Geological Survey of Finland

Bulletin 290

Progressive metamorphism of the metapelites in the Rantasalmi-Sulkava area, southeastern Finland

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Geologinen tutkimuslaitos · Espoo 1977



PROGRESSIVE METAMORPHISM OF THE METAPELITES IN THE RANTASALMI—SULKAVA AREA, SOUTHEASTERN FINLAND

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WITH 46 FIGURES AND 12 TABLES IN THE TEXT

GEOLOGINEN TUTKIMUSLAITOS ESPOO 1977 Korsman, K. 1977: Progressive metamorphism of the metapelites in the Rantasalmi-Sulkava area, southeastern Finland. *Geological Survey of Finland*, *Bulletin 290.* 82 pages, 46 figures, 12 tables.

The Rantasalmi—Sulkava area is located within the Precambrian schist belt of Savo, in the transitional zone between the Svecofennian and Karelian schists. The metamorphism was studied by observing the alterations in the mineral assemblages of the metapelites with reference to the total chemical composition and by noting the changes in the compositions of biotite, cordierite and garnet.

The grade of metamorphism increases in the area from north to south. Progressive metamorphism has produced the following zonality in the metapelites: mica schist zone, K-feldspar—sillimanite zone, K-feldspar—cordierite zone migmatised by potassium granite, and the garnet—cordierite—sillimanite zone intensely migmatised by the potassium granite. The mica schist zone underwent metamorphism under conditions of the amphibolite facies, and the garnet—cordierite—sillimanite zone under conditions of the granulite facies. Grading from the amphibolite to granulite facies takes place between the K-feldspar—cordierite and garnet—cordierite—sillimanite zones. The water content in the migmatising potassium granite was high enough for the granite to produce retrogressive metamorphism in the garnet—cordierite sillimanite zone. In the northern part of the K-feldspar—cordierite zone, the crystallisation temperature was 660°C and the pressure 4 kb, assuming that water pressure approached total pressure. The metamorphism in the Rantasalmi—Sulkava area is classified as a low-pressure intermediate type.

Helsinki 1977. Valtion painatuskeskus

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INTRODUCTION

Georg Barrow (1893), the first man to demonstrate that the grade of metamorphism in the Dalradian schist zone in Scotland increases in a certain direction, was the first to show that metamorphism was locally progressive. Today, there are several areas where progressive metamorphism has been established and it can be stated that the metamorphism within an area of a certain size often exhibits progressive features. By studying the progressive metamorphism in a certain area it is possible to draw conclusions concerning the temperature-pressure relationship in the metamorphism. Progressive metamorphism is not always the same in this respect and the crystallisation temperature that corresponds to a certain pressure may vary from one area to another (Miyashiro 1961). In areas that have undergone progressive metamorphism, the crystallisation conditions tend to change gradually and can thus be presented as a curve on the PT-coordinate system. The Abukuma type of low-pressure metamorphism in Japan (Shido 1958) and the Barrovian type of mediumpressure metamorphism in Scotland (Chinner 1966) are examples of the main types of progressive metamorphism. A type of metamorphism with very high pressure, that is, the jadeite-glaucophane type, may also exhibit zoning (Miyashiro 1961).

Zoning due to progressive metamorphism has been studied mainly by means of changes in the mineral assemblages, that is, by applying the principle of metamorphic facies proposed by Eskola (1915). Progressive metamorphism may also reveal itself as a change in the composition of some minerals (Miyashiro 1953, Engel and Engel 1960, Wenk 1962, Binns 1965). The progress of metamorphism is indicated particularly by changes in the iron—magnesium ratio of the Fe—Mg silicates in certain mineral assemblages (Reinhardt 1968). The P—T conditions of the metamorphism are generally estimated by means of experimentally determined mineral reactions. During the last fifteen years much has been written about the effect of P—T conditions and the grade of metamorphism in general on the Mg—Fe partition coefficients of some mineral pairs (Frost 1962, Sen and Chakraborty 1968, Perchuk 1969).

Zoning due to progressive metamorphism has been studied in detail in the Baltic Shield, e.g. north of Lake Ladoga in USSR (Glebovitskiy 1970, Velikoslavinskiy 1972). Velikoslavinskiy has determined the crystallisation temperatures for the metamorphic zones in the above-mentioned area by means of the Mg—Fe partition

coefficient for the mineral pair biotite—garnet. Gorbatschev (1968), Saxena (1968) and Saxena and Hollander (1969) have studied the influence of crystallisation conditions on the partition coefficients of mineral pairs in some Swedish and Finnish metamorphic rocks.

It has been demonstrated that in southeastern Finland the metamorphic grade of the metapelites increases from the eastern margin of the Karelian zone westwards (cf. Simonen and Vorma, in press). In describing the accessory minerals in limestone, Eskola (1919) showed that the metamorphic grade increases gradually from Karelia to southwestern Finland (e.g. between Soanlahti—Parainen). According to Eskola, the increase in the grade of metamorphism in the area from east to west is reflected in the following order of limestone types: quartz—limestone, tremolite—limestone, diopside—limestone and wollastonite—limestone.

Nykänen (1971) has shown that the late-Karelian granite at Kitee was the cause of the progressive contact metamorphism in the Karelian schist belt at Tohmajärvi. Korsman (1972 and 1975) has determined partition coefficients for some trace elements of the mineral phases in samples of the granulite facies from Lapland, West Uusimaa and Sulkava, and also in samples of the amphibolite facies from the Savo schist belt. Paavola (1976) has worked on the variation in the metamorphic grade in the Haukivesi area of the Savo schist belt by applying the Mg—Fe partition coefficients for the mineral pair biotite—garnet.

The Rantasalmi-Sulkava study area is located in the Savo schist belt within the transitional zone between the Svecofennian and Karelian formations (Fig. 1). The Svecofennian and Karelian schists were metamorphosed in the Svecokarelian orogeny 1800 to 1950 Ma ago (cf. Simonen 1971). The progressive metamorphism in the study area was brought to light during geologic mapping for the Sulkava (sheet 3144, Lehijärvi 1966) and Rantasalmi (sheet 3233, Korsman 1973) map sheets. The geologic map covered by these sheets and the location of the study area are shown in Fig. 2. The progressive metamorphism of the area is also revealed in some fashion on the geologic general maps of Savonlinna (sheet D2, Hackman and Berghell 1931) and Mikkeli (sheet C2, Frosterus 1900). These maps show that the northern study area is occupied by mica schists, mica schist-like gneisses or mica gneisses, whereas the southern half is characterised by veined gneisses. The study area has also been referred to in papers by Hackman (1933), and Korsman and Lehijärvi (1973). Gaál and Rauhamäki (1971) have described the geology around the northern half of the area. Parkkinen (1975) has constructed a model of structural geology for the Joutsenmäki intrusive and its environment 20 km east of the study area. Applying the fission track method, Lehtovaara (1976) has dated the apatites from the Joutsenmäki intrusive, and others.

Progressive metamorphism in the Rantasalmi—Sulkava area was investigated by tracing the changes in the metapelitic mineral assemblages with reference to Eskola's facies concept, and by examining the variation in the compositions of biotite, cordierite and garnet. Partition coefficients for some elements were determined



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Fig. 1. Structural elements of the Finnish bedrock after Simonen (1971, p. 1411). Presvecokarelidic: 1 a — schists and paragneisses, 1 b — granulite, 1 c — orthogneisses. Svecokarelidic: 2 — Karelidic schist belt, 3 — Svecofennidic schist belt, 4 — orogenic plutonic rocks. Prekambrian younger than the basal complex: 5 — anorogenic rapakivi intrusions, 6 — Jotnian sediments. The Rantasalmi and Sulkava map-sheet areas are marked by a rectangle.

between mineral pairs, and the suitability of partition coefficients as reflectors of crystallisation conditions discussed. Crystallisation conditions during metamorphism were estimated by means of some mineral reactions and partition coefficients. Progressive metamorphism is often accompanied by migmatisation (cf. Read 1957). Hence, in the Rantasalmi—Sulkava area, studies were made to delineate the correlation between migmatisation and progressive metamorphism, and further the influence of migmatisation on regional metamorphism in general. In the geologic description of the area, emphasis is placed largely on the role played by progressive metamorphism.

METHODS

The samples analysed were taken from a given layer or part of one in schists or gneisses. The samples extracted from migmatites were either gneiss or migmatising granite. Homogeneous samples were aimed at, and hence the specimens submitted to analyses were taken either from drill cores or road cuts.

Using the X-ray fluorescence method (Philips vacuum X-ray spectrometer), the present author analysed the samples for Si, Al, K, Ca, Fe and Ti in the Geological Department at the University of Turku; he further analysed them by AAS (Perkin Elmer 290 and Perkin Elmer 360) for Mg, Mn and Na and checked them for Al, K, Ca and Fe. The chemical compositions of some samples were determined by wet chemical methods by R. Saikkonen in the Chemistry Department of the Geological Survey. Biotites, cordierites and garnets were partially analysed by Tuula Paasivirta at the Geological Survey with the aid of a Geoscan electron microanalyser using natural biotite, cordierite and garnet as standards. The determinations were made from polished thin sections by assaying at least three adjacent points within a given part of a mineral grain. The determinations were repeated.

Using separated fractions, Ringa Danielsson and A. Puisto analysed biotites and garnets for vanadium and cobalt by means of an emission spectrograph (Jarrel-Ash, 3.4 m plane grating spectrograph) in the Chemistry Department of the Geological Survey. Duplicate analyses were performed. Biotites and garnets were separated by the present author by means of an isodynamic magnetic separator, heavy liquids and hand picking. The following symbols are used in the present study:

T = temperature

P = pressure

 $\begin{array}{l} P_{fluid} = total \mbox{ fluid pressure (Sobolev and Dobretsov 1972)} \\ P_{H_2O}, P_{CO_2} = pressure \mbox{ of } H_2O \mbox{ and } CO_2 \\ X_{Mg}, \ X_{Fe} = mole \mbox{ fraction of magnesium } \left(\frac{MgO}{FeO + MgO} \right) \mbox{ and } iron \end{array}$

METAPELITES AND METAMORPHIC ZONES

In the study area, the metamorphic grade increases from north to south. Owing to progressive metamorphism the following zones are present in the metapelites: mica schist, K-feldspar—sillimanite, K-feldspar—cordierite and garnet—cordierite sillimanite. The metamorphic zones are depicted in Fig. 2.

The location of the samples is shown in Fig. 3; the samples investigated, together with their mineral compositions are listed in Table 1. The Table does not include retrogressive alteration products. The occurrence of some minerals in the meta-morphic zones is given in Table 2; Table 3 lists the mineral assemblages typical of metapelites; the relative abundances of certain minerals are given in Table 4 and Table 5 shows the chemical compositions of the samples analysed.

Particularly in the northern part of the zone, the mica schists have preserved structures such as graded bedding (Fig. 4) which, according to the classification by Bouma (1962), are typical of turbidites. These structures have been described by Gaál (Gaál and Rauhamäki 1971) from east of the study area. In the northern part of the mica schist zone, quartz and plagioclase occur as fragments, 0.3 to 1.5 mm in size (Fig. 8), in the psammitic bottom part of the metapelite layer. The pelitic top part of the layer is in general even-grained, the grain size ranging from 0.01 to 0.02 mm. In the southern part of the zone, these clastic features disappear from the bottom part of the layer, and the top part is less fine-grained than it is in the northern part of the zone. Scattered tourmaline-bearing granite veins have been encountered in the southern part of the mica schist zone. The composition and grain size vary in the tourmaline granite, which is often pegmatitic with high Na₂O (Table 5, samples 65KS and 66KS).

The boundary between the mica schist and K-feldspar—sillimanite zones is easily delineated because of the absence from the mica schists of sillimanite, a mineral readily recognisable in K-feldspar—sillimanite gneiss outcrops (Fig. 5) as protuberances composed of accumulations of fibrolite (Fig. 9). The grain size of sillimanite, indeed of most other minerals as well, grows with the increase in metamorphic grade (Figs 8—11). K-feldspar occurs as poikiloblastic grains, 2.5 to 5 mm in size, in the metapelites of the zone. The K-feldspar—sillimanite gneisses have retained their bedding well, but not their graded bedding. Tourmaline-bearing granite veins are more numerous in the K-feldspar—sillimanite zone than in the mica schist zone.

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Fig. 2. Geological map of the Rantasalmi and Sulkava areas. The map is simplified from geological maps compiled by Lehijärvi (1966) and Korsman (1973). The Rantasalmi—Sulkava study area is outlined. Isograds: I — the K-feldspar—sillimanite isograd and the southern boundary of the mica schist zone, II — the K-feldspar—cordierite isograd and the southern boundary of the K-feldspar—sillimanite zone, III — the hypersthene isograd, IV — the garnet—cordierite—sillimanite isograd and the southern boundary of the K-feldspar—cordierite zone.

Table 1

Type and mineral composition of samples from the Rantasalmi-Sulkava metamorphic zones

Sample	Type of sample	Mineral composition
Northern	part of the mica schist zone	
1MS	Mica schist, bottom part of the layer	Qtz, Plag, Bio
2MS	»	Qtz, Plag, Musc, Bio
3MS	»	»
4MS	Mica schist, upper part of the bottom interval	Qtz, Plag, Bio
5MS	Mica schist, top part of the layer	Qtz, Plag, Musc, Bio
6MS	»	»
Southern	part of the mica schist zone	
7MS	Mica schist, top part of the layer	Otz, Ksp, Plag, Musc, Bio
8MS	Mica schist, bottom part of the layer	»
9MS	Mica schist	Otz, Plag, Bio, Cord, Gar
10MS	»	Otz, Plag, Bio, Cord, Gar, Staur
11MS	»	»
12MS	»	Otz, Plag, Bio, Cord, Gar
13MS	»	»
K-feldspa	ur—sillimanite zone	
14KS	Biotite-sillimanite gneiss	Otz. Plag. Sill. Bio
15KS	K-feldspar—sillimanite gneiss	Otz, Ksp. Plag, Musc. Sill, Bio
16KS	»	»
17KS	»	»
18KS	»	>>
19KS	Biotite—garnet gneiss	Otz. Plag. Bio. Gar
20KS	»	»»
21KS	»	»
22KS	»	>>
23KS	»	»
Northern	part of the K-feldspar—cordierite zone	
24KC	K-feldspar-cordierite gneiss	Otz, Ksp. Plag. Musc. Sill. Bio. Cord.
25KC	Biotite—garnet gneiss	Otz. Plag. Bio. Gar
26KC	K-feldspar—cordierite gneiss	Otz, Ksp. Plag. Sill. Bio. Cord
27KC	»	Otz, Ksp. Plag. Sill, Bio, Cord. Gar
28KC	»»	»
29KC	»	Otz, Ksp. Plag. Sill. Bio. Cord
30KC	»»	»
31KC	»»	»
32KC	»»	Otz, Ksp. Plag. Sill, Bio, Cord, Gar
33KC	»»	»»
34KC	»	»
Southern	part of the K-feldspar—cordierite zone	
35KC	K-feldspar-cordierite gneiss	Qtz, Ksp, Plag, Bio, Cord, Gar
36KC	»	Qtz, Ksp, Plag, Sill, Bio, Cord. Gar
37KC	»	»
38KC	»	»
39KC	»	»
40KC	Biotite-garnet gneiss	Qtz, Plag, Bio, Gar
41KC	Biotite-cordierite gneiss	Qtz, Plag, Bio, Cord, Gar
42KC	»	»
43KC	K-feldspar-cordierite gneiss	Qtz, Ksp, Plag, Sill, Bio, Cord. Gar
44KC	Biotite-cordierite gneiss	Qtz, Plag, Bio, Cord, Gar
	0	· ,,,,,,,,,,,,,,,,,,,

Table 1	(continued)
Table I	(continueu)

Sample	Type of sample	Mineral composition
Southern	part of the K-feldspar—cordierite zone	
45KC	K-feldspar-cordierite gneiss	Otz Ksp Plag Sill Bio Cord Gar
46KC	»	Otz Ksp Plag Bio Cord Gar
47KC	>>	Otz Ksp. Plag. Sill. Bio. Cord. Gar
48KC	»»	»
49KC	Biotite-cordierite gneiss	Qtz, Plag, Bio, Cord, Gar
Garnet-	–cordierite—sillimanite zone	
50GC	Garnet-cordierite-sillimanite gneiss	Qtz, Ksp, Plag, Sill, Bio, Cord, Gar
51GC	»	»
52GC	»»	»
53GC	»	»
54GC	»	»
55GC	»	»
56GC	»	»
57GC	»»	»
58GC	»	»
59GC	»	»
60GC	»	»
61GC	Garnet-cordierite gneiss	»
62GC	Garnet—hypersthene gneiss	Qtz, Plag, Bio, Gar, Hyp
63GC	»	»
64GC	»	»
K-feldst	bar—sillimanite zone	
65KS	Tourmaline granite	Qtz, Ksp, Ab, Musc, Tourmaline
66KS	»	»
Norther	rn part of the K-feldspar—cordierite zone	
67KC	Microcline granite	Qtz, Ksp, Plag, Bio
Southern	n part of the K-feldspar—cordierite zone	
68KC	Microcline granite	Otz, Ksp. Plag. Bio
69KC	»	»
UTILO	<i>17</i>	
Garnet-	-cordierite-sillimanite zone	
70GC	Microcline granite	Otz Ksp Plag Sill Bio Gar
71GC	w	Otz Ksp Plag Bio
7260		w
1200	"	"

Ine	occurrence of	some inmerals	in the Kantasam	m—Sulkava metamo	orphic zones
	Mica schist zone	K-feldspar— sillimanite zone	Northern part of th K-feldspar—cordierit zone	e Southern part of the e K-feldspar—cordierite zone	Garnet—cordierite— sillimanite zone
K-feldspar	rare	common	common	common	common
Plagioclase	common (An 25—35)	common (An 25—35)	common (An 25—35)	common (An 25—35)	common, but amount often small
Muscovite	common	vanishing	rare and only as r	etrogressive phase	common, but only as retrogressive phase
Sillimanite	not found	common	common	common, but only as inclusions in cordierite and garnet	common
Biotite	common	common	common	not in association with sillimanite	common, but amount often small
Cordierite	not in assoc K-feldspar	iation with	common	common	common
Garnet	not in assoc K-feldspar	iation with	common	common	common

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The occurrence of some minerals in the Rantasalmi-Sulkava metamorphic zones

Table 3

Common mineral assemblages of metapelites in the Rantasalmi-Sulkava metamorphic zones

Mica schist zone	Qtz—Plag—Musc—Bio
K-feldspar—sillimanite zone	Qtz—Ksp—Plag—Sill—Bio \pm Musc
Northern part of the K-feldspar-cordierite zone	Qtz-Ksp-Plag-Sill-Bio-Cord-or Gar
Southern part of the K-feldspar-cordierite zone	Qtz—Ksp—Plag—Bio—Cord—Gar
Garnet—cordierite—sillimanite zone	Qtz—Ksp—Sill—Cord—Gar \pm Plag

The metapelite outcrops in the K-feldspar—cordierite zone are migmatic veined gneisses (Fig. 6). Bedding is distinct in the northern part of the zone but in the southern part it is not easily recognisable. As a rule, the abundance of cordierite is fairly high in metapelite outcrops. Cordierite and garnet occur as porphyroblasts, whereas K-feldspar grains are often poikiloblastic (Fig. 10). Practically all the metapelite samples investigated contained garnets, although the content was low: in the northern part of the zone it is less than one percent and in the south from 1 to 1.5 percent.



