

Zeitschrift:	Schweizerische mineralogische und petrographische Mitteilungen = Bulletin suisse de minéralogie et pétrographie
Band:	66 (1986)
Heft:	3
Artikel:	The metamorphic evolution of garnet-cordierite-silimanite gneisses of NW Spitsbergen (Svalbard)
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DOI:	https://doi.org/10.5169/seals-50897

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The metamorphic evolution of garnet-cordierite-sillimanite gneisses of NW Spitsbergen (Svalbard)

by *Eva M. Klaper*¹

Abstract

Metapelitic gneisses occurring within the Caledonian migmatite complex of NW Spitsbergen (Svalbard archipelago) are most typically represented by schistose biotite-bearing cordierite—garnet—sillimanite rocks.

A consistent pressure-temperature range of 5 to 6 kbar and 650°C increasing to 700°C from the south to the north has been deduced for the last metamorphic equilibration using standard geothermobarometry with water-saturated cordierite. Calculated with dry cordierite the pressure reaches 4 to 5 kb at the same temperature of metamorphism.

The textural and chemical features are interpreted in terms of a set of (retrograde) mineral reactions. From the textural evidence as well as from the deduced mineral reactions it may be concluded that the rocks were at higher metamorphic conditions prior to the PT-conditions calculated from five published geothermobarometers.

Although the maximum PT-conditions reached are unknown, it is suggested that they were close to granulite facies conditions with low water activity. The H₂O required to form the late stage assemblages containing hydrated minerals such as biotite and muscovite was presumably released by the crystallizing migmatitic melts in the quartzo-feldspathic portions of the rocks. Therefore, the water activity increased significantly during the initial decompression and cooling of the rocks.

The well preserved portion of the PT-path described in this study is interpreted as a part of essentially one main Caledonian metamorphic cycle culminating in regional upper amphibolite facies metamorphism, “in situ” partial melting and migmatization. Additional heating may be postulated in connection with the intrusion of syntectonic gray granites in the northern part of the study area. Weak late heating is related to the intrusion of the post-tectonic Hornemannoppen Batholith.

Keywords: Metapelites, migmatites, geothermobarometry, Caledonian, Spitsbergen.

Introduction

During the summer of 1983 a Swiss expedition visited NW Spitsbergen, Svalbard. As a member of this expedition I was given the opportunity to carry out fieldwork.

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The studied area is located north of $79^{\circ}30'$ north and west of 12° east and includes Magdalenefjorden, Smeerenburgfjorden and Vasahalvoya. The region visited consists of a Caledonian migmatite and gneiss complex. Granites, migmatitic and metapelitic gneisses form the vast majority of rock types exposed in this area which had been systematically studied by a number of expeditions of the Polar Institute of Norway. A compilation of most available data has been given by HJELLE and OHTA (1974).

The primary purpose of this study is to describe the observed textures and the mineral chemistry of metapelitic cordierite - garnet - sillimanite gneisses occurring in the lower Hekla Hoek succession of northwestern Spitsbergen. The metamorphic conditions under which these rocks equilibrated will be deduced from phase petrology and standard geothermobarometry.

Geological setting

The Caledonian gneiss and migmatite area of northwestern Spitsbergen is characterized by the predominance of various banded gneisses and granitic migmatites. Amphibolites, hornblende gneisses as well as calcsilicate marbles

1	Virgohamna	17, 18
2	Kobbefjorddalen	1, 2, 3, 4, 5, 6
3	Magdalenehuken	53, 54, 55, 56, 57, 58
4	Aasefjellet	59, 60
5	Trankollane	52
6	Fugleholmen	61
7	Losvikfjellet	62, 63, 64, 65, 66
8	Sverdrupfjellet	15, 16
9	Blessingberget	7, 8, 9, 10, 11, 12
10	Holmgrenfjellet	13, 14
11	Likneset	19, 20
12	Drottenfjellet	37
13	Birgerfjellet	51
14	Indre Norskoya	21, 22, 23, 24, 25, 26, 27, 28, 29
15	Ytre Norskoya	30, 31, 32, 33, 34, 41
16	Klovnningen	42, 43, 44, 45
17	Risen	38, 39, 46
18	Fuglesangen	47
19	Jarlifjellet	48, 49
20	Bjelbofjellet	35, 36, 40
		50

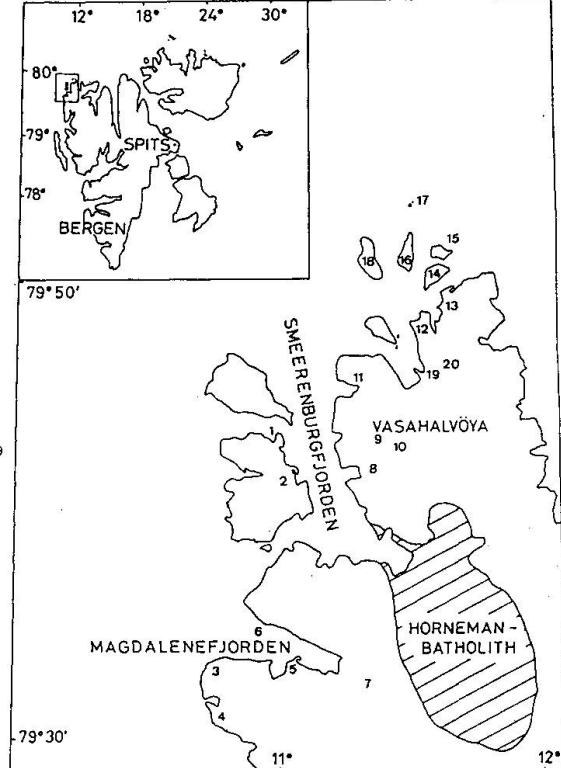


Fig. 1 Map of sample localities. Locality numbers correspond to those in Table 1. Geology: migmatitic gneisses and granites (white) and post-tectonic hornblende-biotite granodiorite intrusion (dashed).

and dolomitic marbles occur as small lenses and bands within the migmatite complex. This metasedimentary rock sequence represents part of the lower Hekla Hoek succession of Precambrian age (OHTA 1974a). A large post-tectonic hornblende-biotite granodiorite pluton (Hornemantoppen Batholith) intruded the central part of this area (Fig. 1).

A complex deformation history with three main phases of folding has been described by HJELLE and OHTA (1974). The general trend of early formed fold axes is from north to south, while later minor folds show east-west trending fold axes. Metamorphism reached upper amphibolite facies during the main phases of Caledonian deformation and was followed by recrystallisation which may be correlated with later deformation phases (OHTA 1974b). Formation of an open, regional N-S trending anticlinorium with south plunging fold axis (HJELLE 1979) in NW Spitsbergen is associated with later migmatization and synchronous formation of grey granites. Late stage chloritization can be related to the intrusion of the Hornemantoppen granodiorite body.

