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HIGH-LEVEL METAMORPHISM OF PELITIC ROCKS IN NORTHERN NEW HAMPSHIRE

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ABSTRACT

Lower Paleozoic pelitic rocks in the Errol quadrangle, northern New Hampshire and Maine, have been regionally metamorphosed to the biotite, garnet, staurolite and sillimanite zones. Andalusite is common in the staurolite zone, and began to form only slightly after staurolite during progressive metamorphism. Contact metamorphism by a granitic stock in the garnet zone produced staurolite, andalusite, cordierite and sillimanite in the surrounding schists. Retrograde metamorphism has altered some of the andalusite and staurolite and added chlorite to many assemblages. The mineral assemblages generally contain too many phases according to the phase rule, the excess being attributed to one or more of the following: retrograde reactions, incomplete prograde reactions, Mn and/or Ca in almandite, incomplete mobility of H_2O, or restricted volume of chemical equilibrium. Six new chemical analyses of coexisting almandite, staurolite and biotite are presented; their Fe/Mg ratios decrease in that order. The mineral assemblages are considered with the aid of Thompson's "AFM" projection. The temperatures of metamorphism are believed to have been less than about 500° C., and the pressures between 4 and 8 kilobars. The heat was principally derived from adjacent and subjacent granitic intrusions, superimposed on that conducted up from lower levels of the crust and mantle.

INTRODUCTION

In a recent study of the geology of the Errol quadrangle in northern New Hampshire and Maine (Fig. 1) the writer found the stratified rocks to show a large range in metamorphic grade. An investigation of the mineral assemblages of these rocks was initiated, with the intent of discovering the degree of approach to equilibrium and of elucidating the conditions of metamorphism. Because well over half of the quadrangle is underlain by metamorphosed pelitic sediments, the petrology of these rocks will be the major subject treated here. The mafic metamorphic rocks will be considered elsewhere.

The Errol quadrangle is underlain by metamorphosed stratified rocks of probable Cambrian to Devonian age, intruded by Devonian granitic stocks and part of a batholith (probably Ordovician) and minor Triassic (?) basaltic dikes. The stratigraphy and structure are treated in detail elsewhere (Green, *Geol. Soc. Amer., Spec. Paper*, in press). The metamorphic grade in the quadrangle ranges from the low biotite zone to the sillimanite zone (Fig. 1), and the metamorphism is believed to have taken place during the Acadian orogeny.

About 430 rock specimens were collected and studied from this area. From these samples 285 thin sections were examined; refractive indices and other optical properties were determined for a large number of min-

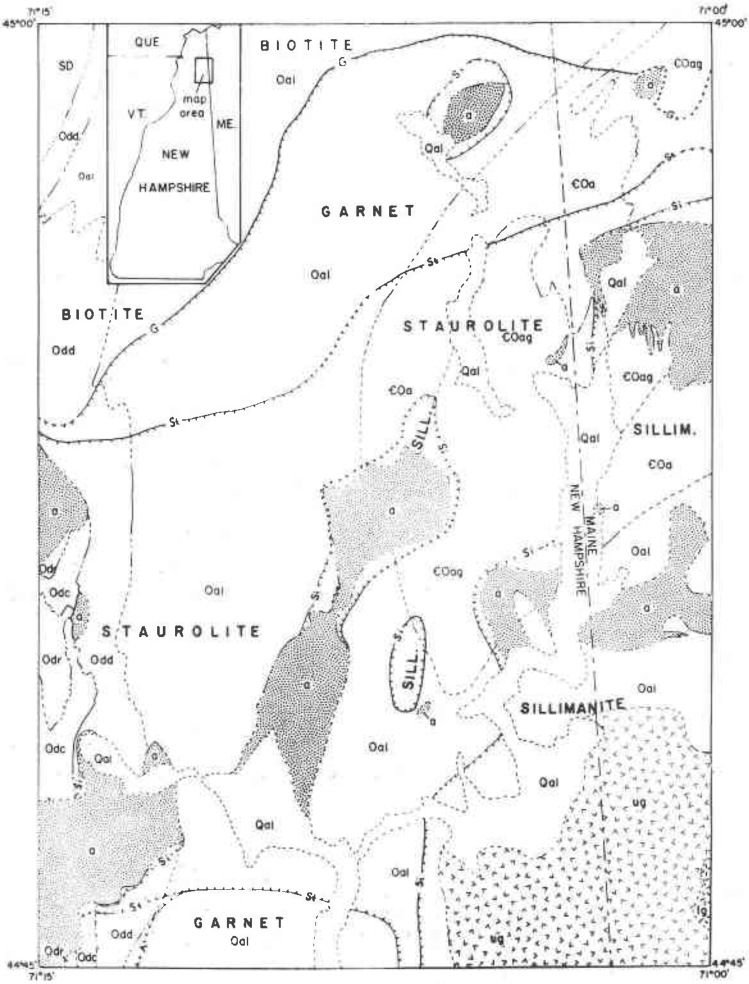

FIG. 1. Index map and geologic sketch map of Errol quadrangle, showing isograds. Formation symbols: SD, un-named Siluro-Devonian formation: Odd, Odc, and Odr, Dixville Formation; Oal, Albee Formation; €Oa and €Oag, Aziscohos Formation; Qal, alluvium; a, adamellite; ug, Umbagog Granodiorite; lg, leucocratic granite. Isograds: G, garnet; St, staurolite; Si, sillimanite.

erals; ten metamorphic minerals were separated for chemical analysis using heavy liquids, magnetic separator, and hand picking; and *x*-ray diffraction investigations were made of many specimens, mainly for mineral identification in very fine-grained rocks.

For the general approach in this study the writer is indebted to the

personal comments and lectures at Harvard University of Prof. James B. Thompson, Jr. The thin sections were supplied by the Department of Mineralogy and Petrography of Harvard University, and Jun Ito of that Department performed the chemical analyses. Field work during three summers (1955, 1956, 1957) was supported by the New Hampshire Planning and Development Commission, Harvard University, and Dartmouth College.

Argillaceous sediments undergo a great number of mineralogic changes, both continuous and discontinuous, during progressive metamorphism. The isograd method of Barrow (1893) and Tilley (1925) has long been used to describe the "grade" of metamorphism of phyllites and schists by noting the appearance of certain aluminous "index minerals." The occurrence of these minerals is partly a function of the bulk composition of the rock, and several petrologists (*e.g.*, Goldschmidt, Eskola, Turner) have attempted to deal with this problem by the use of multicomponent diagrams showing the mineral assemblages to be expected for different bulk compositions, as controlled by the phase rule, in different grades of metamorphism. Most of these diagrams, however, are constructed for rocks of basaltic or impure-colomitic composition, and most of them treat both MgO and FeO together as a single component. Thompson's graphical treatment for metamorphosed pelites (1957), by a series of reasonable assumptions and approximations, shows the compositions and associations of the major minerals in terms of three projected components: Al_2O_3, FeO and MgO. This approach removes the dangers of uncritical interpretation of the isograd system which were due to variations in bulk rock composition, and provides a much more detailed picture of the mineralogic changes constituting equilibrium metamorphism of pelitic rocks.

Theoretically it will be possible to assign specific physical conditions to many pelitic rocks using Thompson's projection once an admittedly great amount of experimental work is done, but not to all pelitic rocks since only a certain fraction of them satisfy the assumptions and approximations on which the projection is based. One object of the present study is to apply Thompson's projection to the pelitic rocks of the Errol quadrangle, and to determine to what extent these assumptions actually are justified in a thick sequence of supposedly "normal" metamorphosed eugeosynclinal sediments.

The isograd scheme of Barrow and Tilley is used in the present study, as it has been elsewhere in New Hampshire (*e.g.*, White and Billings, 1951; Lyons, 1955; Billings, 1956). Because it does not involve any special assumptions and requires merely a variety of aluminous sediments, the isograd scheme has been found to work well and to be of considerable use in studying the pelitic rocks in the present area. The use of isograds brings out broad zones of increasing metamorphic grade. Of the five classical index minerals used in this scheme (biotite, almandite, staurolite, kyanite and sillimanite) only four are found in the Errol quadrangle. Kyanite was not observed anywhere in the area. Other minerals which might be used as index minerals are chloritoid, andalusite and potassium feldspar. Of these only andalusite is present in these metamorphic rocks. The first appearance of andalusite and its general distribution correspond closely to the first appearance and general distribution of staurolite; hence an andalusite isograd would nearly coincide with the staurolite isograd, and is not separately indicated on most of the map (Fig. 1). The widespread coexistence of andalusite and staurolite places these rocks in the low-pressure intermediate group of metamorphic facies series of Miyashiro (1961).

PROGRESSIVE REGIONAL METAMORPHISM

General statement. According to the Gibbs phase rule ($f=c+2-\varphi$) the number of phases in equilibrium cannot exceed the number of components plus two. Since the maximum number of phases coexist only when the variance (f) is zero and the P and T are therefore fixed, the general case as pointed out by Goldschmidt (1911) for a closed system at a randomly specified P and T will be $\varphi \leq c$ (*i.e.*, the number of phases will be equal to or less than the number of components). This is known as Goldschmidt's mineralogical phase rule (Thompson, 1959).

Thompson's projection, which reduces the significant components to three, utilizes the following assumptions and restrictions: (a) quartz and muscovite are present in the rock; (b) CaO, Na_2O, MnO, Fe_2O_3 and TiO_2 are negligible in amount in the minerals appearing on the triangular projection; and (c) H_2O is able to move as a component into or out of the rock to form minerals of greater or lesser degree of hydration, according to an externally controlled activity (a_{H_2O}, essentially the same as the *humidity*, η; Thompson, 1957, p. 844). The remaining common oxides SiO_2, Al_2O_3, FeO, MgO, and K_2O, plus H_2O, form most of the metamorphic minerals of aluminous rocks and are the basic components used in the projection. Since only rocks that contain quartz and muscovite are considered, the system is in effect projected from the quartz and muscovite compositions, which gives a plane having three variable components represented by the corners of a triangle (Thompson, 1957, Figs. 1, 3, 4). No more than three of the minerals of this projection should be present at any random P, T and humidity.

Because quartz and muscovite are practically omnipresent in metamorphosed pelitic sediments, this restriction seems fully justified and does not materially detract from the applicability of the projection. The influence of the minor oxide components, CaO, Na_2O, MnO, Fe_2O_3, TiO_2, etc., each of which if significantly present in any of the minerals considered would allow an extra mineral according to the phase rule, is a major factor in limiting the usefulness of the projection, as is the mobility of H_2O during metamorphism.

Thompson's projection will be referred to below as the "AFM" projection. The graphical relationships in some parts of the triangle are uncertain because of the somewhat restricted bulk composition of the rocks in the area; the uncertainty is greatest for the three corners and the AF and AM edges.

