UNITED STATES DEPARTMENT OF THE INTERIOR
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PEGMATITE GEOLOGY OF THE SHELBY DISTRICT NORTH CAROLINA

by

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This report is preliminary and has not been edited or reviewed for conformity with Geological Survey standards or nomenclature.

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ABSTRACT

The Shelby district is divided into a northwestern and a southeastern province. The rocks in the southeastern province include various units in the Battleground schist formation and the Yorkville granodiorite. Those in the northwestern province include the Carolina gneiss, with its Shelby gneiss member, and the Toluca quartz monzonite. The Cherryville quartz monzonite forms a batholith that is just west of the boundary between the two provinces. Pegmatites related to both the Toluca and the Cherryville quartz monzonites lie in the Carolina gneiss and many dikes of pegmatite that are related to the Cherryville quartz monzonite are in the tin-spodumene belt that lies along the boundary between provinces. The rocks of the southeastern province have been bent into steep isoclinal folds; those of the northwestern province were bent into open folds and gently-dipping isoclinal folds. The rocks to the southeast have been metamorphosed in the epidote-amphibolite facies whereas the rocks to the northwest represent the amphibolite or granulite facies.

The pegmatites related to the Toluca quartz monzonite form sills, dikes, and concordant lenses in the Carolina gneiss, as well as dikes in the Toluca quartz monzonite. The bodies are unzoned and consist mainly of gneissic microcline-plagioclase-quartz pegmatite. The pegmatites related to the Cherryville quartz monzonite form dikes and disconformable lenses in the Carolina gneiss and the Toluca quartz monzonite. These pegmatites range widely in composition and many are zoned. The dikes west of the Cherryville batholith are rich in muscovite and plagioclase and may contain no microcline or only a moderate

amount of microcline. Quartz cores and microcline-rich intermediate zones are common. Similar pegmatite forms dikes along the west edge of the tin-spodumene belt. The tin-spodumene belt contains albite-microcline-spodumene-quartz pegmatite. These dikes of albitic pegmatite are largest and most nearly parallel to one another south of Kings Mountain. Farther north they lie in fractures of several sets.

The main factors that affected the development of zones in the pegmatite dikes are composition of magma, temperatures of magma and wall rock, and deformation during crystallization. For best development of zones and the growth of the largest muscovite books the rate of cooling and the rate of crystal growth must be rather fast. Complete consolidation probably was reached in most dikes within a few years after the start of crystallization.

PEGMATITE GEOLOGY OF THE SHELBY DISTRICT

NORTH AND SOUTH CAROLINA

INTRODUCTION

Location and physical features

The Shelby pegmatite district is in the Piedmont Plateau. Its eastern edge is marked by the concentration of pegmatite dikes in the tin-spodumene belt, for east of that belt there are few pegmatite dikes. The southern, western, and northern limits are less definite, as the abundance of pegmatite decreases more gradually in those directions. The town of Kings Mountain lies near the middle of the tin-spodumene belt and is about 40 miles west of Charlotte, North Carolina. The positions of the pegmatite areas and the major towns in and near the Shelby district are shown in figure 1.

The old Piedmont upland surface rises, in crossing the Shelby district, from an altitude of about 800 feet in the southeastern part of the Kings Mountain quadrangle to about 1,050 feet at the base of the South Mountains in the northwestern part of the Shelby quadrangle. The Pinnacle, altitude 1,705, Crowders Mountain, altitude 1,624 feet, Carpenter Knob, altitude 1,619, and Henry Knob, altitude 1,200 feet, are the most conspicuous hills that rise above the upland surface. The upland surface has been incised to a depth of 250 feet by the largest streams--the Broad River and the South Fork River.

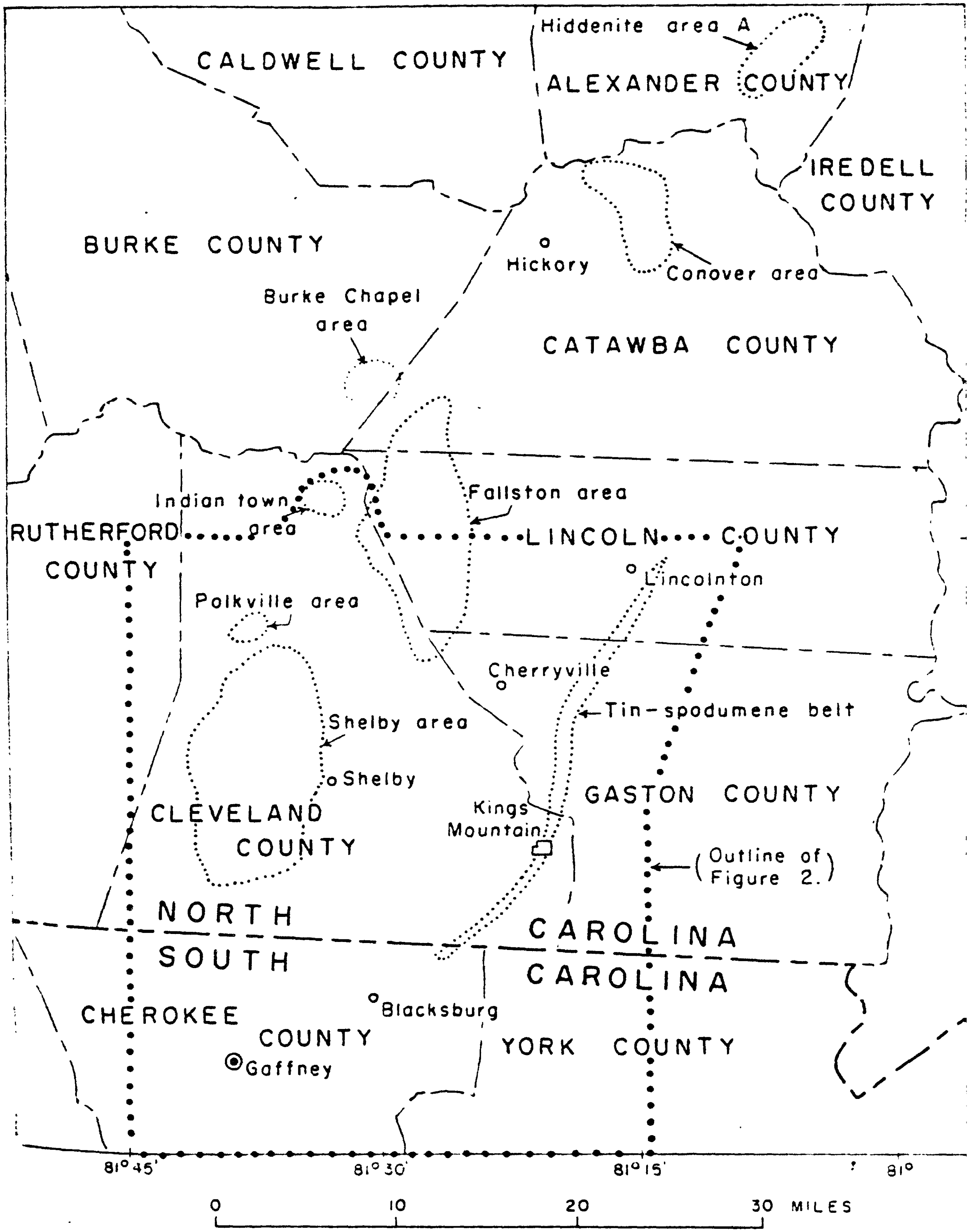


Figure 1.-Index map of the Shelby district.

Field work and acknowledgments

The first effective geologic work in the district was the brief visit of Titus Ulke in 1892 (Ulke, 1894). L. C. Graton, in 1904, spent a longer time in a more detailed reconnaissance of the tin deposits (Graton, 1906). D. B. Sterrett completed a reconnaissance of the tin deposits for the North Carolina Geological Survey (Pratt and Sterrett, 1904) then, in 1907, started to map geologically the Gaffney, Kings Mountain, and Lincolnton quadrangles (Keith and Sterrett, 1917, 1931). In 1939 and 1940 a party under the direction of T. L. Kesler mapped several parts of the tin-spodumene belt and examined deposits in other areas (Kesler, 1942), and in 1943 J. C. Olson examined nearly 100 mica mines in various parts of the Lincolnton, Shelby, and Kings Mountain quadrangles. Olson returned to the district, accompanied by me, in July 1944 and we continued to make mine examinations there until October 1945 (Griffitts and Olson, 1953).

In 1948 a party led by R. G. Yates began mapping the Shelby quadrangle. I began sampling the spodumene-bearing pegmatites near Kings Mountain in January 1949. The following summer I continued the sampling and joined the party that was mapping the Shelby quadrangle. We prepared some of the results of this work for publication (Griffitts and Overstreet, 1952, and Griffitts, 1954). I continued field work in the tin-spodumene belt until the spring of 1952. During this time D. B. Potter studied the geology of the kyanitic quartzites and associated rocks. In preparing this report I have drawn upon the information obtained by all the parties that were active in the area between 1907 and 1953. The field notes of D. B. Sterrett, the maps of Kesler, the maps and notes of Olson, and the data of Yates' party have all been of great value. These sources provided the material shown on the geologic map (fig. 2). A manuscript map by D. B. Potter was used in revising the outcrop areas of sillimanitic and kyanitic quartzite in the Kings Mountain quadrangle. I remapped the pegmatite belt, which passes through the Kings Mountain and Lincolnton quadrangles and into the Gaffney quadrangle as well as parts of the Gastonia and Hickory quadrangles, and checked the rest of the map in the field. The map of the Shelby quadrangle by Yates, Overstreet, and Griffitts was reduced and the geologic units revised to conform with those of the other quadrangles.

Inasmuch as this report is largely restricted to a discussion of the nature and origin of pegmatites related to the Toluca and Cherryville quartz monzonites, many details of structure, metamorphism, and surficial geology are omitted.

This report is based on work which is part of a program that the U. S. Geological Survey is conducting on behalf of the Division of Raw Materials of the U. S. Atomic Energy Commission.

GEOLOGY

Metamorphic rocks

Carolina gneiss

General features.--The Carolina gneiss, which includes most of the metamorphic rocks that are northwest of the tin-spodumene belt, underlies most of the Shelby quadrangle, much of the northwestern two-thirds of the Lincolnton quadrangle, and most of the northwestern two-thirds of the Gaffney quadrangle (fig. 2). One distinctive unit, a biotite gneiss member, was recognized in the Carolina gneiss during mapping of the Shelby quadrangle and was found to extend into the Gaffney quadrangle and through the Lincolnton quadrangle.

Lithology.--The Carolina gneiss is mainly a series of complexly interlayered biotite and biotite-sillimanite schists and gneisses, with small amounts of other rocks. The most common variety is a medium- to fine-grained rock that consists of biotite in flakes that rarely exceed 1/20 inch in diameter, and quartz and oligoclase or andesine in grains about the same size as and smaller than the biotite flakes. The biotite content probably ranges from 15 to 50 percent. In some layers the biotite flakes are larger and the biotite content of the rock is higher. Garnet may form spherules up to 1/10 inch in diameter in this rock. It also forms nearly massive layers up to an inch in thickness that lie in quartz-biotite schist rich in quartz. These garnet layers generally were rather strongly folded and in some places they have been sheared (fig. 3A). Tourmaline forms thin black selvages separating some of these layers from the neighboring layers.

Figure 3A. Contorted garnet layer in Carolina gneiss. Tourmaline forms a thin black coating on the sides of the garnet layer. Specimens obtained about 6 miles northwest of Shelby, North Carolina.

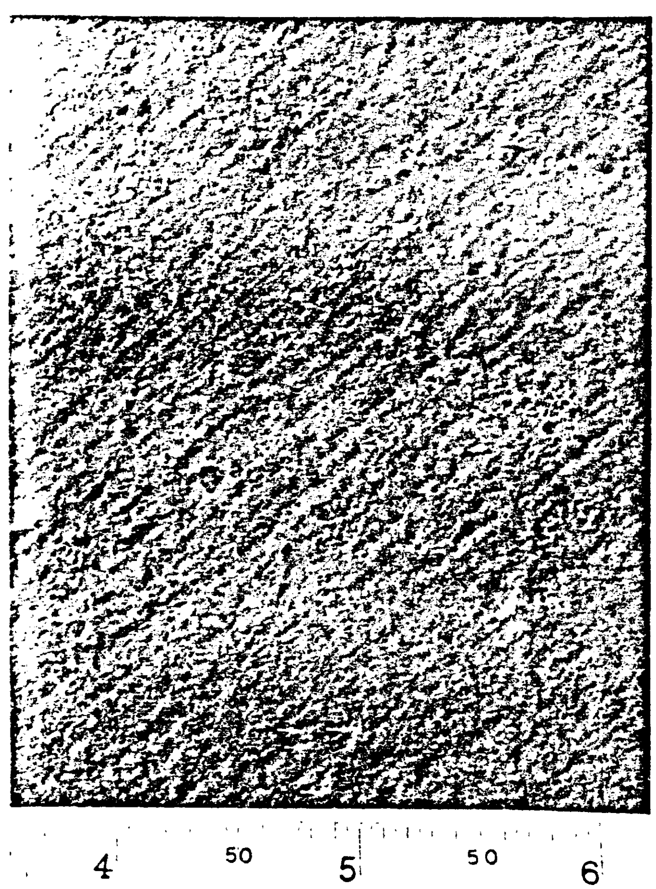


Figure 3B. Grooves on surface of a lens of fine-grained lime-silicate rock from the Carolina gneiss. Obtained about 4 miles northwest of Shelby, North Carolina.

Some of the sillimanitic rocks are very similar to the common biotitic rocks; others are much richer in quartz or garnet. The common type of schist that contains 10 to 30 percent sillimanite contains a few quartzose lenses of rock that are very rich in sillimanite. Where the sillimanite content of the rock is rather low the mineral forms needles disseminated through the rock, somewhat more abundantly in some layers than in others. In the central part of the Shelby quadrangle the schists contain layers of biotite and sillimanite that wrap around eyes of garnet or of feldspar. In that area such eyes are much less common in the biotite schist and gneiss, though eyes of feldspar are moderately common in such rocks elsewhere. The sillimanite-rich schist appears to have deformed plastically near these eyes, a behavior that may also account for a crosscutting dike-like body of sillimanitic schist in the north-central part of the Shelby quadrangle. In one place the schist apparently contained layers of greater and lesser plasticity, as blocky fragments 2 inches across or less and without uniform orientation lie in a sillimanitic matrix. The more brittle layers presumably broke while the surrounding more plastic layers flowed. Similar structures were observed on a much smaller scale in thin sections. The small scale of the geologic map of the district does not permit the sillimanitic and non-sillimanitic rocks to be shown separately.

Rocks that consist predominantly of diopside and andesine are moderately widespread in the Carolina gneiss though they constitute a small fraction of 1 percent of the formation. In most places they form lenses a few inches thick and a foot to 3 feet long. These lie parallel to the cleavage and compositional layering of the gneiss and commonly have strongly striated contacts (fig. 3B). The rock in the lenses is generally fine-grained, dense, and black. It is resistant to weathering but may have a yellow-brown friable rind. About 6 miles northwest of Shelby a series of lenses of a coarse-grained diopside-andesine-calcite rock may be traced for several hundred yards parallel to the planar structures in the enclosing rock. The maximum thickness of these lenses is about 2 feet and the maximum length is about 100 feet. Garnet forms reddish spheres about 1/2 inch across, graphite forms flakes about half that size, and calcite forms masses that may reach an inch in width. The diopside and andesine are finer-grained than these other minerals. It is inferred that the series of lenses represents a thin bed of dolomite or a series of lenses or concretions of dolomite in the original shales.

The parallelism of the lenses of diopside-andesine rock, the other compositional layers of the Carolina gneiss, and the cleavage, suggests that the compositional layers may, in other places, be parallel to original bedding. The layering is parallel to the cleavage in most places; the outstanding exception is the dike-like body of sillimanite schist that has already been mentioned. In some other places a contact between sillimanitic and non-sillimanitic schist, as mapped from float, seemed not to be exactly parallel to the cleavage as measured on outcrops. This may result from the mapping procedures more than from actual discordance in structures.

Biotite gneiss member.--The biotite gneiss is a well foliated rock that consists largely of biotite, quartz, and andesine. Microcline, garnet, and hornblende locally are prominent constituents; monazite, titanite, sillimanite, rutile, and ilmenite are common accessory minerals. Tourmaline, staurolite, epidote, and chlorite are present sporadically as accessory minerals. The proportions of the minerals and the texture of the rock vary greatly from place to place. The most distinctive variety is a dark gray, biotite-rich rock that contains discontinuous layers and lenses of light minerals surrounded by darker material. With increasing biotite content and increasing size of biotite flakes the rock becomes more schistose. A feldspar-rich variety that contains biotite, microcline, oligoclase, or andesine, and quartz in proportions similar to those of the Toluca quartz monzonite is rather common in the Shelby and Gaffney quadrangles. The dark-colored varieties of the gneiss weather to a dark brown saprolite and soil that resemble the weathering products of the darker varieties of biotite schist and some varieties of hornblende schist or gneiss. The lighter, feldspathic varieties of the gneiss weather to light-colored saprolite and soil similar to those of the Toluca quartz monzonite.

The biotite gneiss member was first recognized by W. C. Overstreet during the mapping of the Shelby quadrangle in the late 1940's. The feldspathic varieties in the northwestern part of the Lincolnton quadrangle and in the Gaffney quadrangle had earlier been mapped by Keith and Sterrett (1931 and unpublished) as "Whiteside granite" and the darker varieties in the north-central part of the Lincolnton quadrangle had been mapped by them as "Roan gneiss", their term for hornblende schist and gneiss. Hornblende is, of course, found in the gneiss in this area and according to Overstreet (oral communication, 1955), it is prominent in the gneiss as exposed west of the Shelby quadrangle.

The destruction during metamorphism of most of the original structures in the rocks has made it difficult to determine their origin. At one place, about 5 miles southeast of Cherryville, a rock was found that appears to be an amygdaloidal basalt. In most of the float blocks the white amygdaloids have been stretched but in some the deformation was slight (fig. 4). This rock now consists largely of green hornblende and andesine feldspar and the amygdules are largely andesine. The matrix is homogeneous and shows no sign of altered olivine or pyroxene. If these minerals were originally present they must have been fine-grained and well distributed through the rock, as they might well be in the vesicular upper parts of basalt flows. The wide variation in lithology of the biotite gneiss as well as the northeastward thickening of the formation, coincident with an increase in the interlayering of different rock types, suggest that volcanic and clayey sedimentary materials may have been intermingled and also deposited in separate layers. Neither material contained many structures that were large enough to persist through the metamorphism. The rest of the Carolina gneiss formation might similarly contain both volcanic and sedimentary material. The lenses of lime-silicate rock in the Carolina gneiss were originally carbonate-rich sediments, as shown by the content of calcite as well as by the calcium- and magnesium-rich silicate minerals they contain.

The original sediments that were metamorphosed to form the Carolina gneiss were rich in silicate minerals but contained very little clear quartz sand. The predominance of biotite and quartz in most of the schists suggests that silt and illitic and montmorillonitic or chloritic clays predominated, containing silica, potash, and magnesia.

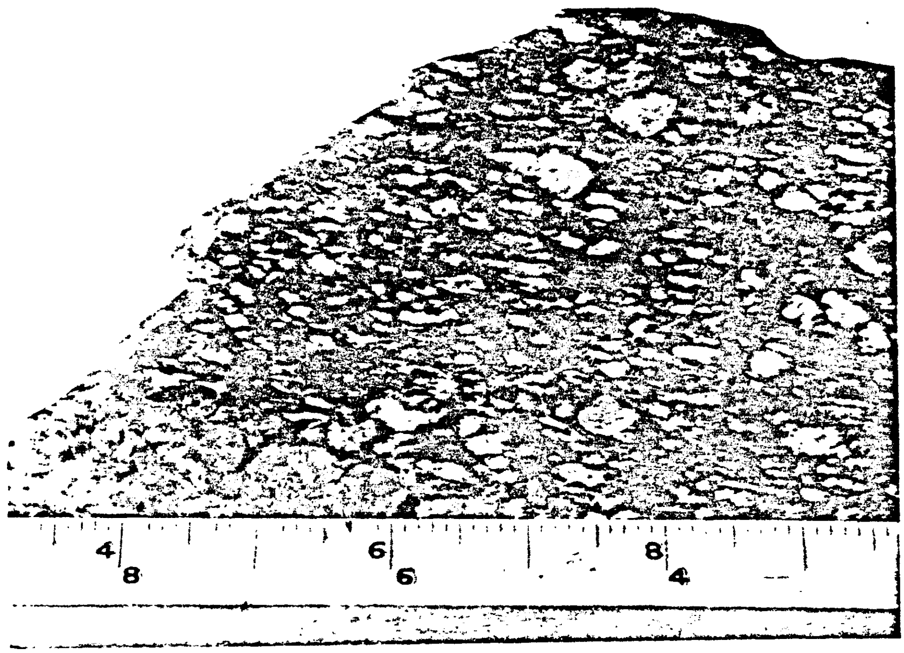


Figure 4. Metabasalt with relict amygdaloidal structure from the biotite gneiss member of the Carolina gneiss.

Battleground schist

General features.--The Battleground schist formation was described originally by Arthur Keith and D. B. Sterrett (1931, p. 4-5), who considered it to be of late Precambrian age. As parts of the formation, they included a sericite-quartz schist that grades locally into quartzite, conglomerate, metamorphosed tuff, and a manganiferous schist member. They believed that the Draytonville conglomerate, the Gaffney marble, and the kyanitic quartzites unconformably overlie the Battleground sericite schists. Hence, they included those rocks in the Cambrian Kings Mountain quartzite formation. Abundant good exposures in road cuts now show quite clearly that the conglomerate and the quartzite beds are essentially conformable with the Battleground schist. In this report, therefore, the kyanitic quartzite, Gaffney marble, and Draytonville conglomerate are treated as parts of the Battleground schist formation, and the formal names for the marble and conglomerate are not used.

Lithology.--Sericite-quartz schists constitute much of the Battleground schist formation. The color of the rock ranges from white to bluish-black, although medium gray to white or pale cream colors are predominant. The color depends largely upon the contents of graphite and iron oxides. Quartz and sericite are by far the most abundant minerals in this rock, amounting to 95 percent or more in most places. The quartz content in a few places may drop to a few percent; in others it may rise above 95 percent. The quartz-rich rock are fine-grained quartzites that contain small amounts of sericite. In the area between the north end of Crowders Mountain and the State line the Battleground schists commonly contain a small fraction of 1 percent of magnetite in well-formed octahedra a millimeter or less across.

Much of the schist contains staurolite, indicating that it was metamorphosed within the conditions of the lower amphibolite facies.

The deformation that was associated with metamorphism of the rocks has not obscured the bedding, which remains evident in the schist in an alternation of quartz-rich and quartz-poor layers 0.1 inch or less in thickness. The bedding naturally is not distinct in the schists with very low quartz content. The distinction between bedding and cleavage can be made in the Battleground schist as the primary and secondary planar structures are not everywhere parallel, although they are parallel in most places. The sericite flakes are oriented parallel to the schistosity and the quartz grains are nearly equidimensional or are slightly flattened parallel to the schistosity. Ottrelite, staurolite, and andalusite grains are not consistently parallel to the bedding or cleavage. Tiny wrinkles, chevron folds an inch or so across, and fractures in the Battleground schists are discussed under structure.

Bedding in the marble is shown by thin layers or seams of biotite, by jointing parallel to the seams, and by interlayering of beds with different colors (fig. 5).

Chlorite schists are not conspicuous in the Battleground schist. They are similar to the darker varieties of sericitic schists but contain very small flakes of chlorite as well as sericite. A few chloritic schists may contain less quartz than the average sericitic schist.

Graton (1906, p. 16) and Keith and Sterrett (1931, p. 4) inferred from the presence in some beds of broken fragments of rocks and minerals that the Battleground schist is tuffaceous, though nothing as conclusive

Figure 5. Bedding in marble in Battleground schist (Gaffney marble). Campbell quarry, Gaffney, South Carolina.

as shard structure or scoriaceous material was found. It is impossible to determine how much of the sedimentary material from which the schist was derived was of volcanic origin. The Battleground schist may be part of the series of slates that crop out in the "Slate belt" east of Charlotte and contains a large proportion of volcanic material.

The sedimentary rocks that were metamorphosed to form the Battleground schist, like the protoliths of the Carolina gneiss, included volcanic materials and shales. In addition, they included sandstones, conglomerates (with pebbles either of quartz or of basalt or rhyolite porphyry), manganiferous and iron-rich shales, and limestone.

Hornblende schist and gneiss

General features.--The rocks indicated on the map as hornblende schist and gneiss include the gabbros and the "Roan gneiss" that were mapped by Keith and Sterrett (1931) as well as small bodies of more basic rocks composed mainly of pyroxene, serpentinite, and talc. There are three general types of rock in the mapped unit: gabbroic rocks that are more or less gneissic, hornblende schist, and pyroxenite.

Gabbroic rocks.--The gabbroic rocks form round or oval masses a few hundred yards across. The marginal parts of some of these bodies are somewhat gneissic, and some bodies are gneissic throughout. In the western part of the district Keith and Sterrett classed the foliated rocks as "Roan gneiss" and the unfoliated rocks as "gabbro", a distinction that no longer seems appropriate. In mineral composition, the rocks here called gabbro range from dioritic to gabbroic, as the feldspar may be either andesine or labradorite.