Description of rock types and mineral textures

The most common rock types considered in this study are dark biotite-rich metapelitic schists or gneisses and lighter, also biotite-bearing gneissose rocks with cordierite recognizable in hand specimen. The biotite-rich samples are strongly foliated and banded while the lighter coloured rocks usually display only a weakly to moderately developed foliation.

The mineral assemblages found in metapelitic gneisses are listed in table 1.

Garnet: In most samples garnet displays irregular and lobate outlines (Fig. 2c). The grainsize typically reaches 1–3 mm. Idiomorphic crystals are extremely rare. Cordierite may replace garnet pseudomorphically. Observed inclusion minerals in garnet are quartz, biotite, plagioclase and, in one sample (Spi 66), also cordierite. Some samples contain web-like garnets which reach diameters of 2–4 cm. Polyphase garnets are rare, but do exist in the northern and western, tectonically deeper parts of the studied area (Fig. 2e, 2f). In some samples garnets are broken apart with muscovite growing in the gaps (Fig. 2b).

Cordierite: Cordierite is very abundant in some rocks. It either forms the rock matrix together with quartz and plagioclase (Fig. 2b, 2d, 2f) or occurs as a late phase replacing garnet (Fig. 2c). Cordierite may be overgrown by fibrolitic sillimanite and is often cracked and pinitized or almost completely transformed to a sericitic felt. Cordierite also forms reaction rims around spinel (Fig. 2a).

Sillimanite: Sillimanite occurs in two different textural types a) as small prismatic needles closely associated with biotite, b) overgrowing cordierite (Fig. 2c) which may replace garnet. Sometimes, sillimanite occurs as radial fibers on

Tab. 1 Mineral assemblages of metapelitic gneisses.

Sample	Locality	Qtz	Mu	Bi	Gt	Sill	Crd	Plg	Sp	other
SPI 1	2	x		x					o	Amph, Chl
SPI 2	2	x	x	x	x			x		Chl, Serizite
SPI 4	2	x	x	x				x		Amph, Chl
SPI 5*	2	x	o	x	x			x		
SPI 6	2	x	o	x	x			x		Chl, Ep
SPI 7	9	x	o	x		x		x		Chl
SPI 8	9	x	o	x		x		x		Kfs, Chl
SPI 9	9	x	o	x	x	x		x		Chl
SPI 12*	9	x	x	x	x	x	x	x		Kfs, Chl
SPI 14	10	x		x				x		Chl
SPI 15*	8	x		x	x			x		Chl, Amph
SPI 16	8	x	x	x		x	x	x		Kfs
SPI 18*	1	x		x		x		x		Chl
SPI 20	11	x		x				x		Kfs
SPI 21	14	x	o	x	x			x		Kfs
SPI 25	14	x		x	x			x		Chl
SPI 27	14	x	o	x				x		Kfs, Chl
SPI 29	14	x	o	x				x		Kfs
SPI 30	14	x	x	o				x		Kfs, Chl
SPI 33*	14	x		x	x			x		Kfs
SPI 34	14	x		x	x			x		Kfs
SPI 36*	19	x	o	x	x		x	o		Chl
SPI 37*	12	x	o	x	x ²	x	x			Chl
SPI 38*	16	x		x	x	x	x	x	o	Kfs, Chl, Ky
SPI 39*	16	x	o	x	o	o	x	o		Chl
SPI 40	19	x	x	o	x		x	x		
SPI 43*	15	x		x	x			x		
SPI 47	17	x		x			x	x		Kfs, Chl
SPI 49	18	x	x	x		x	x	x		Kfs, Chl
SPI 54	3	x		x	x			x		Chl
SPI 55	3	x		x				x		Chl
SPI 59*	4	x		x	x ²			x		Kfs, Chl
SPI 60	4	x	x	x		x				Chl
SPI 64	7	x	x	x	x	x		x		Chl
SPI 65	7	x		x	x	x		x		Chl
SPI 66*	7	x	x	x	x	x	x	x		Kfs, Chl
Oht 403		x	x	x	x			x	x	Co
Oht 403				x	x		x		x	Co, Amph
Oht 440		x		x			x	x	x	Co

Oht = Ohta (1974a), x: essential mineral, o: accessory mineral,
2: two or more generations, r: retrograde

*: Samples analyzed on the electron microprobe

Abbreviations of mineral names:

Alm	Almandine	Crd	Cordierite	Mu	Muscovite
Amph	Amphibole	Ep	Epidote	Plg	Plagioclase
An	Anorthite	Gt	Garnet	Qtz	Quartz
Bi	Biotite	Kfs	K-feldspar	Sill	Sillimanite
Chl	Chlorite	Ky	Kyanite	Sp	Spinel
Co	Corundum				

completely pinitized cordierite. It is difficult to decide whether these sillimanite fibers were formed from muscovite or decomposed to muscovite.

Spinel: Green spinel was observed as a modally subordinate mineral. Spinel only occurs as inclusion in cordierite (Fig. 2a) and in the neighbourhood of sillimanite overgrowing cordierite.

Biotite: The modal abundance of biotite varies strongly with the rock type, but it is a major constituent in all samples. Biotite was found in contact with all other minerals.

Muscovite: Muscovite is a rather late mineral replacing cordierite (Fig. 2f) and often coexists with garnet (Fig. 2b).

