A PETROGRAPHIC STUDY OF THERMALLY METAMORPHOSED PELITIC ROCKS IN THE BURKE AREA, NORTHEASTERN VERMONT

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ABSTRACT. High-grade metamorphic rocks in the Burke area of northeastern Vermont include staurolite-, andalusite-staurolite-, sillimanite-andalusite-staurolite-, sillimanite-garnet-staurolite-, sillimanite-garnet-potash feldspar (rare), kyanite-sillimanite-staurolite-, and kyanite-sillimanite-andalusite-staurolite-bearing assemblages. These rocks are interpreted as having been formed from low-grade schist and phyllite under load conditions higher than those characterizing normal hornfels aureoles. The various assemblages indicate occurrence of a kyanite-sillimanite-andalusite-bearing assemblage suggests that these polymorphs were probably formed under closely similar conditions during thermal metamorphism but not necessarily simultaneously. Change of temperature is easier to envisage than change of pressure during the crystallization of this assemblage. Temperature and pressure conditions must have hovered near to the assumed triple point of the alunino-silicates.

INTRODUCTION

During the course of field work in the Burke quadrangle of northeastern Vermont large areas of metamorphic rocks clearly affected by thermal metamorphism and related to intruded granitic rocks were mapped. Figure 1 is a generalized geological map of the area with isograds of progressive metamorphism. The thermal effects diminish away from the granites, and the rocks pass into regionally metamorphosed phyllite and schist with no obvious evidence of thermal alteration. Study of thin sections of the numerous specimens collected has shown interesting metamorphic assemblages and textures that indicate the polymetamorphic nature of the rocks and their metamorphic trends.

The purpose of this paper is to describe the petrography of these assemblages and to discuss the evidence bearing on their parageneses.

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PETROGRAPHY OF THE PELITIC ROCKS

The Gile Mountain formation, which occupies a wide belt of country trending northeastwards through the Burke area, has as its characteristic lithology a banded alternation of light gray quartzose phyllite and dark gray pelitic or slaty phyllite. The bands range in thickness generally from a few millimeters to several centimeters.

The pelitic rocks are fine grained, with a well developed schistosity commonly cut by a later cleavage. Muscovite is an abundant constituent. Fine-grained quartz (0.03 to 0.08 mm) is always present, increasing in amount as the rock grades into more quartzose types. Opaque minerals include graphitic dust, pyrite, and some iron ores. Groundmass chlorite occurs only in the lowest metamorphic grade, but red-brown biotite is ubiquitous, both aligned with the schistosity and unoriented.

1 Granite as used here includes rocks with compositions ranging from true granite to granodiorite and monzonite.
Higher metamorphic grades are indicated by garnet and staurolite porphyroblasts, frequently idioblastic. The appearance of wine-colored to deep red
almandine (up to 6 mm in diameter) marks the garnet zone, although garnet also occurs in the staurolite and higher grade zones. The garnets usually have inclusions of quartz and some show alteration to chlorite. Garnets from the garnet zone have a refractive index of 1.795 (± 0.004) and a specific gravity of 4.05, indicating a composition of about 70 percent almandine and 30 percent pyrope (Winchell and Winchell, 1951). Staurolite (up to 22 mm in length) is commonly "spongy" with inclusions and may have a corona of sericite; or it may be completely replaced by white mica or partly altered to chlorite.

Adjacent to granite areas and within areas of granite-hornfels complex the pelite has been converted to andalusite- and sillimanite-bearing hornfels in...
Fig. 3. Large porphyroblast of andalusite, largely altered to white mica but with relict mass in the center; prisms of sillimanite occur within the andalusite relict and in the replacing white mica. Staurolite is present in the andalusite. Biotite around the periphery of porphyroblast is altered to fibrolite, which also occurs within the white mica. Quartz, plagioclase, and abundant opaques are present, and some chlorite elsewhere in the thin section. From outcrop at 1460 ft elevation, north-northwest slope of Burke Mountain.

which the banding, schistosity, and cleavage are commonly still clearly evident; the hornfels may be cut by many granite stringers. In places, however, the phyllite or schist has been converted to more thoroughly recrystallized and granitized fine-grained biotite-rich rocks with no obvious relicts of previous structures. In these rocks plagioclase is usually present in notable amounts as twinned individuals or as xenoblastic, void porphyroblasts full of opaque inclusions.