The successive configurations for the "AFM" projection for the regional metamorphism in the Errol quadrangle, as deduced from petrographic observation, are shown in Figure 2. Observed mineral assemblages and compositions for successively increasing metamorphic grade are shown in the figures following. In these figures mineral compositions determined optically are shown by dots, those determined by chemical analysis by crosses. Observed mineral assemblages are shown by tie-lines. Because no potassic feldspar was found in these rocks, the projection is cut off below the biotite composition. Garnets, whose compositions were estimated by refractive index, are assumed to be pyralspites; the position on the diagram shows the $(Fe^{VI}+Mn)/(Fe^{VI}+Mn+Mg)$ ratio. Biotite compositions were determined using the β index and Winchell's diagram (1951, p. 374), assuming the formula $K(Fe,Mg)_{2\frac{2}{3}}Al_{\frac{1}{3}}(Al_{1\frac{1}{3}}Si_{2\frac{2}{3}}O_{10})(OH)_2$ as suggested by Thompson (oral comm.) Chlorite compositions were determined by a graph showing the variation of β with the ratio $(FeO+2Fe_2O_3+MnO)/(FeO+2Fe_2O_3+MnO+CaO+MgO)$ (Albee, 1962).

Biotite zone. It is apparent from Fig. 3 that the phase relations in the biotite zone are extremely simple as far as pelitic rocks are concerned. Among the minerals of this system only chlorite and biotite, and in many cases chlorite alone are found with quartz and muscovite in these rocks. In the three assemblages studied the biotite has a higher Fe/Mg ratio than coexisting chlorite, in agreement with Ramberg's statement (1952, p. 141). Eight phyllites and slates were analyzed by *x*-ray diffractometer for the presence of aluminous phyllosilicates, but no trace of pyrophyllite or paragonite was found. Aluminum-rich rocks of this metamorphic grade might be expected to contain pyrophyllite, and its absence suggests that the bulk composition of these rocks lies to the low-alumina side of the chlorite composition on this projection. It is possible that

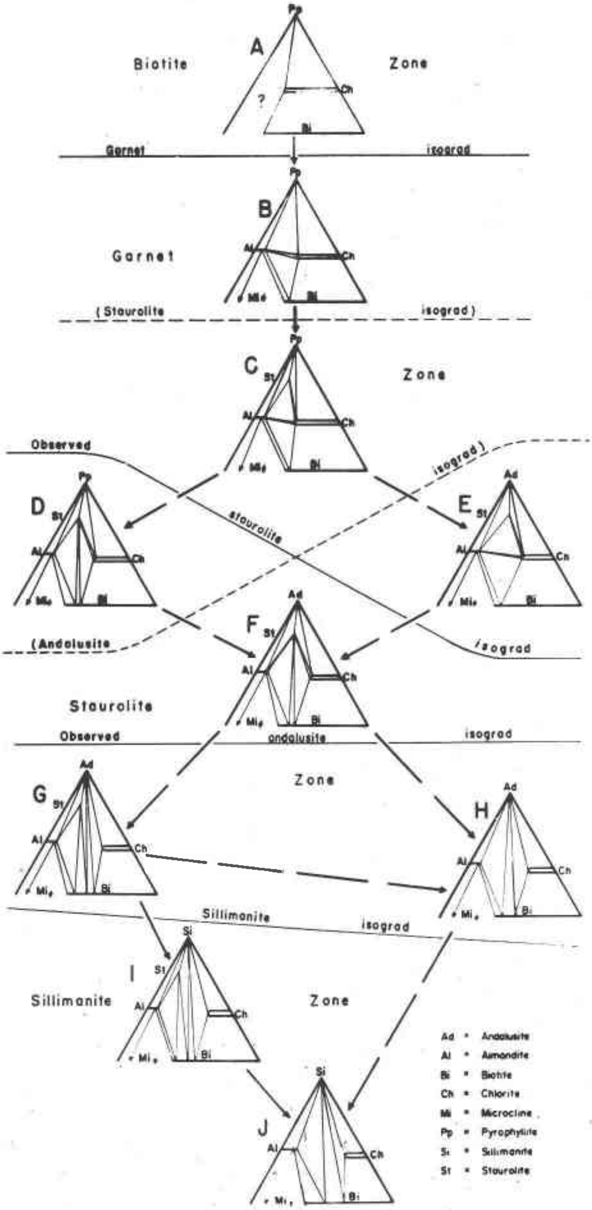

FIG. 2. Deduced sequence of mineral facies in regional metamorphism of pelites, Errol quadrangle. Thompson's "AFM" projection, quartz and muscovite present.

some kaolinite may be present, but its *x*-ray pattern would have been masked by that of chlorite, and it is not anticipated in rocks of this grade.

A few small, poikilitic garnets were found in phyllites a mile or more northwest of the mapped garnet isograd (within the biotite zone), and they are conspicuous in a peculiar quartz-andesine-garnet-hornblende granofels which also occurs in the area "below" the garnet isograd. Lithologically similar garnetiferous granofels was found by chemical analysis to have a greater-than-average manganese content (MnO =0.85%), which implies that the garnet in these rocks (the analyzed specimen is 8% garnet) is somewhat spessartitic. Manganese has long been known to lower the metamorphic grade at which garnet first appears. Because the garnet isograd is based on the first appearance of almandite, it is justifiable to place these rocks in the biotite zone. However, it is difficult to determine quickly the whole composition of any garnet, and even those which do appear at the garnet isograd probably contain a small amount of manganese as well as calcium and magnesium. The observed fact that the first appearance of macroscopic, euhedral garnet does follow a fairly smooth line across the area indicates that even if these garnets are not entirely free of the spessartite component, they

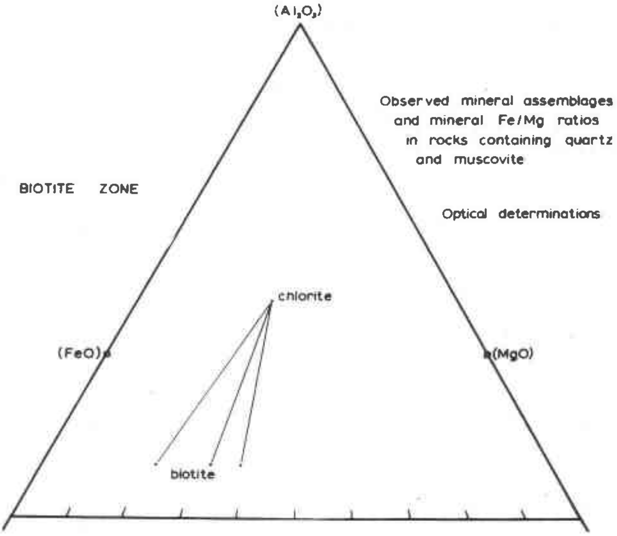

FIG. 3. Observed mineral assemblages and Fe/Mg ratios, biotite zone. Compositions determined by refractive index. Thompson's "AFM" projection.

probably all contain approximately the same amount, and therefore form a useful isograd.

Garnet zone. Figure 4 shows mineral assemblages for aluminous rocks of the garnet zone for which Fe/Mg ratios have been determined optically. There is no evidence from these data that the compositions of any of the minerals tend to change with increasing metamorphic grade. However, chlorites in equilibrium with garnet and biotite in the staurolite zone (Fig. 5) have a lower Fe/Mg ratio than those in the garnet zone. Recent experimental work (Turnock, 1959) has shown that the substitution of Fe for Mg markedly lowers the maximum temperature at which chlorite can occur. In other words, at lower temperatures chlorites can hold more iron than at higher temperatures. This is in agreement with the observations cited above. The three-phase assemblage garnet-biotite-chlorite (plus quartz and muscovite) is very common, as is each of the two-phase assemblages bounding it. In all determined pairs, the Fe/Mg ratio decreases from garnet to biotite to chlorite. Magnetite porphyroblasts occur in many outcrops throughout the garnet zone and are found in association with garnet and chlorite, with chlorite and biotite, with chlorite alone, and commonly with all three. Chloritoid was never observed (except in glacial erratics) in the area. The aluminum silicate

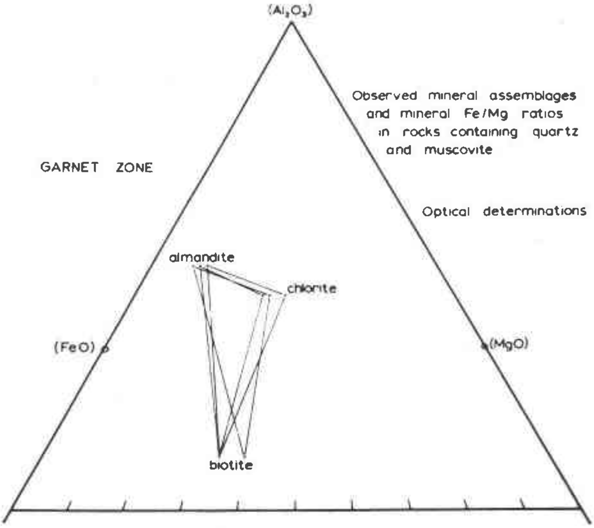

FIG. 4. Observed mineral assemblages and Fe/Mg ratios, garnet zone. Compositions determined by refractive index. Thompson's "AFM" projection.

most likely present in highly aluminous rocks of this metamorphic grade is pyrophyllite. Its absence (twelve samples were checked by x-ray diffractometer) again implies that the bulk compositions of these pelites lie below the line traced by the garnet and chlorite compositions and the most aluminous garnet-chlorite tie-line on the projection. The lack of chloritoid may also be due to this lack of highly aluminous bulk compositions.

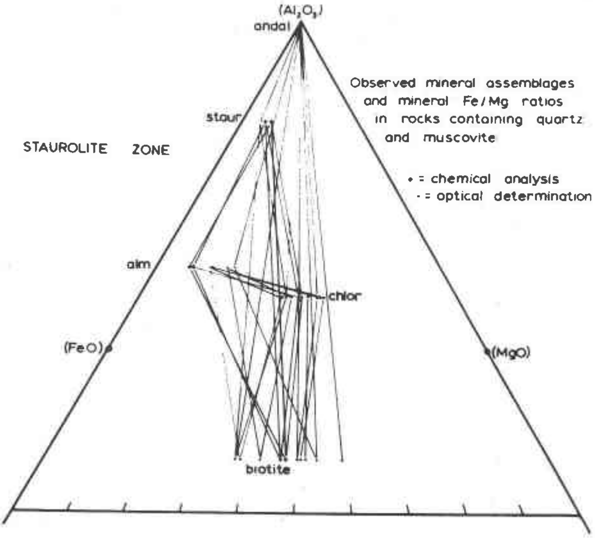

FIG. 5. Observed mineral assemblages and Fe/Mg ratios, staurolite zone. Compositions marked + determined by chemical analysis; those marked · determined by refractive index; others not determined. Thompson's "AFM" projection.

