THE AMERICAN MINERALOGIST, VOL. 49, SEPTEMBER-OCTOBER, 1964

MINERALOGY AND PARAGENESIS OF AMPHIBOLES FROM GIBSON PEAK PLUTON, NORTHERN CALIFORNIA

Peter W. Lipman, *U. S. Geological Survey, Denver, Colorado.*

Abstract

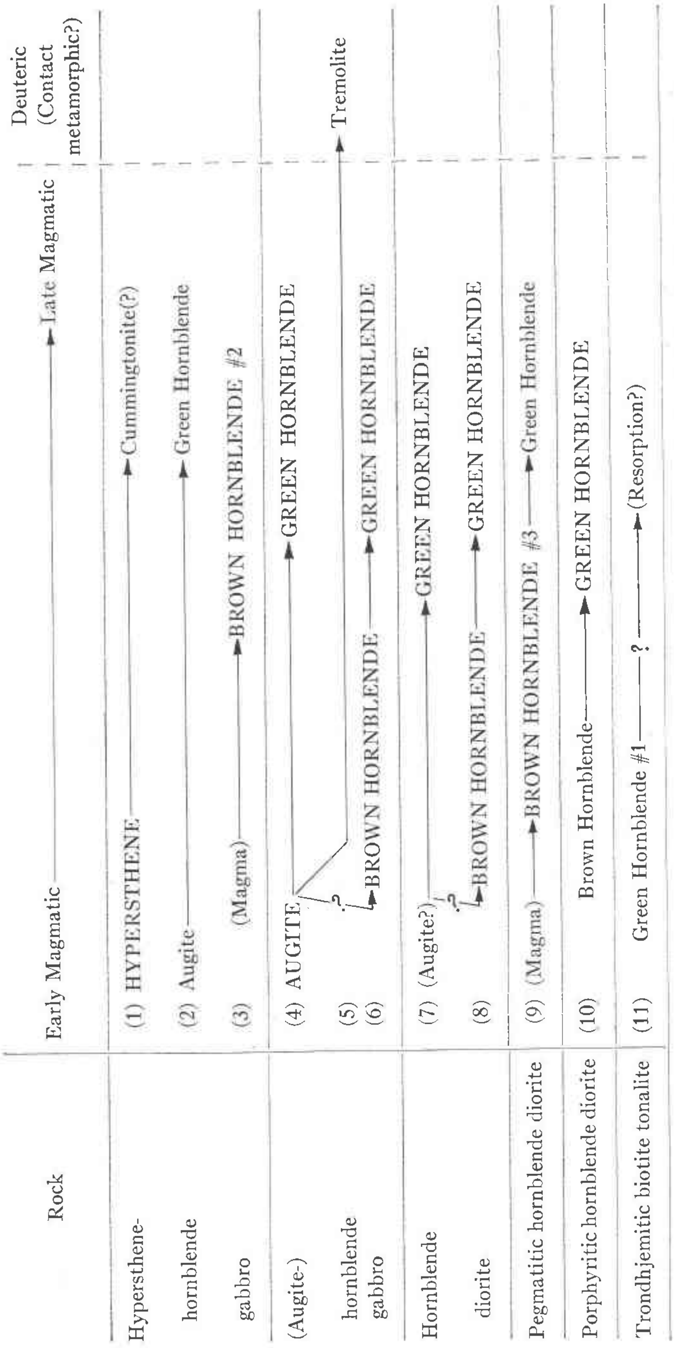
The compositions of hornblendes from Gibson Peak pluton, a composite intrusion ranging from hypersthene gabbro to trondhjemitic tonalite, appear related as much to position within crystallization sequence as to bulk rock composition. Early-formed hornblende in trondhjemitic tonalite has the composition $(Na, K)_{0.3}Ca_{1.8}(Mg, Fe^{2+}, Mn)_{4.7}Fe^{3+}_{0.6}Si_{7.0}Al_{0.8}Ti_{0.1}O_{22}(OH, F, Cl)_{2.0}$. Late poikilitic hornblende from hypersthene gabbro has the composition $(Na, K)_{0.5}Ca_{1.8}(Mg, Fe^{2+}, Mn)_{4.6}(Fe^{3+}, Ti)_{0.7}Si_{6.5}Al_{1.3}Ti_{0.2}O_{22}(OH, F, Cl)_{1.9}$. Hornblende from a pegmatitic residual segregation in diorite has the composition $(NaK)_{0.7}Ca_{1.9}(Mg, Fe^{2+}, Mn)_{4.5}(Fe^{3+}, Ti)_{0.8}Si_{6.1}Al_{1.7}Ti_{0.2}O_{22}(OH, F, Cl)_{1.9}$.

Introduction

Gibson Peak pluton, a 3-square mile composite intrusion in the Trinity Alps of northern California, is particularly suitable for investigation of relations between amphibole paragenesis and igneous crystallization because several distinctive amphiboles are important constituents of genetically related rocks that range from gabbro to trondhjemitic tonalite. This paper describes the sequence of amphibole crystallization in different parts of the intrusion and relates the compositions of three newly analyzed amphiboles to crystallization sequence and composition of the enclosing rock. The main conclusion is that compositions of the investigated amphiboles are as dependent on time of crystallization within their respective rocks as on bulk rock composition.

Petrographic Interpretation of the Amphibole Paragenesis

The general structural and petrologic features of Gibson Peak pluton are described elsewhere (Lipman, 1963), and only relations bearing on the origin of the amphiboles are summarized here. The pluton is composite, and five discrete intrusive units have been recognized on the basis of field relations. In order of intrusion these are hypersthene-hornblende gabbro, (augite-)hornblende gabbro, hornblende diorite, porphyritic quartz-bearing diorite, and trondhjemitic biotite tonalite. All units show intrusive contacts with the preceding rocks, are petrographically distinctive, and contain at least one amphibole. An interpretation of the complex paragenesis of the Gibson Peak amphiboles, based mainly on the textural features described below, is presented in Fig. 1. The evidence is clear on the occurrence of the indicated reactions, but the rela-



Principal minerals in capital letters

#2—Number of an analyzed hornblende. Analysis shown in Table 1.

Cummingtonite (?)—Questionable identification

(Augite?)—Questionable reaction component

FIG. 1. Paragenetic sequence of amphiboles in Gibson Peak pluton.

TABLE 1. COMPOSITION AND PHYSICAL PROPERTIES OF AMPHIBOLES FROM GIBSON PEAK PLUTON, MINERSVILLE (15′) QUADRANGLE, CALIFORNIA

[Analyses by W. H. Herdsman, Glasgow, Scotland]

	1	2	3
SiO_2	47.28	44.08	41.32
Al_2O_3	4.46	7.61	9.42
TiO_3	.94	2.47	2.51
Fe_2O_3	5.74	5.03	6.65
FeO	10.92	10.34	8.61
MgO	14.13	14.79	15.29
MnO	1.61	.44	.09
CaO	11.26	11.48	11.68
Na_2O	.93	1.54	2.14
K_2O	.39	.39	.29
H_2O^+	2.04	1.92	1.84
H_2O^-	.12	.00	.08
F	.00	.06	.08
Cl	.03	.03	.05
Total	99.85	100.18	100.05
Less 0 for F, Cl	.01	.03	.04
Total (corrected)	99.84	100.15	100.01
Specific gravity	3.18	3.26	3.26
α	1.647	1.656	1.657
γ	1.671	1.679	1.682
$\gamma-\alpha$	0.024	0.023	0.025
$2V\alpha$	64 ± 1	77 ± 2	88 ± 2
$Z\wedge c$	23 ± 2	19 ± 2	17 ± 1
Orientation	$Y=b$	$Y=b$	$Y=b$
X	pale straw brown	straw	tan
Y	pale green brown	green brown	pale brown
Z	green	deep green brown	yellow brown
Absorption	$Z>Y=X$	$Z>Y>X$	$Z>Y>X$

1. Green hornblende from trondhjemitic biotite tonalite; lat. 40°52′20″N., long. 122°52′21″W.
2. Brown hornblende from hypersthene-hornblende gabbro; lat. 40°51′42″N., long. 122°52′43″W.
3. Brown hornblende from pegmatitic hornblende diorite; lat. 40°51′56″N., long. 122°52′51″W.

tive positions of different reactions in the crystallization sequence are more interpretive. The analyzed amphiboles (Table 1) are:

(1) an early-formed green hornblende from the trondhjemitic tonalite,
(2) a late-formed brown hornblende from the hypersthene gabbro, and
(3) a brown hornblende from a pegmatitic phase of the hornblende diorite.

All three hornblendes seem to have formed by primary cystallization from their respective magmas; difficulties inherent in the interpretation of amphiboles that have formed by recrystallization from earlier minerals are absent.