Rarely occurring as fresh porphyroblasts (fig. 2), andalusite is generally found as pleochroic relics in muscovite pseudomorphs (fig. 3) called “shimmer-aggregates” by Barrow (1893). These shimmer aggregates, which retain the shape of the replaced mineral and range in length up to 40 mm, may replace staurolite, as seen in interpenetrating pseudomorphs, but usually their
form indicates that they are after andalusite or sillimanite. Those enclosing andalusite relicts also contain staurolite, fibrolite, and prisms of sillimanite as well as quartz, biotite, and tourmaline. Occasionally the andalusite is penetrated by fibrolite. Where andalusite relicts are absent (figs. 4 and 5), it is more difficult to be certain of the identity of the original mineral. When the sillimanite prisms have their "c"-axes parallel to the long direction of the muscovite pseudomorph, the pseudomorph appears to be after sillimanite, but randomly oriented needles and fibrous masses of fibrolite are always present as inclusions in the white mica. Although the fibrolite appears to replace the mica, they are more likely to have grown simultaneously. Thin-section evidence suggests that the stout pseudomorphs are after andalusite, which perhaps had partially inverted to sillimanite, and the remainder had then been replaced by white mica along with the growth of new fibrolite. In some cases less robust
aggregates of white mica, sillimanite, and fibrolite seem to be after sillimanite porphyroblasts, which have been largely replaced.

Sillimanite occurs in the hornfels: (1) in the shimmer-aggregates described above, (2) as porphyroblasts up to 22 mm long, (3) as needle inclusions in quartz, plagioclase, and white mica, (4) as matted fibrous aggregates in white mica, and (5) as matted fibrous aggregates associated with and passing into dark brown biotite, which is commonly bleached. The latter aggregates are quite abundant, and in a number of thin sections are the only type of sillimanite present (fig. 6). That the fibrolite is after biotite is obvious in many thin sections (figs. 2, 3, 5, 6, and 7) from its intimate association with biotite and from the direct passage of the latter into fibrolite accompanied by much iron ore released in the change. Of common occurrence, too, is the bleaching of biotite with the concomitant release of ores in close juxtaposition with the...
Fig. 6. Abundant biotite altered in part to masses of fibrolite, with quartz, oligoclase, muscovite, and iron ores in a spotted, incipiently granitized schist from East Branch, Passumpsic River, 4¾ miles north of East Haven.

appearance of fibrolite (fig. 5). In a few instances there are some colorless blebs present which may be potash feldspar.

Kyanite has been detected in only two outcrops: (1) as small plates (fig. 7) with xenoblastic staurolite, biotite (in part bleached and altered to matted fibrolite), muscovite, quartz, accessory ores, and tourmaline and (2) as large brilliant blue porphyroblasts (fig. 8), up to 87 mm long, in a pegmatitic vein together with large pink andalusite crystals and with quartz and plagioclase containing needle sillimanite. The vein is in contact with a biotite-rich schist composed of biotite (bleached in part and altered to masses of fibrolite), staurolite, quartz, muscovite, and ores.

Staurolite in the hornfels is usually xenoblastic, but in one case it occurs in large porphyroblasts within a fine-grained groundmass of quartz, decussate biotite, and opaques. The staurolite is full of inclusions of quartz and opaques
Fig. 7. Kyanite-staurolite schist, with biotite, in part altered to masses of fibrolite, common small green tourmaline, quartz, plagioclase, and iron ores. Muscovite occurs elsewhere in the section. The staurolite is xenoblastic and the kyanite occasionally intergrown with biotite. From track, 1 3/4 miles north of Center Pond, Newark township.

Regularly arranged as in chiastolite. The area of inclusions is interrupted by smaller lath-shaped, inclusion-free portions that have a size and arrangement similar to the biotite outside of the porphyroblasts. Apparently the absorption of biotite into staurolite has been a late development in a rock already possessing a hornfelsic texture.

Large randomly oriented chlorite porphyroblasts in the pelitic rocks, particularly in the garnet and staurolite zones and even in the andalusite zone, appear to have been a late development and perhaps contemporaneous with the chloritization of amphibole, garnet, staurolite, and biotite. Many determinations of refractive index have given an average $\beta$ of 1.638 (± 0.002) for these porphyroblasts: they have a low birefringence and are usually uniaxial positive. These data indicate that the mineral is ripidolite (Winchell and Winchell, 1951) with an Fe/Mg ratio of about 1:1 (Hey, 1951). It is difficult to give
Fig. 8. Large andalusite, much fractured (with patchy extinction) and altered slightly to sericite around the periphery; intimately associated with kyanite, which is altered to white mica around the border not in contact with the andalusite. The white mica contains inclusions of needle sillimanite. Biotite largely altered to masses of fibrolite lies adjacent to the andalusite. Some quartz is also present. From a pegmatitic vein in a roadcut on Route 114, 5 miles north of East Haven.

reasons for their growth, but it is possible that they formed under conditions of waning temperature and under the influence of introduced iron and magnesia driven out from the higher grade rocks.