In the southern and northeastern parts of the Shelby quadrangle the gabbroic rocks are enclosed in biotite schists and gneisses of the Carolina gneiss. They form elliptical bodies that are oriented with their long axes about parallel to the foliation of the gneiss. In the rest of the district the gabbros are in or near Battleground schist, Cherryville quartz monzonite, and Yorkville granite. They are the wall rocks of spodumene-bearing pegmatite in the Indian Creek and Beaverdam Creek areas.

The biotite gneiss member of the Carolina gneiss, as mapped in the Lincolnton quadrangle, includes many masses of hornblendic rocks. Some of these appear to be gabbros and quite likely are intrusive; others appear to be metamorphosed sedimentary or volcanic rocks.

The gabbros are dark gray and are composed mainly of hornblende and feldspar which, because of their strong color contrast, give the rock a spotted appearance. The grain size ranges from 0.02 inch to 0.5 inch but it is about 0.1 inch in most places. Microscopically the rock is seen to contain hornblende, andesine or labradorite, with minor amounts of hypersthene, biotite, olivine, and magnetite or ilmenite. The hornblende is green in most places, brown in a few. There is a suggestion that olivine tends to be associated with brown hornblende more than with green hornblende. The hypersthene is generally nearly colorless in thin section. It is disseminated through the rock and at one place forms pyroxenite veins or dikes in gabbro. At the Old Plantation emerald mine both hypersthene and hornblende were formed by the alteration of olivine as well as by primary crystallization. Evidence for a secondary origin of hornblende and hypersthene is lacking elsewhere. Sterrett reports in his unpublished notes that olivine gabbro, that crops out 2-1/2 miles N. 70° E. of Lincolnton, contains violet augite, olivine, calcic labradorite, biotite, black iron ores, pleonaste, and pyrrhotite. Serpentine formed radiating reaction rims around other minerals.

Hornblende schists.--Hornblende schists are very rare in the Shelby quadrangle but are found in abundance in and near the tin-spodumene belt. They form the country rocks of many pegmatite dikes south of Kings Mountain and near Lincolnton. The schist forms more or less tabular layers that are elongated parallel to the regional and local trend of the foliation. Interlayered with it are various rocks of the Carolina gneiss and Battleground schist.

In many of the schists hornblende is by far the dominant mineral. It is pale green to black in hand specimens and colorless to green in thin-sections. The hornblende-rich rocks very commonly are strongly lineated, with nearly all the hornblende needles parallel with one another. Thin pencils of light-colored minerals in these lineated rocks are also parallel to the hornblende crystals. Plagioclase feldspar is a rather common accessory mineral and in generally oligoclase or andesine. Sphene is abundant in some rocks, including some that also contain abundant light brown biotite. Glaucophane constitutes most of several thin layers, not more than 1/8 inch thick, in hornblende schists in the Kings Mountain area and is a major constituent of uncommon quartz-glaucophane schist layers several inches thick in the same area.

Pyroxenite and associated rocks.--Small bodies of medium-grained green rock, composed largely of ferro-magnesian minerals, are moderately common in hornblende gneiss near Lincolton and near Gaffney, as well as in a small area a few miles south of Shelby. These rocks were mapped as peridotite by Keith and Sterrett (1931). Some of the bodies are nearly tabular, as the length is several times the width; the outcrop areas most carefully examined, however, are ellipses.

South of Shelby small oval bodies of pyroxenite are in biotite schist but are near bodies of gabbro and hornblende gneiss of similar size and shape. Elsewhere the pyroxenite most commonly is enclosed by the hornblende schist and gneiss.

The pyroxenite is green, the shade varying with the content of minerals other than hypersthene. Increasing content of chlorite, generally in flakes 1/50 inch or more across, has little effect on the color; increasing talc content naturally lightens the rock. The rocks richest in talc are greenish gray. Increasing hornblende content darkens the rock. The hornblende content may vary directly with the feldspar content.

The hornblende of the pyroxenites is green, in some places with a bluish cast, in other places grayish or brownish. Brown hornblende is apparently rare. The pyroxenes include hypersthene, which is dominant, and diopside, which is uncommon.

Schist and granite complex

About 100 years ago Oscar Lieber, the State Geologist of South Carolina, made a geologic reconnaissance of an area that includes the Kings Mountain quadrangle. He found a broad area in which the rocks upon weathering, yielded a very light-colored soil, similar in color to that which forms from granite in many places in the Piedmont

(Lieber, 1860). He concluded that the area was underlain by granitic rocks. In this view he was followed by Keith and Sterrett (1931) who found confirming evidence in outcrops of granitic rocks in the southern part of the Kings Mountain quadrangle, and named the unit "Bessemer granite". This unit contains not only granitic rocks of quartz monzonitic or quartz dioritic composition, but also, and probably in larger amounts, sericite schists that are in part metamorphosed clastic volcanic rocks.

Keith and Sterrett used the presence of conspicuous pyroclastic material as their main criterion for distinguishing Battleground schist from "Bessemer granite." Sericite schists in the Kings Mountain quadrangle containing volcanic material were included in the Battleground schist. Most of the other schists to the east and south were included in the "Bessemer granite". This distinction was not consistently made in the Gaffney and Lincolnton quadrangles. I separated several areas of schist from the "Bessemer granite" unit of Keith and Sterrett but was left with an unstudied residue that may be underlain largely by igneous rock. No more was done because further study of this material would contribute little to the subject of this report.

Igneous rocks

Toluca quartz monzonite

General features.--Toluca quartz monzonite occurs in a northeasterly-trending belt that is about 35 or 40 miles wide at the Shelby district. It is found throughout the Shelby quadrangle but is restricted to the western quarter of the Lincolnton quadrangle and the northwestern third of the Gaffney quadrangle. It may be present at the northwestern corner of the Kings Mountain quadrangle though this has not been demonstrated.

Most of the bodies of Toluca quartz monzonite are sills that are parallel to the foliation of the Carolina gneiss although the contacts of the sills locally cross the foliation of the wall rocks. A few bodies of quartz monzonite in the northwestern part of the Shelby quadrangle are dikes that cut the Carolina gneiss. The largest outcrop is in the northeastern corner of the Shelby quadrangle and parts of the adjoining quadrangles. It is about 7 miles long. Half of the total number of bodies of Toluca quartz monzonite mapped in the Shelby quadrangle are 2,600 feet or more in length.

Most of the bodies are tabular, with some pinching and swelling, but others are irregular. The irregularities in the outcrop area in the central and south-central parts of the Shelby quadrangle represent irregularities in the shapes of the bodies of quartz monzonite whereas those in the very large outcrop in the northeastern corner of the quadrangle are in part results of the erosion of a gently folded and nearly horizontal sill.

Lithology.--The Toluca quartz monzonite is a gray, medium-grained, moderately to strongly gneissic gray rock (fig. 6A). The gneissic structure is shown by layers that are rich in either quartz or biotite. Where the segregation of minerals is less pronounced the foliation is shown parallelism of biotite flakes and of leaves of quartz. Linear structures in the quartz monzonite include elongate flat leaves of quartz, and elongate clusters of biotite flakes. The quartz leaves

Figure 6A. Toluca quartz monzonite with strongly developed planar and linear structure. Scale is 6 inches long. Near U. S. Highway 74, five miles west of Shelby.

Figure 6B. Lineation, parallel to 6-inch scale, in Toluca quartz monzonite. Same locality as figure 6A.

generally are less than an inch long, though a few are several inches long (fig. 6B). The width is about one-fifth of the length and the thickness is a small fraction of the width. In places the foliation is folded and cut by faults along which there was only small displacement. The fractures seem to have little directional relation to the foliation and lineation of the quartz monzonite, though they formed rather soon after the consolidation of the quartz monzonite, because pegmatite related to the quartz monzonite was intruded into them.

The major minerals of the Toluca quartz monzonite are oligoclase, microcline, quartz, and biotite. Muscovite is a minor constituent that is intergrown with feldspar and quartz. As an accessory mineral garnet forms spherules about 1/40 inch across embedded in the light-colored aggregates of quartz and feldspar. The feldspar grains are 1/20 to 1/4 inch across. Apatite, zircon, monazite, and ilmenite are widespread accessory minerals. Rutile, sillimanite, blue spinel, and xenotime are less widespread. The proportions of the major minerals in six specimens are shown in table 1.

The order of crystallization of the minerals has not been worked out in detail. The biotite, the feldspars, and part of the quartz apparently crystallized early, followed by the quartz that formed the leaves, then by garnet, which formed after cessation of the movement that oriented the quartz leaves. The plagioclase, like the quartz, may have crystallized over a long interval because sodic plagioclase seems to have formed after most of the more calcic plagioclase.

Table 1--<u>Mineral composition of Toluca quartz monzonite</u>

	1	2	3	4	5	6	Average
Plagioclase	35	40	31	41	35	25	32.8
Microcline	23	28	34	16	20	38	28.3
Quartz	26	31	23	26	38	31	29.1
Biotite	8	1	4	13	5	2	5.5
Muscovite	8	--	--	4		--	2.0
Garnet	--	--	8	--	2	4	2.3

Columns one through six give compositions of six specimens obtained 4 to 6 miles northwest of Shelby, North Carolina.

Modal analyses made on their sections by point-counter method. Values given to nearest percent, except in averages.

Yorkville granite

General features.--The Yorkville granite forms an elongate batholith that has been traced for more than 50 miles from the town of York (formerly Yorkville), South Carolina, northward to Catawba County, North Carolina. The Yorkville granite is widely exposed south of the Gaffney and Kings Mountain quadrangles and between Gaffney and an area a few miles southeast of Whitestone, South Carolina. In most places it is in contact with Battleground schist, though it might cross the strike of the rocks far enough southwest of Gaffney to reach the Carolina gneiss.

The Yorkville granite has not yet been found in contact with the Cherryville quartz monzonite or with any pegmatites related to the Cherryville quartz monzonite. It does not appear to be involved in any way in the origin either of the pegmatites or of the metamorphic rocks that enclose the pegmatites, so it will not be discussed in detail.

The western margin of the main batholith of Yorkville granite is nearly parallel to the regional trend of the metamorphic rocks. The contact is somewhat transgressive, however, as is indicated by septa of metamorphic rocks in the batholith. South of the Gaffney and Kings Mountain quadrangles the rock mass is markedly discordant.

The metamorphic rocks have been affected by the Yorkville granite; the intensity of metamorphism increases near the contact to produce amphibolite facies, whereas the rocks farther west are of the greenschist or albite-epidote-amphibolite facies (Potter, report in preparation 1955). This increase in intensity of metamorphism is shown most conspicuously by the abundance of sillimanite in the kyanitic quartzites or by the presence of sillimanite without kyanite within a mile of the Yorkville granite.

Lithology.--The most common variety of Yorkville granite is a coarse- to medium-grained porphyritic rock. The groundmass is a medium-grained aggregate of quartz, biotite, oligoclase, and potash feldspar. The phenocrysts, of microcline, reach lengths of 2 inches. In most places they are not conspicuously aligned. Magnetite and zircon are widespread accessory minerals. The plagioclase is strongly zoned. This is especially pronounced in some grains that are adjacent to microcline phenocrysts.

The composition of the Yorkville granite is near the dividing line between quartz monzonite and granodiorite: that is, just about one-third of the feldspar is potash feldspar and about two-thirds is plagioclase. The chemical analyses in table 2 reflect the nature of the rock.

Only a few dikes of aplite and of pegmatite have been found in the Yorkville granite. They are composed of plagioclase and microcline feldspar, quartz, and micas. The dikes are generally only a few inches thick.

Cherryville quartz monzonite

General features.--The Cherryville quartz monzonite forms a large, elongate batholith that extends from a point near Blacksburg, South Carolina, northeastward at least as far as Maiden, North Carolina. In addition, it forms a series of stocks in a line that passes through the town of Kings Mountain and on to the northeast (fig. 2), about 2-1/2

Table 2.--Chemical composition of the Yorkville granite

	1	2	3
SiO_2	68.90	70.77	71.55
Al_2O_3	15.75	14.89	14.47
Fe_2O_3	1.16	.75	.46
FeO	1.49	1.24	1.51
MgO	.74	.43	.77
CaO	2.66	2.08	2.00
Na_2O	4.76	4.47	3.72
K_2O	3.49	4.70	4.16
Ign. loss	.18	.19	.34
TiO_2	.36	.36	.40
MnO	tr	tr	.06
F_2O_5	tr	tr	.13
SO_3	tr	tr	--

1. Earle Sloan, analyst. From Jackson quarry, 0.5 mile north of Clover, York County, South Carolina. From U.S.G.S. Bull. 426, p. 174.

2. Earle Sloan, analyst. Muscovite-bearing Yorkville granodiorite, Whiteside quarry, 2 miles west of Filbert, York County, South Carolina. From U.S.G.S. Bull. 426, p. 174.

3. L. N. Tarrant, analyst. From small quarry 0.9 mile southwest of Machpelah Church, Lincoln County, North Carolina. Collected by D. B. Potter, report in preparation.

to 5 miles east of the batholith. About 7 miles southeast of Kings Mountain this eastern belt of stocks coalesces with the main batholith.

The main batholith is made up of three segments. The northern segment, which is at least 25 miles long, trends N. 25° E. parallel to the belt of stocks. The width of this segment ranges from 1/2 to 2-1/2 miles. The thinner parts are relatively clean quartz monzonite, whereas the thicker parts contain inclusions or septa of metamorphic rocks. The southern segment of the batholith is a wedge-shaped mass about 15 miles long that tapers from a width of about 4 miles west of Kings Mountain to its pointed end about 5 miles west of Blacksburg. This wedge-shaped mass contains many large inclusions of biotitic schists and gneisses as well as large septa or pendants that are attached to the wall of the batholith. These are somewhat more abundant along the western wall than along the eastern wall. Connecting the northern and southern segments of the batholith is the middle segment, a swarm of dikes, sills, and irregular small stocks. This swarm, which is about [illegible] miles wide, extends due south from Cherryville to the latitude of Kings Mountain, where it joins the broad base of the wedge-shaped southern segment.

The stocks in the eastern belt of Cherryville quartz monzonite range in width from about 1/5 mile to 1-1/2 miles and in length from about 2-1/2 to 4 miles. They are separated by 4 to 7 miles of metamorphic rocks. The long mass of quartz-monzonite between Blacksburg and Gaffney might be considered a part of the eastern belt.

It is uncertain how far the Cherryville quartz monzonite extends north of Maiden. None of the rock was found while walking over roads a few miles north of that town, but the possibility remains that a narrow belt of quartz monzonite might have been concealed where crossed by the roads or that the quartz monzonite belts might be interrupted, only to recur still farther to the northeast. Near Pacolet, southwest of Gaffney, which is near the apparent southwestward end of the Cherryville quartz monzonite area, a rock is quarried that strongly resembles the Cherryville quartz monzonite. Detailed studies of the rock at Pacolet and of the geology of the area between Pacolet and Gaffney are needed before a correlation can be established. A lenticular body of granitic rock southeast of Gaffney might also be a body of Cherryville quartz monzonite. It has not been studied.

Relations to enclosing rocks.--Both the northern and the southern segments of the main batholith of Cherryville quartz monzonite are essentially parallel to the regional trend of the metamorphic rocks. The middle segment, with the intimate intermingling of igneous and metamorphic rocks, is discordant. The eastern belt of Cherryville quartz monzonite stocks also trends parallel to the regional structures and the individual stocks are elongated parallel to that trend. The discordant middle segment of the batholith, interestingly, is nearly coincident with and parallel to the axis of the Muddy Fork anticline (fig. 11). The metamorphic rocks therefore are very strongly deformed, having undergone the earlier folding as well as the later fracturing and displacement that accompanied the intrusion of the quartz monzonite magma. The complex structure in the metamorphic rocks and the intermingling of igneous and metamorphic rocks made it impossible in the time available to determine the direction of dip or of plunge in the batholith.

South of the main batholith and along-strike from its end the metamorphic rocks are cut by many closely spaced fractures. The fractures bound blocks of schist that are a few feet to several tens of feet across. Many of the blocks have been rotated. The metamorphic rocks in the eastern belt of quartz monzonite are also somewhat fractured, though less intensely than those along-strike from the main batholith and in the tin-spodumene pegmatite belt.

The association of bodies of quartz monzonite with fractured country rocks and the concentration of the quartz monzonite into two belts in the northern part of the area of figure 2 suggested that shattering along two belts, perhaps resulting from region shearing movements, preceded the intrusion. The northwestern fracture zone was disrupted where it intersected the Muddy Fork anticline (fig. 11) and gave way to a transverse breccia zone that connected it with the southern fracture zone. The especially large amount of quartz monzonite that is in and near the transverse breccia zone might be attributed to a southwestward movement of the northwestern side of the northwestern fracture zone. If the movement continued during the rise of the magma a space of relatively low pressure should have developed there. Naturally, if the batholith does not plunge vertically the direction of movement of the fracture walls that would be most favorable for intrusion would not be horizontal; it would deviate from the horizontal by the same amount the plunge deviated from the vertical.

Keith and Sterrett (1917, p. 130-132) point out that micaceous schists were partly absorbed by the magma to yield igneous rocks with an abnormally high mica content and that hornblendic rocks broke up into more block-like masses which, where assimilated, gave rise to more basic granitic rocks. The recent work verified the blocky nature of inclusions both micaceous and hornblendic, but the amount of assimilation and its effect on the composition of the intrusive rock does not seem to be particularly striking.

In the Muddy Creek exposure the quartz monzonite shows very sharp contacts against the sillimanitic gneiss (fig. 7A). The crumples or shear folds in the gneiss plunge S. 10° W. at an angle of about 65 degrees. The quartz monzonite contains many inclusions. Some of these are of gneiss, in which the foliation is not parallel to that in the country rock (fig. 7B) and some are of somewhat gneissic granitic rock (fig. 10B) and may be cognate inclusions. Still later the normal quartz monzonite was fractured and cemented with lighter colored muscovitic quartz monzonite (fig. 8). In this exposure the quartz monzonite varies in composition but the variation could not be correlated with variations in country rock or with evidence of assimilation.

In most places there was little reaction between the Cherryville quartz monzonite magma and the intruded rocks. At Muddy Creek, Cleaveland County, North Carolina, the schist contains unusually large amounts of muscovite and unusually coarse-grained micas and sillimanite has apparently replaced some of the micas. The muscovite appears to be a young mineral because it is concentrated along axes of shear folds that break the biotite flakes, and it forms metacrysts that lie along axialplane fractures in the folds and the sillimanite that replaces the muscovite must be still younger. Farther east, near the eastern edge of the batholith, the schists in many exposures contain much more muscovite than the Carolina gneiss in the Shelby quadrangle. Inasmuch as the boundary between the muscovite-rich Battleground schist and the

Figure 7A. Discordant contact of Cherryville quartz monzonite (upper left) with mica schist (lower right). Scale is 6 inches long. Muddy Creek, Lincolnton quadrangle.

Figure 7B. Foliation in schist truncated by Cherryville quartz monzonite and small inclusions, rotated, in quartz monzonite (to right of scale). Muddy Creek, Lincolnton quadrangle.

Figure 8. Biotitic Cherryville quartz monzonite broken and cemented by later, muscovitic quartz monzonite. Scale is 6 inches long. Muddy Creek, Lincolnton quadrangle.

biotite-rich Carolina gneiss has not been mapped accurately the amount of folding or faulting of one rock into the other is unknown.

<u>Lithology</u>.--Three varieties of Cherryville quartz monzonite have been recognized in the field. The most common variety, which forms 70 to 80 percent of the batholith, is a biotite-muscovite quartz monzonite that contains feldspar grains about 1/8 inch across, with small mica flakes and interstitial quartz. About 2 miles south of Lincolnton and in a small area north of Vantine the rock is slightly porphyritic, with phenocrysts of microcline that are slightly larger than most of the feldspar grains of the rock but the contrast in grain size is not great. The second variety of rock is found in the southern part of the batholith, and in scattered areas near its western margin this variety is moderately to strongly lineated (fig. 9). The irregular body of Cherryville quartz monzonite that extends from Blacksburg to Gaffney is the lineated variety. The assignment of this body to the Cherryville quartz monzonite is uncertain as similar rocks have been noted near the borders of the main batholith of Yorkville granite. The third variety of Cherryville quartz monzonite contains muscovite but little or no biotite. It is found in small areas scattered through the batholith and appears to be most common near its eastern margin.

The main constituents of the Cherryville quartz monzonite, other than the micas, are microcline, oligoclase, and quartz. The relative proportions of the minerals are shown in table 3.

In the common biotite-muscovite quartz monzonite the microcline almost invariably shows gridiron twinning. It generally contains only a few thin spindles of albite and it only rarely contains irregular

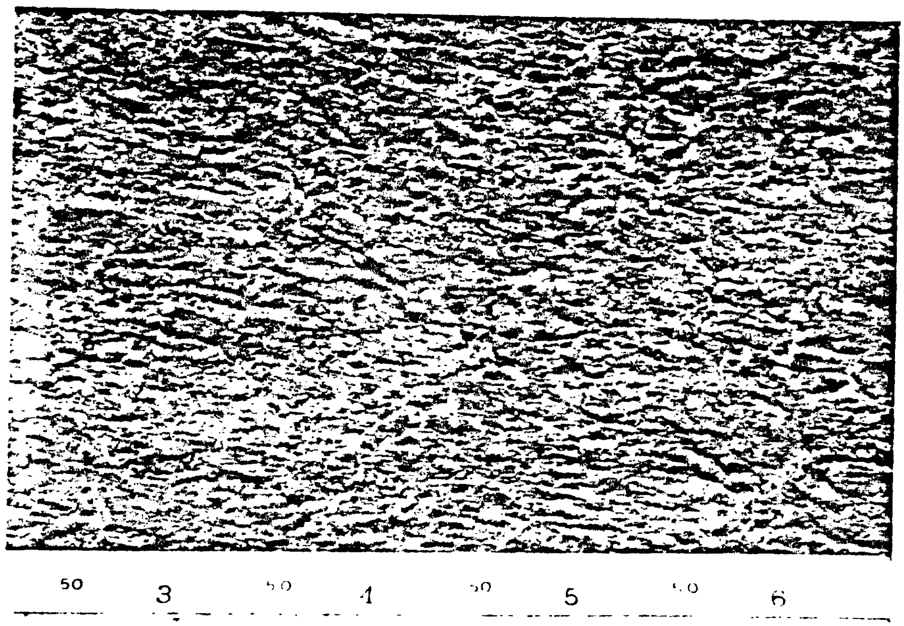


Figure 9. Lineated variety of Cherryville quartz monzonite from Blacksburg, South Carolina. Major scale divisions are 1 inch.

Table 3.--<u>Mineral composition of Cherryville quartz monzonite in percent by volume</u>

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Plagioclase	24	33	30	32	31	24	26	45	26	33	36	32	31.6	32.9
Microcline	1	15	25	18	16	31	38	14	28	33	26	23	23.3	26.8
Quartz	46	32	28	35	42	30	28	26	24	31	30	31	30.6	31.7
Biotite	--	1	6	6	4	4	--	6	3	1	7	4	3.8	4.3
Muscovite	29	19	11	7	7	11	7	9	19	2	1	10	9.4	8.6
Garnet	--	--	--	--	--	--	1	--	--	--	--	--	--	--
Tourmaline	--	($\frac{1}{4}$)	--	--	--	--	--	--	--	--	--	--	--	--
Epidote	--	--	--	2	--	--	--	--	--	--	--	--	--	--

Modal analyses made on thin sections by point-counter method. Values given to nearest percent except for tourmaline and in averages.

1. Sample taken at contact with mica schist. Muddy Creek, Lincolnton Quad.
2. One and one-fourth miles north of Bethel Church, Lincolnton Quad.
3. One and one-fourth miles north of Vantine, Lincolnton Quad.
4. Two and one-half miles north of Vantine, Lincolnton Quad.
5. One mile south of Waco, Lincolnton Quad.
6. Common variety of quartz monzonite, Muddy Creek, Lincolnton Quad.
7. One mile west of Laboratory, Lincolnton Quad. Crushed rock from shear zone.
8. One mile west of Laboratory, Lincolnton Quad.
9. Lineated quartz monzonite, Blacksburg, South Carolina. Weathered.
10. Two and one-half miles north of Oak Grove, North Carolina.
11. One and one-half miles south of Waco, North Carolina.
12. One and one-half miles south of Waco, North Carolina.
13. Average of 2 through 12.
14. Average of biotite-muscovite quartz monzonite, all samples except 1, 7, and 9.

patches of plagioclase in "patch perthite". The larger microcline grains contain round or oval blebs of plagioclase, some of which may have been incorporated in the growing microcline crystal but others have replaced the microcline. A few grains of microcline were largely replaced by plagioclase, leaving only irregular remnants of the rim of the original crystal. The plagioclase occurs in two manners: as blocky grains of oligoclase (ab 70-80) that commonly are twinned and somewhat turbid, and as round grains that generally are less twinned than the blocky grains and are much less turbid. The latter plagioclase contains about 4 to 6 percent more albite than the blocky feldspar. The difference in composition can be determined readily where the clearer plagioclase forms an overgrowth on the blocky turbid plagioclase. Both varieties of plagioclase weather readily; hence, in thin section they contain thin layers of colorless or brown secondary material. The plagioclase is only slightly zoned where the blocky grains have not been overgrown by the later plagioclase. Oscillatory zoning was not seen. The rock is free from fractured grains or other cataclastic effects, though in some samples it contains cracks that cut both microcline and plagioclase feldspar. In one place such cracks are partly occupied by thin veins of quartz. In other places the plagioclase is clear alongside the cracks, though elsewhere it contains many dust-like particles of a brown material. The clear plagioclase apparently has the same albite content as the turbid. The fractures that pass into microcline grains are there filled with tiny particles of brown material resembling the material which was removed from the plagioclase.