The hypersthene-hornblende gabbro contains three different amphiboles: brown hornblende, green hornblende, and cummingtonite(?). Poikilitic brown hornblende grains as much as 15 mm in diameter enclose smaller plagioclase and pyroxene grains, giving the gabbro a distinctive appearance. Enclosed plagioclase and pyroxene are both euhedral, and the tabular plagioclase shows a good planar flow foliation (Lipman, 1963, Pl. 3, Fig. 4), textural features that indicate interstitial formation of the brown hornblende during the later stages of crystallization (Fig. 1, reaction 3). The green hornblende (Fig. 1, reaction 2), which resembles the common hornblende of the more felsic intrusive units, rims augite and in places hypersthene. It nowhere occurs as discrete crystals and was apparently formed by reaction between pyroxene and residual magma, the chemically similar augite having reacted more readily than hypersthene. A third amphibole, colorless cummingtonite(?) with a large positive 2V, forms narrow rims around some hypersthene (Fig. 1, reaction 1). The analyzed poikilitic brown hornblende (Table 1, no. 2) is from a specimen somewhat lower in hornblende and higher in pyroxene than the average hypersthene-hornblende gabbro (Lipman, 1963, Table 1); it has the following mode:[1]

Plagioclase (An_{66-47})	55.5	Hypersthene	12.0
Brown hornblende	21.5	Augite	6.5
Green hornblende	1.0	Opaque minerals	3.0
Cummingtonite(?)	0.1	Apatite	0.5

Most of the green hornblende and cummingtonite(?), atypically sparse in this specimen, was removed from the analyzed material during separation.

The (augite-)hornblende gabbro, the other mafic unit of Gibson Peak pluton, is typically coarse-grained and composed primarily of plagicclase and stubby green-brown hornblende. It everywhere lacks hypersthene, but relict cores of augite have survived in places within the hornblende. These scattered pyroxene relicts and the stubby shape of the hornblende suggest that all hornblende is probably pseudomorphous after augite (Fig. 1, reaction 4). Individual hornblende grains in the gabbro are typically mottled; their pleochroism, birefringence, and refractive indices vary unevenly. Crude zonation of some grains from brownish centers to greenish margins and analogous color variation from

---

[1] Modal compositions are stated in volume per cent and are based on counts of at least 1,500 points in a single 25×40 mm thin section.

specimen to specimen suggest widespread replacement of early brown hornblende by green hornblende (Fig. 1, reaction 6). Within some hornblendes colorless patches ($\gamma = 1.64$) that optically resemble tremolite may have formed by deuteric replacement of augite cores (Fig. 1, reaction 5).

The hornblende diorite intrusive unit, although compositionally intermediate, shows many textural and mineralogical affinities with the (augite-)hornblende gabbro, and these two rocks are locally gradational in the field. Some green or green-brown hornblende in the diorite is subequent and is presumably pseudomorphous after augite, as in the gabbro (Fig. 1, reaction 7). Other hornblendes in the diorite are strikingly acicular, with lengths as much as 5 times widths, and are thought to represent primary crystallization from magma. Color variations in the acicular hornblende of the diorite also suggest reaction of brown to green hornblende (Fig. 1, reaction 8).

Pegmatitic diorite, containing skeletal acicular hornblende crystals as much as 15 cm long (Lipman, 1963, Pl. 2, Fig. 3), is a striking variant of normal hornblende diorite, with which the pegmatitic rock typically intergrades irregularly. Hornblende constitutes as much as 70 per cent of the pegmatitic rocks, which are generally more mafic than adjacent normal diorite. Plagioclase is the only other major constituent of the pegmatite, and its composition is similar to that of the diorite. The texture, composition, and distribution of the pegmatite suggest formation by residual segregation of dioritic magma rich in volatile constituents. High volatile content is indicated especially by interstitial crystallization, apparently late magmatic, of chlorite and analcite in small amounts. Although the pegmatitic diorite as a whole should be regarded as representing a residual magma, textural relations of hornblende with plagioclase indicate that the hornblende formed at an intermediate stage in crystallization of the pegmatite. Typically, (110) crystal faces are well developed in the hornblende, both externally and on the inner surfaces of skeletal prisms. The cores of the hornblende are filled by plagioclase, making the skeletal structure readily visible in hand specimen. Plagioclase also occurs in the angular interstices between hornblendes. In the interstitial plagioclase unzoned calcic cores are typically enclosed by a progressively zoned rim. The zoning encompasses a large compositional range, and one grain that had a core of An_{82} reached An_{25} at its margins. The intra-skeletal plagioclase lacks the distinct calcic cores and is zoned like outer parts of the interstitial plagioclase. These zoning relationships suggest that initial crystallization of unzoned calcic plagioclase was followed by intergrowth of skeletal hornblende and intra-skeletal plagioclase simultaneously with enlargement of the earlier plagioclase that occupied spaces between hornblendes. The pegmatitic hornblende, thus interpreted as representing magmatic crystallization, is dominantly yellow-brown but grades to a mottled green-brown near edges of some grains, apparently indicating partial reaction from brown to green hornblende as described for the other mafic rocks. The analyzed amphibole from the pegmatitic diorite (Table 1, no. 3) is a brown hornblende showing only minor reaction to green hornblende; the green hornblende proved slightly less dense and most of it was removed during separation. The following approximate mode of the pegmatite was estimated visually because of the coarse-grained heterogeneous texture.

Hornblende	55–60	Opaque minerals	2
Plagioclase (An_{82-25})	30	Apatite	1
Biotite	2	Sphene	1
Quartz	5	Analcite	0.5
Chlorite	2		

Hornblende in the intrusive unit of porphyritic quartz-bearing diorite is generally medium-grained as are the other groundmass minerals, but also forms sparse phenocrysts as much as 6 mm long. It is everywhere green or green-brown and acicular in habit, and its crystallization history (Fig. 1, reaction 10) appears analogous to that of petrographically similar acicular hornblende in the diorite (Fig. 1, reaction 8).

The trondhjemitic biotite tonalite, which constitutes the largest part of Gibson Peak pluton, is a weakly foliated homogeneous rock that contains 2 to 6 per cent hornblende. The texture of the tonalite is typically hypidiomorphic, and early inception of hornblende crystallization is indicated by its alignment in the foliation, along with plagioclase. The hornblende forms subhedral prisms of uniform green color, in contrast with the mottled appearance of many amphiboles in the diorites and gabbros. Most prisms are 0.5 to 1.0 mm in length, and are distinctly finer-grained than other minerals in the tonalite or hornblende in the more mafic parts of the pluton. Much hornblende in the tonalite has a ragged, shredded appearance that suggests partial resorption during late stages of crystallization; such hornblende is accordingly interpreted as having crystallized at a relatively early stage in consolidation of the tonalite magma (Fig. 1, reaction 11). The analyzed hornblende from the tonalite (Table 1, no. 1) is from a specimen that is slightly higher in hornblende than average (Lipman, 1963, Table 1); it has the following mode:

Plagioclase (An_{45-16})	62	Opaque minerals	1
Quartz	20	Chlorite	0.5
Biotite	10	Apatite	0.4
Hornblende	5	Sphene	0.2
K feldspar	1		

Chemical and Optical Data

The material submitted for analysis was separated by centrifuging disaggregated rock powders in mixtures of bromoform and methylene iodide after preliminary concentration with a Frantz Isodynamic Separator. Each final concentration contained less than 1 per cent impurities, principally epidote and chorite.

Table 2. Calculated Structural Formulae of Amphiboles from Gibson Peak Pluton

		1		2		3	
	Si	7.00		6.52		6.14	
(Z)	Al	.78	7.88	1.33	8.00	1.65	8.00
	Ti	.10		.15		.21	
(Y)	Ti	—	.64	.12	.68	.07	.81
	Fe^{3+}	.64		.56		.74	
	Fe^{2+}	1.35		1.28		1.07	
(X)	Mg	3.12	4.67	3.26	4.60	3.38	4.46
	Mn	.20		.06		.01	
	Ca	1.79		1.82		1.86	
(W)	Na	.26	2.12	.44	2.33	.62	2.53
	K	.07		.07		.05	
	OH	2.01		1.88		1.82	
	F	—	2.02	.03	1.92	.04	1.88
	Cl	.01		.01		.02	
Total Fe		1.99		1.84		1.81	
Fe^{3+}/Fe^{2+}		.47		.44		.69	
Fe/Mg		.64		.57		.54	

The three chemical analyses, made by W. H. Herdsman, are given in Table 1. Structural formulae (Table 2) have been calculated on the basis of 24(O, OH, F, Cl), with (OH, F, Cl) approximately 2.0, following Warren (1930). In each the calculations correspond reasonably well with Warren's ideal formula, but only if all Al and most of the Ti are included in the Z group. The Z group of analysis 1, which is sufficiently low in Al to be intermediate between actinolite and hornblende, cannot be filled even with inclusion of all Ti.