The following mineral assemblages have been observed in the Burke area:
1. Staurolite ± garnet + biotite + muscovite + quartz ± chlorite
2. Staurolite + andalusite + biotite + fibrolite + muscovite + quartz ± plagioclase ± chlorite
3. Andalusite + biotite + muscovite + quartz
4. Sillimanite + andalusite + staurolite (often relict only) + biotite + muscovite + quartz ± plagioclase ± chlorite
5. Sillimanite + garnet + staurolite + biotite + muscovite + quartz ± plagioclase
6. Sillimanite ± garnet + biotite + muscovite + oligoclase + quartz + rare potash feldspar
7. Staurolite + kyanite + fibrolite + biotite + muscovite + quartz (+ rare orthoclase)
8. Staurolite + fibrolite + andalusite + kyanite + quartz + biotite + muscovite + oligoclase

METAMORPHIC HISTORY

The texture and mineralogical composition of the rocks indicate that the Burke area was regionally metamorphosed during deformation to a relatively low grade, i.e., greenschist facies, quartz-albite-epidote-biotite and perhaps quartz-albite-epidote-almandine subfacies, with a well developed foliation. Continued movement deformed the foliation with little or no obvious recrystallization, except perhaps for limited recrystallization along the newly developed cleavage. A subjacent mass of granitic magma then caused a rise in temperature of the country rocks which probably initiated the development of porphyroblasts. Structurally controlled intrusion of the magma produced further local increases in temperature in the immediate neighborhood of the granitic bodies. The large volumes of metamorphic rock injected by granitic magma formed a mixed complex, a large-scale agmatite, with zones of granite ranging from stringers to large masses intermixed with high-grade hornfels. The petrography of the hornfels indicates that in places it was also metasomatized and "granitized". Little contemporaneous deformation occurred during this metamorphic phase except adjacent to the granite, so that the various mineral assemblages are believed to represent mineral parageneses developed under static conditions of equal load pressure but with a distinct lateral thermal gradient.

The intruded country rocks show lateral temperature variation by the relatively regular order of occurrence of porphyroblasts similar to the sequence described from many regionally metamorphosed areas. Isograds based on these porphyroblasts have been drawn marking the entry of garnet, staurolite, and andalusite-sillimanite (fig. 1). Much new porphyroblastic biotite throughout the lowest zones of the pelites represents the lowest grade of the later episode of metamorphism recognized in the Burke area.

A notable feature of the country rocks is the fine grain size of their groundmass up to the andalusite-sillimanite isograd with little evidence of recrystallization. The minerals present are the ones expected in equilibrium assemblages in these zones except in one instance. This is a chlorite schist with a "strain-slip" cleavage; interrupting the schistosity and cleavage are garnet porphyroblasts, with a "spongy" peripheral zone, associated with staurolite, which has a rim of sericite. Large chlorite flakes occur around the periphery of both the staurolite and garnet. A halo-like zone has developed in the schist surrounding the garnet and staurolite because of the absence of groundmass chlorite. There are also aggregates of sericite and quartz with rims of chlorite that have a similar halo-like zone around them. The thickness of these halo zones indicates the extent of diffusion (0.2 to 1.4 mm) during the growth of the porphyroblasts in the chlorite schist.
At the time of the later rise in temperature the phyllite and schist contained up to a few percent of water in hydrated minerals, such as micas and hornblende. The near absence of intergranular and shearing movements would inhibit the circulation of fluids which promotes recrystallization during dynamothermal metamorphism and would thus tend to make the system a closed one. This may have made the temperature of dehydration reactions somewhat higher than that in normal regional metamorphism (Thompson, 1955), although the rocks were probably “drier” than unmetamorphosed rocks would have been during one-phase metamorphism. These considerations explain why the grain size of the groundmass of rocks in the staurolite zone and even in the andalusite zone remained small and they explain the small range of diffusion of the elements as exhibited by the staurolite-garnet-chlorite schist (but they do not explain why porphyroblasts developed only at relatively widely spaced intervals). Adjacent to the granitic rocks, particularly in the hornfels-complex zones, there is petrographic evidence of metasomatism and of the greater activity of fluids. e.g., the larger general grain size (in some cases notably so), the appearance of large xenoblasts (up to 1.58 mm) of plagioclase, the alteration of staurolite and andalusite to white mica, the appearance of abundant large plates (up to 12 mm) of white mica in some restricted zones, and local granitization of the hornfels.

MINERAL PARAGENESIS OF THE PELITES AND METAMORPHIC FACIES

Staurolite-bearing Rocks

Assemblages of the staurolite zone are shown in figure 9A. utilizing the method described by Thompson (1957) for projecting the system Al₂O₃–FeO–MgO–K₂O in muscovite-bearing assemblages with quartz in excess and the chemical activity of H₂O determined externally. The garnet composition used is that described under petrography. Staurolite apparently has little variation in its FeO/MgO ratio and this variation cannot be reliably obtained from optical data (Juurinen, 1956). Data for the optical determination of the FeO/MgO ratio of natural biotites are not yet reliable. Wones (1958) plots the refractive index (γ) for synthetic phlogopite-anite and also shows that at higher temperatures the biotites produced have a higher MgO/FeO ratio. Applying Wones’ data for variation of refractive index to biotites in a number of Burke specimens from the various metamorphic zones, the Fe/(Fe+Mg) ratio apparently declines from about 57 percent to about 50 percent from the biotite zone to the staurolite zone.