Staurolite zone. The phase relations in the staurolite zone are much more complicated than in lower zones. Two new minerals appear, staurolite and andalusite, but it is not certain from the regional distribution which of these forms at the lower grade. Textural evidence discussed below suggests that staurolite formed first. Staurolite is the more widespread, and in a large area in the west-central part of the quadrangle it occurs where andalusite has not been found, but in other, more restricted areas (*e.g.*, near the southwest corner of the map) andalusite is present where staurolite is not. In general the staurolite and andalusite isograds coincide, but they have been drawn separately in the southwest part of the area. The fact that some local areas were found to contain only one of these two minerals may be the result either of insufficient sampling, differences in bulk rock composition, or differences in metamorphic conditions.

Minerals in the "AFM" projection that are present in the staurolite zone in the Errol quadrangle are garnet, staurolite, andalusite, biotite, and chlorite; practically all combinations of these are observed. Assemblages containing only one of these minerals (with quartz and muscovite) are rare; two-phase assemblages are common but not abundant; three-phase and four-phase assemblages are very abundant; and five-phase assemblages are locally observed. Observed mineral assemblages in the staurolite zone, all seemingly in equilibrium by textural criteria (no mantling, embayment, etc.) are as follows (in addition, all contain quartz and muscovite):

Biotite.
Biotite, chlorite.
Biotite, garnet.[1]
Biotite, andalusite.
Biotite, garnet, chlorite.[1]
Biotite, garnet, staurolite.[1]
Biotite, garnet, andalusite.
Biotite, staurolite, andalusite.
Biotite, chlorite, andalusite.
Biotite, staurolite, chlorite.[1]
Biotite, staurolite, garnet, chlorite.[1]
Biotite, andalusite, garnet, chlorite.
Biotite, staurolite, andalusite, chlorite.
Biotite, staurolite, garnet, andalusite, chlorite.

Biotite is present in all the rocks, whereas the several possible assemblages lacking biotite are not represented. As in the lower-grade zones, these assemblages imply bulk rock compositions that are not highly aluminous. Figure 2 shows the probable successive equilibrium configurations ("mineral facies," Thompson, 1957) of the projection from the biotite zone into the sillimanite zone, as inferred from petrographic study in this area. Because of the restricted bulk composition of the rocks as just mentioned, the staurolite-andalusite isograd in the Errol quadrangle is probably displaced up-grade (between triangles C and D, and E and F, Fig. 2) from its ideal position (between B and C) in more aluminous rocks; the isograd coincides with the first appearance of the pair staurolite-biotite. The garnet-biotite-chlorite assemblage, for example, is generally speaking ambiguous, and can occur in both the garnet and staurolite zones in more aluminous rocks. In the Errol quadrangle its occurrence in the staurolite zone probably means lack of conformity to the restrictions of the "AFM" projecion (see below).

Assemblages for which the mineral compositions have been determined are plotted in Fig. 5. Most of the compositions were determined optically,

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[1] Most common assemblages.

but the biotite, garnet, and staurolite of two biotite-garnet-staurolite-chlorite assemblages were separated and chemically analyzed; their compositions are marked by crosses on the diagram, and the analyses are listed in Table 1. In all cases the Fe/Mg ratios of these minerals decrease from garnet to staurolite to biotite to chlorite, in agreement with Thompson's assumed order (1957, p. 857).

Biotite, garnet and chlorite show a considerable range in composition among these assemblages.

> "The compositions at the corners of the three-phase fields will, in general, show *continuous variation* from place to place in response to variation in P, T, and (a_{H_2O}) at the time of formation of the rocks" (Thompson, 1957, p. 855).

An attempt was made to correlate such changes in composition indicated in Figure 5 with relative distance from the staurolite or sillimanite isograds, but the variations appear to be random; the specimen localities, taken in the order of composition of their biotite and chlorite (which in general do vary together), were scattered without system throughout the staurolite zone. Retrograde reactions may have erased any real original variation.

Because of the general tendency in progressive metamorphism to form less and less hydrous assemblages, the stability field of chlorite, the most hydrous of the phases in this system, would be expected to decrease, thus making chlorites found with two other phases more and more magnesian. An unknown amount of the chlorite in these rocks may be the product of reactions subsequent to the climax of metamorphism. In such rocks, chlorite-bearing assemblages are probably not in equilibrium. Note that all the four- and five-phase assemblages observed contain chlorite. In most rocks the chlorite occurs as clean, platy to sheaf-like porphyroblasts that cut across the schistosity without affecting its attitude. As discussed below, this implies either that the chlorite kept in perfect textural equilibrium with the rest of the rock throughout prograde metamorphism, or that it is the product of late-stage, probably retrograde reactions.

In several of the asemblages plotted, the composition of the garnet is not specified because of the difficulty involved in its determination. The two analyzed garnets from the staurolite zone (D-2, ER-101b) each contain 6 mol. per cent grossularite, and 8.3 and 10.3 mol. per cent spessartite respectively. The position plotted is the Fe/(Fe+Mg) ratio.

Figure 2 referred to above shows the inferred sequence of mineral facies in passing from the garnet zone through the staurolite zone and into the sillimanite zone in the Errol area. Because the distributions of staurolite and andalusite so nearly coincide, the observed andalusite isograd (attained by stabilization of the pair andalusite-biotite) must

Table 1. Chemical Analyses and Physical Properties of Coexisting Staurolite, Garnet and Biotite, Errol Quadrangle[1]

No.	Staurolite		Garnet			Biotite		
	D-2	ER-101b	D-2	ER-101b	ER-142	D-2	ER-101b	ER-142
SiO_2	28.18	28.11	36.68	37.26	35.98	34.95	34.84	35.45
TiO_2	0.75	0.67	0.24	0.55	0.17	1.94	1.79	1.76
Al_2O_3	52.98	53.56	21.49	21.23	22.56	20.71	20.98	21.13
Fe_2O_3	2.89	1.89	0.34	0.51	1.20	—	—	—
FeO	11.16	10.78	32.39	31.45	30.57	19.76	19.40	19.64
MnO	0.09	0.20	3.59	4.39	6.34	0.06	0.05	0.13
MgO	1.60	2.10	2.85	2.64	1.93	9.69	10.01	9.20
CaO	—	0.19	2.07	1.94	1.18	—	—	—
Na_2O	0.08	0.32	—	—	0.10	0.65	0.24	0.31
K_2O	—	—	—	—	0.14	8.87	8.93	8.95
H_2O^+	1.78	1.60	0.32	0.25	0.24	3.21	3.68	3.81
H_2O^-	0.32	0.29				—	—	—
Total	99.83	99.71	99.97	100.22	100.41	99.84	99.92	100.38
αNa	1.743	1.743						
βNa	1.751	1.752	1.809	1.807	1.809	1.639	1.632	1.631
γNa	1.758	1.759						
$2V_{\text{cacl.}}$	−87°	−84°						
a (in Å)			11.53	11.54	11.52			
β associated chlorite	1.626	1.627						

[1] Analyses by Jun Ito, 1957, Harvard University.

Descriptions of rocks and calculated formulas of analyzed minerals

D-2: Staurolite-garnet-biotite-chlorite-muscovite-quartz schist; Diamond Gorge, on road near milepost 2; Aziscohos formation

D-2 Staurolite: $(\mathrm{Fe}^{2+}_{1.30}\mathrm{Mn}_{.01}\mathrm{Mg}_{.34}\mathrm{Ti}_{.08}\mathrm{Fe}^{3+}_{.27})_{2.00}(\mathrm{Fe}^{3+}_{.03}\mathrm{Al}_{8.63})_{8.66}(\mathrm{Al}_{.09}\mathrm{Si}_{3.91})_{4.00}\mathrm{H}_{1.68}\mathrm{O}_{24}$

D-2 Garnet: $(\mathrm{Fe}^{2+}_{2.18}\mathrm{Mn}_{.25}\mathrm{Mg}_{.34}\mathrm{Ca}_{.18}\mathrm{Ti}_{.02}\mathrm{Fe}^{3+}_{.02})_{2.99}\mathrm{Al}_{2.00}(\mathrm{Al}_{.04}\mathrm{Si}_{2.96})_{3.00}\mathrm{O}_{12}$

D-2 Biotite: $(\mathrm{K}_{.87}\mathrm{Na}_{.09})_{.96}(\mathrm{Fe}^{2+}_{1.25}\mathrm{Mn}_{.01}\mathrm{Mg}_{1.11}\mathrm{Ti}_{.11}\mathrm{Al}_{.52})_{3.00}(\mathrm{Al}_{1.34}\mathrm{Si}_{2.66})_{4.00}\mathrm{O}_{10}(\mathrm{OH})_{1.63}$

ER-101b: Silvery staurolite-garnet-biotite-chlorite-muscovite-quartz schist; 1540′ in Corser Brook; Albee formation

ER-101b Staurolite: $(\mathrm{Fe}^{2+}_{1.31}\mathrm{Mn}_{.02}\mathrm{Mg}_{.45}\mathrm{Ti}_{.07}\mathrm{Fe}^{3+}_{.15})_{2.00}(\mathrm{Fe}^{3+}_{.06}\mathrm{Al}_{8.74})_{8.80}(\mathrm{Al}_{.24}\mathrm{Si}_{3.76})_{4.00}\mathrm{H}_{1.65}\mathrm{O}_{24}$

ER-101b Garnet: $(\mathrm{Fe}^{2+}_{2.10}\mathrm{Mn}_{.30}\mathrm{Mg}_{.32}\mathrm{Ca}_{.17}\mathrm{Ti}_{.03}\mathrm{Fe}^{3+}_{.03}\mathrm{Al}_{.01})_{2.96}\mathrm{Al}_{2.00}\mathrm{Si}_{3.00}\mathrm{O}_{12}$

ER-101b Biotite: $(\mathrm{K}_{.88}\mathrm{Na}_{.04})_{.92}(\mathrm{Fe}^{2+}_{1.24}\mathrm{Mn}_{.01}\mathrm{Mg}_{1.16}\mathrm{Ti}_{.10}\mathrm{Al}_{.49})_{3.00}(\mathrm{Al}_{1.41}\mathrm{Si}_{2.59})_{4.00}\mathrm{O}_{10}(\mathrm{OH})_{1.87}$

ER-142: Gray biotite-garnet-sillimanite-andesine-muscovite-quartz gneiss; 1585′ in Meadow Brook;

ER-142 Garnet: $(\mathrm{Fe}^{2+}_{2.13}\mathrm{Mn}_{.45}\mathrm{Mg}_{.24}\mathrm{Ca}_{.10}\mathrm{Ti}_{.01}\mathrm{Fe}^{3+}_{.04}\mathrm{Al}_{.06})_{3.03}\mathrm{Al}_{2.00}(\mathrm{Al}_{.14}\mathrm{Si}_{2.86})_{3.00}\mathrm{O}_{12}$

ER-142 Biotite: $(\mathrm{K}_{.88}\mathrm{Na}_{.04})_{.92}(\mathrm{Fe}^{2+}_{1.23}\mathrm{Mn}_{.02}\mathrm{Mg}_{1.05}\mathrm{Ti}_{.11}\mathrm{Al}_{.59})_{3.00}(\mathrm{Al}_{1.30}\mathrm{Si}_{2.70})_{4.00}\mathrm{O}_{10}(\mathrm{OH})_{1.94}$

have been reached only shortly after the staurolite isograd. Triangle G shows the most probable stable assemblages over most of the staurolite zone. However, the assemblage garnet-andalusite-biotite, as in triangle H, was found at several localities. In such rocks the decomposition of staurolite was accomplished before the andalusite-sillimanite transition occurred. This may have been caused by physical conditions slightly different from those of the surrounding rocks of the staurolite zone, or it may be due to chemical abnormalities such as an unusual amount of Mn compared to Fe^{2+} and Mg in the rock. The excess Mn might have stabilized enough garnet to use up all the Fe and Mg which otherwise would have entered staurolite.

