# MINERALOGICAL RELATIONS IN SOME PELITIC HORN-FELSES FROM THE SOUTHERN SIERRA NEVADA, CALIFORNIA

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## Abstract

An extensive body of pelitic hornfelses ranging in composition from andalusite+biotite +quartz+muscovite to sillimanite+cordierite+perthite+biotite+quartz±almandine occurs in a metamorphic septum near the southern end of the Sierra Nevada batholith east of Isabella, California. Field and laboratory investigations suggest the hornfelses developed at high temperatures over a prolonged period of time. Chemical analyses of some constituent minerals show that the stability field of biotite+andalusite diminishes and biotite and cordierite become richer in $Fe^{2+}$ relative to Mg with increasing temperature. Fe appears to have been reduced from a somewhat oxidized state during prograde thermal metamorphism, suggesting the metamorphic body was "open" to oxygen. The development of garnet with accompanying cordierite, biotite, and sillimanite may be related to excess Ca in some rocks, but this is not definite.

## Introduction

Field mapping in the southern portion of the Sierra Nevada batholith east of Isabella, California (Figs. 1 and 2) has disclosed the presence of a rather extensive mass of pelitic hornfelses unlike those commonly found in other metamorphic pendants and septa in the batholith. This paper presents the results of a detailed investigation of these hornfelses. In studying the progressive thermal metamorphism, emphasis was placed on the (1) phase relations of compatible ferromagnesian minerals, (2) metamorphic reactions which gave rise to the observed mineral assemblages, and (3) variations in chemical composition of constituent minerals. The data presented herein provide information on the general nature of metamorphism adjacent to batholithic intrusions.

Previous geologic studies in the area around Isabella have dwelt mainly on the batholithic rocks (Miller, 1932; Miller & Webb, 1940). A detailed geologic map and a structural analysis of the metamorphic rocks, embodying the results of our investigation, will be published elsewhere.

## General Geologic Setting

The rocks in the Isabella area consist of igneous and metamorphic types similar to those found elsewhere in the Sierra Nevada (see, for ex-

ample, Mayo, 1935; Erwin, 1937; Durrell, 1941; Hamilton, 1956; Parker, 1961). Igneous rocks display a variety of field relations, from dikes, sills and small stocks to the composite batholith surrounding the septum of metamorphic rocks. The composition of the individual intrusions and facies within these constituting the batholith, ranges from
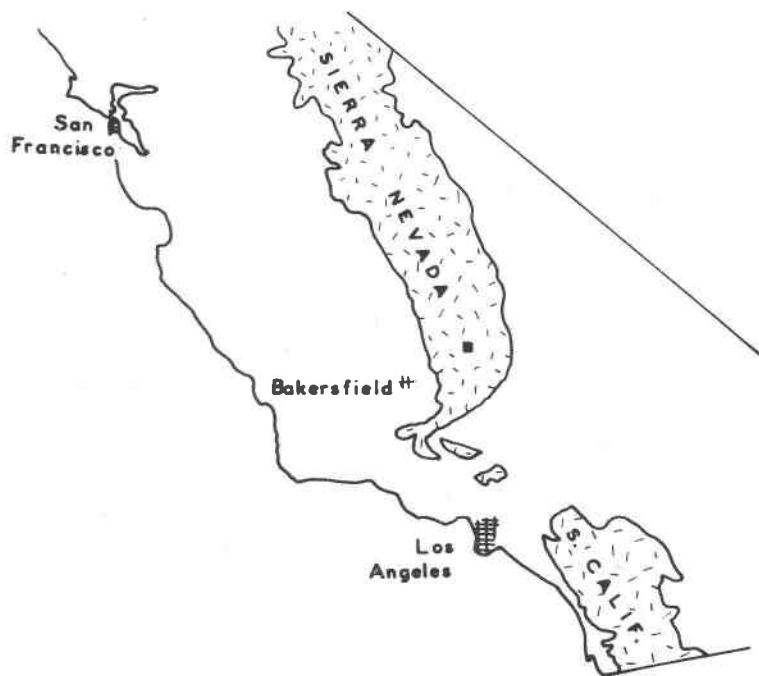


FIG. 1. Map of southern California showing location of the investigated area (black square) and Late Mesozoic batholiths.

granite to diorite but an equigranular biotite granodiorite is dominant. Contacts with the metamorphic rocks are sharp and not blurred by migmatitic or other gradational contacts. The outline of the contact of the batholith in the southeastern part of the mapped area (Fig. 2) is irregular. This feature, together with the occurrence of numerous satellitic apophyses and swarms of pegmatite-aplite dikes in the pelitic rocks, suggests that, in this area, the roof of the batholith lies just below the surface.

The greater part of the exposed metamorphic rocks was derived from pelitic sedimentary rocks and basic to intermediate volcanic rocks, commonly of pyroclastic origin. Marble, calc-silicate rock, and quartzite occur
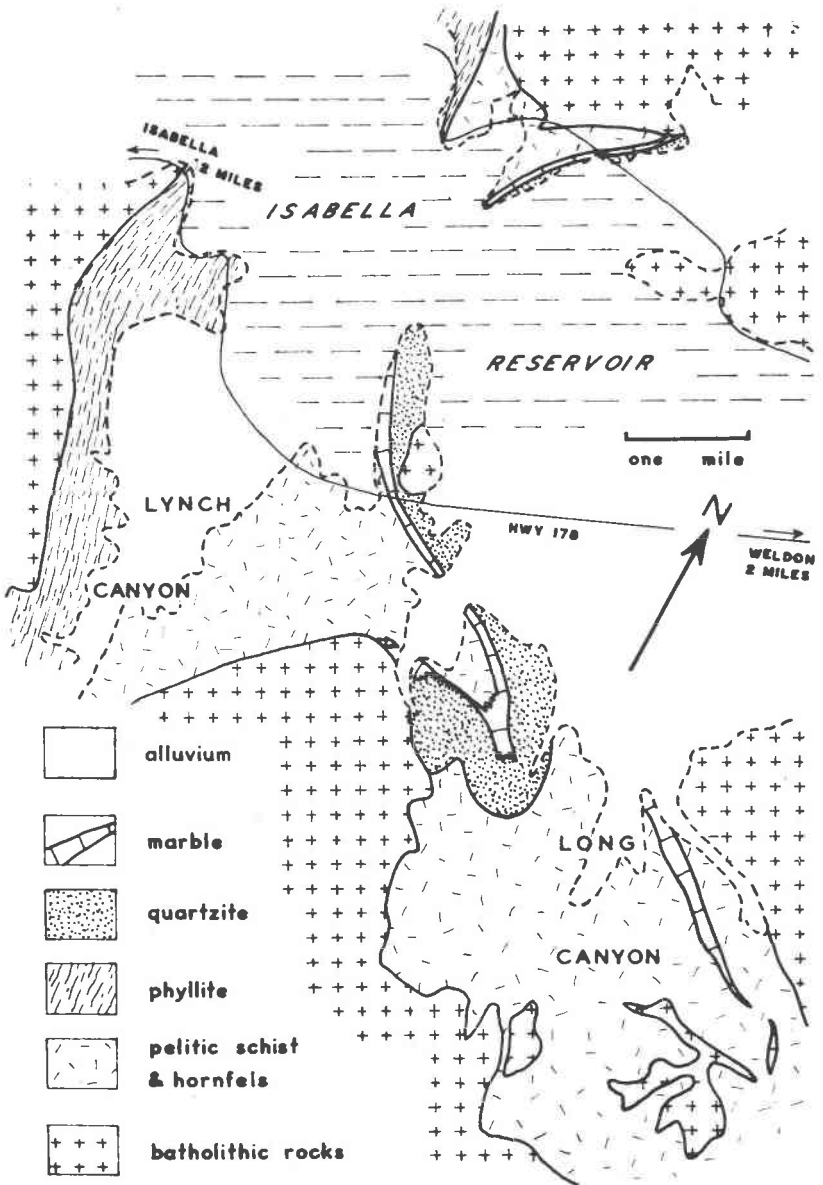
FIG. 2. Generalized geologic map of the Sierra Nevada between
Isabella and Weldon, California.

only in minor amounts, except in the center of the map (Fig. 2) which is the site of a large fold hinge in layers of marble and quartzite. The belt of phyllites, in part derived from volcanic material, along the western edge of the area (Fig. 2) represents a regionally metamorphosed mass apparently unaffected by later thermal metamorphism. Compatible mineral assemblages in these phyllites indicate metamorphism under lower greenschist facies conditions (Fyfe, *et al.* 1958, p. 219). The fabric and mineralogy of metamorphic rocks north of the Isabella Reservoir, on the east side of Lynch Canyon, and in Long Canyon show clearly that an episode of thermal metamorphism, undoubtedly attending emplacement of the batholith, followed after the earlier regional recrystallization and concomitant intense deformation. Metamorphic rocks north of Isabella Reservoir and on the east side of Lynch Canyon consist of hornfelsic and semihornfelsic pelites and lesser amounts of marble, calc-silicate rock, quartzite, and amphibolite. Those in Long Canyon consist almost wholly of pelitic hornfels. Variations in mineral composition[1] of 60 samples of hornfelses do not suggest large variations in bulk chemical composition. Mineralogical differences that do occur (Fig. 3) seem to reflect variations in grade of metamorphism and slight local variations in bulk chemical composition.