Staurolite has been said to be restricted to rocks of somewhat limited chemical composition (Suzuki, 1930) and to have a limited temperature range (Turner and Verhoogen, 1951, p. 454). Reactions suggested for its formation involve chloritoid (Harker, 1939, p. 225; Tilley, 1925, p. 318). However, chloritoid has not been identified in the Burke area, although it (or ottrelite) has been reported from neighboring areas (Dennis, 1956; Eric and Dennis, 1958; Hall, 1959). Presumably it was not formed, although according to Tilley (1925, p. 317), chloritoid is “stable over a wide chemical field, but it also possesses a wide physical field of stability”. Staurolite may then have formed from biotite. Petrographic evidence for this is indicated by the staurolite porphyroblasts in hornfels, as described under petrography.
In rare instances a Burke rock contains staurolite and andalusite, both appearing as stable constituents and to have formed contemporaneously. This presumably represents a staurolite zone assemblage in more aluminous pelite, as indicated in figure 9A. An analogous assemblage with sillimanite in place of andalusite is also found occasionally (the lower sillimanite zone of Thompson, 1957). In general, however, the spatial occurrence of staurolite relative to that of andalusite and sillimanite indicates that the latter two formed in higher temperature zones.

**Andalusite-bearing Rocks**

Mineral paragenesis of the higher temperature zones adjacent to the granite and within the hornfels-granite complex is complicated by reactions involving not only higher temperatures but also the addition of fluids (therefore probably an increase in the chemical activity of water). Equilibrium conditions may thus have varied in these rocks from place to place even down to the microscopic level. The higher temperatures produced andalusite. It is believed that this was formed mainly at the expense of staurolite, because the latter dwindles in amount in these zones, and in a number of Burke specimens it is present only as inclusions within andalusite or andalusite pseudomorphs. Biotite, however, is very abundant. A similar interpretation is proposed by Suzuki (1930) in the following reaction:

\[ \text{Staurolite} + \text{muscovite} + \text{quartz} \rightleftharpoons \text{andalusite} + \text{biotite} + \text{water} \quad (1) \]

He suggests that staurolite formed during a phase of dynamic metamorphism and that andalusite formed later during a phase of thermal and contact metamorphism accompanying pegmatitic intrusion.

Streckeisen (1928) attributes the formation of staurolite, kyanite, and garnet, together with the bleaching of biotite and its alteration to sillimanite, to a stress phase and the formation of andalusite (which encloses staurolite) to a later phase under the influence of the intrusion of granitic magma and injection of the schist. He states that the staurolite is partly consumed by the andalusite, but that the kyanite and sillimanite are also partially absorbed.

Akaad (1956) records staurolite as inclusions in andalusite; the staurolite formed during an earlier phase of thermal metamorphism.

These staurolite-free andalusite assemblages would then plot on the diagram shown in figure 9B, and may be regarded as analogous to the kyanite-muscovite-quartz subfacies of Turner (in Fyfe, Turner, and Verhoogen, 1958). only formed at lower pressures or at a higher partial pressure of water.

Thompson (1955, p. 84) suggests the reaction:

\[ 3 \text{Staurolite} + 2 \text{quartz} \rightleftharpoons \text{andalusite} + 5 \text{andalusite} + 3 \text{water} \quad (2) \]

This is analogous to the reaction proposed for the production of kyanite from staurolite (see, e.g., Francis, 1956), but taking place in the stability field of andalusite. There is no real evidence for this, however, because of the absence or paucity of garnet in Burke andalusite-bearing assemblages.

There is also no petrographic evidence for the reaction:

\[ \text{Muscovite} + \text{quartz} \rightleftharpoons \text{andalusite} + \text{orthoclase} + \text{water} \quad (Turner and Verhoogen, 1951, p. 447) \quad (3) \]

as muscovite is stable and a common constituent of the Burke rocks. The tem-
temperature was never high enough relative to the partial pressure of water to drive the reaction to the right.

Andalusite is normally associated with thermally metamorphosed aureoles in the hornblende-hornfels facies (Turner, p. 207, in Fyfe, Turner, and Verhoogen, 1958), where it is commonly accompanied by sillimanite adjacent to granitic intrusions. The mutual occurrence of garnet, staurolite, andalusite, and sillimanite has been considered to be less common. But Chinner (1962) suggests that such regional andalusite-sillimanite sequences may well be the "normal" sequence in regional zones. Miyashiro (1961) recognizes five metamorphic facies series based on increasing rock pressure. His lowest pressure group is the andalusite-sillimanite type (central Abukuma type), followed by the low-pressure intermediate group in which staurolite occurs along with andalusite-sillimanite. The latter group, he states, is not rare. Read (1923, p. 59) describes the regional occurrence of andalusite schists in Aberdeenshire and Banffshire, Scotland, in which staurolite, garnet, and sillimanite also occur, together with cordierite, and explains (1952) these schists as due to regional thermal metamorphism which he calls the Buchan type.

