranberry Gneiss with cataclastic foliation dipping north overlies metasiltstone of the Grandfather Mountain Formation, which has east-dipping cleavage. The rocks are separated by a gouge zone half an inch thick, and the cleavage in the underblock is disturbed in a zone 2 to 6 feet thick below the fault. The fault dips 35° N.

The fault is locally exposed on the southeast side of the window on the steep slopes of Stone Mountain at an altitude of 2,450 feet, S. 72° E. of Harris Gap (area J–4, pl. 1). There, garnet-biotite-muscovite schist, containing porphyroclasts of muscovite as much as 7 mm in diameter, overlies a few feet of amphibolite and strongly sheared layered quartz-feldspar and biotite-quartz-feldspar gneiss. The schist near the contact has fragments of biotite

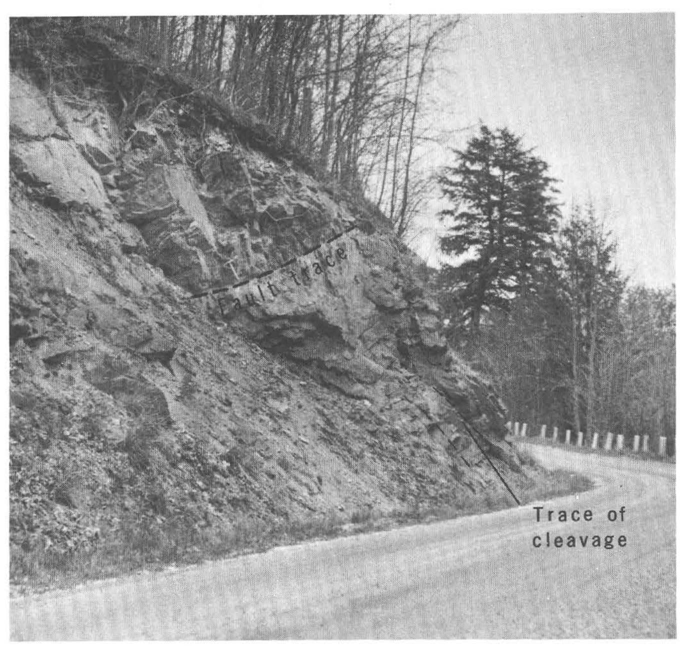



FIGURE 103.—Linville Falls fault in roadcut along North Carolina Highway 194 about one-half of a mile east of Bowers Gap (area E–3, pl. 1). Cataclastic foliation in Blue Ridge thrust sheet is parallel with fault plane, which dips about 35° N. Trace of east-dipping cleavage in siltstone of the Grandfather Mountain Formation shown. Direction of view is northeast. Photograph by Frank G. Lesure, U.S. Geological Survey.

schist 1 to 2 inches long that appear to be tectonic inclusions. Recrystallized plagioclase in rocks on either side of the contact is $An_{15}$. In several places along this segment of the Linville Falls fault, similar exposures occur. Amphibolite is commonly found at the fault and ranges from a few inches to 50 feet thick.

Farther southwest, the fault is exposed on the southeast side of the window in a small quarry and in an adjacent roadcut (area F–8, pl. 1) along North Carolina Highway 181 on the east side of Steels Creek about 0.7 mile northwest of Smyrna Church. In the quarry, layered biotite gneiss overlies felsic metavolcanic rocks. The fault plane is marked by a 10- to 20-foot quartzite slice; several other thin slices are intercalated with the gneiss of the overriding block within a few feet of the fault. In the roadcut, a single 2- to 3-foot quartzite slice separates gneiss from the underlying felsic volcanic rock.

In many places along the fault, lenses of quartzitic blastomylonite or arkosic quartzite a few feet thick resembling rocks of the Chilhowee Group are found with blastomylonite and mylonite derived from feldspathic gneiss.

The sheared rocks along the fault all have strong lineation parallel to cataclastic lineation in nearby parts of the Blue Ridge thrust sheet. In the southwestern part of the window (pl. 1), the fault is locally vertical or overturned, and the lineation along it is horizontal, indicating that in that area the latest displacement along the fault was strike-slip associated with movement along the Brevard fault zone. Roadcuts along North Carolina Highway 80 by the Lake Tahoma Dam (area A–10) expose biotite-bearing arkosic quartzite of the Chilhowee Group and associated blastomylonite and mylonite along the steeply dipping segment of the fault on the west side of the window.

### STONE MOUNTAIN FAULT

The Stone Mountain fault carries rocks of the Blue Ridge thrust sheet over rocks of the Unaka belt in northeastern Tennessee. It occupies a position analogous to the Great Smoky fault to the southwest, but the connection shown on plate 3 is only an interpretation; detailed published maps are lacking between northeast Tennessee (King and Ferguson, 1960), the Hot Springs window (Oriel, 1950), and the Great Smoky Mountains (Hamilton, 1961; Hadley and Goldsmith, 1963; King, 1964b; Neuman and Nelson, 1965).

In the northwest part of the Grandfather Mountain area, the Cranberry Gneiss overrides similar but less metamorphosed plutonic rock and little metamorphosed Cambrian sedimentary rocks along the Stone Mountain fault and related faults (pl. 1). The relations between the faults are complex and somewhat obscure because their exact positions and the configuration of their intersections are obscured by colluvium. A reexamination of parts of the area, however, suggests a somewhat different interpretation than that of King and Ferguson (1960) or Rodgers (1953a).

The gross pattern of faults (pl. 1) resembles that mapped by Rodgers (1953a) more than that by King and Ferguson (1960). Rodgers' map (fig. 104*A*) shows the basement rocks in the vicinity of Dark Ridge Creek (area C–2, pl. 1) within the Mountain City window. He inferred that the Unaka Mountain and Stone Mountain faults were overridden by crystalline rocks along the Snow Mountain fault. King and Ferguson (1960, p. 77–78, fig. 18) offered a "speculative explanation" of the more complex map pattern revealed by further mapping (fig. 104*B*). They suggested a well-defined sequence of faults; the rocks are broken successively along the Stone Mountain fault, the Poga fault, and the Unaka Mountain fault. They believed the strip of basement rocks at the southern foot of Little Stone Mountain (area C–1, pl. 1) to be unconformably beneath rocks of the Chilhowee Group.

We believe that the Unaka Mountain and Snow Mountain faults are connected as shown in figure 104*C* and that the Stone Mountain and Poga faults of King and Ferguson (1960, fig. 18) are an upper branch of the same fault. We have termed the entire system of faults the Stone Mountain fault. The Chilhowee Group rocks on Little Stone Mountain are interpreted as a subsidiary slice below the main fault. The basement rocks in the Dark Ridge Creek area are believed to have been overridden by the Blue Ridge Creek thrust sheet along the lower branch of the Stone Mountain fault and to have overridden the rocks of the Chilhowee Group in the Doe River inner window of the Mountain City window along the complex group of thrusts south of Little Stone Mountain. Thus, they occupy an intermediate sheet between the main Blue Ridge thrust sheet and the Mountain City window. The long westward-protruding tongue of basement rocks shown in the northern part of figure 104*C* is believed to have been carried farther northwest than the main mass of the Blue Ridge thrust

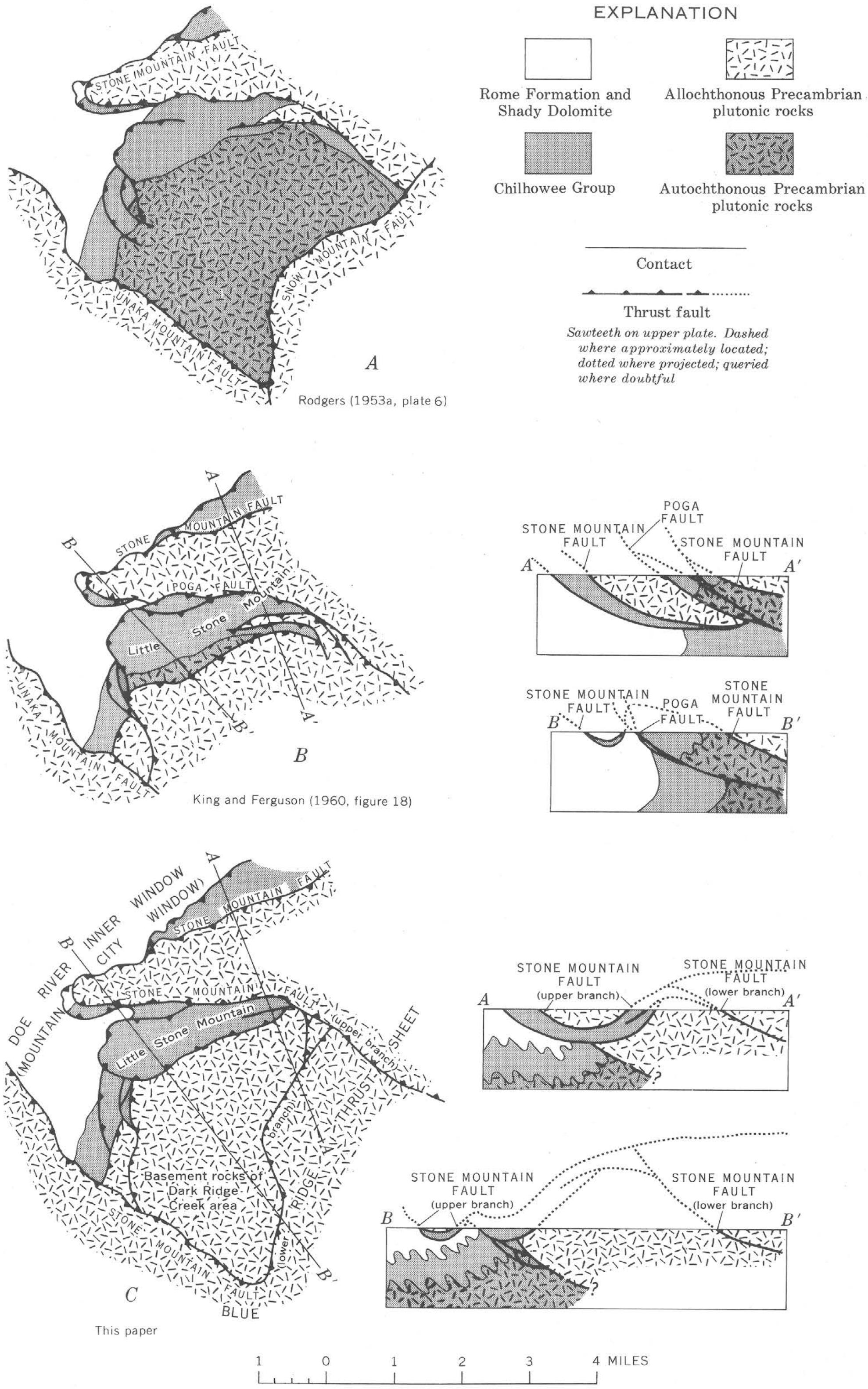



FIGURE 104.—Map and sections of the northwest corner of the Grandfather Mountain area showing various interpretations of the structures on the northwest margin of the Blue Ridge thrust sheet.

sheet by somewhat later movement along the upper branch of the Stone Mountain fault.