Biotite in the quartz monzonite contains black halos about tiny inclusions and is commonly altered to chlorite. A little sphene formed as a by-product of this reaction, but no potash feldspar could be attributed to it. Perhaps the potassium released during the chloritization entered sericite, which accompanies the chloritized biotite in many places. Quartz forms round or irregular grains interstitial to those of feldspar and mica.

During the consolidation of the rock the microcline and blocky plagioclase apparently formed first, probably about at the same time as the micas; then additional plagioclase formed overgrowths on the blocky grains, interstitial masses between the older grains, and replacement masses in the microcline. In most places the replacement formed round blebs; in fewer places it formed patch perthite. Strong zoning of plagioclase alongside some large microcline grains suggests that albite might have been exsolved from the microcline.

Large crystals of perthitic microcline have been found in the biotitic Cherryville quartz monzonite (figs. 10A and 10B). The origin of these is not clear. They have been found mainly in outcrops in which small pegmatite dikes are common and there may be a genetic relationship between the pegmatites and the crystals, which presumably replaced the quartz monzonite. The microcline grains are not accompanied by other coarse-grained minerals, such as the plagioclase, quartz, and muscovite that accompany it in pegmatites, but are embedded in quartz monzonite.

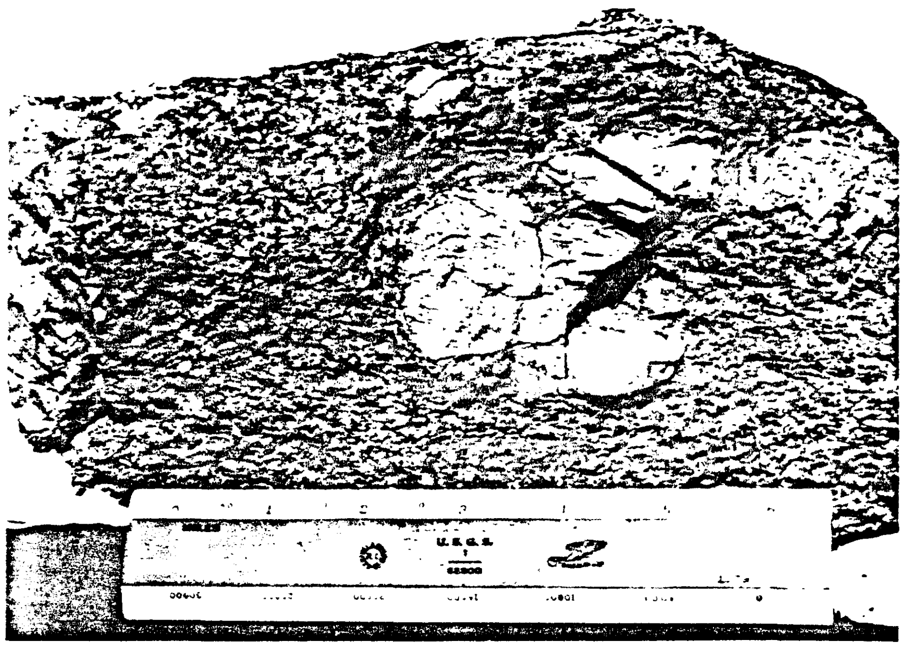


Figure 10A. Perthitic microcline mass in binary Cherryville quartz monzonite. From old U. S. Highway 74 near Buffalo Creek.

Figure 10B. Euhedral white perthitic crystal (below 6-inch scale) in biotitic Cherryville quartz monzonite. Muddy Creek, Lincolnton quadrangle.

The muscovite quartz monzonite is similar to the biotite-muscovite quartz monzonite except for the scarcity of biotite. The relations between the minerals are similar in the biotite-bearing and the biotite-poor rocks, though no perthitic microcline was found in the biotite-poor rocks.

At only one place was the lineated quartz monzonite found to be fresh enough to permit a thin section to be prepared from it. In this place the biotite content seemed to be abnormally low. The relative freshness may result from a low content of both biotite and plagioclase, the two most readily weathered minerals. The modal composition is given in Table 3. Because of the weathering, paragenetic studies and the identification of plagioclase and biotite were difficult. The micas probably formed early during the consolidation of the rock.

Age and origin.--The Cherryville quartz monzonite was emplaced in metamorphic rocks that were brittle immediately before intrusion. This is shown by the blocky shape and abundance of the inclusions as well as by the fracturing in the metamorphic rocks in the zones of intrusion. Farther west the rocks of the Carolina gneiss were brittle at the time of intrusion of the pegmatites related to the Cherryville quartz monzonite. This indicates that the Cherryville quartz monzonite was emplaced long enough after the last major period of metamorphism and plastic deformation for the rocks to have become rigid probably because of a reduction in temperature and pressure. The metamorphism took place about 400 million years ago, or probably in Ordovician time. About 120 million years elapsed between the metamorphism and the intrusion of the Cherryville quartz monzonite if we accept the evidence of a single age determination, made by the Larsen method, of monazite obtained from the rock at Muddy Creek. The age was 285 million years which may correspond to middle Devonian. The monazite was exceptionally rich in uranium, which suggests that the age should be accepted only tentatively, pending confirmatory determinations. Additional samples have not yet been made because of the scarcity of zircon and monazite in the rock. The Devonian (?) eruptive activity in the Piedmont should have left its mark in the sedimentary rocks in eastern Tennessee. Unfortunately, only a small part of the Devonian period is represented in that area. The absence of strata may indicate uplift and crustal unrest during Devonian time, which would be compatible with a Devonian age for the eruption of the Cherryville quartz monzonite.

In many places bulges in elongate bodies of Cherryville quartz monzonite and the smaller stocks have crowded aside the enclosing gneisses. It is not possible to prove that all the space occupied by the intrusives is accounted for by the warping of rocks near the intrusive bodies. However, the evidence of crowding aside of walls by the quartz monzonite bodies and the absence of evidence for metasomatism indicate an intrusive origin for the Cherryville quartz monzonite.

Diabase

Northwesterly-trending dikes of fine-grained diabase are common in the Gaffney, Kings Mountain, and Lincolnton quadrangles but were not found in the Shelby quadrangle. Several of the dikes appear to be 8 to 12 miles long. These quite likely are interrupted but the barren parts of the guiding fractures are relatively short. Other dikes can be traced only for a few hundred feet, or even less. The thickness ranges from a few inches to several tens of feet. Graton (1906, p. 23) reports a maximum thickness of 200 feet. The dikes cut nearly all the other rocks in the area. They have not yet been found cutting the sheet-mica-bearing pegmatites that are related to the Cherryville quartz monzonite but they do cut the quartz monzonite and non-mica-bearing pegmatites related to it. This apparent late emplacement, together with the abundance of dikes of similar rocks in the sediments of Triassic age in the Deep River area to the east, suggests that the dikes are part of the swarm that was emplaced during Triassic time through much of the Piedmont and Blue Ridge.

The diabase is uniform in composition and is composed largely of calcic labradorite (48 percent), augite (37 percent), olivine (13 percent) and ilmenite (2 percent). In his notes, Sterrett mentioned that pyrite was found in the diabase at one place and chlorite in one place. Graton (1906, p. 23) mentions perovskite and found some olivine altered to micaceous-appearing iddingsite. Of these minerals only chlorite was found during the recent study.

The diabase weathers along joints, resulting in boulders of hard, black diabase in a brown clay matrix. The boundary between fresh and thoroughly weathered rock is commonly very sharp. The hard boulders accumulate at the outcrop as the clay is washed away.

Structure and metamorphism

General features

The Shelby district may be divided along a northeastward-trending line into a southeastern and a northwestern structural province; the dividing line lies close to the tin-spodumene belt and is near the boundary between the metamorphic provinces. The southeastern province is more homogeneous structurally than the northwestern. The provinces differ lithologically as well as in the intensity of metamorphism.

Structure of the southeastern province

Regional features.--The cleavage, bedding, and rock units in the southeastern province trend northeasterly, gently bending from N. 55° E. near Gaffney to N. 25° E. near Lincolnton. The discontinuously exposed beds of quartzite and marble that are near the western edge of the province are the only sedimentary features that can be used for reference or as index beds through the entire area examined, though the manganiferous schist and conglomerate beds and the eastern boundary of the schist unit that contains abundant volcanic materials are useful for reference in the Kings Mountain and Gaffney quadrangles.

Folds.--Folds in the Battleground schist are obvious where they are small enough for the crests or troughs to be exposed in outcrops or excavations. Most of the small folds in marble and many in quartzite seem to be open, although most of those in the sericitic schists as well as a few in the quartzites are isoclinal. The axial planes of the folds are most commonly very steep or vertical and most of the axes of small folds that are not in kyanitic quartzite plunge gently either to the northeast or southwest, rarely more steeply than 30 degrees. The small folds in the kyanitic quartzites are not as consistent in either direction or angle of plunge. Many plunge very steeply or are vertical. Within a few feet distance, the angle of plunge may range through 30 degrees. Folds are difficult to recognize in the sericite schists because there are few key beds. For that reason, most folds in sericite schist probably have escaped notice.

Sterrett has pointed out that the ore body at the Kings Mountain gold mine formed along the flanks of folds in a domomite marble bed that is enclosed in sericitic schists. (Keith and Sterrett, 1931, p. 8 and field notes). Sterrett reports that these folds plunge northeastward at an angle of 20 degrees.

Faults.--Faults are exposed in many places. Some can be recognized as cracks. The rocks on opposite sides of the faults may be different in attitude or in lithology and they may show signs of drag near the faults. Most of these faults strike nearly parallel to the trend of the enclosing rocks. As a result they do not offset beds or produce other recognizable structures where examined on relatively flat-lying surfaces. It is likely that the displacement along most of the faults has been small; the estimate of Keith and Sterrett (1931) that the displacements were a few inches to several thousand feet seems appropriate.

Larger faults can be detected by discordances in trends of beds and abnormal thicknesses or complete absence of beds, as well as peculiar outcrop patterns. There are undoubtedly many more faults than are indicated on the map. Only those for which there is good evidence and which are needed to explain the patterns are shown.

Cleavage.--Cleavage is well developed in all the micaceous rocks and in many of the hornblendic rocks but is not prominent in the marbles and quartzites that are free from micas and amphiboles. The cleavage results from the parallel orientation of the cleavage surfaces of mica flakes and the near parallelism of the cleavage surfaces of amphibole prisms. The effect is, in general, due to a combination of dimensional and crystallographic orientation.

Quartz grains in some quartzites tend to be elongate parallel to the bedding and to the cleavage. In the conglomerate of the Battleground schist formation at Dixon Gap the layering and very poorly developed cleavage in the quartz pebbles are parallel to the long directions of the pebbles. The cleavage in the sericite schists is parallel to the bedding in the thick beds of schist alongside the conglomerate bed. The cleavage is also parallel to bedding in the small flakes that are embedded in the quartzite matrix of the conglomerate. The cleavage is not everywhere parallel to beds, however. On Draytonville Mountains the two structures intersect at a large angle, as can be seen in figure 12. The pebbles in a conglomerate bed a few yards away were not elongated parallel to the line of intersection of bedding and cleavage.

Linear structures.--The most common linear structures in the Battleground schist are tiny wrinkles scarcely 1/20 inch across that offset the cleavage surfaces very slightly and resemble miniature chevron folds (fig. 15A). These result from movements along gently-dipping planes.

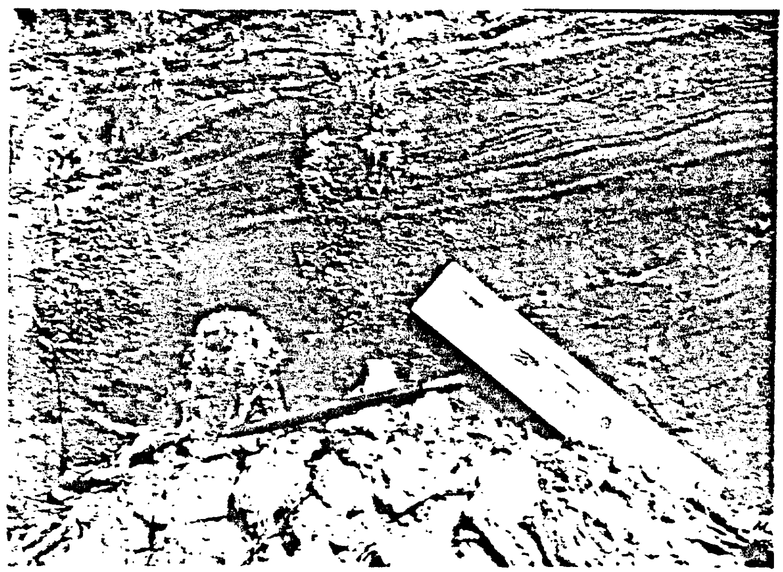


Figure 12. Bedding and cleavage in phyllite of Battleground schist. Bedding is parallel to the pencil and cleavage is parallel to the scale. Draytonville Mountain, Gaffney quadrangle, South Carolina.

Quartzite beds are pencilled where they are enclosed in schist that is wrinkled. The pencils are parallel to the wrinkles, and they result from the intersection of joints parallel to bedding and gently-dipping joints parallel to the planes of movement that formed the wrinkles.

Structure of the northwestern province

Regional features.--The over-all structural pattern in the northwestern province is one of tight isoclinal folds, as shown by the outcrop pattern of layers of sillimanite schist, and of broad areas in which the dips are gentle and the attitudes are uniform. Both strike faults and transverse faults occur there.

Folds.--A broad upfold is evident from the northwesterly trends of rocks and foliations in the northwestern part of the Shelby quadrangle and southwesterly trends in the area farther south. This broad warp is called the Mooresboro anticline (fig. 11). Though the nature of the movements that caused the warp are uncertain, the fold is of great extent. Another large-scale fold is outlined by the biotite gneiss in the Lincolnton, Shelby, and Gaffney quadrangles. This large fold is complicated by smaller folds superimposed upon it. Part of the broadening of the outcrop of Shelby gneiss in the northeastern part of the Lincolnton quadrangle may be attributed to folding. The eastern fold outlined by the gneiss in the north-central part of the Lincolnton quadrangle probably is a syncline and the fold in the northwestern part of that quadrangle is probably an anticline. Both plunge to the north.

Faults.--Faults of several types occur in the northwestern province. Thrust faults apparently bound the elongate body of biotite gneiss in the southern part of the Shelby quadrangle. The gneiss has overridden the sillimanite and biotite schists that dip under it from the west. The total displacement is unknown but it is probably small. The faulting may have resulted from shearing along the edges of the block of relatively competent gneiss inbedded in incompetent schists in which strong lit-par-lit slipping was taking place. The thrust faults thus would be features of local rather than regional extent.

Several northeasterly-trending faults that offset a sillimanitic layer in the Carolina gneiss as much as 3,000 feet were found in the Shelby quadrangle. The faulting took place before or during the metamorphism of the rocks as the fractures have healed. The direction and angle of dip of the faults are not known. Faults with a similar northeasterly trend, but with slight displacement are common in the central part of the Shelby quadrangle. Most of these are occupied by dikes of pegmatite related to the Toluca quartz monzonite (fig. 13). They dip steeply, and along most faults the western wall was upthrown. Movement along the faults caused drag folding of the metamorphic rocks and, in places, grooving of the walls. The movement continued on some faults while pegmatite was crystallizing along them, for quartz and other minerals were drawn out into leaves parallel to the grooving on the walls, giving rise to structures like those shown in figure 14.

The fractures that controlled the emplacement of the mica-bearing pegmatites are younger than those that are mentioned above as being occupied by dikes of pegmatite related to the Toluca quartz monzonite. There is a little drag folding of their walls, though not as much as

Figure 13A. Weathered dike of pegmatite related to Toluca quartz monzonite in fault in Carolina gneiss. Near Washburn Siding, northwest of Shelby, North Carolina.

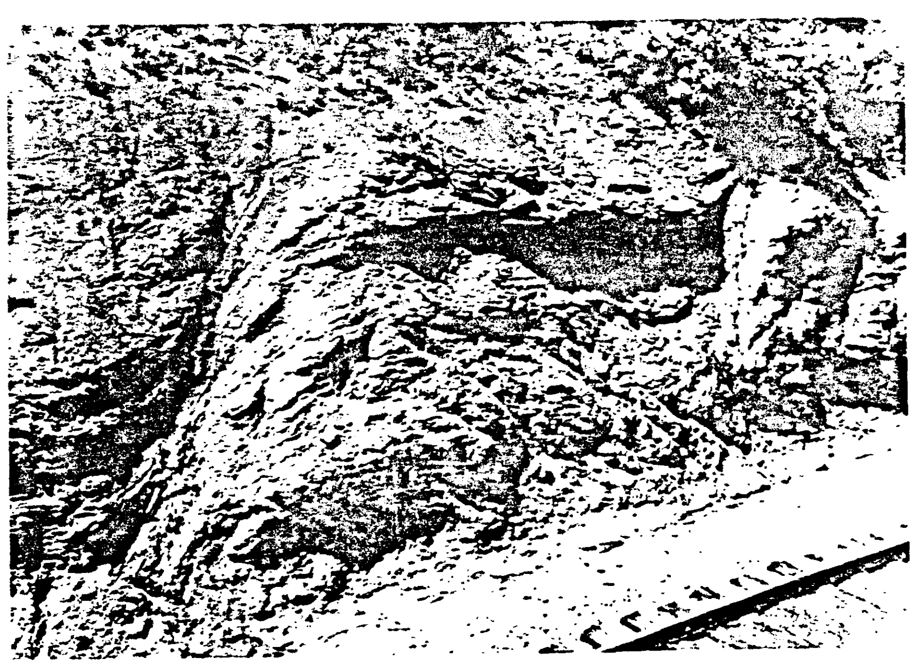


Figure 13B. Pair of faults in Carolina gneiss. Note drag folds in wall rocks. Faults of this type commonly contain dikes of pegmatite related to the Toluca quartz monzonite. Same locality as figure 13A.

along the older faults. These fractures are discussed in the chapter on pegmatites.

Foliation.--The compositional layering and the cleavage in the Carolina gneiss are parallel to one another and to the original bedding. The layers of lime-silicate rock and the rare quartzite beds, the best traceable beds, are parallel to the layers of sillimanite and biotite schist, which are therefore thought also to represent beds. In two places, one near the northwestern corner of the Shelby quadrangle and one in the northwestern outskirts of Shelby, layers of sillimanite schist are clearly discordant to the layers in the enclosing biotite schist and gneiss. These sillimanite schist layers probably flowed plastically into fractures or weak zones in the overlying rocks. The flowage of sillimanite schist was insufficient elsewhere to disturb the normal parallelism between compositional layers.

The cleavage or schistosity of the Carolina gneiss is parallel to compositional layering and bedding everywhere except in the two cross-cutting layers of sillimanite schist mentioned in the paragraph above.

Linear structures.--Linear structures are common in the metamorphic rocks as well as in the Toluca quartz monzonite and the pegmatites related to it (figs. 6 and 14). In the quartz monzonite, part of the quartz is segregated into elongated leaves that generally are shorter than 1 inch, though they may reach lengths of several inches. The width is commonly about a fifth of the length and the thickness is only a small fraction of the width. These leaves of quartz commonly are consistent in trend, though not in angle of plunge. Inasmuch as they lie in the foliation planes of the quartz monzonite, the direction and angle

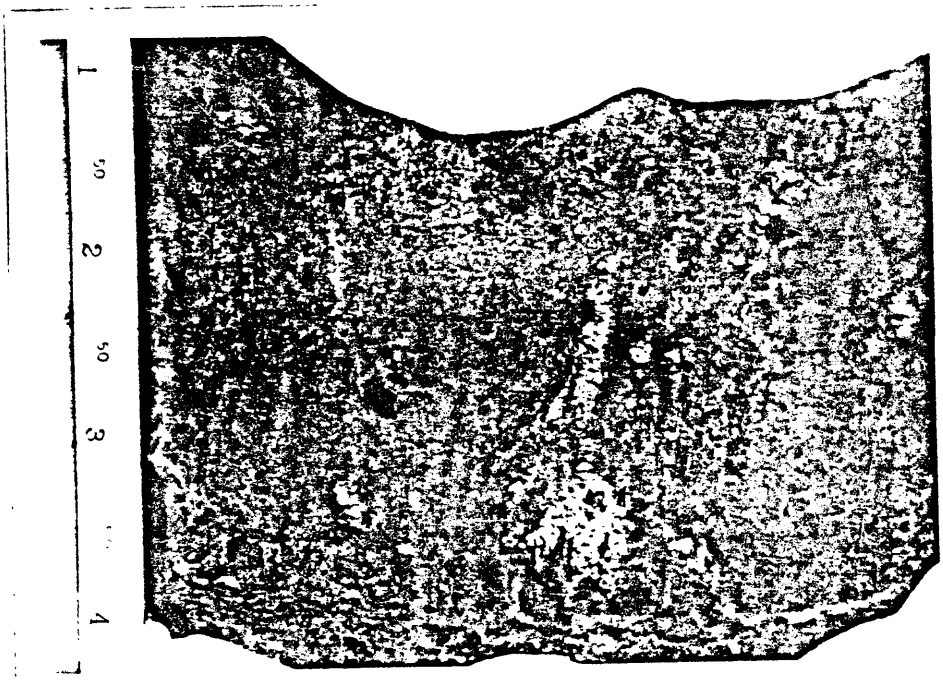


Figure 14A. Grooving of pegmatite at contact of dike related to the Toluca quartz monzonite. Major scale divisions 1 inch. Five miles northwest of Shelby, North Carolina.

Figure 14B. Lineation (elongated leaves of quartz) in pegmatite, same dike as figure 14A. Scale approximately the same as in figure 14A.

of plunge is determined by the local attitude of the foliation. Clusters of biotite flakes may be elongated to form linear structures that are parallel to the leaves of quartz. These are distinctly less common than the leaves of quartz. The lineated Cherryville quartz monzonite also contains spindles of quartz and biotite. The quartz spindles are not as flat as those in the Toluca quartz monzonite and many are interconnected.

The metamorphic rocks that enclose the sills of Toluca quartz monzonite contain linear structures that are similar in nature and attitude to those in the quartz monzonite. Leaves of quartz are uncommon but may be larger than those in the quartz monzonite. One was more than 2 feet long and 2 inches wide. Elongated clusters of biotite flakes are also uncommon.

The more common types of linear structures in the metamorphic rocks are small folds, boudinage, and grooves on the foliation planes. The small folds are generally several feet across and may be several tens of feet across. They plunge gently to the south or southwest in the area south of the Lattimore warp and to the north or northwest north of the warp. The boudinage structures are elongate swellings in layers of schist. These are rarely pinched off from the rest of the layer. They seem to be parallel to areas of the minor folds. The grooves on the foliation planes are conspicuous where the mica schist contains hard, dense layers or lenses of lime-silicate rock (fig. 3B). They probably result from slipping between the layers in the schist.

The linear structures in the Toluca quartz monzonite formed late in the history of the rock. The quartz leaves lie in planes that apparently represent fractures that formed after the feldspars as well as part of the quartz. The last of the quartz was still mobile enough to move into the fractures. The leaves of quartz in the metamorphic rocks also formed late and occupied fractures parallel to the foliation. The folding of the gneiss and the sills seems to have been followed by a gradual stiffening of the rocks, the stiffening increased, the same layers became rigid and the movements gradually became concentrated in the less competent layers. Ultimately most of the slipping took place in the nearly consolidated sills of Toluca quartz monzonite. After movements parallel to the foliation had stopped the metamorphic rocks were cut by joints and faults that crossed the foliation, and the pegmatites related to the Toluca quartz monzonite were emplaced to form dikes. Drag folds and striations on the walls of the dikes resulted from movement along the faults, and elongate masses of quartz in the pegmatite resulted from continued movement of the walls during the consolidation of the pegmatite. The parallelism between the lineation in the pegmatite and the striations on the walls indicates that the movements that formed them were parallel.

Nature and degree of metamorphism

The metamorphism in the Shelby district comprises both textural and mineral changes. The most prominent textural change in the schists is an increase in grain size of the rocks as the areas of stronger metamorphism are approached. Cataclastic features are only locally conspicuous. Mineral transformations are important, and can, to some extent, be correlated with grain size.

The district includes a northwestern province in which the rocks characteristically contain biotite, garnet, sillimanite, plagioclase feldspar, and quartz, and a southeastern province that contains abundant and widespread schists composed largely of sericite and quartz with minor but widespread staurolite, and local garnet, chloritoid, and kyanite. Biotite and hornblende schists also are in the southeastern privince. Broadly considered, the whole region might be included in the amphibolite facies of metamorphism, though most of the southeastern area would be in the lower part or staurolite-kyanite subfacies of Turner (1948, p. 81-85) of the amphibolite range, and the northwestern area would be in the upper part of the range or the sillimanite-almandine subfacies of Turner (1948, p. 85-87).