Sillimanite-bearing Rocks

Further changes in the Burke mineral assemblages were accompanied by a more thorough recrystallization of the country rock, destruction of its lepidoblastic texture, and, finally, by a large increase in its content of white mica, which serves to obscure the preceding steps in the mineral paragenesis. Sillimanite appeared during these stages both in fibrous mats and in robust prisms up to 22 mm long. Assemblages of the sillimanite-muscovite zone are shown in figure 9B: they represent conditions of higher temperatures than those of the rocks containing andalusite without sillimanite.

Fibrolite sheaves and coarser prismatic sillimanite are enclosed in large white mica plates and in intergrowths with aggregates of white mica forming pseudomorphs after andalusite. The andalusite relics in some pseudomorphs are associated and intergrown with fibrolite and prismatic sillimanite (fig. 3). This suggests that the andalusite was inverting to sillimanite. Pseudomorphs containing needle and fibrous sillimanite but no relict andalusite are better explained in this way than considering them as originally sillimanite porphyroblasts. This inversion would thus be a source of the sillimanite. Akaad (1956) records pseudomorphs of fibrolite after andalusite in a narrow transition zone between an outer andalusite and an inner sillimanite zone in Donegal.

Another possible source of sillimanite lies in the following reaction:

\[
\text{Staurolite + quartz } \rightleftharpoons \text{almandine + sillimanite + water} \quad \text{(Turner and Verhoogen, 1951, p. 457; 1960 p. 548)}
\]

No direct evidence of this has been detected in the Burke area, but assemblages containing sillimanite-garnet are present. Although these commonly show fibrolite after biotite, this is not always the case; the origin of the sillimanite may be from reaction (4). These rocks may also contain regenerated biotite, formed by a reaction analogous to equation (1). Perhaps in these cases the rocks passed through the stability field of andalusite too rapidly for that mineral to form.
Sillimanite may also arise by:

\[ \text{Muscovite + quartz } \rightleftharpoons \text{sillimanite + orthoclase + water} \quad (\text{Turner and Verhoogen, 1951, p. 457; 1960, p. 551}) \quad (5) \]

There is little evidence for this reaction, as muscovite is present and often abundant in all of the Burke assemblages, whereas potash feldspar is absent or rare. The latter may occur as minute blebs associated with sillimanite, bleached biotite, and muscovite in one or two thin sections, and it is present or probably present in small quantity in some migmatized specimens, although plagioclase is the usual feldspar in such rocks. There is no petrographic evidence that muscovite forms pseudomorphs after potash feldspar. Evidently the physical conditions for the assemblage sillimanite-potash feldspar were not generally attained in the Burke area.

The fibrolite mats, frequently associated with biotite and bleached biotite but also found in quartz, plagioclase, and white mica, indicate relationships similar to those described by Streckeisen (1928), who emphasizes that the fibrolite is not related to andalusite or kyanite but appears to have formed exclusively from biotite. Tozer (1955) ascribes fibrolite and sillimanite from Donegal to two successive stages of metamorphism: (1) the earlier fibrolite arising from biotite during the main thermoplastic phase and (2) the later fibrolite and sillimanite forming during the subsequent metamorphic phase contemporaneously with and a by-product of the development of later muscovite metacrysts. These metacrysts, induced by late-stage potash-rich fluids from granite, formed at the expense of biotite of much of the earlier fibrolite and probably of quartz (see also, Pitcher and Read, 1960; Akaad, 1956). In the Burke area the appearance together of fibrolite and needle sillimanite commonly included within new muscovite, both the muscovite forming pseudomorphs after andalusite and other muscovite (fig. 4), indicates their close relationship and possible contemporaneous origin. Pitcher and Read (1960) describe very similar occurrences and their figures 1 and 2 of plate III bear a close resemblance to the example illustrated here in figure 3.

Francis (1956, p. 358) suggests the following reaction for producing sillimanite:

\[ \text{Fe-biotite + dissolved “kyanite” } \rightleftharpoons \text{potash feldspar + sillimanite} \]
\[ + 3 \text{FeO} + \text{H}_2\text{O} \quad (6) \]

