Geology and metamorphic evolution of the Roan area, Vestranden, Western Gneiss Region, Central Norwegian Caledonides

CHARLOTTE MÖLLER

Möller, C. 1988: Geology and metamorphic evolution of the Roan area, Vestranden, Western Gneiss Region, Central Norwegian Caledonides. *Nor. geol. unders. Bull. 413,* 1-31.

In the gneiss area of northern Vestranden, kilometre-scale upright domes and basins refold tight and isoclinal folds of banded amphibolite facies orthogneisses and supracrustal rocks. In the Roan area, a major culmination in its core exposes a window, the Roan Igneous Complex (RIC). The RIC is made up of quartz-monzonitic to quartz-monzodioritic gneisses, charnockitic and basic rocks, which are cross-cut by granites, and still younger dolerite dykes. Mafic rocks of Caledonian age form large layered complexes. The RIC is partly remarkably well preserved from the intense Silurian to Devonian deformation and amphibolitization, and retains both primary intrusive relationships and granulite facies parageneses. The RIC was overthrust by the granitic and intermediate orthogneisses, banded amphibolites and supracrustal rocks of the Banded Gneiss Complex (BGC), which now occur interleaved and folded above the RIC. The supracrustal sequences of the BGC are made up of paragneisses, mafic rocks, marbles and calc-silicate rocks.
The rocks within the RIC and the BGC experienced similar tectonothermal evolutions. During the Palaeozoic, Scandian, continent-continent collision, the rocks were depressed to depths of c. 50 km where they equilibrated at peak temperatures of 800-870°C. Locally preserved high-pressure granulite facies tectonites indicate that the thrust zone between the RIC and the BGC was active during this event, in response to telescoping and depression of the western edge of Baltica. The initial stage of uplift involved a near-isothermal decompression into medium-pressure granulite and upper amphibolite facies conditions. Subsequent deformation caused extensive retrogression at amphibolite facies conditions, in part coeval with migmatitization. The rocks then cooled through the andalusite stability field. The preservation of the high-pressure granulite facies assemblages and the symplectite reaction textures, combined with the near-isothermal uplift suggest a rapid tectonometamorphic evolution. The thermal evolution during uplift, and the intense deformation and thrust reactivation at amphibolite facies conditions support a tectonic model where the uplift of the WGR is aided by tectonic unroofing.

Charlotte Möller, Department of Mineralogy and Petrology, Institute of Geology, University of Lund, Sölvegatan 13, S-22362 Lund, Sweden.

Introduction

In western Norway, large-scale culminations expose deep structural levels of the Caledonide orogen. Precambrian igneous rocks form the cores of the deepest windows. These basement rocks are folded together with rocks of supracrustal origin which, at least in part, represent Caledonian allochthonous cover. The Western Gneiss Region (WGR, Fig. 1) southwestern Norway forms the deepest exposed structural level of the Scandinavian Caledonides. The northern part of the WGR, called Vestranden (Kjerulf 1871, Ramberg 1943, 1966, Birkeland 1958), passes from Kristiansund northwards to the Foldereid-Vikna area (Fig. 1).

The southern and central parts of the WGR (south of Kristiansund) are most famous for the occurrences of eclogites and garnet peridotites. Since the region which hosts these high-pressure rocks is considered to be the basement beneath the Caledonian thrust sheets, these rocks will help constraining the tectonic models for the Caledonide orogen as a whole. Important factors are: the structural and tectonostratigraphic position of the high-pressure rocks, the timing of the high-PT metamorphism, the peak-PT conditions, and the PTt-evolution and mechanism for their uplift.

The first known occurrences of similar, ba-

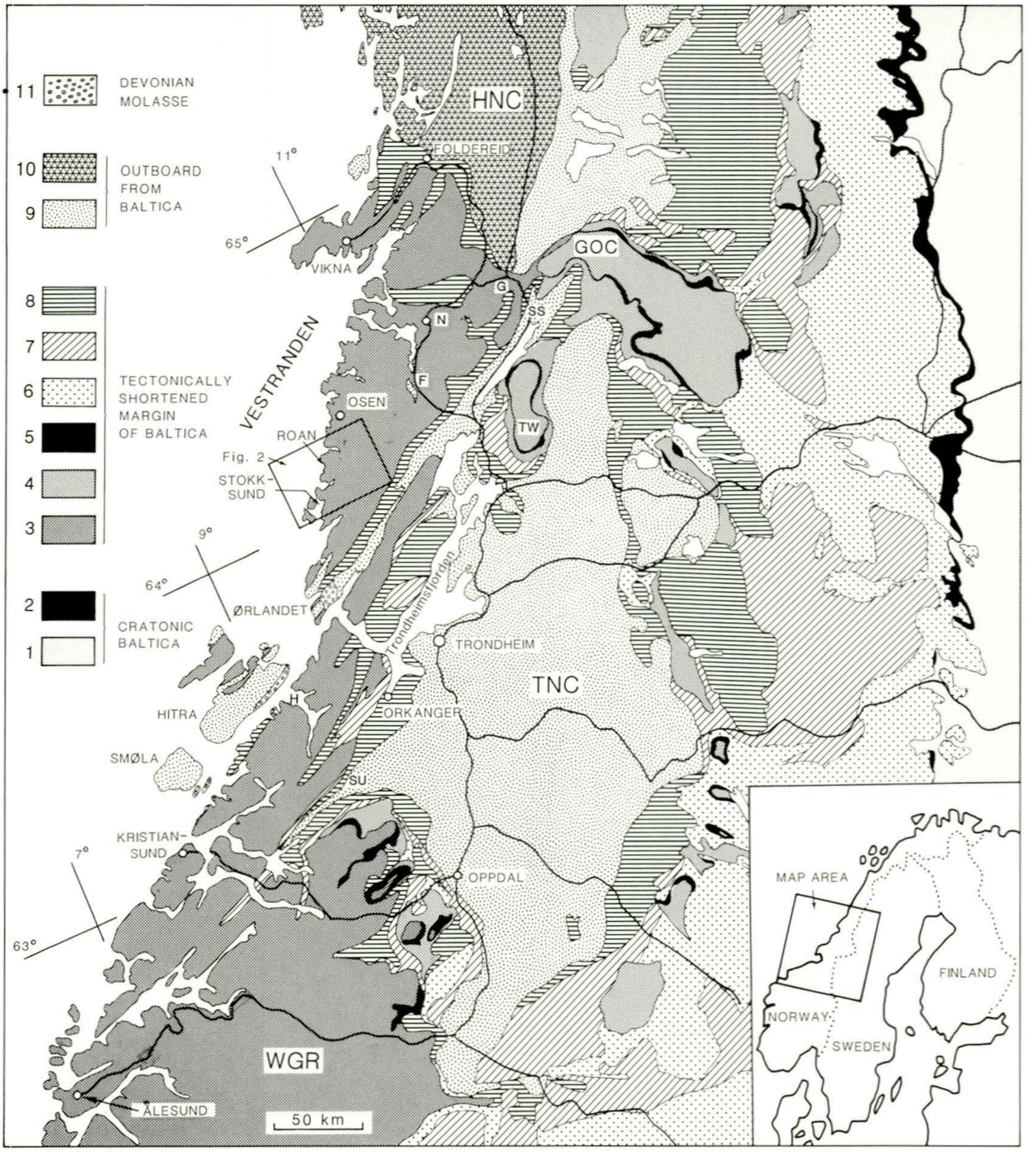

Fig. 1. Geological map of the central Scandinavian Caledonides, simplified from Gee et al. (1985a) with additional information from Sigmond et al. (1984) and Johansson (1986a). Legend: 1. Precambrian crystalline rocks of Baltica; 2. Thrust front: autochthonous platformal Upper Proterozoic to Lower Palaeozoic sedimentary rocks; 3. Western Gneiss Region and Vestranden: undifferentiated Baltoscandian Precambrian crystalline rocks and supracrustal rocks; 4. Windows: Baltoscandian Precambrian crystalline rocks; 5. Windows: platformal Upper Proterozoic to Lower Palaeozoic sedimentary rocks; 6. Miogeoclinal to platformal Upper Proterozoic to Lower Palaeozoic sedimentary rocks; 7. Miogeoclinal dyke-intruded Upper Proterozoic sedimentary rocks and Baltoscandian Precambrian crystalline rocks; 8. Suspect terranes thought to comprise the miogeocline and continent-ocean transition zone: metabasaltic, metasedimentary and Precambrian crystalline rocks; 9. Exotic eugeoclinal terranes: Lower Palaeozoic oceanic sequences of metavolcanic and metasedimentary rocks; 10. Exotic terranes (Laurentia?): continental lithosphere; 11. Late Silurian(?) to Devonian Old Red Sandstone sediments. Name abbreviations. F- Fosslia; G- Geitfjellet; GOC- Grong-Olden-Culmination; HNC- Helgeland Nappe Complex; H- Hemnefjord; N- Namsos; TNC- Trondheim Nappe Complex; TW- Tømmerås Window; SS- Snåsa Synform, SU- Surnadalen; WGR- Western Gneiss Region.

sic, high-pressure rocks in the WGR north of Trondheimsfjorden were found within the Roan peninsula (Fig. 1). From aerial photographs it was evident that the Roan peninsula is structurally different from the intensely banded and folded surrounding gneisses (Fig. 2). Geophysical maps also suggested a difference, with a positive magnetic and gravimetric anomaly centered around Roan. The Roan area was selected for a study of the following aspects: (1) The PT evolution of the high-pressure basic rocks and its significance; (2) the intrusive, structural and metamorphic relationships between the basic high-pressure rocks and their various country rocks; (3) the structural and metamorphic relationships between the rocks within the Roan peninsula and the immediately surrounding gneisses and metasedimentary sequences. The rocks in the Roan area were found to have in part escaped the amphibolite facies deformation which is so pervasive elsewhere in Vestranden. Intrusive relationships, early high-grade parageneses and early structures are locally very well preserved. The Roan area therefore provides an excellent opportunity to unravel the tectonometamorphic evolution of a deep structural level in the Caledonide Orogen. The aims of the present paper are to describe the intrusive and structural history in the Roan area and to relate the thermal evolution to the progressive deformation.

Tectonic setting of the Western Gneiss Region

The geological evolution of the WGR involve several rock-forming episodes and orogenic events extending in time from the Mid Proterozoic to the Palaeozoic (Gorbatschev 1985, Griffin et al. 1985, Kullerud et al. 1986). The origin of the so famous eclogite and garnet peridotites has been the subject of debate for a long period of time. The debate has largely been focussed on whether the eclogites were formed 'in situ' by prograde metamorphism, or whether they are 'foreign' in relation to their host gneisses (Smith 1980, Cuthbert et al. 1983, Griffin et al. 1985 and references therein).

Most eclogites in the WGR give Silurian mineral ages of around 425 Ma (Krogh et al. 1974, Griffin & Brueckner 1980, 1985, Mearns & Lappin 1982, Kullerud et al. 1986, Mørk & Mearns 1986); however, Precambrian mineral ages have also been obtained (the Almklovdalen garnet peridotite south of Ålesund, Mearns 1986). There is evidence that at least some eclogites were formed by prograde metamorphism, and most of them are considered to have been formed during Caledonian, regional, eclogite facies metamorphism of the WGR (Mørk 1985, Griffin et al. 1985, Griffin 1987 and references therein). Tectonic models relate this event to continental subduction of the margin of Baltica during collision with Laurentia (Krogh 1977, Cuthbert et al. 1983, Griffin et al. 1985). The metamorphic array of peak temperature conditions of the eclogites in the southern and central WGR show an increase towards the northwest (Krogh 1977). Eclogites just south of Kristiansund record pressures of 18 to 20 kbar and temperatures up to c. 800°C (Griffin et al. 1985).

In relation to the Caledonian thrust sheets, the Precambrian crystalline rocks of the WGR have generally been treated as parautochthonous basement (Fig. 1; Gee et al. 1985a). During the Silurian the WGR, including Vestranden, was overthrust by units transported from source regions further west. The lower thrust sheets are derived from the western marginal parts of Baltica. Amongst other rocks, they consist of dyke-intruded miogeoclinal Late Precambrian psammites (commonly called the Särv), overlain by amphibolites, schists and Precambrian crystalline rocks of various metamorphic grades (the Seve). Some of the thrust sheets carry high-grade granulite or eclogite facies rocks. Sm-Nd datings of eclogite facies minerals from rocks in the Seve thrust sheet in northern Sweden give Early Palaeozoic ages (c. 505 Ma, Mørk et al. 1988), while Precambrian mineral ages are recorded for granulite facies rocks within the overthrusted Bergen Arcs in southwesternmost Norway (Austrheim & Griffin 1985). The uppermost thrust sheets in the mountain chain comprise units derived from west of Baltica. These include Early Palaeozoic, low-grade, eugeoclinal, island-arc complexes (e.g. the Köli; the Trondheim Nappe Complex, TNC, Fig. 1) and outboard, possibly Laurentian terranes of schists, gneisses and granites (e.g. the Helgeland Nappe Complex, HNC, Fig. 1; Gee et al. 1985a, b, Roberts & Gee 1985, Stephens et al. 1985, Gee 1986, Hossack & Cooper 1986).

The sequence of thrust sheets in the eastern parts of the Caledonide orogen (Gee 1975, Gee & Zachrisson 1979, Gee et al. 1985b) features structural and metamorphic breaks between various superposed nappes (Andréasson & Gorbatschev 1980). The same sequence of thrust sheets has been traced westwards into the WGR where these allochthonous cover units are recumbently and polyphasally interfolded with the basement rocks (Gee 1980, Krill 1980, 1985, Tucker 1986). Here, metamorphic discontinuities and structural breaks are partly erased, and in places both basement and cover were metamorphosed to high-grade conditions (Griffin et al 1985, A. Krill pers. comm. 1987). The Caledonian cover sequence in Surnadalen can be traced to the outermost west coast north of Ålesund (Fig. 1; Krill & Sigmond 1987). However, different types of supracrustal sequences occur interfolded with the Precambrian gneisses of the WGR, and most of them have an unknown age and tectonostratigraphic position (Bryhni 1988). The detailed relationships between the metamorphic and structural evolution of the various basement and cover units within the WGR and Vestranden are as yet poorly known.

During uplift, the WGR experienced extensive retrogression at amphibolite facies conditions. The thermal evolution during uplift is less well described than the early high-PT conditions, and different retrograde trends have been suggested (Cotkin et al. 1988 and references therein). Krogh (1977, 1980, 1982) and Bryhni et al. (1977) proposed an isothermal uplift of the WGR. An early decompression during heating towards the metamorphic maximum, consistent with thermal relaxation during uplift, was inferred from the Eiksundal eclogite south of Ålesund (Jamtveit 1987). The cause of the uplift of the WGR has been discussed by Cuthbert et al. (1983), who proposed that it was controlled by a deep-seated major thrust fault located in the low velocity zone of Mykkeltveit et al. (1980). Griffin et al. (1985) suggested that the uplift was largely due to erosional removal of the overlying plate, and Jamtveit (1987) also suggested an uplift controlled by erosion. On the basis of titanite and zircon U-Pb-dates, Tucker et al. (1987) proposed that the uplift of the Western Gneiss Region may have been assisted by tectonic unroofing.

Tucker et al. (1987) interpreted the time of regional metamorphic resetting and cooling below 500°C in the Orkanger—Hemnefjord area (just southwest of Trondheimsfjorden) to c. 395 Ma. During the late stages of uplift the regional, NE-SW trending, upright synforms and antiforms formed.

In the Late Silurian(?) and Early to Mid Devonian, molasse was deposited in fault-controlled basins in a broadly extensional regime with associated major detachment faults (Hossack 1984, Norton 1987). A subsequent Late Devonian deformation phase has been proposed (Roberts 1983, Torsvik et al. 1986).

The geology of northern Vestranden

In its northern part, Vestranden is overlain structurally by Caledonian allochthonous rocks including the outboard Helgeland Nappe Complex (Fig. 1). In the easternmost part of Vestranden there is a thrust between the Vestranden basement gneisses and the underlying Precambrian granitic and rhyolitic gneisses of the Grong-Olden Culmination (Fig. 1; Johansson 1986a). There are lithological, structural and metamorphic differences between these two basement areas (op. cit.). They are both overlain by far-travelled Caledonian thrust sheets.

The dominant rock-types in Vestranden north of Trondheimsfjorden are Precambrian granitic to intermediate gneisses of plutonic origin. These orthogneisses are, in part, tightly interlayered and folded together with amphibolites and supracrustal rocks. The rocks are refolded on a regional scale into major domes and basins (Ramberg 1943, 1966, Birkeland 1958). The southern part of Vestranden, i.e. the area around and southwest of Trondheimsfjorden, is marked by a consistent ENE-WSW structural trend of the regional fold structures. In addition to basement gneisses, this area comprises allochthonous Caledonian cover units such as e.g. low-grade island-arc igneous and metasedimentary rocks, and Devonian Old Red Sandstone deposits on Ørlandet (Fig. 1. Sigmond et al. 1984, Tucker 1986). The Møre-Trøndelag Fault Zone with prominent, steep ENE-WSW lineaments is located here. The Møre-Trøndelag Fault Zone is considered to be a complex strike-slip fault zone, active from Late Devonian to Jurassic time (Grønlie & Roberts 1988). Towards the north, the marked ENE-WSW structural trend gradually gives way to more irregular domal structures (Roberts 1986).