Fig. 4. Mica schist with turbidite structures from the northern part of the mica schist zone. Length of scale lath 12 cm.



Fig. 5. Bedded gneiss from the K-feldspar—sillimanite zone. Sillimanite knots are visible in the weathered surface. Length of scale lath 3 cm.



Fig. 6. Migmatic bedded K-feldspar—cordierite gneiss from the northern part of the K-feldspar—cordierite zone. Length of scale lath 12 cm.



Fig. 7. Migmatic gneiss from the garnet—cordierite—sillimanite zone. Length of scale lath 12 cm.



Fig. 8. Sketch from photomicrograph showing bottom part of the mica schist layer. The northern part of the mica schist zone. Grey mineral is biotite.



0.5mm

Fig. 9. Sketch from photomicrograph showing K-feldspar—sillimanite gneiss. Sillimanite knot consists of fibrolite. Grey mineral is biotite.

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Fig. 10. Sketch from photomicrograph showing K-feldspar—cordierite gneiss. The northern part of the K-feldspar—cordierite zone. K-feldspar porfyroblast is grey and biotite black. The biotite-foliation visible in K-feldspar is older than the biotite-foliation around the cordierite porfyroblast.



0.5mm

Fig. 11. Sketch from photomicrograph showing garnet-cordierite-sillimanite gneiss. Black mineral is ilmenite.

	Mica schist zone	K-feldspar— sillimanite zone	Northern part of the K-feldspar- cordierite zone	Southern part of the K-feldspar- cordierite zone	Garnet— cordierite— sillimanite zone
Sample	7 MS	18KS	27KC	43KC	52GC
Quartz	16.9	22.0	19.7	20.9	14.5
K-feldspar	4.1	7.8	15.6	24.2	19.0
Plagioclase	17.8	9.5	26.2	12.6	1.5
Muscovite	12.0	2.1			
Sericite and Pinite			0.2	0.1	4.2
Sillimanite		10.9	3.0		6.5
Biotite	47.8	47.3	27.8	19.9	8.6
Cordierite			7.3	21.2	25.5
Garnet				0.9	18.5
Opaques	1.1		_	0.1	1.6
Other minerals	0.3	0.4	0.2	0.1	0.1
	100.0	100.0	100.0	100.0	100.0

Table 4

Modal composition of metapelites in the Rantasalmi-Sulkava metamorphic zones

Metamorphic hypersthene-bearing quartz diorites and hypersthene gneisses occur in the southern part of the K-feldspar—cordierite zone. Hypersthene has not been encountered in corresponding rocks in the northern part of the zone, in the Kfeldspar—sillimanite zone or in the mica schist zone.

The garnet-cordierite-sillimanite zone contains a higher proportion of migmatising granite than does the K-feldspar-cordierite zone (Figs. 2 and 7). Bedding is visible in the garnet-cordierite-sillimanite gneiss outcrops only in those places that are exceptionally rich in calcium. In thin sections made from these gneisses, fairly large portions are composed of garnet, cordierite and sillimanite, and are often separated from each other by granitic segregations (Fig. 11). In these samples, K-feldspar does not occur as poikiloblastic grains, and hence the texture exhibited by the garnet-cordierite-sillimanite gneisses in these sections differs from that of the K-feldspar-cordierite gneisses. The quantity of garnet reaches several percent in the garnet-cordierite-sillimanite gneisses. It is, however, less than 10 percent and also less than that of cordierite. In these gneisses, the cordierite is frequently intensely pinitised, and the K-feldspar sericitised, wheras in the metapelites of the other zones the amount of water-bearing alteration products is low. In the garnetcordierite-sillimanite zone the migmatising granite tends to contain garnets and cordierite. The thin sections reveal that the garnet- and cordierite-bearing portions are gneiss fragments in granite. In the garnet-cordierite-sillimanite zone, hypersthene-bearing quartz diorites and gneisses are more common than in the K-feldsparcordierite zone. The hypersthene gneisses and hypersthene quartz diorites have not been migmatised at all or show only a slight degree of migmatisation. Further, the abundance of water-bearing alteration products in these samples is low.

					Mica	which man	1						K_fe	ldspar_si	llimanite	None		
Sample	1MS	2MS	3MS	4MS	5MS	6 MS	7MS	8 M S	10 M S	11 M S	14KS	15KS	16KS	17KS	18KS	19KS	20KS	22KS
SiO ₂	70.5	70.0	69.1	65.6	54.9	58.3	60.0	72.3	60.6	58.2	59.9	74.6	63.2	59.8	62.4	63.4	60.0	60.7
TiO,	0.5	0.4	0.5	0.8	0.9	0.9	0.7	0.5	1.3	1.6	0.7	0.5	0.5	1.1	0.8	0.7	0.8	0.7
$Al_2 \tilde{O}_3 \ldots \ldots$	14.8	14.9	14.9	14.5	19.2	18.4	18.9	13.9	19.3	17.0	18.2	12.5	17.9	19.3	17.2	14.8	15.2	14.5
FeO (Fe total																		
as FeO)	4.4	4.3	4.6	5.6	8.8	8.5	4.6	2.9	7.2	10.0	7.7	3.9	5.8	7.1	7.5	10.4	9.6	11.2
MnO	0.02	0.03	0.03	0.07	0.08	0.05	0.08	0.04	0.03	0.06	0.06	0.02	0.05	0.07	0.06	0.02	0.09	0.04
MgO	2.2	1.9	1.9	3.1	5.0	4.2	2.9	1.6	2.4	3.4	4.1	1.3	3.1	3.4	2.8	3.3	3.2	3.3
CaO	1.4	1.0	1.3	2.1	1.4	1.3	1.2	1.5	4.0	3.9	1.0	0.8	1.2	0.7	0.4	2.1	1.5	1.8
Na ₉ O	2.7	2.8	3.1	4.0	1.6	3.2	1.9	2.4	3.1	2.5	2.6	1.8	2.6	2.4	2.2	2.7	3.5	2.4
K ₂ Õ	2.8	2.8	2.5	2.6	5.5	5.2	5.5	2.9	1.6	1.7	3.6	3.3	4.4	4.5	5.1	3.4	3.6	3.4
	99.3	98.1	97.9	98.4	97.4	100.1	95.8	98.0	99.5	98.4	97.9	98.7	98.8	98.4	98.5	100.8	97.5	98.0

				Tab	le 5		
Chemical	composition	of samples	in	the	Rantasalmi-Sulkava	metamorphic	zones

			Northern	part of	the K-feld	spar—cord	lierite zon	e		Son	thern par	t of the K	C-feldspar-	_cordieri	te zone		
Sample	24KC	25KC	26KC	27KC	30KC	31KC	32KC	34KC	 35KC	36KC	40KC	41KC	43KC	44KC	45KC	46KC	47KC
SiO ₂	63.2	63.9	62.8	62.6	67.6	71.1	69.5	74.2	66.0	61.4	71.2	73.2	60.4	72.6	56.7	76.9	59.8
TiO ₂	0.7	0.6	0.8	0.9	0.9	0.4	0.6	0.6	0.8	0.7	0.6	0.7	0.9	0.7	1.3	0.5	0.9
Al ₂ O ₃	16.8	15.7	17.9	17.8	16.2	13.4	14.2	12.4	16.7	16.0	12.1	13.3	18.6	12.5	20.4	12.5	19.3
FeO (Fe total																	
as FeO)	6.4	9.4	6.4	7.3	6.2	4.5	4.9	5.1	4.5	10.3	5.5	4.8	8.4	5.3	9.4	3.2	7.3
MnO	0.06	0.14	0.00	0.04	0.04	0.06	0.01	0.02	0.02	0.1	0.03	0.02	0.03	0.02	0.04	0.00	0.02
MgO	2.6	2.8	2.4	3.2	3.0	2.0	1.8	1.6	2.2	2.8	2.4	3.2	4.2	2.4	5.0	1.5	2.8
CaO	0.9	0.6	1.0	1.4	0.8	0.9	0.8	0.4	2.5	0.9	1.5	1.6	0.8	1.1	0.5	1.3	0.6
Na.0	2.1	1.4	2.5	2.8	1.7	2.5	2.4	1.2	3.6	1.9	2.6	3.7	1.4	2.0	2.0	2.2	2.1
K ₂ Õ	4.2	2.2	3.9	4.1	3.6	3.0	2.9	2.7	2.2	4.5	1.9	2.1	4.5	2.1	5.4	1.4	5.3
	97.0	96.7	97.7	100.1	100.0	97.9	97.1	98.2	98.5	98.6	97.8	102.6	99.2	98.7	100.7	99.5	98.1

	Southern	part of t	be		G	arnet—cor	dierite—si	llimanite :	zone		K-feldspar—s	illimanite zone
Sample	K-Jelaspar— 48KC	49KC	zone	50GC	51GC	52GC	53GC	56GC	60GC	61GC	65KS	66KS
SiO,	62.9	67.3		62.6	60.9	61.6	56.7	60.7	61.4	65.4	72.8	72.9
TiO,	0.9	0.8		0.9	1.2	1.1	1.3	0.7	0.8	0.9	0.1	0.1
$Al_2 \tilde{O_3} \dots$ FeO (Fe total	17.5	15.6		21.3	21.0	21.5	21.8	21.3	20.9	15.9	16.8	17.0
as FeO)	7.1	5.3		7.5	8.6	8.0	10.5	7.0	7.6	6.7	0.6	0.9
MnO	0.03	0.03		0.01	0.05	0.03	0.05	0.06	0.04	0.04	0.00	0.06
MgO	3.0	2.1		3.3	3.4	3.0	3.7	3.2	2.8	3.0	0.1	0.3
CaO	0.6	2.1		0.2	0.4	0.6	0.8	0.3	0.2	1.5	0.6	0.7
Na.O	1.5	3.7		0.8	0.9	1.6	1.2	1.2	1.2	1.5	6.5	5.0
K ₂ Ô	5.5	1.5		3.0	3.9	4.7	2.9	5.5	3.9	3.0	0.8	4.5
	99.0	98.4		99.6	100.4	102.1	98.9	99.7	98.8	97.9	98.3	101.5
	K-feldspar	—cordieri	ite zone		Gar	net—cordi	erite—silli	imanite zo	ne		 	
Sample	67KC	68KC	69KC			70GC	71GC	72GC				
SiO	73.8	73 2	73.0			74.2	734	72.6				
TiO.	0.2	0.1	0.02			0.2	0.1	0.3				
Al_2O_3 FeO (Fe total	16.9	17.0	16.5			17.8	16.5	13.4				

2.0

0.7

0.5

3.0

5.3

103.7 102.6

0.00

0.7

0.00

0.5

0.8

3.0

7.6

0.9

0.00

0.7

0.7

2.8

7.5

98.9

logical Survey.

as FeO) ...

MnO

MgO

CaO

 $\begin{array}{c} \text{Na}_2\text{O} \\ \text{Na}_2\text{O} \\ \text{K}_2\text{O} \\ \end{array}$

0.6

0.4

0.4

3.8

7.5

0.01

1.7

0.02

0.5

0.9

3.9

6.2

103.6 103.5 101.2

0.3

0.00

0.2

0.6

3.7

6.9

Table 5 (continued)

Anal. by the author except 7MS, 8MS and 60GC, which were analysed by R. Saikkonen in 1967 and 1973 at the GeoGeological Survey of Finland, Bulletin 290

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Table 4 lists the contents of the minerals in some specimens determined by the point counting method. The compositions thus obtained depend on the part of the sample from which the thin section was made. However, Table 4 clearly shows what changes take place in the amounts of minerals, such as muscovite, K-feldspar, sillimanite, cordierite, garnet and biotite, when the metamorphic grade increases. The chemical compositions of the samples listed in the Table are consistent with those in the upper portion of the mica schist layer. The content of biotite may approach 50 percent in mica schists; in the southern part of the K-feldspar—cordierite zone it is often 15 to 20 percent, whereas in the garnet—cordierite—sillimanite gneisses it is less than 10 percent.

Apatite and zircon are encountered as accessories in the metapelites of all the zones. Mica schists and K-feldspar—sillimanite gneisses frequently contain tourmaline as well. In the garnet—cordierite—sillimanite gneisses, ilmenite and hercynitic spinel are met with as opaque minerals, whereas in samples from other zones ilmenite is often the only opaque. In the garnet—cordierite—sillimanite gneisses the opaques total 1 to 1.5 percent and in the metapelites of the other zones the content of ilmenite is generally less than 0.2 percent. In some samples from the mica schist zone, however, opaques may exceed 0.2 percent.

The quartz diorites and granodiorites are most abundant in the K-feldspar sillimanite zone, where they have folded concordantly with the enveloping schists, and neither cut nor migmatise these schists. The quartz diorites and granodiorites were crystallised under conditions similar to those of the enveloping schists and, like them, their metamorphic grade also increases southwards. In the map sheet areas there are some quartz diorites and granodiorites, such as the Luonterivesi granodiorite on the Sulkava map, that differ from those described above. The Luonterivesi granodiorite crystallised under conditions unlike those of the gneisses in the environment and its zircon exhibits a radiometric age 100 million years younger than the corresponding age of the synkinematic Svecokarelian intrusives (Korsman and Lehijärvi 1973). The age of this granodiorite is also younger than that of the metamorphism of the adjoining garnet—cordierite gneisses and hypersthene-bearing quartz diorites.

The migmatising granite is a microline granite with a high K_2O content exceeding that of Na₂O (Table 5, samples 67KC, 68KC, 69KC, 70GC, 71GC, 72 GC). The grain size varies but is often coarse. This granite is most abundant in the garnet—cordierite—sillimanite zone, and markedly so when compared with its frequency in the southern part of the K-feldspar—cordierite zone. The granite is concordant with the foliation, but also crosscuts other rocks. The microcline granite in the area is similar to the late-kinematic microcline granites described from southern Finland by Simonen (1960).

The subvertical NW—SE trending foliation in the mica schist zone maintains its attitude in the K-feldspar—sillimanite zone, but farther off in the central and southern parts of the K-feldspar—sillimanite zone, the foliation shows a tendency to dip more gently.

Between the K-feldspar—sillimanite zone and the K-feldspar—cordierite zone there is a transition zone that occasionally reveals the mineral assemblage K-feldspar cordierite. Beyond this transition zone the gently dipping and NW—SE trending foliation characteristic of the southern part of the K-feldspar—sillimanite zone fades out and the mode of folding changes in the northern part of the K-feldspar—cordierite zone. Similarly, between the K-feldspar—cordierite zone and the garnet—cordierite sillimanite zone there is a transition zone, a few kilometres in width, in which the quantity of migmatising granite increases southwards. In and beyond this transition zone the E—W striking vertical foliation planes of the K-feldspar—cordierite zone start to bend. Gently dipping foliation predominates in the garnet—cordierite sillimanite zone.

As a rule, the metamorphic grade in the environment of the study area obeys the same rules as it does within the area, that is the grade of metamorphism increases southwards in the areas covered by the Rantasalmi and Sulkava map sheets. However, it also increases northwards, starting from the northern part of the Rantasalmi map sheet area.