Metamorphism in the southeastern province

Potter (report in preparation, 1955), after studying part of the southeastern province, infers that there is an eastward increase in metamorphic grade from greenschist facies near the middle of the schist exposure through epidote-amphibolite facies to amphibolite facies as the Yorkville granite is approached. It is not entirely certain that the rocks most distant from the granodiorite belong to the greenschist facies, as the relations of epidote, chlorite, chloritoid, and the very widespread staurolite are rather uncertain. It is clear, however, that the grade of metamorphism increases eastward toward the Yorkville granodiorite, as the kyanite in the kyanitic quartzites is replaced by sillimanite near the granite. A similar replacement of kyanite by sillimanite was found near the eastern edge of the batholith.

As one passes westward in the western part of the Battleground schist area, the grain size is found to increase over a distance of a mile or more. The coarsening effects certain layers in the schist,

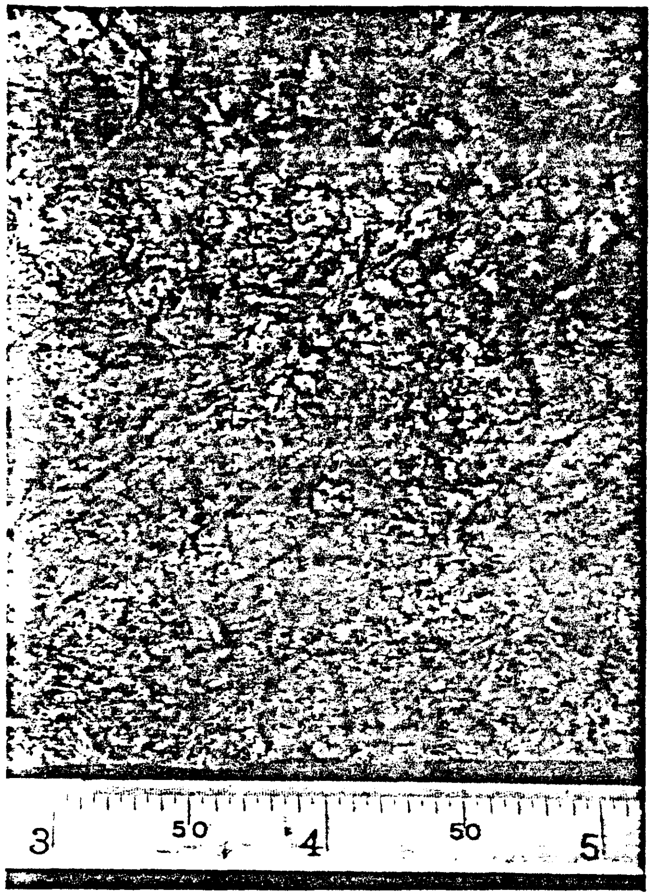


Figure 15A. Tiny wrinkles in phyllite of the Battleground schist.

Figure 15B. Pencilled quartzite in the Battleground schist. Pencils plunge parallel to tiny wrinkles in phyllite nearby.

and, as metamorphism becomes stronger, the number of coarsened layers increases until the rock is largely coarse-grained. At about this point, sillimanite becomes an important constituent of the remaining fine-grained layers. Where the schist is partly recrystallized it has a peculiar puckering, similar to a miniature dome and basin structure (fig. 16). This structure is obscured by complete recrystallization. The general pattern in the southeastern province is one of a relatively low-grade central area (epidote-amphibolite or possibly greenschist facies) bounded on each side by rocks of the amphibolite facies. The nature of the alteration differs on the two sides, as the most prominent feature on the west is a coarsening of grain, whereas the most prominent feature to the east is the substitution of sillimanite for kyanite. The eastward increase may result from something akin to contact metamorphism by the granite, whereas the westward increase seems to be a more normal regional metamorphic gradient.

Metamorphism in the northwestern province

In the northwestern province the Carolina gneiss contains minerals characteristic of the upper amphibolite facies of metamorphism that have been deformed somewhat by the movements accompanying the mineral changes. Sillimanite had formed as a relatively early mineral but remained stable to the end of the orogeny. A little staurolite was found by panning the weathered rock near the center and the northwestern corner of the Shelby quadrangle. This might be relict from the early stages in the metamorphic cycle; retrogressive, late in the cycle; or it might result from a second period of metamorphism, possibly related to the Cherryville quartz monzonite.

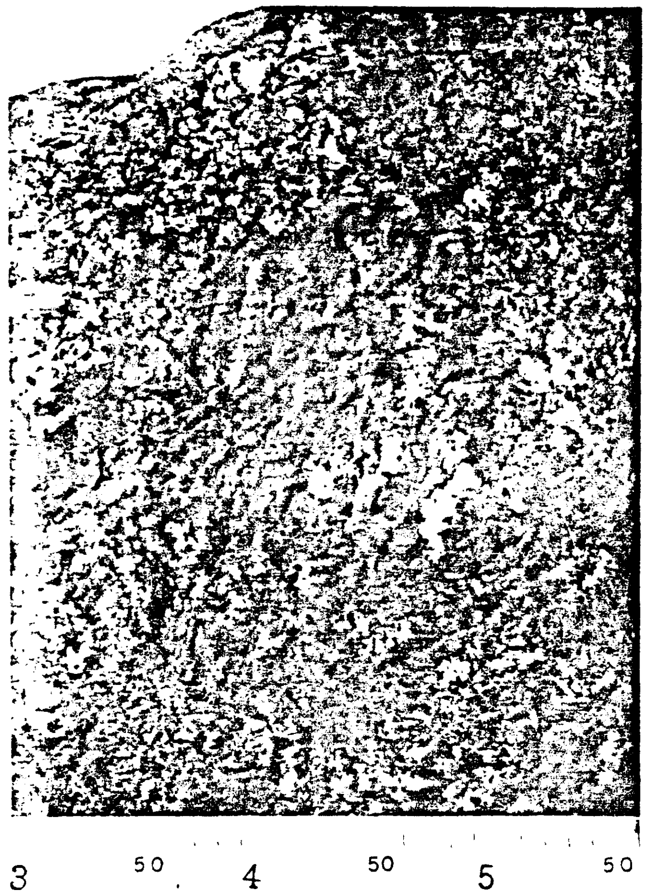


Figure 16. Puckered cleavage in coarsened Battleground schist. This feature develops as some layers in the phyllite are recrystallized to form muscovite schist. Largest scale divisions are 1 inch. Near Lincolnton, North Carolina.

Iron sulfides, with a little feldspar, were segregated locally in the Carolina gneiss late in the main period of metamorphism. In some of these segregations sillimanite forms imperfect crystals or bundles of parallel crystals that reach lengths of 2 inches and widths of 1/3 inch. The amount of coarse sillimanite and sulfide varies. In some places limonite or coarse sillimanite float is moderately heavy over areas of an acre. The sulfides presumably migrated in solution, and the complete lack of deformation of the sulfide masses or of the coarse sillimanite indicates that they solidified after the cessation of the pervasive movements within the rocks, and the formation of coarse sillimanite indicates that the conditions were still those of sillimanite grade, or amphibolite facies, of metamorphism. This is the general condition that prevailed late in the orogeny during and after the formation of the Toluca quartz monzonite and pegmatites related to it.

The fine grain in the Battleground schist as compared with the Carolina gneiss may be attributed to one or more of several factors:

1. Lower pressure and temperature during metamorphism.
2. Slowness of recrystallization and coarsening of muscovite as compared with biotite in the same environment.
3. A great difference in age between the biotitic and sericitic schists, permitting the biotitic schists to undergo more than one period of metamorphism whereas the sericitic schists underwent only one.
4. Retrogressive metamorphism, with inducement of fine grain in a formerly coarser-grained rock.

The third and fourth factors seem to the writer to be the least likely to be dominant. The first two probably were both important.

GEOLOGY OF PEGMATITES

Pegmatites related to the Toluca quartz monzonite

Distribution and mode of occurrence

Pegmatite related to the Toluca quartz monzonite is found in nearly all areas in the district that are underlain by Carolina gneiss. The pegmatite is enclosed in Carolina gneiss, including the biotite gneiss member, and in the Toluca quartz monzonite. In the Shelby quadrangle, the sillimanite schists in the south-central part of the quadrangle probably contain the largest amount of pegmatite, though much pegmatite is also found in the biotite and sillimanite schists near the large bodies of Toluca quartz monzonite near the center.

In quartz monzonite the pegmatite forms pods and irregular tabular masses, generally a foot or less in thickness. In the metamorphic rocks pegmatite forms lenses, elongate parallel to the foliation ranging in thickness from an inch to about a foot. It also forms more nearly tabular layers that are parallel to the foliation and are a few inches to several tens of feet thick and also tabular or lenticular dikes, crossing the foliation, that reach widths of 20 feet. In many places the dikes and sills are clustered into swarms of parallel bodies.

In many places in the south-central part of the Shelby quadrangle the pegmatite dikes are along faults that have slightly offset the rocks of the Carolina gneiss (fig. 13). Only a few of the faults offset the rocks enough to show in the outcrop pattern. In other places, west of Shelby, similar dikes lie along faults that conspicuously affect the outcrop patterns. The hanging wall of most faults was downthrown and most of the faults trend northeast.

Lithology

The pegmatites related to the Toluca quartz monzonite are composed largely of microcline, plagioclase, and quartz. The common accessory minerals are biotite, muscovite, and garnet. Biotite and garnet, the most common of the accessories, rarely constitute more than 10 percent of the rock. Muscovite forms small flakes, rarely exceeding 1 inch in breadth, in these pegmatites. Monazite, zircon, ilmenite, rutile, and sillimanite were each found in more than half of 329 pegmatite samples. Magnetite, staurolite, spinel, dumortierite, and pyrite were found in fewer samples. The only evidence of zoning in the pegmatites related to the Toluca quartz monzonite is the formation of quartz masses surrounded by feldspathic pegmatite. These quartz masses commonly are not elongate parallel to the body of pegmatite but are transverse and resemble veins that formed very late during the crystallization.

The texture of most of the pegmatites is granitoid, with microcline megagrains ranging in length from less than an inch to a foot. These are surrounded by plagioclase, generally in grains smaller than 1/2 inch, and quartz, also in small grains. The result, in places, is a porphyritic texture.

In the pegmatites related to the Toluca quartz monzonite the microcline forms megagrains as broad as 6 inches as well as tiny grains 1/20 inch or less across. It is somewhat translucent and gray. The larger grains in many places are markedly curved, as can be seen from light reflected from cleavage surfaces. The microcline in places has a slightly opalescent luster, forming a rather low-quality moonstone. The opalescence seems to increase with slight weathering and softening of the microcline. Microcline that is identical in appearance forms metacrysts and eyes in the Carolina gneiss. These metacrysts range in breadth from 1/2 inch or less to 1 foot. Some, especially the larger ones, are accompanied by plagioclase. The plagioclase with the metacrysts was not found in unweathered condition so its composition is unknown.

Plagioclase forms fine-grained, sugary-textured masses between grains of microcline and quartz and enters quartz-plagioclase aggregates in which the coarser microcline grains are embedded. Individual grains of plagioclase in these pegmatites generally are smaller than 1/20 inch though aggregates of the small grains may be an inch across. The plagioclase is somewhat more calcic than that from the pegmatites related to the Cherryville quartz monzonite, falling in the calcic oligoclase-andesine range. In some rocks the plagioclase, like the microcline, has undulatory extinction, and the plagioclase that forms films between the microcline grains may be more sodic than that which forms larger masses. In one rock the films contained 83 percent albite, whereas the rest of the plagioclase contained 73 percent. The more sodic plagioclase also forms rinds on plagioclase grains adjacent to microcline grains. These relations suggest that the more sodic plagioclase may have incorporated albite that had been exsolved from the microcline during cooling. This is compatible with the idea of Chayes (1952) that stress stimulates the exsolution of albite from microcline containing a small amount of albite in solid solution. Chayes wrote of the formation of perthites by exsolution, however, and not the complete expulsion of albite from the microcline grain, as may have occurred in the pegmatites related to the Toluca quartz monzonite.

In one thin section from a dike the microcline contained tiny needles that may be sillimanite. The quartz grains are somewhat granulated and show undulatory extinction. Garnet (probably almandite) is a common constituent in spherules as much as 1/4 inch in diameter. It apparently formed late, after the crystallization and deformation of the microcline and quartz.

Origin

The pegmatites related to the Toluca quartz monzonite were formed at the end of an orogenic period as sills parallel to the foliation of the Carolina gneiss as well as in dikes that cross the foliation. There is little evidence as to the manner of emplacement of the sills other than the curving of the foliation around their flanks, which indicates dilation. A little more information can be obtained from the dikes. The foliation of the Carolina gneiss is dragged in folds that extend several feet from the faults along which the pegmatites were emplaced. If the masses of pegmatite formed by replacement of the gneiss along these faults, the parts of the drag folds in the gneiss that was replaced would be destroyed, so that the width of the fold would vary inversely as the thickness of pegmatite in the dike. No such relation was found. The dilation indicated by the sills and the lack of important replacement indicated by the dikes are compatible with an intrusive magmatic origin of the pegmatite. The genetic relationship between the Toluca quartz monzonite and the pegmatite is indicated by the following facts: (1) both were formed at about the same time during the orogeny, (2) the pegmatite is at least in part younger than the quartz monzonite, (3) they both contain monazite and xenotime which are not as abundant in the other rocks, and that (4) both contain a peculiar gray microcline not found in the younger pegmatites.

Pegmatites related to the Cherryville quartz monzonite

Distribution

In 1945, when the results of the mica investigations made during World War II were prepared for publication, the Shelby district of this report was treated as the southern division of a larger Shelby-Hickory district (Griffitts and Olson, 1953, p. 209). The northern division of the Shelby-Hickory district, which extends north from the vicinity of Hickory, is not discussed in the present report.

The main pegmatite areas in the Shelby district are:

1) The tin-spodumene belt; a series of pegmatite dikes in a belt that does not exceed 1-1/2 miles in width extending northeastward from Gover to Lincolnton, a distance of 24 miles. This belt forms the eastern boundary of the Shelby district in Cleveland, Gaston, and Lincoln Counties, North Carolina.
2) The scrap mica belt. This area was divided into the Kings Mountain area and the Cherryville area by Griffits and Olson (1953, p. 209). It is the central part of a belt of muscovite-rich pegmatite that extends 50 miles northeastward from a point at least 7 miles southwest of Gaffney, South Carolina, to Maiden, North Carolina. Where the tin-spodumene belt is absent, the scrap mica belt marks the eastern edge of the district.
3) The Shelby area, like the others listed below, is separated from the tin-spodumene belt and the scrap mica (Kings

Mountain-Cherryville) belt by the Cherryville quartz monzonite batholith. It is in the south-central part of the Shelby quadrangle.

4) The Fallston area lies along the north-northwesterly-trending western boundary of Lincoln County. Most of the pegmatite dikes are in Lincoln County, but some are in Cleveland County and a few are in Gaston County.

5) The Polkville area includes eight dikes that are quite similar to one another in a group that is north of the Shelby area.

6) The Indiantown area includes a compact group of pegmatite dikes in northeastern Cleveland County, northwest of the Fallston area.

7) The Burke Chapel area in southeastern Burke County and the adjacent southwestern corner of Catawba County, contains at least 11 pegmatite dikes that have been explored for sheet mica.

The northern division of the Shelby-Hickory district, according to Griffitts and Olson (1953, p. 209), contains the three main groups of pegmatite dikes listed below:

1) The Conover area in northern Catawba County, with at least 7 dikes.

2) The Hiddenite area, in central Alexander County, which contains at least 6 mica and beryl prospects in addition to the famous mine from which hiddenite (emerald-green spodumene) and emerald were once obtained.

3) The northwestern Iredell County area, with at least 6 mica prospects.

Pegmatite dikes in the tin-spodumene belt are largest and most numerous south of the town of Kings Mountain in what is referred to as the "Kings Mountain area." A second concentration, near the north end of the belt is called the "Beaverdam Creek area" because it includes parts of Beaverdam and Little Beaverdam Creeks. These differ from the areas listed under the muscovite-bearing pegmatites in that they are not surrounded by essentially pegmatite-free terrane, but are the two greatest concentrations of pegmatite dikes in a continuous belt.

The tin-spodumene belt, as can be seen in figure 2 forms a broad "S"; its northern and southern parts trend northeasterly and the middle part trends about north. The southern part is along the eastern side of the Cherryville batholith where the quartz monzonite approaches most closely to the Gaffney marble and other sedimentary rock units in the Battleground schist. The pegmatite there lies largely in muscovitic schist, hornblende schist and, to a very minor extent, in Cherryville quartz monzonite. Most of the muscovitic schist that encloses the pegmatite is moderately coarse-grained, but that found on Whittaker Mountain, near Blacksburg, South Carolina, is a fine-grained sericite schist. This is of some interest, inasmuch as Keith and Sterrett (1931) considered the gneissic pegmatites in the region to be of Precambrian age and the schist on Whittaker Mountain to be of Cambrian age, yet the pegmatite enclosed in the latter is somewhat gneissic.

Northward the belt bends sharply at the town of Kings Mountain, whence it continues nearly due north for a distance of 5 miles, where it bends eastward again and continues in a northeasterly direction--the northern part of the broad "S". In the middle part of the "S" the pegmatite dikes lie largely in hornblende schist and gabbro, although a few are in muscovite schist or dark-colored biotite gneiss. Most of the dikes that are enclosed in Cherryville quartz monzonite are in this area and are near Long Creek Church. As the pegmatite belt bends eastward, it passes around the smaller body of Cherryville quartz monzonite that is in the east-central part of the Lincolnton quadrangle. Thus, in the middle and northern parts, the tin-spodumene belt lies between two bodies of Cherryville quartz monzonite.

The total length of the tin-spodumene belt is about 35 miles, but if the part that is rich in pegmatite is considered, the length is 24 miles. The width ranges from 1/2 mile to 1-1/2 miles. The narrowest part, which is south of Kings Mountain, contains 100 to 650 feet total thickness of pegmatite distributed across a width of 1,500 feet; farther north the pegmatite belt as a whole is twice as wide or more but the total thickness of pegmatite rarely exceeds 200 feet and commonly does not exceed 100 feet. There is no regular decrease in the amount of pegmatite with increasing thickness of the belt, however, but rather a separation of the belt into two parts, a narrow part relatively rich in pegmatite and a wide part, relatively poor in pegmatite.

The attitude of the pegmatite dikes varies according to position in the northeastward-trending pegmatite belts. Thus, in the tin-spodumene belt and the scrap mica belt the dikes strike mainly north or northeast. In the eastern parts of the Shelby and Fallston areas most dikes strike easterly and in the western parts of those areas most dikes strike northeast, as do those in the westernmost belt of mica-bearing pegmatites. The dikes near Polkville in the northern part of the Shelby area strike northwest.

The trends vary markedly along the tin-spodumene belt; south of Kings Mountain most of the dikes are parallel to the belt, that is, they trend northeast. North of Kings Mountain the trends are less uniform, some are northeast, parallel to the belt, some northerly, and some in other favored directions.

Relation to host-rock lithology

Pegmatite has not shown any pronounced preference for particular host rocks other than the Cherryville quartz monzonite. In the tin-spodumene belt pegmatite is found in Cherryville quartz monzonite, muscovite schist, hornblende schist, and gneiss. The number of dikes in each type of rock probably is in direct proportion to the relative outcrop areas of the rocks in the pegmatite belt. The same is true for the pegmatite dikes in the scrap-mica pegmatite belt. Mica schist is the host rock for the largest proportion of the scrap-mica pegmatites and of the dikes in the tin-spodumene belt, because mica schist is the predominant rock in the belt. The pegmatite dikes in the sheet muscovite pegmatite areas west of the batholith of Cherryville quartz monzonite are largely in biotite and biotite-sillimanite schist. No preference for one or the other of these rocks has been found. The Toluca quartz monzonite seems to contain fewer dikes in proportion to its area than the other rocks in the Shelby and Fallston areas. A similar scarcity of pegmatite seems to characterize the biotite gneiss. This is apparently a result of the differences in competency between the weak schists and the rigid quartz monzonite and gneiss.

The Cherryville quartz monzonite probably contains the largest number of pegmatite bodies in proportion to its area. Inasmuch as the bodies are substantially smaller than those in the metamorphic rocks the amount of pegmatite would not be as impressive as the number of pegmatite bodies. These small bodies of pegmatite are most common in the part of the batholith south of Cherryville.

The lack of a clear relationship between the Cherryville batholith and the distribution of the different types of pegmatite in the Shelby district differs from the findings reported by geologists working elsewhere or who wrote about pegmatites in general. Among them Williams (1895, p. 683) pointed out that "Their (the pegmatite bodies) size and abundance are directly proportional to their nearness to some eruptive granite mass. At many localities they can be seen to decrease steadily in both number and size as they recede from such a granite boundary." Gevers used the position of pegmatite bodies with respect to granite contacts as the basis for a geologic classification of pegmatites (Gevers, 1936, p. 339-340) that has been adopted by Jacobson and Webb (1946, p. 116) and Heinrich (1948, p. 434, 442-448). According to this classification the division is into "interior pegmatites", which are within bodies of granitic rocks, "marginal pegmatites" which are near the margins of granitic bodies, and "exterior pegmatites", those that are outside of and some distance from the granitic bodies. The relation between mineralogic features of the pegmatites and the position of bodies of pegmatite with respect to bodies of granitic rocks has been brought out by Sterrett (1923, p. 7, 9), Olson (1944, p. 36), [~~1949, p. 13~~)] and Maurice (1940, p. 173-179). In the lack of close and important relations between the position of the pegmatite with respect to bodies of granitic rock the Shelby district resembles some other districts in the Piedmont.

Lithology

There are three main types of pegmatite in the Shelby district. In order of decreasing age they are: muscovite-oligoclase-quartz pegmatite, with or without perthitic microcline; albite-microcline-quartz pegmatite; and albite-microcline-spodumene-quartz pegmatite. Muscovite-oligoclase-quartz pegmatite is cut by weathered pegmatite that contains pale yellow-green muscovite but does not contain pseudomorphs of clay after spodumene. This probably was albite-microcline-quartz pegmatite. Several masses of albite-microcline-quartz pegmatite are cut by dikes of spodumene-bearing pegmatite. The muscovite-rich pegmatite is found in the sheet-mica mining areas west of the Cherryville batholith and in the scrap-mica pegmatites along the west side of the tin-spodumene belt. The pegmatite forms dikes and discordant lenses that may be more than 500 feet long. They generally are well zoned, though many are unzoned. The unzoned masses consist of plagioclase-quartz-muscovite pegmatite in which the plagioclase megagrains are 2 to 6 inches across, and the muscovite books are a fraction of an inch to about a foot in length. The zoned pegmatites have wall zones that are very similar to the pegmatite in the zoned dikes, though the muscovite books are larger, a few being 2 feet or more in length. There is more variety in the interior of the dikes, as some contain cores of perthite-plagioclase-quartz pegmatite or of massive quartz and others have cores of quartz with intermediate zones rich in either perthite or plagioclase. Several common zonal assemblages are listed below.

A.

1. Wall zone of plagioclase, quartz, and muscovite, with granitic texture.

2. Core of massive quartz.

B.

1. Wall zone of plagioclase, quartz, and muscovite with granitic texture.

2. Intermediate zone of perthitic microcline, with a blocky texture.

3. Core of massive quartz.

C.

1. Wall zone of plagioclase, quartz, and muscovite, with granitic texture.

2. Intermediate zone of plagioclase, perthitic microcline, quartz, and muscovite, with blocky texture.

3. Core of massive quartz. This is very small in some deposits.

D.

1. Wall zone of plagioclase, quartz, and muscovite, with granitic texture.

2. Intermediate zone of plagioclase, quartz, and muscovite with blocky texture. This commonly contains cavities lined with feldspar, mica, or calcite crystals.

3. Core of massive quartz.

The longest dikes generally contain zones of assemblages A or B. Some pegmatite along the west side of the tin-spodumene belt contains relatively little microcline and is exceptionally rich in muscovite but this is in books that cleave only with difficulty, differing from that in the pegmatite west of the Cherryville batholith.

The grains in the wall zones are devoid of crystal faces, except in one exceptional dike, at the E. R. Self mica mine, where some muscovite books have clear rhombic or hexagonal outlines. The blocky texture in the intermediate zones results from a subhedral to euhedral habit of perthitic microcline or of plagioclase. The plagioclase-bearing intermediate zones also have muscovite crystals that show a few faces. Around primary vugs in the intermediate zones the plagioclase and muscovite forms good faces. The quartz cores apparently filled cavities because in some deposits they are bounded by terminated feldspar crystals in an over-all structure similar to that of geodes that are partly filled with crystals. A microcline crystal that projected into a quartz core, and lined quartz from the core, are shown in figure 19B.

The albite-microcline-quartz pegmatite dikes in the tin-spodumene belt contain albite (generally Ab_{95} or higher) as the dominant constituent. It forms sugary aggregates with quartz that are interstitial to the microcline megagrains and the uncommon coarse quartz masses. The grains in this interstitial material rarely exceed 1/10 inch in size. The microcline megagrains are a fraction of an inch to a foot or more in length. Muscovite forms flakes a fraction of an inch across that amount to 1 or 2 percent of the rock. It has a light yellowish green color that contrasts with the dark red-brown color of the muscovite from the mica-rich pegmatites. Beryl is a widespread constituent and probably amounts to nearly a half of one percent of the pegmatite, but it can rarely be seen on outcrops because it is white and most grains are small and devoid of crystal faces.

ones are very poorly developed in albite-microcline-pegmatite. ost dikes are unzoned, others contain inner zones that are a little oarser than the outer parts or contain, as discontinuous cores, clots f gray quartz up to 3 feet long, with or without fringes of muscovite n books as long as 5 inches. One of these quartz masses was found to ontain a few pale green beryl crystals.