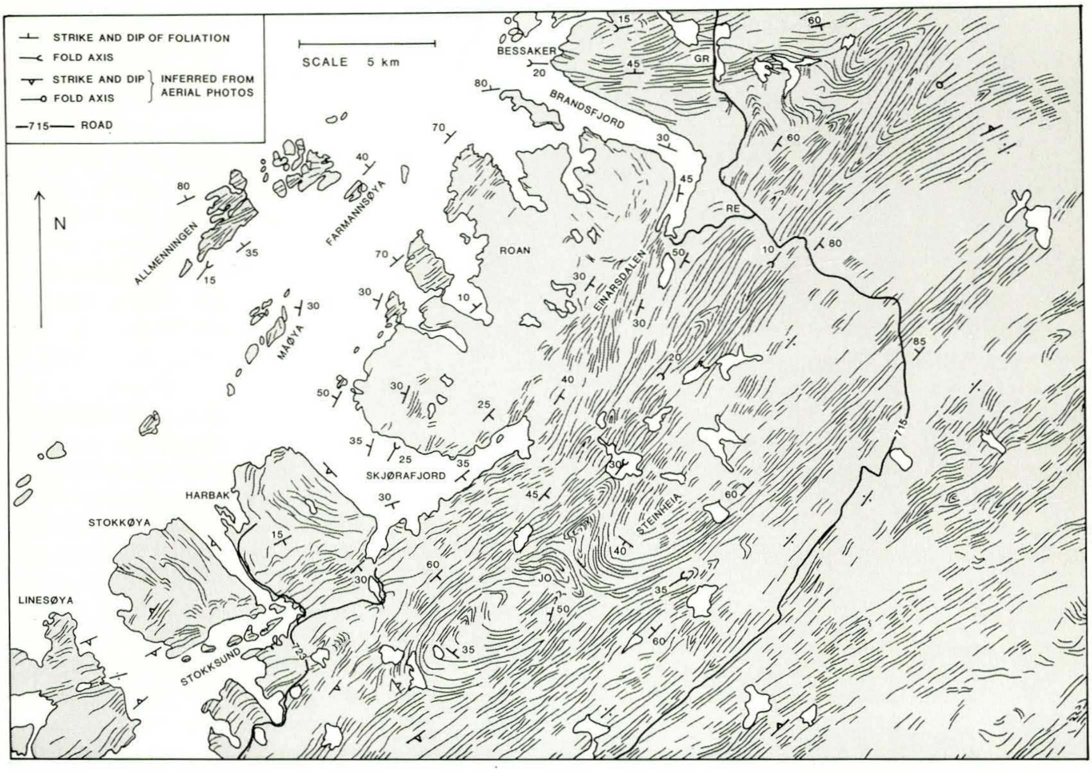

Fig. 2. Regional structures in the Roan-Stokksund area. The traces of banding and foliation are interpreted from aerial photographs. JO- Jotjørnheia, RE- Reppkleiv, GR- Granholvatnet lens. Parts of the structural data are taken from NGU's preliminary 1:50,000 geological maps.

Geological mapping (1:50,000) of Vestranden north of Trondheimsfjorden has started recently (Boyd 1986). Very little detailed petrological or structural work has yet been published. The results of recent and ongoing work in this area are summarized briefly below, though the Roan area is treated in more detail in the following sections.

Protolith ages, interpreted from radiometric zircon U-Pb upper intercept data, for the plutonic rocks in northern Vestranden range from around 1820 Ma (Geitfjellet granite: Johansson 1986a, tonalite gneiss in the Foldereid-Vikna area: Schouenborg et al. in prep.) to c. 1630 Ma (granite at Osen, Schouenborg et al. in prep.). In the Roan area, granites similar to that at Osen are cut by younger dolerites.

The Caledonian thrust sheets can be traced westwards into Vestranden from the Snåsa Synform (Fig. 1; Andréasson & Johansson 1983). In the Snåsa Synform, tectonic breaks and metamorphic discontinuities between the different thrust sheets and the basement still persist. Further west, such differences are to a great extent erased by the regional penetrative amphibolite facies deformation. Recent mapping by the Geological Survey of Norway has revealed a wide extent of supracrustal rocks (Boyd 1986). A supracrustal sequence at Fosslia (Fig. 1) was correlated with the Gula Nappe of the TNC (Johansson et al. 1987a). The thrust emplacement of the Fosslia cover sequence post-dates c. 430 Ma (age of dyke intrusion, op.cit.). Schouenborg (1986, 1988) proposed the existence of allochthonous cover sequences in the northernmost part of Vestranden, in the Foldereid-Vikna area (Fig. 1). The time for the emplacement has been bracketed to between 450 and 400 Ma. For the subsequent, regional, NE-SW trending upright fold phase a maximum age of c. 400 Ma has been obtained (Schouenborg 1988). The existence in northern Vestranden of autochthonous cover of Proterozoic or Early Palaeozoic supracrustal rocks has not been demonstrated but cannot yet be excluded.

Prior to amphibolite facies metamorphism and deformation, the metamorphic conditions

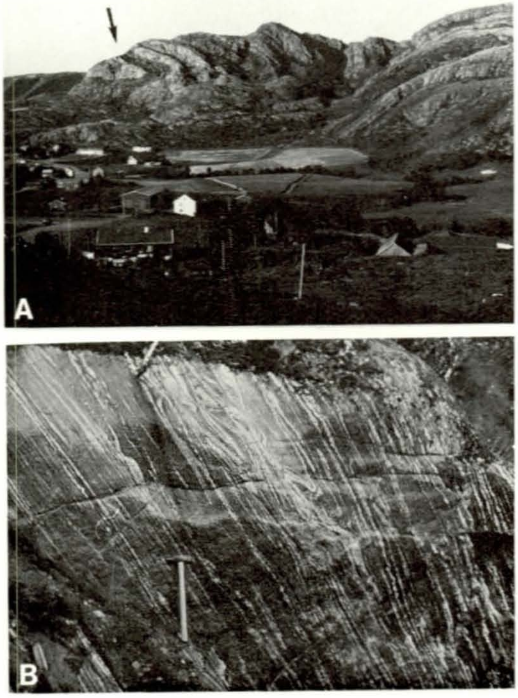

Fig. 3. (A) Megascale isoclinal fold (arrow) of banded orthogneisses, amphibolites and supracrustal rocks, Bessaker. The photo is taken looking towards the east-northeast. (B) Migmatite gneiss at Reppkleiv (RE in Fig. 2), with hammer (55 cm) for scale.

in western Vestranden (the Roan area) reached high-pressure granulite facies (Johansson & Möller 1986). A Sm-Nd mineral isochron suggests a Silurian age of c. 425 Ma for this event (Johansson et al. in prep.). The regional extent and distribution of this very high-grade metamorphism is not yet known in detail, but relics of granulite facies rocks are found throughout the area west of Namsos and south to Trondheimsfjorden (L. Johansson pers. comm. 1983). A probable, major, NW-SE trending metamorphic discontinuity occurs northwest of Namsos (R. Boyd pers. comm. 1987), where the very high-PT rocks are confined to the area to the southwest.

General geology of the Roan—Stokksund area

The regional folds in the Roan—Stokksund area (Figs. 1 & 2) form spectacular patterns with large-scale, NE-SW trending domes and basins, earlier refolded folds, and sheared-out fold limbs. The deepest structural level, the Roan Igneous Complex (RIC) which geographically coincides with the Roan peninsula (Fig. 2), is exposed as the core of a culmination. The rocks above the RIC will, in this paper, collectively be referred to as the Banded Gneiss Complex (BGC).

The Banded Gneiss Complex is composed of migmatitic orthogneisses, thick banded amphibolites, and supracrustal rocks. The orthogneisses are granitic to intermediate in composition, and contain smaller pods of basic rocks such as amphibolites, metagabbros and metadolerites. The orthogneisses form both the cores of large culminations which have less banded and deformed central parts (e.g. Steinheia, Fig. 2) and sheets concordant with the banded amphibolites and supracrustal rocks. Refolded isoclinal folds with cores of migmatitic orthogneiss are found e.g. at Jotjørnheia (JO, Fig. 2) and Bessaker (Fig. 3a). The supracrustal sequences and the banded amphibolites are generally less than 1 km thick but may be continuous for tens of kilometres. The banded amphibolites are found in close proximity to supracrustal rocks or 'solitary' within the orthogneisses. They are dark with lighter layers of more plagioclase-rich compositions, and in places contain lenses of ultramafic rocks. They grade into lighter-coloured, intermediate compositions. The supracrustal rocks consist mostly of garnet-bearing biotite-rich schists and gneisses with or without kyanite and/or sillimanite, hornblende-biotite-quartz gneisses, amphibolites, and to a lesser extent marbles and calc-silicate rocks. Graphite-bearing and sulphide-rich rocks are found locally.

The regional foliation is subparallel to the compositional banding, and is defined by amphibolite facies minerals and concordant migmatite veins (Fig. 3b). The compositional banding and the migmatite veins are locally isoclinally folded, and boudinage structures are common. Stretching and intersection lineations are subhorizontal to gently plunging with a NE-SW trend in most parts of the area shown in Fig. 2. The degree of deformation varies over short distances, but in general the intensity of the deformation is strongest at the contacts between different rocks. Relatively well preserved rocks in places form lenticular bodies, which can vary in size from the micro-

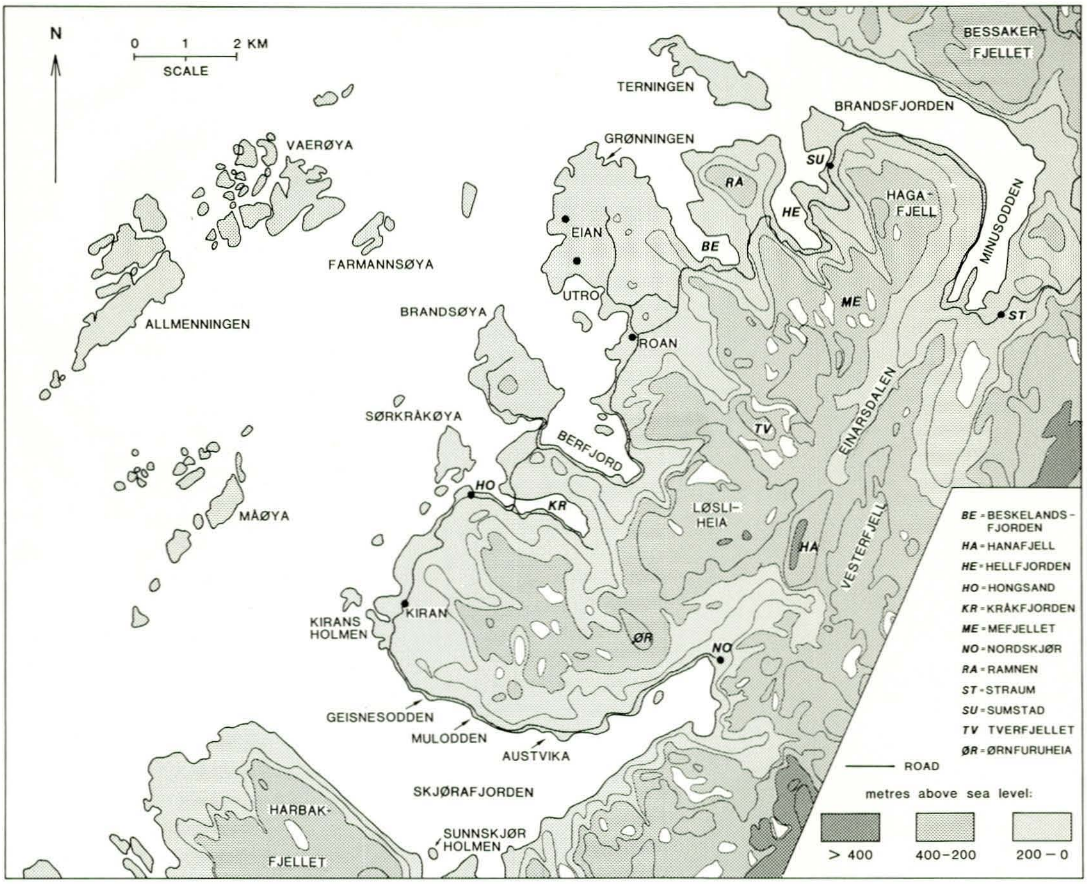

Fig. 4. Location map of the Roan area. Topographic contour interval = 100 metres.

scale to lengths of some kilometres. The gneisses in the Granholvatnet Lens (GR, Fig. 2) exhibit early structures, such as weakly deformed cross-cutting migmatite veins (Johansson 1986b). Granulite facies parageneses are preserved sparsely throughout the Roan—Stokksund area, most abundant in competent mafic rocks.

The rocks within the RIC are better preserved than in the overlying BGC. The RIC is made up of quartz-monzonites, quartz-monzodiorites, granites, charnockitic and basic rocks. Intrusive relationships, early granulite facies parageneses and early structures are still recognizable. These relationships are treated in more detail in the following sections.

The RIC is enveloped by a deformation zone at the contact to the BGC. The contact zone contains two thin sheets of supracrustal rocks, separated by a sheet of orthogneisses. These two lowermost sheets of supracrustal rocks within the BGC can be traced along all exposed parts above the RIC (Figs. 4, 5 & 6). In this paper they are treated in some detail and are referred to as the Einarsdalen Supracrustal Unit (ESU). The structural geometry implies that the RIC forms a window.

The culmination that exposes the RIC — or the Roan Window — as the core, is geophysically characterized by a positive gravimetric and magnetic anomaly (Fig. 7). Geophysical maps suggest that the RIC continues below the present exposed level southwestwards to Linesøya. The rocks within the RIC have a susceptibility high enough to explain the observed positive magnetic anomaly, whereas

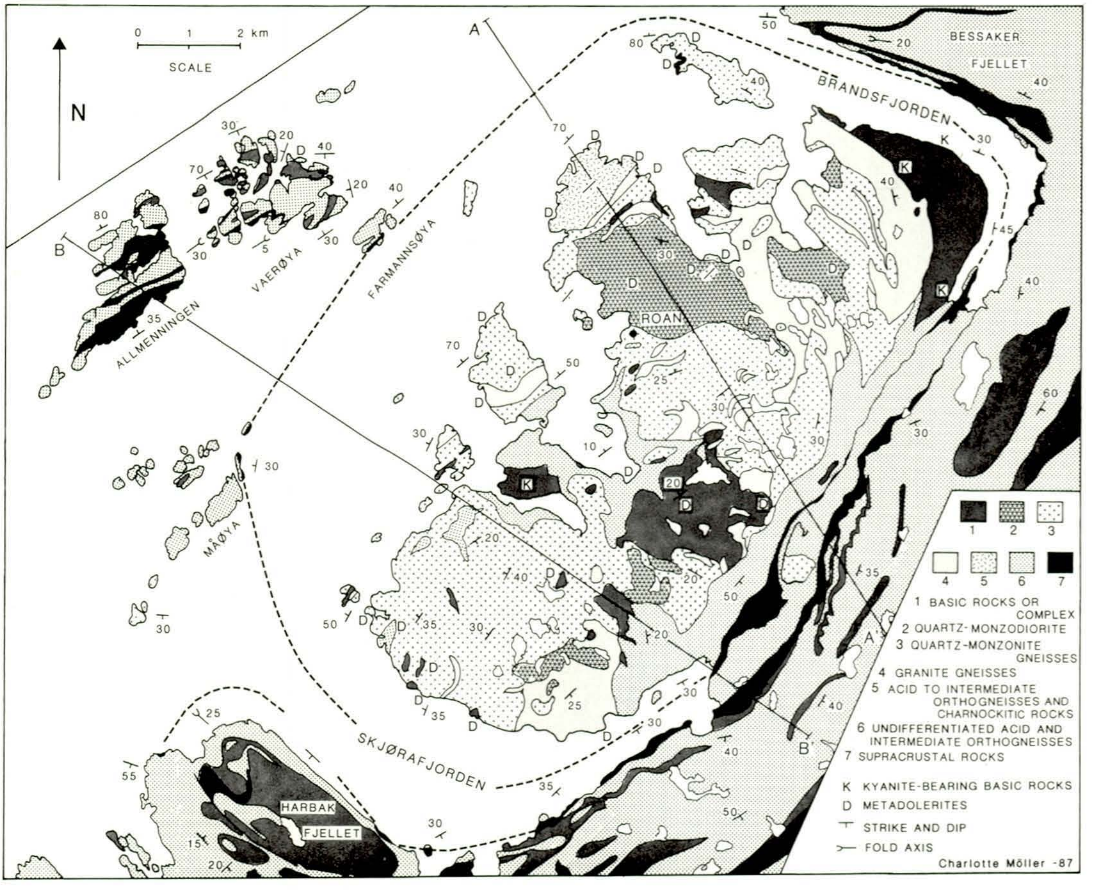

Fig. 5. Lithological map of the Roan area. The area south of Skjørafjord is in part taken from NGU's preliminary 1:50.000 geological maps. The dashed lines mark the probable continuations of the supracrustal rocks (the Einarsdalen Supracrustal Unit), where the innermost line represents the boundary between the Roan Igneous Complex and the Banded Gneiss Complex. The lines A-A' and B-B' mark the cross-sections in Fig. 6.

the overlying supracrustal rocks and orthogneisses have lower susceptibilities (Skyseth 1987). On the basis of petrophysical data, Skyseth (op.cit.) suggested that the RIC extends down to a depth of between 14 and 16 km.

Lithology

The different rock-types in the Roan area are described according to the legend of the lithological map (Fig. 5). Locations are shown in Fig. 4. The description below is divided into three sections: igneous rocks within the Roan Igneous Complex and the Banded Gneiss Complex, the Einarsdalen Supracrustal Unit, and pegmatite dykes.

Igneous rocks within the Roan Igneous Complex and the Banded Gneiss Complex

Thin-sections were stained for feldspar identification according to the methods of Broch (1961) before point-counting (1000 points/sample). The relative proportions of the minerals in the rocks of the RIC are shown in Table 1 and Fig. 8. The mineral compositions given below are based on EDX-analyses made at the Institute of Geology, University of Lund.

Quartz-monzonite to quartz-monzodiorite gneisses

The most common rock-type within the RIC is a medium-grained grey quartz-monzonite (Fig.