The upper branch of the Stone Mountain fault extends southeastward into the Linville quadrangle and apparently dies out north of Long Ridge (area E–2, pl. 1). It is marked by slices of Chilhowee quartzite along much of its length, and where it crosses Beech Creek (area D–1, pl. 1), it dips north rather than south as shown by King and Ferguson (1960, pl. 1 and fig. 18). The same is true at several other places where its dip can be determined.

The Snow Mountain fault of Rodgers (1953a) connects with the Unaka Mountain fault of Rodgers 1953a) and King and Ferguson (1960, fig. 18) near the site of the former Dark Ridge School (area D–2, pl. 1) and extends north to Flat Springs Branch (area D–1) in the Elk Mills 7½-minute quadrangle. This extension cannot be traced with certainty, but the fault is generally marked by phyllonite and locally, by slices of Chilhowee quartzite. The nature of the intersection between this fault and the upper branch of the Stone Mountain fault is uncertain. There is no marked difference in lithology across the segment of the Stone Mountain fault between Dark Ridge school and Flat Springs Branch (Snow Mountain fault of Rodgers), but west of the fault, shearing is generally confined to phyllonite zones, and some of the intervening rock is not affected by shearing and retrogressive metamorphism. Lineation is much less conspicuous in this block than in the Blue Ridge thrust sheet (fig. 33).

We were unable to recognize the difference in the basement rocks immediately south of Little Stone Mountain and those farther south. We therefore believe that the fault separating the Chilhowee rocks on Little Stone Mountain from the basement rocks to the south lies along the contact between the Chilhowee and the basement rocks, rather than in the basement as shown by King and Ferguson (1960, pl. 1). Map relations near the head of Flat Springs Branch indicate that this fault dips north (fig. 104*C*) rather than south as they believed (fig. 104*B*).

No complete exposures of the Stone Mountain fault were found in the Grandfather Mountain area, but there are many exposures of blastomylonite and phyllonite in the fault zone along the steep north side of Dark Ridge (area C–2, pl. 1) and on the west slope of Dividing Ridge (area D–2). Near the northwest corner of the Linville quadrangle, the fault zone contains breccia composed of fragments of sheared quartzite and perthite in a matrix of green iron-rich chlorite (fig. 105).

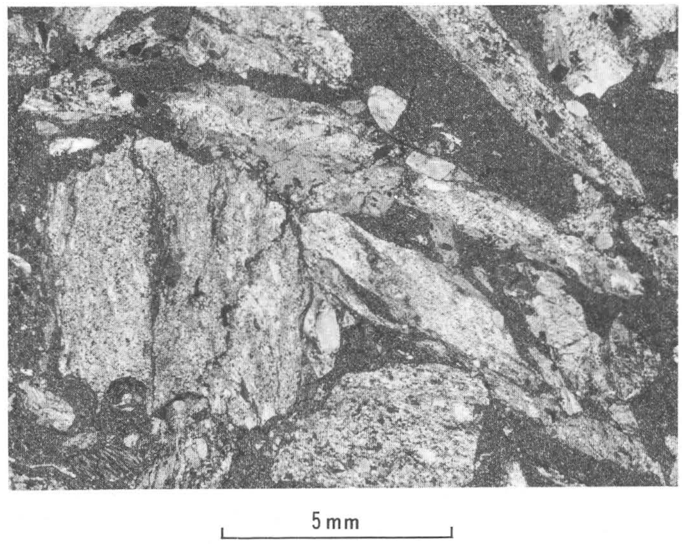



FIGURE 105.—Photomicrograph of breccia from Stone Mountain fault zone 1.0 mile S. 77° E. of northwest corner of the Linville quadrangle (area C–2, pl. 1). Fragments of sheared quartzite and perthite in a matrix of chlorite. Section cut parallel to mineral lineation.

Northwest of Dark Ridge School (area C–2, pl. 1), the Stone Mountain fault dips 5° to 20° SW., except at the Elk River where the dip appears to be as steep as 60°. Discordance between gently south-dipping cataclastic foliation in the Beech Granite above the fault and east-dipping bedding and foliation in rocks beneath the fault is conspicuous along this segment. North of the point where the Stone Mountain fault leaves the contact of the Beech Granite (area D–2), it dips 45° to 70° SE., parallel to foliation in the flanking rocks.

The faults between the basement rocks of the Dark Ridge Creek area and the Doe River inner window (fig. 104*C*) dip from about 40° SE. to nearly vertical. They are marked by phyllonite, phyllonitic gneiss, and slices of quartzite in the crystalline rocks but are more difficult to locate precisely in the quartzites of the Chilhowee Group.

## EXTENT OF THRUSTING

The Grandfather Mountain window lies near the midpoint of the belt of extensive thrusting along the northwestern edge of the Blue Ridge that extends from central Virginia to northern Alabama, where the Appalachian belt is covered by younger sedimentary rocks (fig. 1). The window is perhaps the best available indicator of the minimum amount of northwestward transport of the Blue Ridge thrust sheet. At least 35 miles of movement is required to bring the plutonic rocks of the northwest edge of the Blue Ridge thrust sheet from southeast of the Grandfather Mountain window. Discordant relations between medium- and high-grade metamorphic rocks and the structurally underlying low-grade plutonic rocks around the Grandfather Mountain window suggest the presence of another major fault along which the overlying higher grade rock in the Blue Ridge thrust sheet moved at least an additional 20 miles northwestward.

Little information is available on the amount of transport of the Blue Ridge thrust sheet elsewhere along strike. The map pattern south of Roanoke, Va. (Woodward, 1932), indicates a minimum displacement of 8 miles across an allochthonous footwall block. Estimates based on detailed mapping in the Great Smoky Mountains indicate a minimum of 10 miles displacement on the Great Smoky fault (Neuman and Nelson, 1965, p. 52) and total postmetamorphic northwestward movement of the Blue Ridge thrust sheet of 12 to 24 miles (Hamilton, 1961, p. 46; King, 1964b, p. 121). The northwestward bulge of the Blue Ridge belt in the Great Smoky Mountain region (fig. 1), however, suggests as much as 50 miles of transport. The greatest concentration of thrust faults mapped in the southern Appalachians lies northwest of the Great Smoky Mountains (fig. 1) and may indicate that maximum northwestward transport of the Blue Ridge thrust sheet took place there.

Kessler (1950, p. 30–33) concluded that no major thrust occurs along the margin of the Blue Ridge province in northwest Georgia, but other work just to the north shows that the Great Smoky fault extends into Georgia, where it separates metamorphosed rocks of the Ocoee Series from nonmetamorphosed Paleozoic rocks (Salisbury, 1961, p. 49–50). One of the thrust faults along the western margin of the Blue Ridge belt in Alabama is estimated to have a minimum displacement of 15 miles (Butts, 1940a).

### DIRECTION OF THRUSTING

The pervasive northwest-trending lineation in the rocks of the Grandfather Mountain window and the overlying Blue Ridge thrust sheet is interpreted as an *a* lineation, parallel to the direction of tectonic transport, because of its relation to other structures in the Grandfather Mountain area.

This interpretation is strengthened by regional geologic relations which require northwestward transport of the crystalline rocks of the Blue Ridge thrust sheet over the nonmetamorphosed Paleozoic rocks of the Valley and Ridge belt. To the northwest, basement rocks are concealed beneath nonmetamorphosed Paleozoic sedimentary rocks, so that a large sheet of Precambrian crystalline rocks metamorphosed during the Paleozoic could not have been derived from the northwest. Movement northeastward or southwestward perpendicular to the pervasive lineation and parallel to the regional structural trend would not allow Precambrian rocks metamorphosed to medium or high grade during the Paleozoic to be superposed over Precambrian, upper Precambrian, and lower Cambrian rocks of low metamorphic grade in the Grandfather Mountain window. Minor folds in the window rocks and the later set of minor folds and crinkles in the Blue Ridge thrust sheet are generally asymmetric and overturned to the northwest, and all indicate yielding of the rocks in that direction.

### AGE OF THRUSTING

Direct geologic evidence in the Grandfather Mountain area shows only that thrusting occurred after deposition of the Lower Cambrian Shady Dolomite and before emplacement of the diabase dike of Triassic(?) age. Geologic evidence in other areas and radiometric measurements, however, suggest more precise ages for the thrusts.

West of the Great Smoky Mountains, 60 miles west of the Grandfather Mountain area, movement along the Great Smoky thrust apparently took place during Mississippian time or later (Hadley and others, 1955, p. 406–407). The regional map pattern (pl. 3) suggests that the faults bounding the Shady Valley and Bald Mountain thrust sheets are either approximately the same age as or older than the Great Smoky and that they are in turn overridden by the Stone Mountain fault. If so, the Stone Mountain fault is the youngest fault in the Unaka belt, and movement along it must have occurred in the late Paleozoic or later. The Stone Mountain fault may be equivalent to the Linville Falls fault, although the relations between them may be more complicated.