## PELITIC HORNFELSES

*Petrography.* The transition from regionally metamorphosed rocks into those affected by thermal metamorphism cannot be observed anywhere in the map area. East of Lynch Canyon and north of Highway 178 there are a few exposures of muscovite-chlorite-quartz-hematite phyllite but the next significant exposures approximately along the strike of this rock type, about one quarter mile south and across the highway, consist of rocks which lack chlorite and contain biotite and andalusite. The idiomorphic flakes of biotite, which are interleaved with muscovite, are generally yellow brown or dark brown in the Y and Z directions although in some thin sections, or parts thereof, the Y and Z colors are reddish brown. The foliation, so well developed in phyllites to the north, is partially obliterated owing to the presence of large prisms of andalusite containing abundant inclusions of quartz, opaques and some shreds of biotite. Tiny aggregates of needle-like sillimanite(?) appear locally around margins of biotite in two thin sections. Opaque grains are scarce but if present are thin plates of optically homogeneous ilmenite. A typical sample of this rock type is B50 (Table 1), and an analysis of the biotite from it is given in Table 2.

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[1] All mineral compositions (except. for B116) are based on visual estimates. The fine and very irregular grains in most hornfelses preclude accurate point counting.
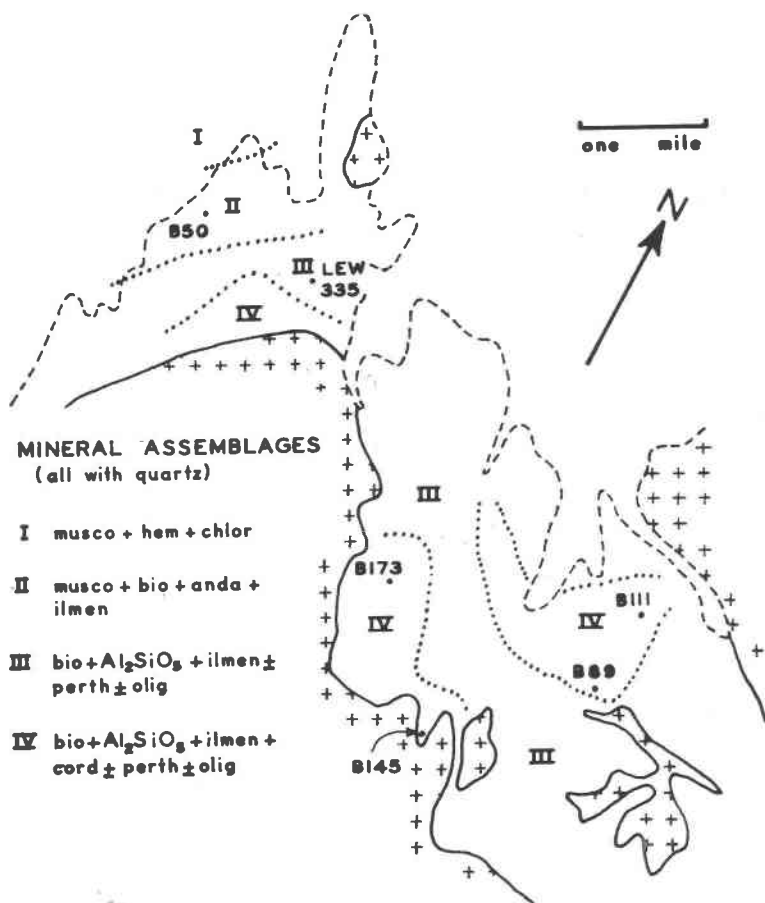
Fig. 3. Generalized distribution of mineral assemblages in the metamorphic rocks between Isabella and Weldon, California. Base map is the same as Fig. 2. Locations of samples whose constituent minerals have been analyzed are indicated.

Pelitic rocks (*e.g.* LEW 335) closer to the exposed contact of the batholith east of Lynch Canyon contain poikilitic patches of slightly perthitic microcline ($2V_x$ 68°) but are otherwise similar to rocks such as B50. Idiomorphic flakes of muscovite in the rock are reduced to small wisps in and near the perthite (Fig. 4). Rope-like masses of fibrolite are developed very locally adjacent to biotite and, in some cases, andalusite.

Cordierite was expected in hornfelses near the outer limits of thermal recrystallization, but none was found by careful search of thin sections. As a further test, the leucocratic fractions of B50 and LEW 335 were separated from biotites, and then these fractions were split into magnetic

and non-magnetic portions in a magnetic separator and $x$-rayed. No diffraction peaks characteristic of cordierite were found in any portions.

Sillimanite is usually an important constituent in pelitic hornfelses still closer to the exposed batholithic contact east of Lynch Canyon as well as in all pelites in Long Canyon (see areas denoted III and IV in Fig. 3). Muscovite is absent except for local poikilitic plates apparently in textural disequilibrium with the other minerals. There are essentially three distinct assemblages of compatible minerals in these hornfelses:

(a) quartz+biotite+$Al_2SiO_5$[1]+ilmenite$\pm$oligoclase
(b) quartz+biotite+$Al_2SiO_5$+ilmenite+perthite$\pm$oligoclase
(c) quartz+biotite+$Al_2SiO_5$+ilmenite+cordierite$\pm$perthite$\pm$oligoclase.

Assemblage (a) occurs in a band extending away from the marble fold hinge in a southeasterly direction and, in some instances, within a few hundred feet of igneous contacts. In general these rocks display a more thorough recrystallization than pelites containing muscovite. Hornfelsic texture is typical, although locally a rock may still possess a slight schistosity. Sillimanite, usually more common than stout prisms of andalusite, occurs as rope-like bundles of fibrolite intimately associated with strongly pleochroic biotite (pale yellow-brown to deep red-brown) or as crisscross needles within biotite (similar to Fig. B in Plate of Chinner, 1961). Rarely it is developed in individual long prisms, some of which penetrate and partially take the place of andalusite. Oligoclase appears as xenomorphic grains difficult to distinguish from quartz. With the single exception of one hematite-bearing rock, all samples of assemblage (a) contain ilmenite as the sole Fe-Ti oxide; the grains are nearly equant, locally hexagonal, and are concentrated in the bundles of fibrolite. Accessory minerals, in these as well as in all hornfelses to be described, include brown, zoned tourmaline, zircon, late or secondary white mica, and, uncommonly, graphite.

Assemblage (b) was found in only four samples. Two of these, both with regard to location and mineralogy, are transitional between muscovitic and cordieritic rocks east of Lynch Canyon. The other two are nearly identical to the cordierite-bearing hornfelses, assemblage (c).

These cordieritic hornfelses are, with few exceptions, found directly adjacent to the exposed area of the batholith, in the migmatitic area (see discussion below) in western Long Canyon, or as country rock in the swarm of aplite-pegmatite dikes at the mouth of Long Canyon. The cordierite in assemblage (c) seldom displays sector or lamellar twinning; a

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[1] The occurrence of both sillimanite and andalusite in the same hand specimens is nearly universal in assemblages lacking stable muscovite. Hence, for brevity, the presence of these phases will be indicated by $Al_2SiO_5$.