in which aluminosilicate in solution may act as a catalyst for the bleaching of biotite and its sillimanitization. The failure to identify potash feldspar, except very rarely, in Burke metamorphic rocks seems to argue against this reaction, but it may be present as an occult constituent. There is biotite in all of the assemblages, but commonly it is partly altered to a bleached mica or sillimanite, accompanied in a few instances by possible blebs of orthoclase. The petrographic evidence indicates that the alteration of biotite produces sillimanite and muscovite together in some cases and one or the other alone in other cases. As muscovite is stable throughout the area, the reactions involving the breakdown of biotite must have taken place within the stability field of muscovite, and the conditions for the breakdown of the latter were not generally attained. While magnesium-rich biotite should survive into higher grades than muscovite (compare the experimental results of the system muscovite-quartz
(Segnit and Kennedy, 1961) with those of phlogopite (Yoder and Eugster, 1954). MacKenzie (1960) states that this is not always the case because of interdependent reactions. Iron-rich biotites have much reduced stability fields compared to magnesium-rich biotites (Wones and Eugster, 1959); the addition of quartz also restricts considerably the stability of annite (Eugster and Wones, 1962) and thus presumably the stability of iron-rich biotite. Chinner (1961) ascribes fibrolite intergrown with biotite to the latter acting as a nucleating agent for the growth of the fibrolite, aluminum and silicon being derived from the solution of unstable kyanite. The biotite does not actually break down but becomes temporarily unstable because of changes in the partial pressure of water and oxygen. The presence of much magnetite associated with the fibrolite and altered biotite in the Burke area suggests, however, that here the biotite is actually in the process of breaking down.

The biotite of the high-grade hornfels (sillimanite zone) is frequently a dark “dirty” brown in contrast to the rich red-brown typical of the lower metamorphic zones. This color change indicates a change in composition. If Wones’ (1958) diagram of variation of refractive index in synthetic phlogopite-annite is utilized, the biotites of the sillimanite zone appear to have an annite content ranging from 55 percent to 65 percent. Contrary to expectations, this is an increase in the FeO/MgO ratio compared to the lower metamorphic zones. Either refractive index is an unreliable indicator of FeO/MgO in natural biotites [Suwa (1961) suggests that Ti+Fe2+ content strong affects the refractive index of biotite], or the biotites of the sillimanite zone are more iron-rich and variable in composition. In addition to the ample petrographic evidence for the incomplete breakdown of biotite and release of iron ores, there is also evidence for the possible regeneration of biotite. Conditions such as the partial pressures of H2O and of oxygen may have varied from place to place and from time to time in the sillimanite zone. It seems likely that the alteration of iron-rich biotite in the Burke area was a late reaction and took place under metasomatic conditions at about the same time as the potash metasomatism of andalusite and staurolite and the growth of the large muscovite plates.

The kyanite-bearing Rocks

Kyanite-schist.—The assemblage of the kyanite-bearing schist, which occurs adjacent to granitic rocks, is kyanite-staurolite-biotite-muscovite-quartz-fibrolite plus accessory ores and tourmaline (fig. 7). The fibrolite is not related to the kyanite nor to the staurolite but appears to be solely after biotite; fibrolite also occurs as needles in quartz. The biotite appears to be in part bleached to a white mica. Kyanite is the unusual member of the assemblage. It may be that composition permitted its appearance, analogous to andalusite coexisting with staurolite (fig. 9A). It may, however, have developed in place of the andalusite of other assemblages, perhaps by a reaction similar to equation (1) (fig. 9B). In either case, one explanation of its occurrence may be that the reaction took place on the higher-pressure side of the kyanite-andalusite inversion curve (fig. 10). It is difficult to account for local increase of pressure unless directed stress was operative. An alternative explanation is that the de-
hydration curve, representing the reversible reaction, may have been displaced as a result of locally reduced partial pressure of water, thus allowing the appearance of kyanite (Thompson, 1955).

Kyanite-pegmatite.—The vein assemblage of kyanite-andalusite-quartz-plagioclase in contact with biotite-rich schist containing biotite-muscovite-staurolite-fibrolite-quartz and with tourmaline and sphene as accessories is unique for the Burke area. The fibrolite, entirely after biotite which shows bleaching, is accompanied by much iron ore. Kyanite and andalusite are slightly altered to muscovite. Biotite-muscovite-fibrolite aggregates occur only adjacent to the vein and in some cases in contact with kyanite. Fibrolite needles are contained within the quartz and plagioclase of the vein margins. Andalusite and kyanite are intergrown in some cases (fig. 8) but there is no definite petrographic evidence to suggest that either was formed from the other. Rather, it appears that both crystallized simultaneously, but kyanite may be in part after andalusite. Some, if not all, of the alteration of biotite to fibrolite seems to have taken place prior to the crystallization of the vein minerals, as fibrolite needles occur in the quartz and plagioclase and as biotite, altered in part to fibrolite, is included within andalusite. However, restriction of the alteration to a narrow zone adjacent to the vein indicates that the fibrolite is contemporaneous with and was formed as a result of emplacement of the vein.