CHANGES IN MINERAL ASSEMBLAGES

Metamorphic facies

Eskola (1939, p. 339) gives the following definition of metamorphic facies: »In a definite facies are united rocks which for identical bulk composition exhibit an identical mineral composition, but whose mineral composition for varying bulk composition varies according to definite laws». (Translated by Turner; Fyfe, Turner and Verhoogen 1958, p. 9). J. B. Thompson (1955 and 1957) and Korzhinskii (1959) have studied metamorphism on the basis of closed and open systems. The facies concept can be stated as follows (Korzhinskii 1959, p. 64): »Rocks which formed under such similar external conditions that the relation between their chemical composition in terms of the inert components, and their mineralogical composition is marked by the same regularities, can be considered as belonging to the same mineral facies». The inert components are those for which the system is closed, and whose relative abundances control the formation of an equilibrated mineral assemblage. Eskola (1915) maintained that water did not need to be taken into consideration in studies on equilibrated mineral assemblages, because water was available in excess. According to Korzhinskii and Thompson, water is a mobile component and for it the system is open. Thus, the influence of water on metamorphic equilibrium depends on external conditions and, in this respect, it is comparable with pressure and temperature.

The relationship between mineral assemblages and bulk chemical composition can be illustrated on ACF and A'KF triangular diagrams (Eskola 1915). Only A'KF triangles are used in the present study because the presentation of the composition of metapelites on ACF triangles requires corrections owing to muscovite and biotite (Eskola 1939, p. 350 and Winkler 1967, p. 49).

The data on the samples analysed were plotted on A'KF diagrams in accordance with the instructions given by Eskola (1915) and Winkler (1967). The results were not corrected for apatite because the majority of the samples were not assayed for phosphorous. Most of the samples were analysed for total iron, and so, in all samples iron was computed as FeO. The A'KF diagrams are given in Figs. 12—16.

According to Eskola (1915, p. 123), the variation in occurrence of garnet and cordierite in the Orijärvi region is a result of the variation in the bulk FeO/MgO ratio. Although the bulk FeO/MgO ratio may be decisive for the crystallisation of the Fe—Mg silicates, FeO and MgO are considered as one component on the A'KF diagram. Hence, A'KF triangles alone cannot delineate the crystallisation of the mineral assemblages. Thompson (1957) has constructed a projection from an AKFM tetrahedron, whose AFM plane allows the presentation of the Fe—Mg silicates and K_2O -bearing minerals of the quartz-bearing metapelites. Muscovite (Thompson 1957) or K-feldspar (Barker 1962, Reinhardt 1968 and Dallmeyer 1972) have been used as the projection points. In the present study the AFM values were calculated as follows:

$$\begin{split} A &= \frac{[Al_2O_3] - [3K_2O]}{[Al_2O_3] - [3K_2O] + [FeO] + [MgO]} & \text{muscovite as projection point} \\ A &= \frac{[Al_2O_3] - [K_2O]}{[Al_2O_3] - [K_2O] + [FeO] + [MgO]} & \text{K-feldspar as projection point} \\ F &= \frac{[FeO]}{[FeO] + [MgO]} & \text{muscovite or} \\ M &= \frac{[MgO]}{[FeO] + [MgO]} & \text{K-feldspar as projection point} \end{split}$$

The formulas in brackets refer to the molecule percentages. The Al_2O_3 incorporated in plagioclase was subtracted from Al_2O_3 by assuming that all the CaO and Na₂O is bound to plagioclase. Iron is as total Fe. The use of the AFM projection implies that the rock contains either muscovite (projected from the muscovite point) or K-feldspar (projected from the K-feldspar point). The AFM projections for the present study are given in Figs. 17–21.

According to several facies classifications the mineral assemblage quartz—placioclase $(An_{25}-An_{35})$ —muscovite—biotite, typical of the mica schist zone, belongs to the amphibolite facies (Eskola 1939, Winkler 1967, Miyashiro 1973). Winkler (1967, p. 119) maintains that the crystallisation of staurolite and cordierite best



Fig. 12. Calculated chemical compositions and observed mineral assemblages or mineral compositions of analysed mica schists shown on A'KF diagram. Staurolite has been met with only as armoured relic.



Fig. 13. A'KF diagram for metapelites of the K-feldspar—sillimanite zone.

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Fig. 14. A'KF diagram for metapelites of the northern part of the K-feldspar—cordierite zone. Sillimanite may be in contact with biotite (Fig. 22).



Fig. 15. A'KF diagram for metapelites in the southern part of the K-feldspar—cordierite zone. Sillimanite occurs only as inclusions in cordierite or garnet (Fig. 23).



Fig. 16. A'KF diagram for melanosomes of the garnet—cordierite—sillimanite zone.

indicates the transition from green-schist facies to amphibolite facies. In the mica schist zone in the study area, garnet, cordierite and staurolite were encountered in only one exposure close to the southern boundary of the zone. Garnet and cordierite are detectable within the same thin section, but not in contact with each other. Staurolite occurs only as an armoured relic in either cordierite or garnet. These garnet-, cordierite- and staurolite-bearing samples do not contain muscovite.

According to Winkler (1974, p. 217), with the increase in metamorphic grade staurolite disappears earlier from rocks containing muscovite than from those devoid of that mineral. Hence, staurolite is not obviously stable in samples containing muscovite, not even in the northern part of the mica schist zone. As shown by the A'KF diagram and the AFM projection (Figs. 12 and 17), the composition of the muscovite-bearing mica schists prevented the crystallisation of sillimanite, cordierite, garnet and staurolite in these samples. Since garnet and cordierite are lacking from the muscovite-bearing samples, the fields of these minerals cannot be presented on an AFM projection. In several other areas, where crystallisation conditions approximate those of the mica schist zone, cordierite and garnet have been established as phases in equilibrium in samples containing muscovite, garnet being typical of comparatively high pressure metamorphism and cordierite of rather low pressure metamorphism (Winkler 1967, Turner 1968, Miyashiro 1973).

The mineral assemblage K-feldspar-sillimanite is characteristic of the K-feldspar-sillimanite zone. This assemblage has not been observed in the metapelites



Fig. 17. AKFM projection through muscovite onto the AFM plane for metapelites of the mica schist zone. Calculated chemical compositions of samples and generalised compositions of biotite, cordierite and garnet (Figs. 17-21). Symbols are as in Fig. 12. Garnet, cordierite and staurolite have not been met with in muscovite-bearing samples.

of the mica schist zone, although, as shown by the A'KF triangle (Figs. 12 and 13), the chemical compositions of the samples analysed are similar in both the mica schist and K-feldspar-sillimanite zone. Thus, the assemblage sillimanite-K-feldspar indicates a change in crystallisation conditions. Neither garnet nor cordierite have been detected in samples containing K-feldspar in the K-feldspar-sillimanite zone. On the AFM projection of this zone (Fig. 18) the biotite-sillimanite field is so large that garnet or cordierite could only have crystallised in those K-feldspar-bearing samples whose X_{Mg} atomic ratio was either exceptionally high or exceptionally low.

Hence, the tie line joining sillimanite to biotite on the A'KF triangle is justified. According to several facies classifications, the assemblage quartz-K-feldsparoligoclase-biotite-sillimanite of the K-feldspar-sillimanite zone (Winkler 1967, Miyashiro 1973) belongs to a subfacies of the amphibolite facies whose rocks are of a higher metamorphic grade than that represented by the corresponding subfacies of mica schists.

The samples taken from the northern portion of the K-feldspar-cordierite zone often exhibit garnet and cordierite in the same thin section, but these minerals are seldom in contact with each other. The composition of the biotite in contact with garnet tends to differ from the composition of the biotite that borders cordierite.

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Fig. 18. AFM projection through K-feldspar for metapelites of the K-feldspar—sillimanite zone. Garnet and cordierite have not been met with in muscovite-bearing samples. Symbols are as in Fig. 13.



Fig. 19. AFM projection through K-feldspar for metapelites of the northern part of the K-feldspar—cordierite zone. Sillimanite may be in contact with biotite. Symbols are as in Fig. 14.







Fig. 21. AFM projection through K-feldspar for melanosomes of the garnet—cordierite—sillimanite zone. Symbols are as in Fig. 16.

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Fig. 22. Sketch from photomicrograph showing biotite in direct contact with sillimanite. The northern part of the K-feldspar—sillimanite zone. The pair sillimanite—biotite is stable. Grey mineral is biotite.

Particularly the abundance of TiO_2 is invariably higher in the biotite adjacent to cordierite than in that in contact with garnet (Table 6 and Fig. 27). The samples manifest the mosaic equilibrium defined by Korzhinskii (1959, p. 19). The three-phase field biotite—cordierite—garnet on the AFM projection (Fig. 19) was constructed on the basis of those samples in which garnet and cordierite are in contact (Samples 33KC and 34KC). If garnet, cordierite and biotite are assumed to be merely Fe—Mg-aluminosilicates, then, according to the phase rule (Korzhinskii 1959, p. 63), their compositions are fixed in siliceous samples, when they are in equilibrium and external conditions remain constant. The metapelites in the northern part of the K-feldspar—cordierite gneiss zone contain cordierite, but garnet may be absent or sparse. The AFM projection and the analytical data suggest that garnet is lacking, because the size of the garnet field is very small compared with the total X_{Mg} ratio. In the northern part of the K-feldspar—cordierite zone, sillimanite may be in contact with biotite (Fig. 22).

In the southern part of the K-feldspar—cordierite zone, sillimanite occurs as inclusions in cordierite or garnet, and has not been observed in contact with biotite (Fig. 23). Accordingly, the pair sillimanite—biotite is not in equilibrium in the southern part of the K-feldspar—cordierite zone. Kays (1976), who noted that in the Merimasku—Naantali area garnet, cordierite or K-feldspar separate sillimanite from biotite, maintains that the sillimanite—biotite pair is in disequilibrium. In the



0.5 mm

Fig. 23. Sketch from photomicrograph showing sillimanite inclusions in cordierite. The southern part of the K-feldspar sillimanite zone. The pair biotite—sillimanite is not stable. Grey mineral is biotite.

southern part of the K-feldspar—cordierite zone, in the Rantasalmi—Sulkava area, the garnet field has expanded further towards the AM side of the AFM projection (Fig. 20) than it has in the northern part of the zone. Thus, in the southern part of the zone, the metapelite samples invariably contain garnet, although, here too, the amount is low (1 to 1.5 percent). The three-phase field biotite—cordierite—garnet on the AFM projection of the K-feldspar—cordierite gneiss zone was constructed on the basis of the samples in which these minerals are in contact (samples 39KC, 43KC and 48KC).

The garnet—cordierite—sillimanite zone is characterised by the assemblage quartz—K-feldspar—garnet—cordierite—sillimanite \pm plagioclase. All the samples studied also contain biotite. As shown by the AFM projection (Fig. 21), the chemical composition of the garnet—cordierite—sillimanite gneisses differs from that of the metapelites in the other zones. The composition points of the garnet—cordierite—sillimanite gneisses lie in or close to the garnet—cordierite—sillimanite field on the AFM projection.

Sample 61GC does not contain the assemblage sillimanite—garnet—cordierite and its composition falls within the biotite—cordierite—garnet field. Thus, the assemblage garnet—cordierite—sillimanite need not necessarily reflect changes in crystallisation conditions compared with the K-feldspar—cordierite gneiss zone; primarily, the assemblage indicates changes in the chemical composition of the metamorphosed material. On the basis of sample 61GC, also the garnet—cordierite field has been constructed on the AFM projection of the garnet—cordierite—sillimanite zone. This field is lacking from the corresponding projections for the K-feldspar—cordierite zone because the assemblage garnet—cordierite—sillimanite was not found there. Hence, it cannot be established whether or not the garnet field of the garnet—cordierite—sillimanite zone is larger than that of the southern part of the K-feldspar—cordierite zone. The compositions of the cordierite and garnet in sample 61GC suggest, however, that the garnet field has expanded towards the AM side of the AFM projection. Although all the garnet—cordierite—sillimanite gneisses contain biotite, the AFM projection shows that sillimanite is unstable with biotite. The assemblage quartz—K-feldspar—sillimanite—garnet—cordierite—biotite \pm plagioclase is problematic in terms of the phase rule.

Barker (1962) has presented a 4-phase biotite—cordierite—sillimanite—garnet field on an AFM projection. According to Khlestov (1973 a, pp. 20 and 24), the assemblage quartz—orthoclase—sillimanite—garnet—cordierite—biotite is divariant in relation to pressure and temperature. Reinhardt (1968) maintains that plagioclase and sillimanite are incompatible phases in the presence of cordierite, garnet, biotite, alkali feldspar, quartz, and opaque oxide. The X_{Mg} ratio of the biotite in the garnet cordierite—sillimanite zone in the study area may vary considerably, even in contact with the same garnet grain (Table 6, samples 52GC, 54GC and 57GC). Retrogressive alterations have also been intense in this zone and once more biotite may occasionally be in contact with sillimanite. These observations show that the equilibrium of biotite is not self-evident in the garnet—cordierite—sillimanite zone.

According to Eskola (1952), the garnet—cordierite gneisses are equilibrium over an extensive P—T field and they may belong to either the amphibolite or granulite facies. In many facies classifications, the granulite facies is distinguished from the amphibolite facies by the dehydration of biotite and hornblende and, above all, by the crystallisation of hypersthene in the granulite facies. Khlestov (1973 b, p. 99) holds that hypersthene is in equilibrium in the amphibolite facies only in the high-temperature parts of this facies. Turner (1968, p. 320) has noted that biotite and hornblende may dehydrate over a fairly large temperature range. According to Miyashiro (1973, pp. 306—307), there are extensive transitional zones between the amphibolite and granulite facies because the dehydration of hornblende and biotite into hypersthene is a gradual process. On the other hand, the ideal granulite facies, in which biotite and hornblende are not equilibrium phases, is questionable and certainly rare (Turner 1968, Khlestov 1973 a, Miyashiro 1973).

Random occurrences of hypersthene quartz diorites and hypersthene gneisses have been noted in the southern part of the K-feldspar—cordierite zone but no hypersthene-bearing rocks have been encountered in the northern part of the zone. In the garnet—cordierite—sillimanite zone hypersthene occurs almost invariably in rocks with a small excess of aluminium over alkalis and calcium. In this zone, hyper-

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sthene is in equilibrium in some garnetiferous rocks as well. The crystallisation of hypersthene in the study area has been restricted not only by external factors but also by the total chemical composition of the metamorphosed matter. As a rule, hypersthene has only been crystallised in rocks with a small aluminium excess. Especially in the K-feldspar—cordierite zone the abovementioned rocks are locally of sporadic occurrence.

In the metapelites of the K-feldspar—cordierite zone the abundance of biotite is relatively high (15—20 percent), but in the garnet—cordierite—sillimanite gneisses it is low (less than 10 percent). This low abundance of biotite is partly an outcome of the large excess of aluminium in the bulk composition. Thus, in the garnet cordierite—sillimanite zone the content of biotite may exceed 10 percent in the hypersthene gneisses. Hornblende is in equilibrium throughout the study area.

As indicated by the observations described above, the garnet—cordierite gneisses in the northern part of the K-feldspar—cordierite zone were metamorphosed under conditions of the amphibolite facies. In the southern part of the zone, hypersthene started to crystallise and the conditions changed to correspond to those of the granulite facies.

The garnet—cordierite—sillimanite zone was metamorphosed under conditions of the granulite facies. The facies boundary between the southern part of the Kfeldspar—cordierite zone and the garnet—cordierite—sillimanite zone cannot be accurately delineated, because the bulk composition may have had a decisive influence on the formation of some mineral assemblages.

Isograds

In the Dalradian schist area in Scotland, Barrow (1893) mapped, among others, the following zones: biotite, garnet, staurolite, kyanite and sillimanite. According to him, the zones were developed by progressive metamorphism and each zone is characterised by a certain index mineral. Tilley (1924 and 1925) developed the concept of isograds on the basis of these zones. The schists of the Dalradian are homogeneous in chemical composition and thus, the crystallisation of a new index mineral indicates a change in crystallisation conditions. According to Tilley, an isograd is a line on a map that joins those points at which the index mineral occurs for the first time and where the P—T conditions are approximately equal.

As shown by AKF triangles and AFM projections, the composition of the metapelites in the study area is homogeneous in so far that the K-feldspar—sillimanite isograd for example, reflects similar crystallisation conditions and changes under these conditions. On the other hand, the composition of the metapelites is limited in such a way that the K-feldspar—cordierite isograd does not indicate the first possible crystallisation of the K-feldspar—garnet and the K-feldspar—cordierite assemblages in terms of crystallisation conditions. Hence, in the present study, isograd is used not merely to reflect a univariant reaction, but to describe a new assemblage in general and similar crystallisation conditions by taking into consideration whenever possible the chemical composition of the rocks. Some isograds of the Rantasalmi—Sulkava area are shown in Fig. 2.

The crystallisation of staurolite has been limited not only by external factors but also by the chemical composition of the metapelites. Staurolite has been observed only as armoured relics in some samples from the mica schist zone, and so it has not been possible to delineate accurately the boundary at which staurolite disappears. This is, however, before the K-feldspar—sillimanite isograd.

The K-feldspar—sillimanite isograd has often been interpreted to indicate the reaction suggested by Harker (1932, p. 56):

1) $H_2KAl_3(SiO_4)_3 + SiO_2 = KAlSi_3O_8 + Al_2SiO_5 + H_2O$. In some areas muscovite has been noted to be an equilibrium phase above the K-feldspar—sillimanite isograd (Evans and Guidotti 1966 and Ashworth 1975). This suggests that the isograd reaction is continuous in terms of P—T conditions. According to Evans and Guidotti (1966), the isograd reaction in Bryant Pond, Maine, was as follows:

2) K.₉₄Na.₀₆Al₂Si₃O₁₀(OH)₂ + SiO₂ + 0.1 (NaAlSi₃O₈) \leftrightarrows

1.1 (K. $_{86}$ Na. $_{14}$ AlSi $_{3}$ O $_{8}$) + Al₂SiO₅ + H₂O.