Sillimanite zone. The sillimanite zone covers a broad area in the east-central part of the quadrangle, a wide zone around the large granodiorite pluton, narrow zones around the adamellite stocks, and irregular areas elsewhere within the limits of the staurolite zone. In certain places in the wider sillimanitic areas, the schists and gneisses are penetrated by or mixed with varying proportions of granitic to tonalitic stringers, lenses and dikes, with the result that the rock becomes best described as a migmatite.

The minerals of the sillimanite zone are the same as those of the staurolite zone, with the addition of sillimanite. The mineral facies of the sillimanite zone are shown in Fig. 2, and Fig. 6 shows the compositions of observed mineral assemblages, determined by refractive indices and by chemical analyses of a biotite and a garnet that coexist with sillimanite (ER-142; Table 1). Observed mineral assemblages in the sillimanite zone are listed below (in addition, all contain quartz and muscovite).

Biotite.
Biotite, garnet.
Biotite, sillimanite.[1]
Biotite, garnet, sillimanite,[1] ±(andalusite).
Biotite, garnet, sillimanite, chlorite.
Biotite, garnet, sillimanite, staurolite, ±(andalusite).
Biotite, chlorite, sillimanite, staurolite,[1] ±(andalusite).
Biotite, chlorite, sillimanite, ±andalusite.
Biotite, staurolite, sillimanite, (andalusite).

The phase relations in the sillimanite zone are only slightly simpler than in the staurolite zone. Only two discontinuous changes are involved: the polymorphic transition of andalusite to sillimanite and the disappearance of staurolite. Staurolite was not seen close to the large granodiorite pluton, but only in the outer part of its aureole and elsewhere in the sillimanite zone. The three-phase assemblage garnet-biotite-sillimanite is

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[1] Most common assemblages. Parentheses indicate anadalusite rimmed by muscovite.

stable relative to staurolite in the highest grade of metamorphism reached in the area.

Andalusite, where present in the sillimanite zone, is in nearly all cases surrounded by a rim of coarse muscovite, and the sillimanite occurs independently as fibrolite in muscovite, biotite, chlorite, quartz, garnet and plagioclase, or as coarse, prismatic crystals. In only one rock was any sillimanite seen in direct contact with fresh andalusite, and there most of the sillimanite was included in muscovite and chlorite. Because of their polymorphic relationship, only one of these two aluminum silicates should be found in the same rock, but the transitions between kyanite, andalusite and sillimanite are well known to be sluggish, and the andalusite in the present rocks must be counted as a relict phase out of equilibrium with the rest of the rock. It is interesting that the sillimanite has not grown as pseudomorphs of older andalusite, but as independent grains. In a study of the growth of sillimanite in the Scottish Highlands Chinner (1961) finds a similar relationship except that the lower-temperature polymorph is kyanite. He believes that the sillimanite was nucleated by the atomic structure of mica, particularly biotite, and this explanation would fit most of the sillimanite in the Errol rocks. However, it does not explain the not uncommon occurrence of fibrolite in quartz, plagioclase

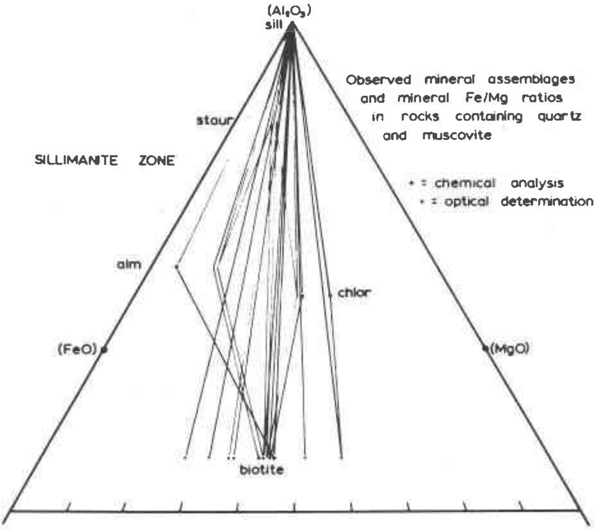

FIG. 6. Observed mineral assemblages and Fe/Mg ratios, sillimanite zone. Compositions marked + determined by chemical analysis; those marked · determined by refractive index; others not determined. Thompson's "AFM" projection.

and garnet. A curious fact is that whereas the andalusite in the sillimanite zone has nearly everywhere been centripetally altered to muscovite, the sillimanite seems to have escaped. This may be due to its being armored by the other minerals in which it lies. It is difficult to imagine how the coarse muscovite rims on the andalusite could have formed except by retrograde reactions; the major prograde reaction within the sillimanite zone (staurolite→garnet+biotite+sillimanite) involves the destruction of muscovite. One possible explanation is that the stability field of sillimanite was not entered until after both the production of andalusite and its partial muscovitization. This would imply two somewhat distinct pulses of metamorphism, separated by either a lowering of temperature or a rise in the humidity, and very probably an influx of potassium (Green, *Jour. Geol.*, in press).

Figure 6 shows that the biotite and chlorite found together in the sillimanite zone have very similar Fe/Mg ratios, but that the biotite is never more magnesian than the coexisting chlorite. Chlorite in the sillimanite zone is in general more magnesian than in lower zones (Figs. 4 and 5), but none was found to have an Fe/(Fe+Mg) ratio of less than 40 per cent. Much of it, but not necessarily all, may be the result of retrograde metamorphism. Thompson (1957, Fig. 5) does not show any chlorite in the "lower sillimanite zone of west-central New Hampshire" (p. 851), but the highly magnesian (but muscovite-bearing) rocks that might show it are not common. The biotite-sillimanite pair is found over a large range of Fe/Mg ratios, including the area where garnet would be expected to form and indeed does form in other rocks. The writer has no explanation for the lack of garnet in these rocks, unless it be that the garnet of this Fe/Mg ratio actually requires for its formation some minimum amount of Ca and/or Mn that is not available in these particular rocks. The analyzed garnet of a garnet-biotite-sillimanite assemblage (ER-142) contains only 3.6 mol. per cent grossularite but 15.3 mol. per cent spessartite.

The occurrence of the metamorphic minerals in each metamorphic zone is summarized in Fig. 7.

Reasons for excess phases. Assuming that equilibrium has been maintained during the progressive metamorphism of these schists, there must be some explanation for the observed excess of phases over the theoretical maximum of three, assuming random P, T and humidity. Actually, there are several possible factors which could tend to exempt these rocks from strict conformity to this rule.

1). A discontinuous change in configuration of the projection (appearance or disappearance of a mineral, or a cross-over in tie-lines to give different pairs of compatible minerals among a four-mineral group) would give an extra phase to a previous assemblage. The P, T and humidity at which this can occur are not, however, independent because the equilibrium is divariant. Transitions of this sort are ideally suited for use as isograds (given rocks of appropriate bulk composition) for the very reason that such a reaction can occur only on a specific plane which intersects the topographic surface in a line. Regions between such isograds are characterized by P, T and humidity conditions which vary independently over a certain range, thus reducing, not increasing, the number of phases. Because the extra-phase assemblages in the schists of the Errol quadrangle are widespread between the Barrovian isograds, they cannot be attributed to such discontinuous reactions of progressive metamorphism as long as equilibrium is assumed to have been maintained. The existence of assemblages such as staurolite-garnet-biotite-sillimanite, which strongly imply such discontinuous changes, points instead to a lack of equilibrium in the form of incomplete consummation of such reactions.

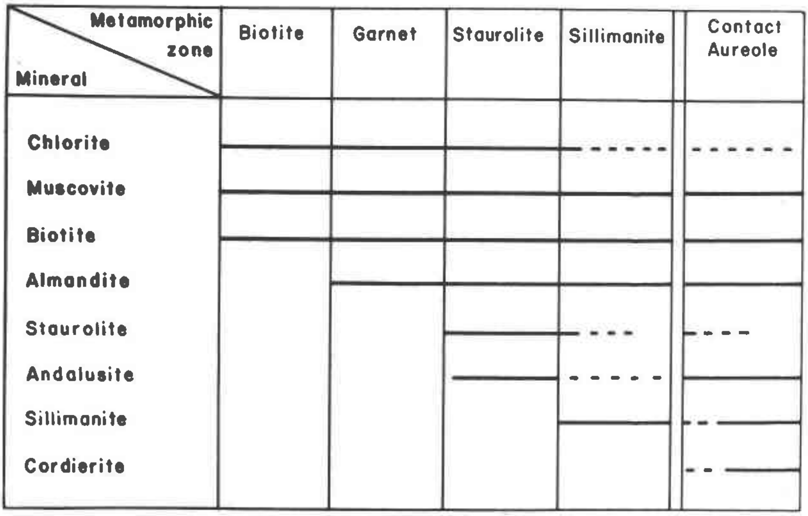

Fig. 7. General occurrence of pelitic metamorphic minerals in Barrovian zones, Errol quadrangle. Dashed bar shows metastability as indicated by resorption, or retrograde formation.

2). Incomplete polymorphic transitions, such as that from andalusite to sillimanite, would produce an excess phase, but the assemblage would of course not be in equilibrium. This pair is common throughout much of the sillimanite zone, but the andalusite, as just mentioned, nearly everywhere has been at least partially replaced by muscovite.