FIG. 4. Photomicrograph of a pelitic hornfels showing development of perthitic microcline (m) at the expense of muscovite (colorless flakes). Biotite, quartz (q), and poikilitic andalusite (upper right) also occur in section. Note droplets of quartz in microcline (upper left). Plain light. Scale is 0.5 mm. long.
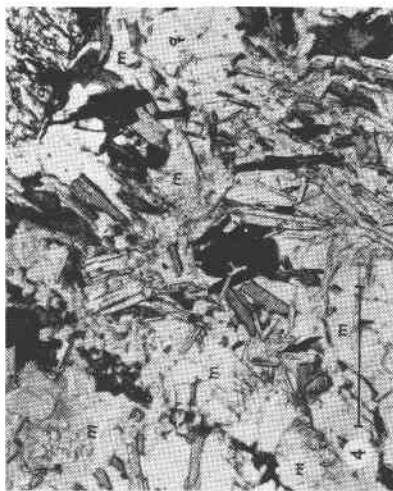
FIG. 5. Photomicrograph of pelitic hornfels B89. Microcline (m) formed by breakdown of biotite and andalusite appears on right. Tufted bundles of sillimanite and irregular particles of graphite are common. Plain light. Scale is 0.5 mm. long.

FIG. 6. Photomicrograph of coarse pelitic hornfels B145. Cordierite (c), with incipient "pinitic" alteration, quartz (q), and microcline perthite (m) constitute most of the rock. Ragged biotite, sillimanite and ilmenite (upper right) occur together. Note needles of sillimanite concentrated around cordierite. Plain light. Scale is 0.5 mm. long.

more reliable criterion of identification is the presence of "pinitic" alteration (*cf.* Schreyer and Yoder, 1961) of grains along margins and cleavage planes. In finer-grained rocks (such as B89, Tables 1 and 2, Fig. 5) in which cordierite occurs as very small irregular grains (1%) it is usually extensively or completely replaced by an isotropic orange "pinite." As the grain size of the cordieritic rocks increases (compare Figs. 5 and 6) there is a concomitant increase in the amount of cordierite, perthite (K-feldspar with a large negative optic angle as host), sillimanite, and il-

TABLE 1. COMPATIBLE MINERAL ASSEMBLAGES IN ROCKS FROM WHICH MINERAL ANALYSES HAVE BEEN MADE. MODAL COMPOSITION (VOL. %) OF THE COARSE HORNFELS B116 IS INDICATED

|  | B50 | LEW 335 | B89 | B145 | B173 | B116 | B111 |
|---|---|---|---|---|---|---|---|
| quartz | X | X | X | X | X | 11.4 | X |
| biotite | X | X | X | X | X | 30.7 | X |
| muscovite | X |  |  |  |  |  |  |
| andalusite | X | X | X |  |  | 15.0 |  |
| sillimanite | ? | X | X | X | X |  | X |
| cordierite |  |  | X | X | X | 26.4 |  |
| garnet |  |  |  |  |  | 7.7 | X |
| plagioclase |  |  |  |  |  | 7.6 |  |
| (composition) |  | oligoclase | $An_{20}$ | $An_{18-23}$ | $An_{21-30}$ | $An_{24-37}$ | $An_{25}$ |
| perthite |  | X | X | X | X |  | X |
| ilmenite | X |  | X | X | X |  | X |
| magnetite |  |  |  |  |  | 1.2 | X |
| pyrrhotite |  |  |  |  |  |  |  |
| graphite |  |  | X |  | X |  |  |

Sample B116 from road cut, batholith-hornfels contact on north side of Isabella Reservoir. Other sample locations shown in Fig. 3.

menite and a decrease in biotite (YZ = deep red-brown) and andalusite. In coarse rocks (such as B173 and B145, Tables 1 and 2) andalusite occurs as clean equant grains.

A few samples of assemblage (c), collected from within the swarm of dikes at the mouth of Long Canyon, contain from 2 to 10 grains of garnet per thin section. The amount was too small to obtain any physical properties, other than the refractive index. which ranges from 1.809 to $1.814 \pm .003$. In one sample the garnet occurs principally as thick, incomplete mantles around andalusite.

Other garnetiferous hornfelses crop out near the batholith north of Isabella Reservoir (Fig. 2). One exceptionally coarse sample, B116, collected in the roadcut at the contact, contains porphyroblastic garnet and

| | Wt. % oxides | | | | | |
|---|---|---|---|---|---|---|
| | B50 | LEW 335 | B89[1] | B173 | B116 | B111 |
| $SiO_2$ | 37.4 | 36.26 | 36.0 | 34.43 | n.d. | 35.0 |
| $TiO_2$ | 2.2 | 2.99 | 3.1 | 3.48 | 2.83 | 3.3 |
| $Al_2O_3$ | 17.1 | 18.40 | 19.2 | 19.87 | n.d. | 19.5 |
| $Fe_2O_3$ | 4.21 | 2.57 | 2.03 | 0.63 | 1.04 | 1.05 |
| $FeO$ | 17.31 | 18.58 | 19.07 | 21.85 | 22.46 | 24.95 |
| $MnO$ | 0.2 | 0.35 | 0.2 | 0.21 | 0.14 | 0.1 |
| $MgO$ | 7.78 | 6.66 | 6.70 | 6.20 | 5.76 | 3.05 |
| $CaO$ | 0.16 | 0.15 | 0.13 | 0.16 | 0.13 | 0.29 |
| $Na_2O$ | 0.25 | 0.15 | 0.19 | 0.17 | n.d. | 0.17 |
| $K_2O$ | 8.5 | 8.90 | 8.9 | 8.30 | n.d. | 8.5 |
| $H_2O^+$ | } 4.30 | } 4.34 | } 3.66 | 4.21 | n.d. | } 3.19 |
| $HO_2^-$ | | | | 0.02 | n.d. | |
| F | 0.21 | 0.19 | 0.23 | 0.26 | n.d. | 0.30 |
| Total | 99.7 | 99.54 | 99.4 | 99.79 | | 99.6 |
| less O | | | | | | |
| for F | 99.6 | 99.46 | 99.3 | 99.68 | | 99.5 |

| | atoms (per 24 O, OH, F) | | | | | |
|---|---|---|---|---|---|---|
| Si | 5.62 } 8.00 | 5.47 } 8.00 | 5.49 } 8.00 | 5.24 } 8.00 | | 5.48 } 8.00 |
| Al | 2.38 | 2.53 | 2.51 | 2.76 | | 2.52 |
| Al | 0.64 | 0.74 | 0.93 | 0.80 | | 1.08 |
| Ti | 0.24 | 0.34 | 0.35 | 0.39 | | 0.39 |
| $Fe^{+3}$ | 0.47 } 5.28 | 0.29 } 5.25 | 0.22 } 5.47 | 0.06 } 5.46 | | 0.11 } 5.57 |
| $Fe^{+2}$ | 2.17 | 2.34 | 2.43 | 2.79 | | 3.27 |
| Mn | 0.02 | 0.04 | 0.02 | 0.02 | | 0.01 |
| Mg | 1.74 | 1.50 | 1.52 | 1.40 | | 0.71 |
| Ca | 0.02 | 0.02 | 0.02 | 0.02 | | 0.05 |
| Na | 0.07 } 1.72 | 0.04 } 1.77 | 0.05 } 1.79 | 0.04 } 1.67 | | 0.04 } 1.78 |
| K | 1.63 | 1.71 | 1.72 | 1.61 | | 1.69 |
| F | 0.10 } 4.40 | 0.09 } 4.47 | 0.11 } 3.83 | 0.12 } 4.39 | | 0.14 } 3.47 |
| OH | 4.30 | 4.38 | 3.72 | 4.27 | | 3.33 |

| Refractive index (±0.003) in Na light | | | | | |
|---|---|---|---|---|---|
| 1.658 | 1.656 | 1.653 | 1.653 | 1.659 | 1.668 |

| Oxidation ratio (O.R. = $2Fe_2O_3 \times 100 / 2Fe_2O_3 + FeO$ mol) | | | | | |
|---|---|---|---|---|---|
| 17.8 | 11.0 | 8.3 | 1.9 | 4.2 | 3.3 |

For location of samples see Fig. 3 and Table 1.
[1] Corrected for 0.35 wt. % carbon.

plagioclase about 8 mm in diameter. (Petrographic and chemical data for B116 are given in Tables 1–4.)

A highly garnetiferous sillimanite-quartz-biotite hornfels (B111, Tables 1, 2, 4) with traces of K-feldspar occurs on the east side of Long Canyon between the layer of marble and the batholith. The garnets are irregular in outline and enclose very numerous granules of minerals found elsewhere in the rock.