Medium- and high-grade metamorphism of the Blue Ridge thrust sheet took place about 350 m.y. ago (latest Devonian according to Holmes, 1959). The fact that the sedimentary and volcanic rocks in the Grandfather Mountain window never underwent metamorphism of that grade indicates that the main period of transport over the Grandfather Mountain window took place after the metamorphism of 350 m.y. ago.

#### ORIGIN OF THE GRANDFATHER MOUNTAIN WINDOW

Any theory of the origin of the Grandfather Mountain window must explain the following facts: (1) It is the only known window completely surrounded by Precambrian crystalline rocks of the Blue Ridge belt, (2) cleavage in the window dips to the southeast in the northwest part of the window as steeply as or more steeply than cleavage elsewhere in the window, and (3) no pervasive cleavage of similar orientation cuts the rocks of the Blue Ridge thrust sheet. If the structural high represented by the window were due to warping of the thrust sheet, the cleavage in the window should have gentler southeast dips on the west side of the window than elsewhere, because the Blue Ridge thrust sheet and the Linville Falls fault dip as much as 40° NW. on the west side. If the deformation were due to slip folding along the pervasive cleavage in the window, rocks of the Blue Ridge thrust sheet should be cut by similarly oriented cleavage. In the absence of such cleavage in the Blue Ridge thrust sheet, the different orientations of cleavage in the window and of cataclastic foliation in the Blue Ridge thrust sheet apparently indicate that folding and formation of cleavage in the window rocks and arching of the thrust sheet over the window took place during the main episode of movement along the Linville Falls fault.

One can speculate why the Grandfather Mountain window is unique. The thick sequence of upper Precambrian rocks in the northwest part of the window probably thins both to the northeast and southwest beneath the Blue Ridge thrust sheet. The window may exist because this thick sequence formed an original structural high that was emphasized by movements along the Brevard fault and perhaps by later doming. Some late deformation is indicated by the gentle northwest-trending fold in the northwestern part of the window and in the overlying Blue Ridge thrust sheet.

### BREVARD FAULT

Reed and Bryant (1964b) have discussed the Brevard fault zone and have summarized evidence that it marks a strike-slip fault of great magnitude. The following section is largely summarized from that paper, but a different interpretation of the relationship between thrusting and strike-slip faulting is set forth.

The Brevard zone is a narrow belt of low-grade metamorphic rocks that emerges from beneath the Coastal Plain deposits northeast of Montgomery, Ala., and has a remarkably straight and continuous trace northeastward for at least 325 miles, passing just southeast of the Grandfather Mountain window (fig. 1). It undoubtedly extends northeastward into Virginia, but it has not yet been traced in detail. The Brevard zone marks the southeastern edge of the Blue Ridge geologic belt and separates it from the Inner Piedmont belt to the southeast (King, 1955).

Keith (1905, 1907b) named the low-grade metamorphic rocks in the zone the Brevard Schist and believed that they were of Cambrian age and occupied a narrow synclinal infold in flanking Precambrian rocks of higher metamorphic grade. Jonas (1932) recognized the continuity and unity of the belt and pointed out that many of the low-grade rocks were products of retrogressive metamorphism of the flanking rocks. She concluded that the belt marked a great overthrust fault that carried rocks of the Piedmont northwestward over rocks of the Blue Ridge.

We concur with Jonas that the Brevard is a major fault zone, but believe that it is a strike-slip fault rather than a thrust because of its long straight trace, the width of the belts on either side showing structural and metamorphic effects related to it, tectonic lenses of exotic rocks in the zone, the contrast between rocks on opposite sides of the zone, and the subhorizontal cataclastic $a$ lineation in the zone. Because no sundered geologic features can be matched across the fault in the 135-mile segment that we examined between central Georgia and the Grandfather Mountain area, we inferred that displacement along it must exceed 135 miles and may be much more.

In our previous paper (Reed and Bryant, 1964b, p. 1188), we interpreted the Brevard fault as younger than the Linville Falls fault; the abrupt clockwise swing from northwest-trending cataclastic lineation in the Blue Ridge thrust sheet and Grandfather Mountain window to subhorizontal northeast-trending lineation in the polymetamorphic rocks along the Brevard zone was attributed to drag. We therefore concluded that movement along the Brevard was right lateral and suggested that the parallelism between the Linville Falls fault and the Brevard zone was due to deformation of the older structure during strike-slip movement along the Brevard.

Burchfiel and Livingston (1967) have recently called attention to the close analogy between the Brevard zone and the root zones of the alpine nappes such as the Urseren zone, the Pusteria-Insubric line and others. They infer that the thrust sheets of Blue

Ridge are rooted in the Brevard zone rather than being truncated by it as we had suggested.

Reevaluation of our data and more thorough analysis of the structural geometry indicate, however, that the northwest-trending cataclastic *a* lineation in the rocks northwest of the Brevard is contemporaneous with the cataclastic lineation in the Brevard zone itself and in the polymetamorphic rocks to the southeast. This suggests that northwestward movement of the Blue Ridge and Tablerock thrust sheets was largely concurrent with and directly related to movement along the Brevard. Apparently the thrust faults are rooted along the northwest side of the Brevard zone.

King (1964a, p. 12–14) has pointed out that major srtike-slip faults like the Brevard parallel the structural grain of major mountain systems in many parts of the world and has suggested that there may be a genetic connection between strike-slip faulting and structures generally attributed to lateral compression.

Reed, Bryant, and Myers (1970) have suggested a possible reinterpretation of the relations between strike-slip faulting and thrusting in the Grandfather Mountain area (fig. 106). This interpretation envisions the Brevard as a first-order wrench fault (Moody and Hill, 1956, p. 1213) developed in response to drift of the crustal block southeast of the fault toward the north, in the direction of the heavy arrows on the top surface of the block (fig. 106). The original position of the fault (*A–A′*, fig. 106) was an unknown distance southeast of its present location

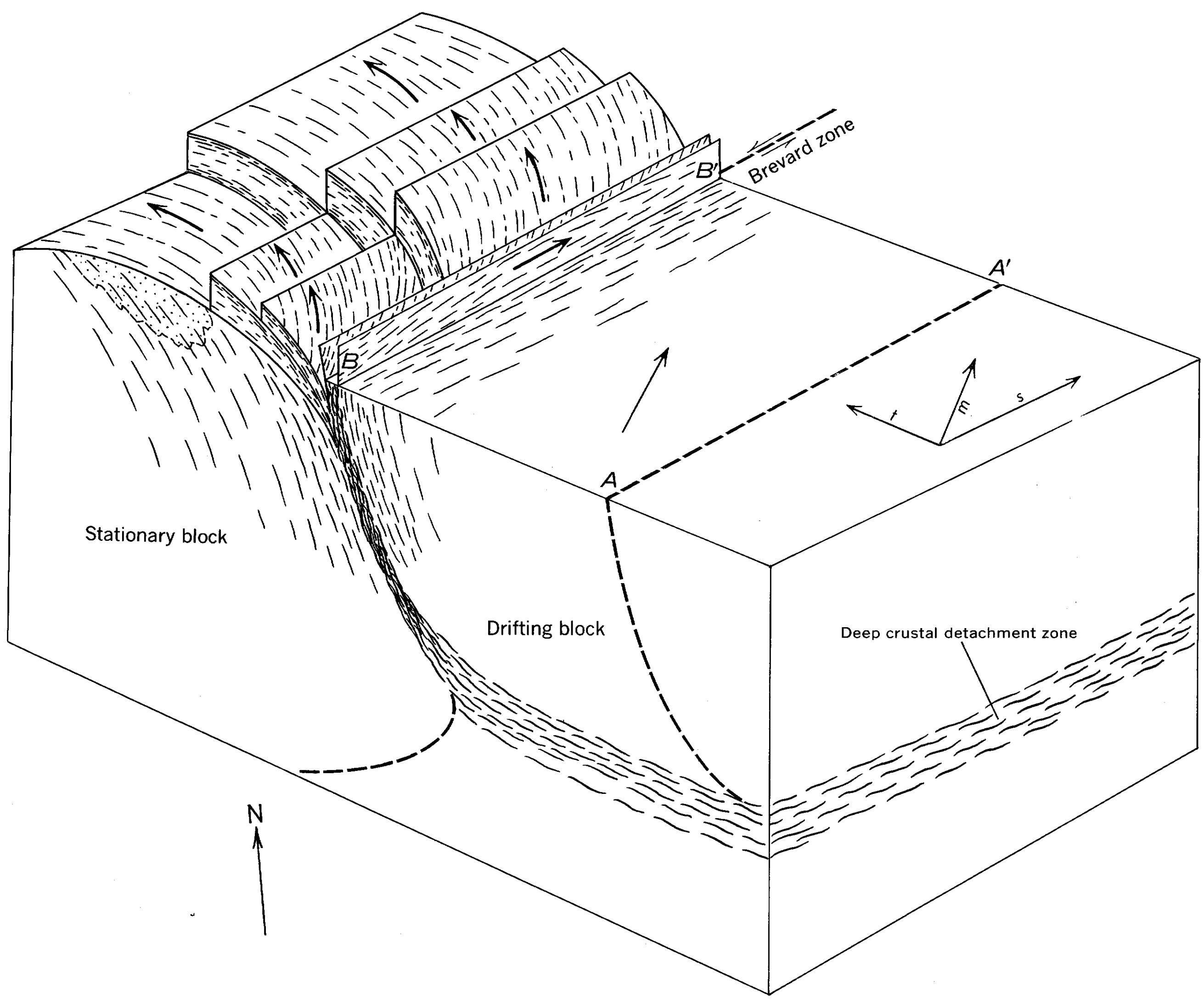



FIGURE 106.—Diagrammatic sketch illustrating posible relation between strike-slip faulting and thrusting.