Petrographic evidence of postcrystallization stress is provided by the deformed and strained albite twin lamellae of the oligoclase, undulose extinction in the quartz, one strained and fractured kyanite porphyroblast, and the shattered and strained appearance of the andalusite, as well as the contorted nature of the biotite-rich schist bordering the vein which contrasts with the

![Diagram](image)

Fig. 9. Projections of the system SiO_2-Al_2O_3-K_2O-FeO-MgO-H_2O, showing phases stable with quartz, muscovite, and plagioclase: A, in staurolite-bearing assemblages (kyanite and sillimanite replace andalusite under changed conditions of temperature, pressure, or partial pressure of H_2O); B, in sillimanite-bearing assemblages (andalusite replaces or accompanies sillimanite under different physical conditions). Assemblages of diagram B are separated from those of diagram A by the breakdown of staurolite.
regularity of the foliation in the surrounding schists. The strains appear to be related directly to the formation of the vein itself rather than to a general stress affecting a wider environment. Some of the biotite grains contain a fine helicitic structure marking a relict wavy schistosity and indicating postregional deformation crystallization. The time of origin of the vein and its associated biotite-rich border relative to the metamorphic history of the Burke area would seem to have been after the regional development of schistosity and later fracture cleavage, and presumably to have been contemporaneous with the thermal metamorphism. Granitic outcrops occur within about 30 yards of the vein, and the nearby schists contain fibrolite and muscovite-biotite pseudomorphs. The vein probably developed by metamorphic segregation (with or without addition of other components from external sources) of the adjacent schist, producing a biotite-rich schist zone around an aluminum-rich quartz-oligoclase vein. The andalusite and kyanite crystallized more or less contemporaneously while the biotite altered to fibrolite. Conditions thus appear to have been close to the presumed triple point of the three polymorphs of Al₂SiO₅. Kyanite may have been the earliest to crystallize, followed by fibrolite and andalusite when the temperature rose. Or, alternatively, kyanite may have formed later and in part after andalusite as a result of local stress, which caused the strain effects noted above.

Kyanite-quartz segregations are not uncommon (Stuckey, 1932; Read, 1933; Heinrich, 1948; Espenshade and Potter, 1960) nor are kyanite-pegmatites. Andalusite-quartz segregations are also frequently described in the literature (Streckeisen, 1928; Kerr, 1932; MacKenzie, 1949; Pitcher and Read, 1960), as well as andalusite-pegmatites (MacDonald and Merriam, 1938; Webb, 1943; Seager, 1944). Sillimanite in pegmatites has been described by Heinrich (1948). Various origins for such segregations and pegmatites have been suggested, including regional metamorphism, thermal metamorphism, metamorphic segregation, metasomatism, and the influence of heated fluids or gases, whereas the source of alumina has been interpreted as being from the adjacent country rocks, partly or wholly introduced from magmatic sources or added to moving fluids from more distant alumina-rich rocks. For the Burke vein, the source of the alumina is believed to have been mainly the adjacent country rocks, and the vein itself to have been formed as the result of the action of introduced fluids catalyzing segregation in the immediately adjacent schists and crystallizing under conditions of high temperature and hydrostatic pressure (? and stress) near to the invariant triple point of the three polymorphs of Al₂SiO₅.

STABILITY RELATIONS OF Al₂SiO₅ POLYMORPHS

The pressure-temperature relationships of the three Al₂SiO₅ polymorphs are still not completely known, nor are the influences of other factors, such as the chemical composition of the phases involved. A number of different phase diagrams have, however, been proposed, based on the interpretation of field data, e.g., Korzhinskii (1937), Miyashiro (1949), Thompson (1955), Schuiling (1957) assigns values of temperature, pressure, and depth to his diagram mainly by correlating the facies of each reported occurrence with
assumed values for the various metamorphic facies. Clark, Robertson, and Birch (1957) have obtained experimentally the kyanite-sillimanite equilibrium curve between temperatures of 1000° and 1300° C and pressures of 18,200 and 21,000 bars, and extrapolate the curve to lower temperatures and pressures. They construct a tentative phase diagram for the polymorphs which is similar to that of Miyashiro, Thompson, and Schuiling except that they suggest that kyanite is not the stable form under surface conditions. Their triple point lies at a lower temperature and a considerably higher pressure than that of Schuiling’s diagram. A later diagram by Miyashiro (1961) is quite similar to that of Clark, Robertson, and Birch (1957). Clark (1960, 1961) records later experimental determinations of the kyanite-sillimanite curve that are very close to the previous results. The phase diagram in figure 10 is based on the diagram of Clark, Robertson, and Birch (1957) and its modifications by Clark (1961).

Clark concludes (1960, p. 55) that the stable formation of kyanite “apparently requires pressure of ten kilobars or more. It may be inferred from this that pressures in this range are commonly attained during metamorphism. This is considerably more pressure than most previous investigators have assumed, and it indicates the need of substantial revision of ideas about the metamorphic process”.

He suggests (1961) that great depths of burial (e.g., a pressure of 10 kilobars is equivalent to a depth of more than 35 kilometers) may be appreciably lessened by “tectonic overpressure” of 1 kilobar or more equal to nearly 4 kilometers of overburden.

