Metamorphism of the Müllerneset Formation, St. Jonsfjorden, Svalbard

JAY JAMES AGUE and ALAN PAUL MORRIS

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The metamorphism of upper greenschist facies metasediments exposed in the extreme southwestern portion of St. Jonsjorden, Svalbard, is described. The rocks form part of the Müllerneset Formation of the late Precambrian age Kongsvegen Group and constitute a portion of the central-western Spitsbergen Caledonides. Four deformations (D_1–D_4) and two metamorphic episodes (M_1 and M_2) have affected the rocks of the Müllerneset area. M_1 was a prograde event which was initiated prior to the onset of the D_1 and continued through this deformation. Pre-D_1 metamorphism reached biotite grade whereas garnet grade was attained syn-D_1. M_2 was a lower-middle greenschist facies metamorphism associated with D_2. The results of quantitative geothermometry in the pelitic rocks show that peak M_1 metamorphic temperatures decrease southwards across the field area from about 540°C to 510°C. Geobarometry and estimates of depth of burial indicate that M_1 pressures were in the range of 5–7 kb. The data are consistent with geothermal gradients in the range of 21 ± 4°C/km to 24 ± 5°C/km. M_2 metamorphic conditions are not precisely determinable but temperatures and pressures were probably less than those attained during M_1. It is suggested that the rocks of central-western Spitsbergen were originally deposited in an aulacogen before the initiation of Caledonian diastrophism.

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Introduction

The pre-Carboniferous rocks of Oscar II land and Prins Karls Forland, Svalbard, constitute the Western and Forland Complexes, respectively, and are believed to be northern extensions of the Appalachian/Caledonian orogen (Harland et al. 1979) (Fig. 1). A number of workers (e.g. Holtedahl 1911, 1933; Tyrell 1924; Orvin 1934; Weiss 1953; Dineley 1958; Atkinson 1960; Cutbill & Challinor 1965; Winsnes 1965; Horsfield 1969; Scrutton et al. 1976; Manby 1978a; Harland & Wright 1979; Harland et al. 1979; Hjelle et al. 1979; Ohta 1979) have made important contributions to our understanding of the general geology and tectonic evolution of the region. In this paper, we describe the metamorphism of upper greenschist facies (garnet zone) meta-sediments from Oscar II Land and discuss the tectonic implications of the metamorphic episodes recorded in the rocks. The examined lithologies comprise the Müllerneset Formation of the Late Precambrian age Kongsvegen Group (stratigraphic terminology of Harland et al. 1979).

Geologic setting

The pre-Carboniferous rocks of central western Spitsbergen may be divided into two packages. The lower package, of late Precambrian to Cambrian age, is dominantly miogeoclinal in nature and consists of glaciogenic mixtites, platform carbonates, pelites, psammites, and lesser amounts of mafic to intermediate igneous material (Harland et al. 1979). The upper package, which is exposed within Oscar II Land, represents a more active tectonic environment and comprises massive conglomerates, calcareous psammites, pelites and carbonates. This sequence is interpreted to be a syn-orogenic molasse deposit (Morris et al. 1983) and is of Siluro-Ordovician age (Scrutton et al. 1976). The Forland and Western Complexes have been affected by four deformational events (D_1–D_4), all of which are represented within the Müllerneset Formation. A major progade metamorphism (M_1) took place prior to and synchronous with D_1 and a later retrograde event (M_2) was broadly synchronous with D_2. The lower package records D_1 through D_4 and both metamorphic

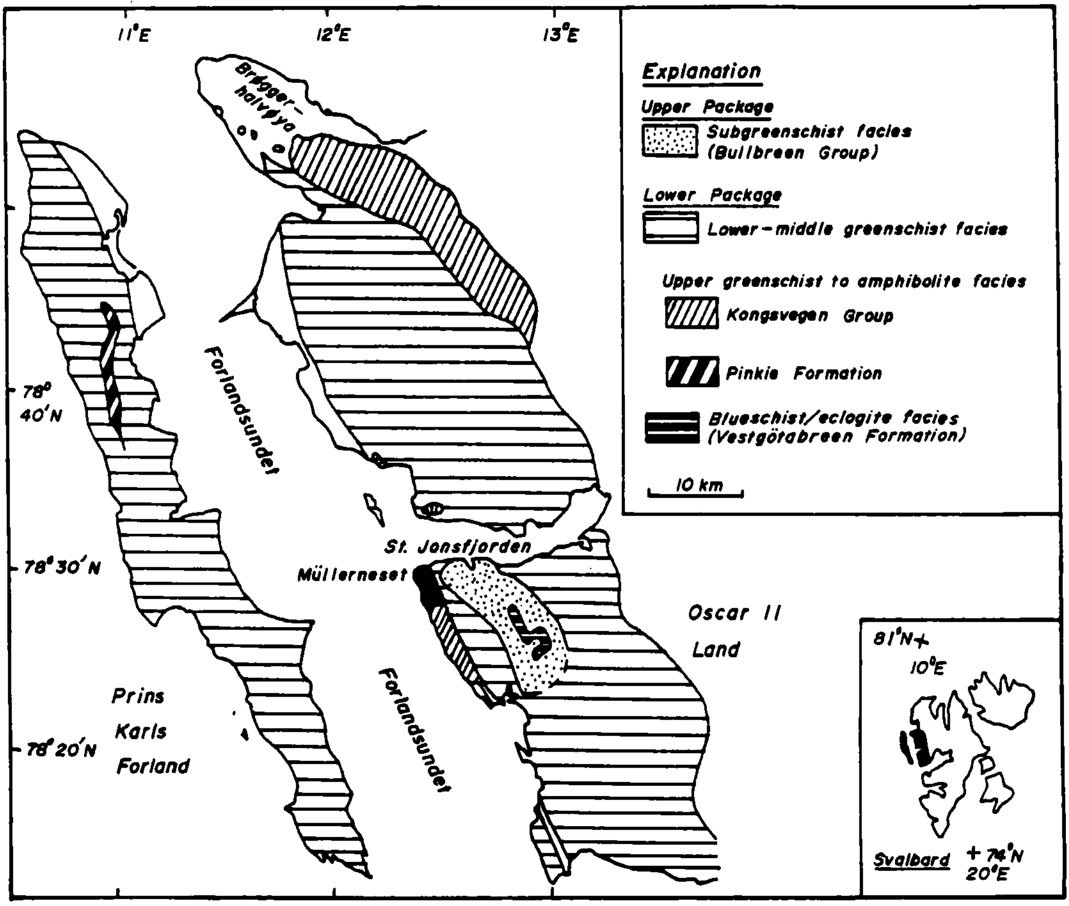


Fig. 1. Map showing the distribution of metamorphic facies within central-western Spitsbergen. Data sources: Harland et al. 1979; Hjelle et al. 1979; Manby 1983a.