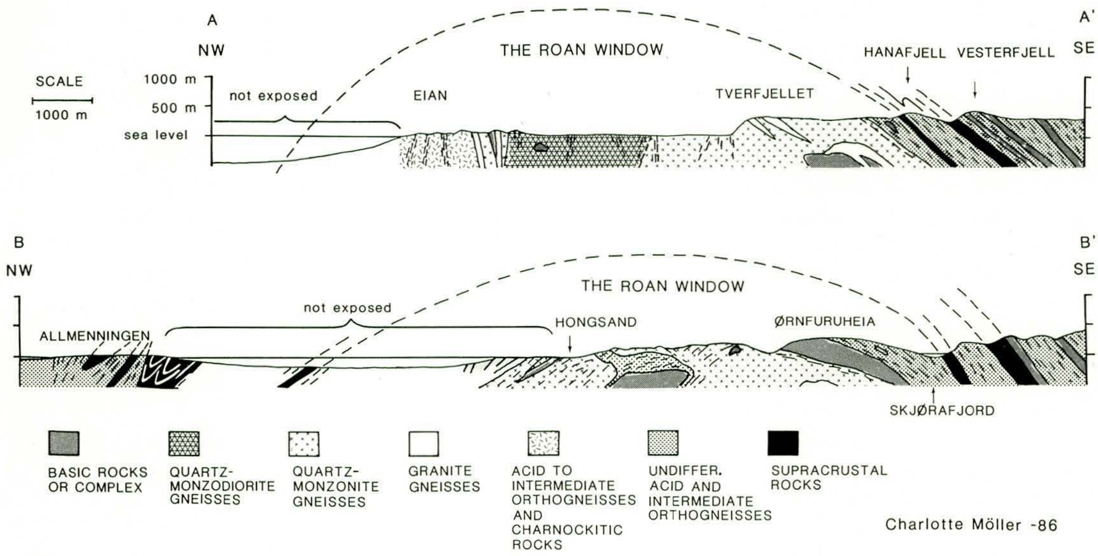

Fig. 6. Profiles across the Roan culmination.

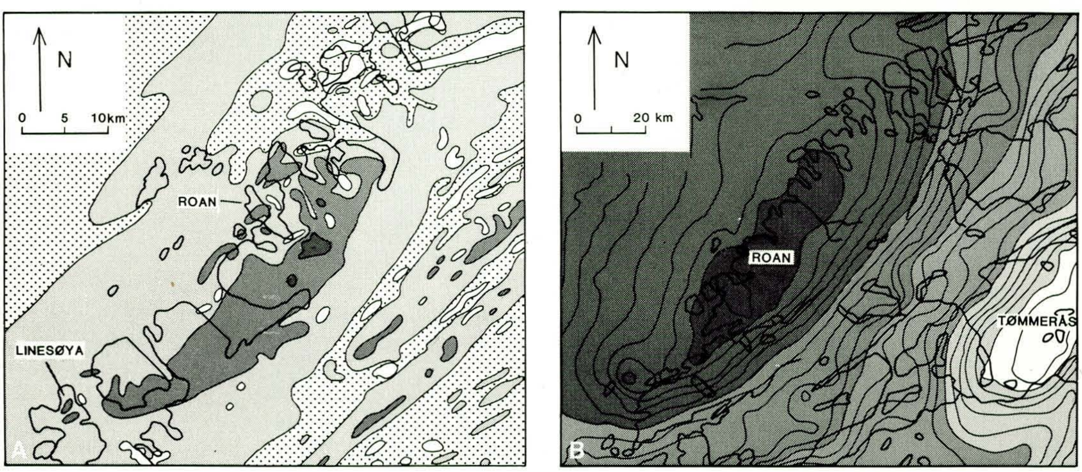

Fig. 7. Geophysical maps of the Roan-Stokksund area. (A) Aeromagnetic map. Contour interval - 500 gamma. Maximum—more than 51 700 gamma. White—local minima; (B) Bouguer anomaly map. Contour interval—5 mGal. Maximum- 45 mGal, minimum- -70 mGal. Simplified from Norges Geologiske Undersøkelse 1983 and 1985, respectively.

5). The rocks mapped as quartz-monzonites grade continuously into quartz-monzodiorites and, in some places, to grey or reddish-grey granites (Fig. 8, Tables 1 & 2). In the field, the latter compositional difference could not be mapped out, which is why this rock-type will be referred to simply as quartz-monzonites. The quartz-monzodiorites were in the field distinguished from the quartz-monzonites by their greater modal-% of dark minerals and the lack of K-feldspar visible in hand specimen. The boundaries between quartz-monzonites and quartz-monzodiorites marked on the map (Fig. 5) are gradational. Relatively strongly deformed, more fine-grained, heterogeneous, migmatitic grey gneisses dominate south of Roan, around Hongsand and Kiran. They have gradational contacts to the homogeneous, medium-grained quartz-monzonites.

The quartz-monzonites and quartz-monzodiorites are composed of feldspars + quartz + hornblende + biotite + clinopyroxene + garnet + sphene + magnetite + Fe-sulphide + apatite + zircon ± scapolite. The rocks are inequigranular. Feldspars show abundant perthitic textures and intergrowths, and myrmekites are common along rims of large feldspars. Antiperthite hosts have a composition An_{25-32} while the exsolved K-feldspars contain a 12-19 mole-% albite-component. Clinopyroxene and garnet commonly occur as relics.

The quartz-monzodiorites are in places massive or only weakly foliated. Completely undeformed varieties occur at Mefjellet and north of Roan in the central parts of the large quartz-monzodiorite massif. Elsewhere, nearly all quartz-monzonites and quartz-monzodiorites have a gneissosity defined by amphibolite facies minerals. The gneissosity is commonly folded by decimetre-scale open to tight folds separated by an axial-planar spaced foliation. Migmatite veinlets, with or without hornblende, are oriented along these axial surfaces. In the veins, the antiperthite host has a composition c. An_{25} and contains around 40 volume-% K-feldspar exsolutions, while the K-feldspar contains c. 20 mole-% albite-component. Some hornblende-bearing veins may also contain biotite ± scapolite ± minor opaques, zircon and garnet. In general, the quartz-monzodiorites are less migmatitic than the quartz-monzonites.

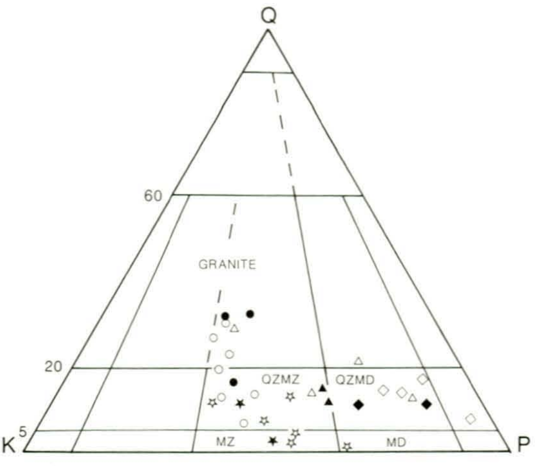

Fig. 8. Modal compositions of plutonic rocks within the Roan Igneous Complex, based on counting 1000 points per thin-section, plotted in a Streckeisen (1976) diagram. Abbreviations in the composition fields: QZMZ— quartz-monzonite, MZ— monzonite, QZMD— quartz-monzodiorite, MD— monzodiorite. The samples were selected to cover the variations within the rock units mapped as: circles—granite gneisses; stars—charnockitic rocks; triangles—quartz-monzonite gneisses; diamonds—quartz-monzodiorites. Filled symbols are modal compositions of the analysed rocks G1-G8 and G12 in Table 2.

Granite gneisses

The rocks mapped as granite gneisses are red, reddish-grey, light grey or yellowish, fine- to medium-grained rocks, and grade into quartz-monzonitic compositions (Fig. 8, Tables 1 & 2). Granite gneisses are mixed with grey quartz-monzonites in the area around Beskelandsfjord and Hellfjord. Clear cross-cutting

Rock mapped as:	LIGHT GRANITES	DARKER GRANITES	QUARTZ-MOZONITES	QUARTZ-MONZODIORITES	CHARNOCKITIC ROCKS
QUARTZ	12-32	7-22	11-29	5-12	1-12
K-FELDSPAR	36-49	42-48	11-41	3-18	31-46
PLAGIOCLASE	24-31	28-38	27-59	43-52	25-61
GARNET	<1	<1	0-1	1-13	0-11
PYROXENES	<1	<1	0-3	1-16	0-5
HORNBLENDE	0-1	2-4	0-11	0-16	0-11
BIOTITE	1-3	2-4	1-10	3-17	1-9
SCAPOLITE	<1	<1	0-2	0-5	<1
OPAQUES	1-3	1-2	1-2	0-2	1-3
OTHER	0-1	<1	<1	0-1	<1
NUMBER OF SAMPLES	5	5	6	6	8

TABLE 1. Mineral contents (%) in rocks from within the Roan Igneous Complex, based on point-counting of thin-sections (1000 points/sample).

relationships can be seen, however, where dykes and veins of light red granite cut the quartz-monzonites, quartz-monzodiorites and also some amphibolites. Very fine-grained, aplitic granites are abundant among the charnockitic rocks in the Eian-Grønningen area. With the exception of aplitic rocks, the granite gneisses generally contain migmatite veins.

The granite gneisses are composed of perthitic K-feldspar ± microcline + plagioclase + quartz + biotite + Fe,Ti-oxides + allanite + zircon + apatite ± hornblende ± sphene ± garnet ± clinopyroxene. The medium-grained quartz-monzonitic varieties commonly contain hornblende. Feldspar-quartz intergrowths are abundant, and myrmekite rims K-feldspar. The plagioclase is An_{7-17}. The microperthite has a *bulk composition with c. 14-25 mole-% albite*-component. Allanite and Fe,Ti-oxides are locally abundant, and may be surrounded by bright red spots or lamellae of Fe-oxides in the neighbouring feldspars.

The granitic rocks are mostly gneissic with foliations defined by amphibolite facies minerals. Decimetre-scale folds with a spaced axial surface foliation are present only locally. There are folded contacts between quartz-monzonites and granites, which carry migmatite veinlets located along the axial surfaces of the folds. Mylonitic foliations are more common in the granitic gneisses than in the more competent quartz-monzonites and quartz-monzodiorites.

Charnockitic rocks

Charnockitic rocks are abundant in the area around Eian-Grønningen. They also form a few small lens-shaped pods within the granitic gneisses at Hagafjell. In the field, the charnockitic rocks are recognized as acidic to intermediate plutonic rocks containing conspicuous grey or brown orthoclase megacrysts. They are medium- to coarse-grained, generally lack foliation and have a yellowish, greenish-grey or brownish lustre. The charnockitic rocks are quartz-monzonites to monzonites in composition (fig. 8, Tables 1 & 2).

The charnockitic rocks are made up of K-feldspar + plagioclase + quartz ± clinopyroxene ± biotite ± hornblende ± garnet ± orthopyroxene + magnetite + Fe,Ti-oxides + apatite + zircon. K-feldspar, plagioclase and clinopyroxene form megacrysts. Perthites and antiperthites are common, as are intergrowths between the feldspars. Myrmekite is seen along the rims of K-feldspar. In monzonitic varieties, plagioclase is approximately An_{25}.

In places, the orthoclase megacrysts are zoned with respect to the amounts of exsolved plagioclase which vary between c. 15 and 35 volume-%. The orthoclase host contains around 25 mole-% albite component, and the plagioclase exsolutions have a composition c. An_{25}. The zoning defines idiomorphic crystal planes, and therefore presumably has a magmatic origin. Clinopyroxene inclusions were found, though only locally, in these orthoclase crystals.

Garnets commonly form coronas around mafic minerals. The coronas and other reaction textures will be described elsewhere (Möller *in prep.*).

The charnockitic rocks are generally unfoliated and devoid of migmatite veins. Where a gneissosity is developed, the charnockitic rocks pass gradually into 'normal' reddish-grey and grey medium-grained gneisses containing microcline instead of orthoclase, and hornblende and biotite instead of pyroxenes and garnets. There are also statically retrogressed rocks.

The charnockitic rocks occur together with red to grey, fine-grained, granitic to aplitic gneisses and grey to reddish-grey, quartz-monzonitic gneisses, both of the types described above. The charnockitic rocks cross-cut metabasites, some of which are brecciated, angular fragments, decimetres to half a metre across, set in a charnockitic matrix. Dykes of aplitic granites locally cut the charnockitic rocks.

Undifferentiated orthogneisses

The undifferentiated orthogneisses shown in Fig. 5 are mostly mixtures and/or deformed varieties of the quartz-monzonitic and granitic gneisses described above. Compositional differences are too varied to be separated on a scale of 1:50,000. The contacts between undifferentiated gneisses and quartz-monzonite and granite gneisses are commonly transitional.

The undifferentiated orthogneisses are, in general, strongly and penetratively deformed. Along Einarsdalen, there is a compositional banding of the gneisses with concordant migmatite veins, which is locally folded isoclinally. Mylonites carry lens-shaped porphyroclasts of e.g. perthitic feldspars, hornblende and scapolite. The penetrative foliation is defined by amphibolite facies minerals such as quartz + feldspars + biotite ± hornblende.

Layered basic complexes within the Roan Igneous Complex

The largest layered basic rock complexes are located along Brandsfjorden, at Kråkfjorden and at Løsliheia. Smaller massifs occur, e.g. at Ramnen, Ørnfuruheia and Kiransholmen. The massifs comprise several types of basic and intermediate rocks, layered on a scale of centimetres to tens of metres. The contacts between the layered mafic complexes and the surrounding gneisses are either foliated concordant, being defined by amphibolite facies minerals and subparallel migmatite veins, or consist of a zone of chaotic migmatite structures, with amphibolitized basic fragments partly dissolved in the migmatite gneisses. It is therefore difficult to establish a relative age for these basic rocks.

THE KRÅKFJORD COMPLEX

The best preserved metabasic complex is located at Kråkfjord. The dominant rock-type forms the central part of the massif and consists of clinopyroxene + kyanite + garnet + rutile + apatite + Cu,Fe-sulphides. The fresh rock is conspicuous, with medium-grained light green clinopyroxene, blue kyanite and red garnet; orthopyroxene, plagioclase, amphibole, spinel, sapphirine and corundum form symplectites. Detailed petrography and chemical analyses of this rock are given in Johansson & Möller (1986). The rock is rich in Mg and Al (Table 2), but more Fe-rich varieties without kyanite are found as well. Ultramafic rocks consisting of coexisting garnet + olivine ± clinopyroxene + apatite + Fe,Ti-oxides form up to 4 m thick discrete layers and lenses. They contain symplectites of clinopyroxene, orthopyroxene, amphibole, spinel and magnetite. Subordinate basic rocks consisting of orthopyroxene + garnet + clinopyroxene are present as layers a few decimetres thick.

Other subordinate rock-types, whose origins are difficult to assess, form cm- to dm-thick discrete layers. These are: massive layers of pure garnet, usually within the orthopyroxene-bearing basic rock; fine- to medium-grained quartz-rich rocks with plagioclase + garnet + rutile + kyanite; garnet-rich (more than 50 vol-%) fine-grained layers containing anorthite + rutile + Fe,Ti-oxides ± amphibole; and light-coloured rocks rich in zoisite, with garnet + plagioclase + amphibole + spinel.

Generally, foliations and boudinage are associated with retrogression and defined by amphibolite facies minerals. A preferred orientation of clinopyroxene and kyanite occurs, however, in some well preserved rocks. Foliations are subparallel with the layering, except in late shear zones with amphibolite and greenschist facies minerals.

The margin of the Kråkfjord massif consists of more Fe-rich basic rocks, irregularly mixed and deformed together with migmatitic gneisses. Lenticular pods contain well preserved fine- to medium-grained, dull greenish-grey basic rocks. They consist of garnet + clinopyroxene + plagioclase + hornblende + quartz + rutile + opaques ± orthopyroxene. Deformation of the basic rocks and migmatitic gneisses is associated with amphibolitization.

Late pegmatites and quartz veins cut the compositional layering and foliations in the massif. In the outer parts of the massif, they also cut folded basic rocks and migmatite gneisses.

THE BRANDSFJORDEN COMPLEX

The compositional layering consists of basic rocks of slightly varying composition, intermediate rocks, and cm- to dm-thick, garnet-rich felsic rocks. The rocks of the Brandsfjorden complex are generally more foliated and amphibolitized than those of the Kråkfjord complex. The foliation is subparallel with the compositional layering and generally defined by amphibolite facies minerals; however, in well preserved high-grade rocks a foliation defined by clinopyroxene may be present. NW-SE kyanite-lineations are seen locally in a garnet-pyroxenite in the innermost part of Brandsfjorden. The original layering is severely transposed by deformation, and commonly disrupted and boudined. In places, deformed reddish-grey granitic migmatitic gneisses occur interbanded with the rocks of the complex. Pegmatites cut the banding and foliation.

The dominant rock-types are fine- to medium-grained, dark greenish-grey metabasites consisting of hornblende + garnet + clinopyroxene + plagioclase + quartz + opaques ± rutile ± sphene ± orthopyroxene ± biotite. Dull grey, fine- to medium-grained, intermediate to felsic rocks and light red, fine-grained, garnetiferous felsic rocks are fairly common and contain the same assemblages as the basic rocks, but with more quartz, biotite and perthitic feldspars. Scapolite is abundant in some basic to intermediate rocks, while zoisite is found in some acidic and intermediate rocks.