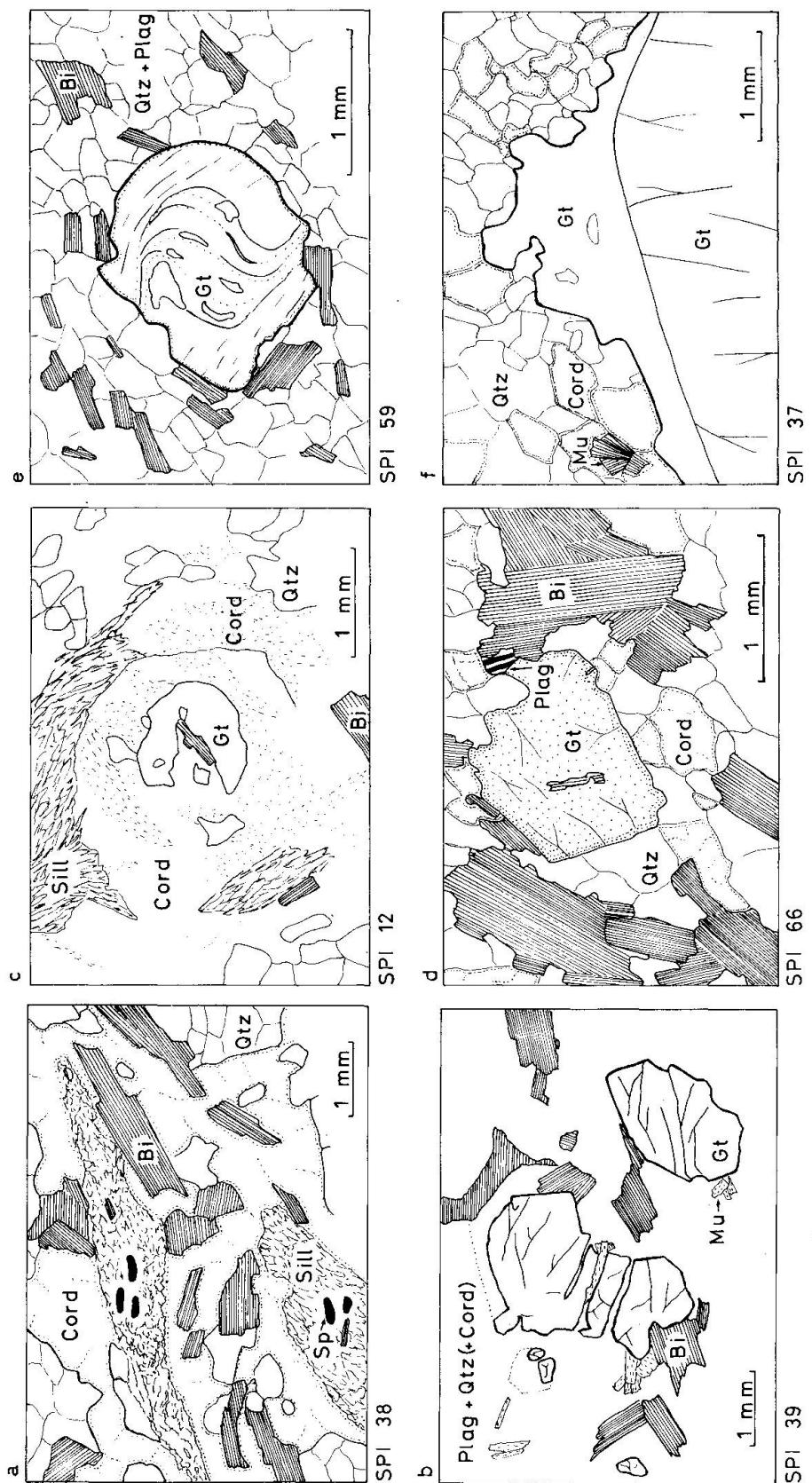


Fig. 2 Typical textures from metapelitic rocks:

a) Spinel mantled by cordierite growing in a cordierite-sillimanite clot (SPI 38).
 b) Garnet replaced by cordierite and biotite, late muscovite growth (SPI 39).
 c) Partially resorbed garnet replaced by cordierite (SPI 12).
 d) Garnet in biotite-quartz-plagioclase matrix (SPI 16).
 e) Polyphase syn- to post-tectonic garnet (SPI 59).
 f) Polyphase post-tectonic garnet (SPI 37).

Tab. 2 Representative microprobe analyses of Spitsbergen minerals.

Sample Comment	Garnet						Biotite*					
	12/7 rim	12/6 core	38/2 rim	38/6 core	66/6 rim	66/2 core	12/1	38/1	38/15 **	66/11	66/2 i	
SiO ₂	37.41	37.59	37.21	38.35	37.09	37.75	34.32	34.91	36.19	34.23	35.33	
TiO ₂							3.89	2.29	4.39	5.20	3.57	
Al ₂ O ₃	21.16	20.86	22.15	20.99	21.34	20.67	19.09	18.02	18.08	17.86	18.27	
FeO	33.89	33.77	34.06	34.90	33.43	33.91	22.44	23.32	20.93	22.80	22.79	
MnO	3.52	4.22	2.12	2.35	5.76	3.51			0.15	0.16		
MgO	2.68	2.79	3.40	2.98	1.52	2.50	6.53	8.00	8.09	5.46	7.08	
CaO	1.26	1.39	0.76	0.93	0.94	0.92			0.14	0.22	0.17	0.16
Na ₂ O							10.15	7.25	8.78	9.34	9.30	
K ₂ O												
Total	99.92	100.62	99.70	100.48	100.08	99.43	96.56	95.12	96.68	96.25	97.29	
Formula	12 (0)						22 (0)					
Si	3.015	3.017	2.931	3.014	3.006	3.054	5.267	5.428	5.441	5.329	5.392	
Ti							0.449	0.268	0.496	0.609	0.410	
Al	2.010	1.973	2.057	1.944	2.039	1.971	3.453	3.303	3.204	3.278	3.286	
Fe++	2.284	2.266	2.243	2.293	2.266	2.294	2.880	3.032	2.631	2.968	2.908	
Mn	0.240	0.287	0.141	0.156	0.395	0.240			0.020			
Mg	0.322	0.334	0.399	0.349	0.184	0.301	1.494	1.854	1.813	1.267	1.610	
Ca	0.109	0.120	0.064	0.077	0.082	0.080			0.042	0.099	0.064	0.051
Na									1.987	1.438	1.684	0.047
K										1.855	1.810	
Total	7.98	8.00	7.84	7.83	7.97	7.94	15.57	15.42	15.33	15.38	15.46	
Alm	77.3	75.4	78.8	79.8	77.4	78.7						
Spess	8.1	9.5	5.0	5.4	13.5	8.2						
Pyr	10.9	11.1	14.0	12.1	6.3	10.3						
Gross	3.7	4.0	2.2	2.7	2.8	2.7						
PLAGIOCLASE												
Sample	12/2	38/3	66/9	12/4	38/1	12/1	38/1	38/2	12/1	38/4	66/8	
SiO ₂	59.14	61.99	62.63	63.66	66.36	36.18	36.80	0.00	47.59	49.05	48.53	
Al ₂ O ₃	25.58	23.37	22.59	18.42	18.06	59.12	59.19	56.35	32.29	32.33	31.91	
FeO						1.18		33.20	11.01	9.32	10.27	
FE ₂ O ₃								1.94				
MgO						0.23	0.07	3.07				
CaO	6.81	5.87	5.48	0.46	0.00				6.24	7.74	5.82	
Na ₂ O	6.97	7.97	7.60	1.40	2.94				0.09	0.18	0.40	
K ₂ O	0.31	0.24	0.16	15.71	11.82				0.04	0.00	0.06	
Cr ₂ O ₃								1.70				
ZnO								3.32				
Total	98.81	99.44	98.46	99.65	99.18	96.71	96.06	99.00	97.26	98.61	96.99	
Formula	12 (0)			8 (0)			5 (0)			16 (0)		18 (0)
Si	2.662	2.763	2.806	2.965	3.033	1.015	1.033	0.000	5.010	5.052	5.090	
Al	1.357	1.228	1.193	1.011	0.973	1.953	1.955	7.690	4.007	3.925	3.945	
Fe++						0.028		3.219	0.969	0.803	0.901	
Fe+++								0.169				
Mg						0.010	0.003	0.530	0.979	1.188	0.910	
Ca	0.328	0.280	0.263	0.023	0.000				0.018	0.034	0.081	
Na	0.608	0.689	0.660	0.126	0.260				0.005	0.000	0.008	
K	0.018	0.014	0.009	0.933	0.689				0.156			
Cr								0.284				
Zn												
Total	4.97	4.97	4.93	4.93	4.96	3.01	2.99	11.99	10.99	11.00	10.94	
An	34.4	28.5	28.2	2.1	0.0							
Ab	63.7	70.1	70.8	11.6	27.4							
Or	1.9	1.4	1.0	86.2	72.6							

* Total includes minor amounts of F and Cl, ** i: inclusion in garnet

Chlorite: Chlorite is clearly a late alteration product and replaces biotite in some samples.