($B$–$B'$). Northward movement of the drifting block by an amount $m$ would result in left-lateral strike-slip movement ($s$) along the Brevard zone and would requie northwestward migration of the entire zone by an amount $t$. Northwestward migration of the zone is accomplished by downbuckling and thickening of the stationary block and by northwestward thrusting of material from along its southeast edge. Once these thrust sheets reach the crest of the tectonic high northwest of the Brevard, further northwestward movement is facilitated by gravity sliding. The swing in the trend of the cataclastic lineations in the rocks immediately northwest of the Brevard is due to the change from predominately strike-slip motion near the zone to predominately thrust motion farther to the northwest. The trend of the lineation at any point (light lines on the fault surfaces, fig. 106) is in the direction of the vector sum of the local strike-slip ($s$) and thrust ($t$) components.

The total cross-sectional area of the thrust sheets, the amount of thickening of the stationary block, and the depth to the detachment zone at the base of the drifting block must be known in order to calculate the amount of northwestward migration of the Brevard zone necessary to explain the thrusting in the Grandfather Mountain area. In the line of the section on plate 4, the thrust sheets in the Blue Ridge are about 60 km wide and probably at least 5 km thick. If the depth of the detachment zone below the drifting block is assumed to be 20 km and thickening of the stationary block is neglected, only 15 km of northwestward migration of the Brevard is required to explain the thrusting; if the detachment zone were at the base of the crust (about 35 km), less than 9 km of migration would be necessary.

This model of simultaneous strike-slip movement and thrusting requires left-lateral displacement along the Brevard, rather than right-lateral as we previously suggested. There is as yet no conclusive evidence as to the sense of movement along the zone, but the reinterpretation seems more nearly compatible with the observed structural relations in the Grandfather Mountain area than our previous interpretation.

An interesting corollary to this hypothesis is that the diabase dike that crosses the Brevard in the Lenoir quadrangle lies in expected orientation of the rotated shear set, conjugate to the left-lateral strike-slip fault (Cloos, 1955, p. 244–245). Swarms of similarly oriented diabase dikes occur southeast of the Brevard in Alabama, Georgia, and South Carolina (King, 1961); some cut the Brevard zone, and several extend entirely across it, but the dike swarms are absent northwest of the zone.

If the above interpretation of the structural relation between the Brevard fault and the northwestward thrusting of the Blue Ridge thrust sheet is correct, most of the movement on the Brevard fault would have been concomitant with the thrusting and could have started in the middle Paleozoic after the climax of regional metamorphism about 350 m.y. ago. However, the Brevard fault may have been active even during early Paleozoic. Movement along the zone must have ended before the emplacement of the diabase dike, presumably in the Late Triassic.

## TECTONICS

### SUMMARY OF METAMORPHIC AND STRUCTURAL HISTORY

Table 30 is an outline of the structural and metamorphic history of the Grandfather Mountain area. The chronologic order of the principal events that have affected rocks of each of the major tectonic blocks is fairly well established on the basis of structural and petrographic relationships discussed in detail in earlier parts of this paper. However, the dates of many events and the correlations between the histories of the tectonic units prior to their juxtaposition by faulting depend largely on radiometric dating of minerals and on inferences drawn from broader regional relationships that are discussed below.

A brief summary of the metamorphic and structural history of the southern Blue Ridge by Bryant and Reed (1970a) was based on some of the conclusions below but encompassed a larger region.

### REGIONAL SYNTHESIS

Plate 4 is a cross section of the western part of the Appalachians extending through the Grandfather Mountain area and northwestward across the Unaka and Valley and Ridge belts into the Appalachian Plateau. It shows what we believe to be the most reasonable structural interpretation in the light of present knowledge. The segment of the section northwest of the Grandfather Mountain area is based principally on Rodgers (1953a); King and Ferguson (1960); Butts (1940b); Virginia Division of Mineral Resources (1963); and on regional syntheses and interpretations by King (1951, 1955, 1959, 1964a); Rodgers (1950, 1953b, 1964); and Colton (1961). Modern geologic information southeast of the Grandfather Mountain area is not yet sufficient to allow extension

TABLE 30.—*Structural and metamorphic history of the Grandfather Mountain window and vicinity*

| Age | Blue Ridge thrust sheet | Grandfather Mountain window: Tablerock thrust sheet | Grandfather Mountain window: Autochthonous rocks | Inner Piedmont | Major fault movements: Thrusting | Major fault movements: Brevard fault |
|---|---|---|---|---|---|---|
| Upper Triassic(?) | Intrusion of diabase. | | Intrusion of diabase. | Intrusion of diabase. | | |
| Late Paleozoic or Early Mesozoic | Gentle warping of thrust sheet northwest and northeast of window(?).<br>Formation of northwest-trending joints in Brevard fault zone (and elsewhere?).<br>Formation of northeast-trending (and north-west-trending?) joints.<br>Incomplete low-grade metamorphism southeast of window.<br>Formation of open folds and crinkles.<br>Low-grade partial metamorphism of plutonic rocks.<br>Formation of mineral lineation.<br>Cataclasis and recrystallization at medium grade southeast of window.<br>Formation of isoclinal folds. | Formation of joints.<br>High-angle faulting in southwestern extension of window.<br>Formation of mineral lineation and late open folds and crenulations; deformation of earlier fold axes.<br>Metamorphism at low grade.<br>Formation of tightly appressed folds and associated cleavage. | Formation of joints, slip cleavage, and small-scale folds.<br>Metamorphism climaxed 350 m.y. ago but may have lasted from 450 to 250 m.y. ?—?—?——?—?—?<br>Formation of phyllonite zones in granitic rocks and formation and deformation of folds in layered gneiss: {Reactivation of all existing cleavage; formation of mineral lineation.<br>Medium-scale folding; formation of new cleavage; obliteration of early cleavage in northwestern part of window; possible reactivation of cleavage in southern part and in basement rocks. | Formation of north-west-trending joints in Brevard fault zone.<br>Latest movement along Brevard fault; formation of blastomylonite and partial recrystallization at low metamorphic grade along narrow belt.<br>Northwestward migration of Brevard fault.<br>Shearing, recrystallization, folding, and formation of lineation under medium-grade conditions associated with movement along Brevard fault. | ?—?<br>?——?<br>?—?<br>?—? | ?——?<br>?—?<br>?—? |
| Middle Paleozoic | Medium-grade dynamothermal metamorphism and folding at about 350 m.y. ago.<br>Intrusion of granodiorite and pegmatite no later than 350 m.y. ago and possibly as early as 450 m.y.<br>—— Metamorphism? —— | | Formation of large-scale folds that control present trends of map units in Grandfather Mountain Formation; formation of early cleavage.} | Probably not earlier than 450 m.y. or later than 350 m.y.: Emplacement of granitic rocks, migmatite (and, possibly, Henderson Gneiss); folding and metamorphism to sillimanite grade. | | |
| Early Paleozoic | Emplacement of ultramafic rocks(?). | Cambrian(?) and Cambrian: Deposition an unknown distance southeast of present position. | | ?—?<br>Emplacement of ultramafic rocks– | | |
| Late Precambrian | Intrusion of Bakersville Gabbro and other mafic dikes. | | Deposition of sediments and volcanics on older Precambrian rocks virtually in present position. | Deposition of sediments and volcanics.<br>?—?—? | | |
| Early Precambrian | 1100–1000 m.y.: {Emplacement of Beach Granite and other intrusive granites—possibly 800–900 m.y. ago. ?—?—?<br>Metamorphism of sediments and volcanics; formation of Cranberry Gneiss and migmatitic features in gneiss southeast of window.<br>Emplacement of ultramafic rocks(?).}<br>Deposition of sediments and volcanics—younger than 1,270 m.y.(?) old. | | 1100–1000 m.y.: {Intrusion of Brown Mountain Granite.<br>Emplacement of Wilson Creek Gneiss and Blowing Rock Gneiss.<br>Folding and metamorphism of preexisting rocks.}<br>Emplacement of diorite and gabbro.<br>Deposition of mafic volcanics(?) and associated sediments. | | | |

of the section southeastward across the remainder of the exposed width of the Appalachians. The approximate thickness of the crust is from Pakiser and Steinhart (1964).

Above the geologic section is plotted the gravity profile taken from the Bouguer gravity map of the United States (Am. Geophys. Union, Spec. Comm. Geophys. and Geol. Study Continents, 1964). The broad gravity low in the Blue Ridge and Unaka belts in the line of section is part of a well-defined negative anomaly that extends along the Appalachian miogeosyncline from Vermont to central Virginia where it begins to transgress southeastward into the crystalline rocks and reaches the Piedmont in western North Carolina. Individual lows on the regional gravity low in western North Carolina and eastern Tennessee are among the lowest negative anomalies in the eastern United States. King (1964a) has pointed out coincidence between these large negative gravity anomalies and the major windows in the southern Appalachians—the Mountain City window, the Hot Springs window, and the Grandfather Mountain window.

The average density of rocks exposed in the Grandfather Mountain window is 2.73 (weighted average of 75 determinations by Zui Yuval, U.S. Geol. Survey). There is apparently no significant difference in density between the upper Precambrian sedimentary and felsic volcanic rocks and the underlying plutonic basement rocks. The average density of rocks in the Blue Ridge thrust sheet in the Grandfather Mountain area is 2.80 (weighted average of 51 determinations), significantly higher than that of rocks in the window. Evidently, the pattern of gravity anomalies in this part of the Blue Ridge belt is strongly influenced by the thickness of the Blue Ridge thrust sheet. Near Asheville, N.C., 40 miles southwest of the Grandfather Mountain window, the lowest negative anomalies are only —50 milligals. The low ridge separating gravity lows over the Grandfather Mountain and Mountain City windows may be due to large bodies of mafic volcanic rocks (density about 2.99) in the Grandfather Mountain Formation in the northwestern part of the Grandfather Mountain window.

The increase in Bouguer gravity values northwestward across the Valley and Ridge belt is probably due to decreasing depth to basement, as average densities of the unmetamorphosed miogeosynclinal rocks are presumably less than those of the underlying basement rocks.