Light green garnet-clinopyroxene rocks, commonly with kyanite, sapphirine and corundum among other minerals, form layers or lenses in at least three localities (K, Fig. 5). In a few places, Mg-hornblendes occur instead of clinopyroxenes, together with garnet and kyanite. Garnets can be as large as 4 cm in diameter, and contain several inclusions such as hornblende, zoisite, kyanite, orthopyroxene, plagioclase, rutile, quartz, biotite, FeTi-oxides and Fe-sulphides. However, many of the garnets are cracked and in places it is difficult to distinguish between primary and secondary inclusions. Some garnets are zoned, with pyrope contents increasing from 26 mole % in the core to 53 mole % in the rim. Zoisite, kyanite, rutile and quartz form primary inclusions in the cores of these zoned garnets. Orthopyroxene, plagioclase, hornblende and spinel form symplectites along the rims and cracks in garnets.

THE LØSLIHEIA COMPLEX

The Løsliheia complex is not as well exposed as those at Kråkfjord and Brandsfjord. The rocks are severely amphibolitized, deformed and intercalated with felsic rocks, and foliations are more or less chaotic. Some of the structural complexities can be seen in the steep walls at the head of Berfjord (Fig. 9).

The rocks are dark, fine- to medium-grained, mafic and intermediate in composition, with similar assemblages as the dark basic and intermediate rocks at Brandsfjorden. Brown, medium-grained, felsic rocks with-quartz + perthitic feldspars + garnet + opaques are present locally. Light garnetiferous felsic rocks and kyanite-bearing basites have not been found here. Quartz-monzodiorite gneisses, metadolerites (described below) and light granite gneisses are found within the Løsliheia complex. Relationships between the rocks are unclear due to poor exposure and retrograde deformation which has caused structural concordance.

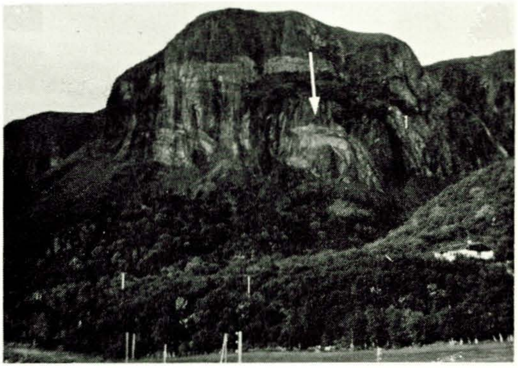

Fig. 9. Fold structures (arrow) in the Løsliheia basic complex, as seen on the steep walls at the head of Berfjorden; looking towards the east.

Banded amphibolites within the Banded Gneiss Complex

The amphibolites in the BGC vary in thickness up to ca. 1 km and can usually be traced for several kilometres. Banded amphibolites are found either in close proximity to supracrustal rocks or 'solitary' within the orthogneisses. They are commonly medium-grained, black, with cm- to m-thick, lighter, more plagioclase-rich layers and bands. The most mafic parts of the amphibolites in places contain lenses of ultramafic rocks, up to tens of metres long. The banded amphibolites grade into plagioclase-rich, intermediate, lighter-coloured, in places quartz-bearing composition. Subordinate amphibolites contain thin calc-silicate bands.

A foliation defined by hornblende and thin quartz-feldspar veins is generally concordant with the compositional layering. Both the layering and the veins are locally isoclinally folded, and the more competent layers are often boudinaged. The amphibolites contain hornblende + plagioclase + sphene + magnetite + Fe- and Fe-Cu-sulphides ± rutile ± garnet ± quartz ± biotite. Garnet and locally clinopyroxene and zoisite may occur as relics.

Metadolerites

Metadolerites are common in all parts of the RIC (D, Fig. 5) and also within orthogneisses in the less deformed parts of the BGC. They generally form lenticular bodies with deformed and amphibolitized contacts to the surrounding gneisses; however, well preserved intrusive relationships are preserved locally. Cross-

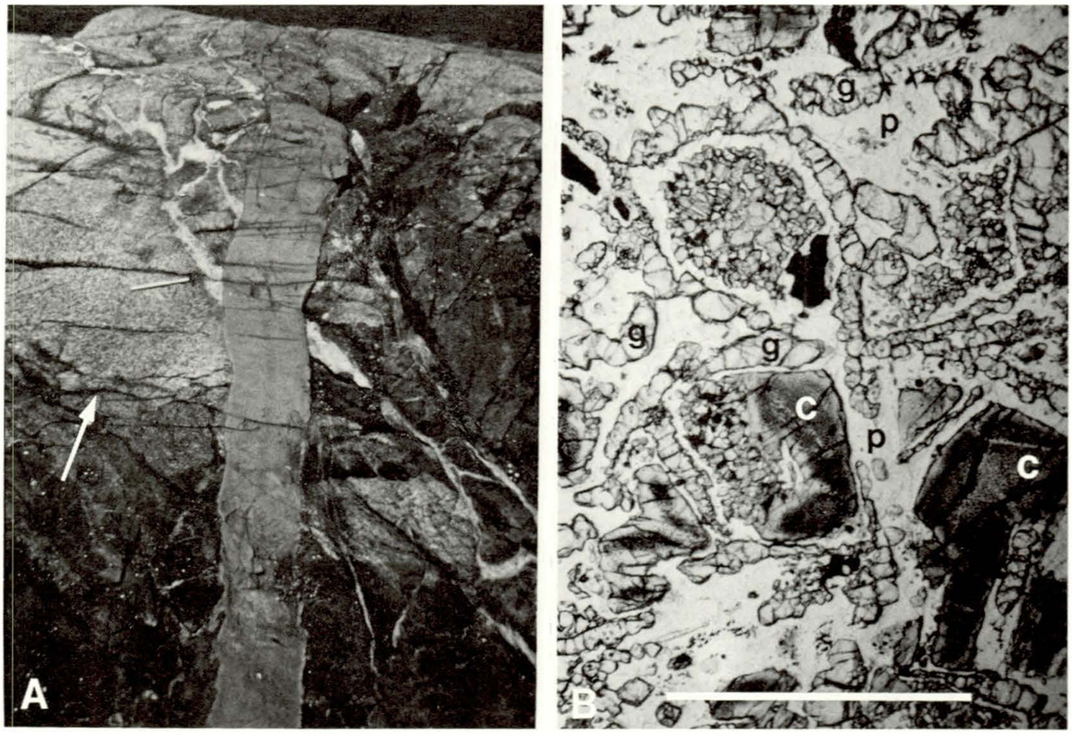

Fig. 10. (A) Dolerite with fine-grained (chilled) margin cutting the contact (arrow) between two basic rocks and a charnockitic vein at Eian, with hammer (55 cm) for scale. All rocks are overprinted by granulite facies assemblages. (B) Photomicrograph of a metadolerite. Pigmented, relict igneous clinopyroxenes (c) are rimmed by clear metamorphic clinopyroxene. In the upper central part of the photo, an igneous pyroxene crystal is completely replaced by a granular aggregate of clear new clinopyroxenes. The plagioclase laths (p) are partly replaced by garnet (g). Spinel forms small needles in the plagioclase. Scale bar = 1 mm.

cutting relationships with fine-grained chilled margins of a dolerite dyke can be seen in Fig. 10a. The dolerites are found within all of the acidic and intermediate rocks and among some basic rocks (e.g. at Løsliheia) and are thus probably the youngest intrusive rocks in the RIC, with the exception of late pegmatites and the layered basic complexes at Kråkfjord and Brandsfjord. Within the latter complexes metadolerites have not been observed.

In the metadolerites, subophitic textures are preserved with relics of magmatic pyroxenes, and, rarely, olivines. The clinopyroxenes are pigmented by numerous Fe- and Fe-Ti-oxides, a few microns across. Plagioclase is generally recrystallized and granoblastic, but the shape of the primary laths is commonly preserved. In a few samples, the primary plagioclase-twins are preserved.

The igneous minerals and textures are partially to completely replaced by granulite facies metamorphic minerals. The magmatic clinopyroxenes have clear, green rims, or are more or less replaced by granoblastic aggregates of clear, green, granoblastic clinopyroxene (Fig. 10b). Garnet forms rims around pyroxenes and plagioclase, and in places plagioclase laths are totally pseudomorphed by garnet. Dark green spinel forms minute needles in the plagioclase. Fe,Ti-oxide, Fe-sulphides, apatite and biotite are present in variable amounts. Orthopyroxene occurs together with clear clinopyroxene, garnet, plagioclase, hornblende, biotite, opaques and, in a few places, perthitic feldspar in some of the non-coronitic (recrystallized?) metadolerites. When retrogressed, biotite and hornblende occur in increasing amounts and the ophitic texture vanishes.

Other basic rocks

Many small (less than 100 m across) amphibolite bodies occur as tectonic lenses in the

SAMPLE	G1	G7	G5	G9	G2	G6	G8	G12	G4	G3	KR3	KR9	KO
wt-%													
SiO_2	74.56	73.18	68.88	66.56	60.84	60.70	60.69	59.46	57.29	56.15	47.65	49.71	46.47
TiO_2	0.28	0.29	0.34	0.36	0.68	0.61	0.71	0.78	0.61	0.75	0.11	1.77	2.36
Al_2O_3	13.00	13.67	15.55	17.39	18.50	17.49	18.12	18.41	16.50	18.85	16.54	14.10	15.71
Fe_2O_3	1.22	1.39	2.04	1.68	4.24	4.81	6.03	5.13	6.62	6.26	0.76	0.96	2.44
FeO	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	3.64	10.95	12.06
MnO	0.06	0.04	0.07	0.03	0.09	0.13	0.11	0.09	0.14	0.13	0.09	0.23	0.22
MgO	0.22	0.28	0.46	0.55	1.37	2.28	1.33	1.42	4.52	3.05	12.04	7.74	5.91
CaO	0.62	0.89	1.55	2.05	4.59	4.46	3.42	3.88	6.40	6.88	16.63	10.16	8.54
Na_2O	3.72	3.38	4.08	3.92	4.44	4.26	3.75	4.51	3.77	4.15	1.15	2.70	3.13
K_2O	5.47	6.24	6.34	6.88	4.37	4.51	5.20	5.41	2.99	2.83	0.05	0.16	1.12
P_2O_5	0.24	0.03	0.09	0.07	0.25	0.29	0.23	0.23	0.34	0.41	0.02	0.12	0.47
LOI	0.32	0.24	0.27	0.19	0.35	0.24	0.11	0.48	0.43	0.24	0.82	1.50	1.52
Sum	99.71	99.63	99.67	99.68	99.72	99.78	99.70	99.80	99.61	99.70	99.50	100.10	99.95
Fe_2O_3tot	1.22	1.39	2.04	1.68	4.24	4.81	6.03	5.13	6.62	6.26	4.80	13.13	15.92
ppm													
Ba	93	309	945	2079	1363	1988	1885	1949	1123	1275	n.a.	n.a.	n.a.
Cr	8	8	6	11	24	30	12	17	216	78	2175	194	n.a.
Nb	21	11	11	6	25	9	10	11	7	8	5	8	14
Rb	219	164	153	159	133	96	128	108	56	80	4	8	27
Sr	44	101	237	647	657	705	455	545	771	981	86	71	535
Y	48	46	25	17	37	26	42	27	26	23	5	45	34
Zr	223	269	379	209	215	242	612	533	153	143	4	109	101

TABLE 2. Bulk major and trace element chemistry of rocks within the Roan Window. Analyses G1-G12 are in the table listed in order of decreasing silica contents. G1— light red granite gneiss, Austvika; G2— grey quartz-monzonite gneiss, Mulodden; G3— grey quartz-monzodiorite gneiss, Roan; G4— dark quartz-monzodiorite gneiss, Roan; G5— red quartz-monzonite gneiss, Beskeland; G6— grey quartz-monzonite gneiss, Beskeland; G7— light red granite gneiss, Brandsøya; G8— yellow charnockitic quartz-monzonite, Eian; G9— grey granite gneiss, Skjørafjord; G12— brown charnockitic monzonite, Grønningen. KR3— kyanite-bearing metabasite, Kråkfjord; KR9— metabasite, Kråkfjord; KO— mean value of 14 coronitic metadolerites: n.a.— not analyzed. LOI— loss on ignition.

gneisses, or as boudined or folded sheets. Remnants of pyroxenes and garnet are commonly present in the cores of these bodies.

In the Eian area, there are small bodies of other types of high-grade metabasic rocks, such as metagabbroic and pyroxene-rich rocks (Möller in prep.).

Geochemistry of rocks within the Roan Igneous Complex

Chemical analyses of some of the different rock-types are presented in Table 2. The samples were carefully selected from the most homogeneous parts of the characteristic rock units. Migmatitic varieties of the rocks were thus avoided. Sample sizes were 5-15 kg. Samples G1 to G12 were analyzed at Midland Earth Science Associates, U.K., using X-ray fluorescence spectrometry. The metabasite samples were analyzed at the Geochemical Laboratory of the Institute of Geology in Lund using X-ray fluorescence and atomic absorption spectrometry (Solyom et al. 1984).

The acidic to intermediate intrusive rocks (samples G1—G9 and G12) are metaluminous to near peraluminous in composition based on the molar Al_2O_3 versus Na_2O+K_2O+CaO plot. Using the log $(CaO/(Na_2O+K_2O))$ versus SiO_2 plot, they are alkali-calcic, according to Brown (1981). In the trace element discrimination diagrams of Pearce et al. (1984), the analyses cluster in the field for volcanic arc granites (Rb versus Y+Nb) and in the field for volcanic arc granites + syn-collisional granites (Nb versus Y), in both diagrams close to the field for within-plate granites.

The two analyses from the Kråkfjord basic complex are representative for the most abundant rock-type within the central layered body, the kyanite-bearing garnet-pyroxenite (KR3), and the more Fe-rich basic rocks surrounding it (KR9). The kyanite-bearing garnet-pyroxenite is high in Mg, Al, Ca and Cr, and likely originated as anorthite-rich layers within a layered intrusion. The Fe-rich rocks can be classified chemically as MORB-type tholeiites (Solyom et al. in prep.).

An average of 14 chemical analyses of

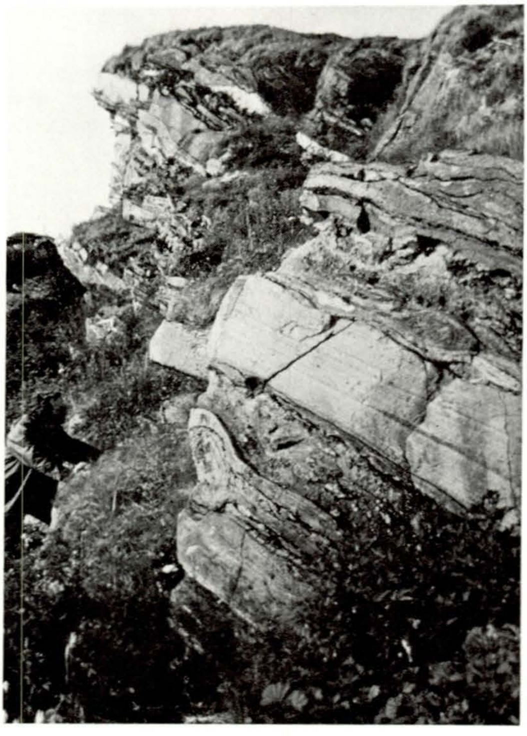

Fig. 11. Compositional banding of paragneisses, marbles, calc-silicate and mafic rocks in the Einarsdalen Supracrustal Unit, Sunnskjørholmen.

metadolerites is presented as KO in Table 2. The metadolerites can be classified as continental tholeiites on the basis of their trace element chemistry; however they have high K-contents (Johansson & Solyom in prep.).

The Einarsdalen Supracrustal Unit

Metasedimentary rocks with concordant amphibolites directly overlie the RIC. They are best exposed along Einarsdalen (e.g. at Minusodden, Fig. 4), and are referred to as the Einarsdalen Supracrustal Unit (ESU). They form two thin sheets separated by a sheet of orthogneisses within the lowermost level of the BGC, and they can be traced continuously around all exposed parts above the RIC (Figs. 5 & 6). Similar supracrustal rocks on the islands west of Roan (Farmannsøya, Klokkarholmen, Allmenningen and Vaerøya) are also included in the descriptions below. Rocks similar to the ones within the ESU are found also in other supracrustal sheets within the BGC.

The thickness of the ESU varies from less than a metre to about 200 metres. Despite the small thickness, it is possible to trace these rocks continuously for several kilometres. The unit is composed of different rocks, individual layers varying in thickness from a decimetre (Fig. 11) to (only locally) up to 40 metres. Because of the small-scale layering and rapid variation along strike, it was not possible to establish a stratigraphy. Locally, metre-thick slices of migmatitic gneisses are present within the supracrustal rocks and vice versa.

The most common and distinctive rock-types are garnet-kyanite-bearing paragneisses, marbles, calc-silicates and mafic rocks. A Mg,Al-rich gedrite rock was found at one place. The minerals in the different rock-types are listed in Table 3. Most rocks have a penetrative fabric which is subparallel with the composi-

	QZP	BIP	MAF	CAR	GED
QUARTZ	XXX	XXX	XX	X	
PLAGIOCLASE	XX	XX	XXX	XX	XX
K-FELDSPAR	XX	XX			
SCAPOLITE				XXX	
ZOISITE			X	XX	
BIOTITE	XX	XXX	XX		
PHLOGOPITE				XX	XXX
WHITE MICA	X	X		X	
HORNBLENDE			XXX		
GEDRITE					XXX
CLINOPYROXENE			XXX	XXX	
GARNET	XXX	XXX	XXX	XX	XXX
KYANITE	XXX	XX			XX
SILLIMANITE	XX	XX			XXX
ANDALUSITE	X				
STAUROLITE	X				XX
CORDIERITE					XX
SPINEL				X	XX
SAPPHIRINE					X
CARBONATES				XXX	
SPHENE			XX	XXX	
RUTILE	XX	X			
OPAQUES	XX	XX	XX	XX	X
ZIRCON	X	X			X
MONAZITE	X				
APATITE	X		X	X	X
GRAPHITE	X		X		

TABLE 3. Minerals in the rocks within the Einarsdalen Supracrustal Unit. QZP— quartz-rich paragneisses, BIP— biotite-rich paragneisses, MAF— mafic rocks, CAR— marbles and calc-silicate rocks, GED— gedrite-rich rock. XXX— abundant, XX— common, X— observed. Relationships between the different parageneses are described in the text.

tional banding and the regional foliation. Mafic rocks occur as both as small lenses and boudins and as several metres thick layers. Locally, isoclinal folds are found concordant with the banding and foliation. In the paragneisses, the foliation is generally defined by biotite ± sillimanite. The marbles commonly show polyphase non-cylindrical folding. With the exception of pegmatites, no cross-cutting igneous rocks have been found within the supracrustal rocks.