Mineral chemistry

The chemical composition of the minerals was obtained by using an automated ARL microprobe equipped with a Tracor Northern energy dispersive detector system and five crystal X-ray spectrometers at the Institut für Mineralogie und Petrographie at ETH Zurich. Analyzed samples are marked with a subscript * on Table 1 and representative mineral analyses are given in Table 2. The extent of Mg/Fe-1 substitutions is shown in the histograms in Fig. 3.

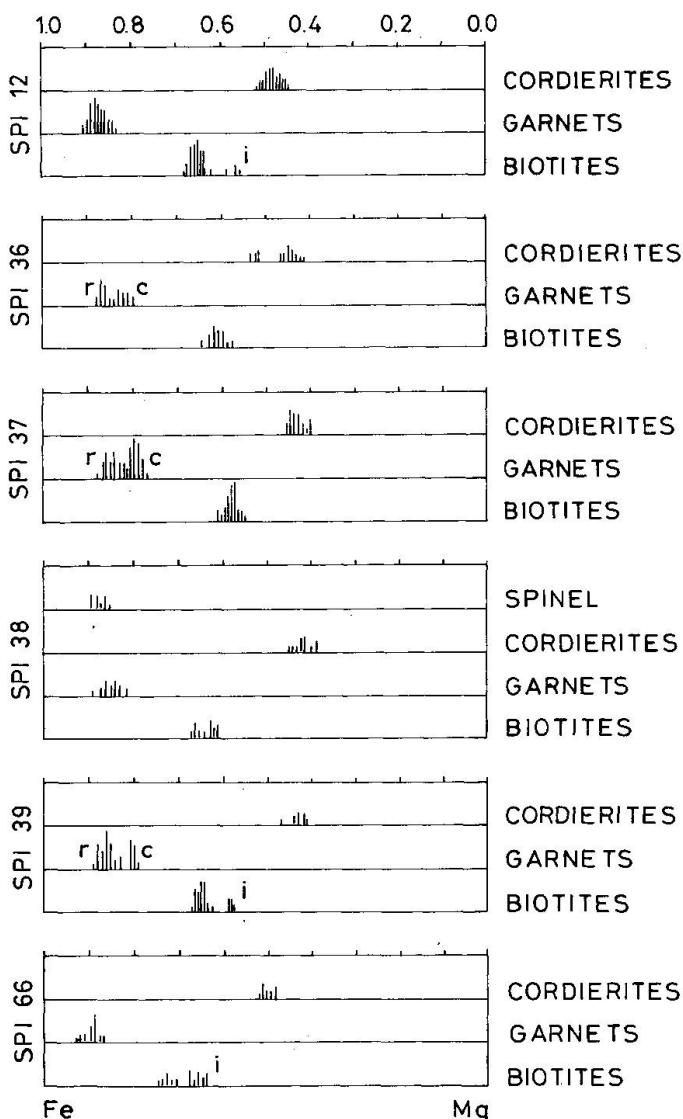


Fig. 3 Iron-magnesium variation histograms for biotite, garnet, cordierite and spinel (SPI 38). Smallest dash represents one analysis. r: rim, c: core, i: inclusion in garnet.

Garnet: The single phase garnets show no or only a very weak chemical zonation (Fig. 4) with some garnet cores slightly enriched in pyrope and depleted in almandine component. Most garnets (except SPI 39 and 59) show no zonation with respect to the grossular component and show only a slight variation in the spessartine component which may be enriched in the outermost parts of the garnets. This enrichment in manganese can best be observed in garnets which display characteristic resorption features and have very irregular outlines. This pattern of Mn enrichment has been reported to be a typical resorption feature by GRANT and WEIBLEN (1971) and DEBETHUNE et al. (1975).

In some samples (e.g. SPI 37, SPI 59) two growth phases of garnets can be observed (Fig. 2e and f). The analyzed polyphase garnets may or may not show a zonation. There is, however, no discontinuity in the chemical composition from core to rim. Examples are given in Figure 4. No zoning can be observed in polyphase garnets from cordierite bearing samples (e.g. SPI 37) from the northern, higher temperature part of the area studied. WOODSWORTH (1977) has shown that prograde garnet zoning profiles may be smoothed or homogenized at high grade. Such homogenization could have taken place at the temperatures attained (700 °C) during metamorphism (see below). However, polyphase garnets from biotite gneiss samples from the south-western part of the area (e.g. SPI 59) show a clear zoning pattern with respect to their Ca, Fe and Mn zoning.

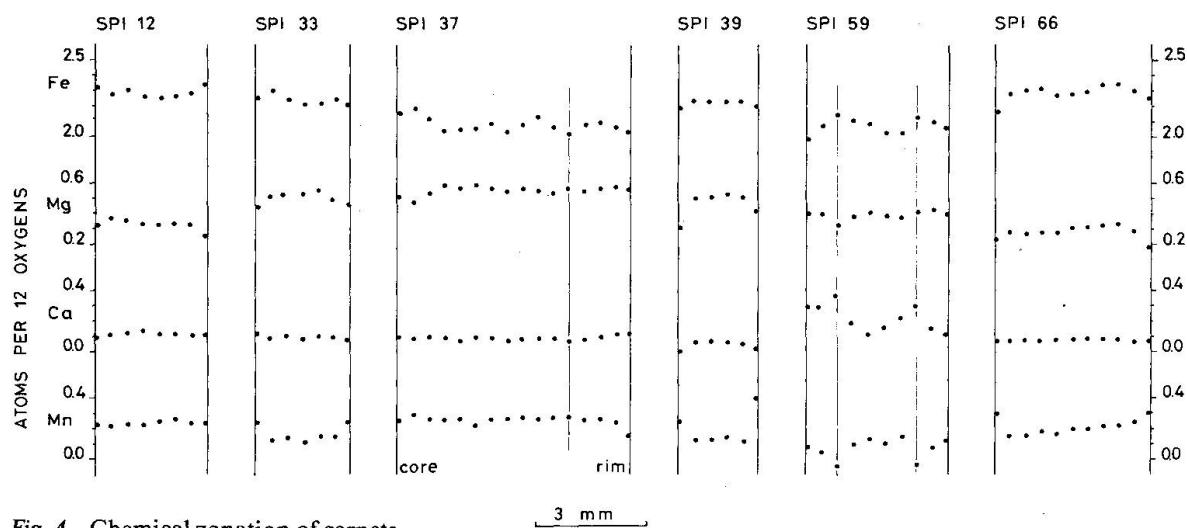


Fig. 4 Chemical zonation of garnets.