The heterogeneity within individual amphibole specimens is the main limit on precision of optical and other physical measurements. The specific gravity was determined by pycnometer measurement of the heavy liquid in which the bulk of the final sample could be suspended. Each analyzed sample shows ranges in specific gravity of about 0.02 to 0.04. Analysis 1 probably closely represents the average amphibole of the rock, but a small amount of less dense, greenish amphibole was removed during the separation of specimens 2 and 3. All refractive index determinations were made by immersion methods, on part of the samples submitted for analysis. The range of error is difficult to estimate because of the variability of the samples but the indices are probably accurate within 0.002. 2V and $Z \wedge c$ were measured by universal stage from thin sections of the rocks from which the samples were separated; after separation the final sample was too fine-grained for universal-stage measurement. The denoted limit of error is the range of values that occurred repeatedly during measurement of at least 10 grains.

INTERPRETATION OF THE ANALYSES

The analyses of the three hornblendes fall into a compositional series, involving significant variation of Si, Ti, Al, Fe, Mg, Mn, Na, halogens, Fe^{3+}/Fe^{2+}, and Fe/Mg. Although this series correlates with neither the sequence of intrusive units in the Gibson Peak pluton nor their bulk compositions,[1] it may be related to the position of each hornblende in the crystallization sequence of its containing rock. Petrographic evidence indicates that the hornblende of the tonalite crystallized early (Fig. 1), whereas the poikilitic hornblende of the hypersthene gabbro crystallized late from a fractionated residual melt. The pegmatitic hornblende appears to represent the greatest degree of fractionation: it formed relatively late in a rock which itself developed by differentiation from hornblende diorite.

If this interpretation of the mineral paragenesis is valid, the hornblende series is a sensitive indicator of chemical changes during crystallization of the plutonic rocks. The increasing Na, which should reflect differential residual concentration, is particularly interesting because of the ease with which this element can be followed in other mineral phases of the same rocks. Normally zoned plagioclase demonstrates a residual increase in Na in all the rocks, and the extreme fractionation in the hornblende pegmatite yielded a late-magmatic sodic zeolite. Comparison of Na in hornblendes from gabbro and tonalite provides further evidence of fractionation: the late hornblende of the labradorite-bearing gabbro

---

[1] The general trend in bulk compositions should be evident from the modal data already cited; chemical compositions computed from averaged modes of the hypersthene-hornblende gabbro and the trondhjemitic biotite tonalite are given elsewhere (Lipman, 1963, Table 2).

is nearly twice as sodic as earlier hornblende in tonalite containing oligoclase.

Systematic variation of several elements in the hornblende series suggests residual concentration of the volatile constituents during crystallization. Although the halogen contents are low and close to analytical limits, total halogens from gabbro and pegmatite hornblendes are respectively four and six times that of the hornblende in tonalite. The higher Fe^{3+}/Fe^{2+} ratio of the pegmatitic hornblende suggests increased PO_2 which is probably related to increased water content. The actual values for $(OH)^-$ seem less significant, both because of the analytical difficulties in this determination and also because of the possibility that some $(OH)^-$ may go into Z positions (C. O. Hutton, oral comm., 1959). Nevertheless, the $(OH)^-$ is higher for all three hornblendes than in many published analyses. Aside from the hornblende composition, high volatile concentration is indicated during crystallization of the pegmatite by the late magmatic chlorite and zeolite.

The higher Ti in the late-crystallizing hornblendes of the gabbro and pegmatitic diorite than in the early hornblende of the tonalite indicates residual concentration of this element. In the pegmatitic diorite residual concentration of Ti is indicated independently by the presence of late interstitial sphene. The changes in total Fe and Fe/Mg ratios, though systematic, are more difficult to interpret because of the effects of coexisting minerals such as hypersthene, biotite, and opaques. In view of the wide range in bulk composition of the three rocks, the limited range of the Fe variation is striking.

The variation of Si and Al appears to reflect several factors. Increased substitution of Al for Si with rising temperature, as occurs in many hornblende assemblages (Harry, 1950; Compton, 1958), was not the dominant control for the present series. Although hornblende may have crystallized from the gabbro at slightly higher temperatures than from the tonalite, the pegmatitic hornblende which contains the most Al in the Z position crystallized later than the gabbro, and low temperatures are indicated by magmatic chlorite and zeolite. The presence of chlorite and zeolite indicates concentration of Al in the residual melt, and the Al-Si substitution may also have been controlled largely by the composition of the residua from which the hornblendes crystallized.

Mn systematically decreases from the unusually high value for the hornblende of the tonalite, but the basis for this distribution is not clear.

Acknowledgments

This paper represents one part of a general study of the geology of the southeastern Trinity Alps, northern California (P. W. Lipman, 1962, unpub. Ph.D. thesis, Stanford Univ.). I am particularly grateful to

Profs. W. R. Dickinson and C. O. Hutton for advice on this part of the study. A grant from the Shell Research Fund at Stanford covered the cost of the analyses. The manuscript has benefited greatly from constructive criticism by B. F. Leonard, C. O. Hutton, and Theodore Botinelly.

REFERENCES

COMPTON, R. R. (1958) Significance of amphibole paragenesis in the Bidwell Bar region, California. *Am. Mineral.* **43,** 890–907.

HARRY, W. T. (1950) Aluminium replacing silicon in some silicate lattices. *Mineral. Mag.* **29,** 142–150.

FAIRBAIRN, H. W. AND OTHERS (1951) A cooperative investigation of precision and accuracy in chemical, spectrochemical, and modal analyses of silicate rocks. *U. S. Geol. Survey Bull.* **980.**

LIPMAN, P. W. (1963) Gibson Peak pluton: a discordant composite intrusion in the southeastern Trinity Alps, northern California. *Bull. Geol. Soc. Am.* **74,** 1259–1280.

WARREN, B. E. (1930) The crystal structure and chemical composition of the monoclinic amphiboles. *Zeit. Krist.* **72,** 493–517.

Manuscript received, January 15, 1964; accepted for publication, April 4, 1964.

THE AMERICAN MINERALOGIST, VOL. 49, SEPTEMBER–OCTOBER, 1964

THE GARNET ISOGRAD IN PELITIC ROCKS AND ITS RELATION TO METAMORPHIC FACIES

M. P. Atherton, *Department of Geology, University of Liverpool, Liverpool, England.*

Abstract

A study of pelitic rocks from the biotite and lower garnet zone of Barrow in the Dalradian shows that: (a) most rocks carrying garnet have chlorite as the only other co-existing ferromagnesian phase; (b) rocks in the biotite zone rarely carry biotite. Comparison of garnet analyses on the isograd from the Dalradian and other low grade areas shows that the composition does not vary greatly with respect to MnO content. The general similarity of MnO content of the garnet and its association with chlorite suggests that it does not "grow" in the chlorite zone and reversal of the isograd does not take place. The association of chloritoid with chlorite-muscovite rocks often carrying garnet implies that chloritoid is stable only at the top of the green schist facies, *i.e.*, on and within the garnet isograd.

Introduction

The concept of zones of progressive regional metamorphism is familiar to all metamorphic geologists. Work by Barrow (1893) and later Tilley (1925) and others has substantiated the principle of correlation of mineralogical variation with physical parameters, more especially temperature. Degrees of metamorphism are recognized by index minerals characteristic of a particular rock composition. Work by Barrow and Tilley relates specifically to the pelitic rocks found in the Scottish Dalradian, where zones are defined in terms of the first appearance of the index mineral. Other sequences of minerals have been used, and it is by no means certain that the Barrovian suite is typical or even usual.

However, the use of index minerals is still the foundation for work both on facies subdivision and more complex problems such as partition between coexisting phases and its relation to temperature. For example, Turner and Verhoogen (1960) define the sub-facies of the greenschist and amphibolite facies in terms of the classic isograd reactions of Barrow. Mapping of these sub-facies will therefore depend on recognition of assemblages equivalent to those found in the Barrovian sequence or on mapping by means of index minerals. In the former case much depends on the definitions of equivalence used, and difficulty exists as to the precise limits of the assemblages and their correlations, since different mineral assemblages will overlap with respect to their stability fields.

These problems may be sorted out with reference to the changing pattern of chemical composition of index minerals, studied either on their own or more rigidly in terms of host rock composition and coexisting phases. In the latter case it may be possible to define more precisely the equivalence of various assemblages, though compositional variation of the host must be eliminated if the variation in mineral composition is to be related to physical factors. The study of mineral assemblages is related to the "study of the genesis of the critical index minerals of the facies in an isochemical rock series" (Tilley, 1925, p. 35) and it is in this field that more detailed work is required. Briefly further elucidation of facies and metamorphic grade must still be concerned in the main with index minerals and their relationship with other coexisting phases and associated assemblages. Particular interest here is centered on the garnet isograd, its relation to the incoming of biotite, and the sub-facies of the greenschist facies.