The albite microcline-spodumene-quartz pegmatite differs from the lbite-microcline-quartz pegmatite by its content of 10 to 20 percent of spodumene. The texture of the albite-quartz matrix and the contents of accessory minerals are essentially the same as in the spodumene-free pegmatite. The spodumene forms crystals that reach lengths of 2 feet, rarely longer. These generally are thick in proportion to their length in the area south of Kings Mountain and much more slender north of that town.

The internal structure of the spodumene-bearing pegmatite varies. South of Kings Mountain much of the pegmatite and associated aplite is gneissic. This is particularly evident on weathered surfaces, because the albite weathers away, leaving gashes, whereas the quartz and microcline stand out in strong relief and the spodumene in moderate relief but is generally rusty (fig. 17A). The pegmatite structure is not as clear on freshly-broken rock surfaces as on polished surfaces of fresh rock. In these, some details may be seen in the albite and spodumene that are obscured on the weathered surfaces (fig. 17B). The matrix in which the spodumene and microcline crystals are imbedded consists of alternating layers of albite and quartz, with a few very thin layers, rely more than 1/50 inch across, of dark blue apatite.

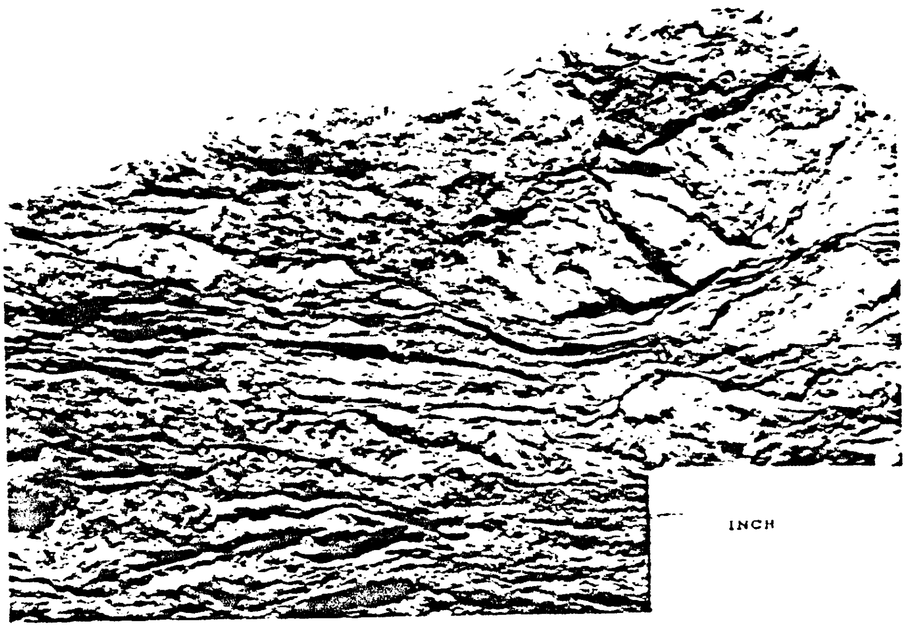


Figure 17A. Gneissic pegmatite. Albite layers have weathered, allowing the quartz layers in the matrix and the microcline and spodumene crystals to stand in relief. Foote Mineral Company quarry.

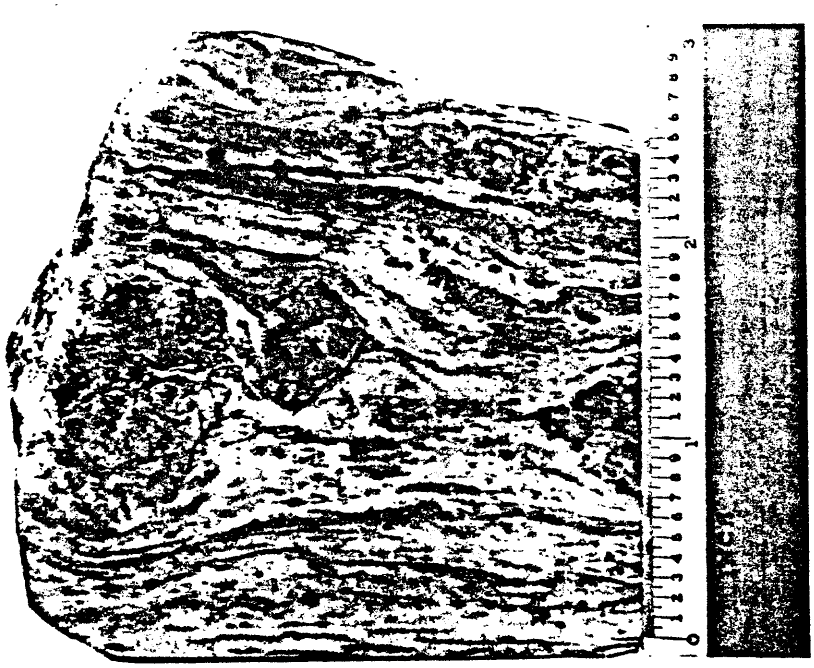


Figure 17B. Polished slab of gneissic pegmatite. Layers of albite and quartz wrap around coarse spodumene crystals. Apatite forms thin black (dark blue) lines at bottom of specimen and near largest spodumene grains. Foote Mineral Company quarry.

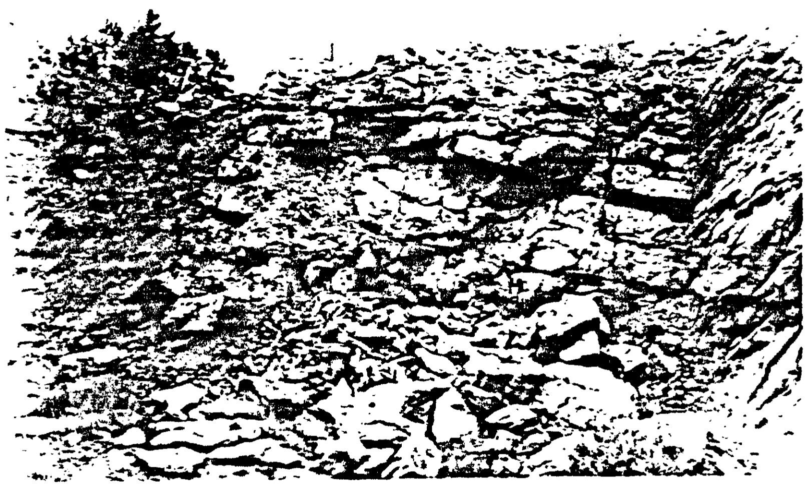


Figure 18A. Pegmatite dike cutting hornblende schist. Foote Mineral Company, Kings Mountain.

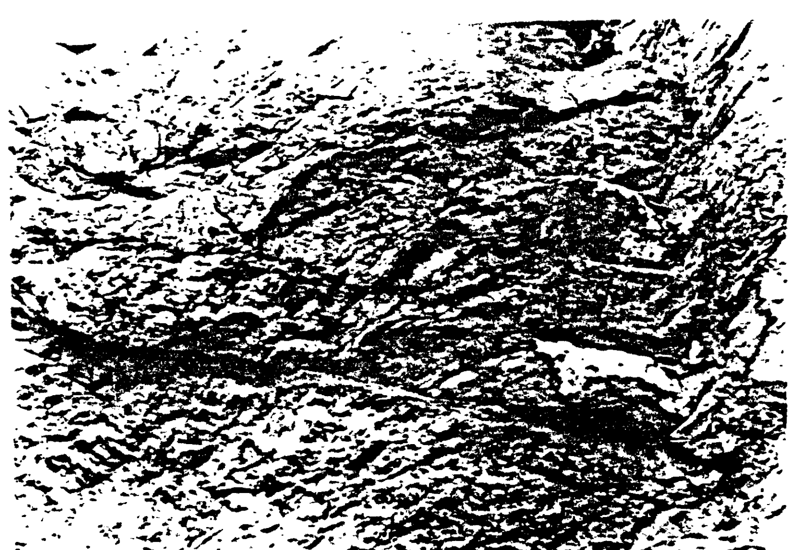


Figure 18B. Curved and cracked spodumene crystals (inclined from upper right to lower left) in aplitic albite-quartz matrix. Near right (east) wall of dike shown in figure 18A. View is about 2 feet wide.

Many of the crystals of spodumene and of microcline in the pegmatites south of Kings Mountain have curved cleavage surfaces or have been broken. The intact crystals as well as the broken pieces lie in the fine-grained albite-quartz matrix. This matrix material also fills cracks in the crystals. Some broken spodumene crystals have been cemented together with microcline as well as with albite and quartz. At the east wall of one dike the spodumene crystals extend into the dike from the wall but are bent so as to curve downward to the west. The most bent parts have cracked and matrix material fills the cracks. The rock that contains these long spodumene crystals is terminated abruptly to the west, apparently by a shear plane. Farther west, the dike consists of gneissic pegmatite. Most of the layers of pegmatite, as well as the layering of the rock, are parallel to the length of the dike. Several, however, cross the dike and the other layers of pegmatite. Most of these are only a few inches thick and consist mainly of sugary-textured quartz and albite.

A very late event in the formation of the gneissic pegmatite south of Kings Mountain was the destruction of spodumene, with the formation of porous masses of albite and bluish fluorapatite near coarser crystalline masses of blue to lavender apatite in fractures and cavities. The apatite is about the same color as fluorite in some other districts and it has been misidentified as fluorite. Another late event was the fracturing of the pegmatite, then the deposition of dark-brown sphalerite, pyrite, and probably pyrrhotite in the fractures. The more persistent fractures contain pyrite as the predominant sulfide, whereas the less persistent ones contain sphalerite (fig. 20). This fracturing and the late deposition of apatite took place after

Figure 19A. Quartz core, flanked by weathered plagioclase-quartz pegmatite of the wall zone. C. C. Blanton mica mine. Core is about 5 feet thick.

Figure 19B. Healed fractures in core quartz and euhedral perthitic microcline crystals from margin of quartz mass. A. F. Hoyle mica mine. Scale divisions are 0.1 inch.

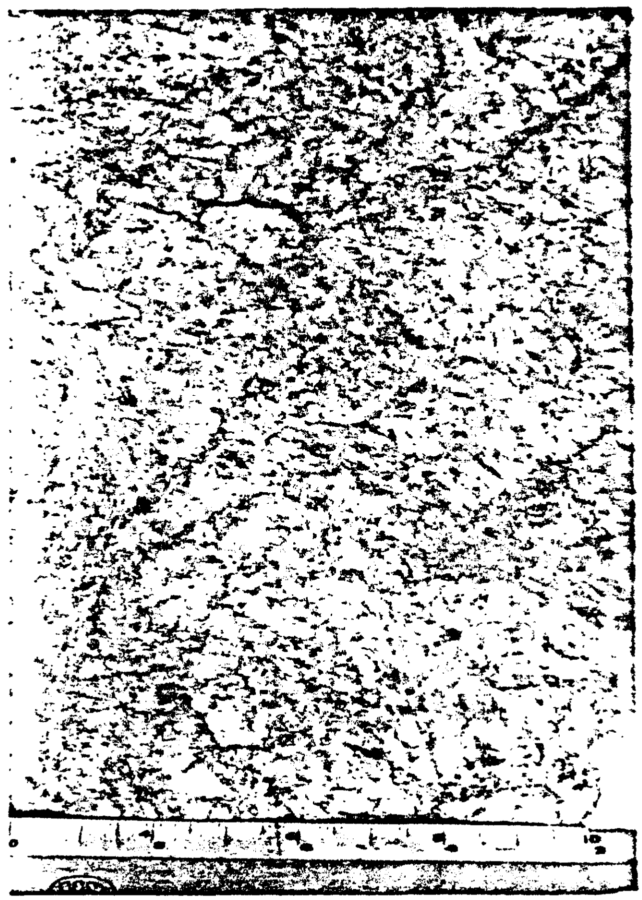


Figure 20. Sulfide-filled fractures in pegmatite. ersistent fractures trending from lower left to upper right contain pyrite. Less persistent fractures trending from wer right to upper left contain dark brown sphalerite. Lar st scale divisions are 1 inch. Foote Mineral Company quar .

the pegmatite had completely solidified to form a rigid mass, but their relative ages are not known. The recovery of sphalerite and pyrrhotite in the mill of the Foote Mineral Co. suggests that the minerals are rather widespread but are present in only very small percentages, probably not exceeding 0.01 percent by weight.

The pegmatite in the tin-spodumene belt north of Kings Mountain contains the same minerals as those south of the town but the gneissic structure is rare. Zones are present in some dikes but they are not particularly obvious. The most common zonal feature is a change in orientation of the spodumene crystals from approximately normal to the walls of the dike in the border and wall zones to essentially random in the cores. There is also a rather small increase in microcline content inward, with concurrent decrease in contents of albite and spodumene. The crystals of microcline and spodumene are several inches long and are embedded in a matrix of albite and quartz. Muscovite forms a few small flakes in the matrix. Broken crystals of microcline are not nearly as common as in the area south of Kings Mountain, where intact crystals are exceptional.

A rough estimate was made of the proportions of Cherryville quartz monzonite and the various lithologic types of pegmatite in the Shelby district related to it. The results, expressed as square feet of outcrop area, are:

Cherryville quartz monzonite	$6{,}000 \times 10^5$
Sheet-mica-bearing pegmatite	4×10^5
Scrap-mica-bearing pegmatite near the tin-spodumene belt	8×10^5
Albite-microcline-quartz pegmatite in the tin-spodumene belt	11×10^5
Spodumene-bearing pegmatite	32×10^5
Total pegmatite	53×10^5

It is striking that the amount of pegmatite increases regularly with decreasing age. Even though the errors in the estimates are undoubtedly large, the increase in amount is too large to be accounted for by errors. The total amount of pegmatite, including all types, is about one percent of the amount of quartz monzonite.

Wall rock alteration

On the whole, wall rock alteration near the pegmatite bodies of the Shelby district has been slight. Of course, a pegmatite magma that is crystallizing to an aggregate of muscovite, biotite, quartz, plagioclase, and potash feldspar would be in equilibrium with them and should effect little alteration in country rocks composed mainly of the same minerals. Thin layers of schist at the Mill Race and Martin mines are rich in muscovite which apparently was formed largely at the expense of the feldspars; the biotite flakes in the schist were greatly coarsened. The largest muscovite crystals found in the wall rock of any pegmatite body were in a large block on the dump of the Mill Race mine. They are sievelike flakes as long as 4 inches and as wide as 1-1/2 inches that enclose small grains of biotite and quartz.

Sillimanite in schist near the pegmatite at the Gettys mine was altered largely to fine-grained sericite in very soft masses that preserve the original form of the sillimanite needles.

No alteration of the Toluca quartz monzonite by mica-bearing pegmatite has been noted although the Cherryville quartz monzonite at the E. R. Self mica mine has been extensively altered near such pegmatite. The pegmatite contact at that mine is very irregular, and in some places euhedral feldspar crystals 1 to 4 feet square replace the quartz monzonite as much as 5 feet from the nearest pegmatite. The nature of the feldspar could not be determined exactly. Inasmuch as the perthite in the pegmatite, though badly weathered, still contained gritty particles of microcline, the thorough weathering of the large crystals in the quartz monzonite suggests that they were plagioclase rather than potash feldspar.

The alteration of hornblende schist near its contact with spodumene-bearing pegmatite has been described by F. L. Hess and R. E. Stevens (1937). They found that biotite formed by the alteration of hornblende occurring within 1-1/2 feet of the pegmatite contained a relatively large amount of caesium and rubidium, which presumably were introduced from the pegmatite. The alteration of hornblende to mica in the walls of pegmatites is characteristic in the Spruce Pine district and elsewhere, but it is not a universal type of alteration in the Shelby district. It is unfortunate that no mica-bearing pegmatite has yet been found in unweathered hornblendic country rock in the Shelby district. The alteration of hornblendic rocks by spodumene-bearing or microcline-albite-quartz pegmatite has generally been slight. On the Foote Mineral Co. property south of Kings Mountain hornblende seems to have remained unaltered or little altered at the contact with pegmatite; glaucophane crystals in the schist actually are unaltered although occurring in tiny wisps of schist that were included in the pegmatite magma. An analysis of glaucophane obtained at a contact with pegmatite showed essentially the same content of lithium (about 3-1/2 percent Li_2O) as in a sample obtained at a greater distance, so there is little evidence that the mineral formed by alteration of the schist by the pegmatite.

Of the various wall-rock alterations that have been noted, only the complete alteration to muscovite has required much change in chemical composition. The formation of feldspar metacrysts, coarsening of biotite, and even the replacement of hornblende by biotite may not have required much addition of material, and in places the alteration probably consisted largely of the reconstitution of the original components of the rock. An introduction of alkalies seems to be the most common chemical change. The formation of large amounts of muscovite requires not only the introduction of alkalies and water, but the removal of moderately large amounts of iron and magnesium.

A little beryllium may have been introduced into the walls of one dike on the Foote Mineral Co. property. Two samples of schist at the contact contained 0.006 and 0.001 percent Be, whereas five samples of schist taken farther from the contact contained 0.0001 percent or less. A sample taken adjacent to a chloritic vein in the schist contained 0.00X percent Be and one taken an inch farther away contained only 0.000X percent. The vein filling contained only 0.000X percent. These results are similar to those of Thurston (Thurston, 1955, p. 55) who presented evidence for only slight enrichment of beryllium in granite near beryl-bearing pegmatites in the Crystal Mountain district in Colorado, but differ from those of Stoll (1945), who inferred that beryllium was introduced into the walls of pegmatite bodies in New England.

Tourmalinized schist and quartz-tourmaline veins are common in a belt just east of the tin-spodumene belt and at scattered places in the rest of the district. There is no strongly tourmalinized rock near pegmatite bodies and no evidence of a direct or close relationship between pegmatite formation and tourmalinization.

Origin of pegmatite

Inasmuch as few detailed studies were made of individual bodies of mica-bearing pegmatites during the recent investigation, little can be added to the discussion of paragenesis that was prepared in 1945 (Griffitts and Olson, 1953). There is a general similarity in the internal features of the mica deposits through the district, which suggests that all resulted from the same general geologic processes. With this in mind it was possible to work out an over-all paragenetic sequence and infer something about the origins of the mica deposits.

The manner of occurrence of the pegmatite dikes and the poor exposures of their wall rocks has made it impossible to demonstrate whether or not the walls were displaced by pegmatite. Many feet of actual contact between the pegmatite and wall rock were examined and, except at the E. R. Self mine, there was no evidence of replacement of wall rock by pegmatite and little evidence of reaction. The similarity between the mica-bearing pegmatites in the Shelby district and those in the Spruce Pine and other districts, in which there is clear evidence that the walls were displaced by the pegmatites suggests that an intrusive mode of emplacement for the dikes in the Shelby district is at least plausible.

Two dikes have offshoots from quartz cores that clearly cut the feldspathic pegmatite in the outer zones and one offshoot enters the wall rock, clearly indicating that the cores formed later than the outer zones. Additional evidence that the cores are younger than the feldspathic pegmatite is suggested by the comb structure formed by feldspar crystals facing quartz core segments. On the other hand, no evidence was found in any mica deposit that the material in the outer zones was younger than the cores. The preponderance of evidence thus indicates that the pegmatite dikes crystallized from the walls inward.

The only conclusive evidence for replacement of older minerals by younger minerals in the mica-bearing pegmatite was found in plagioclase grains in a few intermediate zones that were cut by veinlets of sericite or calcite or were replaced along crystallographic directions by sericite, and in masses of pyrrhotite that replaced feldspar in a few intermediate zones.

The general nature of the magma or solution from which pegmatites form and the processes by which the rock crystallizes have been speculated upon for many years. The interpretations cited below are those that seem to have greatest usefulness in studying the origin of the pegmatites in the Shelby district.

Pegmatite magmas have been considered to be similar to granitic magmas except that they contain more water and other mineralizers. A hydrous solution has reacted with the pegmatite in some places, to form replacement deposits. Few authors have indicated where the solutions came from, but Landes has inferred that they probably formed from the pegmatite magma and were not introduced from outside (1933, 1937). He also thought that the coarseness of crystallization of pegmatites was a result of the rather high water content of the original pegmatite magma. Zavaritski modified this interpretation somewhat, as he attributed the coarseness of pegmatites to the separation of a vapor phase that was in equilibrium with the coexisting magma and with the minerals that had already formed, and thought that the original magma was not notably different from those that formed granitic rocks. Zavaritski attributed the coarseness of the pegmatite minerals to recrystallization induced by the vapors during a metamorphic period during pegmatite formation. The first major alteration was the albitization of microcline; later the rarer minerals formed. (Cited by Saukov, 1953, p. 230-231.)

The interpretation favored by the writer for the origin of the pegmatites in the Shelby district has several points in common with those of Landes and Zavaritski. Like Landes, the writer believes that the common minerals, that is, the feldspars, quartz, muscovite, and beryl, formed large crystals by crystallization from a magma, and that in some places an aqueous solution exsolved from the pegmatite magma. Like Zavaritski, he believes that the aqueous solution and the magma coexisted for an appreciable interval during the consolidation of some dikes. In the Shelby district the solutions evidently were in equilibrium with the solids because there was no reaction between them. In general, the common major feature that might be attributable to the appearance of two fluid phases is the change from granitic-textured early pegmatite in the wall zones to blocky-textured later pegmatite in the intermediate zones.

The contrasting grain size of different minerals indicates either different processes of formation or different rates of seed formation and of crystal growth under the same conditions. There is no field evidence that the modes of origin of plagioclase and microcline are different. In the Kings Mountain area the spodumene crystals that have been bent, broken, and healed or cemented together must have been supported by some non-rigid material like a crystal mush. The matrix that encloses the coarse crystals is a fine-grained aggregate of albite and quartz, similar to some of the aplites in the area. At the time of deformation, prior to the complete solidification of the pegmatite, microcline, spodumene, and albite must therefore have all been present. It seems unlikely that they could have formed under markedly different conditions.

Tammann (1925) found, in cooling melts, that the rate of crystal nucleus formation reaches a maximum with undercooling and that the number of grains formed was affected greatly by the presence of suspended solid particles. Some foreign materials speeded and some slowed the rate of nucleus formation. He also found, though, that linear velocity of crystal growth passed through a maximum as the melts were undercooled. The temperature of this maximum is somewhat lower than the temperature at which the rate of nucleation is at a maximum. These effects might explain the development of small grains of albite and quartz during the growth of large grains of microcline and spodumene; at a certain temperature the rate of nucleus formation of albite and quartz might greatly exceed that of microcline and spodumene. The smaller number of grains of the latter minerals would grow to a larger size than the more abundant grains of albite and quartz unless albite and quartz greatly predominated in amount. The pegmatite magma undoubtedly picked up particles of the wall rocks through which it moved, which presumably would promote the crystallization of quartz, feldspar, and mica.

Wall and intermediate zones in some mica-rich dikes composed of the same minerals pose an interesting problem. The pegmatite in the outer or wall zone has the granitic texture found in most wall zones in the district whereas the pegmatite in the intermediate zone has a blocky texture resulting from the development of crystal faces on many grains of feldspar and muscovite. The vugs found in some of the intermediate zones are interstitial to feldspar grains that are not corroded. The cavities must have been filled with some fluid when the faces on the surrounding crystals grew to their final form and when calcite, sericite, tourmaline, and muscovite were deposited in some cavities. The fluid has since escaped from the cavity, but a silica-rich fluid did not escape from the pegmatite mass, but accumulated to form the quartz core. This suggests that the cavities were filled with water, carbon dioxide, and possibly other materials that are liquid or gaseous at low temperature and pressure. The presence of the cavities in pegmatite dikes that also contain quartz cores indicates that a silica-rich fluid was following the usual course of crystallization even though the water-rich fluid had separated; hence there must have been two fluid phases in equilibrium with the feldspar, quartz, and muscovite, and presumably with one another.

The marked similarity between some blocky intermediate zones without cavities and those with cavities suggests that the change in texture between wall and intermediate zones may be attributable to the development of a system with two fluid phases but that the water-rich fluid was able to escape from the exposed pegmatite, leaving no direct evidence that it was ever there.

Several ideas on the origin of the tin deposits in the tin- and spodumene-bearing pegmatites were proposed before 1893, but, as they were largely based upon grossly erroneous data, they need not be seriously considered. Titus Ulke (1894) first determined that the tin-bearing rock was pegmatite and first identified beryl and spodumene in it. Graton (1906) concurred with Ulke and discussed the deposits in more detail, considering them to be of strictly magmatic origin. Sterrett (1923) discussed the pegmatites of the area and considered them to be a result of aqueo-igneous intrusion, although he also believed that pegmatites could form through recrystallization of older rocks, or, once formed, could be modified by solutions from the same source as the original pegmatite magma. F. L. Hess (1936, 1940) developed in great detail his theory that the spodumene-bearing pegmatites were formed in large part by the replacement of older material, largely older pegmatite. Kesler (1942), like Hess, believed that there was a continued flow of solutions through the deposits, with a succession of materials deposited by replacement of country rock or pegmatite.