TABLE 3. CHEMICAL ANALYSES OF CORDIERITES FROM THE ISABELLA HORNFELSES

| | Wt. % Oxides | | | | Atoms (per 18 Oxygens) | |
|---|---|---|---|---|---|---|
| | B145 | B173 | B116 | | B116 | |
| $SiO_2$ | n.d. | n.d. | 49.66 | Si | 5.196 | 6.000 |
| $TiO_2$ | 0.15 | 0.19 | 0.09 | Al | .804 | |
| $Al_2O_3$ | n.d. | n.d. | 30.70 | Al | 2.980 | 2.987 |
| $Fe_2O_3$ | 4.55 | 3.42 | 2.68 | Ti | .007 | |
| FeO | 7.02 | 6.56 | 8.34 | Mg | .873 | |
| MnO | 0.32 | 0.42 | 0.28 | [1]$Fe^{2+}$ | .939 | |
| MgO | 5.57 | 5.35 | 5.60 | Mn | .025 | 1.985 |
| CaO | 0.11 | 0.24 | 0.16 | Na | .124 | |
| $Na_2O$ | n.d. | n.d. | 0.61 | Ca | .018 | |
| $K_2O$ | n.d. | n.d. | 0.04 | K | .006 | |
| $H_2O^-$ | n.d. | n.d. | 0.02 | | | |
| $H_2O^+$ | n.d. | n.d. | 1.62 | Optics | B116 | |
| Total | | | 99.80 | | | |
| | | | | $\alpha$ 1.545 | | |
| | | | | $\beta$ 1.552 | $2V_x 80$ | |
| | | | | $\gamma$ 1.557 | | |

For location of samples see Fig. 3 and Table 1.

[1] All Fe expressed as $Fe^{2+}$.

*Quartzo-feldspathic Segregations.* A notable feature of the pelitic hornfelses in Long Canyon is the presence of pegmatitic segregations of quartz, milky fibrolite, and, locally, perthite. These segregations are seldom lenticular, such as those commonly found in high-grade pelitic rocks, but more commonly appear as ramifying networks, several inches in diameter, in places with a thick central trunk lying transverse to a weak relict foliation. The segregations do not occur adjacent to the exposed contact of the batholith, but instead are developed in a small circular area central to the swarm of aplite-pegmatite dikes at the mouth of the Canyon and over a more widespread area south of the fold hinge in marble. In the bottom of the side canyon of the latter area the rocks are coarser than other pelitic hornfels and commonly migmatitic.

*Analyses and Compositions of Minerals.* Ideally it would be desirable to have data on the chemical composition of rocks in bulk and of each phase in the several pelitic assemblages described, so as to obtain a complete picture of the paragenesis of minerals during thermal metamorphism. In some instances a general notion of mineral compositions can be obtained by comparing physical properties (mainly optical) with those of minerals of known compositions. This approach is unfortunately sub-

TABLE 4. CHEMICAL ANALYSES OF GARNETS FROM THE ISABELLA HORNFELSES

| | Wt. % oxides | | Atoms (per 24 oxygens) | | |
|---|---|---|---|---|---|
| | B116 | B111[1] | | B116 | |
| $SiO_2$ | 36.36 | 39.73 | Si | 5.846 | 6.000 |
| $TiO_2$ | 0.59 | 0.53 | Al | 0.154 | |
| $Al_2O_3$ | 21.66 | 17.75 | Al | 3.950 | 4.012 |
| $Fe_2O_3$ | 3.14 | 7.86 | Ti | 0.062 | |
| FeO | 32.69 | 28.77 | $Fe^{3+}$ | 0.380 | |
| MnO | 1.99 | 2.43 | $Fe^{2+}$ | 4.395 | |
| MgO | 2.61 | 0.98 | Mg | 0.625 | 5.840 |
| CaO | 0.98 | 2.04 | Mn | 0.271 | |
| Total | 100.02 | 100.09 | Ca | 0.169 | |
| RI($\pm$.003) | 1.812 | 1.813 | almandine | 81 | |
| | | | pyrope | 11 | |
| $a$ ($\pm$.006 Å) | 11.530 Å | 11.541 Å | spessartite | 5 | |
| | | | andradite | 2 | |
| | | | grossularite | 1 | |

For location of samples see Fig. 3 and Table 1.

[1] Includes about 10% quartz and sillimanite contaminant.

ject to great uncertainties and ambiguity in the case of biotite, cordierite and garnet, the very minerals of interest in the Isabella hornfelses. For example, analytical studies by Engel and Engel (1960) on biotites and Leake (1960) on cordierites have demonstrated the uncertainty of making direct correlation between compositional variations and physical properties.

In the present investigation complete or partial analyses were made of representative biotites, cordierites and garnets. Mineral separations were made with a Frantz magnetic separator and heavy liquids. After repeated passes in the magnetic separator the biotites were estimated to be better than 98% pure. For cordierites (Table 3) fine grinding, which probably accounts for the high ratio $Fe_2O_3$/FeO (Chinner, 1962, Table 3), and repeated centrifuging in heavy liquid were necessary to produce a rela-

tively pure concentrate. In addition to sillimanite there were a few per cent of "pinite" remaining in samples B145 and B173, but this contaminant probably did not greatly influence the partial analyses. Cordierite B116 was about 98 pure; the number of atoms, based on 18 oxygens, in it was calculated by including both $Fe^{3+}$ and $Fe^{2+}$ and also by expressing all Fe as $Fe^{2+}$. The latter procedure gave a much better balance in the structural formula, and so this result is the one given in Table 3. Grinding to fine powder and centrifuging failed to remove the last 10% of quartz and minor sillimanite in garnet B111. Several recalculations were tried in an attempt to compensate for these contaminants and the unusually high $Fe_2O_3/FeO$ ratio, but none gave a really satisfactory mineral formula.

Most of the analytical work was done in the laboratory of E.L.P. Mercy in the Grant Institute of Geology, Edinburgh. $SiO_2$, $TiO_2$, $Al_2O_3$, MnO, and $K_2O$ in biotites B50, B89, B111 were analyzed by Robert N. Jack using an $x$-ray fluorescence spectrograph in the Department of Geology and Geophysics, University of California at Berkeley. The mineral sample was fused with lithium tetraborate and lanthanum oxide prior to examination in the spectrograph. Analytical standards used in this work included G-1 and W-1 (Fairbairn *et al.*, 1951), a split from biotite LEW 335 analyzed by Mercy, plus an analyzed pelitic schist BE-811 and a metagabbro BL-964 on hand in the Department.

Although only a limited number of analyses of biotites from the Isabella hornfelses were made, certain general comments regarding chemical variations relative to increasing metamorphic grade seem warranted. Some of the variations obvious in Table 2, *i.e.* $Fe^{2+}$ increasing and Mg decreasing, are closely related to mineral paragenesis and so will be considered in the next section. Other chemical variations, with respect to increasing metamorphic grade (left to right in Table 2), include increases in the $TiO_2$ content and in the molecular ratio.

$$\frac{Al_2O_3 \times 100}{Al_2O_3 + SiO_2},$$

but a decrease in the oxidation ratio (Chinner, 1960)

$$\text{O.R.} = \frac{2Fe_2O_3 \times 100}{2Fe_2O_3 + FeO}.$$

These molecular ratios are plotted together as Fig. 7. The possibility should be assessed as to whether the presence of $Al_2SiO_5$ contaminants in the higher grade biotites would have produced their higher values of

$$\frac{Al_2O_3 \times 100}{Al_2O_3 + SiO_2}.$$

B50, LEW 335 and B89 contain only minute amounts of $Al_2SiO_5$ impurity, but B111 and B173 contain somewhat more. However, even if

the analyzed B173 held 5% $Al_2SiO_5$—a liberal estimate—the ratio value
of pure biotite B173 would still be more than the lower grade biotites;
thus the variations depicted in Fig. 7 appear to be real. Whether biotites
from other pelitic aureoles also show increased Al with respect to Si with
increased grade is uncertain. Pitcher and Sinha (1957) analyzed four
biotites from an aureole in Donegal Co. but no consistent trend in ratio
values exists in their data. Moore's (1960) analyses of twelve biotites
from the Onawa aureole in Maine exhibit no significant variations. In
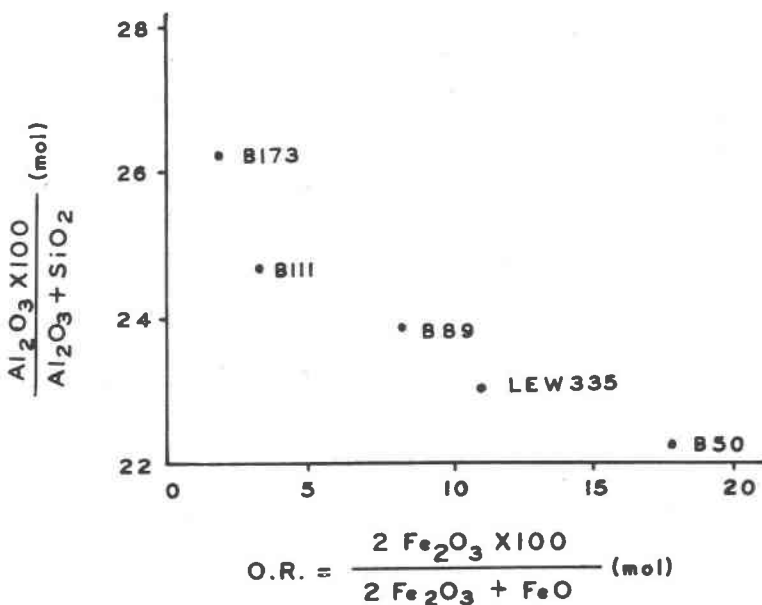