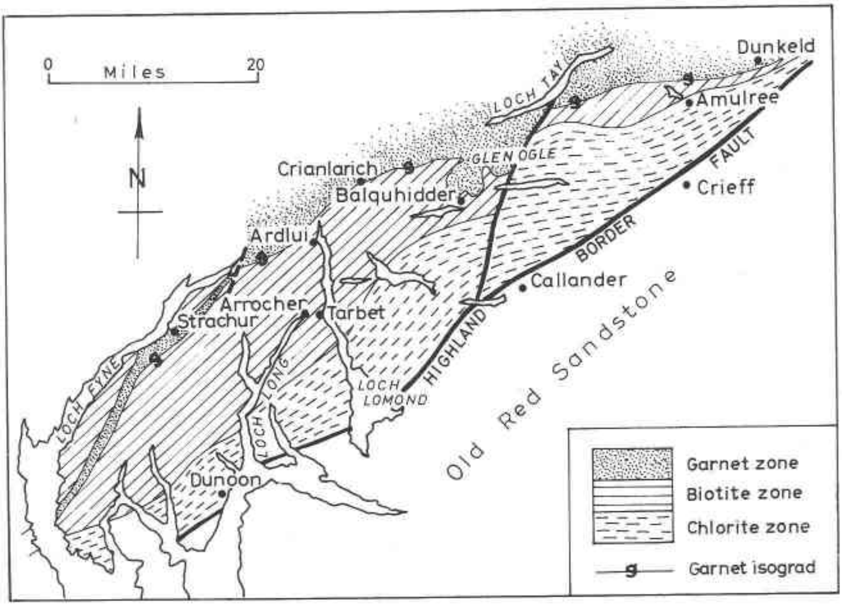


FIG. 1. Zonal map of the southwest Highlands of Scotland, showing the locations of the areas studied.

ROCKS STUDIED

The rocks studied lie in a broad belt, including the garnet isograd, which runs southwest from Dunkeld to Loch Lomond (Fig. 1). Particular attention was paid to the areas around Dunkeld, Balquihidder, Glen Ogle, Loch Lomond and Crianlarich. The rocks are pelites in the sense that they contain various combinations of the following minerals: biotite, chlorite, garnet and muscovite. Hornblende and epidote are absent, and calcite, where present, is of small amount. Quartz is ubiquitous, and with increasing modal amount the rocks grade into semi-pelites. A more rigid definition of the term pelite would obviously involve chemical analysis of the rock and its relation to the mode. As this paper is concerned with the use of index minerals in the field, this subject will be limited to a qualitative discussion, though the ideas expressed here suggest that a more quantitative account of chemical composition in relation to mineral assemblages is necessary.

A thin section study of the rocks from these areas shows that many of the rocks from the "biotite zone" do not contain biotite. Of 67 rocks studied from the biotite zone and on the garnet isograd, 16 contained biotite and of these only 11 contained biotite in excess of 1%. The dominant ferromagnesian mineral is usually chlorite, which even on the garnet isograd almost invariably exceeds biotite. Of the 41 specimens on or to the high grade side of the garnet isograd mostly containing garnet, 30 carry chlorite alone, while the remaining 11 contain chlorite plus some biotite, usually of small amount (in only two rocks does biotite exceed 5% by volume). The results are summarized:

	No. of rocks examined	Biotite present in	
Biotite zone	26	5	(2 less than 1%, usually less than 2%)
Biotite/garnet	8	3	(2 less than 1%)
Garnet zone (lower)	33	8	(2 greater than 5%)

It may be concluded that garnet on and immediately to the high grade side of the isograd in rocks loosely defined as pelitic, occurs more commonly in association with chlorite than biotite. It should be emphasized that the choice of samples in the field was somewhat arbitrary in that specimens which were definitely considered pelitic were chosen on the assumption that index minerals were present. The presence of biotite in quartz-felspar rocks at this grade suggest the index mineral host may have to be more closely defined.

The presence of primary chlorite well to the high grade side of the garnet isograd is of particular interst. Although both Harker and Tilley discuss the genesis of garnet from a rock commonly carrying chlorite, there seems to be some need to re-emphasize this as general ideas on compatible assemblages appear to ignore it. Turner and Verhoogen (1958) for example, do not include chlorite in any of their pelitic assemblages for the quartz-albite-epidote-almandine sub-facies; "Only magnesian chlorites persist in this sub-facies and their occurrence is mainly in magnesian schists" (p. 224).

The relationships of such rocks is well seen in a sample from the Ben

Ledi Grit, just to the high grade side of the garnet isograd near Craig Head, north of Blairgowrie. In this region garnet is very common in phyllitic and quartzose rocks as well as in associated green beds. In a single specimen A 5, two rocks may be seen sharply divided. Although neither rock contains muscovite, they are essentially pelitic in character and do not contain calcium rich minerals such as hornblende or calcite. Yet the assemblages are very different. A5(i) contains quartz, garnet and chlorite whereas A5(ii) carries quartz, chlorite and biotite. The modes and analyses are given in Table 1, where the different composition and mineral assemblages are clearly seen. From the analyses both rocks are seen to be rich in Na_2O and poor in K_2O, and in this sense are related to the albite schists of the same horizon in the southwest. From the mode it is apparent that some of the modal quartz must be albite (in such rocks the albite is rarely twinned and can often be recognized readily only by its commonly carrying many small inclusions). This is confirmed when the SiO_2 content of $A5_{ii}$ is compared to the modal quartz, when an

TABLE 1. CHEMICAL ANALYSES AND MODES OF PELITES FROM THE GARNET ISOGRAD, BLAIRGOWRIE, SCOTLAND

	A5(i)	A5(ii)
SiO_2	41.12	66.02
TiO_2	2.43	0.84
Al_2O_3	22.56	14.11
Fe_2O_3	1.85	1.16
FeO	12.68	5.92
MnO	0.49	0.11
MgO	7.13	4.12
CaO	1.54	1.09
Na_2O	3.49	4.02
K_2O	1.31	1.31
H_2O^+	4.97	1.89
P_2O_5	0.03	0.04
Total	99.60	100.63
Quartz	30.2	60.0
Chlorite	60.5	15.3
Biotite	—	19.1
Garnet	2.3	—
Feldspar	0.6	4.2
Ore	4.2	0.9
Accessories	2.3	0.5

excess of free quartz over the SiO_2 content is observed. The plots of these two rocks on AKF and AKFM diagrams (Fig. 2) is interesting. Both rocks plot near the AMF plane as expected as muscovite is absent. However, the rocks have similar F and M values and differ markedly only in the A content. A5$_{i}$ plots almost on the muscovite-garnet join, while A5$_{ii}$ plots well inside the chlorite-biotite field.

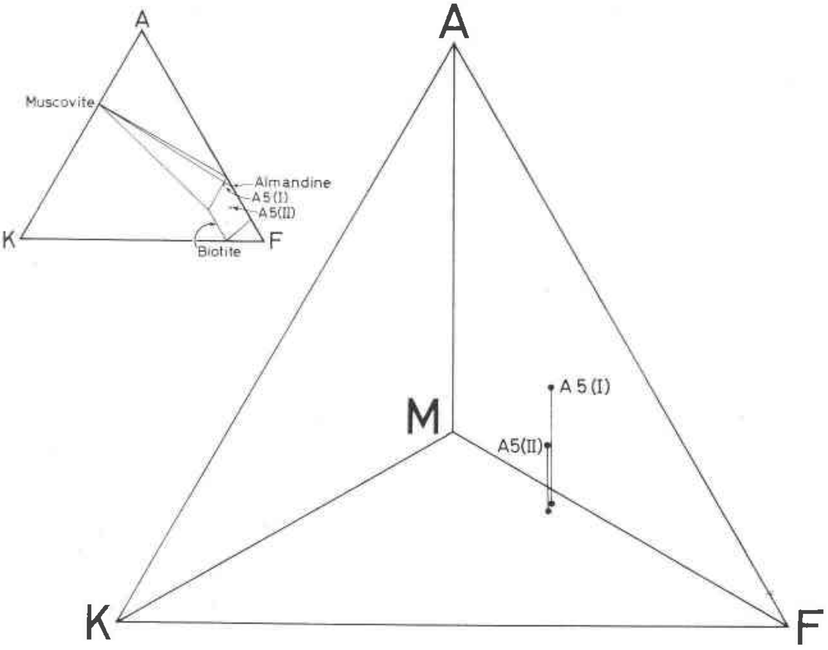


Fig. 2. AKF and AKFM plots of analysed rocks A5I and A5II, from Blairgowrie, Scotland.

Here, therefore, the assemblage chlorite-garnet is metamorphically equivalent to chlorite-biotite. Noticeably the rock containing biotite also contains some chlorite, and is the more siliceous of the two rocks. In general the rocks carrying garnet along the isograd are essentially chlorite-quartz ± muscovite schists and are without biotite. Rocks with biotite and garnet do occur at this grade but they are of much more limited amount.