Hess believed that the pegmatites of the tin-spodumene belt developed through a long sequence of fracturing and replacement. The history of the deposits in schist was summarized as follows (1940, p. 951-952):

1. Formation of pegmatite in cracks and along foliation planes from solutions expelled from the Cherryville quartz monzonite.

2. Movement along the original cracks produced a layered structure in the pegmatite.

3. Replacement of early pegmatite by microcline.

4. Movement along the cracks crushed and broke the microcline.

5. Albite aplite formed by replacement and filling of fractures with the deposition of albite, quartz, spodumene, and muscovite. Microcline was partly replaced by spodumene, as was early pegmatite.

6. Formation of greisen and introduction of cassiterite.

7. A third interval of fracturing.

8. Introduction of large amounts of quartz, which replaced some spodumene and which may have been followed by amblygonite.

9. A fourth period of fracturing and vein filling by quartz and sphalerite.

Hess thought the history of the pegmatites in granite to be simpler than that of pegmatites in schist. He summarized it as follows:

"1. Cracks were formed in the granite, probably while the movement that made it gneissic was taking place, and

"2. The cracks were filled with white microcline (stage 3, of origin), evidentally from solution, because the granite was replaced at the same time, as shown by the sudden irregularities in the pegmatite and the rough frozen sides. These, apparently, are pegmatite bodies to which the term 'veins' can be properly applied.

"3. Another movement, in which the sides moved unevenly as the granite was given further gneissic structure, cracked the microcline diagonally and was accompanied by solutions that replaced the feldspar by spodumene, quartz, and a little muscovite, with less amblygonite and still less green apatite.

"4. Later cracks more nearly normal to the sides allowed quartz to replace microcline, spodumene, and mica and some of the apatite was carried into the granite 2 inches or more from the pegmatite."

The writer believes that Hess's primary interpretation of the mineral relations is questionable and that there is little evidence for a long-continued flow of solutions or a long series of mineral replacements. In addition the age sequence of the minerals seems to the present writer to be different from that inferred by Hess. There is nothing to indicate the nature of the original pegmatite postulated by Hess, unless it is either the albite aplite or the spodumene-free, mica-poor pegmatite of the present report. The relations between these rocks and the spodumene-bearing rocks are not well explained by replacement. There is little evidence that they were merely fractured and that microcline and spodumene were introduced. There is little indication that the microcline is generally older than spodumene, that either mineral has been deposited largely by replacement, or that either has been extensively replaced by other minerals. It is not clear why the irregularities in the pegmatite and rough frozen contacts can be taken to indicate that the pegmatite was deposited from solution or why one can infer extensive replacement of the "granite". The gneissic structure in the Cherryville quartz monzonite, which Hess thought was formed by renewed movement along the fracture seems better to be explained as drag associated with the formation of the original fracture, inasmuch as the pegmatite shows little sign of deformation.

If this is the case, the amount of replacement of the country rock must be thinner than the gneissic layers alongside the fractures, which generally are only a few inches thick.

Kesler (1942, p. 257) summarized his ideas on the origin of the tin-spodumene pegmatites very clearly. He thought that the order of deposition of minerals in the pegmatites was the same throughout the pegmatite belt, and listed the stages as follows:

Stage 1. Tourmaline, beryl, and apatite.

Stage 2. Cassiterite and columbite-tantalite.

Stage 3. Quartz (most of the greisen formed).

Stage 4. Spodumene.

Stage 5. Microcline.

Stage 6. Medium-grained albite (minor alteration of spodumene and microcline to muscovite).

Stage 7. Fine-grained albite and quartz (minor quantities of sulfides and apatite).

"The minerals of stages 1 and 2 developed in and adjacent to prominent joints and fissures forming deposits of three types: (a) Veins and lenticular masses of compact crystalline tourmaline in gneisses and schists; (b) lodes of disseminated tourmaline crystals in muscovite schists; (c) veins and lodes or disseminations of cassiterite, columbite-tantalite, and probably ilmenite in muscovite schist, between layers of muscovite schist and hornblende gneiss, and in joints that cut obliquely across hornblende-biotite gneisses."

These materials were fractured and quartz entered, being emplaced in part by replacement of mica schist, the mica of which was recrystallized to form the mica of greisen. Later materials were deposited largely through replacement of schist and filling of joints. The solutions from which the minerals were deposited changed in composition between stages but the composition was thought to be simple at all times.

The writer does not believe that a consistent age relation can be established between all the minerals, as was attempted by Kesler, as the early minerals continued to crystallize through much of the time of consolidation. Tourmaline seems to be a very persistent mineral and may have been the earliest mineral. It also seems to be a late mineral, contemporaneous with spodumene and microcline. In general, the early minerals in the spodumene-bearing pegmatites continued to crystallize until the end of pegmatite consolidation. For example, near Long Creek Church and Indian Creek, albite, spodumene, quartz, and muscovite were the early minerals, then microcline also began to solidify. The ratios changed somewhat, with a decrease in the albite content with time. Beryl seems to have formed in about the same amount throughout the period of pegmatite consolidation.

Only slight modification of Kesler's sequence of minerals is needed to make it conform to the writer's data. If the various stages are not considered to be distinct episodes, sharply separated from the others, and if albite and quartz are considered to crystallize through the entire interval of pegmatite formation, Kesler's stages indicate the order in which crystallization of additional minerals began.

The history of the pegmatites in the tin-spodumene belt is more complex than that of the mica-rich pegmatites. The pegmatites north of Kings Mountain were intruded into fractures in quartz monzonite or metamorphic rock and were relatively undisturbed during consolidation. Zones therefore formed as a result of orderly crystallization. South of Kings Mountain the pegmatites entered fractures along which movements took place as the pegmatite crystallized. In a few places (fig. 18B, for example) zones started to develop, only to be disrupted by shearing of the partly consolidated pegmatite. The spodumene and microcline crystals must have been surrounded by some material with enough strength to transmit stresses, otherwise they would not have been broken and the pieces of broken crystals would not have been suspended in their present positions. The surrounding material could not have been completely rigid, however, as it filled the cracks in the large grains. These conditions seem best satisfied in a crystal mush. As the amount of fluid decreased during crystallization the mush became more rigid, and movement was concentrated in the weakest material. The last shearing was along surfaces that cross the other layers in the dikes and presumably took place after the mass was solid.

Length of time of crystallization

It may be possible to estimate the approximate length of time needed for the complete crystallization of a pegmatite dike. The optimum rate of growth of synthetic phlogopite from a melt was said by Noda to be 0.1 to 1 mm per minute (Noda, 1955) and by Buckley to be about 2 mm per minute (1951, p. 91-92). A rate of about 1 mm per minute thus is reasonable. If we assume that muscovite growth rates are comparable, the mineral would grow at a rate of about 5 feet per day if dilution did not affect the rate. Von Pickardt found that the velocity of crystal growth of a substance crystallizing from a melt is lowered in proportion to the square root of the concentration of the diluting material and that equal molecular amounts of different dilutants have similar effects (cited by Partington, 1952, p. 520). If we assume that the rate of growth that gives largest crystals is similarly affected the decrease in velocity of growth might be tenfold as a maximum, reducing the rate of growth of muscovite books to about 0.5 foot per day. Inasmuch as most mica books are imbedded in feldspar and quartz, their minerals must have been growing so as to support the mica. If they grew at about the same rate, that is, near 1/2 foot per day, a dike 20 feet thick should crystallize in a few months. The rate of cooling and therefore of crystallization probably decreases during the consolidation, but the extension in the time from this cause probably would not be great. Vogt pointed out long ago (1923) that crystals of minerals may grow in the laboratory at rates of about 1 mm per minute or 1.4 M per day and that coarse-grained pegmatites might therefore crystallize in a short time. It now seems that a moderately rapid rate of growth may be necessary and not merely possible if large, unflawed crystals of mica are to form.

If values are assumed for the thermal conductivity and diffusivity of the wall rock, the pegmatite, and the pegmatite magma, and for the difference in temperature between the walls and the magma as well as for the temperatures at which crystallization began and was completed, we can calculate the time needed to reach the temperature at which crystallization is completed. Some of the assumed features are not critical such as the properties of the wall rock, but others are, so a rigorous mathematical treatment seems unjustified. A graphic solution is possible with the use of Lovering's charts (1935, 1936). These suggest that a few years would suffice for the crystallization of a dike 20 feet thick and that the time is about of the same order of magnitude with a wide range in the initial assumptions. The times obtained by the use of Lovering's charts do not include the time needed to dissipate the heat of crystallization. Even if this heat is half of the total heat in the original magma, doubling the time needed for complete consolidation, the estimate of "a few years" may still be reasonable. The significance of the estimates of time needed for complete crystallization lies in their refutation of the assumption that formation of a coarse-grained pegmatite requires extremely slow crystallization over an interval of hundreds or thousands of years. The "governor" that determines the cooling rate may be the maximum rate at which the heat of crystallization can be removed from the growing crystals.

Structural control of pegmatite

General

The pegmatite bodies in the Shelby district are parts of large regional belts. The interpretation of the mode of emplacement of the masses of pegmatite must accordingly take into account the regional setting of the districts. The three main types of pegmatite, those of the tin-spodumene belt, those of the sheet-mica areas, and those related to the Toluca quartz monzonite, were emplaced under somewhat different conditions and different controls. These will be discussed in the order of age of the pegmatites; that is, first, the controls affecting pegmatites related to the Toluca quartz monzonite, second, those of the pegmatites related to the Cherryville quartz monzonite.

Pegmatites related to the Toluca quartz monzonite

The bodies of pegmatite related to the Toluca quartz monzonite that are enclosed in the quartz monzonite are in part irregular layers that seem not to be controlled by any strong structure and in part clearly defined dikes that are controlled by joints and faults. The nearly vertical irregular bodies in some gently-dipping sills of quartz monzonite may have accumulated in tensional zones in the nearly crystallized quartz monzonite. The dikes in fractures in the quartz monzonite probably are not basically different from dikes that are in the Carolina gneiss. They must have been emplaced after the quartz monzonite had solidified and after cessation of all slipping along the leaves of quartz in the quartz monzonite.

The pegmatites related to the Toluca quartz monzonite form lenses and sills that are parallel to the foliation of the Carolina gneiss, and also clearly defined dikes that cut the foliation. The pegmatite along the foliation planes probably formed while the deformation that accompanied the metamorphism was active; it entered the less competent schists preferentially. Some of the pegmatite is more conspicuously deformed than the quartz monzonite of nearby sills. This might indicate that the pegmatite is somewhat older than the quartz monzonite and therefore was subjected to shearing during a longer time, or perhaps the effects of deformation are merely more noticeable in coarser-grained rocks.

Dikes of pegmatite were emplaced along fractures in the Carolina gneiss that generally have a northerly or northeasterly trend. The fractures were formed after the gneiss was injected by the concordant lenses and sills of pegmatite. Drag folds in the gneiss walls of many fractures indicate that they are normal faults. Only a few of these faults had large enough displacement to affect the patterns of outcrops on the geologic map. In about three-fourths of the faults examined in the central part of the Shelby quadrangle the western wall of most faults was upthrown. None of the pegmatite dikes was found to be in a thrust fault, even though some are very close to the inferred position of the fault. The thrust faults probably are older than the pegmatite but were too tight to be intruded by the magma.

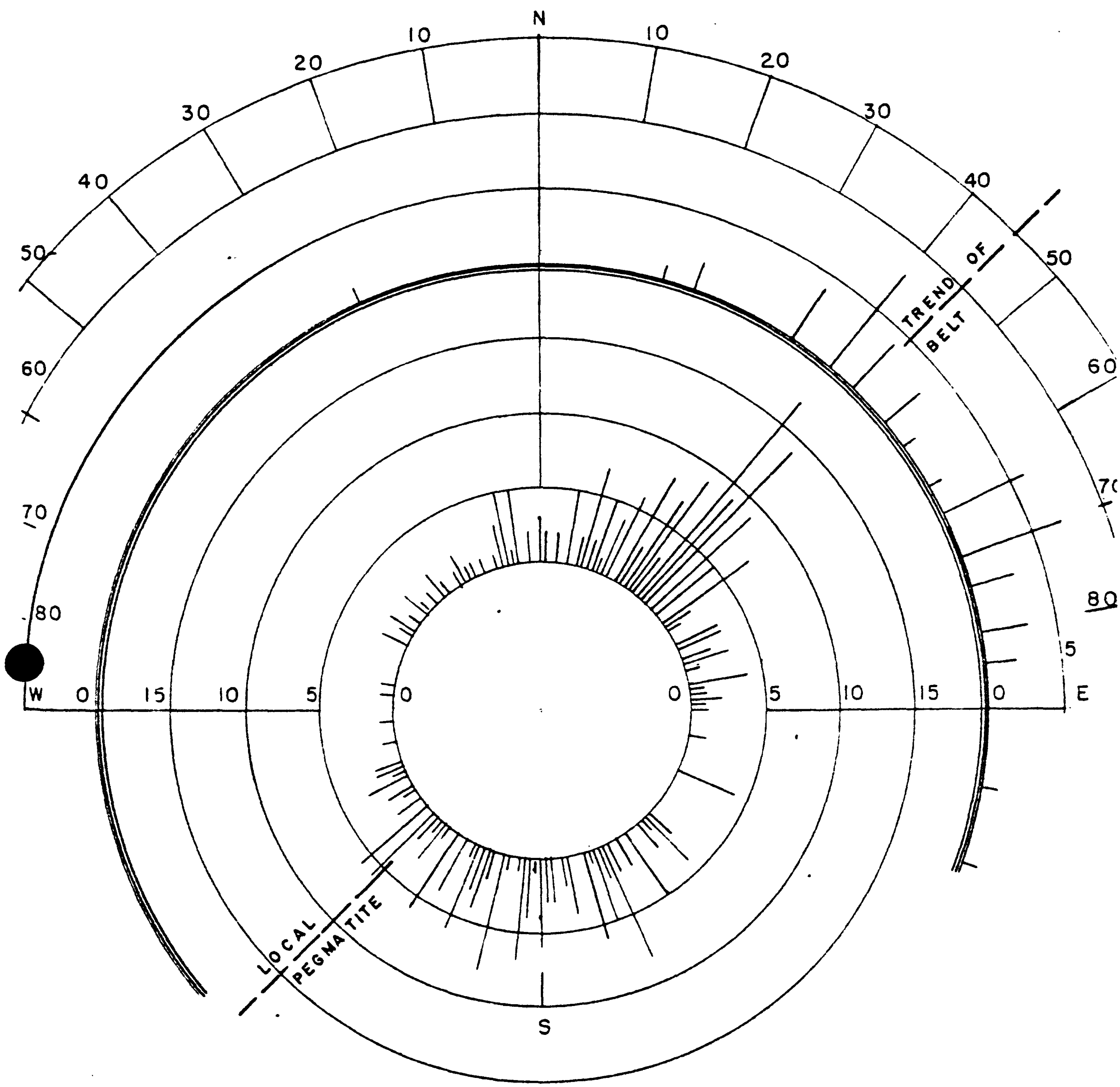
In summary, pegmatite related to the Toluca quartz monzonite was emplaced along foliation planes of the biotite and sillimanite schists while important interlayer slipping and thrust faulting was taking place. The pegmatite was introduced in largest amounts in the least competent rocks, in which the slipping and plastic deformation were strongest. Later, after the interlayer slipping, there was a period of relaxation, with the formation of fractures and small normal faults, into which additional pegmatite magma was intruded. The Toluca quartz monzonite was emplaced before or after the conformable bodies of pegmatite. The shape, size, and attitude of the bodies and the texture of the pegmatite are thus related to the stage of the orogeny at which intrusion took place.

Pegmatites related to the Cherryville quartz monzonite

The pegmatites related to the Cherryville quartz monzonite were emplaced in abundance in three general places: within the batholith of quartz monzonite, in a broad belt of sheet-mica-bearing pegmatites 5 to 10 miles west of the batholith, and in and alongside the tin-spodumene belt east of the batholith. The tin-spodumene belt and scrap-mica pegmatites just west of it are near the margins of the batholith and might, therefore, be classed as "marginal" pegmatites, following the classification of Gevers (1936). The bodies within the batholith would then be considered to be internal pegmatites and the mica-bearing pegmatites west of the batholith would be exterior pegmatites. It is not at all certain that this classification reflects the origin of the pegmatites. The tin-spodumene zone as a whole is not actually marginal to the batholith but forms a curved belt that is alongside the main batholith south of Kings Mountain but is between the main batholith and a series of stocks farther north. The belt of scrap mica pegmatites west of it is largely in the batholith south of Kings Mountain and in the metamorphic rocks farther north. The tin-spodumene belt is a little shorter than the batholith, especially if one disregards two isolated cassiterite deposits near Gaffney; the scrap mica belt is more nearly coextensive with the batholith. This similarity in extent suggests that the batholith has influenced the over-all distribution of the pegmatites. The western pegmatite belt is actually part of a series of pegmatite districts in a belt that can be traced from central Georgia to Virginia, and perhaps also as far north as Maryland. The localization of pegmatites in it therefore cannot be attributed entirely to local factors. The pegmatites in the tin-spodumene belt and the scrap mica belt probably were similarly influenced by both regional and local features. The mica deposits in the Shelby district are concentrated in places where the northeasterly-trending regional belt of pegmatites crosses areas where the metamorphic rocks trend northwesterly.

The pegmatite magmas in the tin-spodumene belt were guided in their emplacement by a fracture zone that extends from the vicinity of Gaffney, South Carolina, to the Catawba River west of Statesville, North Carolina, a total length of 60 miles. It may extend 15 miles farther south to Whitestone, South Carolina, and 55 miles farther north to Winston-Salem, North Carolina. Whether the doubtful extensions are included or not it is evident that the belt is important in the regional geology and not merely a local feature. The width of the structural zone is rather uncertain, though the pegmatite belt in it is about a mile wide at Gaffney, South Carolina, and several miles wide near Statesville, North Carolina. The uncertainty of the width of the belt reflects, to some extent, its history. The type of pegmatite within the tin-spodumene belt varies in an east-west direction. The older pegmatites were emplaced on the western side and the younger on the eastern side which naturally indicates that fracturing progresses from the western side to the eastern side of the pegmatite belt.

The most common trends of pegmatite dikes in the Beaverdam Creek area, as is shown in figure 21, are near N. 40° E. and N. 20° W. This pattern is that of a conjugate fracture system. Plotting the trends of the younger spodumene-bearing pegmatite dikes separately from those of the older spodumene-free dikes shows that the trends of the two types are about the same but that the concentration of the older dikes in preferred directions is not as pronounced as the concentration of trends of the younger dikes. A third favored trend (N. to N. 10° E.) shown by the older or spodumene-free dikes coincides in direction with few spodumene-bearing dikes.



Upper semicircle, spodumene-bearing dikes.

Lower semicircle, spodumene-free dikes.

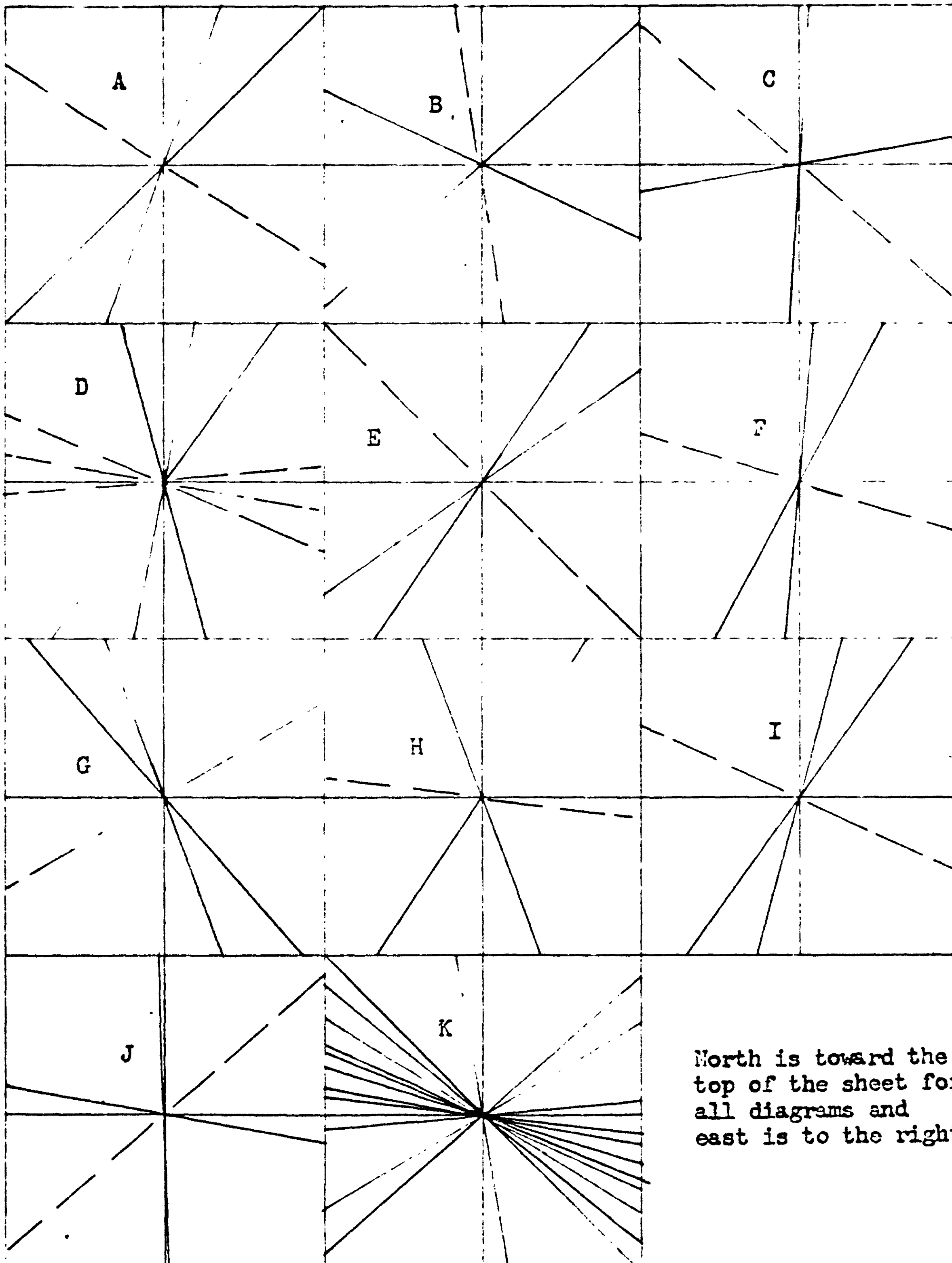
Outer diagram, foliation.

Figure 21. Trends of pegmatite dikes and of foliation of schist. Beaverdam Creek area.

Sharply bent pegmatite dikes are thought to indicate pairs of fractures that were open at the same time and might therefore be used in determining the origin of the fracture pattern. Figure 22 shows the trends of the limbs of ten bent dikes in the Beaverdam Creek area and the bisectrices of the obtuse angles between limbs. Diagram D on that figure shows the trend of the limbs of a dike that is sharply bent in two places and diagram K is a composite that shows the trends of all the obtuse bisectrices determined for bent dikes in the Beaverdam Creek area as well as the average direction and the trend of this part of the pegmatite belt. The bisectrices are concentrated largely in an east to S. 45° E. direction, transverse to the pegmatite belt.

The emplacement of dikes most commonly along fractures whose trends subtend an angle of about 60° suggests a conjugate fracture system as already mentioned. The dikes that are not in one or the other of the fracture sets might be attributed to the effects of mechanical heterogeneity or anisotrophism in the rocks or, at least in part, to subsidiary fractures formed through the action of the same stresses that formed the more important fractures.

Conjugate fractures have been attributed either to compression, generally parallel to the bisector of the obtuse angle between fractures, or to shearing. The fracturing that preceded and accompanied the emplacement of the pegmatite in the tin-spodumene belt may be attributed, accordingly, to either compression or to shear. If the fractures formed by compression there should be moderately open shear fractures in two directions, oblique to and forming the same angles with the direction of compression. These fractures should be readily injected by pegmatite magmas. Tension fractures parallel to the direction of



North is toward the top of the sheet for all diagrams and east is to the right.

A–J. Solid lines show trend of segments of pegmatite dikes. Dashed lines show trend of obtuse bisectrices of angle between segments.

K. Composite diagram showing all obtuse bisectrices.

Figure 22, Trends of bent dikes, Beaverdam Creek area.

compression would probably be more open than, and shorter than, the shear fractures; these should be susceptible to injection but might be expected to enclose short, thick dikes. Thrust faults may develop normal to the direction of compression. As the walls of these fractures would be held tightly together they should contain relatively little injected material; such thrust faults are notoriously unfavorable for the entrance of mineral-bearing solutions.

In the Beaverdam Creek area, the compression would presumably be in a N. 80° W. - S. 80° E. direction, parallel to the obtuse bisectrix of the two most common sets of fractures. Accordingly, one would expect to find steeply-dipping pegmatite-filled tension crqcks parallel to that direction and also to find few if any dikes normal to that direction; that is, with a N. 10° E. trend. Actually, as can be seen on figure 21, dikes with N. 80° W. trend are very scarce, only one having been found in the area. Dikes parallel to the possible N. 10° E. trending thrust faults ought to be very rare. Only five pegmatite dikes have been found with trends near N. 10° E.