Paragneisses
Biotite-rich paragneisses are the most abundant rocks in the ESU. They are fine- to medium-grained and dominated by reddish-brown biotite, pink to red garnet, quartz, feldspar and Al-silicates. Quartz-feldspar veins are common in these biotite-rich varieties.

There are gradations between biotite-rich and quartz-dominated compositions. The most quartz-rich paragneisses generally form individual layers. They are light-coloured with abundant blue kyanite and pink garnet. Garnets up to 2 cm in diameter occur locally. The rocks are rich in heavy minerals such as rutile, Fe,Ti-oxides, Fe-sulphide, apatite, zircon and monazite.

Coexisting minerals in the paragneisses are garnet + kyanite + quartz ± K-feldspar ± plagioclase ± biotite + rutile ± graphite ± Fe,Ti-oxides. In some quartz-rich paragneisses, elongate garnet and K-feldspar define a foliation together with kyanite and quartz. Biotite and sillimanite, which elsewhere commonly defines the foliation, were in some rocks formed at the expense of garnet and K-feldspar. Analyzed garnets have the compositions $alm_{55-65}pyr_{25-40}gro_{5-15}$ $spess_{0-3}$. Compositionally zoned garnets have inclusion trails of minute (<10 µm thick) needles of Al-silicate, probably sillimanite, together with quartz, rutile and Fe,Ti-oxides. An attempt to identify the needles by means of XRD was not successful due to their small volume compared to the garnet. Andalusite locally replace kyanite in the quartz-rich paragneisses.

Mafic rocks
Diffent types of mafic rocks occur as layers and lenses of varying thickness up to several tens of metres. The most common type is dark grey, fine- to medium-grained and is composed mainly of amphibole, biotite, plagioclase and red garnet. The relative amounts of biotite and amphibole vary greatly. Another type is richer in Ti, with abundant sphene, yellowish-brown amphibole and reddish-brown biotite. Garnet compositions in these rocks are approximately $alm_{50}pyr_{20}gro_{30}$. Greyish-green, fine- to medium-grained, dm-long basic lenses consist of garnet + clinopyroxene + plagioclase + quartz ± zoisite + sphene + Fe-sulphides + Fe,Ti-oxides. In these latter rocks, garnet compositions are c. $alm_{45}pyr_{20-25}gro_{30-35}spess_{0-2}$, and clinopyroxene is c. $diop_{70}$-$hed_{25}tsch_{5}$.

Banded amphibolites, similar to the large 'solitary' banded amphibolites within the BGC, are in places present next to the supracrustal rocks of the ESU. Locally they contain up to one metre long lenses of light green diopside.

Marbles and calc-silicate rocks
Marbles vary widely in colour and mineralogical composition. Thicknesses are between a few centimetres and 40 metres. The thicker occurences (e.g. at Nordskjør, Minusodden and Allmenningen) consist of a coarse- to medium-grained white marble. On the island Allmenningen, it contains dolomite with diopside, scapolite and cummingtonite (Ramberg 1943). Parts of the marble contain lenses and bands of coarse diopside, up to a metre thick. The white marble on Allmenningen was quarried in the beginning of the century and used as paving stone in the Nidaros Cathedral in Trondheim. On Allmenningen and east of the Harbak peninsula (Fig. 2), a red marble is present, a few dm thick. It consists of calcite with diopside, scapolite, phlogopite and white mica. Yellowish, impure marbles are common in many places among the metasedimentary rocks. Besides carbonates, they contain abundant quartz, plagioclase, scapolite, diopside, amphibole and sphene.

Calc-silicate rocks mostly form dm-thick or thinner layers, or lenses of coarse minerals. They are rich in diopside, garnet, scapolite, zoisite, sphene, plagioclase and calcite. Coexisting minerals are garnet + diopside + zoisite + sphene + scapolite + calcite. There are spectacular symplectite and corona textures, where anorthite is formed at the expense of zoisite and new garnets form coronas around the earlier minerals. Analyzed garnets have the compositions $alm_{20-45}pyr_{3-8}gro_{50-80}andr_{0-5}$.

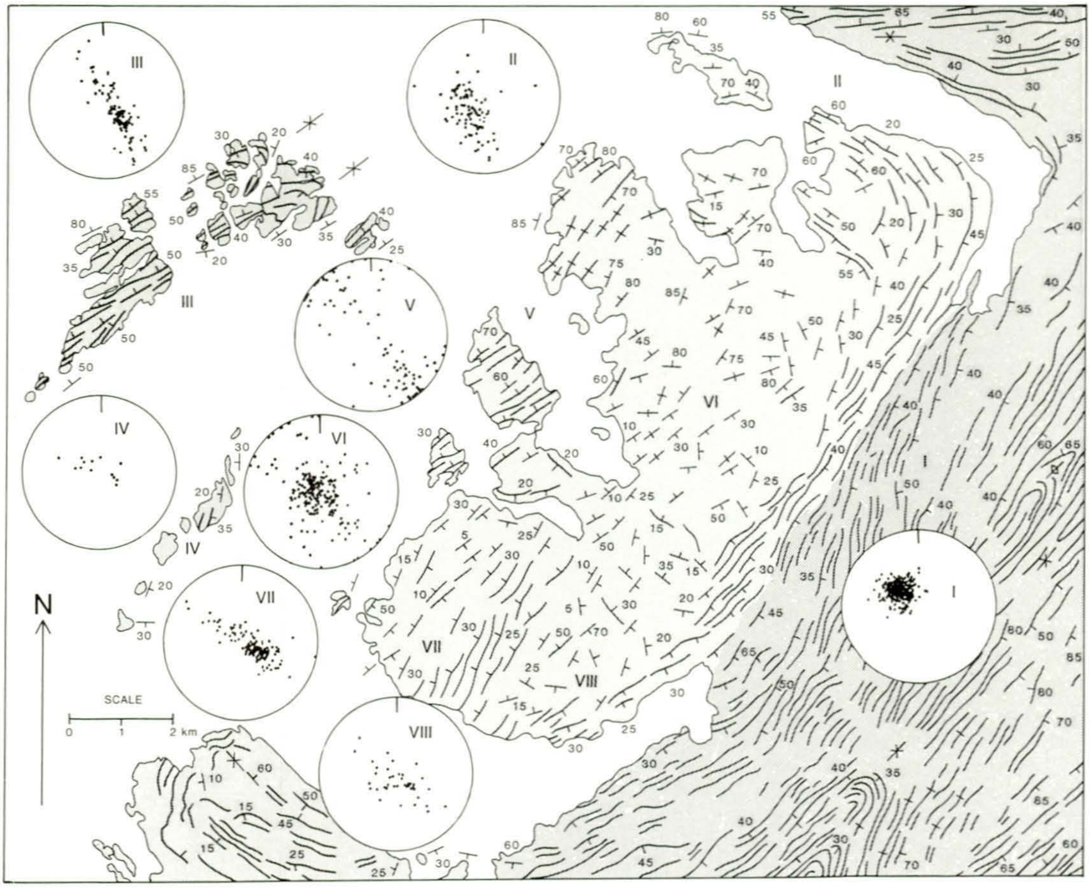

Fig. 12. Strike-line map showing the orientation of penetrative foliations in the Roan area. The Roan Window is shown in light grey. The numbers at the stereographic equal-angle projections (poles to foliations) refer to the domains shown in the inset map of Fig. 13.

Gedrite rocks
Gedrite rocks were found as a lens, only a few metres long, within mylonitized migmatitic gneisses. They are medium- to coarse-grained and rich in dark gedrite, garnet, phlogopite and sillimanite. Coexisting minerals are garnet + gedrite + kyanite ± Mg-rich staurolite ± phlogopite. These minerals break down to different parageneses including cordierite, spinel, sapphirine and sillimanite, in part as symplectites (Möller in prep.).

Pegmatite dykes
Pegmatites are composed of quartz + K-feldspar + plagioclase with minor biotite, muscovite and magnetite. They are 5 cm to 2 m wide and generally persistent for 2 to 10 m. In rare instances, pegmatite dykes contain graphic intergrowths. Pegmatites are found in the four associations listed below. Associations 1-2 and 3-4 correspond to at least two different generations of pegmatites.

1. In the Eian area, pegmatites are generally undeformed and are in places rooted in the charnockitic or granitic rocks. **2.** In the Utro and Berfjorden areas, pegmatites are boudined concordant with the D2 foliation (see Structural geology section below). **3.** Pegmatites cross-cut and retrograde high-grade parageneses in metabasites e.g. at Gejsnesodden, Mulodden and Kråkfjord. **4.** Abundant pegmatites cross-cut early structures (D2 and D3, see below) in the gneisses. They are generally vertical and commonly form the loci for

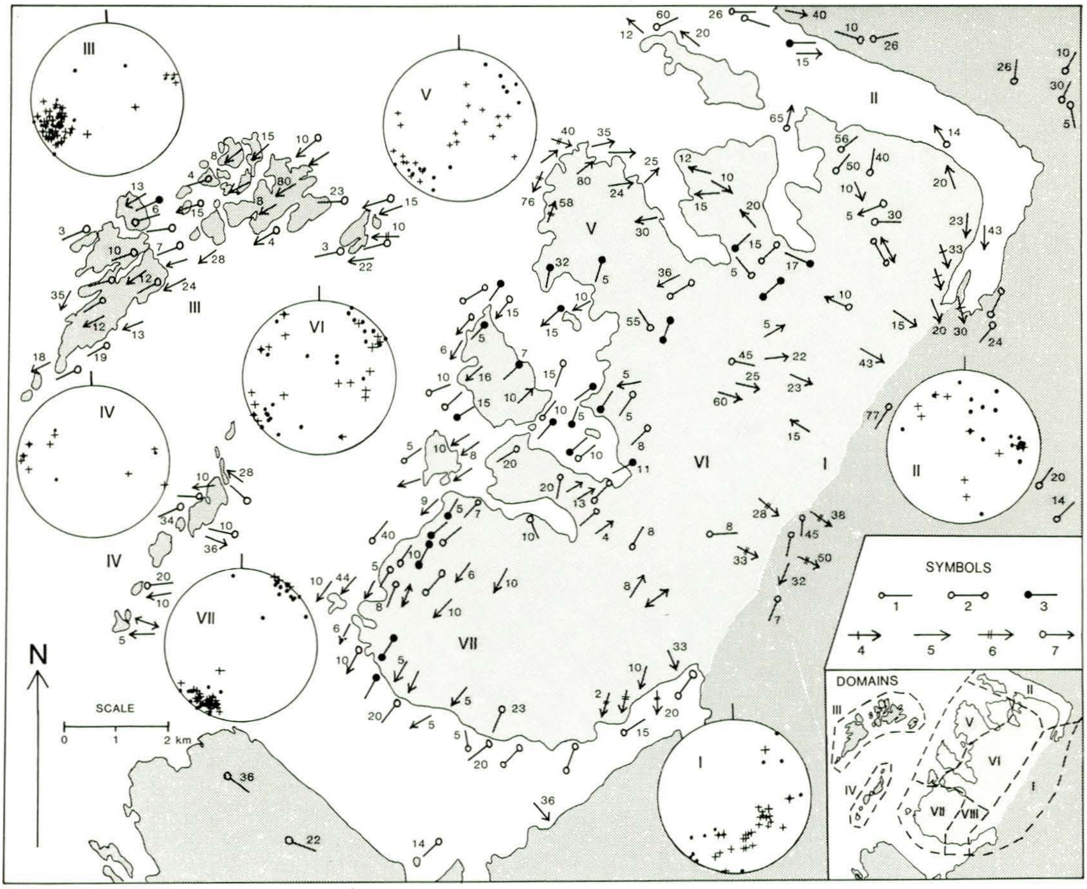

Fig. 13. Linear structures in the Roan area. Symbols: 1: fold axis; 2: horizontal fold axis; 3: fold axis of folds with axial-planar migmatite veinlets; 4: kyanite lineation; 5: amphibolite facies lineation; 6: lineation in amphibolite facies mylonitic rocks; 7: coaxial fold axis and lineation. The inset map shows the domains for the stereographic equal-angle projections of fold axes (points) and mineral lineations (crosses).

minor, mostly sinistral, shear zones, where the pegmatites and/or the country rock are partly mylonitized. Within the ESU, pegmatites cross-cut the compositional layering and foliation, though in places the dykes are disrupted.

Structural geology

The NE-SW trending culmination in the Roan area exposes the Roan Igneous Complex beneath the Banded Gneiss Complex (Figs. 2, 5, 6 & 7). The RIC and the BGC are separated by a contact zone containing two sheets of the Einarsdalen Supracrustal Unit. The structural geometry above the RIC, with repetitions of orthogneisses and supracrustal rocks, implies a thrust relationship, either as large recumbent folds or as detached thrust sheets. The RIC thus constitutes a window which contains better preserved rocks, tectonically overlain by the orthogneisses, amphibolites and supracrustal rocks of the BGC.

The rocks in the contact zone between the RIC and the BGC were intensely deformed, in part mylonitized, at amphibolite facies conditions. High-pressure granulite facies tectonites occur, locally, in this deformation zone, both within the RIC and the ESU, indicating initiation of thrust motion during the early high-PT event. The amount of displacement along the deformation zone has not been possible to ascertain, but a NW- or SE-directed motion is suggested by the lineations. Textures in mylonites indicate that the latest movements in this zone were directed northwestwards.

Several supracrustal sequences similar to the ESU occur within other parts of the BGC; they may or may not correlate with established Caledonian allochthonous units. The roof thrust of the Roan Window is suggested to be located at the contact between the ESU and the RIC since there are similar supracrustals within higher levels of the BGC. Alternatively, the ESU may represent a cover sequence originally deposited in the RIC, with a thrust separating the ESU from the BGC. In either case, the juxtaposition of a banded gneiss sequence over the better preserved igneous complex requires displacement on the order of tens of kilometres.

Structures within the Roan Igneous Complex

The RIC is an intrusive complex partly overprinted by deformation. Along the margins of the window, the rocks are retrogressed and contain a penetrative foliation concordant with the contact to the BGC. In the central parts of the window, the degree of deformation and the orientations of foliations vary (Fig. 12). The least deformed, in some places unfoliated, rocks are found in the Eian-Roan area where intrusive contacts and high-grade parageneses are also well preserved. The degree of retrograde deformation varies locally, but is generally strongest where there are competency contrasts between different rocks, as along margins of basic rocks.

Within the RIC, there are no well defined structural markers which can be traced for more than a few tens of metres. This makes it difficult to relate the structural elements of the various rocks to different deformation phases. High-pressure granulite facies structural elements were found only locally, and most deformation is associated with amphibolitization subsequent to peak-PT conditions. Four retrograde deformation phases (D2 to D5 below) have been distinguished within the orthogneisses, by using small-scale structures and migmatite veinlets as structural markers. In the field, the S2 to S4 amphibolite facies foliations were indistinguishable where a simple gneissosity occurs without veins. Five recognizable phases of progressive deformation are presented below.

D1: High-pressure granulite facies preferred orientations defined by clinopyroxene and, in places, kyanite were found in some basic rocks. In the innermost part of Brandsfjorden, kyanite defines a NW-SE lineation (Fig. 13). Here, the S1 foliation is subparallel to younger amphibolite facies foliations.

D2: The S2 foliation in the acidic to intermediate rocks is a gneissosity defined by brown biotite and brown-green hornblende together with quartz and feldspars. Clinopyroxene and garnet are found as relics. Isoclinal folds (F2?) and early migmatite veins are concordant to this foliation.

D3: Isoclinal folds (F2), foliations (S2) and the early migmatite veins are folded by small-scale open to tight folds, commonly with steep axial surfaces and subhorizontal fold axes. Neosome veinlets, which are regularly spaced and up to a few decimetres in length, developed parallel to the axial surfaces (Figs. 14a & b). These veins contain microperthitic K-feldspar, mesoperthitic plagioclase and quartz. Mineral lineations are generally coaxial (mainly NE-SW oriented) to the small-scale fold axes; however, lineations cut by such veins also occur. The orientations of these veins and related fold axes and lineations are shown in Figs. 13 & 15. In places, a spaced foliation without veinlets was developed, defined by the same minerals as in D2. Both S2 and S3 foliations have poorly oriented individual grains and may be recrystallized. The S3 spaced foliation locally grades into a penetrative foliation.

Neosome veins, containing hornblende crystals up to 5 cm long, exhibit two different structural relationships: **1.** Long veins generally cut the S3 foliations, the F3 folds, the spaced S3 veins, and show a parallel orientation on local scale (Fig. 14c). **2.** Hornblende-bearing veinlets which are axial-planar to small-scale folds are also found.