The average density of exposed rocks in the Inner Piedmont belt in the Grandfather Mountain area is 2.74 (weighted average of 21 determinations), nearly the same as that of rocks exposed beneath the Blue Ridge thrust sheet in the Grandfather Mountain window. It appears that the regional gravity profile in the Piedmont must be influenced by crustal thickness, basement lithology, or other factors not obvious from the surface geology.

## PRINCIPAL TECTONIC EVENTS

### PRECAMBRIAN

The earliest recorded metamorphic event in the southern Appalachians was an episode of plutonic activity about 1,100 m.y. ago during which the complex of granitic, migmatitic, and metamorphic rocks now exposed in the Blue Ridge belt was formed. The basement now concealed beneath Paleozoic sedimentary rocks northwest of the Blue Ridge was presumably subjected to the same plutonic episode, for similar dates have been recorded for minerals from basement rocks penetrated by deep wells as far west as the crest of the Cincinnati arch and from basement rocks exposed in Virginia, Maryland, New Jersey, and New York (Tilton and others, 1960). Some granitic plutons such as the Beech Granite and the small intrusive body near Crossnore (Davis and others, 1962, p. 1990) may have been emplaced in the basement complex as recently as 800 to 900 m.y. ago. The lithology of the basement rocks in the northwestern part of the line of section (pl. 4) is entirely unknown, for the basement is nowhere exposed at the surface nor has it been penetrated by drilling. Basement rocks now exposed in the Grandfather Mountain window are more uniformly granitic than the rocks in the overriding Blue Ridge thrust sheet, a fact suggesting that plutonic activity generally diminished southeastward, but the original position and orientation of the boundary between plutonic and nonplutonic rocks is unknown.

In late Precambrian time, sedimentary and volcanic rocks of the Grandfather Mountain Formation and the Mount Rogers Formation were deposited in one or more deep local basins. Farther southeast and south, rocks to the Ocoee Series were laid down in a more extensive basin, which was probably not connected with the Grandfather Mountain basin in the line of section but may have connected farther south. Siltstone and crossbedded sandstone of the Snowbird Group were deposited near the northwest margin of the Ocoee basin, whereas father southeast, conglomerates, sandstones, siltstones, and shales of the Great Smoky Group were laid down. Volcanic rocks are lacking in the Ocoee Series, but rocks of the Snowbird Group somewhat resemble sedimentary rocks in the Grandfather Mountain Formation. The Great Smoky Group commonly has graded bedding and was probably deposited in deeper water than either the Snowbird Group or the Grandfather Mountain Formation. Clearly, these deposits record extensive tectonic activity, but no radiometric age or structural evidence suggests the date or type of tectonism that resulted in deposition of the upper Precambrian rocks.

Sedimentation probably occurred in the Piedmont in late Precambrian time, for metasedimentary rocks are cut by subsequently metamorphosed plutonic rocks having lead-alpha ages of 505 to 565 m.y. (Overstreet and Bell, 1965, p. 100).

### EARLY CAMBRIAN TO EARLY ORDOVICIAN

Sandstones, shales, and conglomerates of the Chilhowee Group are the basal clastic deposits of the sequence of miogeosynclinal Paleozoic rocks northwest of the Blue Ridge. The stratigraphy of these rocks is exceedingly complex in detail, but gross stratigraphic units can be recognized for hundreds of miles along strike and for tens of miles across strike, even where facies have been telescoped by thrusting. In a few places, the uppermost beds of the Chilhowee Group contain identifiable Lower Cambrian fossils, but great thicknesses of beds conformably below the lowest known fossiliferous horizons are classed as Lower Cambrian(?). The Chilhowee Group rests nonconformably on plutonic basement rocks that are 1,100 m.y. old, but no unconformity has been proven to exist between the Chilhowee Group and the upper Precambrian sedimentary and volcanic rocks. Differences in stratigraphy and environment of deposition suggest that at least local disconformities must be present.

The Chilhowee Group forms a clastic wedge that thickens southeastward from a few hundred feet in deep wells near the northwestern end of the section (pl. 4) to more than 5,000 feet in the Shady Valley thrust sheet in the line of section. As much as 7,500 feet of Chilhowee beds are preserved elsewhere in the Shady Valley thrust sheet (King and Ferguson, 1960, p. 33).

The presence of rocks of the Chilhowee Group and of the conformably overlying Shady Dolomite in the Tablerock thrust sheet in the Grandfather Mountain window shows that Chilhowee deposition must have extended southeastward, completely across the present site of the Blue Ridge belt and at least 12 miles southeast of the present position of the Tablerock thrust sheet. Bloomer and Werner (1955, p. 599) suggested that the Chilhowee Group in Virginia may pass southeastward into eugeosynclinal rocks in the Piedmont. If such a transition occurred in the line of our cross section, it took place southeast of the present position of the Brevard zone, and all evidence for it has been obscured by subsequent deformation and metamorphism and by movement along the Brevard fault.

None of the radiometric mineral ages from the Piedmont or Blue Ridge indicate tectonic activity corresponding to deposition of the Chilhowee clastic wedge, nor is there any geological evidence that such an event affected any pre-Chilhowee rocks now exposed. Hadley (1964, p. 41) inferred that the Chilhowee sediments were derived from basement rocks to the northwest that are now concelaed beneath the Appalachian Plateau and pointed out that the lack of evidence for tectonic activity during latest Precambrian and Early Cambrian time in the Piedmont and Blue Ridge supports this view.

Cambrian and Lower Ordovician deposits overlying the Chilhowee Group in the Valley and Ridge and Unaka belts are chiefly carbonate rocks interbedded with shale and locally, with sandstone. They were deposited under stable conditions, probably in a slowly subsiding trough; the clastic rocks were apparently derived from the continental interior (Rodgers, 1953a). The occurrence of Shady Dolomite in the Tablerock thrust sheet indicates that at least the lower part of the Cambrian and Ordovician carbonate sequence was deposited at least as far southeast as the present site of the Brevard zone. This occurrence and the lack of clastic rocks derived from the southeast in the Cambrian and Ordovician carbonate sequence in the Valley and Ridge belt suggest that rocks now exposed in the Blue Ridge were not subjected to deformation or metamorphism during Cambrian or Early Ordovician time.

### MIDDLE ORDOVICIAN TO EARLY SILURIAN

A thick wedge of marine clastic rocks of Middle Ordovician age rests discomformably on the Cambrian and Ordovician carbonate sequence in the Valley and Ridge belt and furnishes the first stratigraphic evidence of Paleozoic tectonism southeast of the Appalachian miogeosyncline.

Most of the wedge in the line of cross section consists of sales and sandstones that thin rapidly northwestward and intertongue with carbonate rocks. Red beds are extensive in the upper part of the Middle Ordovician clastic wedge. The southeasternmost exposures of Middle Ordovician clastic rocks in the line of section are just northwest of and below the Holston Mountain fault, where more than 5,000 feet of Middle Ordovician shale and sandstone are preserved in a synclinorium below the Shady Valley thrust sheet (King and Ferguson, 1960, p. 57). The upper part of the Middle Ordovician sequence there contains conglomerate beds and lenses as much as 30 feet thick. Pebbles in the conglomerates are poorly sorted and variously rounded; they are predominantly derived from Ordovician and Cambrian carbonate rocks but also include pebbles of quartzite probably derived from the Chilhowee Group and pebbles of vein quartz, feldspar, and volcanic rocks (Kellberg and Grant, 1956). Similar congolmerates occur elsewhere in Tennessee, Virginia, and Georgia.

The Middle Ordovician clastic rocks must have been derived from land to the southeast. King and Ferguson (1960, p. 61) concluded that the conglomerates were originally deposited near shore but were later carried into their present position by turbidity currents flowing down an oversteepened submarine slope. The conglomerates must ultimately have been derived from an area where the Chilhowee Group and older rocks were exposed to erosion.

Accumulation of clastic rocks, chiefly shales and sandstones, continued in the miogeosyncline in Late Ordovician and Early Silurian time, but these deposits are very thin in the line of cross section. These later clastic deposits are largely derived from sources to the northeast and represent the distal edge of a thick clastic wedge centered in southern Pennsylvania.

Rodgers (1953a, p. 94) referred to the tectonism indicated by the Middle Ordovician clastic deposits as the Blountian phase and to the Upper Ordovician and Lower Silurian clastic rocks as representing the main phase of the Taconian orogeny.

Current geologic time scales (Holmes, 1959; Kulp, 1961) indicate that the Ordovician and Silurian clastic rocks were deposited during the time span from about 460 to 420 m.y. ago. Zircon from bentonite interbedded with Middle Ordovician rocks in Tennessee and Alabama has $U^{238}/Pb^{206}$ ages of about 445 m.y.; biotite from the same rocks has similar ages (Kulp, 1961).

There is no definite indication that rocks now exposed in the Grandfather Mountain window were subjected to deformation or metamorphism during this interval, although some of the early folds and cleavage in the Grandfather Mountain Formation may have started to form at this time. During the Taconian orogeny, rocks in the Blue Ridge thrust sheet lay many miles southeast of their present position. A few mineral ages from basement rocks in the Blue Ridge thrust sheet suggest that they may have undergone shearing and metamorphism 420 to 450 m.y. ago (Hadley, 1964). In the eastern Great Smoky Mountains, folding of the Ocoee Series, formation of foliation in basement rocks, and extensive thrusting occurred prior to Late Devonian regional metamorphism, probably during the Ordovician (Hadley and Goldsmith, 1963, table 16). Biotite in metamorphosed rocks of the Ocoee Series at Ducktown, Tenn., has an apparent potassium-argon age of 435 m.y., suggesting that at least part of the Ocoee Series was subjected to metamorphism at about this time (Long and others, 1959).

Scattered isotopic uranium-lead age determinations on zircon indicate emplacement of granitic rocks in the Inner Piedmont at about this time. Zircon from the Toluca Quartz Monzonite at its type locality in the Piedmont of North Carolina has discordant ages of 405 to 480 m.y. (Davis and others, 1962); zircon from granitic rocks in Georgia gives discordant ages of 415 to 490 m.y. (Grunenfelder and Silver, 1958).