events whereas the upper contains evidence of D_2, D_3, and D_4 and shows only incipient metamorphism (Fig. 1). Lithologies of the Müllerneset Formation were affected by all of the deformational and metamorphic episodes and therefore constitute part of the lower package. Although the ages of diastrophism are not yet precisely known, it is thought that D_1 and M_1 are Caledonian *sensu lato* in age whereas the later episodes are Tertiary (Challinor 1967; Morris 1979). However, recent work indicates that D_2 and M_2 may in fact be manifestations of late-stage Caledonian tectonism (Hjelle et al. 1979; Morris et al. 1983).

Garnet-biotite schists of the Müllerneset Formation outcrop in a narrow (1 × 13 km) thrust slice along Forlandsundet (Fig. 1). This work deals with the metamorphism of the northern portion of the sequence, exposed on the south shore of St. Jonsfjorden, in the vicinity of Müllerneset (Fig. 2). To facilitate study, the field area has been divided into three subareas. Pelites, semi-pelites, and psammites are the predominant lithologies, although lesser amounts of intercalated carbonate material are also present. The total thickness of the Müllerneset Formation exposed within the field area is approximately 850 m. The rocks have been subjected to four deformational events, the effects of which are summarized in Table 1. M_1 metamorphism produced mineral assemblages characteristic of the garnet zone in pelitic lithologies, although there is a decrease in metamorphic grade southwards from subarea 1 to subarea 3. The Müllerneset is in thrust contact with a group of unmetamorphosed Permo-Carboniferous limestones and dolomites (Cutbill & Challinor 1967) which outcrop in the easternmost portion of the field area. Farther east, within St. Jonsfjorden, lower package meta-sedimentary and meta-igneous rocks, metamorphosed to chlorite grade, and subgreenschist facies upper package lithologies are exposed. Five kilometres south of St. Jonsfjorden and ten kilometres inland of Forlandsundet, blueschist/eclogite facies rocks and serpentinites of the Vestgötabreen Formation outcrop (Horsfield 1969; Manby 1978b; Ohta 1979). Stratigraphic schemes for the St. Jonsfjorden region are presented in Krasil'scikov &

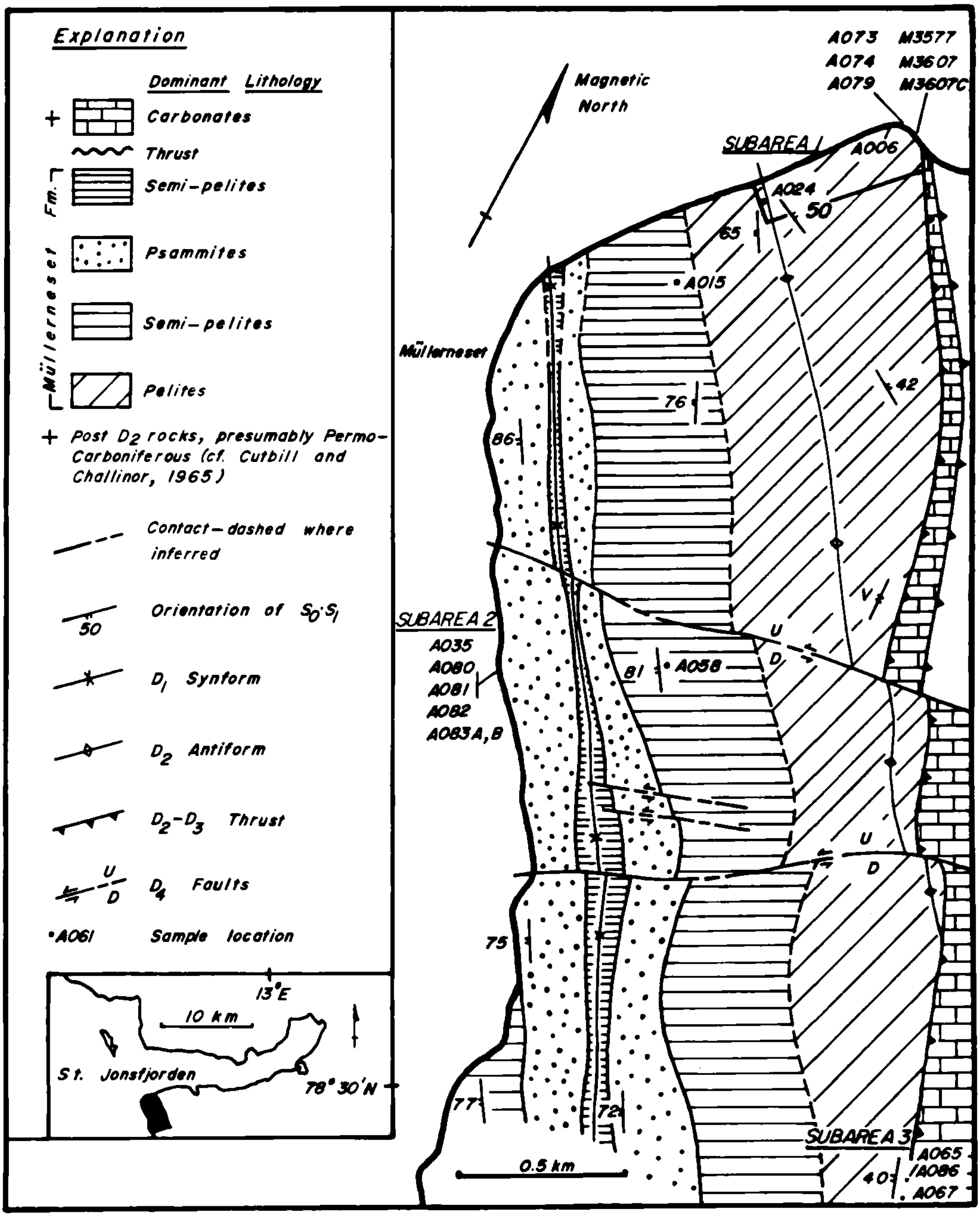


Fig. 2. Generalized geologic map of the field area.