3). It has already been pointed out that the common garnet in these rocks is not a pure Fe-Mg garnet but contains appreciable proportions of Ca and Mn. Both of these elements tend to remove the garnet from the plane of the projection. Because the garnet structure readily takes in Ca and especially Mn, almandite garnet is stabilized beyond the range of conditions and bulk compositions in which it would occur were its only divalent metals Fe and Mg. This is possibly a reason for the large number of four- and five-phase assemblages involving garnet. The more "impure" in this respect the garnet becomes, the more assemblages it can occur with, as its field of stability expands toward the Al-Mg side of the projection. Miyashiro (1953) suggests that almandites in regions of low-pressure metamorphism (characterized by andalusite instead of kyanite) typically contain relatively large amounts of Mn. Although the analyzed garnets from the Errol quadrangle do not contain as much Mn (Table 1) as Miyashiro proposes, it seems likely from his suggestion as well as from the analyses that all the garnets in the Errol area contain appreciable Mn. It would be interesting to find if garnet-bearing assemblages in other, and still lower-pressure metamorphic areas, contain excess phases that cannot be attributed to retrograde reactions or other factors. Actually as discussed below, most of the excess-phase assemblages in the Errol quadrangle contain chlorite, which may be retrograde in origin. If, as appears probable, this is so, the most common mineral assemblages are reduced to two- or three-phase assemblages, and the garnet need not be counted as an excess phase. It thus appears that the compositions of the garnets in this area are not in general far enough from the almandite-pyrope system to stabilize an excess phase.

4). Although many of the excess-phase assemblages show no textural evidence of reaction or disequilibrium, others do, in the form of chlorite rims around staurolite and andalusite, and muscovite rims around andalusite. Chlorite rims on garnet are rare. Chlorite, mixed both with anhedral oxide minerals (rutile, magnetite, or ilmenite) and with minute lenses of potassic feldspar (?), is also found in some rocks intergrown with and pseudomorphous after biotite. The chlorite in many schists, however, occurs as large, clean porphyroblasts, showing no textural relationships to other ferromagnesian minerals. Since these cut across the schistosity, do not deform it and do not appear to have been rotated or otherwise strained, these too may be retrograde products. Since most of the excess-phase assemblages contain chlorite as one of the phases, this is a likely explanation, especially in the four-phase assemblage lacking garnet.

5). An extra hydrous phase such as chlorite may also be stabilized if the content of H_2O is locally fixed, that is, if the H_2O component is not completely free to migrate away from the scene of a prograde, dehydrating reaction. In such a case the humidity would be internally determined, instead of externally controlled. This possibility would be difficult to test petrographically, and really amounts to a particular case of (1) above.

Some particular examples of excess-phase assemblages might here be considered. Four other phases besides andalusite are present in many rocks of the sillimanite zone. Of five rocks examined which show the four-phase assemblage garnet-chlorite-biotite-sillimanite, only one is free of textural evidence of disequilibrium. In two rocks some of the garnet is rimmed by chlorite, and in two others biotite is partly altered to chlorite. Retrograde reactions clearly account for an excess phase (chlorite) in these four rocks. The fifth, with no rimming or pseudomorphic textures, contains fresh and fairly coarse grains, and the excess phase may be due either to Mn and/or Ca in the garnet or to independent growth of retrograde chlorite porphyroblasts.

Of the five sections studied that contain the assemblage staurolite-chlorite-biotite-sillimanite, only one seems free of reaction textures. The others show biotite or staurolite altering to chlorite, or staurolite armored by andalusite. In the latter rock, since the andalusite is obviously relict and out of equilibrium with the surrounding assemblage, the staurolite within it may well be also.

The assemblage staurolite-garnet-biotite-sillimanite cannot be attributed to retrograde reactions. In one such rock the staurolite is included within larger andalusite crystals, suggesting that it is a relict phase. However, two rocks with this assemblage show fresh, euhedral, un-mantled staurolite. Abnormally high Mn and/or Ca in the garnet might account for the excess phase in these rocks, but the true explanation is unknown.

Isochemical Contact Metamorphism

In the north-central part of the quadrangle (Fig. 1) a small stock of biotite-muscovite adamellite has penetrated metasedimentary phyllites and schists of the garnet zone. In a zone about one-eighth to one-fourth mile thick surrounding the stock, the phyllites have been converted to schists and gneisses containing the following minerals characteristic of considerably higher grades of metamorphism: andalusite, staurolite, sillimanite and cordierite. No hornfels, in the sense of a fine-grained, hard, granular, baked contact rock, is present; even up to the knife-sharp contact, the metasediments are medium-grained and micaceous, although some of their schistosity is reduced by the haphazard crystallization of andalusite prisms, knots of cordierite, and sericite flakes. The adamellite is not chilled at the contact.

Andalusite occurs throughout the contact zone, even in rocks containing sillimanite. In some outcrops the andalusite prisms are rimmed by biotite in a way that suggests growth of the andalusite rather than its alteration to biotite. Staurolite is widespread but not abundant in the contact zone. It occurs as small, anhedral grains but has no mica or chlorite rims. In one section all the staurolite was enclosed in andalusite.

Cordierite was identified in only two thin sections, but chlorite pseudomorphs after cordierite are common in the phyllites and schists of the contact zone. In thin section the cordierite appears as extremely poikilitic, equant to oblong masses about 2 to 3 mm in diameter, containing minute flakes of muscovite (similar to those in the schist groundmass) and abundant grains of magnetite. Some grains show the characteristic sector twinning, and it is rimmed by a yellowish-brown chloritic mineral.

The pelitic rocks of this aureole are characterized by a surplus of minerals over the equilibrium number which should occur at random P, T and humidity. Observed mineral assemblages are as follows (in addition, each contains quartz and muscovite):

Andalusite, sillimanite, garnet, biotite.
Andalusite, sillimanite, chlorite, biotite.
Andalusite, staurolite, garnet, chlorite.
Andalusite, staurolite, garnet, biotite.
Staurolite, biotite, chlorite, cordierite.
Andalusite, staurolite, garnet, biotite, chlorite.[1]
Andalusite, sillimanite, staurolite, biotite, cordierite.

The ubiquitous occurrence of four or five phases of the projection in these rocks indicates either chemical deviation of phases such as garnet from the simplified system, restricted mobility of H_2O, or the non-attainment of equilibrium. Each of these factors may play a part. The coexistence of andalusite and sillimanite implies non-attainment of equilibrium during the polymorphic transition, unless the andalusite, much of which has pleochroic (pink) cores with anomalous interference colors, contains enough Mn or Fe^{3+} to stabilize it outside its normal limits. The irregular, "resorbed" appearance of the staurolite suggests that the upper stability limit of that mineral was reached but that it was not entirely converted to garnet, aluminum silicate, and biotite. The configurations of the triangles in Fig. 8 show the general order of the mineral reactions in this contact zone and several possible sequences of reactions by which the observed assemblages might have been produced assuming that equilibrium was maintained. Because of the extra phases it cannot be established with certainty which of the indicated paths was followed, but

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[1] Most commonly observed assemblage.

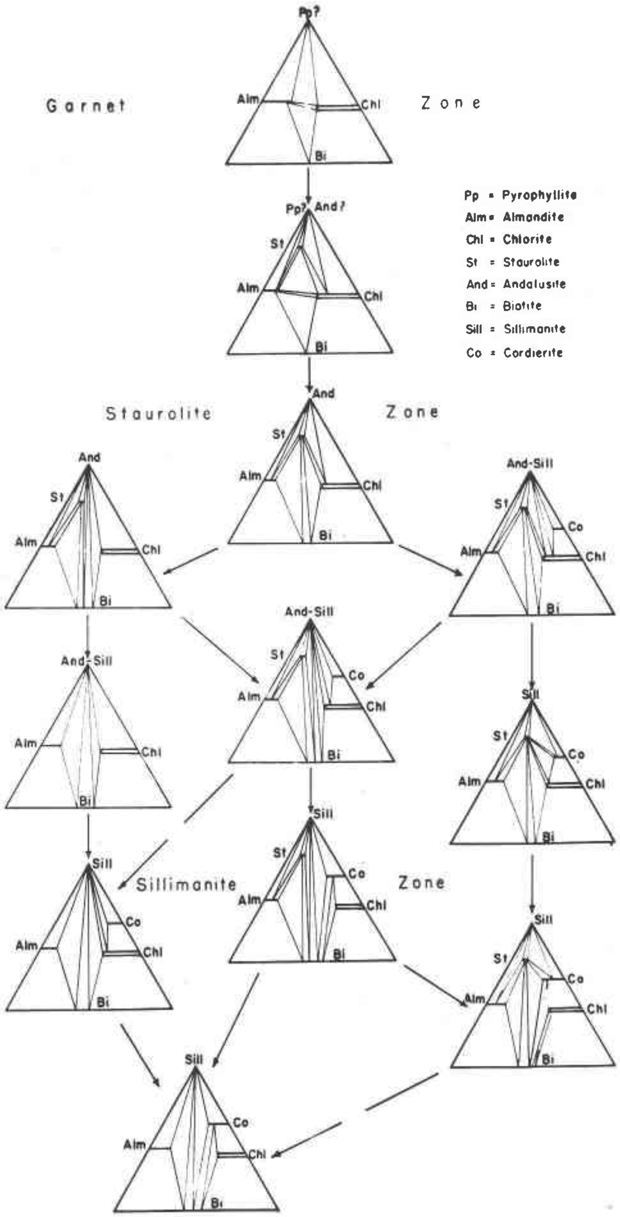

Fig. 8. Deduced sequence of mineral facies in contact metamorphism of pelites, Errol quadrangle.

since both of the observed cordierite-bearing assemblages contain staurolite and biotite, the right-hand paths seem more likely than the paths on the left. It is probable that equilibrium was never actually attained, in which case none of the sequences shown in Fig. 8 may have been strictly followed. Some of the chlorite is certainly attributable to retrograde reactions.

Aside from the presence of cordierite, the major difference between the rocks of this contact-metamorphic zone and those regionally metamorphosed to a comparable grade elsewhere in the area is a difference in the attainment of equilibrium, as shown by the number of coexisting minerals in the "AFM" diagram. In the contact zone four- and five-phase assemblages are the rule, whereas in the regionally metamorphosed rocks three- and four-phase assemblages are most common. This difference implies a difference in the conditions of metamorphism, a difference not in pressure or temperature themselves, but in the rate of increase of temperature and in the mobility of H_2O. The thermal gradient from the intrusive body into the phyllites must have been steep, producing rapid heating of the adjacent rocks, whereas the thermal gradient in the regionally metamorphosed rocks must have been very gentle, as indicated by the wide separation of the isograds.