D4: Near the contact to the overlying ESU, the D2 and D3 structures in the RIC are overprinted by later deformation. A strong to mylonitic foliation is developed, with an orientation that is subparallel to the margin of the Roan Window. Migmatite veins are elongated in this foliation, which may locally be isoclinally folded (F4: Fig. 14d). This S4 foliation is most strongly developed in a km-wide zone along Einarsdalen; but similar mylonitic foliations are also

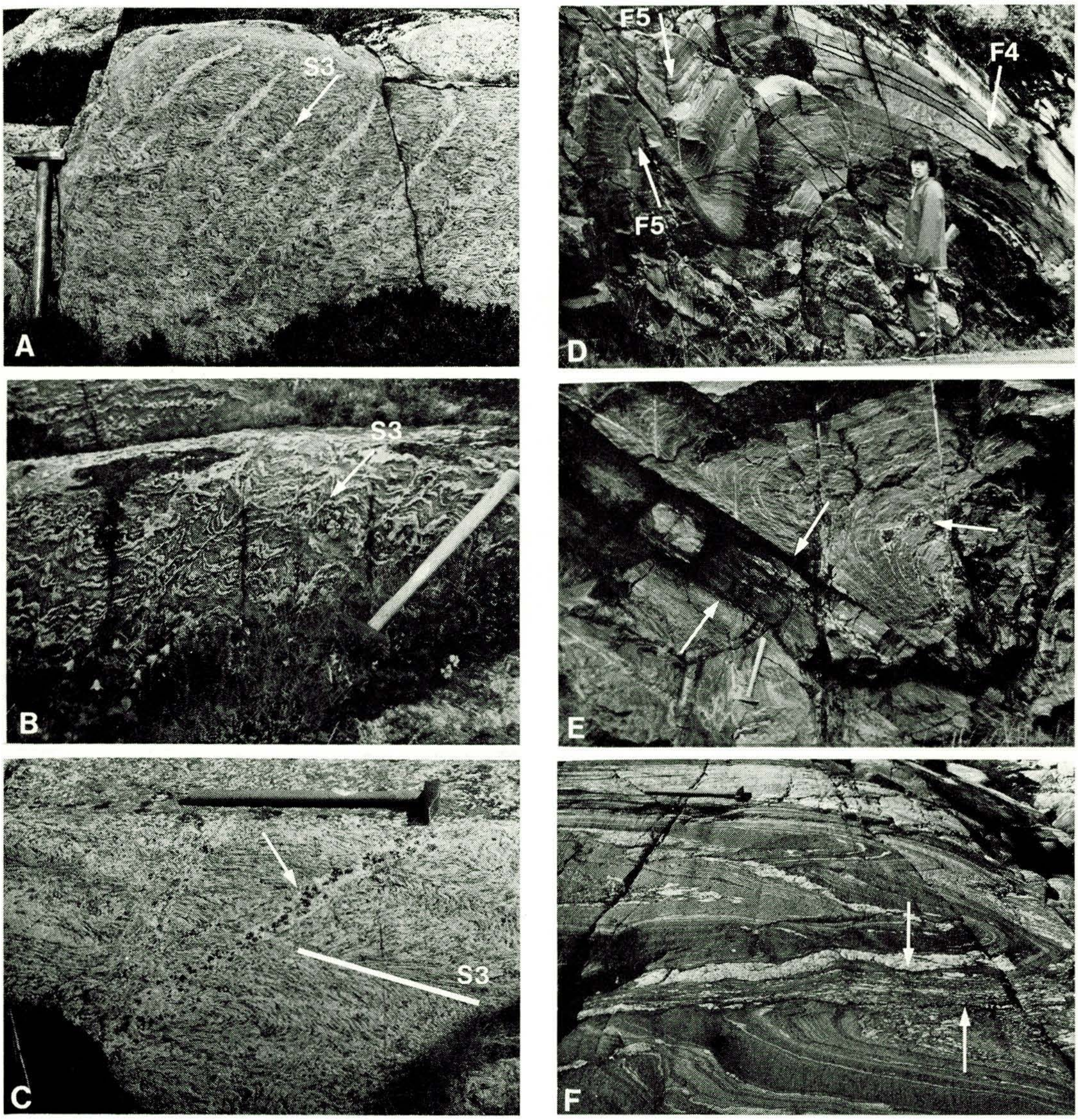

Fig. 14. Photographs of minor structures in the Roan area to illustrate the deformation phases, with hammer (55 cm) for scale. (A) Migmatite veinlets subparallel to the axial surfaces of small-scale F3 folds, which folds the S2 gneissosity. Quartz-monzonite gneiss at Utro. (B) Early, folded migmatite veins, with cross-cutting veins along axial surfaces of F3 folds, gneiss at Berfjorden. (C) Hornblende-bearing migmatite vein (arrow) cutting F3 folds, quartz-monzonite gneiss at Terningen. (D) Mylonitic gneiss with isoclinal folds (F4) refolded by a westward verging noncylindrical fold train (F5), Nordskjør. (E) Shear zone (left arrows) developed along limbs of F5 folds, Nordskjør. A deformed hornblende-bearing vein at the right arrow. (F) Disrupted folds, Værøya. Arrows mark the shear zone boundaries.

encountered elsewhere in the window. In the Einarsdalen zone, mineral lineations in the mylonites are NW-SE oriented. Feldspar, hornblende and biotite occur as asymmetric augens and as smaller grains defining the foliation. The sense of movement inferred from the asymmetry of porphyroclast tails varies; however, in the SE-dipping contact zone in Einarsdalen, most samples indicate a reverse, westward movement. A systematic study of kinematic indicators is in progress.

D5: The S4 foliations and related isoclinal

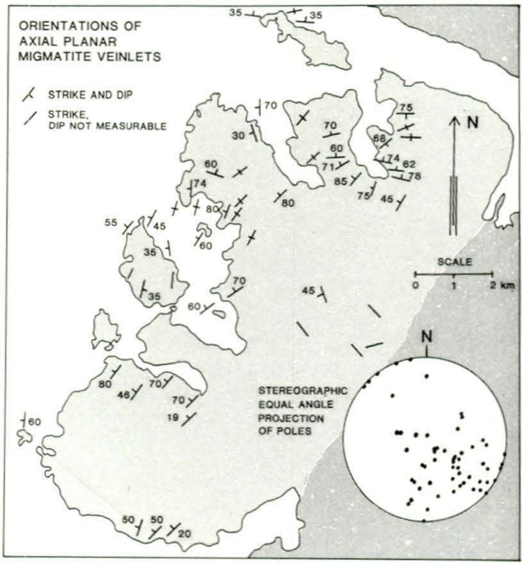

Fig. 15. Orientations of axial planar migmatite veinlets within the Roan Igneous Complex.

folds were refolded by upright, open to tight, noncylindrical folds (Fig. 14d), with dominantly NE-SW trending subhorizontal to moderately plunging axes. These folds were found mainly at the southeastern margin of the window. Locally developed axial-surface foliations are defined by the same minerals as in D4. Intersection and mineral lineations are commonly coaxial with fold axes. Shear zones disrupting F5 folds (Fig. 14e) are probably contemporaneous with the D5 event.

Late structures: Most pegmatites have clear cross-cutting relationships to D2 and D3 structures in the gneisses. These, generally vertical, pegmatites are partly mylonitized by late, mostly sinistral, shear zones.

Vertical joints trending NE-SW, NW-SE and N-S are common. Faulting along the same systems is of only minor importance within the window.

Structures within the Banded Gneiss Complex

Compared to the RIC, the BGC was more intensely deformed and folded (Fig. 2) at amphibolite facies conditions. The banding of orthogneisses, amphibolites and supracrustal rocks was refolded at least twice into mega-scale isoclinal and tight to open folds, most of the later ones upright with NE-SW trending horizontal to moderately plunging axes. Just north and south of the RIC the synform axes are oriented E-W, which suggests that the orientations of some of these folds were adjusted to the shape of the RIC. Limbs of major folds were sheared out, e.g. north of Jotjørnheia (Fig. 2), and on Værøya (Fig. 5) where two map-scale synforms are juxtaposed. Similar sheared-out structures can be seen on a small scale (Fig. 14f).

In general, the acid to intermediate gneisses, banded amphibolites and supracrustal rocks have an amphibolite facies gneissosity with attenuated and locally isoclinally folded migmatite veins (Fig. 3b). The layering in the supracrustal rocks is isoclinally folded, and mafic rocks commonly form lenses and boudins. The foliations and the veins are subparallel to the compositional banding. Intersection and amphibolite facies mineral lineations are generally coaxial with the NE-SW fold axes. However, attitudes of the early fold axes and associated lineations vary, e.g. SSW of Jotjørnheia (Fig. 2). In the hinge zone just northwest of Jotjørnheia, sillimanite lineations plunge WNW, apparently coaxial with the early fold axis. E-W lineations parellel to fold axes are present just north and south of the Roan Window.

In most parts of the BGC, all migmatite veins are concordant with the pervasive amphibolite facies foliation. However, within the refolded isoclinal fold of orthogneisses and amphibolites at Jotjørnheia, the banding of migmatite veins and the foliation in the gneisses are refolded by small-scale NE-SW upright folds. Discordant migmatite veins are regularly spaced along axial surfaces of these small-scale folds, similar to the D3 structures within the Roan Window. Cross-cutting migmatite veins as well as unoriented, patchy veins are present in less deformed areas, e.g. the Granholvatnet Lens (Johansson 1986b), where granulite facies assemblages are also partly preserved in metadolerites.

High-pressure granulite facies foliations and NW-SE oriented lineations defined by elongate garnet and K-feldspar together with kyanite and quartz (Fig. 16a) were found at the contact to the RIC, at Minusodden. These structures may be equivalent to the D1 structures in the RIC. Foliations defined by kyanite, sub-

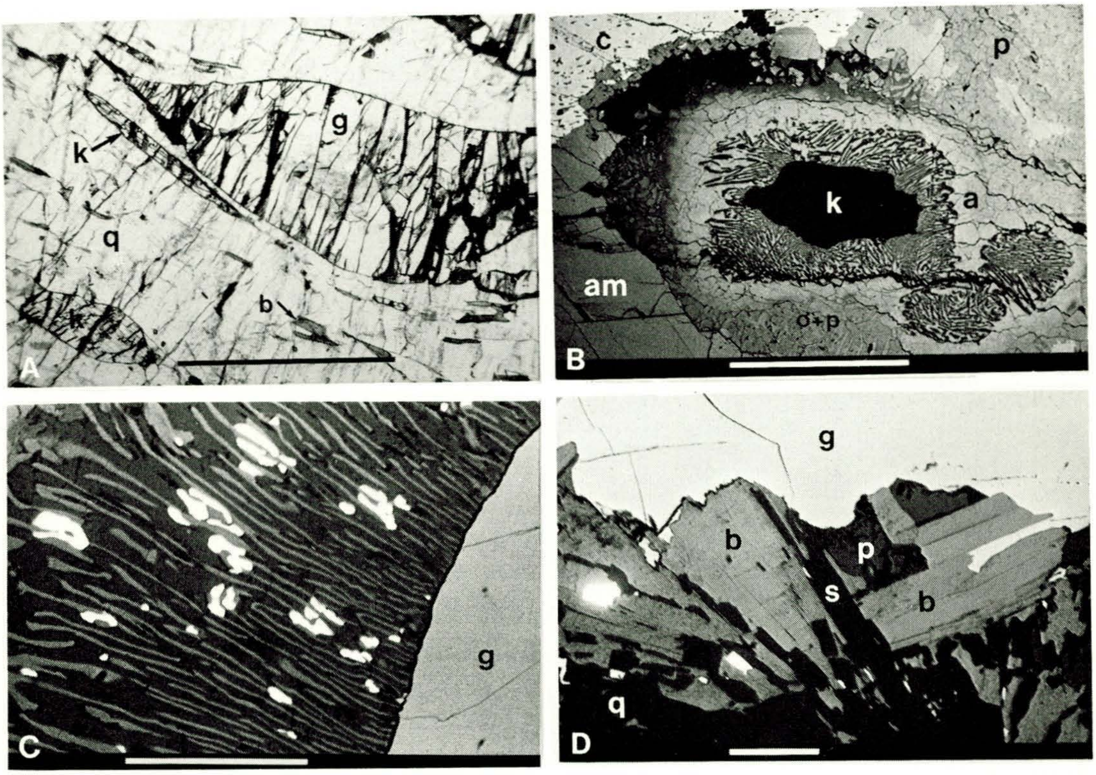

Fig. 16. Photomicrographs of textures in rocks from the Roan area. (A) Lens-shaped garnet (g) and kyanite (k) which define a high-pressure granulite facies foliation together with K-feldspar and minor biotite (b) in a recrystallized quartz-matrix. (q) Quartz-rich paragneiss from Minusodden. Scale bar = 1 mm. (B) Breakdown of kyanite, clinopyroxene and amphibole in a Mg,Al-rich metabasite at Kråkfjord. Kyanite (k, centre) is rimmed by symplectites of corundum, sapphirine and anorthite (a). Clinopyroxene (c, upper) and amphibole (am, lower) are rimmed by symplectites of orthopyroxene (o) and andesine plagioclase (p). The black part below the clinopyroxene is a hole in the thin-section. Scale bar = 1 mm. (C) Garnet (g, right) breakdown to a symplectite of anorthite (dark grey), clinopyroxene (light grey) and magnetite (white blebs). Amphibolite from Vesterfjell. Backscatter electron image. Scale bar = 100 microns. (D) Garnet breakdown in a biotite-rich paragneiss from Einarsdalen. Garnet (g) is rimmed by biotite (b), sillimanite (s), plagioclase (p) and quartz (q). Note the tiny remnant of garnet within biotite (right). Backscatter electron image. Scale bar = 100 microns.

parallel to nearby sillimanite-foliations, were also found within the biotite-rich paragneisses in the narrow synform east of Steinheia (Fig. 2).

Metamorphic evolution

The supracrustal and igneous rocks in the Roan area were all, except for the late pegmatites, depressed to depths of at least 50 km where they equilibrated at temperatures of 800-870°C. The initial stage of the uplift was a near-isothermal decompression into medium-pressure granulite and upper amphibolite facies conditions. Most rocks reequilibrated during deformation at amphibolite facies conditions; however, where the rocks escaped deformation and hydration they retain the early minerals and assemblages. Relatively high temperatures prevailed throughout the uplift.

The metamorphic assemblages and textures in the different rocks are described below, divided into separate stages which may be viewed as a continuous evolution. The Roan Igneous Complex and the Banded Gneiss Complex are treated separately. The relationships between the metamorphic stages and structures are outlined schematically in Table 4. The peak temperature estimate and an outline of the uplift path are shown in Fig. 17. Further work on the metamorphic (PTt) evolution and more detailed metamorphic studies of certain rocks within the RIC and the ESU are in progress.

Stage	The Roan Igneous Complex	The Banded Gneiss Complex
A Early or prograde stage.	Kyanite, zoisite, quartz and rutile in cores of zoned garnets.	Inclusion trails of Al-silicates, quartz, rutile and Fe,Ti-oxides in zoned garnets.
B High-pressure granulite facies stage. *D1* foliations.	Ky + Gt + Cpx ± Amp, Gt + Cpx ± Plag ± Amp, Gt + Opx + Cpx in basic rocks, in places foliations/lineations. Gt + Cpx + Fsp ± Amp ± Opx in acid to Intermediate rocks. Gt + Ol ± Cpx in ultramafic rocks.	Gt + Cpx ± Qz ± Plag ± Amp in basic rocks within orthogneisses. Gt + Cpx ± Zoi ± Plag ± Amp in banded amphibolites. Gt + Cpx ± Plag ± Qz in mafic rocks within paragneisses. Ky + Gt ± Kfsp + Qz ± Plag in paragneisses, in places as foliations/lineations. Gt + Ged + Ky ± Stau ± Phl in gedrite rocks.
C Early low PH_2O decompression stage. No deformation?	Symplectites of: Opx + Plag ± Spi ± Mag ± Amp after garnet in basic and intermediate rocks. Opx + Plag ± Amp after Cpx in basic rocks. Corundum + anorthite + sapphirine after kyanite in basic rocks. Cpx + Opx + Spi ± Amp after Gt + Ol in ultramafic rocks. Early pegmatites here?	Symplectites of: Cpx + Plag + Mag after garnet and zoisite in banded amphibolites. Sill + Spi, Cord + Spi ± sapphirine after Mg-rich staurolite, kyanite and gedrite in gedrite rocks. Formation of biotite + sillimanite after garnet and K-feldspar in paragneisses. Sillimanite after kyanite.
D Retrograde high PH_2O stage. *D2* regional foliation and isoclinal folds.	Hydration of assemblages of stage B-C. Foliations and lineations of Amp + Fsp ± Bio ± Qz in orthogneisses and basic rocks. Deformation of migmatite veins.	Foliations and lineations of Bio ± Sill in paragneisses. Foliations of Amp + Fsp ± Bio ± Qz in orthogneisses and basic rocks. Deformation of migmatite veins.
E Retrograde high PH_2O stage. *D3* folds and foliations.	Assemblages as in stage D. Formation of new migmatite veins as an axial planar spaced cleavage in intermediate gneisses. NE-SW lineations here? Later Hbl-bearing migmatite veins.	Assemblages as in stage D. Formation of new migmatite veins as an axial planar spaced cleavage in intermediate gneisses. NE-SW lineations here?
F Retrograde high PH_2O stage. *D4* foliations and isoclinal folds. *D5* upright noncylindrical refolding.	Deformation of D3 migmatite veins. Foliations of Amp + Fsp ± Bio ± Qz in orthogneisses and basic rocks. NW-SE lineations in mylonitic gneisses in Einarsdalen. Shear zones in fold limbs.	Deformation of D3 migmatite veins. NE-SW lineations here? Foliations of Amp + Fsp ± Bio ± Qz in orthogneisses and basic rocks. Shear zones in fold limbs.
G Late features.	Pegmatites here, at or before stage F? Minor shear zones.	Pegmatites here, at or before stage F? Andalusite after kyanite in paragneisses.

TABLE 4. Schematic presentation of the metamorphic and structural stages within the Roan Igneous Complex and the Banded Gneiss Complex.

The Roan Igneous Complex

Magmatic textures and minerals are partly preserved, such as clinopyroxene and plagioclase in coronitic dolerites, and K-feldspars and, probably, clinopyroxene in charnockitic rocks.