Farther east, in the Carolina slate belt, recent lead-alpha determinations on zircon from felsic volcanic rocks give ages of 440 to 470 m.y. (White and others, 1963), showing that at least some of the volcanic, pyroclastic, and sedimentary rocks probably accumulated during the Taconian orogeny. Sedimentation was, at least in part, tectonically controlled (Conley and Bain, 1965, p. 133–134).

Thus, during the Taconian orogeny, which resulted in deposition of a thick wedge of Middle Ordovician through Lower Silurian clastic rocks in the miogeosyncline, rocks in the Blue Ridge thrust sheet (which then lay southeast of the present position of the Grandfather Mountain window) were probably subjected to shearing and metamorphism. Still farther southeast, granitic rocks now exposed in the Inner Piedmont belt were emplaced in the deeper parts of the eugeosyncline. At the same time, volcanic and sedimentary rocks were being deposited in higher parts of the eugeosyncline; these rocks now compose the Carolina slate belt.

## LATE DEVONIAN AND EARLY MISSISSIPPIAN

The next major episode of orogenic activity is recorded in the miogeosyncline by clastic deposits of Late Devonian and Early Mississippian age. These deposits rest on a regional unconformity that bevels beds ranging in age from Early Devonian to Early Ordovician (Rodgers, 1953a). The Chattanooga Shale, which rests on the unconformity in northeastern Tennessee, is largely of Late Devonian age. The Chattanooga and equivalent shales, siltstones, and sandstones in southwestern Virginia represent the distal edge of the Catskill delta, the great clastic wedge of Middle and Late Devonian age centered in New York and Pennsylvania. This wedge marks the Acadian orogeny of the central and northern Appalachians. Lower Mississippian clastic rocks are chiefly shale, siltstone, and sandstone apparently derived from the southeast. They indicate a period of uplift in the crystalline belt of the southern Appalachians somewhat later than the Acadian orogeny farther north.

Potassium-argon and rubidium-strontium ages on biotite from a bentonite layer in the Chattanooga Shale and whole rock uranium-lead ages on the shale establish the absolute age of the Chattanooga Shale at about 350 m.y. (Faul, 1960; Kulp, 1961). Radiometric mineral ages in the range 335 to 350 m.y. seem to record metamorphic and plutonic events in the Blue Ridge and Piedmont during deposition of the Upper Devonian and Lower Mississippian clastic rocks in the miogeosyncline. A rubidium-strontium age on biotite from Wilson Creek Gneiss (Davis and others, 1962) indicates the low-grade retrogressive metamorphism and formation of cataclastic foliation in basement rocks in the Grandfather Mountain window took place about 350 m.y. ago. Geologic evidence shows that the late Precambrian rocks were folded, sheared, and metamorphosed at about the same time (table 30). Mica ages (Long and others, 1959; Kulp and Eckelmann, 1961) indicate widespread medium-grade regional metamorphism of rocks in the Blue Ridge thrust sheet during this interval. Granodiorite and pegmatite in the Spruce Pine district were emplaced prior to or during this metamorphism.

Scattered radiometric ages of minerals from rocks in the Inner Piedmont belt fall in the 335 to 350 m.y. range and suggest at least some metamorphic and plutonic activity at that time.

Hart (1964) found that potassium-organ and rubidium-strontium ratios are affected by reacting to temperatures as low as 200°C in an environment of contact metamorphism in already crystalline rocks. Hadley (1964) has pointed out that mineral ages based on these ratios may therefore record the last time the rock cooled below some critical temperature, rather than the date of crystallization or metamorphism. Thus, the 335 to 350 m.y. dates may record a period of uplift, erosion, and cooling following an episode of metamorphism, rather than the date of the climax of the metamorphism.

## LATE MISSISSIPPIAN, PENNSYLVANIAN, AND PERMIAN

### STRATIGRAPHIC AND GEOCHRONOLOGIC RECORD

Upper Mississippian and Pennsylvanian clastic rocks form a wedge which thickens and coarsens southeastward in the line of section (pl. 4). The Upper Mississippian clastic rocks are chiefly varicolored marine shale and sandstone but locally include thin beds of limestone. They rest conformably on older Mississippian limestones and shales and are overlain by continental clastic rocks of Pennsylvanian age. In northeastern Tennessee, there is no evident disconformity between Mississippian and Pennsylvanian rocks, but to the north and northeast the Pennsylvanian rocks rest disconformably on Mississippian strata (Rodgers, 1953a, p. 125).

The Mississippian and Pennsylvanian clastic rocks constitute the last preserved stratigraphic record in the miogeosyncline in the Southern Appalachians; farther north, in Pennsylvania and West Virginia, continental clastic rocks as young as early Permian are preserved.

Recently revised geologic time scales (Holmes, 1959; Kulp, 1961) indicate that the Upper Mississippian clastic rocks began to accumulate about 320 m.y. ago, and the youngest preserved Permian clastic rocks may be as young as 260 m.y.

Published radiometric age determinations do not suggest that metamorphism or plutonism affected rocks in the Blue Ridge at this time, but structural evidence (see below) indicates that much of the major thrusting took place during this interval. Many micas from metamorphic and igneous rocks in the eastern Piedmont have potassium-argon and rubidium-strontium ages in the range from 310 to 240 m.y.; Long, Kulp, and Eckelmann (1959) have suggested that these ages indicate a major episode of metamorphism and plutonic activity that reached a peak about 250 m.y. ago in the eastern Piedmont in Georgia, South Carolina, and North Carolina. Alternatively, these dates may reflect regional uplift and cooling at that time. As Hadley (1964) has pointed out, many of the youngest mica ages in the Piedmont are younger than the youngest preserved deposits in the carboniferous clastic wedge. Perhaps the preserved clastic deposits represent material removed by erosion which eventually allowed rocks now exposed in the eastern Piedmont to cool below the critical temperature for retention of argon and strontium in micas.

### STRUCTURAL EVENTS

Evidence from geochronology indicates a long and complex history of deformation and metamorphism for the rocks of the Blue Ridge and Piedmont belts during the Devonian and Mississippian, but tectonic events in these areas did not directly affect the partly synchronous miogeosynclinal rocks to the northwest until late in the Paleozoic. Prior tectonic activity in the Blue Ridge and Piedmont is recorded in the rocks of the miogeosyncline by unconformities and by wedges of clastic deposits derived from the southeast, but the rocks themselves were apparently little deformed, although local stratigraphic variations suggest that some folds formed in them during Paleozoic sedimentation (Cooper, 1964).

Thrust faults in the Cumberland Plateau and in the northwestern part of the Valley and Ridge belt involve beds as young as Early Pennsylvanian. Latest movement on these faults is thus later than Early Pennsylvanian, but how much later is not established. Thrust faults in the southeastern part of the Valley and Ridge belt and in the Unaka and Blue Ridge belts involve only older rocks. In the Grandfather Mountain area, the latest movement on the Linville Falls fault was probably pre-Late Triassic, for the fault is cut by a diabase dike of Late Triassic(?) age.

The available stratigraphic and structural evidence thus serves only to set broad upper or lower limits on the time of latest movement of a few of the major thrust faults. In no case is the time of latest movement of any single fault closely dated, and no evidence is available as to the date of initial movement or the span of time involved in thrusting. As King (1964a, p. 25) has pointed out, deformation must have proceeded for a long time, and initial movement on the thrusts may have begun well before deposition of the youngest beds known to be involved. Nevertheless, the major thrust faults shown on plate 4 were apparently all formed during an orogenic episode that reached a climax in late Paleozoic or earliest Mesozoic time.

Imbricate thrust faults that displace rocks in the Valley and Ridge belt (pl. 4) are probably all closely related in age and origin. They generally dip southeastward; some extend along strike for several hundred miles and remain nearly parallel to bedding in the overriding block but commonly cut across formations in the rocks beneath. Incompetent shales of Middle and Early Cambrian age (Conasauga Group and Rome Formation) are brought to the surface along the soles of these thrusts for many miles, but nowhere do older rocks appear at the surface. This has led to the interpretation (Rich, 1934; Miller, 1945; Rodgers, 1953b, 1964) that the thrust faults in the Valley and Ridge belt are rooted in a décollement horizon in the Paleozoic sequence and that older rocks below are not involved in the thrusting, although they may be somewhat deformed. Geophysical work just south of the line of section in Tennessee and Kentucky (Watkins, 1962, 1964) and northeast of the line of section near Blacksburg, Va. (Sears, 1964), supports the hypothesis that basement rocks are not involved in the imbricate thrusting of Paleozoic sedimentary rocks.

Beneath the Valley and Ridge belt, the décollement horizon is thought to be incompetent shales in the Conasauga Group and Rome Formation; northwest of the Powell Valley anticline, the principal décollement takes place in the Chattanooga Shale, which forms the sole of the Pine Mountain fault.

The décollement must continue southeast of the Valley and Ridge belt, but its location is uncertain. It may continue in the Rome Formation and emerge as an unrecognized folded bedding-plane thrust as shown on plate 4, or it may turn downward into basement rocks beneath the Unaka and Valley and Ridge belts. Gwinn (1964) has recognized a similar detachment thrust beneath the central Appalachians and has suggested analogous alternatives for its southeastward continuation. Unfortunately, at present, no geological or geophysical evidence tends to favor one or the other of these alternatives.

The Shady Valley and Buffalo Mountain thrust sheets (pl. 3) are the principal thrust masses in the Unaka belt. They cannot have been rooted in a décollement within the Paleozoic sequence, for they are composed principally of rocks older than the Rome Formation and contain late Precambrian rocks and older plutonic basement rocks in their lower parts. In the line of cross section (pl. 4), the Shady Valley thrust sheet is preserved in the Shady Valley syncline, a shallow syncline younger than the thrusting; southwest of the line of section, the Buffalo Mountain thrust sheet occupies the trough of the syncline. In the cross section, it is tentatively suggested that the Spurgeon and Bristol faults are continuous beneath a similar syncline northwest of the Shady Valley syncline, but no modern mapping is available there to support this interpretation.