Table 1. Deformational history of the Müllerneset Formation.

Event	Major structures	Minor structures	Age
D_1	Folds with amplitudes and 1/2 wavelengths of approximately 500 m Structural "grain" NW–SE	Penetrative schistosity (s_1) Tight folds (f_1) Fold-axis-parallel biotite lineations Quartz veins/segregations Boudinage	Caledonian
D_2	Folds with amplitudes and 1/2 wavelengths of the order of 300 m; approximately coaxial with D_1 folds High angle thrusts	Zonal crenulation cleavage (s_2) Folds with angular geometries (f_2)	Late Caledonian or Early Tertiary
D_3	Some reactivation of D_2 thrusts	Brittle shear zones	Tertiary
D_4	Normal faults	Joints	Late Tertiary?

Kovaleva (1979), Harland et al. (1979), Hjelle et al. (1979), and Waddams (1983).

Previous work

Relatively little attention has been focused on the Müllerneset Formation in recent years. The structural characteristics of the rocks have been investigated by Weiss (1953), Hjelle et al. (1979), and Ague & Morris (1983). The metamorphism of the Müllerneset lithologies has not been discussed in detail previously although Horsfield (1969), Manby (1978b), Harland et al. (1979), Hjelle et al. (1979), and Ohta (1979) have commented upon the mineral assemblages present in the rocks. Horsfield (1969), and Manby (1978b) give estimates of Kongsvegen Group metamorphic conditions which range from T = 300°C maximum (Horsfield) to T-550°C maximum and P = approximately 8 kb (Manby).

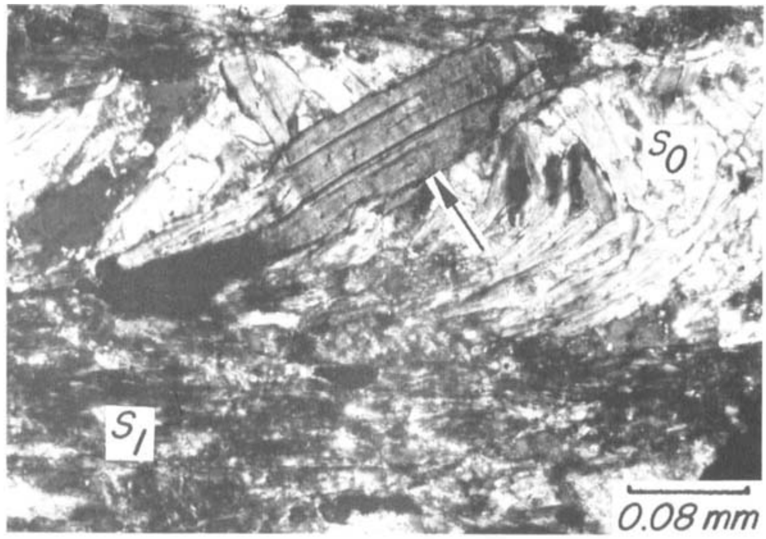
Metamorphic textures

Textural relationships between metamorphic mineral phases provide a detailed record of the timing of metamorphism and deformation within the Müllerneset Formation. Unless otherwise noted, the discussion of metamorphic textures which follows is limited to pelitic rocks because they contain the most complete record of periods of crystallization.

M_1 textures

There are three phases of mica growth preserved within the rocks. The first two are pre-tectonic whereas the third crystallized synchronous with D_1 (Fig. 3A). The s_1 schistosity is well developed in the pelitic rocks of the field area and is penetrative at hand specimen scale. Under the microscope, this foliation is seen to post-date two earlier generations of mica growth. The first fabric (s_0) could be a mimetic recrystallization of mica after the original sedimentary bedding or a relict foliation derived from an earlier (pre-D_1) deformation event. However, the lack of any other structures attributable to deformation older than those ascribed to D_1 (such as refolded folds) argues against the latter interpretation. Overprinting this early foliation is a phase of mica growth marked by randomly oriented books and decussate clots of biotite. Both of these generations of mica are deformed by the s_1 schistosity which is defined by relatively strain-free plates of muscovite and biotite aligned sub-parallel to one another. The above observations indicate that the

A.



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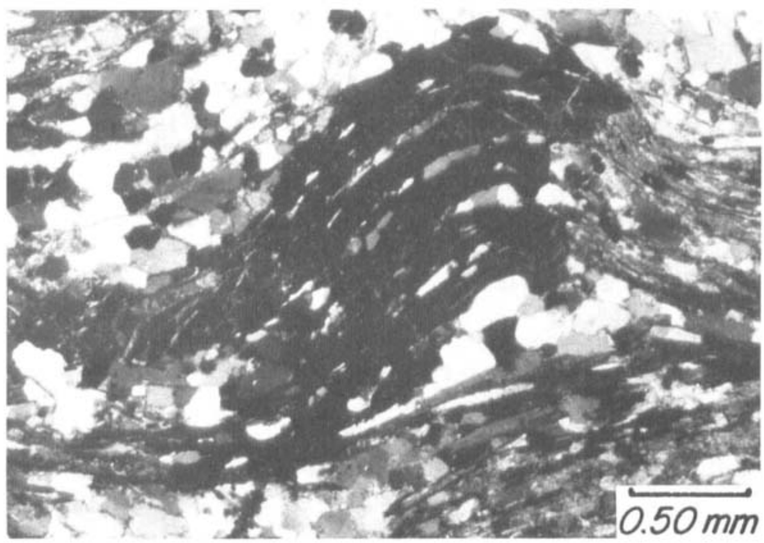

Fig. 3. Metamorphic textures. A. Three generations of mica growth preserved within the pelites. Note that the earliest foliation (s_0) has been overprinted by a later phase of biotite growth (arrow). The s_1 schistosity represents the last period of crystallization (sample M3607; NX). B. Syn-tectonic garnet (sample M3607; NX).

first two episodes of mica crystallization occurred under hydrostatic stress conditions before the onset of deformation while the third phase was syn-tectonic.