If we assume that the coexistence of sillimanite and andalusite, of kyanite and sillimanite, and of kyanite, sillimanite, and andalusite indicates physical conditions at the time of their origin near to their respective stability fields, then the range of temperatures and pressures during their formation must have been quite restricted. The presence of staurolite in all of these assemblages, although sometimes only relict, suggests that conditions were not far removed from its stability field. According to Clark, Robertson, and Birch (1957, p. 638), the field around the triple point of the aluminum silicates is one of high pressures and moderate temperatures. The temperature is on the low side of the average geothermal gradient, shown in their diagram (see fig. 10). Geothermal gradients would not, of course, be the same in various belts where different metamorphic facies series are produced. The gradient (ΔT/ΔP) may be lower, for example, in zones where the kyanite-sillimanite series is produced than in zones of andalusite-sillimanite. The petrographic evidence from the metamorphic assemblages of the Burke area, in comparison with areas of the Dacian type of metamorphism, indicates that the temperature was relatively high in proportion to the depth in the crust, whereas the position of Clark’s triple point would imply a relatively low geothermal gradient.

Chinner (1962) suggests that the triple point lies in the vicinity of 10 to 11 kilobars and 500° to 600° C. The latter is a much higher temperature than Clark’s. [It may be noted here that Aramaki (1961) suggests that the andalusite-sillimanite boundary of Clark, Robertson, and Birch (1957) should cross the temperature axis at a higher temperature.] Chinner’s triple point lies on the kyanite-sillimanite inversion curve near its intersection with the average geothermal gradient. Higher temperatures required to push the field into the
sillimanite zone would thus be even greater than 500° to 600° C and would approach the minimum melting curve for granite (Tuttle and Bowen, 1958), but not the reaction curve of muscovite + quartz which lies near 800° C at such pressures. Even allowing for tectonic overpressures as suggested by Clark (1961) (and petrographic evidence indicates that the second Burke metamorphism took place under static conditions), the depth would still be in excess of 30 kilometers. It appears reasonable to assume that there was no great change in depth of burial between the time of the regional metamorphism to the level of greenschist facies and the time of the intrusion of granite and the accompanying thermal metamorphism. This implies that the upper greenschist facies developed at depths of over 30 kilometers and at concomitant temperatures. (The geothermal gradient then may well have been relatively low in a thick geosynclinal sequence.) Such depths are greatly in excess of previous published estimates. Although no independent check of depth is yet available, if the depth is excessive, the pressure of the triple point of kyanite-andalusite-sillimanite suggested by Chinner is too high. A position on Clark's (1961) kyanite-sillimanite curve towards lower pressures, e.g., to the position as shown by Clark, Robertson, and Birch (1957), leads to lower temperatures and to the difficulty previously mentioned. A position within Chinner's temperature range but at lower pressures is inconsistent with the extrapolated kyanite-sillimanite inversion curve. Further data on pressures operating at the temperatures of metamorphism, e.g., more data on the P-T diagram for the polymorphs of Al$_2$SiO$_5$, are required for the resolution of these problems.

In figure 10 equation (1) is expressed by two possible curves, B and C, representing different partial pressures of water. The possible form of a curve D, representing the breakdown of biotite to sillimanite, is shown. This is assumed to take place within the stability field of muscovite. Curves B, C, and D are schematically placed relative to the triple point (Clark, Robertson, and Birch, 1957) to agree with the petrographic evidence on paragenesis in the Burke area. There are no other data on actual values of temperatures and pressures for these curves. If the position of the triple point is altered, e.g., to the position suggested by Chinner, the three curves would also be shifted. Curve E represents the reaction in equation (5) after Segnit and Kennedy (1961). Curve F is the minimum melting curve of granite after Tuttle and Bowen (1958). The values of temperature and pressure are taken from the diagram in Clark, Robertson, and Birch (1957) and their implications have been discussed. The isobaric line on the diagram represents the metamorphic trends and assemblages observed in the Burke area as listed under petrography. Assemblages 2 through 4 are related to curve B, and assemblage 5 may be related to curve C. Assemblages 7 and 8 and 2 and 2 may be present because of compositional control; or assemblage 8 may be separated from 2 by a curve similar to curve B but displaced to the left because of a lower partial pressure of water. The fibrolite after biotite was a late development at higher temperatures and perhaps under local metasomatic influence.

Relict minerals provide evidence of incomplete reactions and suggest that the assemblages did not arise synchronously in spatially arranged zones but tended to develop successively in time as the temperature continued to rise.
Fig. 10. Phase diagram of Al₂SiO₅ polymorphs after Clark, Robertson, and Birch (1957) and Clark (1961). The dotted lines show the same relationships but after Schilling (1957): An—andalusite field, K—kyanite field, S—sillimanite field. Curve A is the geothermal gradient from the diagram in Clark, Robertson, and Birch (1957). Curves B to F and the isobaric line with numbers 1 to 8 are explained in the text.

This is particularly true of the metamorphic rocks near the granite and in the granite-hornfels complexes where favorable conditions for andalusite and sillimanite formation moved outwards and enveloped staurolite-bearing rocks.

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