Fe/(Fe+Mg) ratios vary within narrow limits in each sample and between 0.75 to 0.94 on the whole. Some samples show a bimodal iron-magnesium distribution (Fig. 3) in garnet which is related to differing core and rim compositions.

Biotite: Fe/(Fe+Mg) ratios of 0.5–0.75 were found in all analyzed samples (Fig. 3). Biotite inclusions in garnet have lower Fe/(Fe+Mg) ratios than the matrix biotites.

Cordierite: Cordierites are chemically homogeneous with Fe/(Fe+Mg) ratios between 0.4 and 0.5, irrespective of local mineral assemblage or grain size. In some samples cordierite shows a strong alteration to yellow pinite. Most analyses show up to 0.50 wt% Na₂O. The water content in cordierite has not been determined.

Spinel: All spinels analyzed are essentially hercynite—Mg-spinel solutions with up to 13 mol. % spinel endmember. Gahnite (up to 7 mol. %), magnetite (up to 3 mol. %) and chromite (up to 2 mol. %) components total less than 12 mol. %.

Plagioclase: The anorthite content of plagioclases varies about 10 mol. % within each sample but shows no systematic zoning within each grain and lies between An(20) and An(35) in the different samples analyzed. No regional changes in plagioclase composition were observed.

Alkalifeldspar: The composition of K-feldspar (microperthite) ranges from Or(70) to Or(90).

White mica: Most analyzed muscovites contain 4 to 7% celadonite solid solution.

Chlorite: The chlorite which is a late stage mineral and formed as alteration product from biotite is of ripidolitic composition.

Geothermometry and geobarometry

The widespread mineral assemblage cordierite – biotite – garnet – plagioclase – sillimanite – quartz can be used to deduce estimates of pressure and temperature conditions under which these rocks equilibrated. The calibrations of the equilibria listed below were used for PT-calculations (exchange component notation adopted from THOMPSON (1982)).

(1) MgFe-1(bi) = MgFe-1(gt)	THOMPSON (1976)
(2) MgFe-1(crd) = MgFe-1(gt)	INDARES and MARTIGNOLE (1985)
(3) an(plag) = gross(gt) + sill + qtz	THOMPSON (1976)
(4) Fe-crd(crd) = alm(gt) + sill + qtz	ARANOVICH and PODLESSKIY (1980)
	GHENT et al. (1979)
	THOMPSON (1976)
	NEWTON and WOOD (1979)

The coexistence of K-feldspar and plagioclase in sample SPI 38 permits the temperature estimation from the additional equilibrium

$$(5) \text{NaK-1(K-fsp)} = \text{NaK-1(plag)} \quad \text{STORMER and WHITNEY (1977).}$$

Temperatures calculated from equilibrium 1 are shown in Fig. 5 and those from equilibrium 2 in Fig. 6. The scatter of the calculated garnet-biotite rim temperatures is of about 100°C around mean temperatures of 650 to 700°C in the southern part of the studied area and 700 to 750°C in the central and northern part. To deduce statistically valid mean temperatures, the average value from several samples has been taken for the northern and southern part of the studied area. Garnet-cordierite temperatures generally overlap well with garnet-biotite temperatures but show a slightly smaller scatter. Garnet core temperatures calculated from biotite or cordierite inclusions in garnet (e.g. SPI 18,

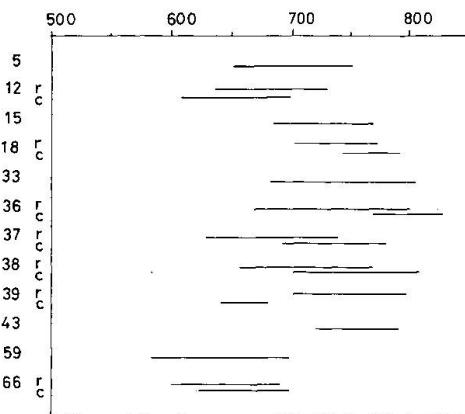


Fig. 5 Calculated temperatures from garnet-biotite Fe-Mg exchange thermometer at 5 kbar. Calibration of THOMPSON (1976). r: rim, c: core.

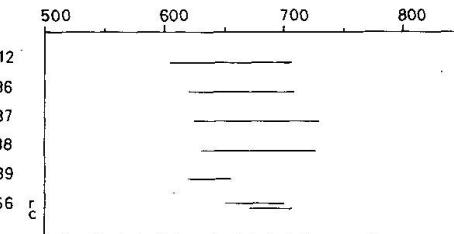


Fig. 6 Calculated temperatures from garnet-cordierite Fe-Mg exchange thermometer at 5 kbar. Calibration of THOMPSON (1976). r: rim, c: core.

36, 37, 38, 66) are about 50°C higher than rim temperatures. There are, however, exceptions (e.g. SPI 12, 39) which show the opposite trend with higher rim temperatures.

The Ca content in the garnets is rather low and can, therefore be ignored but the rather high Mn content in the garnets indicates that the data have to be interpreted very carefully (NEWTON and HASELTON 1981). Temperatures up to 800°C are, however, unrealistically high, because these rocks do not contain the assemblage orthopyroxene + quartz and the calculated phase relations also contradict such high temperatures. Temperatures calculated according to the calibration of INDARES and MARTIGNOLE (1985) which accounts for deviations from ideality in both garnet and biotite solid solution, are about 50°C lower than the values given above according to the THOMPSON (1976) calibration and slightly decrease the scatter in the data to about 60°C. The former data are consistent with temperature estimates of 640 to 680°C from dolomitic marbles from the same area (BUCHER 1981).

At an average metamorphic temperature increasing from the south towards

the north from 675 to 725 °C, garnet-plagioclase pressures of 4 to 5 kbar (at 700 °C) have been deduced according to the calibration of ARANOVICH and PODLESSKIY (1980) as well as GHENT et al. (1979).

Plagioclase inclusions in garnet allow the deduction of garnet core pressures assuming that an aluminosilicate (sill or ky) was also present in the assemblage when the garnet cores formed. This assumption may be valid because there are rocks showing sillimanite patches pseudomorphous after kyanite, which indicates that kyanite was present at an earlier stage of metamorphism. Kyanite relicts have also been reported from neighbouring areas (OHTA 1978). Some samples with a suitable assemblage (e.g. SPI 66) indicate retrograde garnet growth with core pressures of up to 2 kb higher than garnet rim pressures at decreasing metamorphic temperatures (see above). However, there are two samples (SPI 12, 39) indicating garnet growth with increasing temperature but there is no information from inclusion assemblages about the pressure under which the garnet cores formed.¹

In contrast to equilibrium 3 which is strongly temperature dependent, the assemblage cordierite + garnet + sillimanite + quartz represents an excellent geobarometer. The water content of cordierite, however, plays an important role in the positioning of equilibrium 4. Assuming dry cordierite, using the calibration of NEWTON and WOOD (1979), the regional pressures are between 4 and 5 kbar, which is consistent with the calibration of ARANOVICH and PODLESSKIY (1980) and GHENT et al. (1979). Assuming H₂O saturated cordierite the pressures are between 5 and 6 (-7) kbar which is consistent with the calibration of THOMPSON (1976).