The growth of garnet on the isograd is of great importance, particularly in understanding the inter-relationships of minerals in the Barrovian sequence. Harker (1939, p. 192) presents fine pictures of garnet "growing" from the existing chlorite, some of which it appears to pseudomorph.

The same relations were observed by Tilley (1926) who suggested growth "from laths and scales of chlorite interspersed among quartz grains" (*op. cit.*, p. 40). Extensive examination of garnet on the isograd indicates that garnet growth is often related to the quartz boundaries, and commonly there is little evidence of chlorite near the "newly" growing garnet; indeed garnet is often seen in quartz stringers with the production of curious apophyses and skeletal shapes (Fig. 3). Elsewhere, particularly in albite schists, beautiful small euhedral garnets are present inside large albites. In phyllitic rocks sub- or euhedral garnet grows in a chlorite groundmass without evidence of pushing or pseudomorphing. Although there is often no contact relationship between chlorite and garnet, there is a reaction relationship between the two minerals, which is not simple as chlorite contains little CaO or MnO. It is of interest to note that hornblende too, when present in very small amount in these rocks, is isolated in the quartzose parts of the rock, and it is considered that these sites, *i.e.*, quartz-quartz boundaries, towards the close of the deformation probably form high energy positions suitable for the nucleation of a new phase.

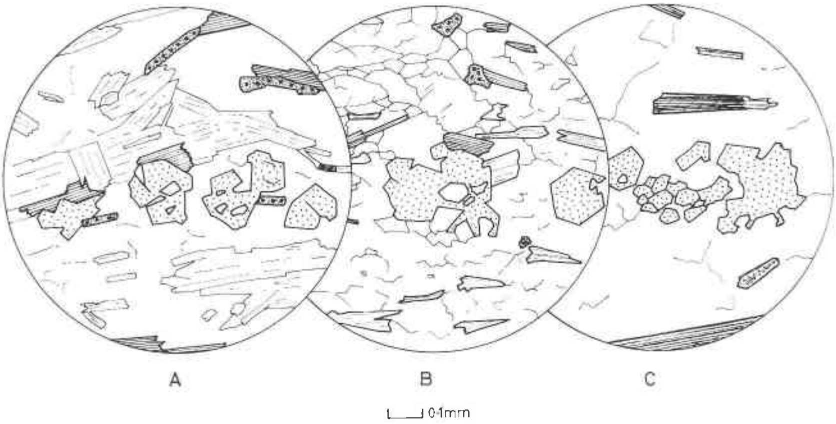


FIG. 3. Garnetiferous schists: A. Muscovite chlorite garnet quartz schist with small subhedral garnet. B. Chlorite-quartz garnet schist with small garnets showiug growth along quartz grain boundaries. C. Chlorite quartz garnet schist with small clustered garnets in a quartz matrix.

The composition of garnets from the isograd are given in Table 2, together with analyses of garnets from the Moines (Lambert, 1959), Start (Tilley, 1923) and Devon (Phillips, 1928). In general the analyses reveal a common trend. They are all almandines, with relatively poor

TABLE 2. GARNET ANALYSES FROM ON THE ISOGRAD

	1	2	3	4	5	6	7	8	9	10	11	12	13
SiO_2	35.93	36.62				38.00	37.23			38.0	38.44	38.25	36.05
Al_2O_3	20.13	19.66				20.80	21.31			21.0	20.62	19.57	20.82
TiO_2	0.11	0.30				0.79	0.29			—	—	0.52	1.37
Fe_2O_3	3.85	3.83	29.1[1]	28.2[1]	24.2[1]	1.35	2.79			2.5	7.39	3.02	1.43
FeO	25.70	25.68				20.83	27.04	(over 20.0)		23.7	17.09	21.12	22.73
MnO	3.84	4.04	5.8	5.7	5.3	5.01	1.51	5.8	5.7	3.3	4.22	5.48	4.76
MgO	2.00	1.84				1.08	1.39			4.5	2.51	2.31	0.77
CaO	8.47	8.17				12.20	8.37			7.5	10.08	7.56	11.28
H_2O^+						0.11	0.10					2.48	1.16
H_2O^-												0.07	
P_2O_5												0.15	
	100.03	100.14				100.36	100.29			100.5	100.35	100.53	100.37

[1] Total iron as FeO.

1. Garnet on the isograd, Perthshire, Dalradian. (Atherton, in press).
2. Garnet on the isograd, Perthshire, Dalradian. (Atherton, in press).
3., 4., 5. Micro analyses of garnet on the isograd, Perthshire, Dalradian. (Atherton in press).
6. Garnet from "biotite zone." (Lambert, 1959). Total includes 0.19% alkalis.
7. Garnet from garnet zone. (Lambert, 1959). Total includes 0.28% alkalis.
8. Garnet from Start. (Tilley, 1923).
9. Garnet from North Devon. (Phillips, 1928).
10. Garnet from chlorite schist, County Mayo, Eire.
11. Garnet from isograd amphibolite. (Wiseman, 1934).
12. Garnet from "chlorite zone" rock containing glaucophane. Japan. (Miyashiro and Banno, 1958).
13. Garnet from glaucophane containing rock. (Banno, 1959).

MnO and MgO contents (2–6%). The CaO content is of interest particularly in the Dalradian garnets, which confirms Tilley's suggestion that low grade garnet may be stabilized by a relatively high CaO content. The Moine garnets, too, are rather similar though the lowest grade garnet has a slightly higher CaO content than those from the Dalradian which must be close to the maximum amount found in pyralspites. In general these garnets are different from the garnets found at higher grade in the Dalradian which often have higher FeO and MgO and lower MnO and CaO values (Sturt, 1963; Atherton, in press). Similar variations have been observed elsewhere (Engel and Engel, 1960), though care must be taken to ensure these changes are not direct functions of host rock composition.

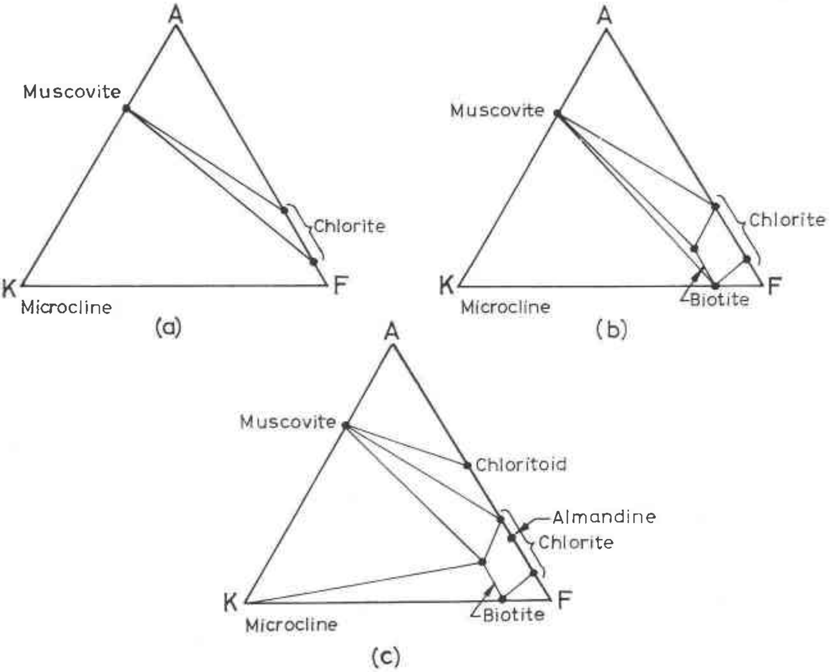


FIG. 4. AKF plots of the subfacies of the greenschist facies.

DISCUSSION

The problem of the garnet isograd and more especially its relation to the biotite isograd has caused much discussion. In the Dalradian biotite appears at lower grade than garnet, although commonly garnetiferous rocks on the isograd are devoid of biotite. Elsewhere the absence of biotite in rocks containing chlorite and garnet has been considered due to the stabilization of garnet at lower grade than in the Dalradian. Thus in Start (Tilley, 1923) and North Devon (Phillips, 1928) garnet is present with chorite and without biotite and here a reversal of the normal sequence has been suggested.

Primary vs. secondary chlorite. It may be argued that the chlorite associated with garnet seen in the Dalradian on and above the garnet isograd is secondary, possibily after biotite. This would account for the lack of the latter mineral and the dominance of chlorite. Indisputable criteria for recognizing secondary chlorite are rarely present; however the following evidence suggests the chlorite is primary.

(1) In rocks such as A 5 chlorite-garnet assemblages are associated with biotite-chlorite assemblages, and in these cases biotite shows no alteration and is not involved texturally with chlorite.

(2) The chlorite is clear, without rutile needles or any other evidence of biotite breakdown.

(3) Almost invariably the garnets associated with chlorite show no alteration to chlorite; they have clear peripheries and are often sub- or euhedral. Had biotite retrogressed one would expect garnet to show some alteration too.