The lower the An content in plagioclase, the earlier the isograd reaction begins. In the Bryant Pond area, however, the four-phase assemblage muscovite—K-feldspar —sillimanite—plagioclase is in practice univariant in relation to P—T conditions. According to Evans and Guidotti, the isograd reaction may have controlled P_{H_2O} . On account of the increased P_{H_2O} , the equilibrium area of muscovite may have been expanded to above the isograd. Muscovite occurs as a fine-scaled retrogressive alteration product in all the zones of the Rantasalmi—Sulkava area south of the K-feldspar—sillimanite isograd. In addition to this muscovite, the K-feldspar—sillimanite zone also contains coarse-grained muscovite that is often discordant to the biotite foliation. Ashworth has demonstrated that the coarse and discordant muscovite is an equilibrium phase in the K-feldspar—sillimanite zone of the Dalradian schists.

The K-feldspar—sillimanite isograd does not indicate the immediate and easily detectable extension of the garnet and cordierite field in the study area (cf. Winkler 1974, pp. 214 and 216). As shown by A'KF triangles and AFM projections, the K-feldspar—cordierite isograd is an indication of reaction:

3) 6 sillimanite + 2 biotite + 9 quartz = 2 K-feldspar + 3 cordierite + 2 $\rm H_2O$ or reaction

4) sillimanite + biotite + 2 quartz = K-feldspar + garnet + H_2O .

Reactions 3 and 4 are continuous and take place over a P—T interval (Reinhardt 1968, Blümel and Schreyer 1976). These authors propose, however, that the appearance of the assemblage quartz—K-feldspar—biotite—garnet—cordierite is due to the discontinuous reaction:
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5) biotite + sillimanite + quartz = K-feldspar + garnet + cordierite.

In the northern part of the K-feldspar—cordierite zone within the Rantasalmi— Sulkava area garnet and cordierite, and biotite and sillimanite may be in mutual contact, whereas in the southern part of the zone it has not been noted that the last mentioned minerals touch each other. Nevertheless, the X_{Mg} ratio in garnet and cordierite is higher in the samples from the southern part of the zone than in those deriving from the northern part. On the basis of what has been said above, reaction 5 may well have been continuous in the study area.

The isograd reactions are, of course, more complicated than as delineated above. Muscovite also takes part in the isograd reactions, since the coarse muscovite disappears south of the K-feldspar—cordierite isograd. It is probable that plagioclase participates in reaction 4 at least. The fact is that garnet contains CaO and often displays plagioclase inclusions. The K-feldspar—cordierite isograd denotes the start of migmatisation.

Hypersthene is an important mineral in a discussion of metamorphic facies. Consequently, the hypersthene isograd has been added to the figure even though hypersthene is not a mineral typical of the metapelites. Although accurate delineation of the hypersthene isograd is hampered by the bulk composition of the rock, it can be concluded that the hypersthene isograd is distinctly to the south of the K-feldspar—cordierite isograd. South of the hypersthene isograd the quartz—horn-blende assemblage becomes a rarity. Thus, there is every justification for presuming that at least one of the isograd reactions is that suggested by Parras (1958, p. 108):

6) hornblende + quartz
$$\longrightarrow$$
 plagioclase + $\xrightarrow{\rightarrow}$ biotite
 \rightarrow orthopyroxene

Accurate localisation of the garnet—cordierite—sillimanite isograd is difficult, because the assemblage cannot be traced on the basis of observations made on outcrops alone. South of the isograd, garnet and cordierite are often in contact with each other and sillimanite occurs as individual grains, even abundantly in places.

The isograd reaction can be formulated as:

7) 3 cordierite = 4 sillimanite + 2 garnet + 5 quartz.

This reaction as well takes place over a P—T interval. As stated by Reinhardt (1968), in progressive metamorphism the garnet—sillimanite—cordierite field on the AFM projection moves towards the AM side. In the light of experimental studies (Currie 1971, Hensen 1971, Hensen and Green 1973), the X_{Mg} ratio of garnet and cordierite increases in the garnet—sillimanite—cordierite—quartz assemblage, particularly when the crystallisation pressure increases.

As shown by AFM projections, the chemical composition of the garnet—cordierite —sillimanite gneisses differs from that of the other zones, in such a manner that it may have had a decisive influence on isograd reaction 7.

PARTITION OF ELEMENTS BETWEEN COEXISTING MINERALS

Mg-Fe partition between biotite and garnet

Ramberg and DeVore (1951) and Kretz (1961) have made theoretical studies on the effect of crystallisation conditions on the partition of iron and magnesium between mineral pairs. The Mg—Fe partition coefficient for the coexisting mineral phases is:

$$K_{D} = \frac{X_{1} (1 - X_{2})}{X_{2} (1 - X_{1})}$$

where K_D is the partition coefficient, X_1 the X_{Mg} ratio for mineral phase 1, and X_2 the X_{Mg} ratio for mineral phase 2. If equilibrium between phases is achieved and the partition is ideal, the numerical value of the partition coefficient depends on crystallisation pressure and temperature alone. Hence, the partition coefficient reflects crystallisation conditions and indicates the degree of equilibrium attained between phases. The partition coefficient may, however, also refer to changes in the compositions of the phases. The numerical values of the partition coefficients have often been noted to depend on crystallisation temperature while the effect of pressure is negligible (Ramberg and DeVore 1951, Kretz 1961, Albee 1965, Perchuk 1969).

In the present study the Mg—Fe partition coefficient between biotite and garnet was determined by substituting X_{Mg} in biotite for X_1 , and X_{Mg} in garnet for X_2 . The partition coefficients taken from other papers were converted into this form. It has been demonstrated by many studies that, provided certain conditions are met, the Mg—Fe partition coefficient between the mineral pair biotite—garnet decreases as the metamorphic grade increases (Frost 1962, Sen and Chakraborty 1968, Albee 1965, Saxena and Hollander 1968, Hietanen 1969, Perchuk 1970, Dougan 1974). Albee (1965) has shown that when the Mn/Mn + Mg + Fe ratio in garnet approaches zero, the numerical value of the partition coefficient changes with the increase in metamorphic grade as follows: in the garnet zone $K_D = 5.0$, in the staurolite zone $K_D = 4.6$, in the kyanite zone $K_D = 4.3$ and in the K-feldspar—sillimanite zone $K_D = 2.7$ —3.3.

Partial electron microprobe analyses of biotite, garnet and cordierite are shown in Tables 6, 7 and 8, and numerical values of the partition coefficients in Table 9. The Mg—Fe partition points and the graphs of the partition functions for the mineral pair biotite—garnet are plotted in Figs. 24, 25 and 26. The compositions of biotites and garnets (represented in Figs. 24—26) were determined from the cores of biotite and garnet grains in contact with each other.

The partition points of the samples from the mica schist zone are close to the line $K_D = 4.1$. The partition points of the samples from the K-feldspar—sillimanite zone are only slightly above this line, the value of the partition coefficient being 3.8. The K_D values of some samples from the K-feldspar—cordierite zone are fairly

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Table 6

FeO- (total Fe as FeO), MgO-, TiO_2 -microprobe determinations and X_{Mg} ratios of biotites in the Rantasalmi—Sulkava metamorphic zones. Anal. Tuula Paasivirta. Symbols: a = biotite in contact with cordierite, b = biotite in contact with garnet, c = biotite in contact with cordierite and garnet

San	nple	FeO	MgO	TiO_{2}	X _{Mg}
Mica s	schist zone				
	11/6	10.1	0.0	1.0	0 427
	IMS	19.1	8.0	1.8	0.427
	4115	20.6	9.5	2.5	0.3/4
	5M5	18.7	9.6	1.2	0.478
b	9MS	22.8	8.9	2.8	0.410
a	10MS	24.7	6.9	2.8	0.332
Ь	10MS	21.4	8.8	2.2	0.423
a	11MS	24.9	6.3	1.8	0.311
b	11MS	22.7	7.7	2.6	0.380
a	12MS	23.1	7.5	2.5	0.367
Ь	13MS	22.2	9.2	2.7	0.425
K-feld.	spar—silliman	ite zone			
	14KS	20.6	87	2.0	0.429
	16KS	19.4	9.1	24	0.455
h	19KS	22.7	7.6	17	0.374
b	20KS	22.0	7.0	1.0	0.365
b	21165	21.6	6.8	1.7	0.305
b	2115	21.0	0.0	1.4	0.339
D	2210	22.0	7.0	1.7	0.375
D	2385	21.5	/.4	1,4	0.380
North	ern part of th	be K-feldspar—co	rdierite zone		
a	27KC	22.5	7.4	4.1	0.370
b	27KC	21.6	7.2	3.3	0.373
a	28KC	20.5	7.2	3.9	0.385
b	28KC	21.8	7.5	2.7	0.380
2	29KC	20.9	7.0	39	0.374
2	30KC	22.5	7.5	3.5	0.373
0	32KC	21.3	7.0	3.1	0.369
h	32KC	22.5	6.7	2.0	0.309
D	32KC	22.5	0.7	2.9	0.347
С	SSKC	23.4	5.0	3.0	0.299
a	34KC	24.2	5.5	3.3	0.288
b	34KC	25.4	5.7	3.0	0.286
Souther	rn part of the	e K-feldspar—cor	dierite zone		
a	35KC	21.8	8.5	5.8	0.410
Ь	35KC	22.3	7.9	4.3	0.387
a	36KC	22.9	7.9	5.3	0.381
b	36KC	22.5	8.4	3.9	0.400
a	37KC	21.2	74	5.2	0.384
b	37KC	19.6	87	33	0.442
3	38KC	19.9	8.1	5.1	0.420
h	38KC	19.7	85	12	0.420
0	30KC	22.3	7.8	5.5	0.435
a L	30KC	20.1	0.2	2.0	0.384
D	JOKC	20.1	0.0	5.9	0.424
D	40KC	18.7	11.7	2.4	0.527
a	41KC	18.7	9.7	3.5	0.480
b	41KC	17.6	11.0	2.1	0.527
a	42KC	18.6	10.1	3.9	0.492
b	42KC	17.7	11.5	2.4	0.537
С	43KC	20.9	8.8	4.2	0.429

San	nple	FeO	MgO	TiO_2	X _{Mg}	
a	44KC	18.0	10.6	5.6	0.512	
b	44KC	18.4	10.0	5.0	0.492	
с	48KC	20.8	8.3	5.7	0.416	
a	49KC	20.5	9.1	5.5	0.442	
b	49KC	19.9	8.8	4.1	0.441	
Garnet	—cordierite—	sillimanite zone				
a	50GC	17.2	14.0	1.9	0.592	
b	50GC	17.6	13.0	2.4	0.568	
a	52GC	19.7	9.3	5.2	0.457	
b	52GC	18.1	11.5	4.3	0.531	
C	52GC	19.9	9.3	4.4	0.454	
а	53GC	19.3	9.2	3.5	0.459	
с	53GC	17.1	11.4	2.9	0.543	
a	54GC	20.7	9.4	4.4	0.447	
b	54GC	18.4	12.1	4.1	0.540	
с	54GC	19.2	9.9	4.1	0.479	
a	55GC	18.8	10.1	3.6	0.489	
b	55GC	17.4	12.4	3.4	0.551	
a	57GC	21.4	9.2	4.3	0.434	
b	57GC	18.2	12.8	2.8	0.556	
с	57GC	18.7	9.7	4.3	0.480	
a	58GC	20.2	7.9	3.8	0.411	
с	58GC	19.4	8.6	3.9	0.441	
a	59GC	21.9	8.5	4.4	0.409	
с	59GC	19.7	9.9	4.1	0.473	
a	61GC	20.7	12.0	3.9	0.508	
b	61GC	20.0	13.5	3.4	0.546	
b	62GC	21.4	9.1	5.5	0.431	
b	63GC	18.2	10.0	5.9	0.495	
b	64GC	19.2	11.0	6.0	0.505	

Table 6 (continued)

high (samples 40KC, 41KC and 42KC). Nevertheless, most of the partition points lie close to line $K_D = 3.1$. The partition points of the samples from the garnet—cordierite—sillimanite zone are greatly scattered. A hypersthene-bearing sample from this zone (sample 62GC) shows the lowest value encountered in the study area. Hence, the numerical value of the Mg—Fe partition coefficient for the mineral pair biotite—garnet decreases with the increase in metamorphic grade. This regular trend in the value of the partition coefficient is, however, not without exceptions, for the highest numerical values of the partition coefficients have been observed in samples of garnet—cordierite—sillimanite gneiss.

Several investigations show that the Mg—Fe partition coefficient for the mineral pair biotite—garnet depends on the MnO or CaO abundance in garnet (Kretz 1959, Frost 1962, Albee 1965, Sen and Chakraborty 1968). Hence, the increase in MnO in garnet, for example, increases the numerical value of the partition coefficient even though the conditions under which the samples crystallised had remained the same. As a rule, the MnO and CaO abundances in the garnet investigated were



Fig. 24. Mg-Fe partition between biotite and garnet in the mica schist zone (squares) and in the K-feldspar-sillimanite zone (triangles).



Fig. 25. Mg-Fe partition between biotite and garnet in the northern part (open circles) and in the southern part (grey circles) of the K-feldspar-cordierite zone. Grey circles with inclined crosses refer to samples 40KC, 41KC and 42KC.



Fig. 26. Mg-Fe partition between biotite and garnet in the garnet-cordierite-sillimanite zone. Dark circles with crosses refer to the hypersthene-bearing samples.

too low to have a marked effect on the values of the partition coefficients. Particularly the CaO and MnO contents in the garnet of the garnet—cordierite—sillimanite gneisses are very low and the variations insignificant.

Dallmayer (1974 a and b) has observed that K_D decreases as the TiO₂ in biotite increases. The effect of titanium in biotite on the partition coefficient is not, however, completely clear. In progressive metamorphism the abundance of biotite decreases, the TiO₂ tenor in biotite then increases provided that the bulk chemical composition varies slightly and biotite is practically the only titaniferous phase. Likewise, Miyashiro and Shido (1973) have suggested that the MnO content in garnet decreases with an increase in metamorphic grade.

Within the study area the breakdown reaction of biotite starts at the K-feldspar cordierite isograd. Accordingly, the abundance of titanium in biotite is higher in samples from the northern part of the K-feldspar—cordierite zone than in samples from the mica schist and K-feldspar—sillimanite zones. This all requires a constant chemical composition for the metapelites. The advance of the breakdown reaction of biotite in the K-feldspar—cordierite zone is also reflected in the titanium content of biotite, which is higher in the southern than in the northern part of the zone. Hence, the titanium content of biotite is dependent not only on the bulk composition but also on the metamorphic grade. The excess aluminium in the garnet—cordierite sillimanite gneisses is so high that the breakdown reactions of biotite in these gneisses is low. Nevertheless, the titanium tenor may be comparatively low in biotite and the increase in metamorphic grade is not manifest in the TiO₂ content in the biotite of the garnet—cordierite—sillimanite gneisses.

Fig. 27 shows the TiO₂ content in the biotite of the metapelites in the metamorphic zones within the study area. The garnet-bearing samples from the K-feldspar—sillimanite zone and the mica schist zone are not shown in the figure, neither are the hypersthene-bearing samples from the garnet—cordierite—sillimanite zone, because the chemical composition of these samples differs from that of the other metapelites. In the garnet—cordierite—sillimanite gneisses, the X_{Mg} ratio in biotite may vary, even in the contact of the same garnet grain (Table 6, samples 52GC, 54GC and 57GC). In the samples from the north-eastern Reading Prong area, Dallmayer (1974 a) believes that, close to the contact with garnet, the X_{Mg} ratio in biotite has increased and the TiO₂ content decreased owing to retrogressive metamorphism.

The X_{Mg} ratio in the centre of the garnets in the garnet—cordierite—sillimanite gneisses is fairly homogeneous (Fig. 28), whereas in the samples from the garnet cordierite—sillimanite zone the X_{Mg} ratio is higher in the centre of all the garnets analysed than in the margin (Table 7 and Fig. 28). Such zoning in garnet has been attributed to retrogressive metamorphism (Hess 1971, Grant and Weiblen 1971, Dallmayer 1974 a). Nagaytsev (1973) has noted that the X_{Mg} ratio in garnet decreases from the centre towards the margin whenever garnet is replaced by cordierite or biotite.

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Fig. 27. TiO_2 contents of analysed biotites in the Rantasalmi–Sulkava metamorphic zones. Symbols: open dots — biotite in contact with cordierite, solid dots — biotite in contact with garnet or with garnet and cordierite.

In intensely migmatised garnet—cordierite—sillimanite gneisses the abundance of water-bearing alteration products is high. Further indications of retrogressive metamorphism are retrogressive zoning in garnet and the low TiO_2 content in biotite. Schneider (1975) has experimentally verified a retrogressive reaction associated with anatexis in which garnet is replaced by biotite. Owing to this reaction, the X_{Mg} atomic ratio decreases in the margin of garnet. Grant and Weiblen have demonstrated that retrogressive zoning in garnet is due to migmatisation.