FIG. 7. Oxidation ratio plotted against an alumina-silica parameter for analyzed
biotites from Isabella hornfelses.

regional metamorphosed rocks (Lambert, 1959, Snelling, 1957; Kretz,
1959) Al increases in higher grade biotites. According to Harry (1950)
higher temperature favors more Al in tetrahedral positions. In the Isa-
bella biotites the main variation of Al occurs in the octahedral sites,
though some change also appears in the tetrahedral sites.

Many investigators have found that the biotite in the inner parts of
pelitic aureoles is redder than that in the outer parts. This effect has been
ascribed to variations in Fe and Ti content and in the oxidation state of
Fe. Hayama (1959) reports that high Ti causes red coloration and high
$Fe^{3+}/Fe^{2+}$ green, but that the actual color is a balance of these two influ-
ences. The color change from biotite B50 (yellow-brown) to LEW 335
and other biotites (red-brown) above the K-feldspar isograd correlates

with a jump in $TiO_2$ from 2.2 to at least 3.5 wt.% and a drop in O. R. from 17.8 to as little as 1.9 (Table 2).

*Paragenesis of Minerals.* There are two independent factors governing the type of minerals co-existing stably together in a metamorphic rock (Thompson, 1957). One of these, the bulk chemical composition of the rock or the relative amounts of inert components such as $SiO_2, K_2O$, etc., can not be discussed directly in the case of the Isabella hornfelses because no bulk analyses are available. Nonetheless certain indirect approximations may be made on the basis of petrographic and mineralogic data, and in any case it seems all hornfelses herein described have, with few exceptions (*viz.* the garnetiferous B111), rather similar compositions. This conclusion is based on the absence of large, mapable variations in pelites—visible relict sedimentary beds are rare even in single exposures—together with the gross petrographic uniformity apparent in thin sections.

The second factor influencing the kind of minerals coexisting together in pelites is the set of external variables P, T, and $\mu_m$ (the chemical potentials of mobile components $O_2$[1] and $H_2O$ whose values are fixed by restraints external to the rock body). In discussions of metamorphic rocks, reference is frequently made, as has been done above, to the "grade" or "degree" of metamorphism. In regional metamorphism higher grade rocks are usually taken to mean those rocks recrystallized at higher T; variations in P and $\mu_m$ are generally considered to be of secondary importance. While it is possible to correlate higher grade thermally metamorphosed rocks with higher T there is occasion for uncertainty because particular prograde reactions might have been induced by substantial changes in $\mu_{H_2O}$ and / or $\mu_{O_2}$, with increasing proximity to the igneous intrusion. Little confusion will be caused if at this point it is noted that the higher grade hornfelses at Isabella are generally coarser, have a more pronounced hornfelsic fabric, and occur nearer to exposed igneous bodies than do lower grade rocks.

To expedite the discussion of mineral paragenesis an AFM diagram, derived by projecting compositional points in the tetrahedron AKFM onto one face, will be employed to represent phase relations. This diagram has been developed and used extensively by Thompson (1957) and his students (*e.g.* Zen, 1960; Phinney, 1963; Moore, 1960) in analyzing mineral assemblages in metamorphic rocks, and no mention will be made

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[1] Eugster (1959, p. 423) gives reason for considering hydrogen to be more mobile than oxygen during metamorphism, and hence it may be more appropriate to count hydrogen rather than oxygen. There is however a dependency between the chemical potentials of these two gases and so diffusion of hydrogen into a system producing a certain redox reaction may be viewed as a loss of oxygen.

here of the theory and mechanics of projection. It is worthy of note however that this diagram does not account for the components $Na_2O$, $CaO$, and $MnO$. The results of mineral analyses and the subsidiary occurrence of modal plagioclase indicate the first two components are not generally important in Isabella pelites, the small amount present in the rock being readily taken up by other minerals. $MnO$ will also be taken into ferromagnesian minerals if present in small amounts. In the event that the amount of $MnO$ or $CaO$ in the rock exceeds the solubility limit of these components in constituent minerals at some set of values of external variables, then an independent phase, *e.g.*, garnet, will develop and phase relations cannot be adequately represented in the usual AFM plot (Thompson, 1957). The apex F in Figs. 8, 9 and 10 equals $FeO + 2Fe_2O_3 - TiO_2$ (molecular) which in the Isabella hornfelses takes into consideration the ubiquitous occurrence of ilmenite and the apparent behavior of $O_2$ as a mobile component during metamorphism (see below) so that $FeO$ and $Fe_2O_3$ are regarded as a single independent component (Zen, 1960, p. 164). The AFM plot with $Fe_2O_3$ incorporated into F shows a range of compositional variation of Isabella biotites shorter than in a plot ignoring $Fe_2O_3$.

Compositional relations of phases in the regional phyllites, composed of assemblage I in Fig. 3, are unknown, as are the reactions by which that assemblage gave way to the lowest grade thermal assemblage II. A rock composed of assemblage II would plot at some point such as X along the tie line connecting A and B50 in Fig. 8 if it were expressed in terms of the three components A, F and M.

Rocks closer to the exposed batholith are coarser, more decussate and contain perthitic microcline. Muscovite is clearly destroyed at the appearance of feldspar (Fig. 4), suggesting the reaction.

$$\text{muscovite} + \text{quartz} \rightarrow \text{perthite} + Al_2SiO_5 + H_2O \tag{1}$$
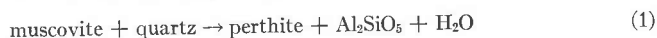
This reaction proceeds to the right at increased temperature and presumably proximity to the batholith is the factor responsible. The alkali feldspar in these rocks is obviously perthitic, a feature which has been noted in other instances of prograde metamorphism (usually of regional character however). Guidotti (1963) studied some pelites relatively rich in $Na_2O$ and $CaO$ and concluded from their phase relations that both of these components were relevant to the reaction whereby muscovite breaks down to alkali feldspar. However the pelites on the east side of Lynch Canyon, *i.e.* those now under discussion, appear to be relatively poor in $Na_2O$ and $CaO$ because (a) plagioclase has not been found in any rocks of assemblages I and II, (b) muscovites are uniformly poor in $Na_2O$, judging from *x*-ray diffraction measurements of the basal lattice spacing, and (c)