(4) Primary chlorite is different in colour and form to secondary chlorite found in rocks of slightly higher grade. Usually the secondary chlorite cuts across the fabric indiscriminately or forms peripheral margins to garnet or biotite. Primary chlorites have anomalous blue polarization colours and are optically negative while secondary chlorite has brown green polarization colours and is optically positive. These criteria are general in the Dalradian, and agree with those suggested by Goldschmidt (1921) for the Stavanger rocks, where "original chlorite is always feebly negative in optical character, while derivative chlorite is as a rule either strongly negative or feebly positive" (Harker, 1939, p. 222).

Reaction involving the production of garnet. As Chinner (1960 p. 213) has said "knowledge of the reactions which give rise to an almandine rich garnet at the garnet isograd has not materially increased since Tilley's (1926) discussion of the problem." In general the following reaction has been envisaged for the isograd:

$$\text{chlorite}_a + \text{quartz} \rightarrow \text{garnet} + \text{chlorite}_b + H_2O$$

The production of chlorite$_b$ which would be more magnesian rich than chlorite$_a$ and its appearance well to the high grade side of the isograd fits in with the evidence presented earlier. Alternatively in rocks which contain biotite, the reaction suggested by Chinner (1960, p. 213) may indicate the stability relations across the isograd:

$$\text{chlorite} + \text{biotite}_1 + \text{quartz} \rightarrow \text{garnet} + \text{biotite}_2 + H_2O$$

although there is no evidence in the Dalradian that chlorite is generally completely consumed at the garnet isograd as Chinner maintains (1960, p. 213). Rocks of such a composition probably contained only chlorite at very low grade and conversion to biotite took place long "before" recrystallization of garnet. However, the growth of new phases such as garnet, appears to involve reactions which are not simple breakdown of pre-existing mineral phases, though obviously these are involved indirectly in the formation of a new phase. For example, garnet on the isograd contains considerable amounts of CaO and MnO not found in chlorite in quantity. Obviously all the components in the rock are involved in the production of a new phase, and it is felt that reactions such as those considered above should be regarded as equations representing the stability relationships rather than genesis. Under the microscope garnet appears to grow in a quartz milieu and its formation could equally well be expressed, probably more realistically, as follows:

$$\begin{tikzcd} & \text{chlorite}_a+\text{quartz etc.} \\ \text{unaltered rock} \arrow[ur] \arrow[dr] & \text{———isograd———} \\ & \text{chlorite}_b+\text{garnet}+\text{quartz} \end{tikzcd}$$

This implies a lack of progressive metamorphism which is in keeping with the textural and microscopic evidence. Here we should be more concerned with looking at the chemical characteristics of the host rock rather than the composition of equivalent lower grade phases.

MnO content of garnet in the Dalradian, Start and Devon. The garnets from Start and North Devon described by Tilley (1923) and Phillips (1928) have frequently been used as examples of garnet occurring in the chlorite zone, thus suggesting the garnet isogard has shifted to a lower grade than its position in the Barrovian sequence. In Start (Tilley, 1923) garnet dodecahedra up to $\frac{1}{16}''$ in diameter occur in quartz-muscovite-chlorite-albite schists, with incipient biotite. Compositionally the garnet contains 5.8% MnO (12.8% spessartite) and FeO in excess of 20%. The garnet from North Devon occurs in mineralogically rather similar rocks. These contain porphyroblastic chlorite and/or white mica with accessory ore; locally chloritoid is present in large amounts, *e.g.*, Woolgarden Phyllite. Phillips' analysis for MnO gives a value corresponding to 12.5% spessartite, which is the same as Tilley's garnet with respect to MnO content. Garnets analyzed from chlorite-muscovite-albite schists in County Mayo, Eire, by the author, have MnO contents of 3–5%; one such analysis is given in Table 2.

The three cases described here are similar both with regard to the mineral association and the MnO content of the garnet, to the garnets analyzed from the isograd in the Dalradian, and there is therefore no reason to suspect that the growth of any one of these garnets is at lower P-T conditions than the rest. If this is correct, the absence of biotite implies no reversal of the isograd, certainly not if one uses MnO content of garnet and host assemblage as a criterion.

The sporadic appearance of biotite is considered to be due to host rock composition. Certainly most low grade rocks in the Dalradian carry little biotite as shown earlier, and a clue to biotite stability may be seen in rocks such as A 5 (Fig. 2) where the two assemblages occur in the same hand specimen. Notably the biotite is present in the more quartzose rocks and is absent in the more aluminous and quartz poor rocks. As chlorite disappears at higher grade, the presence of biotite must be related to host rock composition and chlorite stability. Chlorite is often the only ferromagnesian silicate present in many low grade pelitic rocks, and as its composition varies considerably, the production of biotite is only permitted if the rock composition does not allow chlorite as the only phase. Low grade chlorites are aluminum rich, and perusal of the AKFM diagram for A5 indicates this is the most important difference between the rocks with and without biotite, apart from silica content. Thus biotite in the lower zones will only form if the rock is short of sufficient Al to allow the production of an aluminous chlorite. As most of the schistose rocks used for zonal identification have sufficient Al_2O_3, biotite is commonly absent till the grade is high enough for garnet and biotite to be present in quantity, *i.e.*, amphibolite facies. Perhaps here too the inter-relations of chlorite-biotite-garnet may be readily appreciated as the growth of all three is strongly interconnected through their relative stability fields and host rock composition.

Garnet with a high MnO content. Apart from the examples discussed above, *i.e.*, Start and North Devon, there have been others quoted as evidence of reversal of isograds, particularly Goldschmidt's (1920) Stavanger garnet These garnets are different from those cited above in having higher MnO contents. The argument normally used is that as there is no biotite present, the garnet is of lower grade than normal and this is due to the high MnO content stabilizing it at this grade. The Stavanger rocks grade from quartz-muscovite-chlorite-garnet phyllites to quartz-muscovite-chlorite-biotite-garnet phyllites and both chemically and modally are comparable to the rocks described above. Thus they are aluminum rich phyllites which would be expected to carry chlorite on the garnet isograd as do similar rocks from Start and the Dalradian. With regard to MnO content the evidence is not clear; MnO rich garnets are by no means confined to low grade rocks. Clifford (1960), for example, presents a garnet with 56% spessartite from the sillimanite zone coexisting with biotite. Here, as in most examples of MnO rich garnets from metamorphic rocks, *e.g.*, Hutton, 1940, host rock composition, particularly MnO content, has been much the most important single factor. It is held that in view of the limited data, although MnO does occur in some low grade garnets, it does not necessarily imply that a high MnO content is correlatable with low grade. As Harker has shown (1939, p. 54) manganese rich garnet occurs "after" biotite in many thermal aureoles.

Indirect evidence from felspar composition substantiates the conclusions reached above. Thus six felspars from the garnet isograd in the Dalradian range from An_2—An_6, while Goldschmidt's quartz-muscovite-chlorite-garnet phyllite has felspar of a composition $An_{2.5}$. Although this type of evidence is not conclusive, it does suggest, together with the general equivalence of mineral assemblage, that the rocks recrystallized at "normal" garnet grade.

Apart from the analyses discussed earlier, Table 2 shows analyses of a few selected garnets from rocks of different composition than the pelites, though the grade is similar. Analysis 11 is from a garnet-biotite-clinozoisite-albite amphibolite (Achahaish, Scotland, Wiseman, 1934) which is surprisingly similar in composition to the garnet from neighbouring pelitic rocks.[1] The rock is on the garnet isograd, though the host is of radically different composition to the surrounding rocks. Number 12 is from the Omi schist (Miyashiro and Banno, 1958) which may be divided into chlorite and biotite zones, the former containing rocks with glaucophane, chlorite and rare garnet (the analyzed garnet is from a rock of this type). whereas the biotite zone rocks contain biotite and almandine but lack glaucophane. Chlorite is also present at this grade, but it disappears at higher grade. The most common assemblage is almandine-chlorite-muscovite-biotite-albite-quartz. Miyashiro and Banno (1958) conclude that the biotite and garnet isograds are coincident; indeed their higher grade assemblage is similar to that seen along the isograd in the Dalradian, while the garnet is very close in composition to isograd garnet. Garnet 13, from Kotu, Japan (Banno, 1959) is also from a glaucophane rock and is rather similar in composition to garnet 12, especially with regard to the CaO and FeO values.

The grade relations of the garnets mentioned above are not very clear, with the exception of Wiseman's amphibolite which lies on the garnet isograd as defined by Tilley. However, there is a strong similarity even in garnets from such widely different environments as these, to the garnets discussed above from the Dalradian.

---

[1] It may be noted that the Fe_2O_3 in this analysis is high and probably present in the garnet as FeO.

THE PRODUCTION OF ISOGRAD MINERALS

The presence or absence of a particular mineral or assemblage is dependent on three sets of variables:

(a) The physical parameters specifying the assemblage, *i.e.*, T, P, P_{H_2O} etc.
(b) The kinetics of the reactions producing the assemblage.
(c) The composition of the host rock.