The facts presented above indicate that the deviation and comparatively high values shown by the numerical values of the Mg—Fe partition coefficients for the mineral pair biotite—garnet in the garnet—cordierite—sillimanite gneisses are caused by retrogressive metamorphism. The X_{Mg} ratio in biotite may vary even in a contact with the same garnet, which indicates that equilibrium was not attained during the retrogressive stage. The disequilibrium is further indicated by the fairly large variation in the TiO₂ content of biotite, since in the quartz—K-feldspar—sillimanite—garnet assemblage this content is controlled by the crystallisation conditions (Kretz 1964). Most of the feldspars and cordierite in sample 50GC from the garnet—cordierite—sillimanite zone have altered into water-bearing minerals. Exceptionally intense retrogressive metamorphism is further indicated by the high numerical value, 8.1,



Fig. 28. X_{Mg} zoning of garnet. Symbols: open dots — sample 28KC from the northern part of the K-feldspar—cordierite zone, solid dots — sample 54GC from the garnet—cordierite—sillimanite zone.

of the partition coefficient for this sample, a result of biotitisation, manifested by the low X_{Mg} ratio in the centre of the garnet. The retrogressive zoning in garnet is not, however, unequivocally associated with abnormal values of the partition coefficient. Also some garnets in the K-feldspar—cordierite zone exhibit retrogressive zoning (Fig. 28), which is not reflected in the values of the partition coefficient as it is in the garnet—cordierite—sillimanite samples. The high values of the partition coefficients in these gneisses are presumably largely due to the biotite that crystallised during retrogressive metamorphism.

The numerical values of the partition coefficients are fairly high for the samples (40KC, 41KC and 42KC) taken from an outcrop in the southern part of the K-feldspar -cordierite zone. In these samples X_{Mg} ratio in biotite in contact with garnet is distinctly higher than it is in the biotite in contact with cordierite, thus suggesting that the value of the partition coefficient might have changed during the retrogressive stage. Even so, the garnet in these samples is not zonal. Chemical analyses show that the X_{Mg} ratio of the bulk composition is comparatively high and the Al₂O₃ excess over alkalis and calcium low. The samples do not contain K-feldspar, but biotite is fairly abundant (20 to 25 percent) and the TiO₂ content of biotite low. Consequently, the breakdown reactions of biotite were not as effective in these samples as they often were in samples from the southern part of the K- feldspar -cordierite zone. The high numerical values of the partition coefficients may thus be attributed to the bulk composition of the samples. Sample 43KC, containing K-feldspar, was extracted from the same site as the forementioned samples. Its chemical composition corresponds to the other samples from the K-feldsparcordierite zone. Likewise, the numerical value of the Mg-Fe partition coefficient for the mineral pair biotite-garnet in sample 43KC approaches those of the other samples from the southern part of the K-feldspar-cordierite zone as does the TiO2 content of biotite.

Table 7

FeO- (total Fe as FeO), MgO-, MnO-, CaO-microprobe determinations and X_{Mg} ratios of garnets in the Rantasalmi–Sulkava metamorphic zones. Anal. Tuula Paasivirta. Symbols: a = garnet in contact with biotite, b = garnet in contact with cordierite, c = garnet in contact with biotite and cordierite

5	Sample	FeO	MgO	MnO	CaO	${\rm x}_{\rm Mg}$	
Mic	ca schist zon	,					
1,111	a summer zone	24 7	2 5	0.5	1.0	0.145	
а	9M5	30.7	3.5	0.5	1.9	0.145	
а	IUMS	36.1	4.0	0.5	1.8	0.165	
a	11MS	35.9	2.8	0.4	1.7	0.122	
a	13MS	36.7	3.5	0.5	1.8	0.145	
K-f	eldspar—silli	manite zone					
a	19KS	35.4	3.2	0.3	2.0	0.139	
a	20KS	33.6	2.8	2.5	1.3	0.133	
a	21KS	34.3	2.8	2.1	1.2	0.127	
a	22KS	36.2	3.1	0.3	2.0	0.132	
a	23KS	36.5	3.2	1.0	1.6	0.135	
No	rthern part o	of the K-feldstu	ar—sillimanite	zone			
1.07	OTVC	24.0	2.4	1.0	1 1	0 1 4 9	
а	2/KC	34.9	3.4	1.0	1.1	0.148	
а	28KC	35.7	3.9	1.0	1.0	0.1/1	
а	32KC	34.5	3.4	1.1	0.8	0.149	
С	33KC	38.3	2.7	1.0	0.8	0.112	
С	34KC	37.4	2.7	0.8	0.7	0.114	
Sout	thern part of	the K-feldspan	r—sillimanite ze	one			
а	35KC	35.7	4.0	0.9	1.0	0.166	
a	36KC	36.1	4.2	0.8	1.1	0.172	
a	37KC	36.0	5.0	0.7	1.1	0.198	
2	38KC	35.3	5.1	0.7	0.9	0.205	
2	39KC	35.5	4.9	0.6	1.0	0.197	
h	39KC	36.9	4.8	0.7	1.0	0.188	
0	40KC	34 1	4 7	0.6	1.1	0.197	
a	AIKC	35.6	5.0	0.7	0.0	0.200	
a	4INC	33.0	5.0	0.7	1.0	0.200	
a	42KC	34.9	5.0	0.8	1.0	0.203	
С	43KC	35.7	5.1	0.7	1.2	0.203	
a	44KC	35.0	5.0	0.5	0.9	0.225	
С	48KC	36.2	5.1	0.5	1.0	0.201	
а	49KC	36.3	4.9	0.5	0.8	0.194	
Gar	net—cordierii	te—sillimanite	zone				
С	50GC						
	core	38.4	3.5	0.2	0.8	0.140	
	edge	37.1	2.9	0.3	0.9	0.122	
С	52GC						
	core	34.5	5.0	0.3	0.9	0.205	
	edge	37.1	3.4	0.4	0.9	0.140	
с	53GC						
-	core	36.2	5.0	0.2	1.0	0.198	
	edge	36.5	3.3	0.4	0.9	0.139	
C	5466						
C	COTE	35 2	52	0.3	0.9	0.208	
	edge	38 3	3.8	0.4	1.0	0.150	
	euge	50.5	5.0	0.4	1.0	0.150	

				,			
	Sample	FeO	MgO	MnO	CaO	X _{Mg}	
c	55GC core edge	35.9 37.6	4.7 3.8	0.3 0.4	0.8 0.8	0.189 0.153	
с	57GC core edge	35.2 35.8	4.8 4.0	0.2 0.2	0.6 0.7	0.196 0.166	
с	58GC core edge	34.2 35.1	4.9 3.9	0.1 0.2	0.8 0.8	0.203 0.165	
с	59GC core edge	36.2 37.0	4.9 3.8	0.3 0.4	$1.0\\1.0$	0.194 0.155	
с	61GC core edge	34.9 36.8	6.4 5.0	0.3 0.3	1.0 0.9	0.246 0.195	
a	62GC core edge	32.4 33.3	6.2 4.4	0.6 0.5	3.1 2.0	0.254 0.191	
а	63GC core edge	33.1 35.6	5.7 5.0	1.0 1.1	2.5 2.1	0.235 0.200	
а	64GC core edge	34.0 32.4	6.3 5.7	0.5 0.6	1.9 1.7	0.248 0.239	

Table 7 (continued)

Mg-Fe partition between cordierite and biotite

Information on the effect of pressure and temperature on the Mg—Fe partition coefficient for the mineral pair cordierite—biotite is contradictory. Some workers report that the numerical value of the partition coefficient does not vary with meta-morphic grade (Gorbatschew 1968, Saxena and Hollander 1969, Gable and Sims 1969). Others (Dallmayer and Dodd 1971, and Kamineni 1975) propose that the partition coefficient is affected by changes in the crystallisation conditions. Kamineni has noted that the Mg—Fe partition coefficient for the mineral pair cordierite—biotite decreases in the metagreywackes and metapelites outwards from the contact of the Sparrow Lake granite, in other words, the numerical value of the partition coefficient decrease in the decrease in contact effect. Dougan (1974) has demonstrated that the decrease in the partition coefficient is due to the increase in metamorphic grade. Several authors have reported values for the Mg—Fe partition

TL.	1.1	1.00	0
1a	D.	le	0

FeO- (total Fe as FeO), MgO-microprobe determinations and X_{Mg} ratios of cordierites in the Rantasalmi—Sulkava metamorphic zones. Anal. Tuula Paasivirta. Symbols: a = cordierite in contact with biotite and b = cordierite in contact with biotite and garnet

S	ample	FeO	MgO	X _{Mg}
Mica	schist zone	21.07		
a	10MS	8.8	5.8	0.498
a	11MS	9.1	6.1	0.544
a	12MS	8.1	7.0	0.606
Nori	thern part of the	K-feldspar—sillimanite	zone	
а	27KC	9.0	6.8	0.574
2	28KC	8.5	7.1	0.598
9	29KC	8 5	63	0.569
a 0	30KC	8.9	6.5	0.566
a	32KC	83	73	0.500
a 1-	32KC	0.0	1.9	0.011
	34VC	10.9	4.0	0.467
b	34KC	10.8	5.5	0.40/
South	bern part of the k	K-feldspar—sillimanite	zone	
a	35KC	7.4	7.3	0.637
a	36KC	8.5	7.6	0.614
a	37KC	8.1	7.4	0.620
a	38KC	8.0	7.9	0.638
h	39KC	8.2	7.2	0.626
2	41KC	6.7	8.3	0.688
2	42KC	8.0	8.8	0.692
h	43KC	8.0	8.0	0.641
2	44KC	7.0	9 1	0.699
a h	48KC	8.2	81	0.638
	AOKC	8.0	8.4	0.652
a	+7KC	0.0	0.4	0.032
Garn	et—cordierite—sill	limanite zone		
b	50GC	9.0	6.8	0.574
b	52GC	8.7	6.8	0.582
b	53GC	8.3	6.8	0.594
b	54GC	8.6	6.9	0.588
ĥ	55GC	8.3	6.4	0.579
h	5760	9.6	7 3	0.575
L 1	5866	0.1	6.5	0.575
1	5000	0.2	6.4	0.500
0	6100	7.6	0.4	0.554
		()	N N	

coefficient for the mineral pair cordierite—biotite that are close to 1.86 (Kamineni 1975). The partition coefficient was calculated by substituting X_{Mg} in cordierite for X_1 , and X_{Mg} in biotite for X_2 (formula on p. 37).

The Mg—Fe partition points for the mineral pair cordierite—biotite and the graphs of the partition function are plotted in Fig. 29. The compositions of cordierites and biotites (represented in Fig. 29) were determined from the cores of cordierite and biotite in contact with each other. Most of the partition points lie close to the



Fig. 29. Mg—Fe partition between cordierite and biotite. Symbols are as in Figs. 24, 25 and 26.

line with $K_D = 2.5$. The K_D values of the samples 41KC and 42KC from the southern part of the K-feldspar—cordierite zone do not differ from those of the other samples in the same way as do the Mg—Fe partition coefficients for the mineral pair biotite—garnet.

The lowest values of the partition coefficient were noted in samples from the garnet-cordierite-sillimanite zone. As stated earlier, the biotite in contact with garnet in the garnet-cordierite-sillimanite gneisses is retrogressive in origin. For this reason, the low K_D values of the pair cordierite-biotite in samples from the garnet-cordierite-sillimanite zone may reflect retrogressive metamorphism. The cordierite in the garnet-cordierite-sillimanite gneisses does not exhibit zoning similar to that observed in garnet of these gneisses. The cordierite has, however, often undergone pinitisation. The concentration of TiO₂ in the biotite in contact with cordierite is higher than in biotite adjacent to garnet. Nevertheless, the TiO_2 tenor in biotite is lower that it is in the biotite in contact with cordierite in samples from the southern part of the K-feldspar-cordierite zone (Fig. 27). In the retrogressive sample 50GC, from the garnet-cordierite-sillimanite zone, the numerical value of the partition coefficient (0.9) was the lowest recorded. On the basis of the observations reported above, it is evident that the retrogressive metamorphism in the garnet-cordierite-sillimanite zone also affected the Mg-Fe partition coefficient of the mineral pair cordierite-garnet.

Table 9

Mg—Fe partition coefficients between coexisting biotite and garnet, cordierite and biotite, and cordierite and garnet in the Rantasalmi—Sulkava metamorphic zones, as well as estimated P—T conditions of samples. Symbols used are at the end of Table

Sample	Bio-Gar	Cord-Bio	Cord-	Gar		°C	-	-	kb	
	К _D	К _D	К _D	T ₁	T_2	T ₃	T4	Τ ₅	Pı	P ₂
Mica schist	zone									
9MS	4 1			633	650					
10MS	3.7	21		647	683					
11 MS	1.1	2.4		624	620					
12MS	4.4	2.0		024	629					
12115	12	2.1		(00	(20)					
13115	4.3			628	638					
mean	4.1	2.6		633	650					
K-feldspar–	–sillimanite zone									
19KS	3.7			647	683					
20KS	37			647	683					
2116	3.0			640	666					
2115	3.9			040	000					
2210	5.9			640	666					
23K3	3.9			640	666					
mean	3.8			643	673					
Jorthern p	art of the K-feld	dspar—cordierit	e zone							
27KC	3.4	2.3		660	714					
28KC	3.0	2.4		678	758					
20KC	5.0	2.4		0/0	130					
29KC		2.2								
SOKC	2.0	2.2			-					
32KC	3.0	2.7		678	758					
33KC	3.4	2.0	6.9	660	714	691				
34KC	3.1	2.2	6.8	673	747	695				
mean	3.2	2.3	6.9	670	738	693				
outhern par	rt of the K-felds	spar—cordierite	zone							
35KC	32	2.5	6	660	735					
36KC	32	2.6		660	735					
37KC	3.2	2.0		640	735					
38KC	3.0	2.0		609	750					
JOKC	3.0	2.4	7.0	6/8	/58	(=0				
JAKC	3.0	2.1	1.2	6/8	758	678				
44KC	3.4	2.2		660	714					
48KC	2.8	2.5	7.0	688	786	684				
49KC	3.3	2.4		664	724					
43KC	2.9	2.4	7.0	684	773	684				
mean	3.1		7.1	673	746	682				
40KC	4.5									
41KC	4.5	2.4								
42KC	4 5	23								
12110	T.J									
mean		2.5								

Sample	Bio	Gar	Cord Bio	Cord—Gar			°C			kb	
oumple	K _D	-Gai	K _D	K _D	T ₁	T_2	T ₃	T_4	Тъ	P1	P ₂
Garnet—cor	dierite—sil.	limanit	e zone								
52GC	3.2		1.7	5.4			782	692	760	5.7	8.0
52GC	4.4										
53GC	4.8		1.7	5.9			750	709	720	5.7	8.0
54GC	3.5		1.8	5.4			782	692	760	5.7	8.0
54GC	4.5										
55GC	5.3		1.4	5.9			750	709	720	5.6	8.0
57GC	3.8		1.8	5.6			770	698	780	5.6	7.5
57GC	5.2										
58GC	3.1		1.8	5.0			816	676	800	5.6	7.5
59GC	3.7		1.8	5.1			808	680	780	5.6	7.5
61GC	3.7		2.0	6.3			724				
mean		1.8	5.6				772	694	760	5.6	7.8
50GC	8.1	0.9	8.3								
Hypersthene	-bearing .sa	mples									
62GC	2.2				727	894					
63GC	3.2				669	735					
64GC	3.2				669	735					

Table 9 (continued)

The temperature and pressure determinations in Table 9 were based on the following:

 $T_1 = K_D$ ^{Bio}—Gar</sup> (Perchuk 1970) $T_2 = K_D$ ^{Bio}—Gar</sup> Thompson 1976)

 $T_3 = K_D$ (Thompson 1976)

 T_4 and $P_1 =$ according to data by Currie (1971)

 T_5 and P_2 = according to data by Hensen and Green (1973)

Mg-Fe partition between cordierite and garnet

It has been demonstrated that the Mg—Fe partition coefficient for the mineral pair cordierite—garnet decreases with the increase in metamorphic grade (Gable and Sims 1969, Perchuk 1969, A. B. Thompson 1976). The partition coefficient was determined by substituting X_{Mg} in cordierite for X_1 , and X_{Mg} in garnet for X_2 (formula on p. 37). Gorbatschev (1968) points out that the partition coefficient reflects crystallisation conditions if the variation in X_{Mg} of garnet is taken into consideration. The numerical value of the partition coefficient tends to decrease with the increase in metamorphic grade, or in X_{Mg} of garnet. On the other hand, Saxena and Hollander (1969) have noted that the partition of magnesium and iron between cordierite and



Fig. 30. Mg-Fe partition between cordierite and garnet. Symbols are as in Figs. 25 and 26.



Fig. 31. X_{Mg} content of garnet plotted against Mg—Fe partition coefficient for the pair biotite—garnet. Symbols: crosses — samples studied by Saxena and Hollander (1969), open circles — the northern part and grey circles the southern part of the K-feldspar—cordierite zone, dark circles — the garnet—cordierite—sillimanite zone.

garnet is regular in the epidote—amphibolite facies, whereas in the amphibolite and granulite facies the partition coefficient decreases with the increase in X_{Mg} of garnet. The results obtained by Saxena and Hollander indicate that the partition coefficients cannot be used to estimate crystallisation conditions. Currie (1971) and Hensen and Green (1971, 1972 and 1973) have studied experimentally the partition of magnesium and iron between cordierite and garnet. Their studies show that the partition coefficient depends mainly on temperature. Currie reports that the increase

in the partition coefficient coincides with that of the crystallisation temperature. In contrast, the results by Hensen and Green suggest that the partition coefficient may decrease with rising temperature.