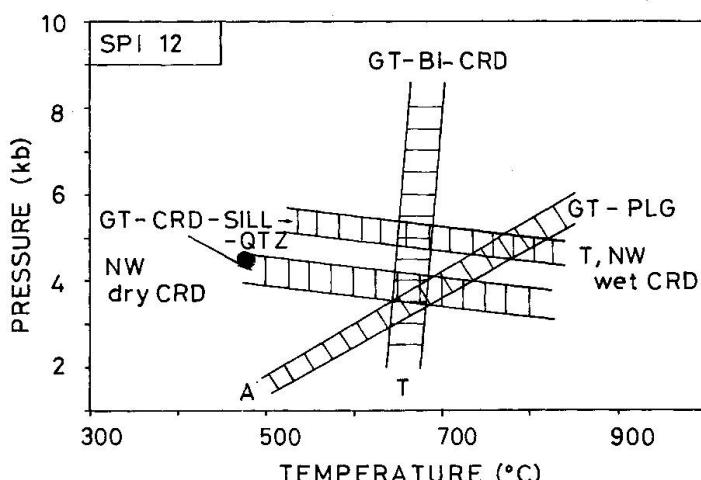


Fig. 7 Calculated pressures from the assemblages garnet + plagioclase according to ARANOVICH and PODLESSKIY (1980) (A) and garnet + cordierite + sillimanite + quartz according to THOMPSON (1976) (T) and NEWTON and WOOD (1979) (NW) for sample SPI 12. Temperatures given according to THOMPSON (1976).

¹ The pressure change indicated by the garnet zoning has been calculated with the program PTPATH (written by F. Spear) and suggests a decompression with heating path of metamorphism for these samples.

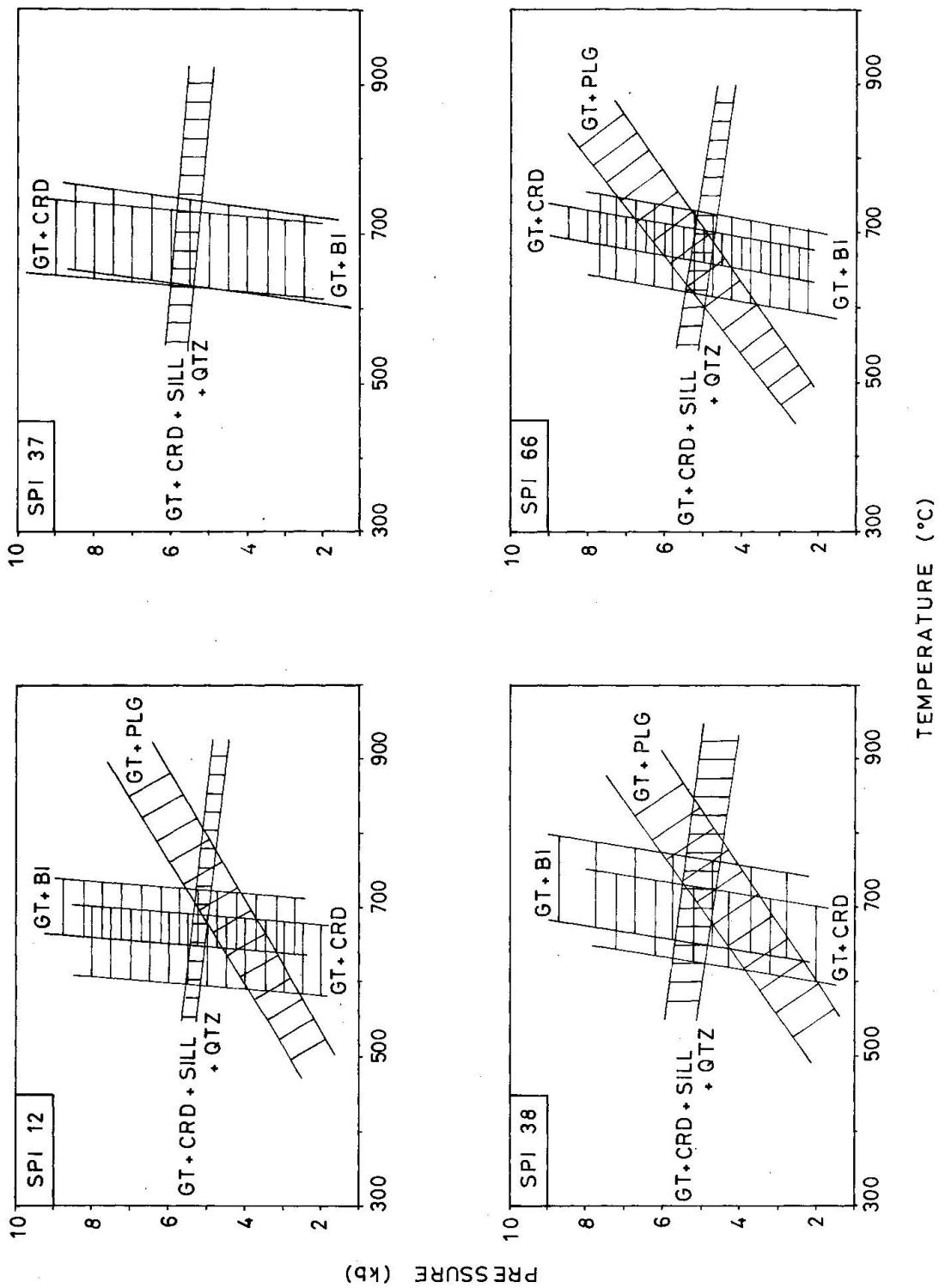


Fig. 8 Summary of thermobarometric data for four different samples. The calibrations of THOMPSON (1976), NEWTON and WOOD (1979, H_2O saturated cordierite), ARANOVICH and PODLESSKIY (1980) have been used.

Coexisting K-feldspar-plagioclase pairs in sample SPI 38 revealed solvus temperatures of 720 °C at 5 kbar using the sanidine calibration of STORMER and WHITNEY (1977). Microcline temperatures are 80–90 °C higher.

Temperatures calculated from calibrations 1 and 2 and pressures calculated from equilibrium 3 and 4 are represented in Fig. 7 for one typical sample (SPI 12).

The results for the thermobarometry are summarized on Fig. 8 for four of the samples separately. The pressure-temperature conditions have been plotted and permit the selection of a reasonably well defined PT-range for the last supposed chemical equilibration of the area studied.

Interpretation of textures and mineral reactions

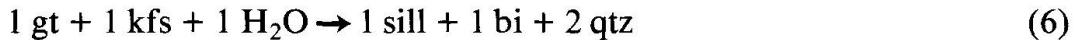
A commonly observed textural relation involving cordierite + partially resorbed garnet + sillimanite + quartz suggests that cordierite does not represent a cogenetic mineral with garnet. The observed strong resorption of retrograde garnet and its replacement by cordierite can be explained as the result of the reaction



The equilibrium conditions of this cordierite-producing reaction have been used in the previous section as a geobarometer and are shown on Fig. 9 for sample SPI 38 forming a low grade boundary to the estimated PT-conditions.