The rapid heating of the rocks in the contact zone produced mineral reactions tending toward much less hydrous assemblages (staurolite, cordierite, and aluminum silicate from muscovite and chlorite). These reactions would liberate considerable H_2O that would have to escape if the new assemblage were to remain stable. If all the H_2O could not escape readily enough, these reactions would be only partly completed, and an extra phase thus would be allowed because of the decrease in variance of the system. On the other hand, in the regionally metamorphosed rocks the loss of H_2O seems in general to have kept pace with the dehydrating prograde reactions. The conversion of andalusite to sillimanite and the destruction of staurolite also did not go to completion, giving extra phases to some assemblages.

Metasomatic Contact Metamorphism

Samples taken from schists immediately adjacent to contacts of the biotite-muscovite adamellite stocks throughout the quadrangle characteristically show simplified mineral assemblages consisting mainly of quartz, muscovite, opaque minerals ± biotite. A trace of sillimanite is present in some specimens. One sample contains 75 percent muscovite, 20 percent quartz, and 5 percent tourmaline; another was estimated to contain 15 percent tourmaline. The intrusives next to these contacts are also characteristically simple, lacking the otherwise typical microcline but containing abundant muscovite. Their biotite is mostly altered to chlorite.

These features point to a transfer of matter between the intrusive magma and the adjacent schist. The intrusive has evidently lost potassium, and the schist seems to have gained it since the schist has more muscovite and less sillimanite than is normal, and typical pelitic rocks in the area contain only one-half percent or less of tourmaline.

The widespread muscovitization of andalusite in the staurolite and sillimanite zones, discussed in another paper (Green, *Jour. Geol.*, in press) may well be related to these solutions given off by the crystallizing adamellite magma.

IMPLICATIONS OF TEXTURES OF THE METASEDIMENTARY ROCKS

The megacrystic nature of the common metamorphic minerals (biotite, chlorite, garnet, magnetite, staurolite, andalusite and rarely sillimanite) indicates that they have nucleated and grown during certain stages in the metamorphism, in most cases after some recrystallization had already occurred. Their mutual relations help to elucidate the sequence of crystallization and the time relations between the metamorphism and the regional and local deformation.

As the porphyroblasts grew, they either pushed the surrounding mineral grains aside, or replaced the surrounding groundmass, or grew by a combination of these processes. By the first method, the foliation would be largely conformable to the surface of the porphyroblasts. With growth wholly by replacement, no such conformable relationship between the porphyroblasts and the groundmass foliation would be produced, unless the porphyroblast were platy and happened to lie parallel to the foliation.

However, the structural history of the rock also plays a role in determining the relation of a porphyroblast to the foliation. If deformation and the production of foliation had already ceased by the time of formation of the porphyroblast, the relation of the porphyroblast to the foliation would be determined only by the method of growth of the porphyroblast. If, on the other hand, deformation continued after the growth of the porphyroblast, either the porphyroblast would be deformed or the foliation would be made to conform more closely to its surface, no matter what the original relations were. Therefore, the texture in which foliation is conformable to a porphyroblast is by itself somewhat ambiguous.

In the Errol quadrangle the minerals biotite, garnet, staurolite, magnetite, ilmenite, and in some rocks, andalusite most commonly exhibit this conformable relationship. Porphyroblastic minerals whose shapes have generally not influenced the surrounding foliation are chlorite, cordierite, plagioclase, and occasionally andalusite and muscovite. These porphyroblasts must have formed after the deformation in their local areas, and must also have grown largely by replacement.

The fact that chlorite occurs in this way in so many middle- and high-grade rocks suggests a later time of formation, in harmony with the conclusion reached above that at least much of the chlorite is the result of retrograde reactions. In several specimens muscovite also occurs as similar cross-plates as well as flakes in the groundmass, which indicates two periods of recrystallization.

The post-deformational growth of cordierite and andalusite in the contact aureole in the north-central part of the area shows that the adamellite which caused the metamorphism was intruded after the regional deformation.

Poikilitic textures can also provide information on the metamorphism and deformation of the rocks. It seems very probable that inclusions in a porphyroblast which were not formed by subsequent, internal reactions (*e.g.*, not sericite nor antiperthitic lamellae in plagioclase) were part of the original groundmass before the crystal started to grow, and that as the crystal grew they were left in place for one of two reasons: either the surface of the growing porphyroblast did not force the grains away (in the case of a crystal growing by one-way, accretive diffusion and displacement), or the grains were for some reason not replaced by a crystal growing by replacement. For instance, granules of carbonaceous matter are normally as abundant inside a porphyroblast as in the groundmass, whereas micas of the groundmass are rarely seen as inclusions in porphyroblasts. In the rocks of the Errol quadrangle the most common inclusions are carbonaceous granules, other opaque minerals, and quartz.

In the Errol area, in a very few garnet and staurolite porphyroblasts post-recrystallization deformation is indicated by curved lines of inclusions; but in the great majority of porphyroblasts, including all those (especially plagioclase, andalusite, chlorite and muscovite) that are unconformable to the foliation, the lines of inclusions pass undeflected through the crystal. Therefore, most of the growth of porphyroblasts must have occurred after the regional deformation. Of course, if a porphyroblast has been rotated, its curved lines of inclusions may not be seen if the section is cut parallel to the axis of rotation.

In some specimens certain porphyroblastic minerals are found as inclusions in other minerals. Again assuming that the inclusion is not the result of a later reaction, this relationship directly implies a time sequence; the inclusion must have formed before the larger, including crystal. A possible exception could be the case in which both formed at the same time. In the Errol quadrangle garnets were seen enclosed in andalusite in three different rocks. In two of these, fresh staurolite was also present in the rock; the garnet and andalusite here did not form contemporaneously from staurolite. Staurolite porphyroblasts were surrounded by andalusite in three other rocks. It is to be expected that garnets, which are known to begin to form at a lower grade of metamorphism than andalusite, might be found inside andalusite crystals, but the occurrence of staurolite in a similar relationship is more informative and indicates that the staurolite must have formed also before andalusite. The first appearance of staurolite thus represents slightly lower-grade conditions from those at which andalusite first occurs in this area. As one of these specimens is from the contact-metamorphic aureole, this conclusion applies to the contact zone as well as to the regional metamorphism.

In summary, the textures of the rocks in the Errol quadrangle indicate that although most of the regional deformation occurred before the growth of the higher-temperature porphyroblasts, and certainly before the chlorite porphyroblasts produced by retrograde reactions, some rock deformation continued in certain areas after the peak of metamorphism (especially in the vicinity of the contacts of the adamellite stocks). Textural relations also indicate that garnet, staurolite, and andalusite probably appeared in that order during prograde metamorphism.

THOMPSON'S "AFM" PROJECTION AND THE PELITIC ROCKS OF THE ERROL QUADRANGLE

In the pelitic rocks of the Errol area, the abundance of mineral assemblages containing more than the three phases allowable by the phase rule demonstrates that the restrictions demanded by Thompson's "AFM" projection were not met in many rocks. The major factor probably is incomplete retrograde metamorphism, giving disequilibrium assemblages most of which involve chlorite. This complication may not, of course, be present in other areas.

Some of the difficulty met in using this projection lies in the composition of common pyralspite garnet, which because of its Ca and especially its Mn content, does not lie strictly in the system of the projection. Garnet therefore may be present as an extra phase in some pelitic rocks. This condition is certainly not a peculiarity of the rocks of the Errol quadrangle. However, as discussed above, if all or most of the chlorite in the extra-phase assemblages of the staurolite and sillimanite zones is assumed to be retrograde in origin, the most common mineral assemblages are reduced to two- or three-phase assemblages, and the garnet need not be counted as an extra phase.

A further obstacle to the strict application of the "AFM" projection lies in incomplete reactions such as andalusite to sillimanite, and the various reactions in the contact aureole. The lack of completion of some of the latter may be the result of restricted mobility of H_2O. These difficulties are not likely to be restricted to the Errol quadrangle, but may not be serious in some areas. High-grade rocks should in general show better attainment of equilibrium than low-grade rocks, and regional metamorphism should in general give more complete equilibrium than contact metamorphism, because of kinetic aspects of the reactions.

A final possible reason for the presence of extra-phase assemblages is that equilibrium, though obtained locally within a specimen, did not extend over the whole volume of a hand specimen or even over the area of a thin section. In a few thin sections different assemblages were found to be confined to restricted parts of the section. As mineral assemblages have been described above for whole thin sections, this areal restriction of equilibrium corresponding to local differences in bulk composition will produce extra phases in the described assemblages. Similarly restricted volumes of equilibrium have been found by W. C. Phinney (1963) in pelitic rocks of Nova Scotia, and are probably not uncommon in other metamorphic terranes.

The various impediments discussed above in no way invalidate Thompson's treatment, but only serve to point out that some of the necessary restrictions are not likely to be met in many metamorphic areas. The "AFM" projection in fact aids in disclosing disequilibrium relationships.

PHYSICAL CONDITIONS AND CAUSES OF METAMORPHISM

General statement. More and more areas are being described in which andalusite is found in regionally metamorphosed rocks, and it is becoming quite evident that andalusite cannot be considered an "anti-stress" mineral in the sense of Harker (1932) and others, since it is found in highly deformed rocks. The coexistence of andalusite and staurolite is less common, although described from, among other places, northeastern Scotland (Read, 1923), Korea (Yamaguchi, 1951), Japan (Ishioka and Suwa, 1956) and Finland (Eskola, 1927). Miyashiro states that this pair also occurs in the Baltic and Canadian shields (1961, p. 283).

The mineral assemblages in the Errol quadrangle correspond roughly to the greenschist facies (quartz-albite-epidote-biotite, and quartz-albite-epidote-almandine subfacies), the hornblende-hornfels facies, and the almandine-amphibolite facies (sillimanite-almandite-muscovite subfacies) of Fyfe *et al.* (1958). Nowhere, however, do these authors make allowances for the coexistence of andalusite and staurolite, nor of staurolite and cordierite. (The latter pair may not be in equilibrium in the Errol quadrangle; they are found together only in the contact aureole, and here the staurolite appears to have undergone resorption.) However, such assemblages are well known from some of the other areas just mentioned.

The production of the interesting combination andalusite, staurolite and cordierite in regionally metamorphosed rocks in Aberdeenshire and Banffshire has been called the "Buchan type" of metamorphism by Read (1952). Miyashiro (1961) recently proposed the use of types or groups of "metamorphic facies series" to designate mineralogically different sequences of progressive metamorphism, thus expanding and elucidating the attempts of Fyfe *et al.* (1958, p. 239). These "types" and "groups" of metamorphism are correlated with different thermal gradients, with the highest gradient according to Miyashiro corresponding to the "andalusite-sillimanite" type, and with successively lower gradients producing in turn the "low pressure-intermediate" group, the "kyanite-sillimanite" type, the "high pressure-intermediate" group, and the "jadeite-glaucophane" type. Thus the andalusite-sillimanite type would involve prograde reactions taking place at relatively shallow depths compared to the others.