If the fracturing is attributed to horizontal shearing, the total movement must have been rather small, permitting the rather weak deformation to be dominated by local features. It would be difficult to devise a uniform homogeneous movement throughout the belt that would cause the opening of fractures in the narrow belt south of Kings Mountain and the development of the fracture patterns farther north, opening the different sets at slightly different times to permit intrusion of different types of pegmatite. It is possible that the fracturing that now can be seen is a minor feature overlying a major zone of movement at depth.

Summary

The migration of pegmatite magmas in this district were determined by two main factors: the competency of the enclosing rocks and the movements in the rocks. These in turn were determined by the nature of the orogeny associated with the pegmatite formation and the time during the orogeny at which the pegmatites were emplaced. The pegmatites that formed early, during the orogeny accompanied by high-grade regional metamorphism and intrusion entered along foliation planes in incompetent rocks along which slipping was taking place. They therefore are now recognizable as pinching and swelling sills and conformable lenses. Pegmatites emplaced slightly later during this orogeny entered more competent rocks that tore and fractured, giving rise to dikes and large, irregular, discordant lenses as well as small discordant lenses. The orogeny that accompanied the emplacement of the Cherryville quartz monzonite did not have the effect on the metamorphic rocks that the earlier orogeny had. At no time in the later orogeny did the rocks of the Shelby district appear to have lost their competency. Therefore the deformation consisted mainly of fracturing, with only slight drag along the walls, and with movement of relatively large blocks of rock in which there was little internal deformation. The pegmatites related to the Cherryville quartz monzonite entered about at this time and found sharply defined fractures and a few breccia zones. The movements along these fractures generally had ceased before the masses of pegmatite magma in them began to crystallize, so the final product was a group of dikes and discordant lenses of non-gneissic pegmatite, with a few clusters of gneissic pegmatite dikes.

Origin of zones

General

The Shelby district contains both zoned and unzoned bodies of plagioclase-microcline-quartz pegmatite that have been the same general range in composition, as well as both zoned and unzoned bodies of spodumene-bearing pegmatite with a distinctly different composition. This indicated that the composition of the parent magma could not have been the dominant factor in determining whether zones developed. An attempt is therefore made below to determine what environmental, as well as compositional factors, influenced zone formation. Naturally, the attaching of the names "Toluca" and "Cherryville" to certain pegmatites has no effect on their origins, so the discussion will consider the characteristics and environments of various types of pegmatite, disregarding the formal names applied to them. The factors considered are:

1. Composition.
2. Temperature.
3. Deformation.
4. Size, shape, and attitude.

The pegmatites of the Shelby district have zones characterized by distinctive textures and by differences in composition. In the mica-bearing pegmatites there are two distinctive textural changes independent of compositional changes that mark zone boundaries. The first is a change from elongate quartz and plagioclase grains and flat mica flakes, all oriented nearly normal to the wall in the border zone, to more nearly equidimentional quartz and plagioclase and randomly-oriented mica books in the wall zone. The second change is from compact plagioclase-quartz-muscovite pegmatite with a granitic texture to vuggy plagioclase-quartz-muscovite pegmatite with a blocky texture in the intermediate zone. The dikes containing vuggy pegmatite in intermediate zones also have well developed border zones, though similar border zones are found in dikes that do not contain vuggy pegmatite. Both of these textural changes are best developed in relatively short, thick dikes.

Zones defined by compositional changes show either (1) an increase, generally abrupt, in micorcline content as the center of the dike is approached, or (2) a sharply defined core of massive quartz. The core may be accompanied by a microcline-bearing intermediate zone. Most of the longer dikes and many of the shorter dikes in the district contain quartz cores, with or without microcline-bearing intermediate zones.

The development of zones from a body of magma may largely be determined by the distribution of nuclei about which crystallization might start. In the poorly zoned bodies, the crystals began to form at the same time throughout the mass, interfering with one another so as to form a mass of relatively small interlocking crystals. The minerals that formed later were obliged to crystallize interstitially to the early formed minerals. In a body of magma that was completely devoid of nuclei, an unlikely situation, the material that reached saturation first would crystallize against the walls of the magma chamber. The texture of the rock formed by such a process would resemble that found near the walls of the border and wall zones of many mica-bearing pegmatites in which mica flakes form a fringe normal to the walls, and grains of quartz and of plagioclase are oriented normal to the walls. These elongate grains become larger away from the walls and become fewer in number as the faster-growing grains engulf the slower growing grains, shutting them off from the magma. Farther away from the walls the degree of orientation of the crystals becomes much less. Similar relations may be found in some spodumene-bearing pegmatites north of Kings Mountain. Along the walls of these the long c-axes of spodumene crystals tend to be normal to the contact. Farther from the walls the grains, though larger, are oriented at random. These spodumene-bearing pegmatites differ from the mica-bearing pegmatites in that the entire body of pegmatite generally consists of the same group of minerals, in proportions that differ only moderately from place to place. In some of the mica-bearing pegmatites, the minerals formed later may be the same as those formed against the wall; in others they may be markedly different as in the dikes that contain microcline-rich intermediate zones or quartz cores.

Relation of zones to composition of pegmatite

The early-formed material in the outer parts of mica-bearing pegmatite dikes is strikingly similar throughout the district. The material in the inner parts varies greatly. In pegmatite magmas low in quartz and microcline, the solid body may be homogeneous throughout. If the content of quartz were abnormal an excess of that mineral remained to form a core after crystallization of substantial amounts of granitic-textured plagioclase-quartz rock. In most pegmatites, microcline, if present, crystallized with the "left overs" after most of the plagioclase had crystallized. The transition from plagioclase and quartz crystallization (plagioclase-quartz stage) to microcline or quartz crystallization (microcline, microcline-quartz, or quartz stage) generally was abrupt, but in a few places the microcline crystallized along with plagioclase and quartz. The reason for this difference is not known though it might be attributed to differences in vapor pressure; it does not appear to lie in the relative proportions of microcline and plagioclase. Thus, the development of zones in the mica-bearing pegmatite was dependent upon the bulk composition of the magma to the extent that zones developed most readily where there was an excess of microcline or an excess of quartz over that in the plagioclase-quartz pegmatite. In a few dikes a zoning was developed even though the microcline and quartz contents were low. These dikes, of which those at the Bowen and Nigara mines are examples, commonly contain wall and intermediate zones composed of plagioclase, quartz, and muscovite. A little biotite may lie near the wall and a little microcline may be in the intermediate zone, without obvious effect on the other features of the rock. The pegmatite in the wall zones of these bodies is the common type of granitic-textured plagioclase-quartz pegmatite whereas the pegmatite in the intermediate zones is blocky in texture and contains vugs, which may constitute several percent of the rock, and the muscovite tends to be greener, and in small books with poorer cleavage than in the wall zones. The vugs show no evidence of an origin as solution cavities but are generally outlined by uncorroded crystal faces on the surrounding feldspar grains and may contain hexagonal crystals of muscovite and masses of fine-grained sericite, or carbonate or sulfide minerals. The quartz content of these pegmatites seems not to have been particularly important as it determined mainly the size of the quartz core that might develop in the later stages of consolidation. The presence of vugs in the pegmatite that formed later and the presence of minerals that contain "mineralizers" in the vugs suggest that water and other hyperfusible materials may have formed discrete bodies in the crystallizing material and that a crystalline phase coexisted with two fluid phases, one a hydrous solution and the other a siliceous solution. At least the vugs must have been filled with some fluid that has escaped leaving no trace, and the silicate minerals in the core must have formed from a fluid that existed at the same time as the vugs. The boundary between the wall zone and the intermediate zone might then represent the place at which the residual fluid became saturated with mineralizers. If the mineralizer was largely water, the content at that point and the amount in solution thereafter probably did not exceed 10 percent (Goranson, 1931). Enhanced fluidity of the magma after it became water saturated might be responsible for the greater tendency for the feldspar and muscovite to form euhedral crystals.

Some bodies of pegmatite do not contain vugs but are otherwise similar to those that do. If the development of the vuggy pegmatite can be taken to show the presence of a hydrous fluid phase along with the siliceous magma it seems likely that the bodies of compact pegmatite with sizes, shapes, and zones similar to the vuggy ones also formed from a two-fluid-phase system. The absence of vugs from the pegmatites would be of little direct evidence if the aqueous solution were not trapped as bubbles between feldspar and mica crystals but escaped to higher levels that have been eroded. The pyritic quartz-garnet-chlorite vein at the A. F. Hoyle mine might represent the filling of a channel by such a solution. It might also be possible that water vapor could escape from a congealing pegmatite magma, the water boiling off as it exsolved from the magma.

Relation of zones to temperature

The development of zones with different mineral composition by fractional crystallization requires that different minerals begin to crystallize at different times and, in some types of zoning, that the precipitation of one mineral be essentially completed while others are still largely in solution. The crystallization of any mineral requires that the solution be saturated and that there be nuclei about which crystal growth may start. If nuclei are absent or are very scarce, the solution may become supersaturated before crystallization starts. In a magma, a scarcity of nuclei probably would result from superheating, which might be brought about by adding heat to raise the temperature of the magma or by adding heated water or other mineralizer to reduce the degree of saturation or the melting temperature. A source of mineralizing solution would be easier to find geologically than a source of heat.

It is inferred that crystallization of the pegmatite magma in the Shelby district was brought about primarily by a decrease in temperature, and that zones might be developed best where the number of nuclei was small at the time crystallization begins. This would cause the early minerals to form against the walls of the dike, where the magma would be coolest and where the mineral grains in the wall rocks would provide nuclei upon which new minerals could grow. A scarcity of nuclei might lead to a certain amount of supersaturation of the later minerals, thus delaying the start of their crystallization until most or all of the earlier minerals had formed, thereby causing the zone boundary to be sharp. An abundance of nuclei and enough internal movement in the magma to keep a steep thermal gradient from developing within the magma along the walls would presumably result in the formation of unzoned or poorly zoned dikes.

The comb structure formed along the walls of the pegmatite dikes can be explained more readily than other features of zones. It has been pointed out that the crystal axis along which the heat conductivity is the greatest tends to be oriented parallel to the lines of heat-flow from a cooling melt, that is, normal to the cooling surface (Buckley, 1951, p. 273 and elsewhere). The effectiveness of this mechanism of orientation should be greater where the thermal gradient of the system and the thermal anisotropism of the crystal are greatest. The thermal conductivity of quartz is greatest parallel to the c-axes and the difference in conductivity between the c- and a-axes decreases with increasing temperature (Birch, 1942, p. 248-249). It might therefore be expected that the comb structure would not develop where either the magma or the wall rock was exceptionally hot. The thermal gradient outward from the body of magma would be greatest immediately after intrusion and would gradually decrease as the wall rocks became warmer and the magma became cooler. The temperature of the wall rock at the contact would increase rapidly to a maximum, where it would remain until the center of the dike began to cool; then it would decrease as the magma cooled. A preferred orientation might develop best in crystals that formed against the wall and immediately after the intrusion of the magma. If the start of crystallization were delayed, the temperature of the rock at and near the contact would rise. As a result the thermal anisotropism of the crystals would be reduced and, at the same time, the thermal gradient would be lowered; both effects would reduce the degree of preferred orientation. This may have taken place in some of the dikes that have no notable orientation of the grains in the border and wall zones. These dikes are generally larger than those with well developed border zones.

The rate of cooling is very important in determining both the orientation and the size of crystals. This has been found by several groups who have investigated the growth of synthetic phlogopite crystals in crucibles that were cooled from the bottom or side. Noda (1955) found that the best phlogopite crystals were obtained when the rate of growth was about 0.1 to 1 mm per minute and that the probability of nucleus formation increased rapidly with supercooling. Noda believes that "The rate of crystal growth depends upon the rate of removal of heat of crystallization from the boundary of the crystal-melt through the crystallized mass." Buckley has summarized the results of phlogopite synthesis, mainly in Europe, in a form that suggests applications to geology (1951, p. 91-92). "Cooling from below is kept within the crystallization velocity of the mica, which has been found to be about 2 millimeters per minute. To obtain as large crystals as possible and to insure their being oriented vertically, the temperature gradient should correspond very closely with the ascertained crystallization velocity. If the cooling rate exceeds this, new crystals form ahead of the main body and, if it is much less the mica is found to separate into small flakes. The latter danger is greater because of the heat of crystallization liberated." The effect of the interplay of these factors in the formation of mica from a melt was strikingly illustrated in the Electrotechnical Laboratory of the U. S. Bureau of Mines at Norris, Tennessee. A 3-ton batch of synthetic phlogopite was fused, then allowed to cool. The mica crystallized first in flakes oriented normal to the walls of the container, then, after a "border zone" several inches thick had formed, the flakes were oriented at random, reproducing very closely the conditions found in the muscovite deposits in the Shelby district. Presumably the growth of large muscovite books in the pegmatites, like the growth of large phlogopite crystals in crucibles, required a cooling rate within a narrow range and that the change from oriented to disoriented crystals during consolidation has resulted from a rate of cooling that was in the higher part of the permissible range. In this range mica flakes began to form ahead of the crystals growing out from the walls but not in large enough numbers to cause the size and mechanical firmness of the crystals to be reduced greatly by mutual interference during growth. The small size of the mica flakes in the Cherryville quartz monzonite and in the early pegmatites formed in the main batholith of quartz monzonite may be attributed to a separation of too many crystals caused by too slow a cooling rate. The inferred narrowness of the range in cooling rate in which large mica books can form probably accounts for the scarcity of sheet mica deposits and for the occurrence of groups of such deposits in districts or parts of pegmatite belts in which the relative temperatures of wall rock and magma were in the favorable range. The preservation of the useful mica books depended upon tectonic quiescence after the pegmatite had formed.

The temperature gradient at the contacts of the dike is determined by the temperatures of the magma and of the wall rock at the time of intrusion. If the temperature of the wall were too low the margins of the magma body would be cooled too rapidly and wholesale precipitation of all the materials in the magma might be induced. Later, after the remaining magma was separated from the cold walls by a layer of newly-formed igneous rock, the gradient and the rate of crystallization might be low enough to permit zoned pegmatite to form in the usual manner. Naturally, only a moderately thick mass of magma would contain enough heat to maintain a sufficient temperature in its inner part to permit pegmatite to form after the heat loss incident to the chilling by the walls. To keep the rate of crystallization in the favorable range, not only for the formation of large muscovite books but also for the formation of zones, the wall rocks must be moderately warm, probably about at the temperature represented by the lower amphibolite or epidote-amphibolite facies of metamorphism which seems to be a common environment for the formation of mica-bearing pegmatites.

The pegmatites related to the Toluca quartz monzonite were emplaced shortly after the main period of metamorphism and while the rocks were still at or near the temperatures that were responsible for the metamorphism of the rocks to an upper amphibolite or lower granulite level. This would tend to inhibit development of zones because the magmas probably were not much warmer than the schists. The thermal gradient at the contact thus would be low and the high temperature would reduce the thermal anisotropism of the crystals. The very slow rate at which the rocks of the region cooled would be less than of the growth rate of many minerals, hence mica books and perhaps other minerals would separate into smaller crystals. In addition, the tectonic agitation of these pegmatites during consolidation would mix the fluid constituents, preventing the concentration of residual fluids in the center of the dike and keeping the temperature of the whole mass relatively even so that nucleation and crystallization could proceed throughout the mass.

Because of the difference in host rock temperature and history of deformation between the pegmatites related to the Toluca quartz monzonite and those related to the Cherryville quartz monzonite it is not surprising that the former are unzoned and the latter are zoned, even though the bulk compositions are quite similar.

The same factors may determine whether or not zones develop in other pegmatites in this district. The pegmatites of the tin-spodumene belt are poorly zoned. Some dikes north of Kings Mountain have outer zones that contain spodumene crystals with their c-axes about normal to the walls and inner zones in which the crystals are oriented at random. The thermal properties of pyroxenes and amphiboles are not well known. Hornblende is anisotropic (Birch, 1942, p. 250), and Wooster has stated that crystals with chain structures, like the pyroxenes, conduct heat most readily parallel to the chains (1949, p. 84). Spodumene, as a pyroxene, contains chains parallel to the c-axis, hence its thermal conductivity is greatest along that direction. The c-axis should be oriented normal to the cooling surface or wall of a magma chamber. The variations in degree of orientation of spodumene crystals north of Kings Mountain are very similar to those of mica books in the mica-bearing pegmatites. Spodumene-bearing pegmatite, very similar in composition to that in the tin-spodumene belt, is normally found in intermediate zones in complexly zoned pegmatites in the Black Hills (Page and others, 1953) where it formed later than the plagioclase-quartz-muscovite pegmatite in the wall zones. The age sequence is thus similar to that in the Shelby district, even though the types of pegmatite are here segregated into separate bodies. It is reasonable to assume that the temperature declined during the consolidation of the complexly zoned Black Hills pegmatites and also during the formation of the unzoned pegmatites in the Shelby district. It may be that good development of zones in a pegmatite magma of spodumene-albite-microcline-quartz composition requires a somewhat different cooling rate than the magmas that yield plagioclase-muscovite-quartz pegmatite. For the cooling rate of the spodumene-bearing pegmatites to be the same, the temperature of the wall rocks would have to be somewhat cooler than alongside the hotter magma of the mica-mining areas.

South of Kings Mountain a few remnants of undeformed pegmatite have been found along the walls of spodumene-bearing pegmatite dikes but many pegmatite magmas in this area, like the pegmatite magmas related to the Toluca quartz monzonite, were deformed during crystallization which disrupted the normal formation of zones.

It is evident from the above discussion that zones are likely to form where the thermal gradient from the pegmatite magma to the wall rock is steep, and that the actual temperature should be below that at which the thermal anisotropism of the crystals is greatly reduced. This suggests that well-zoned pegmatites should most commonly form outside of areas in which extremely deep seated or very high-grade regional metamorphism was taking place. Such zoned pegmatites have been found in many places in rocks that reached a lower amphibolite facies of metamorphism at or shortly before the time of pegmatite emplacement, though they apparently are not particularly common in areas of granulite facies of metamorphism. It is not everywhere easy to determine the conditions under which the pegmatite was emplaced. In the Shelby district most mica-bearing pegmatites were emplaced in biotite-garnet-sillimanite schists and gneisses. These schists had formed long before the pegmatites were intruded, but were not appreciably altered to lower-grade rocks at the time of pegmatite formation except adjacent to the pegmatite dikes. If the geologic history were less well known the environment of pegmatite crystallization probably would have erroneously been assumed to be near that of the metamorphism.

Relation of zones to deformation of pegmatite

Zones have not been developed in pegmatite that was deformed during consolidation. It is unlikely that this could be attributed to the development of zones early in the period of deformation and their destruction later by continued or resumed movement because the deformation does not seem to have been particularly severe. It probably would not have been able to disrupt the relatively massive bodies of microcline and quartz found in the well zoned masses. In addition, at least some of the pegmatites related to the Toluca quartz monzonite were emplaced rather late in the main period of deformation or even during the last rather feeble movement.

The deformation probably prevented the formation of zones during consolidation of the pegmatite. Several mechanisms can be suggested by which such deformation could prevent the formation of zones:

1. Movement would cause a mixing of the solid and fluid materials in the pegmatite, reducing the formation of either concentration gradients or temperature gradients in the pegmatite material near the walls.
2. Movement presumably would break or tear loose particles of minerals in wall rocks, which could provide abundant nuclei about which crystallization might take place. Thus

there would be less tendency for the early minerals to begin to form along the contacts where the undisturbed wall rock provided nuclei.

3. Any mineral grains that might form against the walls would likely be broken loose and mixed with those that formed within the magma body.

Some evidence that the third mechanism was active was found in Kings Mountain area. Spodumene crystals extend into the dike from east wall; they curve downward and are somewhat broken. The fractures are filled with other minerals, mainly albite and quartz, with a little microcline. The inner edge of the mass that contained the distorted crystals had been sheared off and the rest of the pegmatite was strongly deformed and contained broken crystals of spodumene and deformed crystals of microcline. The orientation of the elongated fragments of spodumene crystals indicates that the movements were in the same direction as those that caused the bending and fracturing of the crystals along the walls. The movements continued until the pegmatite was nearly or entirely solid as some layers of deformed pegmatite cross the layers of the older gneissic pegmatite and seem to be shear zones that were open while there was still a little pegmatite magma left to enter them. Thus, only locally did the zones form in the dike in spite of the shearing that took place during the crystallization of the magma.

Relation of zones to size, shape, and attitude of pegmatite mass

The size, shape, and attitude of the pegmatite body were thought by Vlassov to be very important in determining the differentiation material and the rate of crystallization of the minerals (referred to by Saukov, 1955, p. 223). These factors seem not to have been dominant in the Shelby district. Well-zoned dikes of pegmatite related to the Cherryville quartz monzonite range in thickness from several inches to over 100 feet. Similarly, poorly zoned dikes in the tin-spodumene belt and dikes related to the Toluca quartz monzonite range in thickness from less than a foot to nearly 300 feet.

The shapes of the pegmatite dikes in the Shelby district do not vary greatly. Most dikes are more or less elongate lenses. The zoning is quite similar in short and long lenses. The main difference is that only the shorter lenses contain vuggy intermediate zones or intermediate zones consisting of the same minerals as the outer zones. The shape, like the size, may exert a modifying influence but not a dominant influence on the zoning. This is in complete agreement with Heinrich's conclusion that "pegmatite shape and degree of development of internal structure appear to be independent features" (1953).

The vertical and nearly vertical bodies of pegmatite commonly are symmetrically zoned; the gently-dipping dikes are rarely symmetrical. This is more prominent in other districts in the Southeast as very few dikes in the Shelby district dip gently. In the gently-dipping part of the Bess dike, the muscovite content was much higher along the hanging wall than along the footwall. This is a common feature in the Spruce Pine district where many bodies of pegmatite dip gently. The concentration of muscovite along the crests of steeply-plunging dikes in the Shelby district might be related to the hanging wall concentrations in the Bess mine and Spruce Pine district. An upward movement of water dissolved in the magma may account for both types of concentration of mica in upper parts of pegmatite bodies.

Mineralogy

General

The emphasis in the discussion is on the minerals in the pegmatites and on minerals in the enclosing rocks that may be useful in determining the environment of pegmatite formation. Most of the minerals that have been found in the district are shown on table 4. Many additional ore minerals and many supergene minerals have been found but they have been omitted because they have little to do with the pegmatite geology.

Quartz

Quartz is a very common and widespread mineral district as:

1. Massive coarse-grained quartz.
2. Sugary, vitreous, gray to white or nearly transparent, fine-grained quartz.
3. Clear to slightly smoky or color zoned crystals.

Quartz in coarse-grained aggregates is most common in the cores of sheet-mica pegmatites west of the Cherryville batholith. The grains are from 1/4 inch or less to 7 inches in breadth. The average is probably 1 inch or larger. The color ranges from transparent and colorless through translucent white to translucent gray or very lightly smoky.

Rarely the quartz is almost clear and colorless. The luster is most commonly glassy or vitreous but some is oily or greasy. The quartz from most productive mica mines tends to be glassy rather than greasy. Similar coarse-grained quartz forms veins in the Carolina gneiss in the Shelby quadrangle.

The quartz in some cores of mica-bearing pegmatites has been subjected to enough strain to cause an undulose extinction that is visible microscopically. Moderately clear and less milky quartz is crossed by white layers of fine-grained quartz that are rarely as thick as 1/20 inch. Most commonly they form lines 0.01 to 0.03 inch wide on the surfaces of quartz fragments (fig. 19B). The smooth conchoidal fracture of the fragments is offset or roughened by these layers, which are apparently healed fractures. In extreme cases, white quartz that appeared in hand specimens to be coarse-grained had actually been broken into many tiny grains, each of which had been rotated slightly.

Sugary quartz is characteristic of the pegmatites in the tin-spodumene belt and of the quartz veins in that area and within a mile or two west of it. The average grain size of the quartz is near 1/20 inch, and the maximum is little more than an inch. The uncommon coarse megagrains are set in a matrix of the fine-grained quartz in a porphyry-like texture. The luster varies from vitreous to greasy and dull. Under the microscope it can be seen that quartz with a greasy luster has been stained and broken into small particles and that the quartz with a dull or earthy luster has been broken to form a mosaic of even smaller particles. East of the tin-spodumene belt the quartz in veins is coarser and bears little resemblance to that in the pegmatites.

In the pegmatites much of the sugary quartz is interstitial to masses of spodumene, feldspar, and less commonly with muscovite, apatite, or tourmaline; in the veins it may be quite free from other minerals or it may be associated with small amounts of black tourmaline or muscovite. Segregations of sugary quartz in the pegmatites are generally either free from other minerals or are accompanied by as much as 5 percent of muscovite or, rarely, by a little white or pale green beryl.

It is likely that the sugary quartz in the quartz veins is related to the sugary quartz in the pegmatites of the tin-spodumene belt just as the coarse-grained quartz farther west is related to the coarse-grained quartz of the cores of the sheet-mica-bearing pegmatites of the surrounding area. It may be that the movements that opened the fracture zones, permitting the intrusion of the Cherryville quartz monzonite magma and various pegmatite magmas of the tin-spodumene belt, continued after the crystallization of the pegmatites and the veins, but were spread over a broad belt instead of being concentrated in a narrow zone. The late movements may have been the same as or related to those that deformed the pegmatites south of Kings Mountain and represent the waning phase of an orogeny that had earlier caused extensive fracturing.