the analyzed biotites B50 and LEW 335 contain relatively little of these components (compare for example, analyses in Table 13 in Deer *et al.*, vol. 3, 1963). The $Na_2O$ for the perthite must have been derived from the muscovite, albeit a small amount, but possibly augmented by release of
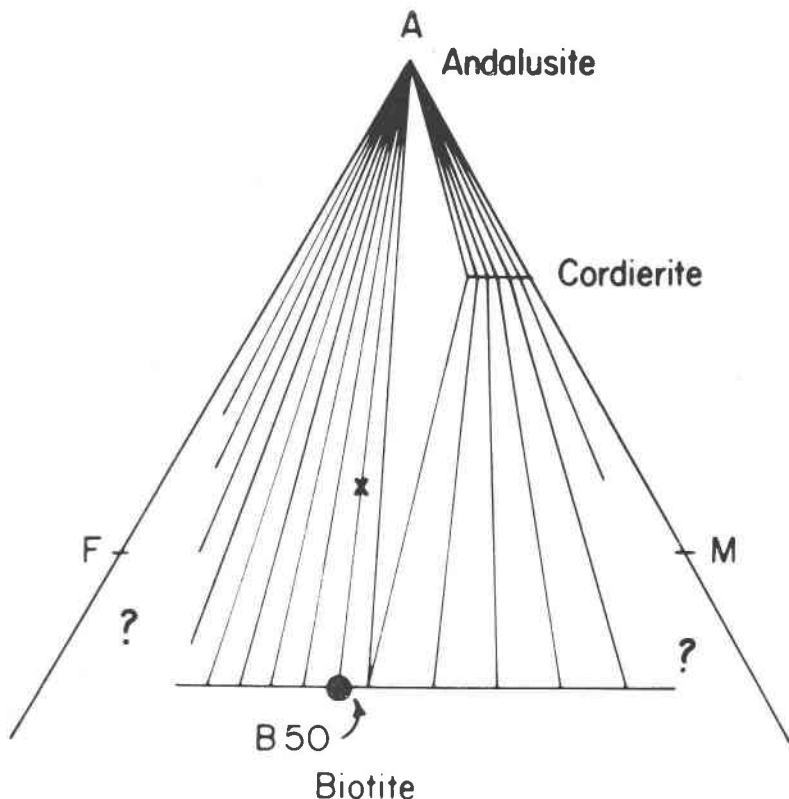


FIG. 8. Possible phase relations (plus quartz, muscovite, and ilmenite) in lower grades of thermal metamorphism of pelitic rocks. Some point such as X represents the bulk composition of the rock B50, the biotite from which is shown at the end of the tie-line through X. The exact position of the limiting andalusite-biotite tie line is not known but it must lie to the right of A-B50. Diagram is slightly modified after Moore (1960, Fig. 18f).

Na from biotite (compare B50 and LEW 335, Table 2). Scrutiny of the immediate area around the perthite metacrysts failed to reveal the location of the additional $Al_2SiO_5$ produced by reaction (1). There is the possibility that the surplus $Al_2O_3$ from breakdown of muscovite migrated to other parts of the rock. An alternative explanation—formation of a biotite at the isograd with less $K_2O$ and this surplus oxide combining with the surplus $Al_2O_3$ and quartz forming more K-feldspar—is not supported

by the analytical data in Table 2 which shows the $K_2O$ content of all Isa-
bella biotites to be essentially constant, regardless of grade.

The absence of cordierite in muscovitic and some perthitic assem-
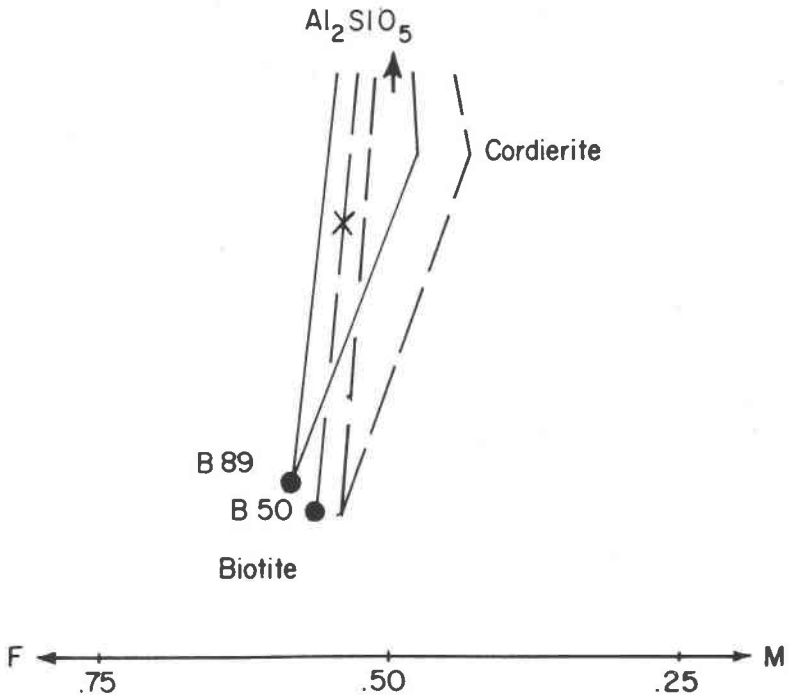blages means either that sufficiently magnesian rocks whose bulk com-



FIG. 9. A portion of the AFM system (plus quartz, K-feldspar, and ilmenite) represent-
ing reaction (2). Phase relations under low grade conditions at which rock B50 formed are
shown by dashed tie lines. At higher grade for rock B89 full lines apply. X represents the
bulk composition of pelitic hornfels which under low grade conditions lies in the two-phase
field $Al_2SiO_5$ biotite. As metamorphic grade increases the three phase field of $Al_2SiO_5$-
biotite-cordierite moves (from some position to the right of B50) to a more Fe-rich position
so that X lies within it. Biotites become richer in Fe with increasing grade. Data pertaining
to sample B50 are plotted *as if* K-feldspar were the K-silicate in the rock.

positions would fall in the cordierite stability field (Fig. 8) do not occur at
Isabella, or that cordierite was not a stable phase at the prevailing set of
external variables, or both. One might expect a stable field of magnesian
chlorite near the AM side of Fig. 8 if cordierite were not present, and pos-
sibly even if it were. Absence of chlorite in thermal assemblages at Isa-
bella suggests that bulk compositions were too rich in Fe, though other
possibilities exist. Moore (1960) observed a prograde transition from

chlorite- to cordierite-bearing assemblages in the outer aureole of the Onawa pluton in Maine.

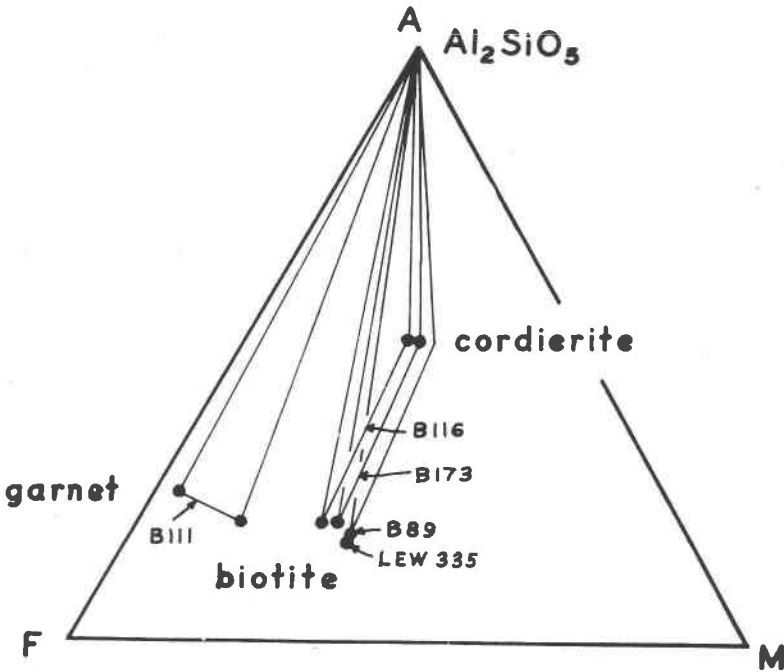The development of cordierite in the Isabella hornfelses may be explained by the endothermic reaction

$$\text{biotite}_1 + Al_2SiO_5 + \text{quartz} \rightarrow \text{K-feldspar} + \text{cordierite} + \text{biotite}_2 + H_2O \qquad (2)$$



FIG. 10. Synoptic phase relations (plus quartz, K-feldspar and ilmenite) in the Isabella hornfelses for *different* values of external variables (*i.e.* non-isofacial conditions). Illustrated is the shift to the left with increasing grade of the three phase field $Al_2SiO_5$-biotite-cordierite. Tie lines involving cordierite B89 are drawn so as to be concordant with points for analyzed cordierite.

where $Mg/Fe^{2+}$ is greater in biotite$_1$ than in biotite$_2$. Although no clear textural evidence for this reaction was found, coarser higher grade hornfelses generally contain (in addition to cordierite) more K-feldspar and less quartz, $Al_2SiO_5$, and biotite than do low grade rocks. Higher grade biotites are richer in $Fe^{2+}$ relative to Mg than those of low grade (Table 2), suggesting that the three-phase field of $Al_2SiO_5$-cordierite-biotite has shifted so as to encompass more Fe-rich compositions which under low grade conditions were located in the two-phase field of $Al_2SiO_5$-biotite. This shift is represented graphically in Fig. 9 which shows a bulk com-

position point X eclipsed by the three-phase field which has moved toward F from a (dotted) position at low grade. X at low grade lies in the two-phase field. The reaction (2) would produce a continuous variation (Thompson, 1957, p. 855) in the composition of the biotites and would produce no major topological change in the AFM diagram. One could envisage a succession of reactions (2), each involving development of a more Fe-rich biotite from one with less Fe, and represented by a succession of $Al_2SiO_5$-biotite tie-lines moving toward the AF side of the diagram (Fig. 10).