The degree and nature of s_1 development varies throughout the field area. The schists from subareas 1 and 2 are generally well differentiated into micaceous and quartzofeldspathic domains whereas those of subarea 3 are poorly differentiated. The micaceous and quartzofeldspathic layers probably developed through metamorphic differentiation processes, although the compositional heterogeneity of the sedimentary protolith may also have been important in some cases. The variations in compositional layering are mirrored by phyllosilicate grain size changes; individual mica plates are visible to the naked eye in the schists from subareas 1 and 2 (0.5–1.5 mm) but are much less obvious (except for occasional clots of biotite visible on s_1 cleavage surfaces) in the rocks from subarea 3 (<0.1–0.5 mm).

In general, garnets within the schists display syn-tectonic crystallization textures (Fig. 3B). For example, most garnets contain sigmoidal quartz inclusion trails which are indicative of syn-tectonic development (Spry 1963; Schonoveld 1978). The presence of quartz pressure shadows adjacent to some of the garnets and the wrapping of s_1 cleavage planes around the porphyroblasts imply that garnet growth may have ended before D_1 ended.

On the basis of textural evidence, the main metamorphic peak (garnet grade) is interpreted to have occurred synchronously with D_1, while biotite grade metamorphic conditions were attained prior to D_1. Pre- to syntectonic metamorphic mineral growth has also been observed in the rocks of Prins Karls Forland (Manby 1978b, 1983a, b; Morris 1979, 1982) and the Vestgötabreen Formation (Manby 1978b) although the latter also contains evidence of post-tectonic crystallization.

M_2 textures

Although D_2 was a relatively 'strong' deformation, structures attributable to this event are irregularly developed throughout the field area. In rocks which have been affected by intense D_2 deformation (as evidenced by well developed s_2 cleavage and f_2 folds) garnet-biotite pairs retrogress to chlorite ± epidote and, occasionally, garnets are completely chloritized. However, biotite by itself retrogresses only rarely. D_2 deformation also produced fractures in many of the garnets, and shear zones (e.g. Gregg 1978) in some of them. In rocks which have not suffered extensive D_2 deformation, retrograde textures are rare. These observations suggest that M_2 retrogression was more or less synchronous with D_2. The introduction of fluids necessary for M_2 retrograde metamorphism may have occurred along the many small quartz and calcite filled fractures (0.2–0.4 mm wide) within the schists. That these fractures are of D_2 age is evidenced by their cross-cutting relationship with D_1 structures, and their approximate parallelism with D_2 fold axial surfaces. Indeed, the introduction of fluids may have triggered D_2 deformation by weakening the rocks, as suggested by many workers (e.g. Fyfe et al. 1978). Whether or not M_2 actually began before D_2 is difficult to determine however.

Post-M_1 retrograde assemblages have been observed in specimens from Prins Karls Forland (Morris 1982; Manby 1983b), and within the Vestgötabreen Formation (Manby 1978b; Ohta 1979) but the retrogression is generally interpreted to have occurred synchronously with D_1 in these rocks.

Petrology

Analytical techniques

Whole-rock x-ray fluorescence analyses were performed by X-ray Assay Laboratories, Don Mills, Ontario. Electron microprobe analysis was carried out by one of us (JJA), under the direction of R. L. Barnett, on an automated Materials Analysis Company instrument at the University of Western Ontario. Between one and four analyses were performed on unzoned minerals and within subdomains of zoned phases. The results of more than one analysis of an individual grain or subdomain were averaged. Representative mineral and whole-rock analyses are presented in Appendices 1 and 2.

Mineral assemblages

The mineral assemblages present within rocks of each of the compositional groups are summarized below:

Pelites: biotite + muscovite + quartz ± almandine ± feldspar ± chlorite ± epidote ± calcite ± opaques ± graphite

Semi-pelites: biotite + muscovite + quartz ± feldspar ± chlorite ± calcite ± ankerite ± opaques

Psammites: muscovite + feldspar + quartz ± biotite ± opaques ± graphite

Carbonates: biotite + muscovite + feldspar + quartz + calcite ± almandine ± dolomite ± chlorite ± epidote ± opaques

Staurolite and the Al_2SiO_5 polymorphs are absent from all of the rocks studied.

Description of pelitic phases

The following are brief descriptions of the physical and chemical characteristics of the mineral phases present within the pelites. The discussion is limited to pelites because these rocks contain the most useful petrologic information.

Garnet. – Garnetiferous pelites which have not been severely retrograded or weathered outcrop at three locations (subareas 1–3) within the field area (Fig. 2). Garnets occur as euhedral to anhedral porphyroblasts which contain inclusions of quartz and, more rarely, feldspar, chlorite, epidote, calcite, and opaques. Garnet size decreases southwards, with average garnet diameters of 5 mm, 3 mm, and 1.5 mm for subareas 1, 2, and 3, respectively.

Compositionally, the garnets are almandine solid solutions which contain varying amounts of Mn, Ca, and Mg. In particular, the spessartine component is the most variable with X_{sp} (garnet rim) ranging from 0.05 to about 0.20. The rims of subarea 2 and 3 garnets are more Mn-rich than those of subarea 1. The unusually manganiferous garnets of subarea 2 occur within graphitic schists which contain appreciable amounts of sulfide material (pyrite and pyrrhotite). It is probable that reactions such as:

pyrite + Fe-silicates + graphite = pyrrhotite + Fe-poorer silicates + CO_2
(Carpenter 1974)

occurred within these lithologies since Mn may serve to stabilize garnet in lieu of Fe in an Fe-depleted system (Mohr & Newton 1983). Garnets from subareas 1 and 2 are normally zoned (e.g. Brown 1969) whereas those from subarea 3 are unzoned. In general, the garnets from subarea 1 schists display the best developed zoning profiles.