Small clear crystals of quartz and muscovite have been found in vugs in the plagioclase-quartz muscovite wall zones of two pegmatites that are apparently related to the Cherryville quartz monzonite. The crystals reach lengths of a quarter inch and are intergrown with muscovite, which distinguishes them from the clear float crystals described above. Mica is much more abundant than quartz in these crystal-lined cavities. The cavities have not yet been found in abundance in wall zones of any pegmatite that has been profitably mined for sheet mica.

Mica-lined cavities are common in intermediate zones of at least four productive sheet-mica mines, but these cavities are generally quartz-free.

The quartz is clear near the tips of crystals as broad as 7 inches in vugs in cores of some of the mica-bearing pegmatites near the main W. T. Foster mica mine. The mode of occurrence is not common. Crystal-lined cavities have been found in the cores of only five mica-bearing pegmatite dikes.

Plagioclase

Plagioclase feldspar is a major constituent of all the types of pegmatite in the district. Its grain size and composition vary with its mode of occurrence.

The plagioclase in the muscovite-bearing pegmatites that are related to the Cherryville quartz monzonite forms irregular to blocky grains that range from a fraction of an inch to about a foot in length. Most commonly the megagrains are 3 to 6 inches long. They are invariably associated with quartz and generally with muscovite. Biotite, tourmaline, garnet, and apatite are less common associates. The plagioclase-quartz pegmatite is found in unzoned masses of pegmatite, in wall zones of zoned masses, and also in a few intermediate zones. In the intermediate zones the texture is coarser than that in unzoned bodies or wall zones and it is blocky--the crystals tending to be subhedral to euhedral. Some of these intermediate zones may contain perthitic microcline or small cavities that are lined with crystal faces of the surrounding plagioclase. The cavities may contain crystals of red-brown muscovite, needles of black tourmaline, calcite, or massive yellow sericite. The plagioclase megagrains that bound quartz cores commonly have crystal faces well developed against the quartz but not against the feldspathic pegmatite. The plagioclase in these muscovite-bearing pegmatites is generally thoroughly weathered, but fresh plagioclase was found in several places: it contained 80 to 85 percent of albite.

In the tin-spodumene belt the plagioclase forms sugary-textured masses of grains that seldom exceed 1/10 inch in breadth. The masses commonly are several inches across. Fine-grained plagioclase and quartz form an aplitic matrix that enclosed coarse grains of microcline and spodumene. In the Kings Mountain and Long Creek Church areas the plagioclase and quartz fill fractures in deformed phenocrysts of microcline and spodumene. Few samples examined contained less than 95 percent of albite. There is no apparent relation between the composition of the albite and the nature of the associated minerals, the position in the dike from which the sample was obtained, or the country rock lithology. The plagioclase was most calcic (An_{80}) in a thin dike, scarcely 2 inches thick, that cut gabbro near Indian Creek. The very small volume of pegmatite and the large area of contact between the pegmatite and the enclosing gabbro accentuated the effects of assimilation of calcium from the walls. The pegmatite was composed entirely of feldspar and quartz and there was no megascopically visible evidence that the gabbro had been altered or that the pegmatite had picked up iron or magnesium from the gabbro. Large pegmatite dikes a few yards away and in the same gabbro body contained highly sodic albite.

Plagioclase is also found in most of the other rocks in the district. It is a major constituent of the Toluca and Cherryville quartz monzonites, the Yorkville granodiorite, and the diabase. It is also present in most varieties of the Carolina gneiss.

Microcline

Microcline is a major constituent of the pegmatites in the tin-spodumene belt and is very common in the muscovite-rich pegmatites. In this report "microcline" refers to non-perthitic microcline or to all microclines in the rocks--context will indicate which meaning is intended--"perthite" refers to perthitic microcline. Perthitic orthoclase has not yet been found in the district.

The microcline in the pegmatites in the tin-spodumene belt is similar in appearance to that in the pegmatites related to the Toluca quartz monzonite. It is gray, somewhat translucent, and commonly forms bent or deformed crystals. The grains of microcline range from particles a fraction of an inch to megagrains about 4 feet in maximum dimension. In only a few places does this feldspar contain perthite layers of albite, though irregular patches and more or less round grains of albite are moderately common inclusions.

Perthitic microcline is found in the muscovite-bearing pegmatites, whether sheet-mica-bearing or scrap-mica-bearing. It may be white, light gray, or very pale salmon in color. In some of the unzoned, fine-grained scrap-mica pegmatites near Kings Mountain are perthite grains, like the grains of other minerals, are generally smaller than 1 inch. In the coarser-grained sheet-mica pegmatites, perthite is nearly everywhere restricted to the intermediate zones, where it forms megagrains 2 inches to a foot across. They are blocky and show crystal faces where they are against quartz cores. In a few poorly zoned pegmatites that have been mined for sheet muscovite, perthite forms masses up to 8 inches across, associated with plagioclase, quartz, and muscovite through most of the pegmatite.

Perthitic microcline forms monogranular round or oblong masses up to 8 inches in length in the Cherryville quartz monzonite that are not associated with other large mineral grains.

The content of perthitic plagioclase lamellae was determined in perthite from a mass in the Cherryville quartz monzonite and from another in a dike of muscovite-bearing pegmatite in the Carolina gneiss. The former contained about 12 percent of plagioclase and the latter about 18 percent. About 10 percent of the plagioclase in the perthite from the dike was in stringers transverse to the (001) cleavage in the microcline and about 8 percent was in stringers parallel to that cleavage.

Muscovite

Muscovite is found in all the pegmatites of the district as well as in the granitic rocks and some schists. It is most abundant in the muscovite-plagioclase-perthite-quartz pegmatites that have yielded the sheet and scrap mica mined in the district. It forms red-brown books several inches across. Muscovite is present in most of the pegmatite dikes in the tin-spodumene belt as small yellow-green flakes.

In the muscovite-oligoclase-perthite-quartz pegmatites the muscovite is found in rock that consists of plagioclase, in 3- to 6-inch megagrains, quartz, in 1/4- to 4-inch grains and muscovite. The muscovite content is very low in some masses of this rock but reaches 10 percent or more in others. In the Cliff Blanton mine, for example, the recoverable mica content of the pegmatite was 7 percent. The oligoclase-muscovite quartz aggregate that contains large mica books most commonly forms the wall zones of well zoned pegmatite bodies and the cores of poorly zoned bodies.

The muscovite books in the plagioclase-muscovite-quartz pegmatite may reach lengths of 2 feet or more, though most are 6 inches long or smaller. Most of the books are free from ruling and reeves. Some contain reeved layers that can be rifted out leaving the adjacent layers free from the defect. The flatness of sheets very close to the reeved layers is surprising. Fractures are found in most of the books, but in many they do not prevent the recovery of a substantial amount of sheet mica.

In most deposits mineral stains are not common in the mica, though an occasional book is found that contains one or two opaque black spots of magnetite(?); the largest spot seen was nearly 1/2 inch long. A few deposits in the district contain more heavily spotted mica. The spots are black or dark brown and commonly are concentrated into lines that may be curved and not parallel to any of the major crystallographic directions.

In the eastern part of the district, in the scrap-mica pegmatites alongside the tin-spodumene belt, the muscovite is generally strongly reeved; much of it is characterized by "A" or herringbone structure. In some deposits the reeved books are large and contain flat sheets. The recovery of sheet from such mica rarely is 2 percent of the mine-run whereas the recovery of sheet from the flat books obtained in the western part of the district commonly is 7 percent.

Some of the mica in the eastern area is similar in mode of occurrence to that in the western part but much occurs in the intermediate zones that are high in plagioclase and low in quartz. The wall zones associated with these intermediate zones generally are plagioclase-quartz pegmatite that contains little mica.

Muscovite also is found in plagioclase-bearing intermediate zones in fewer than one-fourth the total number of pegmatite bodies; these zones are associated with wall zones composed of the more common plagioclase-muscovite-quartz pegmatite. The mica in the intermediate zones is generally inferior to that in the wall zones in that it tends to cleave poorly, whereas that in the wall zones is free-splitting.

Muscovite is found in small amounts in many of the pegmatites in the tin-spodumene belt. It forms yellow-green flakes that commonly are not more than 1/10 inch across scattered through the rock. Muscovite also forms veins or pods of quartz-muscovite rock, or greisen. Some of the veins are in pegmatite and some are in schist. Masses of quartz and muscovite in some of the dikes of albite-microcline-quartz pegmatite may be core segments. The muscovite in these is in reeved books that may be 6 inches long, in which interlocking sheets impair the cleavage.

All the sheet mica produced in the Shelby district has been classed as "ruby" in color. In Professional Paper 225 (Jahns and Lancaster, 1950), most of the samples from the Shelby district were called cinnamon brown or pinkish buff. A few were brown, olive, or greenish. There is no marked zoning of mica colors in the district other than the concentration of yellow-green mica in the tin-spodumene belt. The green mica found in the mica-mining areas may be either clear or spotted. Most of the heavily spotted micas examined were green, though rare, isolated spots were found in the ruby mica. Reeved mica generally is striped, with pale streaks along the reeves. These pale streaks are found also in flat sheets obtained from "flat A" mica books. In the Dycus mine the color variations are particularly striking, as the pale yellowish streaks form a regular grid pattern in the darker red-brown mica. The marginal parts of many red-brown books are pale. The pale color may follow particularly strong fractures into the interior of the books.

The color of mica is much lighter and much greener when the books are examined with light transmitted parallel to the cleavage instead of normal to it. Mica that is dark brown when viewed normal to the cleavage is pale yellowish green when viewed parallel to the cleavage.

The refractive indices of 76 micas were determined by the use of a total refractometer. No regional zoning of indices was found, but within individual pegmatite dikes the mica from intermediate zones had slightly higher refractive indices than that from the wall zones. The difference was 0.003 or less in the micas examined. A large "flat A" book from the Dycus mine had the two lower indices 0.001 higher at the margin than in the interior of the sheets; the third index (the highest) was about the same at the margin and in the interior.

Biotite

Biotite is very widespread. It is a major constituent of all varieties of the Carolina gneiss and in all varieties of Yorkville granite, in Toluca quartz monzonite and in most varieties of Cherryville quartz monzonite. It also forms as much as 2 percent of some pegmatites that are related to the Cherryville quartz monzonite. This mica forms books as wide as 5 inches in the sheet-mica-bearing pegmatites. Inasmuch as biotite weathers readily it is difficult to obtain fresh specimens from the kaolinized pegmatites. The fresh material that was found is red-brown. Most of the fine-grained biotite in the non-pegmatitic rocks is likewise brown or red brown.

Biotite is in pegmatites related to the Cherryville quartz monzonite in all parts of the district except in the tin-spodumene belt. It characteristically occurs in the border zones and wall zones of sheet-mica-bearing pegmatites. It probably is, on the average, more abundant in the outer part of the wall zone than in the inner part. The amount rarely reaches 5 percent. No systematic regional variation in biotite content of the pegmatite or of the properties of the biotite had been noted. Biotite is associated with muscovite in the wall zones and small amounts may be found intergrown with muscovite. Common associated minerals are oligoclase and quartz; less common associates are garnet and apatite.

Spodumene

Spodumene is entirely restricted to the albite-microcline-spodumene-quartz pegmatites of the tin-spodumene belt. It may constitute 20 percent or more of these pegmatites. The spodumene is in very small masses, as compared with those in the Black Hills as the longest crystal reported is 38 inches. The width rarely exceeds 6 inches. Most of the spodumene is in megagrains that are no longer than 6 inches, many of which are fragments that have been broken from larger crystals during the crystallization of the pegmatite. Spodumene, like microcline, is more generally broken in the part of the tin-spodumene belt south of Kings Mountain than in the part farther north.

The spodumene is generally opaque and is greenish white in color. Locally it is translucent to transparent and pale green or pale blue. The transparent blue spodumene almost invariably is associated with blue apatite that fills seams in the spodumene as well as in the surrounding pegmatite. The clearness of the spodumene results from alteration that accompanied the emplacement of the apatite veins. The opacity of most of the spodumene is due to a multitude of tiny inclusions. A small proportion of these are of an unidentified black mineral, the rest are transparent and may be fluids. The process of clarification involved the coalescing and expulsion of many of the inclusions.

The hypogene alteration of spodumene to micaceous minerals that has been found in the Black Hills and elsewhere has not been effective in the tin-spodumene belt. One specimen was found in the Kings Mountain area in which the spodumene was altered to a white micaceous material and one block was found in a quarry in which the spodumene had been completely altered, leaving porous masses of albite, lavender apatite, and minor amounts of micaceous material. The only other alteration that was noted was the reduction of turbidity that accompanied formation of blue apatite. The spodumene was readily altered during the weathering of the pegmatite. It was more resistant to weathering than albite, which is the least stable common mineral, but was less resistant than microcline, quartz, and muscovite. The mineral became dull in luster and tended to separate along cleavage planes during the early stages of weathering, and finally was altered to clay minerals.

Beryl

Beryl forms small white or very pale green grains in the pegmatites of the tin-spodumene belt and larger green crystals in the pegmatites in the rest of the district.

Beryl has been found in fewer than a dozen outcrops in the tin-spodumene belt though it has been recognized by microscopic examination of material from other exposures and has been inferred to be present in all samples of pegmatite from the belt because the beryllium content of the rocks is too high to be accounted for by any other mineral known to occur in the district. The grains that were found in the field were recognized by their cyrstal outlines and by shiny crystal faces. Most of the beryl, unfortunately, does not occur in well formed crystals, but forms irregular grains smaller than 1/10 inch and they are white; hence they are not readily distinguished from the surrounding white albite and quartz. One crystal is embedded in white, sugary-textured quartz, the others are embedded in aggregates of albite and quartz. The beryllium content of the pegmatites in the tin-spodumene belt does not vary with the position in the dikes; there is no consistent change either along the strike or across the width of the dikes. Likewise, the spodumene-free pegmatites are not much different in beryllium content from the spodumene-bearing pegmatites.

The beryl in the muscovite- and oligoclase-rich pegmatite is in larger and darker colored crystals. It most commonly is in the outer parts of massive quartz cores and in the adjoining parts of intermediate zones. The intermediate zones that contain beryl generally also contain perthitic microcline, either as the only feldspar or associated with plagioclase. The largest beryl crystal seen in these pegmatites was about a foot in diameter. It has been broken in mining so that the fragment remaining was not over 1-1/2 feet long. The color of the beryl is pale green, in varying intensity. In the Old Plantation emerald mine the beryl, of course, was rather dark green. Farther north, at the Hiddenite mine in Alexander County, the beryl is also darker colored whereas at the Drum mica mine in Catawba County some beryl is yellow. Blue beryl was not found. Clear and transparent beryl was found in the gem deposits at the Old Plantation and Hiddenite mines and small parts of crystals in other places are clear. The crystals are generally solid, though a large crystal at the Bess mica mine in Gaston County had separated along a well developed prismatic parting and the crystals at the Old Plantation mine contained thin tubes parallel to the c-axes.

Tourmaline

Tourmaline is present in small amounts in each of the widespread rock types in the district. It forms layers up to 1/2 inch thick in tourmalinized rocks, both in the Battleground schist and in the Carolina gneiss. It was recovered in panning the Toluca quartz monzonite and the pegmatites related to it and was found in black prisms as long as 4 inches in the mica-bearing pegmatites. It most commonly lies in the outer parts of the pegmatite bodies, where it is associated with plagioclase, quartz, and muscovite. In the pegmatites of the tin-spodumene belt tourmaline grains, 1/4 inch across or smaller, are embedded in albite-quartz rock. The tourmaline in these pegmatites is dark green. Farther east dark blue tourmaline is in quartz veins, both in veins that contain pyrite and gold and some that contain no recognizable ore mineral.

Garnet

Garnet is common in the Carolina gneiss, the Toluca quartz monzonite, and the pegmatites related to the quartz monzonite. It is uncommon in the pegmatites related to the Cherryville quartz monzonite and, apparently is absent from the quartz monzonite itself. It occurs locally in the plagioclase-perthite-muscovite-quartz pegmatites along the eastern edge of the district and was recovered in very small amounts by panning one albite-microcline-quartz pegmatite. Garnet is not present in the heavy mineral concentrate at the Foote Mineral Co. mill at Kings Mountain. The garnet is predominantly almandite.

Staurolite

Staurolite can be obtained in most places by panning weathered Battleground schist and it occurs in Carolina gneiss at a few places near the center and the northeastern corner of the Shelby quadrangle. It forms grains smaller than 1/20 inch in a weathered spodumene-bearing pegmatite east of Grover. These were seen only in a panned concentrate, so the mode of occurrence in the pegmatite is not known.

Dumortierite

Kesler reported that dumortierite was very scarce in the pegmatites of the tin-spodumene belt and indicated that it was not well disseminated through the pegmatite (Kesler, 1942, p. 255). One or a very small number of grains of dumortierite were recovered from a few dikes of pegmatite related to the Toluca quartz monzonite near the center of the Shelby quadrangle. The dumortierite content of the pegmatite is exceedingly small as each sample examined represented about 20 pounds of rock. The bluish-gray grains were small, rarely reaching 1/15 inch in length.

Sillimanite

Sillimanite was not found in any of the pegmatites related to the Cherryville quartz monzonite, but is present in about half of the pegmatites related to the Toluca quartz monzonite. The highest content in the 329 samples of uncontaminated pegmatite panned was 0.014 percent. In some places, where the pegmatite contains wisps of partly assimilated schist, the sillimanite content may be higher.

Andalusite

Andalusite is present in the coarse, muscovite schist variety of the Battleground schist along the tin-spodumene belt, but it is not in the pegmatites. One quartz vein in the tin-spodumene belt contains 10 to 20 percent of fresh pink andalusite and small amounts of muscovite and of kaolinized feldspar, probably plagioclase.

Zircon

Zircon is a minor constituent of all the rocks except the pegmatites related to the Cherryville quartz monzonite and the basic rocks. In the Carolina gneiss and Toluca quartz monzonite zircon forms small more or less round grains that are colorless and transparent. Most of the samples examined under short-wave ultra violet light fluoresce with a bronze-yellow color. The zircon in the Battleground schist is in small, colorless grains that also fluoresce with a bronze color. The surfaces of the grains in the Battleground schist were frosted, either during transportation before the formation of the original sedimentary rock or during the metamorphism.

The zircon in the pegmatites related to the Toluca quartz monzonite are several times as long as those in the other rocks. They are euhedral, with well developed prism and pyramid faces and are brown and opaque. They do not fluoresce.

Age determinations give rather consistent results of about 400 million years on zircon samples from the Carolina gneiss, Toluca quartz monzonite, and pegmatites related to it. The zircon from the Battleground schist gives a much older age, over 600 million years.

Calcite and dolomite

Calcite was found in three mica deposits in the district that were mined below the depth of weathering. In all of these the calcite forms small crystals in vugs or veins or irregular masses in plagioclase-perthite-quartz or plagioclase-quartz intermediate zones. In one it also forms masses in some segments of a discontinuous core. At one mine certain small crystals proved to be dolomite that was associated with massive calcite.

Sulfide minerals

Pyrite, pyrrhotite, and chalcopyrite are in a few sheet-mica deposits, and pyrite, pyrrhotite, chalcopyrite, and sphalerite are in the spodumene-bearing pegmatites. The pyrrhotite in the mica-bearing pegmatites generally forms masses up to 10 inches across that are in quartz cores or in intermediate zones containing perthite or blocky plagioclase. Masses of pyrrhotite contain veinlets 1/10 inch thick of pyrite and chalcopyrite. The chalcopyrite is distinctly subordinate to pyrite in the veinlets. The vein minerals have replaced the pyrrhotite along fractures and cleavage planes. In the Martin mica mine the pyrrhotite is crumbly because of weathering. The degree of weathering varies directly with the amount of pyrite. This suggests that the pyrite is of supergene rather than hypogene origin. If this is true the chalcopyrite must also be of supergene origin. It is, of course, possible that the pyrrhotite was altered to a more readily weathered form during a partial hypogene replacement by pyrite. The secondary origin of the pyrite is reflected in its high content of nickel, which normally enters pyrrhotite. The nickel presumably was inherited from the parent pyrrhotite.

Pyrite and dark brown sphalerite form fairly straight veinlets in the spodumene-bearing pegmatites. These veinlets clearly replace quartz, albite, and microcline. They seem also to be younger than the veinlets of dark blue apatite, but, like the apatite veins, they have caused no apparent alteration of the pegmatite. The thickest sulfide veinlet is about 1/10 inch across. It contains pyrite mainly, with masses of sphalerite about 1/10 inch long. The fractures seem locally to form an intersecting conjugate system. The more persistent fractures contain pyrite with smaller amounts of sphalerite and the less persistent fractures contain more sphalerite than pyrite. The sphalerite is somewhat coarser grained in these than in the persistent veins.

Pyrrhotite was found in larger amounts than pyrite or sphalerite in heavy mineral concentrates of the spodumene-bearing pegmatites but it has not been found in place.

Cassiterite

Cassiterite forms grains 1/50 to 1/10 inch across that are scattered through the feldspathic pegmatite of the tin-spodumene belt as well as in grains as large as 1/4 inch that are in quartz-muscovite rock or greisen that forms layers a few inches thick in pegmatite and schist. The mineral is dark brown to black in hand specimens. In thin sections it is yellow-brown or reddish brown. Color zones are common. The cassiterite is widespread in the tin-spodumene belt, though it is abundant in only a few places.

Columbite

Columbite, like cassiterite, forms small grains scattered through the feldspathic pegmatites in the tin-spodumene belt. It is even less abundant than cassiterite and may not be as widespread. It was seen only in heavy mineral concentrates during the recent investigations so its relations with other minerals are unknown. M. C. Hardin, of Grover, has found columbite in grains as broad as 1/4 inch in soil that apparently formed from hornblende schist or gneiss about a mile east of Grover. This occurrence was not studied, but it might represent a non-pegmatitic occurrence of the mineral.

Magnetite and ilmenite

Black minerals can be obtained by panning most of the rocks in the district. The Cherryville quartz monzonite and the pegmatites related to it contain very little black material, however. The Toluca quartz monzonite and the pegmatites related to it generally contain ilmenite but magnetite is uncommon, having been found in only 23 of the 329 pegmatites panned and in 14 of the 96 quartz monzonites panned. The ilmenite content reaches 0.027 percent in the pegmatites and 0.036 percent in the quartz monzonites. The maximum magnetite content in the two rocks are 0.0016 and 0.0006, respectively. East of the tin-spodumene belt the Battleground schist and Yorkville granite contain magnetite but very little ilmenite.

Some opaque black material that forms spots in muscovite has been called magnetite. One spot carefully shaved off a mica book from the J. B. Patterson property was attracted by a strong Alnico magnet.

Apatite

In the pegmatites in the tin-spodumene belt apatite forms dark green disseminated grains, dark blue stringers that cut the feldspathic pegmatite, and lavender masses that replace feldspathic pegmatite. The dark green apatite appears to have formed rather early during the consolidation of the pegmatite. It may even be older than some of the albite in the aplitic groundmass. The dark blue apatite forms stringers that generally are parallel to the layers in gneissic pegmatite. In places, moreover, it forms veinlets that are transverse to the layering. Where coarse grains of spodumene are in rock that contains abundant blue apatite the apatite may form veinlets with random orientation in the spodumene. These veinlets obviously formed after the complete consolidation of the rock in which they lie and are therefore younger than the somewhat deformed dark green apatite grains. The spodumene was not altered to other minerals by the solutions that deposited the dark blue apatite. It must therefore have been in equilibrium with the solutions. That was not the case later, when the lavender apatite was deposited. The spodumene near the largest apatite mass was completely altered.

The dark green apatite is by far the most widely distributed of the three varieties; the dark blue is less widely distributed, but has been found in many places in several dikes in the Kings Mountain area. The lavender apatite was found only at one place in one dike.

Monazite and xenotime

Monazite was recovered from all but one of 96 samples of Toluca quartz monzonite that were panned and from all but 18 of the 329 samples of pegmatite related to the Toluca quartz monzonite. It was also recovered from many of the samples of biotite and biotite-sillimanite schist from the Carolina gneiss. In all these rocks the monazite forms grains that rarely exceed 1/20 inch in diameter. Those in pegmatite are largest and those in biotite and sillimanite schist are smallest. The mineral is greenish yellow or greenish brown.

Xenotime was found in several places in each of these rocks. It was recognized by the characteristic pyramidal crystals and a color slightly different from that of the monazite. In several places in the western part of the Shelby quadrangle xenotime was intergrown with zircon, with the c-axes of the two minerals parallel.

The distribution of monazite in the Cherryville quartz monzonite is erratic. It was found in hard rock at one place on Muddy Creek, Lincolnton quadrangle, and in one sample of saprolite in a road cut above the hard outcrop in the creek. Another sample from the same road cut yielded no monazite. Monazite is present in muscovitic quartz monzonite near Grover and apparently is more consistently present near Cherryville than elsewhere. The monazite is similar in appearance to that in the Toluca quartz monzonite but is much richer in uranium. Monazite was in a heavy mineral concentrate made from spodumene-bearing pegmatite at the mill of the Foote Mineral Company. The monazite content of the concentrate was not high and the mineral probably constitutes only a few parts per million of the pegmatite.

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