The Mg—Fe partition points of the samples analysed for the mineral pair cordierite —garnet and the graphs of the partition function are plotted in Fig. 30. The partition points were determined only for those samples in which garnet and cordierite were in contact with each other. The partition points of the samples from the northern and southern parts of the K-feldspar—cordierite zone lie close to line $K_D = 7.0$. As a rule, the partition coefficients of the samples from the garnet—cordierite sillimanite zone are lower than those obtained from the K-feldspar—cordierite zone. Most of the partition points are close to line $K_D = 5.6$. The numerical value of the partition coefficient for the retrogressive sample 50GC is 8.3, which is considerably higher than the value from the other samples.

Fig. 31 illustrates the relationship between the Mg—Fe partition coefficient of the mineral pair cordierite—garnet and the X_{Mg} of garnet in the samples of the present study and in the samples of the amphibolite and granulite facies studied by Saxena and Hollander. The values of the partition coefficient for the samples from the southern part of the K-feldspar—cordierite zone and the garnet—cordierite—sillimanite zone approach the values reported by Saxena and Hollander, whenever X_{Mg} ratios of garnet are equal. However, on the basis of the samples of the present study, it cannot be concluded that the partition coefficient is dependent only on X_{Mg} of garnet.

Partition of cobalt and vanadium between biotite and garnet

The partition of trace elements between a pair of phases has been theoretically studied by McIntire (1963). Studies on trace element partitions relie largely on the application of the Berthelot—Nernst partition law: $\frac{X_1}{X_2} = K_D$. In the present study, X_1 denotes the abundance of the trace element in biotite, and X_2 the corresponding abundance in garnet. If the partition of a trace element is ideal and the phases are in equilibrium, the partition coefficient (K_D) refers to certain crystallisation conditions. Several studies have shown that in the mineral pairs of the metamorphic rocks the partition of the trace elements obeys the partition law (Kretz 1959, Moxham 1965, Doe and Tilling 1967, Leelandam 1970, Annersten and Ekström 1971, Korsman 1975). Dagelayskiy and Krylova (1973) have worked out the effect of metamorphic grade and crystallisation temperature on the partition coefficient. In Finland, Häkli (1968) has studied the differentiation of the Parikkala intrusive by determining the model crystallisation temperatures over the intrusive on the basis of the partition of nickel between olivine and augite.

Table 10

Emission spectrometric determinations of cobalt and vanadium (in parts per million) in biotite and garnet from the Rantasalmi—Sulkava area and partition coefficients between biotite and garnet. * Anal. Ringa Danielsson and ** Anal. A. Puisto

	Sample	Со	V	К _D (Со)	K _D (V)
K-fe	ldspar—sillimanite zone				
	10KS**				
	biotite	53	350	2.3	4.8
	garnet	23	73		
	20KS**				
	biotite	67	430	2.9	6.7
	garnet	23	64		
	23KS**				
	biotite	59	410	2.5	5.5
	garnet	24	74		
				mean 2.6	5.7
Nor	thern part of the K-feld	dspar—cordierite z	zone		
	27KC*				
	biotite	58	650	2.9	11.0
	garnet	20	59	2.7	11.0
	2012 C*				
	biotite	56	660	2.0	12.5
	garnet	19	53	2.7	12.5
	2012 C**				
	52KC**	54	780	2.0	7.0
	garnet	27	99	2.0	1.9
	2212.0**				
	33KC**	53	830	2.0	10.2
	garnet	27	81	2.0	10.2
	8417 C**	_,	01		
	34KC**	50	710	2.5	0.0
	garnet	24	80	2.5	8.9
	Surfect treater	2.1	00		
				mean 2.5	10.1
South	bern part of the K-felds	par—cordierite zo	one		
	35KC**				
	biotite	65	880	2.2	10.7
	garnet	29	82		
	36KC**				
	biotite	68	830	2.2	10.0
	garnet	31	83		
	37KC*				
	biotite	59	820	2.6	10.9
	garnet	25	15		
	39KC*	10	<i></i>		
	biotite	60	840	2.1	11.4
	garnet	28	/4		

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		/	
Со	v	K _D (Co)	K _D (V)
 71	660	2.5	8.4
 28	79		
 68	750	2.5	9.1
 27	82		
 77	930	2.3	10.1
 33	92		
 42	1 200	1.6	8.0
 27	150		
 63	1 300	2.1	8.7
 30	150		
			mean 9.7
 60	820	2.4	4.6
 25	180		
		mean 2.3	
	Co 71 28 27 77 33 42 27 33 42 27 33 63 30 60 25	Co V 71 660 28 79 28 79 28 79 27 82 27 82 33 92 27 1200 27 150 30 150 27 150 27 150 27 150 27 150 27 150 27 150 27 150 30 150 30 150 30 150	Co V $K_D(Co)$ 71 660 2.5 28 79 2.5 68 750 2.5 27 82 2.5 27 82 2.3 33 92 2.3 42 1 200 1.6 63 1 300 2.1 63 1 300 2.1 60 820 2.4 25 180 2.3

Table 10 (continued)

The cobalt and vanadium abundances in biotites and garnets and the K_D values are given on Table 10. Fig. 32 plots the partition points of cobalt between biotite and garnet in samples from the K-feldspar—sillimanite zone and the K-feldspar cordierite zone. Korsman (1975) obtained 1.7 as a value of the cobalt partition coefficient for the mineral pair biotite—garnet in samples of the granulite facies. The Sulkava sampling area referred to in the latter paper is the same as the garnet cordierite—sillimanite zone of the present paper.

Also shown in Fig. 32 is a graph of the partition function ($K_D = 1.7$) for the granulite facies samples. Most of the partition points lie close to the line $K_D = 2.3$, and practically all of them (excluding sample 48KC) plot below the graph of the granulite facies samples. The highest values of the partition coefficient were observed in samples from the northern part of the K-feldspar—cordierite zone and in those from the K-feldspar—sillimanite zone. These samples, however, also exhibit a marked deviation in the values of the partition coefficient. The partition of cobalt between biotite and garnet in the samples from the southern part of the K-feldspar—cordierite zone has reached equilibrium rather well. The samples from Sulkava, that is from the garnet—cordierite—sillimanite zone, contain biotite of retrogressive

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Fig. 32. Partition of cobalt between biotite and garnet in the Rantasalmi—Sulkava area. Symbols are as in Figs. 24, 25 and 26. Scale is logarithmic.



Fig. 33. Partition diagram showing a compilation of data collected earlier on the partition of cobalt between biotite and garnet. Symbols: solid triangles — the Lincoln area (Albee *et al.* 1965), open triangles — the Idaho area (Hietanen 1965), open inclined crosses — the Colton area (Engel and Engel 1960), solid crosses — the Westport area (Wynne-Edwards and Hay 1963). Scale is logarithmic.

origin. Thus, it is not feasible to draw conclusions concerning the crystallisation conditions referred to by the value 1.7 of the partition coefficient of the granulite samples.

Shown in Fig. 33 is the partition of cobalt between biotite and garnet as reported by Albee et al. (1965), Hietanen (1969), Wynne-Edwards and Hay (1963) and Engel and Engel (1960). The samples from the Lincoln area (Albee et al.) contain the mineral assemblage: garnet-biotite-chlorite-muscovite-albite-quartzilmenite, the samples from the Idaho area (Hietanen) derive from the garnetstaurolite, staurolite-kyanite, kyanite, kyanite-sillimanite and muscovite-sillimanite zones. The samples from Colton (Engel and Engel) and Westport (Wynne-Edwards and Hay) originate from the two-pyroxene zone. Hence, the metamorphic grade increases by areas as follows: Lincoln, Idaho, and Colton and Westport. The partition points of cobalt for the samples from Lincoln lie distinctly below the graph of the partition function ($K_D = 2.3$) for the samples from the K-feldspar cordierite zone. Some of the partition coefficients of the samples studied by Hietanen are compatible with those of the Lincoln samples. The majority of the partition points for the Idaho samples, however, are close to line $K_D = 2.3$. The samples studied by Hietanen crystallised under variable conditions. Nevertheless, the deviation in the values of the partition coefficient cannot be explained on the basis of a change in metamorphic grade. The partition points of the samples from Colton and Westport are not far from line $K_D = 2.3$ or $K_D = 1.7$.

From the above, it can be concluded that the partition coefficient of cobalt for the mineral pair biotite—garnet does not unambiguously change with the increase in metamorphic grade, although some observations indicate that, in such cases, the partition coefficient may decrease. Even so, the values of the partition coefficient for the samples that underwent metamorphism in the upper part of the amphibolite facies or under conditions of the granulite facies are close to each other.

Fig. 34 plots the vanadium partition points between biotite and garnet in the samples from the K-feldspar—sillimanite zone and the K-feldspar—cordierite zone as well as in the hypersthene-bearing samples from the garnet—cordierite—sillimanite zone studied by Korsman (1975). According to him, the partition coefficient for vanadium between biotite and garnet is 9.8 in some samples of the granulite facies, excluding those containing hypersthene. The majority of the partition points of the samples from the K-feldspar—cordierite zone lie close to the same line ($K_D = 9.8$) as do those of the granulite facies samples, whereas the partition coefficient of the samples from the K-feldspar—sillimanite gneiss zone average 5.7. Also close to the graph of the partition function for the K-feldspar—sillimanite gneisses are the partition points of the hypersthene-bearing samples from the K-feldspar—sillimanite gneiss zone as well as the partition point of sample 38KC from the southern part of the K-feldspar—cordierite zone. The garnet in the K-feldspar—sillimanite gneisses and in the hypersthene-bearing samples exhibits higher CaO/CaO+MgO+MnO+FeO (mol%) than the garnet in the K-feldspar—cordierite gneisses or in



Fig. 34. Partition of vanadium between biotite and garnet in the Rantasalmi—Sulkava area. Symbols are as in Figs. 24, 25 and 26. Scale is logarithmic.



Fig. 35. Partition of vanadium between biotite and garnet. Solid ellipses represent Grenville samples (Kretz 1959) having Ca/Fe + Mn + Mg + Ca ratio of garnet smaller than 0.11 and open ellipses Grenville samples having Ca/Fe + Mn + Mg + Ca ratio greater than 0.11. Other symbols are as in Fig. 33. Scale is logarithmic.



Fig. 36. Partition coefficient of vanadium between biotite and garnet plotted against Ca content of garnet. Symbols: solid triangles — the Lincoln area, ellipses — the Grenville area, open triangles — the K-feldspar—sillimanite zone, open circles — the northern part and grey circles — the southern part of the K-feldspar—cordierite zone, dark circles with crosses — the hypersthene-bearing samples of the garnet—cordierite—sillimanite zone (Korsman 1975).

the garnet—cordierite—sillimanite gneisses. In contrast, the CaO content in garnet in sample 38KC is low. Kretz (1959) has stated that the increase in the value of the vanadium partition coefficient for the mineral pair biotite—garnet coincides with the decrease in the CaO/CaO + MgO + FeO + MnO ratio of garnet.

Fig. 35 shows the vanadium partition points between biotite and garnet in the Grenville gneiss studied by Kretz (1959) and in samples deriving from the same areas as those in Fig. 33. The samples from Grenville were metamorphosed under conditions of the amphibolite facies, but close to those of the granulite facies. The vanadium partition coefficients of the Lincoln samples, and most of those from Idaho, are lower than the partition coefficients exhibited by the samples from the K-feldspar—cordierite zone. The samples studied by Hietanen show, however, a rather large deviation in the partition coefficient, which does not seem to depend on metamorphic grade. Some of the partition points of the Westport samples lie close to line $K_D = 9.8$ but some plot below the line. All the partition coefficients exhibit significant deviation. Even so, the partition coefficients of the samples in which the CaO/CaO + MgO + FeO + MnO ratio of garnet is less than 0.11 are closest to line $K_D = 9.8$.

Fig. 36 illustrates the relationship between the vanadium partition coefficients for the mineral pair biotite—garnet and the CaO/CaO + MgO + MnO ratio in

garnet. In the samples studied by Kretz, Albee and the present author, the partition coefficients decrease with the increase in the above ratio. The samples investigated by Hietanen, Engel and Engel, and Wynne-Edwards and Hay do not show that the CaO content in garnet has any influence on the partition coefficient.

MIGMATISATION

Sederholm (1907, p. 110) was the first to define migmatites as follows: »For the gneisses here in question, characteristic of which are two elements of different genetic value, one, a schistose sediment or foliated eruptive, the other, either formed by the resolution of material like the first or by an injection from without, the author proposes the name of migmatites». Mehnert (1968, pp. 229–230) defined migmatites thus: »A migmatite is a megascopically composite rock consisting of two or more petrographically different parts, one of which is the country rock generally in a more or less metamorphic stage, the other is of pegmatitic, aplitic, granitic, or generally plutonitic appearance».

The most difficult part in studies on migmatites is to delineate the history of the granitic portion. In this respect, special importance is laid on establishing the equilibrium phases of the system NaAlSi₃O₈—KAlSi₃O₈—SiO₂—H₂O at given pressure and temperature (Tuttle and Bowen 1958), as well as on the data obtained from melting tests of clays, greywackes and gneisses (Winkler 1957, Winkler and von Platen 1958, 1960, 1961, von Platen 1965). The melting experiments showed that greywackes and clays yield granitic melt already under conditions of the amphibolite facies, provided that the P_{H_2O} is high enough. If the normative Ab: An ratio in greywacke is reduced, then the initial melting temperature rises and the composition of the melt first formed becomes more orthoclase-bearing, that is, it moves in the Or—Ab—Q field towards the Q—Or side. The decomposition reactions of muscovite and biotite have a bearing on anatexis. These reactions produce K-feldspar, which may later also melt. Melting experiments show that progressive metamorphism may generate migmatites already at the K-feldspar-sillimanite isograd. On the other hand, it has been noted in many areas that the metamorphic grade rises towards the migmatic zone (Read 1957).

The anatectic model derived from the experiments is fairly readily applicable, provided there is enough water and $P_{H_2O} = P_{fluid} = P_{total}$. Several students, however, claim that anatexis takes place under conditions depleted in water and in which $P_{H_2O} < P_{fluid}$ and $P_{H_2O} < P_{total}$ (Burnham 1967, Wyllie 1971, Fyfe 1970). Maaløe and Wyllie (1975) have made melting experiments on granite under water-saturated and -depleted conditions. If water is present, no matter how little, then under undersaturated conditions the solidus temperature of granite remains the same as it does, when water is in excess. In contrast, if the system is undersaturated, the liquidus



Fig. 37. Normative (CIPW norms) Q: Ab: Or ratios of analysed granites in the Rantasalmi—Sulkava area. Symbols: solid dots — potassium granites, open dots — tourmaline-bearing granites. Solid dots fall close to the field for the microcline granites of southern Finland (outlined area), compiled by Simonen (1960).

temperature rises conspicuously. Hence, the amount of melt may remain small if not enough water is available. The granitic melt generated under undersaturated conditions may intrude to higher level than the water-saturated granite (Cann 1970).

The fluid pressure formed during metamorphism is mainly composed of partial pressure of water and carbon dioxide (Sobolev and Dobretsov 1972). Experiments indicate that the solidus temperature of granite rises significantly if, in the fluid phase, the molar proportion of carbon dioxide increases at the cost of water (Fig. 43) (Wyllie and Tuttle 1959, Novgorodov and Shkodzinskiy 1974). In metamorphic areas, the most intense migmatisation has often taken place under conditions of the amphibolite facies. On the other hand, no signs of melting are seen in some granulite areas even though the rocks exhibit leucocratic mineral composition (Khlestov 1973 a, pp. 37—39). This may be because the samples of the granulite facies.

The composition of the migmatising granite may also pose problems. The migmatising granite in the extensive migmatite areas of southern Finland is frequently microcline granite with a high potassium content and normative Or: Ab ratio (Simonen 1960). These migmatising potassium granites are so vast that, in all likelihood, the areas received alkalies from an external source. Härme (1965, 1966) has demonstrated that there are some areas where the migmatising granite does not derive

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Fig. 38. Garnet—cordierite—sillimanite melanosome and potassium granitic segregations. The amount of potassium granite in the melanosome is also relatively high. Black spots represent garnet.

from the paleosome, the potassium content of which is too low to produce substantial amounts of potassium granites. Although the decomposition of biotite and muscovite may increase the potassium content in the anatectic melt, Büsch *et al.* (1974) do not believe that the formation of the potassium granites can always be explained merely on the basis of the above reactions.

The normative Or:Ab:Q ratios for the analysed granites are shown in Fig. 37. Tourmaline-bearing granite veins have been encountered occasionally in mica schist and K-feldspar—sillimanite gneiss outcrops in the study area. The granite veins are more common in the K-feldspar—sillimanite zone than in the mica schist zone. Most of the tourmaline granite samples show a high Ab:Or ratio, despite variation in the composition of granite. Granite veins are seen almost invariably in the K-feldspar—cordierite gneiss and garnet—cordierite—sillimanite gneiss outcrops, which are actually veined gneiss.