Other phases. – None of the other analyzed phases show any consistent, detectable compositional zoning. Plagioclase is dominantly oligoclase (An_{12-22}) although a few grains of nearly pure albite were noted in several samples. Plagioclase (and quartz) grain sizes are in the 0.1–0.3 mm range. Biotites are universally Fe-rich and muscovites generally contain an appreciable amount of phengite component. No chemical differences were noted between pre- and syn-deformational biotite and muscovite within a single thin section. Mica grain sizes range from 0.5–1.5 mm (subarea 1) to <0.1–0.5 mm (subarea 3). Chlorite occurs in samples affected by strong D_2 deformation as a retrograde mineral, replacing prograde garnet and, more rarely, biotite. In addition, two rocks from subareas 1 and 2 (M3577, A035) contain chlorite which does not display replacive textures and is thus inferred to be a prograde mineral. Both prograde and retrograde chlorites fall within the ripidolite field of Hey (1954) although retrograde chlorite tends to be more Fe-rich than prograde chlorite. It is probable that the breakdown of Fe-rich garnet produced chlorite with a relatively low Mg/(Mg + Fe) ratio (e.g. Albee 1965). Tiny (0.01–0.05 mm) grains of epidote are sometimes found associated with patches of retrograde chlorite which replace garnet. This mode of occurrence strongly suggests that the epidote is also a retrograde mineral. Calcite is quite rare within the pelites. Its most common occurrence is in late-stage quartz-calcite veins (0.2–0.4 mm wide) which transect features produced during M_1 and D_1. Ilmenite (generally altered to leucoxene in subarea 3 samples), magnetite, pyrite, and pyrrhotite are the primary metallic phases present within the rocks. Some hematite also occurs and this is interpreted to have developed through supergene oxidation of pre-existing magnetite. Graphite is found as finely disseminated dust or irregularly shaped aggregates up to about 0.2 mm across. The rocks of subarea 2 are particularly rich in graphite and sulfide material. Trace amounts of tourmaline, apatite, and zircon occur in virtually all the specimens studied.

Equilibrium criteria

Evidence for chemical equilibrium in metamorphic rocks is usually equivocal and, in addition, there is little agreement between workers as to what constitutes as 'equilibrium assemblage'. The following points suggest, but do not necessarily prove, that equilibrium was attained in the rocks of each subarea. Contacts between phases are generally quite sharp and do not show extensive alteration (except in rocks which have been strongly overprinted by M_2). In addition, a given mineral phase does not change appreciably in composition within a single thin section.

AFM projections (Thompson 1957) of garnet (rim)-biotite pairs for each of the three subareas show a systematic displacement of tie-lines which strongly suggests that samples from individual subareas crystallized in equilibrium with each other (Fig. 4). However, the composite plot of all tie-lines (Fig. 4) displays a number of crossing tie-line relationships which indicates that physical conditions and/or values of chemical potentials of projected components were variable throughout the field area during M_1 metamorphism. The diagrams also show that chlorites displaying replacive (retrograde) textures are more Fe-rich than the non-replacive variety.

Garnet producing reactions

Many reactions have been described or proposed

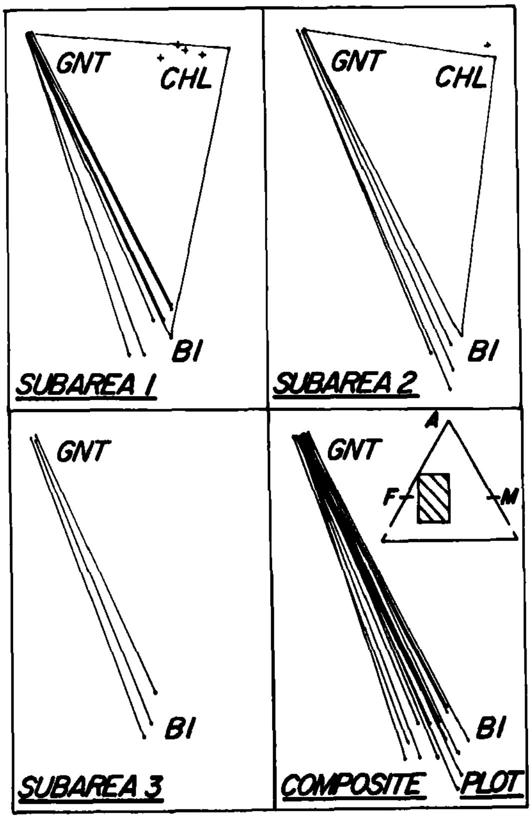

Fig. 4. AFM projections (Thompson 1957) of coexisting garnet, biotite, and chlorite. Crosses indicate retrograde chlorite (no tie-lines). Projections for rocks of each subarea are shown along with a composite plot of all tie-lines between coexisting garnet-biotite pairs. Note the crossing tie-line relationships on the composite plot.

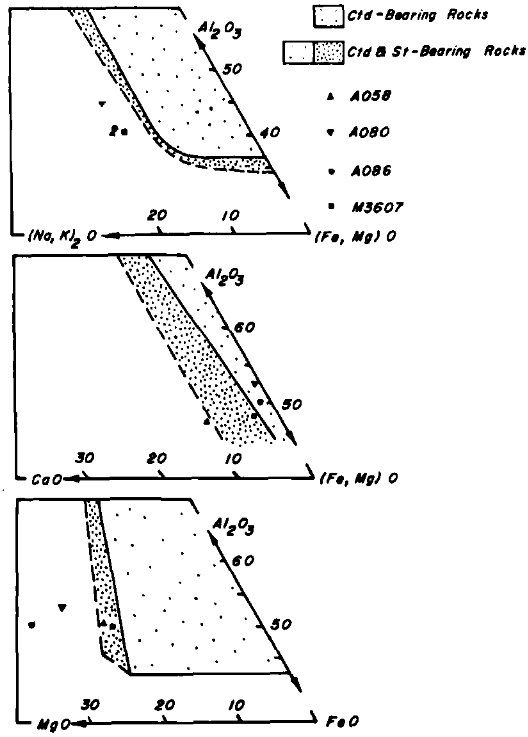

Fig. 5. Whole-rock compositions which favor the formation of staurolite and chloritoid (after Hoschek 1967) and whole-rock compositions of four 'representative' schists.