All reactions calculated in this section are based on the data of HELGESON et al. (1978). The free energy data of almandine, hercynite and Fe-cordierite given in DROOP and BUCHER (1984) as well as their activity models (BUCHER and DROOP, 1983) were used for the calculations.

The observed assemblage garnet + K-feldspar + sillimanite + biotite + quartz indicating simultaneous growth of biotite and sillimanite replacing garnet can be explained by the reaction



The retrograde reaction 6 leading to the simultaneous formation of sillimanite + biotite is shown on Fig. 9 to give a low temperature boundary to the independently inferred PT-conditions. The reaction curve is calculated for sample SPI 12 and $a_{\text{H}_2\text{O}} = 1$. The assumption for $a_{\text{H}_2\text{O}} = 1$ is based on two observations: only PT-estimates using wet cordierite give a consistent PT-box (Fig. 9) and, secondly, the position of the reaction curve indicates that all H_2O stored in a probable melt phase has been expelled to enable re-hydration reactions.

Simultaneous growth of cordierite and biotite observed in some samples (e.g. SPI 38) may be explained by hydration reactions of the type

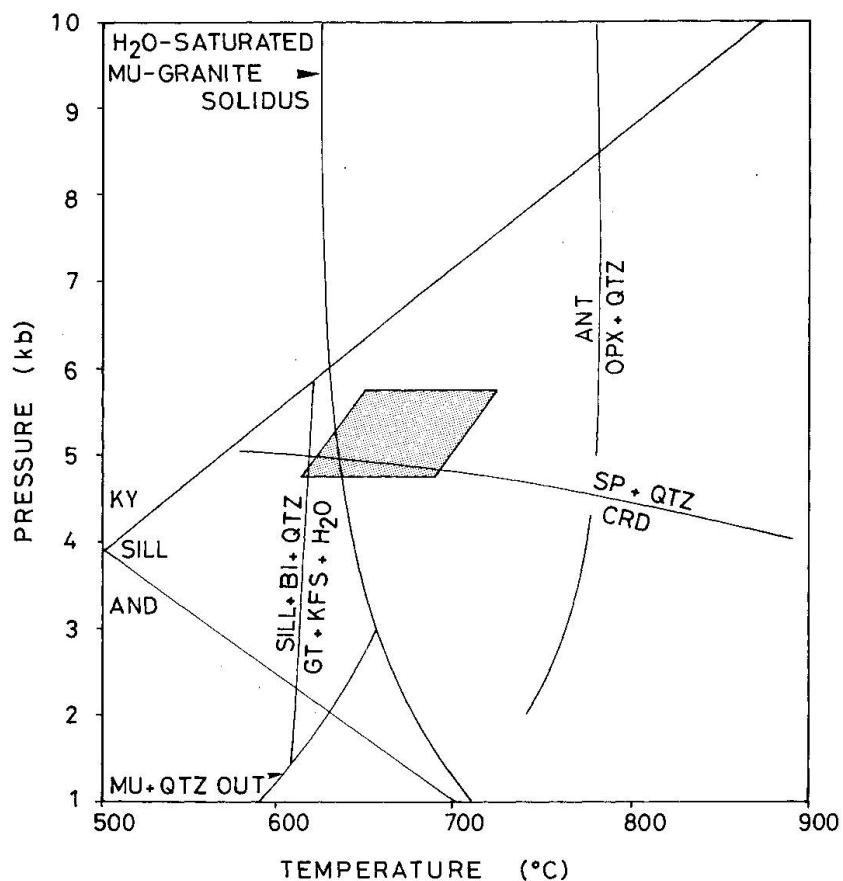
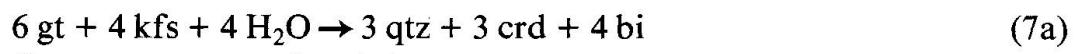


Fig. 9 Summary of the metamorphic conditions. Dashed box: PT-conditions from thermobarometry. Granite solidus after WYLIE (1977), equilibrium conditions of reactions 6 (SPI 12), 8 (SPI 38). The low temperature limit for the reaction assemblage orthopyroxene (OPX) + quartz = anthophyllite (ANT) which has been calculated using the dataset of HELGESON et al. (1978), has not been reached.



(in quartz saturated rocks)



(in quartz deficient rocks)

All these textures indicate that cordierite formation may have resulted from decompression and during a period where re-hydration reactions were active. The conclusion of decompression formation of cordierite is further supported by spinel + cordierite textures (Fig. 2a). Spinel mantled by cordierite has been interpreted as the educt mineral of reaction 8:



Cordierite formation, therefore, requires decompression and the rocks were at higher pressures prior to equilibration at PT-conditions calculated from thermobarometry. The reaction curve calculated for water-saturated cordierite passes through the determined PT-box (Fig. 9) and suggests high H_2O activity for the retrograde equilibration phase.

This interpretation is again supported by the occurrence of kyanite bearing rocks west of the studied area and of rocks with sillimanite patches pseudomorphous after kyanite (HJELLE, in OHTA, 1974a). Uncertain kyanite relicts have been observed in one sample (SPI 38). This supports the suggestion that the rocks also underwent the reaction

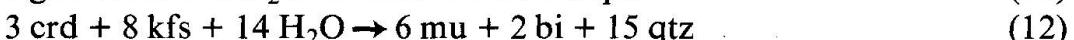
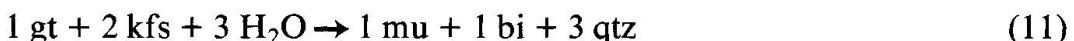


This reaction also indicates that metamorphic pressures reached higher peak values than the ones determined in this study.

The PT-conditions deduced for garnet equilibration are not consistent with the stable coexistence of muscovite + quartz. Experimental work (HOLDAWAY and LEE 1977) also demonstrates that cordierite, almandine and muscovite are not stable together. However, when almandine exists as an excess phase due to high Mn and/or Ca component in garnet these minerals may coexist (OSBERG 1971, OKRUSCH 1971). Texturally, the observed muscovite is clearly a late, retrograde mineral overgrowing cordierite or garnet and has probably formed from the following retrogressive hydration reaction



and possibly also according to the reactions



The equilibrium conditions for reaction 10 are shown on Fig. 9 labelled as mu + qtz - out curve for unit water activity.

The PT-conditions calculated from thermobarometry are above the conditions of the H_2O saturated muscovite-granite solidus which is shown on Fig. 9 according to WYLLIE (1977). Partial melting is in fact indicated by the widespread migmatitic textures in quartzo-feldspathic gneisses of NW Spitsbergen. The reactions involving K-feldspar formulated on the previous pages may thus involve a melt phase rather than solid K-feldspar. This melt phase, containing plagioclase and quartz as well as minor amounts of mafic components, probably stored a large amount of H_2O . This H_2O has been given off by the earlier prograde dehydration reactions such as dehydration of muscovite, biotite and possibly staurolite. Along the cooling and uplift path the H_2O stored in the melt was released and excess H_2O expelled from the migmatites increased the water activity as soon as the rocks came close to the muscovite-granite solidus and was then built into the newly formed minerals.