The migmatising granite is mainly microcline granite with a high Or: Ab ratio. In the northern part of the K-feldspar—cordierite zone the migmatising granite tends to cover only a few percents of the surface area of the exposures and the granite veins are but a few centimetres in thickness. The amount of migmatising granite is not markedly higher in the southern part of the K-feldspar—cordierite zone than in the northern part. In the garnet—cordierite—sillimanite gneiss outcrops the migmatising granite often amounts to several tens percent, and the zone is characterised by large granitic areas (Fig. 2). The thickness of the granite veins and dykes in the garnet—cordierite—sillimanite gneiss outcrops varies from ten or so centimetres to several metres. The garnet—cordierite—sillimanite gneiss portions, or the melanosomes, exhibit microcline granite segregates with ill-defined boundaries (Fig. 38). The terminology of the migmatites is based mainly on definitions by Mehnert (1968). In the present study, melanosome refers to the dark portion of the migmatite. The portions containing garnet—cordierite—sillimanite are either paleosomes, whose composition has passed through metamorphism practically unaltered, or restites, which are depleted in granitic matter. In thin section, however, the garnet—cordierite—sillimanite melanosomes show microcline granite that separates garnet—cordierite—sillimanite-bearing portions from each other.

Hence, the amount of granite in the melanosome is significantly higher than could be deduced from megascopic examination. According to Mehnert (1968, p. 64), melanosomes are seldom completely depleted in light components. It must be noted, however, that during migmatisation the composition of the melanosome may be enriched in potassium (Härme 1965).

In the study area, mica schists are probably the rocks whose composition has changed least from the original. The upper part of the mica schist layer shows the highest K_2O , Al_2O_3 and FeO+MgO abundances and the lowest SiO_2 content. Figs. 39, 40, 41 and 42 demonstrate the relationship between some elements and SiO_2 in the analysed samples. The compositions of the K-feldspar—sillimanite gneisses and K-feldspar—cordierite gneisses are close to those of the mica schists; the variation is mainly due to layering, as it is in the mica schists. According to chemical data, the gneisses with K-feldspar porphyroblasts contain roughly as much K-feldspar as do the corresponding mica schists. Inclusions in K-feldspar porphyroblasts in samples from the northern part of the K-feldspar—cordierite zone show that these porphyroblasts were crystallised during the breakdown of muscovite (Fig. 10).

The chemical composition of the analysed garnet—cordierite gneisses is not unlike that of the upper part of the mica schist layers, excluding sample 61GC. The layers rich in plagioclase and poor in sillimanite (sample 61GC) or the layers resembling the lower parts of the mica schist layers are rare in many garnet—cordierite gneiss outcrops.

Thus, in composition, the garnet—cordierite—sillimanite melanosomes are partly more homogeneous and pelitic than the layered mica schists. Table 11, column 1, shows the average composition of the mica schists, which is the same as the mean of the compositions for samples 1MS, 5MS, 7MS, 8MS, 2MS, and 6MS; in other words, the average composition of the mica schist was calculated by means of the averages for the upper and lower parts of the layers. Column 2 gives the composition



Fig. 39. K_2O content plotted against SiO₂ content for analysed metapelites of the Rantasalmi—Sulkava metamorphic zones. Symbols: squares — the mica schist zone, triangles — the K-feldspar—sillimanite zone, open circles — the northern part and grey circles — the southern part of the K-feldspar—cordierite zone, dark circles — the garnet—cordierite—sillimanite zone.



Fig. 40. Al_2O_3 content plotted against SiO₂ content for analysed metapelites of the Rantasalmi—Sulkava metamorphic zones. Symbols are as in Fig. 39.



Fig. 41. CaO content plotted against SiO₂ content for analysed metapelites of the Rantasalmi—Sulkava metamorphic zones. Symbols are as in Fig. 39.



Fig. 42. FeO + MgO content (total Fe as FeO) plotted against SiO₂ content for analysed metapelites of the Rantasalmi—Sulkava metamorphic zones. Symbols are as in Fig. 39.

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Average chemical composition of mica schist (1), calculated chemical composition of restite (2), average chemical composition of garnet—cordierite—sillimanite gneiss (3)

	1	2	3
SiO,	65.6	59.4	60.7
TiO,	0.7	1.2	1.0
Al ₂ Õ ₂	17.0	18.0	21.3
FeO (total Fe as FeO)	5.7	10.4	8.2
MgO`	3.1	5.8	3.2
СаО	1.3	1.9	0.4
Na ₂ O	2.4	1.5	1.2
K ₂ Õ	4.2	1.8	4.0
	100.0	100.0	100.0

for the restite, calculated by subtracting from the average composition of mica schist 50 percent of the microcline granite whose composition equals the mean composition of the analysed microcline granites. The potassium content in the mica schists is so high that considerable amounts of potassium granites could have formed from it. Column 3 shows the mean composition of the analysed garnet—cordierite sillimanite gneisses. Their Al_2O_3 content is high but the CaO content very low. The CaO content in the calculated restite is significantly higher than the CaO in the mica schists. However, during potassium granite migmatisation, the CaO content in the melanosome may decrease (Härme 1962, 1965) and the Al_2O_3 excess over calcium and sodium increase in the potassium metasomatism accompanying migmatisation (Härme and Laitila 1955). Even so, the calcium content is fairly low in the mica schists, and extremely so in some sillimanite gneisses and cordierite gneisses. It may well be that the low calcium content in the garnet—cordierite—sillimanite gneisses is a primary feature of these rocks.

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In any case, in the garnet-cordierite-sillimanite gneisses the breakdown reactions of muscovite and biotite (reactions 1 to 5, pp. 35-36) have advanced far in the progressive stage of metamorphism. During these reactions, K-feldspar was developed and water liberated. The conspicuous Al₂O₃ and sillimanite content in the garnet-cordierite-sillimanite gneisses paid a marked contribution to the breakdown of biotite. At some stage of metamorphism these reactions were accompanied by anatexis. Part of the anatectic melt was expelled from the melanosome, but part remained in it. Nevertheless, in the garnet-cordierite-sillimanite zone, microcline granites are abundant and the potassium content is fairly high in the melanosomes as well. Thus, it seems likely that the garnet-cordierite-sillimanite zone received more potassium than was present in the original composition. The granite is of intrusive type, a feature particularly well developed in the large granite bodies whose centres are practically devoid of gneiss fragments. The decomposition reactions of biotite and muscovite in gneisses with a composition of mica schists or close to that of the garnet-cordierite-sillimanite gneisses played an important role in the development of intrusive potassium granite magma.

The water content in the microcline granite was fairly high, and thus the granite has caused retrogressive metamorphism in the garnet—cordierite—sillimanite zone. This retrogressive metamorphism is indicated, not only by water-bearing alteration products, but also by some anomalous values of the partition coefficients between the mineral pairs. Khlestov (1973 a, p. 38) maintains that the anatectic magma brings about retrogressive metamorphism in its crystallisation environment owing to the liberation of water. Schneider (1975) noted in melting tests that the water-bearing anatectic granite melt produces retrogressive metamorphism in its contact.

According to Vorma (in press), the rapakivi magma was formed under dry conditions through the dehydration of biotite, muscovite and hornblende. The dry rapakivi magma intruded to a level higher than the water-saturated microcline granite. Vorma (1972) has described the contact aureole around the Wiborg rapakivi granite massif in southeastern Finland. The microcline in the contact aureole has turned into orthoclase, because, in Vorma's opinion, the contact metamorphism took place under dry conditions. In the study area, K-feldspar, in both microcline granite and garnet—cordierite—sillimanite gneisses, is largely cross-twinned microcline. The determination of the thermal state of the K-feldspar by the X-ray diffraction method points in the same direction. Hence, also in this respect, the contact effect of microcline granite differed from that of rapakivi.

The microcline granite obviously had several contact effects. Its temperature may have exceeded that of the environment, and thus it would have caused progressive contact metamorphism in the environment. On the other hand, the granite expelled to its environment the water necessary for anatexis; this stage of contact metamorphism is, however, also accompanied by retrogressive metamorphism. Härme (1962 and 1965) noted that the contact effect of potassium granite aids anatexis, largely on account of its volatile components.

The K-feldspar—cordierite gneisses are migmatic, which is suggestive of anatexis. The mineral assemblages in these samples and melting tests (Winkler 1970, p. 220) indicate that the P—T conditions favoured the partial melting of the rocks. No large-scale anatexis has taken place in the K-feldspar—cordierite zone, however, owing to the lack of water. The mica schist zone and the K-feldspar—sillimanite zone do not show significant signs of anatexis. The tourmaline granite veins in these zones may derive from the granite that migmatised the K-feldspar—cordierite zone and the garnet—cordierite—sillimanite zone.

The composition of the metapelites in the Rantasalmi—Sulkava area has favoured the crystallisation of K-feldspar in progressive metamorphism. Consequently, the potassium content of the anatectic melts derived from these metapelites may have risen fairly high; in this respect the study area differs from the Haukivesi area north of the Rantasalmi map sheet area. The Geological Survey has recently carried out geologic mapping in the Haukivesi area (sheet 3234, Varkaus) under the supervision of the present author. The metasediments in the Central Haukivesi area are mainly migmatic plagioclase—biotite—garnet gneisses with a persistently low content of K-feldspar. The K-feldspar and potassium content of the rock that migmatises these gneisses is also low, owing to its trondhjemitic composition. The results reported by Rauhamäki (Rauhamäki and Gaál 1971) from the Haukivesi area are compatible with the concepts that emerged from the geologic mapping. The water content of the potassium granite intruding the garnet—cordierite—sillimanite zone in the Rantasalmi—Sulkava area was high, as is indicated by the retrogressive metamorphism caused by the granite.

The fairly high water content makes it impossible for the microcline granite to have intruded from a very deep level to the level at which the metamorphism of the garnet—cordierite—sillimanite zone took place. The observations refered to above suggest that the potassium granite is a product of progressive metamorphism. It must be remembered, however, that migmatisation and the events associated with it are not unambiguous. This was shown by Härme (1965), who maintains that the migmatisation caused by potassium granite promotes the crystallisation of garnet and cordierite because the composition of the rock undergoing migmatisation changes. Nykänen (1971), on the other hand, reports progressive contact metamorphism produced by late-Karelian Kitee granite in its environment.

CRYSTALLISATION CONDITIONS

According to J. B. Thompson (1955, 1957) and Korzhinskii (1959), the existence of an equilibrated mineral assemblage depends on temperature and pressure, on the activity or chemical potential of the perfectly mobile components and on the relative amounts of the inert components. If water is a perfectly mobile component

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in the metamorphism of pelites, its partial pressure does not depend on the primary composition in the same way as do the proportions of the inert components. Hence, the partial pressure of water does not vary from one layer to the other and, in a sense, water is an equilibrating factor similar to temperature and pressure in metamorphic processes.

Most of the reactions in the progressive metamorphism of pelites are dehydration reactions. The higher the metamorphic grade, the farther the dehydration has proceeded. Thus, the partial pressure of water has an important bearing on metamorphism: as shown by experimental and theoretical studies, increase of P_{total} at constant P_{H_2O} lowers the dehydration temperature (Thompson 1955, Dobretsov and Sobolev 1972). On the basis of studies of metamorphic rocks, the following generalisations can be made (Sobolev and Dobretsov 1972): The pressure of the fluid phase in metamorphism is mainly composed of the partial pressures of carbon dioxide and water. When pelites are metamorphosed, the partial water pressure is close to the total pressure, that is, $P_{H_2O} \approx P_{fluid} = P_{total}$. During metamorphism, the system is open with reference to water and carbon dioxide, both of which are perfectly mobile components. The increase in temperature, however, reduces the partial pressure of water because the decarbonisation reactions take place at high temperatures.

In the study area, the mineral assemblages of the mica schist, K-feldspar—sillimanite, and K-feldspar—cordierite zones are often regular, as was noted in the discussion on facies. The X_{Mg} ratio in cordierite and garnet, and the TiO₂ content in biotite increase from north to south in the K-feldspar—cordierite zone. Thus, the composition of the minerals change progressively southwards. These facts reveal that, by and large, water was a perfectly mobile component in the progressive stage of metamorphism. The occurrence of muscovite south of the K-feldspar—sillimanite isograd, however, suggests that the system was not unambiguously open for water in the progressive stage of metamorphism.

Evans and Guidotti (1966) have described a similar mode of occurrence in muscovite from the Bryant Pond area. They maintain that the breakdown reaction of muscovite might possibly have controlled the partial pressure of water. The increased P_{H_2O} brought about the pseudo-retrogressive crystallisation of muscovite. But the albite component of plagioclase also participates in the breakdown reaction of muscovite and so the dehydration of muscovite takes place over a temperature interval of about 30°C (Chatterjee and Froese 1975).

Provided muscovite decomposed in the study area in accordance with the model by Evans and Guidotti, P_{H_2O} was lower than the total pressure (cf. Guidotti 1970, p. 320).

Fig. 43 shows the curves of the muscovite breakdown reaction (reaction 1) in terms of P and T for various water contents in fluid phase ($P_{fluid} = P_{H_2O} + P_{CO_2}$) according to Kerrick (1972), and the solidus curves of granite for the corresponding composition of the fluid phase as reported by Novgorodov and Shkodzinskiy (1974).



Fig. 43. P—T diagram showing equilibrium curves in $H_2O + CO_2$ mixtures for muscovite breakdown according to Kerrick (1972), and for melting of granite according to Novgorodov and Shkodzinskiy (1974) (1.0, 0.8 and 0.5 are mole fractions of water). Stability relations of Al_2SiO_5 polymorphs are from Althaus (1969) (dot dash line) and from Richardson *et al.* (1969) (dotted line). Shaded area represents presumed pressure—temperature conditions in the K-feldspar—sillimanite zone and in the northern part of the K-feldspar—cordierite zone.

Marked in the same figure are the stability fields of the aluminosilicates sillimanite, kyanite and andalusite according to Althaus (1969) and Richardson *et al.* (1969). In the Rantasalmi—Sulkava area, muscovite disappears finally in the northern part of the K-feldspar—cordierite zone, that is, where the metapelites grade into veined gneisses and the anatexis presumably started.

The disappearance of muscovite in this area is largely consistent with the reaction proposed by Winkler (1974, p. 84): muscovite + quartz + albite + K-feldspar + $H_2O =$ liquid. According to Winkler, the reaction cannot take place if the water pressure is less than 3.5 kb. Fig. 43 shows the minimum P_{H_2O} at 4 kb. In the K-feldspar—sillimanite zone the breakdown of muscovite does not seem to be accompanied by anatexis. These facts indicate that P_{H_2O} was close to 4 kb at the boundary between the K-feldspar—sillimanite and K-feldspar—cordierite zones. If muscovite disappears and anatexis sets in at pressure of 4 kb, the temperature

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Fig. 44. Relationship between Mg—Fe distribution coefficient for the pair biotite—garnet and temperature according to Perchuk (1970) and according to Thompson (1976)

will then be 660°C as shown by Fig. 43. Carbonate intercalations have not been encountered in the metapelite outcrops examined, which suggests that the proportion of water in the fluid was quite high. On the basis of the stability field of the alumino-silicates constructed by Althaus, muscovite can disappear from the metapelites without anatexis within the stability field of sillimanite, even if water pressure were equal to total pressure. According to data determined by Richardson *et al.*, however, in this case water pressure must be less than total pressure.

Perchuk (1970) and A. B. Thompson (1976) have determined the relationship between the Mg—Fe partition coefficient of the mineral pair biotite—garnet and the crystallisation temperature. The crystallisation temperature was determined by means of a known method, such as the breakdown reaction of muscovite. Fig. 44 plots the natural logarithm of the partition coefficient as the function of the inverse of the absolute crystallisation temperature, since theoretically $\ln K_D = C_1 T_K^{-1} + C_2$ (McIntire 1963), where C_1 and C_2 are constants. Table 9 lists the crystallisation temperatures of the samples studied.

The crystallisation temperatures (670°C) obtained from Perchuk's diagram are consistent with observations on the breakdown of muscovite and the onset of anatexis. Perchuk's diagram gives a temperature of about 640°C for the boundary between the mica schist and K-feldspar—sillimanite zones. Samples 19KS to 23KS are also close to the K-feldspar—sillimanite isograd. According to Fig. 43, the breakdown of muscovite at this temperature takes place at a pressure of 3.3 kb, provided that

 $P_{H_{aO}} = P_{total}$. Since plagioclase participates in the reaction and since $P_{H_{aO}}$ may be less than P_{fluid} , the pressure must exceed 3.3 kb. However, because the K-feldspar sillimanite gneisses are not migmatic, $P_{H_{aO}}$ has to be less than 4 kb. Fig. 43 illustrates the presumed crystallisation conditions in the K-feldspar—sillimanite zone and in the northern part of the K-feldspar—cordierite zone. The crystallisation temperature, 740°C, of the samples from the northern part of the K-feldspar—cordierite zone, evaluated on the basis of Thompson's diagram, is significantly above the minimum temperature that heralds the beginning of anatexis. In terms of such a high temperature it is not easy to explain the simultaneous disappearance of muscovite and the onset of anatexis.

On the other hand, even the slighest variations in the partition coefficients are shown distinctly in the crystallisation temperatures, which were determined on the basis of Thompson's diagram; hence, accurate determination of the temperatures is difficult.

The average crystallisation temperature of the samples from the southern part of the K-feldspar—cordietite zone is only slightly higher than that in the northern part of the zone.

Retrogressive metamorphism prevents the Mg—Fe partition coefficients of the mineral pair biotite—garnet in the garnet—cordierite—sillimanite gneisses from being used for temperature determination. The crystallisation temperature for the hypersthene-bearing sample 62GC from the garnet—cordierite—sillimanite zone, determined by Perchuk's method, is 727°C. The other hypersthene-bearing samples gave crystallisation temperatures similar to those obtained for the K-feldspar—cordierite zone.