Early or prograde stage. A possible prograde PT-path can be traced in zoned garnets within kyanite-bearing basic rocks at Brandsfjorden. The cores have flat zoning profiles with an almandine-rich composition ($pyr_{26}alm_{47}$ $gro{+}and_{25}sp_2$), while the rims are strongly zoned with increasing pyrope contents and decreasing almandine and grossular contents (to $pyr_{53}alm_{28}$ $gro+and_{18}sp_{1}$). Primary inclusions in the cores of these garnets include zoisite, kyanite and quartz, which is a high-pressure assemblage.

High-pressure granulite facies stage. High-pressure granulite facies parageneses are found in all rock-types within the RIC, with the exception of pegmatites. In acidic to intermediate rocks the high-grade assemblages include garnet + clinopyroxene + quartz + perthitic feldspars ± hornblende ± orthopyroxene. Basic rocks contain garnet + clinopyroxene ± orthopyroxene ± plagioclase ± hornblende ± kyanite. Garnet + olivine ± clinopyroxene coexist in ultrabasic rocks.

High-PT parageneses are found preferentially in undeformed, massive rocks. However, in the Brandsfjord and Kråkfjord basic complexes, high-grade minerals may show preferred orientations. Where the high-pressure granulite facies assemblage is well preserved, there is an equilibrium texture and the minerals are generally unzoned. Exceptions are garnet-corona textures in the charnockitic rocks, garnet- and clinopyroxene-producing corona textures in the dolerites and the compositional zoning of garnets in the Brandsfjorden Complex.

Geothermobarometry (using the calibrations of Ellis & Green 1979 and Harley & Green 1982) applied to the assemblages clinopyroxene + kyanite + garnet and orthopyroxene + garnet + clinopyroxene in basic rocks within the Kråkfjord complex gives a PT estimate of 870±50°C and 14.5±2 kbar (Johansson & Möller 1986; Fig. 17).

Scapolite is common together with high-grade minerals in some intermediate to basic rocks and indicates the presence of CO_2-rich fluids. From textural relationships it is not everywhere evident at which stages scapolite was stable. However, it decomposes to plagioclase and form relics at the latest stages of amphibolitization.

Early low-P_{H2O} decompression stage. Decompression caused the development of symplectites of medium-pressure granulite facies parageneses at the expense of the high-pressure phases. The most common reaction textures in both basic and intermediate rocks are very fine-grained symplectites of orthopyroxene and/or amphibole + anorthite ± spinel ± magnetite replacing garnet. The same symplectite assemblages formed along cracks in garnets. Networks of diopside and plagioclase, indicating a former clinopyroxene richer in Na and Al, are common in basic rocks.

The Mg-rich rocks of the Kråkfjord complex preserve several reaction textures related to the early retrogression (Johansson & Möller 1986). Examples are the partial replacement of clinopyroxene and kyanite by orthopyroxene + plagioclase ± amphibole, and anorthite + sapphirine + corundum (Fig. 16b), respectively. Symplectite rims of orthopyroxene + plagioclase on amphibole were found in the Brandsfjord Mg-rich basic rocks. In ultrabasic rocks, the garnet and olivine reacted to form symplectites of spinel + orthopyroxene + clinopyroxene ± amphibole.

The symplectite assemblages represent high temperatures and moderate pressures, and mainly low-P_{H2O} conditions. Dehydration or hydration reactions took place depending on the local partial pressure of water.

Retrograde high-P_{H2O} stage. Subsequent retrogression involved hydration and further destabilization of anhydrous phases including the above-described symplectite assemblages. Biotite + feldspars + quartz ± brown-green hornblende formed in the acidic to intermediate gneisses, and basic and ultrabasic rocks were partially to completely converted to amphibolites.

During amphibolitization, migmatite veins developed in acidic, intermediate and even basic rocks. These veins are commonly regularly spaced along axial surfaces of small-scale folds (D3 above); they contain quartz + mesoperthitic plagioclase + microperthitic K-feldspar ± hornblende. During the later stages of amphibolitization (D4-D5 above) perthitic feldspars and scapolite were replaced by non-perthitic feldspars.

Very late retrogression. Late, open cracks associated with shear zones in basic rocks contain chlorite, green biotite, quartz and calcite.

Chlorite replacing garnet, biotite and other mafic minerals can be found in most rock-types. This replacement is apparently unrelated to deformation, and was developed only locally.

The Banded Gneiss Complex
Within the orthogneisses of the BGC, magmatic textures and minerals are partly preserved within coronitic dolerites.

Early or prograde stage. Zoned garnets in quartz-rich paragneisses of the ESU contain inclusion trails of Al-silicate needles (*probably* sillimanite), quartz, rutile and Fe-Ti-oxides. If the Al-silicate is sillimanite, this is an early high-T/P stage which contrasts with the early stage recorded in basic rocks within the RIC.

High-pressure granulite facies stage. High-grade basic rocks containing coexisting garnet and clinopyroxene are present locally throughout the BGC. Metadolerites with coronas of garnets and pyroxenes as well as small basic lenses with relics of garnet and clinopyroxenes are found among the orthogneisses.

Relics of coexisting garnet, clinopyroxene, rutile and zoisite were found in the large banded amphibolites at Vesterfjell (Figs. 4, 5 & 6). Geothermometry (using the calibration of Ellis & Green 1979 and a pressure of 13 kbar) on garnet and clinopyroxene with mutual, 'clean', straight grain-boundaries gave temperatures around 850°C. Within the ESU, small mafic lenses within the paragneisses preserve clinopyroxene + garnet ± quartz ± plagioclase. Geothermometry on these rocks gave temperatures of 800-850°C. Also within other paragneiss sequences of the BGC, as e.g. in the narrow synform west of Steinheia (Fig. 2), mafic rocks contain the same assemblage.

The quartz-rich paragneisses of the ESU contain the assemblage garnet + kyanite + K-feldspar + quartz ± plagioclase ± red biotite + rutile ± graphite. The garnet and K-feldspar may be elongated and define a foliation and/or lineation together with kyanite (Fig. 16a; D1 above). The presence of kyanite in peak PT textural relationships implies pressures of at least 11 kbar at temperatures of 800°C.

The high-pressure assemblage in the gedrite rocks is garnet + gedrite + kyanite ± phlogopite ± Mg-rich staurolite.

Decompression and retrograde stage. In the banded amphibolites at Vesterfjell, garnet and zoisite reacted to form symplectites of anorthite + magnetite + clinopyroxene (Fig. 16c). In the gedrite rocks, the high-pressure assemblage was partly replaced by symplectites. The reactions involve e.g. replacement of garnet by gedrite + spinel, and replacement of Mg-rich staurolite, kyanite and gedrite by sillimanite + spinel and cordierite + spinel ± sapphirine. The latter assemblages imply high temperature and low to medium pressure, possibly as low as 6 kbar at 700–800°C.

Retrograde amphibolite facies parageneses are by far the most common in the rocks of the BGC. Hornblende and biotite together with quartz and feldspars define the gneissosity in both the orthogneisses and banded amphibolites. Sillimanite is a retrograde phase in the paragneisses of the ESU and was in places formed together with biotite at the expense of garnet and K-feldspar (Fig. 16d). Commonly, sillimanite + biotite define a foliation. Sillimanite growth directly on kyanite was found locally in biotite-poor rocks.

Andalusite replaces kyanite in some of the paragneisses. The presence of andalusite implies that the low-temperature retrogression (500-600°C) took place at pressures lower than 5 kbar.

Very late retrograde stage. Chlorite and white mica were found as very fine-grained aggregates mostly located along grain boundaries of earlier minerals, or as chlorite-lamellae in biotites.

Short synthesis of structures and metamorphism
The orthogneisses of the BGC are derived from protoliths similar to the rocks within the RIC. They were, together with the amphibolites and the supracrustal rocks of the BGC, imbricated and thrusted over the Roan Window. Though the degree of deformation and retrogression varies considerably between the different structural levels, there are strong similarities with respect to both the PT evolution and the small-scale structures. The rocks within the BGC thus experienced a tectonothermal history during and after thrusting which is similar to that of the rocks within the Roan Window. The synthesis outlined below is summarized in Table 4.

A: Early and/or prograde stages are preserved as assemblages within cores of zoned garnets. Contrasts in PT conditions between

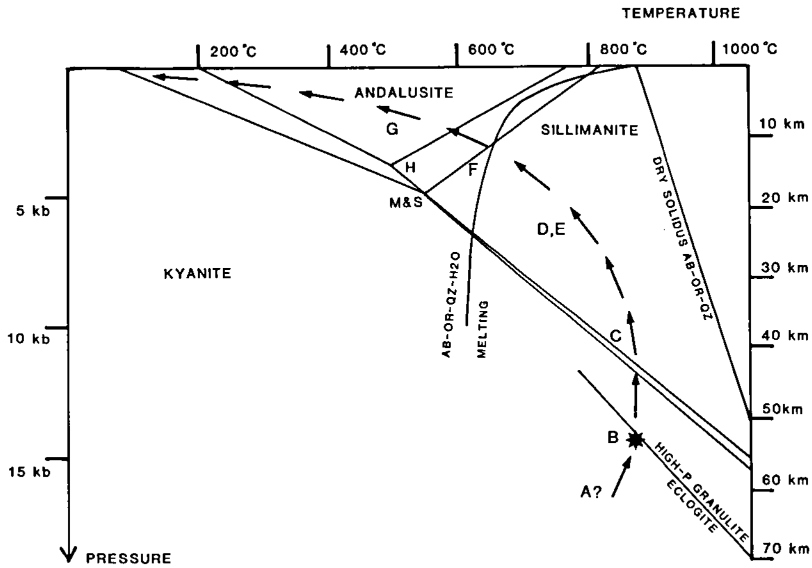

Fig. 17. Rough outline of the PT-evolution of the rocks within the Roan area. Al-silicate stability fields are after Holdaway (1971; H) and Mueller & Saxena (1977; M&S). The eclogite - high-pressure granulite transition for quartz-tholeiite is extrapolated from Green & Ringwood (1967, quartz-tholeiite B, Fig. 7), the beginning of melting in the Ab-Or-Q-H_2O system is from Merill et al. (1970), and the dry solidus Ab-Or-Q is from Mueller & Saxena (1977). A-G are approximate PT-positions of the stages outlined in Table 4.

the RIC and the ESU *might* have existed prior to peak PT conditions.

B: High-pressure granulite facies assemblages are found in all rock-types except for the pegmatites. Coexisting garnet and clinopyroxene occur in basic rocks within the RIC, the ESU and the orthogneisses, banded amphibolites and other supracrustal units of the BGC. Hydrous phases such as amphibole, biotite and zoisite were stable in some rocks. Geothermometry suggests temperatures well above 800°C at the peak of metamorphism. The pressure estimate at peak temperature conditions from the Roan Window is 14 ± 2 kbar. Abundant kyanite in peak PT textural relationships in the rocks within the ESU implies a minimum pressure of 11 kbar at temperatures of 800°C. The (D1) lineations and foliations in the contact zone between the RIC and BGC formed at these conditions.

C: High-pressure minerals were partly replaced by symplectites of high-temperature, medium-pressure phases. These assemblages formed during decompression into the sillimanite stability field, mostly at low-P_{H2O} conditions, in placed by dehydration of hydrous phases. The symplectites developed radially, apparently unrelated to deformation, at least on a small scale.

D: Amphibolitization caused partial to complete replacement of earlier anhydrous, high- and medium-pressure minerals. A regional (D2) foliation developed at amphibolite facies conditions and older migmatite structures were deformed.

E: During refolding (D3), new migmatite veins were developed regularly spaced along axial surfaces. This important structural time marker is common within the RIC, and has been found locally within the BGC. Within the RIC, an even later generation of migmatite veins containing hornblende poikiloblasts was found.

F: Later deformation with local mylonitization and isoclinal folding (D4) took place at lower amphibolite facies conditions. Hornblende-bearing and D3 neosome veins were deformed. Upright refolding (D5), mainly along NE-SW subhorizontal fold axes, took place and in places shear zones developed in fold limbs.

G: Late features include the intrusion and subsequent shearing of pegmatites. The rocks cooled through the andalusite stability field.

Intrusive and tectonic evolution: a summary and discussion

Intrusive evolution of the Roan Igneous Complex

Based on field criteria, the oldest rocks within the RIC are the intrusive sequence of quartz-monzodiorites to quartz-monzonites and possibly some basic rocks. The charnockitic rocks may be of the same age; however, they are younger than some basic rocks which they cross-cut and brecciate. The Na- and Ca-rich, zoned orthoclase-megacrysts in the charnockitic rocks, and the presence of clinopyroxene inclusions in these K-feldspars, suggests that these rocks crystallized in a dry, lower crustal environment. The quartz-monzodiorites, quartz-monzonites and charnockitic rocks are cross-cut by reddish granites. A red granite at Osen (10 km north of Roan), similar to the granites within the RIC, gave an upper intercept U-Pb zircon age of c. 1630 Ma (Schouenborg et al. in prep.).

Younger coronitic dolerites are common in the RIC. Similar dolerites are found over a large part of northern Vestranden as far north as Namsos (Fig. 1; L. Johansson pers. comm. 1983). The former coexistence of plagioclase and olivine in these rocks implies a pressure less than 9 kbar (Ito & Kennedy 1970); thus a crustal depth at intrusion of less than 30 km. Bulk chemistry of the dolerites suggests a correlation with the c. 1200 Ma (Patchett 1978) Central Scandinavian Dolerite Group (Johansson & Solyom in prep.).

Sm-Nd whole-rock isotope data suggest that the protolith age of the rocks within the Kråkfjord Complex is Early Ordovician (Johansson et al. in prep.). The emplacement (in situ intrusion *or* thrust emplacement) of the Kråkfjord and Brandsfjord layered basic complexes into the RIC took place prior to or contemporaneous with the high-pressure granulite facies metamorphism.

Tectonic evolution

The structural geometry of the rocks above the RIC, with repetitions of supracrustal rocks and orthogneisses, implies a thrust relationship, either as recumbent fold nappes or as an imbricate stack of thrust sheets. The RIC thus constitutes a window which contains better preserved rocks, tectonically overlain by the orthogneisses, amphibolites and supracrustal rocks of the BGC.

No metadolerites or cross-cutting basic rocks have been found within the ESU. Such dolerites are common within the RIC and the orthogneisses of the BGC, which suggests that the metasedimentary rocks are either younger than the dolerites or allochthonous, or both. No evidence for deposition of the metasedimentary rocks upon the RIC (unconformities or a basal facies) has been found within the supracrustal rocks. However, the actual amount of thrust displacement is not known, and no stratigraphic or lithological correlations with established Caledonian allochthonous units could be made with any certainty. It is therefore not possible yet to decide whether the ESU represents a cover sequence originally deposited directly upon the RIC, or is a far-travelled thrust sheet.

The RIC may extend to depths of 12-16 km, as indicated by geophysical data (Skyseth 1987). If the thrust zone above the Roan Window extends to these depths, major parts of the orthogneisses and the supracrustal rocks in northern Vestranden must be regarded as allochthonous and this thrust zone may be of major regional importance.

The peak PT-event. The supracrustal and igneous rocks in the Roan area, except late pegmatites, experienced high-pressure granulite facies metamorphism, and went through a similar tectonothermal history during subsequent uplift and deformation. The juxtaposition of these units was thus most likely established before or during the peak of metamorphism. Locally preserved high-pressure granulite-facies foliations and NW-SE lineations in rocks within the RIC and the ESU indicate that the thrust zone above the Roan Window was active at the peak of metamorphism. The thrust zone may have been initiated in response to crustal imbrication at deep levels. The Sm-Nd age of ca. 425 Ma of a high-pressure granulite facies assemblage from the Kråkfjord Complex (Johansson et al. in prep.) is consistent with this event being related to depression of the western margin of Baltica during Caledonian (Scandian) continent-continent collision. This may support the model presented by Cuthbert et al. (1983), which proposes imbrication of the margin of Baltica as the cause of crustal thickening and eclogite facies metamorphism in the WGR.

The high-PT conditions reached 870 $\pm$ 50°C and 14.5 $\pm$ 2 kbar (Johansson & Möller 1986). Assuming an average density of 2.7 g/cm^3 for the overburden, the pressure estimate gives a depth of burial exceeding 50 km. Compared to the peak metamorphic conditions recorded for the WGR south of Kristiansund (Griffin et al. 1985), higher temperatures and lower pressures prevailed in the Roan area.

The uplift. Subsequent to the high-pressure metamorphism, the rocks experienced a marked decompression, generally under low-P_{H2O} conditions. Medium-pressure granulite facies assemblages are found in the form of symplectites. These symplectites developed radially, unrelated to deformation, at least on the small scale. Relatively low pressures, possibly as low as 6 kbar at high temperatures (700-800°C), are indicated by the formation of spinel + cordierite. This implies a near isothermal uplift, which together with the preservation of the peak PT-assemblages and the symplectite reaction textures, despite the high temperatures during the retrogression, point to a rapid uplift.

Subsequent extensive deformation took place at amphibolite facies conditions. The ages of 430 Ma for dyke intrusion in the Fosslia cover sequence (Johansson et al. 1987a) and 425 Ma for the peak of metamorphism in the rocks of the Roan Window (Johansson et al. in prep) set an upper limit for the regional amphibolite facies deformation at Mid Silurian. The retrograde deformation was in part coeval with migmatitization.

The rocks then cooled through the andalusite stability field. High-T/P conditions during the later part of the uplift is a regional feature in northern Vestranden (Johansson et al. 1987b). $^{40}Ar/^{39}Ar$-data from hornblendes give cooling ages (500 °C) scattered around 410 Ma (R.D. Dallmeyer pers. comm. 1987).