in the literature to account for the first appearance of almandine-rich garnet in pelitic rocks:

Chloritoid + Biotite + Quartz
= Garnet + H_2O (1)
(Thompson & Norton 1968)

Chlorite + Biotite + Quartz
= Garnet + Biotite$_2$ + H_2O (2)
(Chakraborty & Sen 1967)

Chlorite + Muscovite + Quartz
= Garnet + Biotite + H_2O (3)
(Thompson & Norton 1968)

Chlorite + Muscovite + Epidote
= Garnet + Biotite + H_2O (4)
(Brown 1969)

It is difficult to determine which reaction(s) operated in the pelites because the rocks lie on the high side of the almandine isograd and, therefore, assemblages transitional between the biotite and garnet zones are not present within the field area. However, some comments can be made. Reaction (1) is unlikely because whole-rock compositions of four representative schists are not suitable for chloritoid formation (Fig. 5). In addition, no evidence of chloritoid has been found in any of the rocks. Since prograde chlorite is extremely rare within the pelites, it was probably consumed in garnet generating reactions such as (2), (3), and (4). Reaction (2) is possible, although it should be noted that biotite compositions are uniform within a single thin section. It is probable that reactions similar to (3) and (4) produced most of the garnet present in the rocks. Reaction (4) is particularly attractive since it provides a mechanism for the entry of Ca (from epidote) into garnet. Epidote (and carbonate phases) may also have been utilized in reactions which produce calcic plagioclase (e.g. Crawford 1966), athough available evidence is inconclusive.

The pelitic mineral assemblages indicate that the rocks were metamorphosed at temperatures of about 500°C or greater and pressures of at least 3–4 kb at the peak of M_1 metamorphism (Winkler 1979; Turner 1981).

Thermobarometry

Since the assemblages present in the rocks only generally fix the P-T regime of metamorphism, thermobarometry was undertaken in the pelites to determine metamorphic conditions more accurately. A number of garnet-biotite thermometers are presently available (e.g. Thompson 1976; Goldman & Albee 1977; Perchuk 1977; Ferry & Spear 1978). The Ferry & Spear (1978) calibration is used here because it corrects for pressure effects, although it should be noted that the garnet-biotite Fe-Mg exchange reaction is highly insensitive to pressure. Temperature calculations were carried out at a reference pressure of 6 kb. Non-ideal solution behavior in garnet was accounted for utilizing the correction proposed by Hodges & Spear (1982). Pressures were calculated using the plagioclase-biotite-garnet-muscovite thermobarometer of Ghent & Stout (1981). Only garnet rim compositions were employed for the purposes of thermometry and barometry (cf. Pigage & Greenwood 1982).

Computed pressures and temperatures are presented in Table 2. Uncorrected temperatures are similar for subareas 1 and 2 (490°C) whereas subarea 3 temperatures are about 20°C lower. Corrected temperatures are 30–50°C higher and show a decrease in metamorphic grade southwards from subarea 1 to subarea 3. These results illustrate the importance of correcting for non-ideal solution behavior in garnet when metamorphic temperatures are calculated.

The computed temperature variations correspond well with qualitative grade indicators. As noted above, both the degree of metamorphic differentiation and phyllosilicate and garnet grain sizes increase northwards. In addition, the most Fe-rich garnets occur within the schists of subarea 1 which strongly suggests that metamorphic temperatures were higher there than elsewhere within the field area (Tracy 1982). Furthermore, the AFM projections indicate that the rocks of an individual subarea did not crystallize in equilibrium with rocks from other subareas. At present it is unclear whether the calculated temperature variations represent a gradual north-south grade change as a result of differential uplift along the east–west trending D_4 faults which cut through the central portion of the field area (Fig. 2), or

Table 2. Calculated pressures and temperatures of metamorphism.

	Sample	T°C (uncorrected)	T°C (corrected)	P bars
Subarea 1	A0006	503	556	6,338
	A024	517	548	
	A073	499	548	
	A074	455	510	
	A079	496	532	
	M3577	460	517	6,922
	M3607	472	539	
	Mean (σ)	486 (22)	536 (16)	6,630 (292)
Subarea 2	A035	492	526	5,570
	A081	460	499	5,620
	A082	503	548	
	A083A	492	523	5,827
	A083B	503	538	
	Mean (σ)	490 (16)	527 (17)	5,670 (111)
Subarea 3	A065	475	507	
	A067	484	523	
	A086	460	499	
	Mean (σ)	473 (10)	509 (11)	

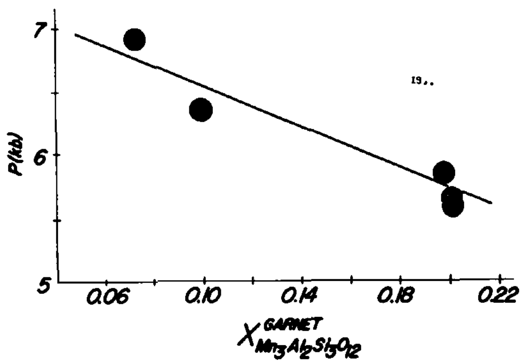

Fig. 6. Plot showing the dependence of computed pressures upon Mn content of garnet.

a southerly decrease in temperature obtaining at the time of metamorphism.

Metamorphic pressures are more difficult to determine. Since computed temperatures span a range of about 30°C, it is reasonable to infer that M_1 pressures did not vary by more than about 0.5 kb (assuming geothermal gradients of the order of 20–30°C/km). It is probable that P_{fluid} approached P_{solid} during metamorphism owing to the common occurrence of syn-metamorphic quartz-mica pods throughout the field area (e.g. Albee 1965). Application of the Ghent & Stout (1982) thermobarometer to five pelites containing coexisting plagioclase-biotite, garnet, and muscovite gives pressure estimates which range from 6.9 to 5.6 kb for the rocks of subareas 1 and 2. This thermobarometer was unsuitable for the rocks of subarea 3 because these lithologies did not contain plagioclase coexisting with garnet. Figure 6 shows that there is a strong negative correlation of calculated pressure with Mn content of garnet. Thus the computed pressures may reflect real differences in M_1 metamorphic conditions and/or the amount of Mn contained within garnet.