Summary and geological interpretation

Metapelitic gneisses from NW Spitsbergen typically contain cordierite + garnet + biotite + sillimanite + plagioclase + k-feldspar + quartz \pm spinel. The PT-range deduced from these rocks by thermobarometry indicate isobaric conditions of 5 to 6 kbar (assuming H_2O saturated cordierite) at a temperature between 650 to 700 °C. A slight increase in the average metamorphic temperature of about 50 °C from the south to the north can be observed.

Figure 10 indicates a possible PT-path of the rocks of the lower Hekla Hoek succession. Only a small portion of this path is well constrained. This part is represented by decompression with increasing (early garnets) or decreasing (late garnets) temperatures, respectively. An additional phase of heating may be postulated because of sillimanite fibers overgrowing partly pinitized cordierite in the northern part of the studied area. This heating can possibly be correlated with the intrusion of grey granites which can be observed prior to the Horne-mantoppen intrusion. A weak late heating related to this intrusion is responsible for late stage chloritization of biotite.

It is clear from the observed textures, relict minerals and deduced mineral reactions that the inferred conditions of metamorphism could not have been the maximum conditions experienced by these rocks and that they have endured

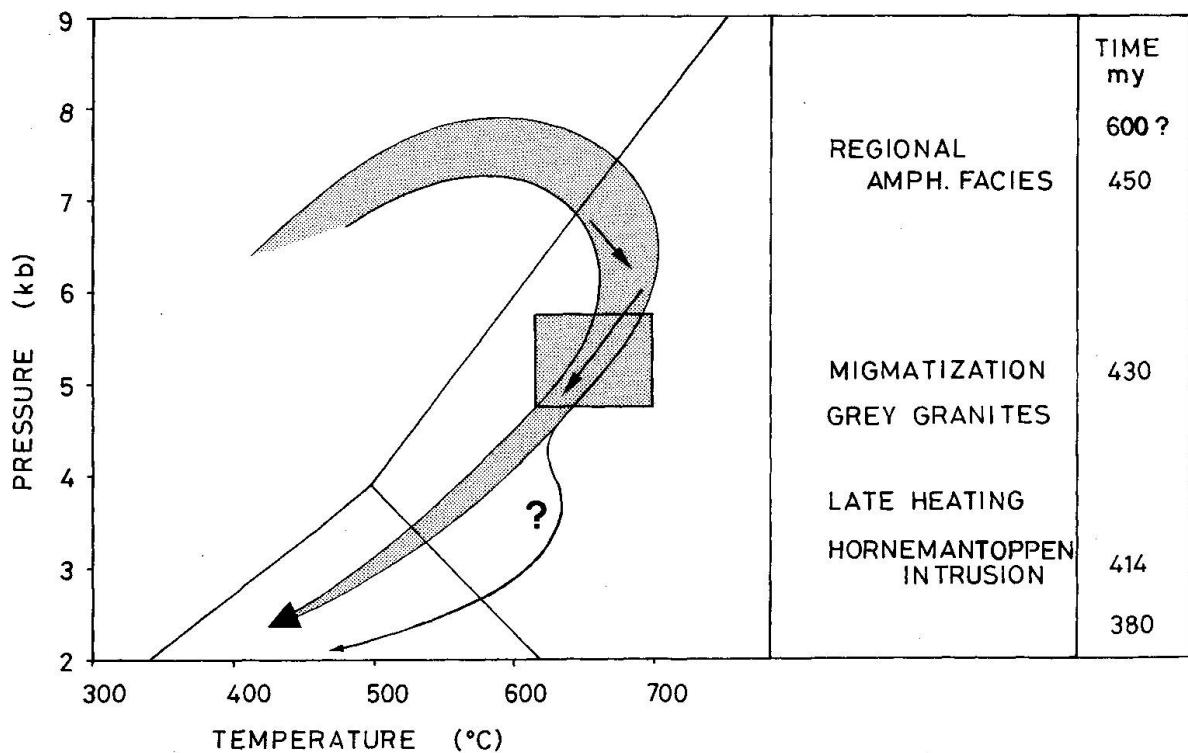


Fig. 10 Possible PT-path for metapelitic rocks of the lower Hekla Hoek formation in northwestern Spitsbergen. Isotopic age data taken from HJELLE (1979) and GAYER et al. (1966).

both, higher pressures and possibly also higher temperatures prior to the PT-conditions calculated from standard geothermobarometry. If, as suggested, the rocks have been at higher PT-conditions at some earlier stages during metamorphism, this implies that conditions inside the PT-field for granitic melt formation prevailed (Fig. 9), because thermobarometry on biotite, cordierite and plagioclase inclusions in most garnet porphyroblasts suggest a retrograde growth path almost parallel to the kyanite-sillimanite transformation curve.

At metamorphic peak conditions of the rocks, probably a low water activity prevailed. The H_2O required to make the high degree of equilibration at 5–6 kbar and 700 °C possible was presumably released by crystallizing migmatitic melts present in the quartzo-feldspathic gneisses.

Although the maximum conditions reached are unknown and earlier stages in the history of the rocks are poorly preserved, it seems reasonable to assume that conditions close to granulite facies have been reached by the lower Hekla Hoek rocks of NW Spitsbergen. It is also significant that corundum as additional petrogenetic indicator mineral has been reported by OHTA (1974b) at one locality occurring in a layered gneiss sequence in the assemblage $co + sp + crd + bi + plag + qtz$.

A metamorphic pressure of about 4 kbar at 640–680 °C was estimated for the positions of equilibrium curves and invariant points in PT-space deduced from siliceous dolomitic marble assemblages within the same area (BUCHER 1981). This pressure of 4 kbar is in agreement with pressures calculated from metapelitic rocks assuming dry cordierite. A pressure of 4 kbar is also compatible with the aluminosilicate triplepoint data of HOLDAWAY (1971). Based on the data of RICHARDSON et al. (1971), OHTA (1974b) deduced a higher pressure of 5.5 kbar at roughly the same temperature (650–700 °C) which is consistent with the calculations for wet cordierite in this study.

Acknowledgements

A travel grant by the "Schweizerische Naturforschende Gesellschaft" is gratefully acknowledged.

I thank all members of the expedition for their support. Particularly, I am indebted to B. and R. Eichin for their company in the field and to the leader of the expedition, Dr. L. Schmutz who also helped collecting the samples. I appreciate the comments on an earlier version of this manuscript by Dr. J. Ridley and Dr. G.T.R. Droop. The critical review of Dr. Y. Ohta is gratefully acknowledged.

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Manuscript received December 8, 1986; revised manuscript accepted January 23, 1987.