Thermal models of overthrusted regions (England & Thompson 1984) suggest that the crust is unlikely to experience such high temperatures at shallow levels if uplift is strictly controlled by erosion. Uplift caused by tectonic unroofing could raise the crust to shallow levels rapidly enough for it to still remain hot. Another possible heat source is that of late plutonic activity; however, the late Caledonian intrusive activity in Vestranden appears to be restricted to the formation of thin pegmatites (c. 390-410 Ma old, Pidgeon & Råheim 1972, Scouenborg 1988). The thermal evolution during uplift, and the extensive deformation and thrust reactivation under amphibolite facies conditions upport a tectonic model where the uplift of the WGR is aided by tectonic unroofing.

Acknowledgements

This paper is a part of a Ph.D. thesis carried out at the Department of Mineralogy and Petrology, Institute of Geology, University of Lund. The fieldwork was financially supported by the Swedish Natural Science Research Council (NFR; grant no. 3559-130). Financial support for mapping was also received from the Geological Survey of Norway (NGU).

My warmest thanks go to Leif Johansson for field company, almost daily discussions, and encouragement during all aspects of my work. Leif introduced me to the Roan area, and inspired much of the present project. He took part in the mapping of all the islands, and was an excellent, patient skipper. My work has also benefited from daily company and discussions at the department with Jan-Erik Lindqvist and Björn Schouenborg. Per-Gunnar Andréasson, David Gee, Jane Gilotti and Roland Gorbatschev are thanked for discussions both in the field and at the department, and for their critical reading and improvement of various versions of the manuscript. Inge Bryhni and another, unknown, referee are thanked for constructive reviews. Rognvald Boyd, Allan Krill and David Roberts kindly read and gave valuable comments on the manuscript.

The geologists participating in the Nord-Trøndelag project of the Geological Survey of Norway (NGU) in Trondheim, including Trond Skyseth who worked on geophysics in the Roan area during the autumn of 1986, are thanked for stimulating discussions and meetings. Christin Andréasson is thanked for the drawing of some of the figures and Josef Gersner for preparation of the thin-sections.

Special thanks go to Karen-Marie and Erling Eian, and their family, for hospitality and company during field work in Roan.

References

Andréasson, P.-G. & Gorbatschev, R. 1980: Metamorphism in extensive nappe terrains: a study of the Central Scandinavian Caledonides. *Geol. Fören. Stockh. Förh. 102,* 335-357.

Andréasson, P.-G. & Johansson, L. 1983: The Snåsa Megalens, west-central Scandinavian Caledonides. *Geol. Fören. Stockh. Förh. 104,* 305-326.

Austrheim, H. & Griffin, W.L. 1985: Shear deformation and eclogite formation within granulite-facies anorthosites of the Bergen Arcs, western Norway. *Chem. Geol. 50,* 267-281.

Birkeland, T. 1958: Geological and petrological investigations in northern Trøndelag, western Norway. *Nor. Geol. Tidsskr. 38,* 327-420.

Boyd, R. 1986: Bedrock mapping within the Nord-Trøndelag Project, with some observations on the Geology of the Namsos area. *Geol. Fören. Stockh. Förh. 108,* 270-271.

Broch, O.A., 1961: Quick identification of potash feldspar, plagioclase and quartz for quantitative thin section analysis. *Am. Min. 46,* 752-775.

Brown, G.C. 1981: Space and time in granite plutonism. *Phil. Trans. Roy. Soc. London A301,* 321-336.

Bryhni, I. 1988: Status of the supracrustal rocks in the Western Gneiss Region, South Norway. *In:* Gayer, R. (ed.) *Caledonian and Related Geology of Scandinavia.* Graham and Trotman, in press.

Bryhni, I., Krogh E.J. & Griffin, W.L. 1977: Crustal derivation of Norwegian eclogites: a review: *Neues Jahrb. Mineral. Abh. 130,* 49-68.

Cotkin, S.J., Valley, J.W. & Essene, E.J. 1988: Petrology of a margarite-bearing meta-anorthosite from Seljeneset, Nordfjord, Western Norway: Implications for the P-T history of the Western Gneiss Region during Caledonian uplift. *Lithos 21,* 117-128.

Cuthbert, S.J., Harvey, M.A. & Carswell, D.A. 1983: A tectonic model for the metamorphic evolution of the Basal Gneiss Complex, Western South Norway. *J. Metam. Geol. 1,* 63-90.

Ellis, D.J. & Green, A.W. 1979: An experimental study of the effect of Ca upon garnet-clinopyroxene Fe-Mg exchange equilibria. *Contrib. Mineral. Petrol. 71,* 13-22.

England, P.C. & Thompson, A.B. 1984: Pressure-temperature-time paths of regional metamorphism I. Heat transfer during the evolution of regions of thickened continental crust. *J. Petrol. 25,* 894-928.

Gee, D.G. 1975: A geotraverse through the Scandinavian Caledonides - Östersund to Trondheim. *Sver. Geol. Unders. Ser. C. 417,* 1-66.

Gee, D.G. 1980: Basement-cover relationships in the central Scandinavian Caledonides. *Geol. Fören. Stockh. Förh. 102,* 455-474.

Gee, D.G. 1986: Middle and upper crustal structure in the central Scandinavian Caledonides. *Geol. Fören. Stockh. Förh. 108,* 280-283.

Gee, D.G. & Zachrisson, E. 1979: The Caledonides in Sweden. *Sver. Geol. Unders. C 769,* 1-48.

Gee, D.G., Kumpulainen, R., Roberts, D., Stephens, M.B., Thon, A. & Zachrisson, E., 1985a: Scandinavian Caledonides: Tectonostatigraphic Map. *In:* Gee, D.G. & Sturt, B.A. (eds.) *The Caledonide Orogen - Scandinavia and Related Areas.* John Wiley & Sons Chichester, 843-857.

Gee, D.G., Guezou, J.C., Roberts, D. & Wolff, F.C. 1985b: The central-southern part of the Scandinavian Caledonides. *In:* Gee, D.G. & Sturt, B.A. (eds.) *The Caledonide Orogen - Scandinavia and Related Areas.* John Wiley & Sons Chichester, 109-133.

Gorbatschev, R. 1985: Precambrian basement of the Scandinavian Caledonides. *In:* Gee, D.G. & Sturt, B.A. (eds.) *The Caledonide Orogen - Scandinavia and Related Areas.* John Wiley & Sons Chichester, 197-212.

Green, D.H. & Ringwood, A.E. 1967: An experimental investigation of the gabbro to eclogite transformation and its petrological applications. *Geochim. Cosmochim. Acta 31,* 767-833.

Griffin, W.L. 1987: 'On the eclogites of Norway' — 65 years later. *Min. Mag. 51,* 333-343.

Griffin, W.L, Austrheim, H., Brastad, K., Bryhni, I., Krill, A.G., Krogh, E.J., Mørk, M.B.E., Qvale, H. & Tørudbakken, B. 1985: High-pressure metamorphism in the Scandinavian Caledonides. *In:* Gee, D.G. & Sturt, B.A. (eds.) *The Caledonide Orogen — Scandinavia and Related Areas.* John Wiley & Sons Chichester, 783-801.

Griffin, W.L. & Brueckner, H.K. 1980: Caledonian Sm-Nd ages and a crustal origin for Norwegian eclogites. *Nature 285,* 319-321.

Griffin, W.L. & Brueckner, H.K. 1985. REE, Rb-Sr and Sm-Nd studies of Norwegian eclogites. *Chem. Geol. (Isot. Geosc.) 52,* 249-271.

Grønlie, A. & Roberts, D. 1988: Dextral strike-slip duplexes of probable Mesozoic age along the Verran and Hitra-Snåsa Faults, Møre-Trøndelag Fault Zone, Central Norway. *Abstract. 18:e Nord. Geol. Vintermøde, København* 143-144.

Harley, S.L. & Green, D.H. 1982: Garnet- orthopyroxene barometry for granulites and peridotites. *Nature 300,* 697-701.

Holdaway, M.J. 1971: Stability of andalusite and the aluminum silicate phase diagram. *Am. J. Sci. 271,* 97-131.

Hossack, J. 1984: The geometry of listric growth faults in the Devonian basins of Sunnfjord, W Norway. *J. geol. Soc. London 141,* 629-637.

Hossack, J. & Cooper, M.A. 1986: Collision tectonics in the Scandinavian Caledonides. *In:* Coward, M.P. & Ries, A.C. (eds.), *Collision Tectonics.,* Geol. Soc. Spec. Publ. 19, 287-304.

Ito, K. & Kennedy, G.C. 1970: Fine structure of the basalt to eclogite transition. *Spec. Pap. Min. Soc. Am. 3,* 77-83.

Jamtveit, B. 1987: Metamorphic evolution of the Eiksunddal eclogite complex. Western Norway, and some tectonic implications. *Contrib. Mineral. Petrol. 95,* 82-99.

Johansson, L. 1986a: *Basement and cover relationships in the Vestranden - Grong-Olden region, Central Scandinavian Caledonides.* Unpubl. Ph.D.-thesis, Univ. of Lund, 142 pp.

Johansson, L. 1986b: Mega-lenses and Scandian deformation in the basement in the northern part of Western Gneiss Region, Vestranden, Central Norwegian Caledonides. *Geol. Fören. Stockh. Förh. 108,* 287-289.

Johansson, L. & Möller, C. 1986: Formation of sapphirine during retrogression of a basic high-pressure granulite, Roan, Western Gneiss Region, Norway. *Contrib. Mineral. Petrol. 94,* 29-41.

Johansson, L., Andréasson, P.G. & Schöberg, H. 1987a: An occurrence of the Gula Nappe in the Western Gneiss Region, Central Norwegian Caledonides. *Nor. Geol. Tidsskr. 67,* 85-92.

Johansson, L., Lindqvist, J.-E. & Möller, C. 1987b: High T/P-gradients during retrograde metamorphism in the central Scandinavian Caledonides. *Geol. Fören. Stockh. Förh. 109,* 347-349.

Kjerulf, T. 1871: Om Trondheims stifts geologi. *Nyt. Mag. for Naturv. Christiania 18,* 1-79.

Krill, A.G. 1980: Tectonics of the Oppdal area, central Norway. *Geol. Fören. Stockh. Förh. 102,* 523-530.

Krill, A.G. 1985: Relationships between the Western Gneiss Region and the Trondheim Region: Stockwerk-tectonics reconsidered. *In:* Gee, D.G. & Sturt, B.A. (eds.) *The Caledonide Orogen - Scandinavia and Related Areas.* John Wiley & Sons Chichester, 475-483.

Krill, A.G. & Sigmond, E.M.O. 1987: Surnadalens dekkelagfølge og dens fortsettelse mot vest. *Abstract. NGF X Landsmøte. Nor. Geol. Foren. Medlemsblad Nr. 21.* 35.

Krogh, E.J. 1977: Evidence for a Precambrian continent-continent collision in western Norway. *Nature 267,* 17-19.

Krogh, E.J. 1980: Compatible PT conditions for eclogites and surrounding gneisses in the Kristiansund area, Western Norway. *Contrib. Mineral. Petrol. 75,* 387-393.

Krogh, E.J. 1982: Metamorphic evolution of Norwegian country-rock eclogites, as deduced from mineral inclusions and compositional zoning in garnets. *Lithos 15,* 305-321.

Krogh, T.E., Mysen, B.O. & Davis, G.L. 1974: A paleozoic age for the primary minerals of a Norwegian eclogite. *Ann. Rep. Geophys. Lab. Carnegie Inst. Washington 73,* 575-576.

Kullerud, L., Tørudbakken, B. & Ilebekk, S. 1986: A compilation of radiometric age determinations from the Western Gneiss Region, South Norway. *Nor. geol. unders. Bull. 406,* 17-42.

Mearns, E.W. 1986: Sm-Nd ages for Norwegian garnet peridotite. *Lithos 19,* 269-278.

Mearns, E.W. & Lappin, M.A. 1982: A Sm-Nd isotopic study of 'internal' and 'external' eclogites, garnet lherzolite and grey gneiss from Almklovdalen, Western Norway. *Terra Cognita 2,* 324.

Merill, R.B., Robertson, J.K. & Wyllie, P.J. 1970: Melting reaction in the system $NaAlSi_3O_8$-$KAlSi_3O_8$-SiO_2-H_2O at 20 kilobars compared with results for other feldspar-quartz-H_2O and rock-H_2O systems. *J. Geol. 78,* 558-569.

Mueller, R.F. & Saxena, S.K. 1977: *Chemical Petrology.* Springer Verlag, New York. 394 pp.

Mykkeltveit, S., Husebye, E.S. & Oftedahl, C. 1980: Subduction of the lapetus Ocean crust beneath the Møre Gneiss Region, Southern Norway. *Nature 288,* 473-475.

Mørk, M.B.E. 1985: A gabbro to eclogite transition in Flemsøy, Sunnmøre, western Norway. *In:* D.C. Smith, G. Granz and D. Gebauer (Guest-Eds.), *Chemistry and Petrology of Eclogites. Chem. Geol. 50,* 283-310.

Mørk, M.B.E. & Mearns, E.W. 1986: Sm-Nd isotopic systematics of a gabbro-eclogite transitions. *Lithos 19,* 255-267.

Mørk, M.B.E., Kullerud, K. & Stabel, A. 1988: Sm-Nd dating of Seve eclogites, Norrbotten, Sweden — evidence for Early Caledonian (505 Ma) high-pressure metamorphism. *Abstract. 18:e Nord. Geol. Vintermøde, København.* 289-290.

Norges Geologiske Undersøkelse 1983: *Aeromagnetic map Namsos. 1:250.000.*

Norges Geologiske Undersøkelse 1985: *Bouger gravity anomaly map South Norway.*

Norton, M. 1987: The Nordfjord-Sogn Detachment, W. Norway. *Nor. Geol. Tidsskr. 67,* 93-106.

Patchett, P.J. 1978: Rb-Sr ages of Precambrian dolerites and syenites in southern and central Sweden. *Sver. Geol. Unders. Ser. C 747,* 1-63.

Pearce, J.A., Harris, N.B.W. & Tindle, A.G. 1984: Trace element discrimination diagrams for the tectonic interpretation of granitic rocks. *J. Petrol. 25,* 956-983.

Pidgeon, R.T. & Råheim, A. 1972: Geochronological investigation of the gneisses and minor intrusive rocks from Kristiansund, west Norway. *Nor. Geol. Tidsskr. 52,* 48-68.

Ramberg, H. 1943: En undersøkelse av Vestrandens regionalmetamorfe bergarter. *Nor. Geol. Tidsskr. 23,* 1-174.

Ramberg, H. 1966: The Scandinavian Caledonides as studied by centrifuged dynamic models. *Bull. Geol. Inst. Uppsala, 43,* 1-72.

Roberts, D. 1983: Devonian tectonic deformation in the Norwegian Caledonides and its regional perspectives. *Nor. geol. unders. Bull 380,* 85-96.

Roberts, D. 1986: Structural - photogeological and general tectonic features of the Fosen-Namsos Western Gneiss Region of Central Norway. *Nor. geol. unders. Bull. 407,* 13-25.

Roberts, D. & Gee, D.G. 1985: An introduction to the structure of the Scandinavian Caledonides. *In:* Gee, D.G. & Sturt, B.A. (eds.) *The Caledonide Orogen - Scandinavia and Related Areas.* John Wiley & Sons Chichester, 55-68.

Schouenborg, B.E. 1986: An allochthonous cover in northern Vestranden, Western Gneiss Region, central Norway. *Geol. Fören. Stockh. Förh. 108,* 127-133.

Schouenborg, B.E. 1988: U/Pb-zircon datings of Caledonian cover rocks and cover-basement contacts, northern Vestranden, central Norway. *Nor. Geol. Tidsskr. 68,* in press.

Sigmond, E.M.O., Gustavson, M. & Roberts, D. 1984: Berggrunnskart over Norge. M. 1:1 million. *Norges Geologiske Undersøkelse.*

Skyseth, T. 1987: Geofysisk og geologisk tolkning av aeromagnetiske og gravimetriske anomalier på og rundt kartblad 1523 II Stokksund og 1623 III Roan, Sør-Trøndelag. *Unpubl. report. Norwegian High-School of Technology (NTH), Trondheim,* 112 pp.

Smith, D.C. 1980: A tectonic melange of foreign eclogites and ultramafites in west Norway. *Nature 287,* 366-367.

Solyom, Z., Andréasson, P.G., Johansson, I. & Hedvall, R. 1984: Petrochemistry of late Proterozoic rift volcanism in Scandinavia. 1. The Blekinge-Dalarna Dolerites (BDD)-volcanism in a failed arm of lapetus? *Lund Publications in Geology,* 23, 56 pp.

Stephens, M.B., Gustavson, M., Ramberg, I.B. & Zachrisson, E. 1985: The Caledonides of central-north Scandinavia — a tectonostratigraphic overview. *In:* Gee, D.G. & Sturt, B.A. (eds.) *The Caledonide Orogen - Scandinavia and Related Areas.* John Wiley & Sons Chichester, 135-162.

Streckeisen, A. 1976: To each plutonic rock its proper name. *Earth Sci. Rev., 12,* 1-33.

Torsvik, T.H., Sturt, B.A., Ramsay, D.M., Kisch, H.J. & Bering, D. 1986: The tectonic implications of Solundian (Upper Devonian) magnetization of the Devonian rocks of Kvamshesten, western Norway. *Earth Planet. Sci. Lett. 80,* 337-347.

Tucker, R.D. 1986: Geology of the Hemnefjord-Orkanger Area, South Central Norway. *Nor. geol. unders. Bull. 404,* 1-20.

Tucker, R.D., Råheim, A., Krogh, T.E. & Corfu, F. 1987: Uranium-lead zircon and titanite ages from the northern portion of the Western Gneiss Region, south-central Norway. *Earth Planet. Sci. Lett. 81,* 203-211.