The rocks in the Shady Valley and Buffalo Mountain thrust sheets are unmetamorphosed and were not deformed prior to thrusting (King and Ferguson, 1960, p. 85). Southeast-dipping cleavage on the southeast side of the Buffalo Mountain thrust sheet is younger than thrusting (Ordway, 1959). Basement rocks in the thrust sheets show no evidence of the pervasive cataclasis that affected rocks in the Blue Ridge thrust sheet and Grandfather Mountain window. The thrust sheet rests on unmetamorphosed rocks as young as Middle Ordovician that were folded prior to or during emplacement of the thrust sheet.

Northeast of the Great Smoky Mountains (pl. 3), the Shady Valley thrust sheet overrides the Pulaski fault, the southeasternmost of the imbricate thrusts in the Valley and Ridge belt. The Pulaski fault involves rocks as young as Early Mississippian in Virginia, but it is probably of the same general age as the thrusts farther northwest which involve Early Pennsylvanian rocks. Whether or not there are any systematic age relations among the imbricate thrusts in the Valley and Ridge belt has not yet been determined. In any case, the Shady Valley and the tectonically higher Buffalo Mountain thrust sheets arrived in their present position in post-Early Mississippian time.

King and Ferguson (1960, p. 79, 83) have shown that the Shady Valley thrust sheet originated southeast of the Mountain City window. Rocks in the Shady Valley and Buffalo Mountain thrust sheets are unmetamorphosed and were not deformed prior to thrusting; clearly, they were not subjected to folding and low-grade metamorphism that affected rocks now exposed in the Grandfather Mountain window about 350 m.y. ago (table 30). In the Shady Valley thrust sheet, rocks of the Chilhowee Group are in stratigraphic contact with plutonic basement rocks. The thrust sheet, therefore, could not have originated in the area of the Grandfather Mountain window, where a thick sequence of upper Precambrian rocks lies between the Chilhowee and the basement. Its source must have been either in the area between the Grandfather Mountain and Mountain City windows or southeast of the Grandfather Mountain window, beyond the southeastern edge of the basin of deposition of the Grandfather Mountain Formation.

Derivation of the Shady Valley thrust sheet from southeast of the Grandfather Mountain window requires more than 50 miles of northwestward transport. In addition, it requires that the rocks in the thrust sheet must have somehow escaped deformation and metamorphism during the tectonic event 350 m.y. ago. It therefore seems more likely that the source of the thrust sheet was between the unmetamorphosed rocks now exposed in the Mountain City window and the metamorphosed rocks exposed in the Grandfather Mountain window. The present distance between the two windows is less than the exposed width of the Shady Valley thrust sheet, but the distance between them may have been telescoped by a major fault now hidden beneath the Blue Ridge thrust sheet.

The principal objection to this hypothesis is that at least 20 miles of displacement is required along such a fault to make room for the rocks of the Chilhowee Group in the thrust sheets of the Unaka Belt, yet the fault must bring rocks of the Grandfather Mountain Formation over basement rock. If the fault has a rather low dip, the Grandfather Mountain Formation may override basement rock that was immediately below Chilhowee Group rocks and structurally higher than the Grandfather Mountain Formation before faulting. It is even possible that lower Paleozoic rocks underlie the fault locally.

The relations between the Linville Falls fault and any fault hidden beneath the Blue Ridge thrust sheet are unknown. If the underlying fault is later than the Linville Falls fault, then the Linville Falls and Stone Mountain faults are not correlative, and the age relations between the Linville Falls fault and faults of the Unaka belt are unknown. The Unaka belt faults might even be splits off the Linville Falls fault which were subsequently overridden by the Stone Mountain fault. In any case the Blue Ridge thrust sheet overrides the thrust sheets of the Unaka Belt along the Stone Mountain and equivalent faults.

The presence in the Grandfather Mountain window of overridden upper Precambrian and Cambrian rocks beneath the Blue Ridge thrust sheet indicates at least 35 miles of northwestward thrusting of the crystalline rocks of the Blue Ridge. If the thrust sheets in the Unaka belt came from between the Grandfather Mountain and Mountain City windows, an additional 20 miles of northwestward displacement of the Grandfather Mountain window and overlying Blue Ridge thrust sheet are required. King (1964a, p. 18) has pointed out the possibility "* * * that the whole body of Precambrian rocks in this segment of the Blue Ridge is allochthonous," an inference that is strongly supported by the amount of thrusting demonstrated in the Grandfather Mountain area.

Latest movement along the Stone Mountain fault system at the northwest margin of the Blue Ridge thrust sheet was post-Early Mississippian, later than the latest movement along faults in the Unaka belt and therefore, later than the latest movement along the Pulaski fault. Movement probably ceased prior to emplacement of the Upper Triassic(?) diabase dike that cuts the Linville Falls and Brevard faults in the Grandfather Mountain area. Unlike rocks in the thrust sheets to the northwest, rocks in the Blue Ridge thrust sheet were deformed and metamorphosed prior to thrusting, and movement of the thrust sheet was accompanied by shearing and recrystallization of rocks near its sole and by the formation of cataclastic lineation in the direction of movement.

Rocks of the Chilhowee Group and the overlying Shady Dolomite in the Tablerock thrust sheet were deformed and metamorphosed prior to or during the early stages of thrusting (table 30). Structures within the Tablerock thrust sheet closely resemble those in the Blue Ridge thrust sheet and indicate that it is a tectonic slice carried from southeast of the Grandfather Mountain window along the sole of the overriding Blue Ridge thrust sheet. Small slivers and slices of rocks of the Chilhowee Group occur along branches of the Linville Falls fault northwest of the Grandfather Mountain window (pl. 1) and along faults of the Stone Mountain family south and southeast of the Mountain City window (King and Ferguson, 1960). These bodies are analogous in tectonic position to the Tablerock thrust sheet and are probably similar in origin.

The Blue Ridge and Tablerock thrust sheets are rooted along the northwest side of the Brevard zone rather than being truncated by it as we have previously suggested (Reed and Bryant, 1964b). Northwestward movement of these thrust sheets was apparently concurrent with and possibly a direct result of strike-slip movement along the Brevard, although movement on the Brevard may have lasted somewhat longer than movement on the Linville Falls fault and have caused local truncation at the southwest corner of the Grandfather Mountain window.

According to the interpretation given above, the Blue Ridge thrust sheet is composed of rocks that originally lay southeast of the present position of the Brevard, but northwest of the position of the Brevard prior to thrusting. Rocks now exposed in the Blue Ridge thrust sheet were probably originally overlain by metamorphosed eugeosynclinal rocks of late Precambrian and Paleozoic age similar to those now exposed in the Piedmont. Upper Precambrian rocks of the Ocoee Series exposed in the higher parts of the Blue Ridge thrust sheet could, despite their lack of volcanic rocks, represent the basal parts of the Piedmont eugeosynclinal sequence that now has largely been removed by erosion.

Movement along the inferred fault now concealed beneath the Blue Ridge belt could have been directly related to the same tectonic mechanism. If the concealed fault dips southeastward below the Grandfather Mountain window, it must either connect with the Brevard at depth or descend to a level in the crust where movement along it is distributed as plastic flow. In either case, northward movement of the crustal block southeast of the Brevard could have caused northwestward thrusting of the block between the Brevard and the fault underlying the window.

The interpretation of the thrust faults in the Valley and Ridge belt according to this mechanism depends in large part on the position of the southeastern extension of the décollement zone in which they are rooted. If it turns downward into basement rocks beneath the Unaka belt, the faults in the Valley and Ridge belt may be analogous to those in the Unaka belt, and they too may be due to compression during northwestward migration of the Brevard.

If, on the other hand, the décollement remains in the lower part of the Paleozoic sequence and is present as an unrecognized folded bedding-plane thrust in the rocks below the Shady Valley thrust sheet, the décollement and the structures rooted in it may be due to gravity sliding off a tectonic high raised during early stages of thrusting. If so, the breakaway zone and the training edge of the sliding block have subsequently been folded and overridden by the thrusts in the Unaka belt.

Rocks overridden by faults of the Unaka belt are exposed in the Mountain City window. The Rome Formation crops out over broad areas, but nowhere in the window are rocks preserved that are younger than the Rome. King and Ferguson (1960, p. 83) believed that:

> The rocks that once overlay the Mountain City window probably formed a thrust in the Appalachian Valley similar to those now preserved there, but it was probably higher and has been long since eroded.

Is it possible, however, that the higher rocks are absent in the Mountain City because they slid northwestward along the décollement before the arrival of the Shady Valley thrust sheet in its present position? If so, they are to be found above the Rome Formation somewhere northwest of the Shady Valley syncline, perhaps in the sheet bounded by the Spurgeon and Bristol faults (pls. 3, 4) which may be a more northwesterly part of the Shady Valley thrust sheet.