The increased proportion of $Fe^{2+}$ in higher grade biotites is an important aspect of the Isabella hornfelses. Moore (1960) also found the same variation in biotites in the Onawa aureole. The fact that other investigators (see the discussion in Deer, *et al.*, vol. 3, p. 74–75) have found the opposite variation merely serves to emphasize that variations in chemical composition of minerals within different metamorphic facies depends upon the reactions giving rise to the minerals and upon the nature of co-existing phases. The decrease of $Fe^{2+}$ in higher grade biotites found by Barth (1936) and Engel and Engel (1960) occurs in garnetiferous rocks whereas data presented herein and by Moore are for cordieritic rocks. Further emphasis on this point is provided by an examination of the left side of Figs. 8 and 10. In Fig. 8 a question mark is shown near F because of lack of pertinent data for such compositions. However, it might be expected that the two-phase field of biotite-$Al_2SiO_5$ extends nearly to the AF boundary because of the rarity (absence?) of almandine in low grade pelitic hornfelses. At higher grades a field of stable almandine and a three-phase region for almandine-biotite-$Al_2SiO_5$ do develop as shown by the data for rock B111 in Fig. 10 (see also Best, 1963, for similar phase relations in another part of the Sierra Nevada). Thus biotites in almanditic hornfelses become more magnesian with increasing grade, concurrent with enrichment of Mg in the almandine.

It thus appears that the field of biotite which can coexist in equilibrium with $Al_2SiO_5$ is shrinking as metamorphic grade increases. For some set of external conditions (not found at Isabella) biotite may no longer coexist stably with $Al_2SiO_5$ and phase relations such as depicted in Fig. 11 (see Chinner, 1962, Fig. 10) might then hold.

Chinner (1962) shows a series of AFM diagrams (his Fig. 10) representing phase relations in pelitic rocks under a variety of external conditions, P and T. It is worthwhile to note here that in none of his diagrams is $Al_2SiO_5$ stable with biotite. On petrographic grounds, at least, it would seem that a field of biotite-$Al_2SiO_5$ does exist in nature,[1] and possibly over

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[1] Excluded here is the remote possibility that all such assemblages are metastable.

a fair range of P and T, because of the common occurrence of this mineral pair (*cf.* Fyfe, *et al.* 1958, p. 207). Although cordierites rich in Fe and with Mg/Fe > 1 are common, intermediate cordierite appears to be non-existent (compare diagrams in Leake, 1960), its place, in the absence of garnet, being taken by the pair biotite-$Al_2SiO_5$ (Fig. 8).

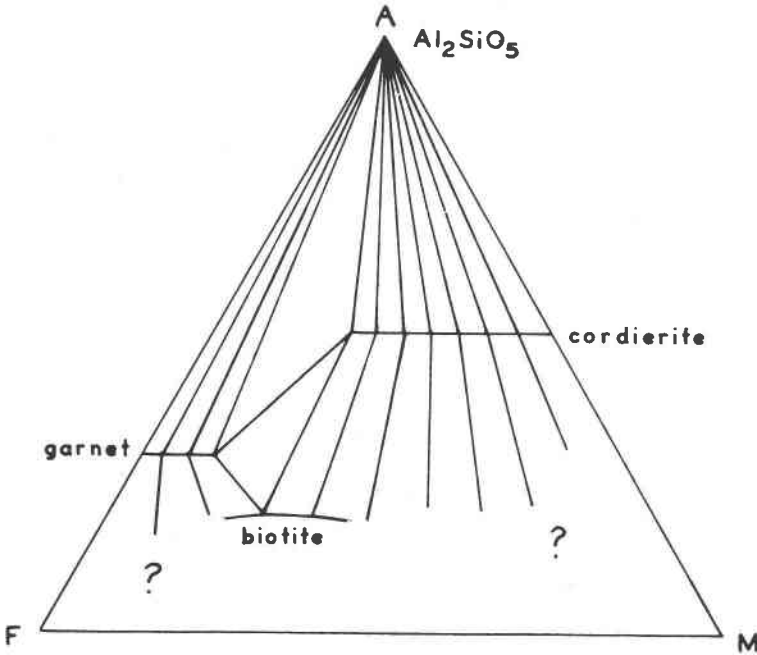Much has been written recently concerning the paragenesis of garnet in pelitic hornfelses and there is no need here to review the details, which



Fig. 11. Possible phase relations in the AFM system at a higher temperature than depicted in Fig. 10.

can be found in Chinner (1962) or Wynne-Edwards and Hay (1963). Assemblage B111, lacking cordierite, is what one might expect to find in an Fe-rich rock, Fig. 10, at some particular set of external conditions P, T, and $\mu_m$. The garnet-poor hornfelses at the mouth of Long Canyon contain perthite and cordierite in addition to the phases occurring in B111. Partial mantles of garnet around andalusite in one sample may indicate an arrested reaction which, generalized, may be of the form

$$\text{biotite} + \text{andalusite} + \text{quartz} \rightarrow \text{garnet} + \text{cordierite} + \text{feldspar} + H_2O \qquad (3)$$
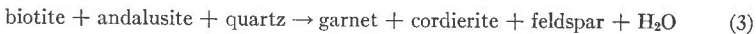
The general validity of this reaction for the rocks in Long Canyon is, however, somewhat uncertain, because in the rock with mantles of garnet

there is retrogressive alteration of some of the feldspars and cordierites, which may have some bearing on the paragenesis of the garnet, and be-cause the composition of the garnet and biotite is not known. The third type of garnetiferous pelite is exemplified by B116. Mineralogically it differs from those in Long Canyon in lacking perthite. It is coarser, reaction relationships are not present, and textural stability seems to have been attained.

If, as is commonly done, we count $Al_2O_3$, FeO, MgO, $K_2O$, $SiO_2$ and $Na_2O$ as the inert components[1] relevant to the pelitic system, then there is a violation of the phase rule if the phases $Al_2SiO_5$, quartz, biotite, cordierite, garnet, perthite and plagioclase (which add up to one more than the number of components) are all present. If perthite is absent, as in B116, the phase rule would be satisfied and one could presume that the presence of biotite is called for because of the component $K_2O$.

The whole interpretation of the paragenesis of garnet can perhaps be better understood if one considers the role played by other components, such as MnO and CaO, in stabilizing almandine garnets. This approach, alluded to by Thompson (1957, p. 583), would hopefully cast some light on the paragenesis of the perthite-almandine-biotite-cordierite-$Al_2SiO_5$ assemblages at Isabella which are found, as well, in many other metamorphic terrains (cf. Barker, 1962). Wynne-Edwards and Hay (1963) present analytical data from regional gneisses in southeastern Ontario suggesting that the appearance of almandine in such assemblages occurs when the Ca content and the ratio Fe/Mg of the rock are both high. Garnetiferous rocks studied by them contain more plagioclase with a greater An content than do similar but garnet-free rocks. A survey of 14 garnet-bearing and garnet-free cordierite-perthite hornfelses from Isabella disclosed that in the former the An content of plagioclases ranged from 21 to 42 with a mean of 30 whereas in the latter group the range is 20 to 29 with a mean of 23. Unfortunately, there are insufficient chemical data from thermally metamorphosed rocks containing the critical assemblage to establish compositional fields in which garnet would or would not be expected. A firmer statement could perhaps be made concerning the stabilizing component in the Isabella garnets if this phase were more abundant and more analytical data were available.