Harland et al. (1979) have estimated tectonostratigraphic thicknesses of slightly over 20 km obtained in central-western Spitsbergen during Caledonian orogenesis. It is reasonable to assume that the Müllerneset Formation was part of the lower-most portion of the rock package owing to its advanced state of metamorphism when compared with most of the other rocks of the orogen (Harland et al. 1979; Hjelle et al. 1979; Waddams 1983). 20+ km of rock material gives overburden pressures which approach 6 kb (crustal density = 2.65 g/cm^3; Ramberg 1967). Thus, on the basis of the above evidence, we estimate peak M_1 metamorphic pressures to have been in the range of 5–7 kb. At present it is impossible to define accurately pressure variations within the field area.

Summary and conclusions

Textural evidence indicates that the beginning of the first metamorphic event pre-dated D_1. The presence of pre-D_1 micas within the rocks strongly suggests that metamorphism to biotite grade was occurring before the onset of deformation. Burial metamorphism may have been initiated by an accumulation of 10–15 km of rock material in a sedimentary-volcanic pile. Although this material may have been deposited in a number of tectonic settings, the possibility that it accumulated in a failed rift (aulacogen; Burke & Whiteman 1973) is particularly attractive for the following reasons:

(1) A failed rift provides a deep basin for the deposition of great thicknesses of rock material.

(2) Elevated geothermal gradients, associated with crustal thinning, commonly persist in failed rifts and this would explain the early initiation of metamorphism (Royden et al. 1980).

(3) The central-western Spitsbergen Caledonides form an angle of approximately 110° with the main Caledonian trend (Birkenmajer 1981).

The peak of M_1 metamorphism, which reached garnet grade, is inferred to be synchronous with D_1. The pressures of metamorphism (5–7 kb) require between 19 and 27 km of overburden (assuming a crustal density of 2.65 g/cm^3; Ramberg 1967). This amount of material may have been built up through the continued input of sediment and volcanic material and also through tectonic thickening. At a reference pressure of 6 ± 1 kb, the various subareas yield geothermal gradient estimates which range from 21 ± 4 to 24 ± 5°C/km. These estimates are consistent with, but slightly higher than, those of Morris (1982) (18–21°C/km) and Manby (1983b) (18–23°C/km) for the middle greenschist facies rocks of the Forland Complex. However, the blueschist and eclogites of the Vestgötabreen Formation contain mineral assemblages consistent with geothermal gradients of the order of 15°C/km (Manby 1978b;

Ohta 1979). Therefore, the lithologies of the Müllerneset Formation and the middle greenschist facies rocks of the Forland Complex were metamorphosed under a different thermal regime from those of the Vestgötabreen Formation. The thermal regimes of metamorphism of the remainder of the Western Complex and the higher grade rocks of the Forland Complex remain to be investigated in detail.

The effects of M_2 retrograde metamorphism are local; generally only rocks which have been strongly affected by D_2 show pronounced retrogression. The close correlation of retrograde textures and D_2 deformation strongly suggests that M_2 occurred synchronous with D_2. These second metamorphic and deformational events appear to have been triggered by the introduction of fluids, perhaps derived from pressure solution or metamorphic dehydration reactions proceeding at deeper levels of the rock package. The pressures and temperatures of M_2 retrogression are poorly constrained but were probably less than those attained during the peak of M_1 metamorphism.

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APPENDIX 1 – *Mineral analyses*

Sample locations Appendices 1 and 2:
Subarea 1: A006, M3577, M3607
Subarea 2: A035, A080
Subarea 3: A086

GARNET ANALYSES

Oxide	A006 (Rim)	A035 (Rim)	A035 (Core)	A086	M3577 (Rim)	M3577 (Core)	M3607 (Rim)	M3607 (Core)
SiO_2	37.50	36.67	37.03	36.60	37.63	37.91	37.50	38.29
Al_2O_3	21.22	21.57	21.72	21.52	21.23	20.89	21.32	21.32
FeO	31.20	28.21	28.06	31.51	31.69	31.53	32.42	28.87
MnO	4.46	9.01	9.52	6.84	3.13	3.25	2.23	4.71
MgO	1.33	1.50	1.48	1.15	1.36	1.17	1.36	0.96
CaO	4.40	3.10	3.33	3.07	4.92	5.18	5.51	6.75
Total	100.11	100.06	101.14	100.69	99.96	99.93	100.34	100.90

				Formula (24 oxygens)				
Si	6.036	5.936	5.933	5.912	6.051	6.099	6.016	6.082
Al	4.027	4.116	4.104	4.103	4.025	3.962	4.031	3.992
Fe	4.200	3.819	3.760	4.262	4.262	4.242	4.348	3.836
Mn	0.608	1.235	1.292	0.937	0.426	0.443	0.305	0.663
Mg	0.319	0.362	0.353	0.277	0.326	0.280	0.324	0.227
Ca	0.760	0.538	0.572	0.532	0.847	0.893	0.949	1.148
				Mole Fractions				
X_{Al}	0.713	0.641	0.629	0.709	0.727	0.724	0.734	0.653
X_{Py}	0.054	0.061	0.059	0.046	0.056	0.049	0.055	0.039
X_{Sp}	0.103	0.207	0.216	0.156	0.073	0.076	0.051	0.113
X_{Gr}	0.129	0.090	0.096	0.089	0.145	0.152	0.160	0.195
$\frac{Mg}{Mg + Fe}$	0.071	0.087	0.086	0.061	0.071	0.062	0.069	0.056