Thompson used other geologic thermometers to establish the relationship between the Mg—Fe partition coefficient of the mineral pair cordierite—garnet and the crystallisation temperature. Currie (1971) and Hensen and Green (1971, 1972, 1973) have studied experimentally the Mg—Fe partition between garnet and cordierite. According to Currie, the increase in the numerical value of the partition coefficient coincides unambiguously to the increase in crystallisation temperature. The compositions of the coexisting garnet and cordierite are, as shown by the above study, fixed at a certain temperature and pressure in the quartz—sillimanite assemblage. The findings of Hensen and Green suggest that the partition coefficient is not entirely dependent on the crystallisation temperature, although the partition coefficient generally decreases with the falling temperature. Hensen and Green also demonstrate that the compositions of garnet and cordierite refer to a certain pressure and temperature in the sillimanite—quartz assemblage or in the hypersthene—quartz assemblage.

The samples from the garnet—cordierite—sillimanite zone only exhibit the quartz—sillimanite—garnet—cordierite assemblage. In the samples from the K-feldspar—cordierite zone the aluminium excess might have been too low to allow the crystallisation of the fore-mentioned assemblage. Even so, the samples

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from the K-feldspa1—cordierite zone exhibit the mosaic equilibrium cordierite—sillimanite and cordierite—quartz.

Assuming that in the northern part of the K-feldspar—cordierite zone the temperature was 660°C and the pressure 4 kb, the crystallisation pressure may have been too low for the reaction cordierite = garnet + sillimanite + quartz to occur (cf. Winkler 1974, Fig. 14—12). Hensen and Green (1973) have demonstrated that X_{Mg} ratios in garnet and cordierite at given P and T are significantly higher in the quartz—hypersthene assemblage than in the assemblage sillimanite—quartz. Consequently, the compositions of garnet and cordierite may vary within a large range in the garnet—cordierite—quartz assemblage.

Shown on Table 9 is the relationship between the Mg—Fe partition coefficients of the coexisting cordierite and garnet, and the crystallisation temperature and pressure. The assemblage cordierite—garnet—sillimanite—quartz has been met with only in the samples from the garnet—cordierite—sillimanite zone, and hence, the P—T conditions of these samples alone were determined by means of the methods developed by Hensen and Green, and Currie.

The temperatures given to the samples from the K-feldspar-cordierite zone by Thompson's method do not differ much from those obtained for them by means of the Mg-Fe partition coefficients of the mineral pair biotite-garnet as proposed by Perchuk. The crystallisation temperatures are the same in both the northern and southern parts of the zone. The crystallisation temperature 694°C of the garnetcordierite-sillimanite gneisses calculated by applying the method developed by Currie is only slightly higher than the crystallisation temperature, 660°-670°C, determined by the other methods for the K-feldspar-cordierite gneisses. In contrast, the crystallisation temperatures (averaging 760°C and 772°C) obtained for the garnetcordierite-sillimanite gneisses by the methods of Hensen and Green, and Thompson are conspicuously higher than those of the K-feldspar-cordierite gneisses. Currie's method gives 5.6 kb for the crystallisation pressure of the garnet-cordieritesillimanite gneisses, whereas Hensen and Green's method suggests 7.5 to 8 kb. This pressure is significantly higher than the assumed crystallisation pressure of 4 kb of the K-feldspar-cordierite gneisses. According to Shkodzinskiy and Kitsul (1974), the X_{Mo} ratio in garnet and cordierite at anatexis is dependent on whether the system is saturated or undersaturated with water. Thus, the geobarometer, which is based on the compositions of garnet and cordierite, must be applied with caution to intensely migmatised areas.

The observations referred to suggest that the crystallisation temperature close to the K-feldspar—sillimanite isograd was 630° to 640°C, and the crystallisation pressure 3.5 to 4 kb, provided that water pressure approached total pressure.

Staurolite has not been found in muscovite-bearing samples from the mica schist zone; in these, staurolite is not stable on the K-feldspar—sillimanite isograd. According to Hoschek (1969), the reaction staurolite + muscovite + quartz = Al_2SiO_5 + biotite + H_2O takes place at 600°C if the pressure is 3 kb and the water

pressure equal to total pressure, but if the pressure is 4 kb, the reaction temperature is 630°C. The metamorphic grade increases at a low rate from the northern part of the mica schist zone to the K-feldspar—sillimanite isograd and no facies discordance has been noted between the northern and southern parts in the mica schist zone.

In the northern part of the K-feldspar—cordierite zone, the temperature was 660°C and the pressure 4.0 kb, provided that the partial pressure of water was close to the total pressure. Although the differences in the compositions of biotite, garnet and cordierite between the northern and southern parts of the zone indicate changes in the crystallisation conditions, this change is only vaguely indicated by the geothermometers. It is to be presumed that the crystallisation conditions differed so little from each other that the change is not readily verified by means of partition coefficients.

Some geothermometers and barometers show that the garnet—cordierite sillimanite zone was crystallised at higher temperature and pressure than was the K-feldspar—cordierite zone. Thus, the garnet—cordierite—sillimanite zone might represent a deeper level of the earth's crust than its environment. This concept is supported by the changes in the chemical composition of the garnet—cordierite sillimanite gneisses in relation to the K-feldspar—cordierite gneisses, by the intense migmatisation, and by the structural features that distinguish the garnet—cordierite sillimanite zone from its environment.

The uplift and development of the zone into a dome may be due to the intruded microcline granite. If the crystallisation pressures were directly proportional to the depth at which metamorphism took place, the pressure in the northern part of the K-feldspar—cordierite zone would correspond to the pressure at a depth of about 15 km, and the pressure of 7.5—8 kb to that at a depth of 25 to 30 km. Had the garnet—cordierite—sillimanite gneisses crystallised at 770°C, the temperature gradient between the K-feldspar—cordierite zone and the garnet—cordierite—sillimanite zone would have only been 10°C/km. Such a low temperature gradient and the sharp drop in it are not easy to explain, particularly since microcline granite has succeeded in producing progressive contact metamorphism in the garnet—cordierite—sillimanite at a temperature as high as 770°C, the crystallisation pressure need not necessarily have been significantly higher, than that in the K-feldspar—cordierite zone.

TYPE OF METAMORPHISM

In his global study, Miyashiro (1961) pointed out that characteristic of regional metamorphism in all areas is the depth at which a given temperature is attained. The geothermal gradient has varied from one metamorphic area to another and the progressive metamorphism in each area can be presented by a curve in P—T


Fig. 45. P-T diagram showing the main types of metamorphism after Miyashiro (1973). Shaded area represents presumed pressure-temperature conditions in the Rantasalmi-Sulkava metamorphic zones.

space. According to Miyashiro (1973), there are three main types of metamorphism: low-pressure, medium-pressure and high-pressure. These are shown on the P-T diagram in Fig. 45. He maintains that and alusite is a characteristic mineral of the low-pressure type and that biotite, cordierite, staurolite and sillimanite are common. Kyanite is typical of the medium-pressure type; biotite, almandine, staurolite and sillimanite occur frequently.

Several investigators are of the opinion that the metamorphism of the Finnish Precambrian bedrock in general should be included in the low-pressure type (Miyashiro 1961, Zwart 1967, Simonen 1973). Miyashiro (1973, p. 74) maintains that the orthopyroxene zones in West Uusimaa and Lapland were metamorphosed under conditions between the medium-pressure granulite facies and the low-pressure pyroxene-hornfels facies. Simonen and Vorma (in press) hold that the Karelian and Svecofennian schist belt was metamorphosed in southern Finland at low pressure. The type of metamorphism shown by the granulites in Lapland may differ from those mentioned above, since, according to Simonen and Vorma, the metamorphism of granulite is either of the low-pressure or intermediate-pressure type.

Meriläinen (1976) considers that the granulite area in Lapland was metamorphosed 2150 Ma ago under conditions of the granulite facies. Most of the granulite area was remetamorphosed 1900 Ma ago at high pressure and temperature under conditions of the low-grade granulite facies and the high-grade amphibolite facies. Hence, garnet altered into cordierite and biotite was recrystallised. However, as stated by

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FeO- (total Fe as FeO), MgO-microprobe determinations and X_{Mg} ratios of garnets and cordierites from cordierite granulites of Lapland and Mg—Fe partition coefficients between cordierite and garnet. Anal. Tuula Paasivirta. Sample 1 = Ola/KK/70 and sample 2 = Old/KK/70 (Korsman 1975)

Table 12

		E-O	24.0	X	К
		FeO	MgO	Mg	-D
1	garnet	28.7	10.0	0.383	6.5
	cordierite.	4.7	10.6	0.801	
2	garnet	28.1	8.8	0.358	6.5
	cordierite.	5.0	10.1	0.783	



Fig. 46. AFM projection showing fields for the mineral assemblage garnet—cordierite—sillimanite in cordierite granulites of Lapland (dashed line) and in garnet—cordierite —sillimanite gneisses in the Rantasalmi—Sulkava area (solid line). The compositions of garnet and cordierite granulites were determined in this study.

Meriläinen, it is unlikely that all the garnet-cordierite gneisses were generated at this stage.

Fig. 45 includes the plot of the Rantasalmi—Sulkava area on the P—T diagram, As shown by the figure, the area belongs to the low-pressure intermediate type. that is, the Buchan type (Read 1952). The study area does not differ from its environment in terms of the type of metamorphism, or generally speaking, the area should be included in the same type of metamorphism as are the Karelian and Svecofennian schists in southern Finland. According to Glebovitskiy (1970), the metamorphism of the northern Ladoga area is also classified as a low-pressure intermediate type.

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Eskola (1961) laid particular stress on the importance of the well-developed Fe—Mg diadochy in the garnet of Lappish granulites. The high pyrope content in these garnets indicates high crystallisation pressure, which ,according to Eskola, distinguishes the Lappish granulites from rocks of the amphibolite facies. If garnet and cordierite are assumed to be purely Fe—Mg-aluminosilicates, their compositions are related to certain crystallisation conditions in the assemblage quartz—sillimanite—cordierite—garnet.

Table 12 lists the FeO and MgO abundances of some garnets and cordierites from the Lappish cordierite granulites, the X_{Mg} atomic ratios and the Mg—Fe partition coefficients between garnet and cordierite. Fig. 46 shows on the AFM projection the fields of the garnet—sillimanite—cordierite paragenesis for the garnet—cordierite—sillimanite gneisses in the Rantasalmi—Sulkava area and the Lappish cordierite granulites. The fields are unalike, which suggests that the areas were metamorphosed under dissimilar conditions. The values of the Mg—Fe partition coefficients for the mineral pair cordierite—garnet in the cordierite granulites of Lapland are slightly higher than the corresponding values for the garnet—cordierite—sillimanite gneisses in the study area, but the values of the partition coefficients for the cordierite are somewhat lower than those obtained from the K-feldspar —cordierite zone.

Experimental studies show that especially increasing pressure moves the field of garnet and cordierite on the AFM projection towards the AM side, whereas the partition coefficient depends largely on the crystallisation temperature. The observations referred to above seem to indicate that the Rantasalmi—Sulkava area may possibly belong to a facies series of lower pressure than do the cordierite granulites in Lapland.

SUMMARY

In the southern part of the Rantasalmi—Sulkava area, metamorphism has almost completely obliterated the structures of metapelites that are still visible in the mica schists in the northern part of the area. The structures were wiped out gradually from north to south as the metamorphic grade increased.

Progressive metamorphism can be traced in the study area by means of changes in the mineral assemblages in the metapelites. The occurrence of the assemblage quartz—K-feldspar—sillimanite—cordierite—garnet was restricted chiefly by the bulk composition of the metapelites. Consequently, it is not easy to conclude to what extent the equilibrium of this assemblage was restricted by crystallisation conditions. Further, it is difficult to establish the role played by migmatisation in the chemical composition of the garnet—cordierite—sillimanite gneisses. At any rate, the composition of some samples taken from unmigmatised outcrops, such as mica schists and K-feldspar—sillimanite gneisses, is nearly similar to that of the garnet—cordierite—sillimanite gneisses. The mica schist zone was metamorphosed under conditions of the amphibolite facies, and the garnet—cordierite—sillimanite zone under those of the granulite facies. The grading from the amphibolite to the granulite facies occurs between the K-feldspar—cordierite zone and the garnet cordierite—sillimanite zone.

Progressive metamorphism is also manifested in changes in the compositions of the minerals, for example, in the increase in TiO_2 in biotite. Characteristic of the area is the fairly low pyrope content in garnet in the assemblage biotite—garnet—cordierite or in the assemblage quartz—sillimanite—garnet—cordierite.

Analytical data reveal that the pyrope content in the garnets of the hypersthene zones in southern Finland tends to be low (Parras 1958, Korsman 1975, Paavola 1976, Kays 1976). According to Glebovitskiy (1970), the pyrope content in the garnets of the granulites in the northern Ladoga area does not exceed 28 percent.

Data on the relationship between crystallisation conditions and the partition of elements between mineral phases and the compositions of coexisting minerals are to some extent contradictory. Hence, P—T conditions estimated by means of partition coefficients and mineral composition must be taken with some reservation. Nevertheless, the retrogressive metamorphism of the garnet—cordierite—sillimanite zone is well indicated by the Mg—Fe partition coefficients for the biotite—garnet pair.

On the basis of observations on the disappearance of muscovite and the onset of anatexis, it has been concluded that in the northern part of the K-feldspar cordierite zone the crystallisation temperature was 660°C and pressure 4 kb, provided that the water pressure approached the total pressure. The crystallisation temperatures determined by means of the Mg—Fe partition coefficients for the mineral pairs biotite—garnet and cordierite—garnet of the samples from the northern part of the K-feldspar—cordierite zone correspond to this temperature. Some observations suggest that the garnet—cordierite—sillimanite zone was metamorphosed at a significantly higher temperature than was the K-feldspar—cordierite zone. According to Paavola (1976), the Central Haukivesi area was metamorphosed partly under conditions of the amphibolite facies and partly under those of the granulite facies.

As shown by Paavola, the crystallisation temperatures obtained from the Mg—Fe partition coefficients for the biotite—garnet pair vary from 600° to 680°C in samples from the Central Haukivesi zone.

Velikoslavinskiy (1972) has estimated, on the basis of the Mg—Fe partition coefficients for the biotite—garnet pair, that the crystallisation temperature in the second sillimanite zone in the northern Ladoga area (sillimanite—almandine—K-feldspar subfacies) was 630° to 695°C, and in the hypersthene zone (hornblende—granulite subfacies) 695° to 760°C. According to Glebovitskiy *et al.* (1972), the rocks of the granulite facies of northern Ladoga were crystallised at a pressure of 4 to 4.5 kb. The crystallisation pressure was determined by means of the compositions of the coexisting biotite, garnet and cordierite. Thus, the P—T conditions in the northern Ladoga area were very similar to those in the Rantasalmi—Sulkava area.

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The metamorphism of the Rantasalmi-Sulkava area, like that of the northern Ladoga area, is of the Buchan type, that is, it is a low-pressure intermediate type.

The samples for the present study exhibit fairly constant partition of vanadium and cobalt between biotite and garnet. However, comparison of the partition coefficients reported in various papers shows that the partition coefficients of vanadium and cobalt between biotite and garnet are ill-suited to delineate the metamorphic grade.

Although migmatising potassium granite shows intrusive features, the granite was formed through the anatexis of metapelites during progressive metamorphism. The fairly high K_2O in the K-feldspar—cordierite gneisses, the poikiloblastic K-feldspar grains and the scarcity of migmatising granite all show that anatexis did not proceed far in the K-feldspar—cordierite zone. Presumably, there was not sufficient water nor high enough P_{H_2O} even if the partial pressure of water was close to total pressure. The water liberated in the breakdown of biotite and muscovite and the water of the potassium granite had a favourable effect on anatexis. The water of the potassium granite contributed decisively to the intense retrogressive metamorphism of the garnet—cordierite—sillimanite gneisses.

The features that characterise the study area, such as the close relationship between migmatisation and progressive metamorphism, the fairly high quantities of migmatising granite, and the mineral assemblages of high temperature but relatively low pressure, are all characteristics of the Hercynian type of orogeny. According to Zwart (1967), the metamorphism of the Svecofennian and Karelian formations corresponds to that connected with the Hercyno-type orogeny. The uplift in this part of the Baltic Shield is apparently rather small, generally not more than 15 km and in many places even less (Zwart *op. cit.*). This depth of erosion, 15 km, thus refers to the depth at which the northern part of the K-feldspar—cordierite zone underwent metamorphism.

ACKNOWLEDGEMENTS

This thesis is the outcome of research carried out in 1974—76 while I was research assistant of the Natural Science Research Council of the Academy of Finland. The laboratory investigations I did in the Geological Department at the University of Turku and at the Geological Survey of Finland.

I am deeply grateful to my esteemed teacher, Professor K. J. Neuvonen of the University of Turku for his continuous intrest in my study and the support and encouragement he gave me.

Professor Nils Edelman and Dr. Atso Vorma were the official readers of the manuscript and I have endeavoured to put their constructive criticism to good use. Dr. Atso Vorma deserves my special thanks for stimulating discussions. I thank Professor Maunu Härme for the assistance he lended to me in the course of the work.

The manuscript was translated into English by Mrs Gillian Häkli.

Mrs. Ritva Ääri helped me in many ways in the geological laboratory at the University of Turku. The drawings are the work of Mrs. Elsa Järvimäki, Mrs. Ritva Forsman, Miss Liisa Siren, Mrs. Anni Vuori and Mr. Pekka Wasenius.

I greatly appreciate the financial support given me by the Outokumpu Oy Foundation.

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Manuscript received February 17, 1977.

ISBN 951-690-072-0