*Status of $Al_2SiO_5$.* The very common coexistence of sillimanite and andalusite, apparently in textural equilibrium in muscovite-free Isabella hornfelses, has been previously mentioned. It is by no means rare that

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[1] Consider here that the system is open to $O_2$ and $H_2O$ and that general conditions of P, T and $\mu_m$ prevail.

both polymorphs are found together, and in fact rocks have been found in some parts of the world which contain all three polymorphs of $Al_2SiO_5$ (Hietanen, 1956; Woodland, 1963). Only if foreign constituents (*e.g.* Mn, Fe) are stabilizing one of the polymorphs or if the external conditions during metamorphism are appropriate to univariant equilibrium in the $Al_2SiO_5$ system can andalusite and sillimanite coexist in equilibrium (Chinner 1961, p. 317). In the Isabella hornfelses, these polymorphs are colorless, and may therefore lack appreciable foreign ions, and external variables P and T must have varied over the body of hornfelses, precluding a chance set of particular values appropriate for univariant equilibrium. It may be concluded then that andalusite persisted metastably (for experimental data bearing on this point, see Weill and Fyfe, 1961) into the P, T field wherein sillimanite was stable. In the higher grade hornfelses, *e.g.* B116, B145, it appears, however, that andalusite was nearly eliminated because it occurs in minor amounts as small clean granules. There may be some connection between the common coexistence of andalusite and sillimanite in hornfelses and a region in P, T space where experimental syntheses yield inconclusive results as to the stable form of $Al_2SiO_5$ (Weill, 1963; Bell, 1963).

Textural relations of sillimanite in Isabella hornfels bearing on its nucleation and growth are similar to those commonly observed elsewhere. Only in rare instances do needles of sillimanite penetrate into andalusite; whisker-like growth of sillimanite on andalusite (Compton, 1960, Pl. 3, Fig. 3), although not common, was noted in a few thin sections. More generally sillimanite is intimately intergrown with biotite; individual needles may be locally aligned parallel to the pressure-figure directions in host biotite (Chinner, 1961). In higher grade cordieritic rocks sillimanite is concentrated along grain boundaries with the longer needles penetrating into adjacent cordierite or quartz. In the absence of compelling evidence to the contrary it appears that breakdown of biotite was not a factor in the development of sillimanite. Rather, biotite seems to have acted as a nucleating agent for epitaxial growth of sillimanite, as shown by Chinner (1961) for some gneisses at Glen Clova, Angus, the necessary Al and Si coming primarily from the destruction of andalusite and carried via pore fluids to other parts of the rock. Of course, the breakdown of muscovite following the reaction (1) furnished $Al_2SiO_5$ in the lower grade rocks.

*Status of Mobile Components.* As far as the graphical representation of mineral assemblages is concerned, it makes no difference whether $H_2O$ during metamorphism behaves as a mobile component or as an inert component in excess (in which case a separate phase, water, appears). Hence

no conclusion can be drawn on this basis as to whether, in the Isabella hornfelses, $H_2O$ behaved as a mobile or inert component. However, it is apparent that $H_2O$ was evolved during progressive metamorphism (note reactions (1) and (2)), and, assuming that the porosity of high and low grade rocks remained essentially constant, there must have been a net migration of $H_2O$ away from the high grade rocks.

Similar ambiguities are involved in interpreting the status of $O_2$ in metamorphism (Zen, 1960, p. 164). Presence in a metamorphic rock of only one Fe oxide, say hematite, could mean that $O_2$ was either mobile (the value of $O_2$ varying within the limits of the stability of hematite) or that $O_2$ was inert and in excess. If two oxides, say hematite and magnetite, coexist then it is more probable that $O_2$ was inert and the initial $Fe_2O_3$—FeO content of the rock influenced the final compatible phases (Chinner, 1960, p. 210).

In most Isabella hornfelses ilmenite is the sole (Fe, Ti) oxide. Magnetite was positively identified only in some of the garnetiferous hornfelses. Polished sections of many hornfelses indicate the ilmenite is homogeneous, except for that in B116 wherein one ilmenite grain contained small blebs of hematite[1]—presumably exsolved during slow cooling of an initially homogeneous $Fe_2O_3$—Fe $TiO_3$ phase.

Chinner (1960, p. 211–212) discusses in some detail the case for an environment "open to oxygen" in certain contact metamorphic aureoles. Examples are cited suggesting that, under favorable circumstances, there tends to be an equilibration between the oxidation state of Fe in the aureole rocks with that in the intrusive magma. Strongly oxidized parent rocks containing hematite may be reduced, owing to the influence of a gas phase whose $\mu_{O_2}$ is "buffered" by the ferrous oxides in the magma. This very condition seems to have occurred in the Isabella rocks during their thermal metamorphism. Regional rocks (assemblage I, Fig. 3) containing hematite as the sole (Fe, Ti) oxide give way to ilmenite hornfels near the batholith. In addition there is more $Fe^{2+}$ relative to $Fe^{3+}$ in higher grade biotites than in lower grade, Fig. 7. The actual number of recorded instances where oxygen has behaved as a mobile component in contact aureoles is not great; this lack may be related to the limited number of detailed studies of thermal aureoles, but it might also reflect very real factors such as insufficient time for attainment of equilibrium of $\mu_{O_2}$ between magma and country rocks or to the impermeable character of the latter.

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[1] Kindly identified by W. Petruk.

## Nature of Metamorphism

The sequence of first appearance of new minerals (andalusite or biotite-K-feldspar-cordierite) at Isabella differs from the sequence recorded in other pelitic aureoles where cordierite precedes K-feldspar, as for example at Onawa (Moore, 1960) and Santa Rosa (Compton, 1960). One explanation is that this reversal merely reflects a difference in bulk composition of the rocks, those from Isabella being as a whole richer in Fe so that cordierite does not appear until its stability field has so expanded to include the rock composition (Fig. 9). However, Isabella biotites as a group are more magnesian than Onawa biotites, thus weakening any argument based on such differences in bulk composition. There may have been significant differences in $\mu_{H_2O}$ or $\mu_{O_2}$ during metamorphism at Onawa and at Isabella, but this is conjectural as no specific arguments can be proposed.

It is difficult to specify in which facies of metamorphism the Isabella pelites should be placed because metamorphic rocks of other bulk compositions are very sparse and because of an ambiguity of correlation with established facies assemblages (*e.g.* Fyfe *et al.*, 1958). Study of available calcareous and mafic rocks in areas III and IV (Fig. 3) suggests metamorphism under hornblende hornfels facies conditions rather than pyroxene hornfels. On the other hand, Turner (in Fyfe, *et al.*, 1958, p. 212) notes the coexistence of cordierite and $Al_2SiO_5$ with K-feldspar under pyroxene hornfels conditions, not hornblende hornfels.

The sequence of mineralogical changes—embodying as well chemical variations in biotites and cordierites—in the Isabella hornfelses is related to a gradient of one or more of the external conditions P, T, $\mu_{O_2}$ $\mu_{H_2O}$. A gradient in P (*i.e.* load pressure) seems unlikely in this thermal aureole as well as in other where the vertical and horizontal extent of the exposed metamorphosed rocks is limited. The possibility of a gradient in $\mu_{O_2}$ toward smaller values nearer the Isabella batholith has already been discussed. One would expect that in an aureole of pelitic rocks around a granitic intrusion that $\mu_{H_2O}$ and T would both increase toward the intrusion. Reactions (1) and (2) would proceed to the right for increasing T or for decreasing $\mu_{H_2O}$. Hence, in the Isabella hornfelses increasing T must have been more dominant than increasing $\mu_{H_2O}$ since the reaction products did develop in rocks nearer to the batholith. It thus seems that the gradient in T was the most significant factor which produced the observed prograde mineralogical changes in the pelites.

Any attempts to fit the mineralogical changes in the hornfelses to experimentally determined phase relations can be at best only approximate. A complete correlation would have to include evaluation of mobile

components as well as P and T. The upper T limit of the metamorphism is set by the minimum melting of granitic rocks, a limit which seems to have been nearly or just reached in migmatitic rocks south of the fold in marble (Fig. 2). The lower T limit for assemblages III and IV is set by the curve representing reaction (1) (*cf*. Signet and Kennedy, 1961). The shift in composition of coexisting biotite-cordierite (Fig. 10) in hornfelses above the K-feldspar isograd suggests that a substantial range in T existed during metamorphism between rocks at the isograd and rocks of highest grade, such as B116 and B173. Evaluation of P, a confining pressure commonly assumed to be approximately equal to the partial pressure of $H_2O$, is difficult. It would seem however that P was less than that commonly existing during regional metamorphism because staurolite and kyanite do not occur in Isabella hornfelses of appropriate composition. If Bell's (1963) data on $Al_2SiO_5$ polymorphs are reliable P must not have exceeded 6 or 7 kilobars.

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