BIOTITE ANALYSES

Oxide	A006	A035	A086	M3577	M3607
SiO_2	35.42	34.67	35.67	35.35	35.66
TiO_2	1.38	1.42	1.45	1.41	1.48
Al_2O_3	18.49	18.65	18.40	19.19	18.77
Cr_2O_3	—	—	—	—	0.15
FeO	23.46	22.01	23.83	21.67	23.20
MnO	0.09	0.09	0.12	—	0.07
MgO	7.29	8.94	7.37	8.05	7.95
CaO	0.04	0.16	—	0.06	—
Na_2O	0.06	0.09	0.05	0.29	0.22
K_2O	8.84	8.88	8.65	8.89	9.36
Total	95.07	94.91	95.54	94.91	96.86
		Formula (22 oxygens)			
Si	5.488	5.363	5.501	5.441	5.432
Al	2.512	2.637	2.499	2.559	2.568
ΣIV	8.000	8.000	8.000	8.000	8.000
Al	0.867	0.765	0.847	0.921	0.804
Ti	0.160	0.164	0.168	0.164	0.169
Cr	—	—	—	—	0.018
Fe	3.043	2.848	3.074	2.789	2.956
Mn	0.012	0.012	0.016	—	0.009
Mg	1.684	2.063	1.694	1.848	1.805
ΣVI	5.766	5.852	5.799	5.722	5.761
Ca	0.008	0.027	—	0.012	—
Na	0.020	0.027	0.015	0.090	0.064
K	1.750	1.754	1.702	1.746	1.820
ΣXII	1.778	1.808	1.717	1.848	1.884
$\frac{Mg}{Mg + Fe}$	0.356	0.420	0.355	0.399	0.379

MUSCOVITE ANALYSES

Oxide	A006	A035	M3577	M3607
SiO_2	49.07	49.44	47.74	46.69
TiO_2	0.29	0.32	0.25	0.29
Al_2O_3	32.21	31.95	35.76	35.19
Cr_2O_3	0.03	—	—	0.04
FeO	2.35	2.38	1.54	1.90
MnO	—	—	—	—
MgO	1.68	1.90	0.84	0.80
CaO	—	0.03	—	—
Na_2O	0.82	0.86	0.84	0.62
K_2O	9.41	9.66	9.15	9.60
Total	95.86	96.54	96.12	95.13
		Formula (22 oxygens)		
Si	6.469	6.484	6.242	6.203
Al	1.531	1.516	1.758	1.797
ΣIV	8.000	8.000	8.000	8.000
Al	3.477	3.425	3.754	3.715
Ti	0.027	0.031	0.023	0.029
Cr	0.004	—	—	0.005
Fe	0.258	0.262	0.168	0.211
Mn	—	—	—	—
Mg	0.332	0.371	0.164	0.158
ΣVI	4.098	4.089	4.109	4.118
Ca	—	0.004	—	—
Na	0.211	0.219	0.215	0.160
K	1.582	1.617	1.527	1.627
ΣXII	1.793	1.840	1.742	1.787
$\frac{Mg}{Mg + Fe}$	0.563	0.586	0.494	0.428

CHLORITE ANALYSES

Oxide	A006	A035*	M3577*	M3607
SiO_2	25.00	26.04	24.84	24.75
TiO_2	0.05	0.08	0.05	0.06
Al_2O_3	21.22	20.18	22.53	20.76
Cr_2O_3	—	0.03	—	0.06
FeO	32.05	29.23	28.71	33.38
MnO	0.25	0.02	0.15	0.25
MgO	10.41	12.74	12.55	9.28
CaO	0.04	0.02	0.06	—
Na_2O	—	—	0.03	0.02
K_2O	0.05	0.04	0.02	0.28
Total	89.07	88.38	88.94	88.84
Formula (28 oxygens)				
Si	5.376	5.547	5.248	5.392
Al	2.624	2.453	2.752	2.608
ΣIV	8.000	8.000	8.000	8.000
Al	2.757	2.617	2.859	2.723
Ti	0.008	0.013	0.008	0.010
Cr	—	0.005	—	0.010
Fe	5.765	5.208	5.072	6.082
Mn	0.045	0.004	0.027	0.046
Mg	3.337	4.046	3.952	3.012
Ca	0.018	0.010	0.014	—
Na	—	—	0.013	0.008
K	0.013	0.010	0.005	0.079
Σ	11.943	11.913	11.950	11.970
$\frac{Mg}{Mg + Fe}$	0.367	0.437	0.438	0.331

* Prograde chlorite

FELDSPAR ANALYSES

Oxide	A006	A035	A035	M3577
SiO_2	63.27	67.65	63.27	63.69
Al_2O_3	23.54	19.75	23.10	23.22
CaO	4.51	—	4.03	3.88
Na_2O	9.16	12.45	10.03	9.60
K_2O	0.07	0.01	0.10	0.13
Total	100.55	99.86	100.53	100.52
Formula (8 oxygens)				
Si	2.781	2.969	2.789	2.800
Al	1.219	1.023	1.199	1.203
Ca	0.211	—	0.191	0.183
Na	0.781	1.059	0.855	0.818
K	0.004	—	0.004	0.007
Mole Fractions				
X_{An}	0.212	—	0.182	0.182
X_{Ab}	0.784	1.000	0.814	0.812
X_{Or}	0.004	—	0.004	0.007

APPENDIX 2 – *Whole rock analyses*

WHOLE-ROCK ANALYSES

Oxide	A058*	A080	A086	M3607
SiO_2	62.75	45.32	61.80	61.99
TiO_2	0.98	1.66	1.01	0.92
Al_2O_3	17.12	26.14	17.54	17.13
FeO**	6.40	8.31	7.31	7.13
MnO	0.06	0.08	0.06	0.08
MgO	2.86	3.41	2.42	2.60
CaO	1.97	0.25	0.45	1.24
Na_2O	2.31	1.16	2.22	2.10
K_2O	3.97	7.67	3.63	3.86
P_2O_5	0.10	0.17	0.14	0.17
L.O.I.***	1.86	5.13	2.72	2.33
Total	100.38	99.30	99.30	99.55
$\frac{Mg}{Mg + Fe}$	0.44	0.42	0.37	0.39

* Non-garnetiferous semi-pelite
** All Fe as FeO
*** Loss on ignition