According to Miyashiro's analysis, the "Buchan type" of metamorphism as described by Read belongs to the low-pressure intermediate group. The occurrence of cordierite in the high-grade regionally metamorphosed rocks of southern New Hampshire (Heald, 1950; Billings, 1956; Barker, 1961) would probably place this area in the low-pressure intermediate group also, except that kyanite is the stable aluminum silicate at lower metamorphic grades just to the north, instead of andalusite (Chapman, 1939). Miyashiro does not give enough information on pelitic assemblages in his low-pressure intermediate group to allow a strict assignment of these rocks. The role of chloritoid is also uncertain in Miyashiro's scheme, but it appears typical of the kyanite-sillimanite type of metamorphism in central Vermont (*e.g.*, Albee, 1957). Chloritoid appears to be absent throughout New Hampshire (Billings, 1956), but this may be due to restricted bulk compositions as well as to metamorphic conditions.

In the Errol quadrangle the widespread occurrence of andalusite and staurolite place the rocks in the low-pressure intermediate group, but since cordierite is found only in the contact aureole, the pressure is assumed to have been slightly higher than that characterizing the Buchan metamorphism. The maximum temperature attained probably was not as high as in southern New Hampshire, where cordierite is common in regional gneisses and muscovite has over wide areas decomposed to sillimanite and potassium feldspar (Heald, 1950; Barker, 1961).

Implications of experimental work. If equilibrium is maintained during progressive metamorphism, the resulting mineral assemblages and compositions will be determined solely by the original composition of the rock, the temperature, the pressure, and the activities of any mobile components. Since every mineral and mineral assemblage has a finite stability range, the mineral composition of a rock provides internal evidence of the conditions of crystallization, provided the actual stability ranges can be accurately determined in the laboratory. Because of the difficulties involved in the synthesis of many of the common metamorphic minerals, and the variable compositions thereof, there are few experimental data on which to base quantitative assignments of physical conditions to the recrystallization of these rocks. Eventually it is hoped that a quantitative grid can be developed which will correlate any equilibrium mineral assemblage (assuming one can be identified) with the specific physical conditions of its formation (Thompson, 1961).

Meanwhile estimates of metamorphic conditions must be based mainly on geologic relationships, guided by experimental work. The maximum stability fields of those metamorphic minerals of pelitic rocks which have been investigated in the laboratory [*e.g.*, muscovite (Yoder and Eugster, 1955), annite (Eugster and Wones, 1962), phlogopite (Yoder and Eugster, 1954), pyrite (Kullerud and Yoder, 1957), chloritoid (Halferdahl, 1957), cordierite (Schreyer and Yoder, 1959), chlorite (Turnock, 1959), and the aluminum silicates (Clark, Robertson, and Birch, 1957; Clark, 1960; and Bell, 1963)] generally extend beyond the limits reasonably expectable in regional metamorphism. This results principally from the fact that most of the experimental systems were unlike natural occurrences in one or more of the following respects: (1) the experimental system was saturated with H_2O; (2) the system was not saturated with SiO_2 nor with other minerals of natural assemblages; and (3) the ferromagnesian minerals were pure end members. These restrictions were of course necessary for the basic work on these minerals, and it can be expected that further work will further approach natural systems. Many of the results of these experiments can be anticipated qualitatively by thermodynamic considerations as well as empirical observations of mineral assemblages and compositions. It is to be expected, for instance, that both saturation in SiO_2 and undersaturation in H_2O will lower the maximum stability temperatures of most of these minerals.

A few estimates have been made (Fyfe *et al.*, 1958; Turner and Verhoogen, 1960; Miyashiro, 1961; Thompson, 1961) of the physical conditions corresponding to particular metamorphic and mineral facies, based on such considerations, but they differ widely (Fig. 9). The minimum

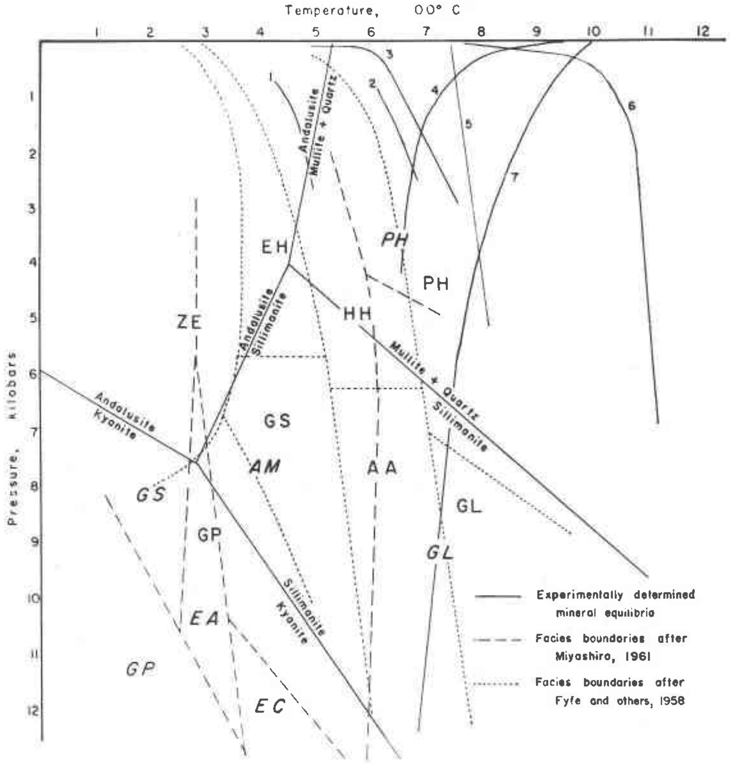

FIG. 9. Experimental mineral equilibria and physical conditions of metamorphism. Metamorphic facies: ZE = zeolite; EH = epidote hornfels; HH = hornblende hornfels; PH = pyroxene hornfels; GS = greenschist; AM = amphibolite; AA = almandine amphibolite; GL = granulite; GP = glaucophane schist; EA = epidote amphibolite; EC = eclogite. Slanted letters refer to Miyashiro, 1961; straight letters to Fyfe *et al.*, 1958. Mineral equilibria: 1—annite $\rightleftharpoons$ sanidine + magnetite + hematite + vapor (hematite-magnetite buffer); 2—annite $\rightleftharpoons$ sanidine + magnetite + vapor (magnetite-quartz-fayalite buffer); 3—muscovite $\rightleftharpoons$ sanidine + corundum + vapor; 4—beginning of melting of water-saturated Quincy and Westerly granites; 5—pyrite $\rightleftharpoons$ pyrrhotite + liquid; 6—phlogopite $\rightleftharpoons$ forsterite + leucite + $KAlSiO_4$ + vapor; 7—"curve of fusion of water-saturated granite." References: Curves 1,2,5—*Ann. Rept. Geophys. Lab.*, 1956–57; curve 3—Yoder and Eugster, 1955; curve 4—Tuttle and Bowen, 1958; curve 6—Yoder and Eugster, 1954; curve 7—Fyfe and others, 1958. Aluminum silicate curves from Clark *et al.* 1957 and Clark, 1961.

melting curve of granite (Tuttle and Bowen, 1958; Fyfe *et al.*, 1958) provides a useful upper limit for metamorphic conditions, but even here different authors disagree on the physical conditions of melting. One of the major reasons for the disparity in interpretation of the experimental data seems to be in divergent opinions on the partial pressure or activity of H_2O during metamorphism. Whereas Thompson (1955, 1957) and

Miyashiro (1961) consider rocks undergoing prograde metamorphism to be at least in large part open to the escape of H_2O, with the activity of H_2O externally determined and generally less than unity, Fyfe *et al.* consider the total pressure to be equal to the P_{H_2O} (the rock being constantly saturated with H_2O). Largely for this reason, the estimates of temperature given by Fyfe *et al.* (p. 237) are higher than those of Miyashiro (1961, p. 285, 286).

One of the most applicable mineral systems to receive study is the composition Al_2SiO_5. The polymorphic transitions between kyanite, andalusite and sillimanite are independent of other components, and the major problem has been with reaction rates and metastable phases (Clark *et al.*, 1957; Clark, 1961). Bell's recent experimental work (1963), as well as geologic evidence and thermodynamic considerations (Miyashiro, 1953; Thompson, 1955; Clark, et al., 1957) show that the polymorphic relations are similar to those shown in Fig. 9 (taken from Clark *et al.*, 1957 and Clark, 1961) with andalusite the low-temperature, low-pressure polymorph.

The sillimanite isograd in the Errol quadrangle corresponds to the andalusite-sillimanite phase boundary, and the rough estimates of Clark *et al.*, thus provide limits of pressure and temperature for this surface. As no kyanite is present, the low-temperature limit is about 300° C, and the high-temperature limit is about 450° C. The pressure corresponding to these triple-point temperatures are respectively estimated at about 8 kilobars and about 4 kilobars. If the pressure is purely hydrostatic, these figures in turn imply depths in the range 29 to 15 kilometers, assuming an average density of overlying rocks of 2.79 gm/cm^3.[1] Birch's (1955, p. 113–114) estimates of the geothermal gradient show that the andalusite-sillimanite isograd suggested by Clark *et al.* could be reached and crossed simply by hydrostatic burial, if reasonable assumptions are made as to radioactivity and conductivity in the crust and assuming time enough for thermal equilibrium to be established. According to Hatch (Ph.D. thesis, Harvard University, 1961) about 14,000 feet of stratified rocks overlie, in the area immediately to the west, the rocks exposed in the Errol quadrangle, and more may well have been present before post-metamorphic erosion began. A thickening of this upper part of the crust by a factor of at least three seems necessary, and may not be unreasonable in such strongly folded rocks. Another alternative has been suggested by Clark (1961, p. 647); "tectonic overpressure," produced by orogenic forces in rocks of appreciable strength, could decrease the depth neces-

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[1] The average specific gravity of 22 metamorphic rocks from the Errol quadrangle, measured by Wm. Joyner.

sary to attain these total pressures. Although such tectonic overpressure may be a necessary condition for the formation of kyanite in many regionally metamorphosed rocks, it does not seem to be necessary in the present instance for the transformation of andalusite to sillimanite.

Miyashiro (1960) has calculated that the temperature of the decomposition of muscovite to sillimanite plus potassium feldspar, a widespread phenomenon in south-central New Hampshire but not attained in northern New Hampshire, lies in the range 450° C to 590° C, depending on the activity of H_2O. Because this reaction did not occur in the Errol quadrangle the maximum temperatures must have been within or below this range.

In summary, the mineral facies in the Errol quadrangle were produced at temperatures probably below about 500° C and under pressures not necessarily entirely hydrostatic, ranging between 4 and 8 kilobars. There appears to have been a fairly rapid increase in temperature with depth during the metamorphism.