Generally variations due to (b) (particularly with reference to index minerals) and (c) in the type of rock suite studied, are not considered to be important factors in the production of a mineral assemblage. Thus all rocks generally called "pelitic" are considered to show successively chlorite, biotite and garnet. However, the variations within this group indicate that the chemical variation is sufficient to produce different assemblages at the same grade, *e.g.*, chlorite-biotite, garnet-chlorite. The absence of a biotite phase in this case is entirely a function of host rock composition at this grade. Generally the lack of biotite in many Dalradian "pelites" of greenschist facies is due to the fact that the rocks plot above the chlorite-muscovite join of the AKF diagram, *i.e.*, they are aluminum rich sediments. Such rocks at higher grade produce the kyanite, staurolite rocks common in the Dalradian. The crystallization of biotite in such rocks will be inhibited until the amphibolite facies. In the sense that we are using two different host compositions to define one mineralogical sequence, we could consider isograd inhibition, and this is presumably what Harker (1939, p. 218) meant when he said "perhaps we should see here rather a postponement of biotite than a promotion of garnet."

At this point it is pertinent to discuss isograds in a more general way. Although some isograds involve simply polymorphic change, most are dehydration reactions of the type discussed by Thompson (1955) and Francis (1956). Such reactions when related to a particular composition may be considered univariant if the mobile phase, *i.e.*, H_2O in this case, is dependent on the pressure and temperature distribution in the crust. If the mobile phase is independent of pressure and temperature the curve or boundary becomes bivariant. General opinion indicates that the chemical potential of H_2O has some relation to the temperature and pressure in the earth's crust, in which case we can regard the curves as univariant, when referred to a given composition (Thompson, 1955). Considering the production of garnet as discussed here, which may be a dehydration reaction, there is little evidence of variation in the position of the dehydration curve from area to area. Generally indirect evidence of mineral assemblage and felspar content suggests any variation in the absolute sense is not discernible, at least by present methods. However, this argument refers principally to the incoming of garnet in the Dalradian and Start etc.

where the composition is the same, suggesting that the reaction took place under the same P, T conditions and in rocks of the same composition. It is perhaps interesting to refer to Wiseman's (1934) garnet from a basic rock, which is of similar composition to the garnet from pelites. In both basic and pelitic rocks garnet "comes in" with the same composition. The similar composition of garnets at this grade, their definite entry in the field and their appearance in most cases with a simliar assemblage point to the univariant character of the boundary and its similar position in different terrains.

The sensitivity of such reactions as opposed to the constancy of their field appearance strengthens the view that P_{H_2O} is not an independent variable and its effect on the relative position of dehydration reactions is small (see Francis, 1956). As isograd reactions are commonly dehydration reactions, they are most sensitive to T and P_{H_2O}. If one assumes (Turner and Verhoogen, 1960, p. 506) that P and T increase together in the earth's crust and $P_{H_2O} = P_f$ and $P_l = P_f$ approximately, then these curves will have steep positive slopes and will be equivalent to an isothermal. As the relative importance of P and the relation of P_f to P_l is as yet little understood, the isograds are here treated as isothermals.

This being so, the displacement of an isograd is a function of the bulk chemistry of the rock, *i.e.*, variations in MgO, FeO, Al_2O_3, MnO, etc. The incoming of biotite is certainly a function of bulk chemistry, in the sense that in rocks with high aluminum content biotite does not occur till a much higher grade than in less aluminum rich rocks. As most of the rocks studied in the Dalradian are rich in Al_2O_3, rocks with biotite are not common in the greenschist facies. Barrow's zonal sequence is developed on aluminum rich pelitic rocks, and in these the biotite and garnet isograd are almost coincident. A more rigid definition of pelites in terms of chemical composition would eliminate confusion due to host rock variation. Garnet appearance is not the same; it comes in abruptly in rocks rich in Al_2O_3, and also in more quartz rich rocks. The MnO, FeO, MgO values have little effect on the appearance of garnet in A5, where the F and M (including MnO) values are essentially similar. Again the most important factor appears to be the ratio of Al_2O_3 to FeO+MgO, although the presence of garnet is much less limited than biotite in the greenschist facies of the Dalradian.

The composition of garnet is the same in Eire, Scotland and S.W. England, where MnO content does not appear to have to be necessarily over 5% or so to stabilize garnet at low grade. In conclusion it may be said that no reversal of isograds in the same rock type is required to explain the data observed, which is merely a reflection of initial host rock composition and the stability of chlorite, biotite and garnet.

GARNET, BIOTITE STABILITY AND THE GREENSCHIST FACIES

This interpretation of low grade isograd reactions has some bearing on the subdivision of the greenschist facies which is based on the two isograd reactions discussed here (Turner and Verhoogen, 1960). It is not surprising that facies boundaries are defined in the main by equations involving the production of pelitic isograd minerals as this host rock is particularly sensitive to external physical conditions. The difficulty in facies correlation lies in defining the boundaries across a varying composition field. Turner and Verhoogen use reactions (1) and (2) below to subdivide the greenschist facies into the following three sub-facies:

(a) quartz—albite—muscovite—chlorite.
(b) quartz—albite—epidote—biotite.
(c) quartz—albite—epidote—almandine.

The reactions defining the boundaries between (a) & (b) and (b) & (c) are:

(1) $\text{muscovite}+\text{chlorite}_1 \rightarrow \text{biotite}+\text{chlorite}_2$
(2) $\text{chlorite}_a+\text{quartz} \rightarrow \text{garnet}+\text{chlorite}_b+H_2O$

respectively. In both cases the chlorite on the high grade side of the equation will be more MgO rich than the lower grade chlorite. The products of these reactions have been seen to be facies equivalents and although reaction (1) may occur at lower grade it does so only within a limited composition field. This reaction is common in rocks with an $Al_2O_3/MgO+FeO$ ratio lower than in rocks not carrying biotite (Ambrose, 1936). The general appearance of biotite in aluminum rich rocks is rather similar to that of garnet. Miyashiro and Banno (1958) reached a similar conclusion in their work on the rocks of the Omi district of Japan, where they thought the biotite and garnet isograds were coincident.

In the Dalradian (Harker, 1939) the general sequence of pelitic assemblages is as follows:

- 1
  - (a) quartz—sericite—chlorite (—albite)
  - (b) quartz—muscovite—chlorite (—albite)
  - (c) quartz—muscovite or sericite (—albite—chlorite)
- 2
  - (a) quartz—muscovite—chlorite (—albite)
  - (b) quartz—muscovite—chlorite—biotite
  - (c) quartz—muscovite—chlorite—green biotite
  - (d) quartz—biotite—albite (—muscovite)
- 3
  - (a) quartz—muscovite—chlorite—garnet (—albite)
  - (b) quartz—muscovite—biotite—garnet (—albite)
  - (c) quartz—biotite—chlorite—garnet (—albite)
  - (d) quartz—biotite—chlorite (—albite)
  - (e) chlorite—muscovite—chloritoid (—albite—quartz)

Numbers 1, 2 and 3 refer to the sub-facies of the greenschist facies roughly equivalent to Turner and Verhoogen's sub-facies. In particular 2a is the common pelitic assemblage in the Dalradian. 2b and 2c are of much less frequent occurrence, and 2d seems confined to more quartzo-felspathic rocks.

Apart from the status of biotite in the subdivision of the greenschist facies, the place of garnet and chloritoid is of interest. The garnet referred to above is almandine garnet of a composition as indicated in Table 2 and there is little doubt as to its relative position. Spessartite garnet is placed by Turner and Verhoogen (1960) at much lower grade than almandine garnet, *e.g.*, in the quartz-albite-muscovite-chlorite sub-facies. This is based on isograd work in New Zealand where the garnets are from manganiferous schists (Hutton, 1940). The occurrence of spessartite is not confined to rocks of low grade (see earlier, Clifford (1960)). Hutton (1962) cites an example with two different types of garnet from the same locality, one as small euhedral garnets in a quartz-glaucophane-stilpnomelane schist, the other large garnet porphyroblasts from a glaucophane-pumpellyite-epidote schist. These two analyses are:

	1	2
SiO_2	35.95	38.05
Al_2O_3	20.79	21.73
Fe_2O_3	.46	.31
TiO_2	.15	trace
MgO	.31	4.58
FeO	5.86	21.08
MnO	33.72	4.26
CaO	2.56	9.57
H_2O^+	0.09	0.07
H_2O^-	0.17	0.15
	100.06	99.80

1. Small euhedral garnet. San Juan Batista Mine, San Jose.
2. Large porphyroblastic garnet. Locality as for garnet 1.

It is not easy to generalize from Hutton's data as these rocks are of basic character and hence difficult to correlate with the pelitic paragenesis. Hutton (1962) suggests that the two different types may have formed "under diverse conditions of temperature and pressure," though both come from the same locality. Hutton, however, also states that "determination of the physical properties of coarse and fine garnets from the *same* specimens, in several instances, also indicates the coexistence of garnets of dissimilar composition in rocks of the glaucophane schists facies" (1962, p. 131). General geological data does not favor this interpretation, especially with regard to mineral composition and coexisting phases, which show remarkably uniform characters over large areas, *e.g.*, Kretz (1959) and Mueller (1960). It seems likely that host rock control is much more important (see also Chinner, 1960). In low grade rocks an "early" garnet of the composition discussed in this paper, is here considered equivalent to a spessartite. The garnets quoted by Hutton (1962) are good examples of this. Garnet 2 is exactly comparable to the "early" garnets from the Dalradian, while garnet 1 is typical of garnet from metamorphosed manganiferous sediments (Clifford, 1960).