TABLE 31.—*Location of typical exposures of rock units*

| Map unit | Quadrangle | Area on plate 1 | Location |
|---|---|---|---|
| **Mountain City window** | | | |
| Chilhowee Group | Linville | C–2 | South side of Nowhere Ridge along road between Poga School and the Elk River, and along Dark Ridge Creek. |
| Shady Dolomite | Linville and Elk Mills 7½. | C–1, C–2 | Valley of Elk River. |
| Rome Formation | Elk Mills | C–1 | Many exposures in roadcuts. |
| **Blue Ridge thrust sheet** | | | |
| Biotite-muscovite schist | Linville Falls | C–7 | Roadcuts along Blue Ridge Parkway. |
| | Linville | C–5 | Valley of Plumtree Creek. |
| | Blowing Rock and Grandin 7½ min. | J–2, K–3 | North side of Valley of Elk Creek. |
| | Blowing Rock | J–5 | Along Buffalo Cove Road. |
| Amphibolite and hornblende schist. | Linville | C–5 | Along Squirrel Creek. |
| | Blowing Rock | G–2, H–2 | South slopes of Howard Knob, Rich Mountain, and Snakeden Mountain. |
| Mixed rocks | Linville | C–4 | Along North Toe River between Minneapolis and Newland. |
| | Blowing Rock | I–2 | Along north fork of Elk Creek between altitudes of 2,600 and 2,800 feet. |
| | do | H–2 | Along road northeast of the New River south of U.S. Highway 421. |
| Beech Granite | Linville | C–2, D–2 | Upper reaches of Beech and Buckeye Creeks and along the Elk River. |
| Quartz monzonite gneiss | do | D–3 | Roadcuts along North Carolina Highway 194 along Elk River and along secondary road along east fork of Curtis Creek. |
| Aegerine-augite granite gneiss | do | C–5 | Inactive quarry along road between Crossnore and Mt. Pleasant, 0.7 mile northwest of Crossnore. |
| Bakersville Gabbro | do | C–3, C–4 | Slopes of Hump Mountain. |
| Chilhowee Group | do | C–3 | Near head of Cooper Branch. |
| | do | D–3, F–3 | Along North Carolina Highway 194 northwest of Whitehead Creek and along strike in the Elk River. |
| | Marion | A–10 | Roadcut on North Carolina Highway 80 south of Lake Tahoma dam. |
| | Linville Falls | F–7 | Roadcut along a quarry south of North Carolina Highway 181 at the southeast boundary of the Grandfather Mountain window. |
| Layered gneiss southeast of Grandfather Mountain window. | Blowing Rock | J–5 | Along Buffalo Creek Road and in valley of Licklog Branch. |
| | Lenoir | G–7 | Along North Carolina Highway 90 west of Collettsville. |
| | Linville Falls | F–8 | Along North Carolina Highway 181 near Smyrna Church. |
| | do | D–9 | Along Linville River west of Lake James. |
| Cranberry Gneiss | Blowing Rock | J–3 | Along Elk River. |
| | Linville | E–2, F–2, and D–3. | Along Watauga River and Elk River. |
| | Linville Falls | C–6 | Along Blue Ridge Parkway northeast of Humpback Mountain and along U.S. Highway 221 south of village of Linville Falls. |
| Ultramafic rocks (see table 11). | | | |
| Granodiorite and pegmatite | Linville Falls | C–6 | Saprolite exposures in clay pits in valley of Brushy Creek. |
| | Linville | C–5 | Mines and prospects in valley of Brushy Creek. |
| Quartz monzonite and pegmatite in Deep Gap area. | Blowing Rock | J–2 | Roadcuts along Blue Ridge Parkway south of Deep Gap. |

TABLE 31.—*Location of typical exposures of rock units*—Continued

| Map unit | Quadrangle | Area on plate 1 | Location |
|---|---|---|---|
| **Autochthonous rock units, Grandfather Mountain window** | | | |
| Metagabbro and metadiorite | Blowing Rock | G–4, G–5 | West side of Billy's Knob and in vicinity of Upton. Exposures poor. |
| Wilson Creek Gneiss | Linville Falls | E–6 | Roadcuts along North Carolina Highway 181 south of road-metal quarry. |
| | Linville Falls, Linville | F–5, F–6 | Roadcuts along Wilson Creek Road. |
| Wilson Creek Gneiss (layered phase). | Blowing Rock | I–4 | Along Buffalo Creek Road. |
| | do | J–5 | In roadcuts along U.S. Highway 321. |
| Wilson Creek Gneiss (phyllonite phase). | Linville, Linville Falls. | F–5, F–6 | Along North Harper Creek below uranium prospect. |
| | Linville | F–5 | Along lower reaches of Rockhouse Creek. |
| | Linville Falls | E–7, F–7 | In roadcuts along North Carolina Highway 181 in vicinity of bench mark 1661 (weathered exposures). |
| Wilson Creek Gneiss (blastomylonite phase). | Blowing Rock | I–4 | Along Yadkin River, Bailey Camp Branch, and Dennis Creek. |
| Blowing Rock Gneiss | do | H–3, H–4 | Many large exposures along U.S. Highway 321 both north and south of village of Blowing Rock. |
| Brown Mountain Granite | Lenoir | G–7 | Large continuous exposures along Wilson Creek for about 2 miles upstream from Brown Mountain Beach. |
| Grandfather Mountain Formation (arkose units). | Linville | E–5 | Along Blue Ridge Parkway east of Pineola. |
| | do | E–4, F–4 | Along U.S. Highway 221 on southeast flank of Grandfather Mountain. |
| Grandfather Mountain Formation (siltstone units). | Blowing Rock | H–3 | Valley of Flannery Fork. |
| | Linville | F–3 | Roadcuts along North Carolina Highway 105 in vicinity of Foscoe. |
| Grandfather Mountain Formation (felsic volcanic rocks). | Blowing Rock | G–2 | Roadcut along North Carolina Highway 105, 0.2 mile south of bridge across Watauga River. |
| | do | H–3 | Roadcuts along Blue Ridge Parkway near first overlook northeast of the Arts and Crafts center northwest of Blowing Rock. |
| | Linville Falls | F–6 | Cuts along old railroad grade on east side of Wilson Creek opposite bench mark 1410. |
| | do | F–7, F–8 | Saprolite exposures in roadcuts along North Carolina Highway 181 north of bench mark 1195; fresh exposures along Steels Creek a few hundred yards to the west. |
| Grandfather Mountain Formation (mafic volcanic rocks). | Blowing Rock | G–4 | Along Phillips Creek about 0.9 mile southeast of Pack Hill School. Chiefly flows. |
| | do | H–3 | Roadcuts northeast of Flattop Mountain. Tuffs and interbedded sediments. |
| | Linville Falls | F–6 | Outcrops on east side of Wilson Creek about 0.1 mile upstream from bench mark 1410. Mafic flows (not mapped separately) interbedded with felsic volcanic rocks. |
| Grandfather Mountain Formation (Montezuma Member). | Linville | D–4, D–5 | Outcrops in vicinity of village of Montezuma. Type locality. |
| | do | E–3 | Outcrops and roadcut along North Carolina Highway 194 about 0.6 mile southeast of Banner Elk. |
| Linville Metadiabase | do | E–4 | Outcrops along U.S. Highway 221 south of Pilot Knob. |
| | do | E–4 | Outcrops along North Carolina Highway 105 at Linville Gap. |
| **Tablerock thrust sheet** | | | |
| Chilhowee Group (lower quartzite unit). | Linville Falls | D–6 | Roadcuts along North Carolina Highway 183, 0.8 to 1.7 miles northeast of bridge over Linville River above Linville Falls. |
| | do | D–7 | Cliffs at Wisemans view. |
| Chilhowee Group (phyllite unit). | do | D–6 | Roadcut along North Carolina Highway 183 about 0.8 mile northeast of bridge over Linville River above Linville Falls. |

TABLE 31.—*Location of typical exposures of rock units*—Continued

| Map unit | Quadrangle | Area on plate 1 | Location |
|---|---|---|---|
| **Tablerock thrust sheet—Continued** | | | |
| Chilhowee Group (upper quartzite unit). | Linville Falls | D–6 | Roadcuts along North Carolina Highway 183, 0.4 and 0.8 mile northeast of bridge across Linville River above Linville Falls. |
| | do | D–6 | Outcrops along National Park Service trails in vicinity of Linville Falls. |
| Shady Dolomite | do | C–7 | West side of North Fork of Catawba River in vicinity of Linville Caverns and for 1 mile to north. |
| | Little Switzerland 7½. | B–9 | Road-metal quarry east of U.S. Highway 221 about 0.5 mile south of Woodlawn. |
| **Inner Piedmont belt** | | | |
| Biotite gneiss | Linville Falls | E–9 | Saprolite exposures in wave-cut cliffs on both sides of northeastern arm of Lake James. |
| | Lenoir | I–7 | Roadcuts along country road from Lenoir to Collettsville for about 1 mile norhtwest of Abingdon. |
| Mica schist | Linville Falls | D–9 | Outcrops on both sides of Catawba River arm of Lake James south of South Mountain Institute. |
| | Lenoir | I–7 | Roadcuts on east side of Husband Creek near bench mark 1080 southeast of Oakwood Church. |
| Sillimanite schist | do | I–9 | Cliffs on north shore of Rhodhiss Lake east of Huffman Bridge. |
| | do | I–9 | Roadcut on south side of Cayah Mountain, 1.4 miles north of Baton. |
| Henderson Gneiss | Linville Falls | E–9 | Outcrops along southeast side of arm of Lake James south of Lake James Church. |
| | Lenoir | J–6 | Outcrops and roadcuts along road over Indian Grove Gap, both north and southwest of gap. |
| Granitic rocks | Linville Falls | F–9 | Roadcuts south of Canoe Creek on road from Oak Hill to Willow Tree School. |
| | Lenoir | J–7 | Roadcuts along bypass road west of Lenoir in valley of Spainhour Creek. |
| | do | I–9 | Abandoned quarry on west side of McGalliard Creek south of Lakeview Church. |
| Migmatite | do | G–9 | Saprolite exposures in roadcuts along North Carolina Highway 18 bypass west of Morganton. |
| | do | J–7 | Roadcuts along U.S. Highway 321 (business route) near southern city limit of Lenoir and in abandoned quarry to west. |
| Quartz monzonite gneiss | Blowing Rock | I–5 | Roadcut along road on west side of Yadkin River at south edge of quadrangle. |
| Ultramafic rocks | Lenoir | G–7 | Asbestos prospect on southeast side of Johns River, 0.4 mile northeast of confluence with Wilson Creek. |
| **Brevard belt** | | | |
| Blastomylonite | Marion | B–10 | Cuts along Clinchfield Railroad southeast of Hankins. |
| | Linville Falls | E–9 | Borrow pit south of Canoe Creek, 0.5 mile west of Mt. View Church. |
| | Lenoir | G–7 | Roadcuts along road on northwest side of Johns River, 1.2 miles southwest of Collettsville (area H–7). |
| Phyllonitic schist and gneiss | Blowing Rock | J–5 | Outcrops on northeast side of Buffalo Creek north of bench mark 1206 and roadcuts along road on southwest side of creek. |

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