Causes of metamorphism. One problem in any metamorphic terrane is the source of heat for the endothermic reactions of progressive metamorphism. Four major sources of heat may be considered: frictional heat from crustal deformation, heat from inherent radioactivity, heat from adjacent rocks of the mantle, and magmatic heat.

Shearing stress certainly plays a role in developing the textural character of metamorphic rocks, and also in catalyzing and speeding otherwise sluggish reactions, but, as concluded by Turner and Verhoogen (1960, p. 660), it does not appear to be an appreciable source of heat for metamorphism.

A small amount of the heat undoubtedly was supplied by internal radioactivity, but there is no evidence that metamorphic rocks contain any greater amount of radioactive isotopes than other rocks. Crustal thickening above would of course prevent rapid conduction away of what normal radiogenic heat was generated.

Some of the heat must also have been conducted into these rocks from below, as the sediments were gradually sinking and the crust was thickening. This amount may have been enough to produce mineral facies appropriate to the greenschist or at least the zeolite facies. Fyfe *et al.*,

> "suggest subject to revision when more data on actual gradients in metamorphic rocks and of their radioactive content become available, that metamorphism occurs only under conditions of greatly increased heat flow. Metamorphism . . . is not a 'normal' phenomenon in the sense that it occurs whenever sediments have been buried to sufficient depths" (1958 p. 196).

This seems to contradict the implications of Birch's estimates of "reasonable" geothermal gradients (1955, p. 113). Also, low-grade metamorphism might well be possible under geosynclinal heat gradients if the rocks are under the catalytic influence of intense orogenic deformation, without any contribution from magmatic heat.

Heat derived from crystallizing and cooling magma and igneous rock is obviously the principal cause of metamorphism in the clear-cut cases of contact metamorphism. In the Errol quadrangle the aureole around the adamellite stock in the north-central part, and the wider zone around the granodiorite body in the southeastern part, obviously can be attributed to magmatic heat, as can the narrow sillimanite zones around some of the other adamellite stocks. In fact, the entire area within the regional staurolite isograd coincides with an irregular band of adamellite stocks that trends from southwest to northeast across the quadrangle (Fig. 1). This correlation between metamorphic grade and abundance of similar intrusive bodies implies that the stocks may be cupolas of an elongated batholith at some depth. Because of the contact metamorphism and the intrusive relations of these stocks, it is clear that they contributed heat to the country rocks, and are not simply the product of partial melting in place. Being of peraluminous granitic composition near the minimum melting composition (Tuttle and Bowen, 1958), these adamellites are most likely the products of partial fusion of the deeper parts of the sialic crust depressed in the Acadian tectogene. Turner and Verhoogen (1960, p. 668) find it hard to believe that palingenetic granites would be able to form from a tectogene in time to intrude and supply heat to overlying geosynclinal rocks, stating that "a perfectly insulated piece of granite would require some 30 million years to melt itself." Was this calculation made starting at 0° C or at temperatures appropriate to the base of the normal, sialic crust? Also, if the convection theory of orogeny and mountain roots is valid, there is a large amount of heat being supplied by conduction from the mantle into the root. Furthermore, the world-wide evidence of the association of large volumes of granitic rocks with orogenic zones must be recognized, and it points strongly toward partial melting during orogeny.

The intrusive rocks remain as the most probable major source of heat for the medium-grade and high-grade metamorphism in the Errol area. The movement of palingenetic magmas to higher levels in the crust would necessarily have set up steeper heat gradients than "normal" in the geosynclinal rocks. With the thick rock cover at such depths, and with the surrounding sediments having already been somewhat heated according to a "normal" geothermal gradient, this magmatic heat would have been insulated to some degree, and temperatures could build up to those appropriate to the middle and higher metamorphic grades throughout a considerable volume of rock.

REFERENCES

ALBEE, A. L. (1957) Bedrock geology of the Hyde Park quadrangle, Vermont. *U. S. Geol. Surv. Quad. Map,* GQ 102.

——— (1962) Relationships between mineral association, chemical composition and physical properties of the chlorite series. *Am. Mineral.* **47,** 851–870.

BARKER, FRED (1961) Phase relations in cordierite-garnet-bearing Kinsman quartz monzonite and the enclosing schist, Lovewell Mountain quadrangle, New Hampshire. *Am. Mineral.* **46,** 1166–1176.

BARROW, G. (1893) On an intrusion of muscovite-biotite gneiss in the southeast Highlands of Scotland. *Geol. Soc. London Quart. Jour.* **49,** 330–358.

BELL, PETER M. (1963) Aluminum silicate system: Experimental determination of the triple point. *Science* **139,** No. 3539, 1055–1056.

BILLINGS, M. P. (1956) *The Geology of New Hampshire, part 2, Bedrock geology.* N. H. State Planning and Development Comm.

BIRCH, A. F. (1955) Physics of the crust, *in* A. Poldervaart, ed., Crust of the Earth—a symposium. *Geol. Soc. Am. Special Paper* **62,** 101–117.

CHAPMAN, C. A. (1939) Geology of the Mascoma quadrangle, New Hampshire. *Bull. Geol. Soc. Am.* **50,** 127–180.

CHINNER, G. A. (1961) The origin of sillimanite in Glen Clova, Angus. *Jour. Petrol.* **2,** 312–323.

CLARK, S. P., JR. (1961) A redetermination of equilibrium relations between kyanite and sillimanite. *Am. Jour. Sci.* **259,** 641–650.

——— E. C. ROBERTSON AND A. F. BIRCH, (1957) Experimental determination of kyanite-sillimanite equilibrium relations at high temperatures and pressures. *Am. Jour. Sci.* **255,** 628–640.

ESKOLA, P. (1927) Petrographische Charakteristik der kristallinen Gesteine von Finland. *Fortschr. Mineral.* **11,** 57–112.

EUGSTER, H. P. AND D. R. WONES (1962) Stability relations of the ferruginous biotite, annite. *Jour. Petrol.,* **3,** 82–125.

FYFE, W. S., F. J. TURNER AND J. VERHOOGEN (1958) Metamorphic reactions and metamorphic facies. *Geol. Soc. Am. Mem.* **73.**

GOLDSCHMIDT, V. M. (1911) Die Kontaktmetamorphose in Kristianiagebiet. *Oslo Vidensk. Skr.,* I, *Mat.-Naturv. Kl.,* **11**

GREEN, J. C. (1963) Stratigraphy and structure of the Boundary Mountain anticlinorium in the Errol quadrangle, New Hampshire and Maine. *Geol. Soc. Am. Special Paper,* (in press).

——— (1963) Alkali metasomatism in a thermal gradient: two possible examples. *Jour. Geol.* (in press)

HALFERDAHL, L. B. (1957) Chloritoid. *Carnegie Inst. Wash. Year Book* **56,** 225–228.

HARKER, A. (1932) *Metamorphism.* Methuen, London.

HEALD, M. T. (1950) Structure and petrology of the Lovewell Mountain quadrangle, New Hampshire. *Bull. Geol. Soc. Am.* **61,** 43–89.

ISHIOKA, K. AND K. SUWA (1956) Metasomatic development of staurolite schist from rhyolite in the Kurobe-gawa area, central Japan (a preliminary report). *Jour. Earth Sci. Nagoya Univ.* **4,** 123–140.

KULLERUD, G. AND H. S. YODER (1957) The Fe-S system. *Carnegie Inst. Wash. Year Book* **56,** 187–195.

Lyons, John B. (1955) Geology of the Hanover quadrangle, New Hampshire-Vermont. *Bull. Geol. Soc. Am.* **66,** 105–146.

Miyashiro, A. (1953) Calcium-poor garnet in relation to metamorphism. *Geochim. Cosmochim. Acta,* **4,** 179–208.

——— (1960) Thermodynamics of reactions of rock-forming minerals with silica, Part IV, Decomposition reactions of muscovite. *Japan. Jour. Geol. Geog.* **41,** 113–120.

——— (1961) Evolution of metamorphic belts. *Jour. Petrol.* **2,** 277–311.

Phinney, W. C. (1963) Phase equilibria in the metamorphic rocks of St. Paul Island and Cape North, Nova Scotia. *Jour. Petrol.,* **4,** 90–130.

Ramberg, H. (1952) *The origin of metamorphic and metasomatic rocks, a treatise on recrystallization and replacement in the earth's crust.* Univ. Chicago Press.

Read, H. H. (1923) The geology of Banff, Huntly and Turriff. *Geol. Surv. Great Britain Mem.* **86.**

——— (1952) Metamorphism and magmatisation in the Ythan Valley, Aberdeenshire. *Trans. Edinburgh Geol. Soc.* **15,** 265–279.

Schreyer, W. and H. S. Yoder (1959) Cordierite-water system. *Carengie Inst. Wash. Year Book* **58,** 100–104.

Thompson, J. B., Jr. (1955) The thermodynamic basis for the mineral facies concept. *Amer. Jour. Sci.* **253,** 65–103.

——— (1957) The graphical analysis of mineral assemblages in pelitic schists. *Am. Mineral.* **42,** 842–858.

——— (1959) Local equilibrium in metasomatic processes, *in* P. H. Abelson, ed., *Researches in Gochemistry.* John Wiley & Sons, New York, 427–457.

——— (1961) Mineral facies in pelitic schists (English summary). *D. S. Korzhinsky Comm. Vol.,* 313–325.

Tilley, C. E. (1925) Metamorphic zones in the southern Highlands of Scotland. *Geol. Soc. London Quart. Jour.* **81,** 100–112.

Turner, F. J. and J. Verhoogen (1960) *Igneous and Metamorphic Petrology.* McGraw-Hill Book Co., New York.

Turnock, A. C. (1959) Stability range of iron chlorite (abs.). *Bull. Geol. Soc. Am.* **70,** 1690–1691.

Tuttle, O. F. and N. L. Bowen (1958) Origin of granite in the light of experimental studies in the system $NaAlSi_3O_8$-$KAlSi_3O_8$-SiO_2-H_2O. *Geol. Soc. Am. Mem.* **74.**

White, W. S. and M. P. Billings (1951) Geology of the Woodsville quadrangle, Vermont-New Hampshire. *Bull. Geol. Soc. Am.* **62,** 647–696.

Winchell, A. N. and H. Winchell (1951) *Elements of Optical Mineralogy—an Introduction to Microscopic Petrography.* John Wiley & Sons, New York.

Yamaguchi, T. (1951) On the so-called Yonchon system and its regional metamorphism (in Japanese). *Jour. Geol. Soc. Japan,* **57,** 419–438.

Yoder, H. S. and H. P. Eugster (1954) Phlogopite synthesis and stability range. *Geochim. Cosmochim. Acta,* **6,** 179–182.

——— (1955) Synthetic and natural muscovites. *Geochim. Cosmochim. Acta,* **8,** 225–280

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