The appearance of chloritoid is more difficult to understand than that of garnet and its relation to the Barrovian isograd sequence somewhat obscure. Harker (1939) and Turner and Verhoogen (1960) place it in the chlorite zone and quartz-muscovite-chlorite sub-facies.

In the Stonehaven section, it occurs first in Perthumie Bay, occasionally accompanied by small garnets. "By the time chloritoid reaches its best development the biotite has completely given place to garnet, the schists now being definitely in the garnet zone" (Williamson, 1953, p. 354). In the chlorite phyllites of the upper Devonian of North Cornwall, chloritoid is developed and also garnet which contains 12.5% spessartite. This is exactly the same as the garnets from Start (Tilley, 1923) and the Dalradian analyzed here. Harker (1939, p. 213) places this chloritoid schist in the chlorite zone, though garnet is present in the nearby Delabole slate as well as in the associated chlorite phyllites (Phillips, 1928, p. 550). In County Mayo, chloritoid and garnet coexist and in Duchess County chloritoid and garnet porphyroblasts occur in a ground mass of quartz, sericite and albite (Barth, 1936). It would seem that chloritoid is stable at the top of the greenschist facies at approximately the same grade as garnet.

The stability relations of the greenschist minerals from pelitic rocks differ in detail from those presented by Turner and Verhoogen (1960), particularly with respect to chloritoid, biotite and garnet-chlorite stability fields. The limitations of AKF diagrams with respect to mineral equivalents such as garnet-chlorite, are obvious and suggest such diagrams should serve as guides only to mineral paragenesis. The limited knowledge of the compositional variation of chlorite and biotite in the greenschist facies makes the positions of the tie lines somewhat tentative, while it is impossible to put garnet on a three component diagram, as low grade garnet contains considerable CaO and MnO, neither of which is represented on the diagrams. The sub-facies division of Turner and Verhoogen is kept with the minor modifications discussed above (Fig. 3), though commonly in pelitic rocks the diagnostic assemblages carrying biotite are rare, especially in Dalradian type rocks.

CONCLUSIONS

In pelitic rocks from the Dalradian the simple index mineral sequence as suggested by Barrow requires some qualification. Thus in many rocks from the biotite zone the index mineral is absent, although this of course does not invalidate the use of the isograd, which is defined in terms of the advent of biotite. In aluminous rocks biotite "comes in" about the same grade as garnet so that the usual assemblage on the garnet isograd is quartz-chlorite-garnet-muscovite-(biotite). In rocks richer in SiO_2 and poorer in Al_2O_3 the assemblage is quartz-biotite-chlorite-(muscovite). Here then the assemblages chlorite-garnet and biotite-chlorite are facies equivalents. The association of garnet with chlorite therefore in aluminous rocks, does not mean that the garnet is stable at a lower grade than normal. The MnO content of garnet on the isograd from the Dalradian is similar to that of garnets from Start and North Devon, both of which occur in chlorite rich schists of supposedly chlorite grade. It is held that there is no reversal of isograds in these areas, if the MnO content of garnet and coexisting phases and assemblages are considered.

Frequently the presence of MnO as a stabilizing influence in the garnet structure has been used to explain the presence of garnet in a chlorite zone assemblage; however the presence of MnO at varying grades and even showing considerable variation at the same grade suggests host rock control is of far greater importance. Essentially the problem of isograd sequence is in defining a strictly isochemical rock series, for rock composition controls for example, not only biotite composition, but also its presence or absence at a particular grade, within rocks usually termed pelites.

ACKNOWLEDGMENTS

This work forms part of a larger program of study on the chemical mineralogy of metamorphic rocks, carried out at the University of Liverpool with the aid of a grant from the Department of Scientific and Industrial Research.

REFERENCES

AMBROSE, J. W. (1936) Progressive kinetic metamorphism of the Missi Skerries, near Flinflow, Manitoba. *Am. Jour. Sci.*, 5th Ser., **32,** 257–286.

BANNO, S. (1959) Notes on rock forming minerals. (10) Glaucophane and garnet from Kotu district, Sikoku. *Jour. Geol. Soc. Japan,* **65,** 658–663.

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

BARTH, T. F. W. (1936) Structural and petrologic studies in Duchess County, New York. Part II. Petrology and metamorphism of the Palaeozoic rocks. *Bull. Geol. Soc. Am.* **47,** 775–850.

CHINNER, G. A. (1960) Pelitic gneisses with varying ferrous/ferric ratios from Glen Clova, Angus, Scotland. *Jour. Petrology* **1,** 178–217.

CLIFFORD, T. N. (1960) Spessartite and magnesium biotite in coticule bearing rocks from Mill Hollow, Alstead Township, New Hampshire, U.S.A. *Neues Jahrb. Mineral. Abh.* **94,** 1369–1400.

ENGEL, J. AND C. S. ENGEL (1960) Progressive metamorphism and granitization of the major paragneiss, north-west Adirondack Mountains, New York. Part II. Mineralogy. *Bull. Geol. Soc. Am.* **71,** 1–58.

FRANCIS, G. H. (1956) Facies boundaries in pelites at the middle grades of regional metamorphism. *Geol. Mag. London* **93,** 353–368.

GOLDSCHMIDT, V. M. (1921) Die Injektionsmetamorphose im Stavanger-Gebiete. *Skr. Nordke Vidensk Akad., Mat. nat. Kl,* **10.**

HARKER, A. 1939. *Metamorphism.* 2nd ed. Methuen & Co., London.

HUTTON, C. O. (1940) Metamorphism in the Lake Wakatipu region, Western Otago, New Zealand. *New Zealand Dept. Sci. Ind. Research, Geol. Mem.* **5.**

——— (1962) Composition of some garnets in low grade schists. *Trans. Roy. Soc. New Zealand* **1,** 129–133.

KRETZ, R. (1959) Chemical study of garnet, biotite and hornblende from gneisses of Southwestern Quebec, with emphasis on distribution of elements in coexisting minerals. *Jour. Geol.* **67,** 371–402.

LAMBERT, R. ST. J. (1959) The mineralogy and metamorphism of the Moine schists of the Morar and Knoydart districts of Inverness-shire. *Trans. Roy. Soc. Edinburgh* **63,** 553–588.

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

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

——— AND S. BANNO (1958) Nature of glaucophanatic metamorphism. *Am. Jour. Sci.* **256,** 97–110.

MUELLER, R. F. (1960) Compositional characteristics and equilibrium relations in mineral assemblages of a metamorphosed iron formation. *Am. Jour. Sci.* **258,** 449–497.

PHILLIPS, F. C. (1928) Metamorphism in the upper Devonian of North Cornwall. *Geol. Mag. London* 550–556.

RAMBERG, H. (1952) *The Origin of Metamorphic and Metasomatic Rocks.* The University of Chicago Press.

STURT, B. A. (1962) The composition of garnets from pelitic schists in relation to the grade of regional metamorphism. *Jour. Petrology* **3,** 181–191.

THOMPSON, J. B. (1955) The thermodynamic basis for the mineral facies concept. *Am. Jour. Sci.* **263,** 65–103.

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

TILLEY, C. E. (1923) The petrology of the metamorphosed rocks of the Start area, South Devon. *Quart. Jour. Geol. Soc. London* **79,** 172–204.

——— (1925) Metamorphic zones in the southern Highlands of Scotland. *Quart. Jour. Soc. London* **81,** 100–110.

——— (1926) Some mineralogical transformations in crystalline schists. *Mineral Mag.* **21,** 34–43.

TURNER, F. J. AND J. VERHOOGEN (1960) *Igneous and Metamorphic Petrology.* McGraw-Hill Book Co., New York.

WILLIAMSON, D. H. (1953) Petrology of chloritoid and staurolite rocks north of Stonehaven, Kincardineshire. *Geol. Mag. London,* **90,** 353–361.

WISEMAN, J. D. H. (1934) The central and south-west Highland epidiorites: a study in progressive metamorphism. *Quart. Jour. Geol. Soc. London,* **90,** 354–417.

Manuscript received, February 14, 1964; accepted for publication, May 21, 1964.