The Tectonostratigraphy and Structural Evolution of the Sitas Area, North Norway and Sweden (68° N)

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The Sitas area contains rocks derived from the pre-Caledonian Scandinavian craton and from an accreted Caledonian fore-arc complex. These rocks form the par-autochthonous(?) Rombak Complex which can be tied stratigraphically to Baltica and three allochthons which from the base upwards are: the Storriten Complex, the Langvatn nappe and the Marko nappe. The Storriten Complex is a schuppen-zone composed of rocks derived from hinterland equivalents of the Rombak Complex while the Langvatn and Marko nappes are part of an accreted fore-arc terrane which can be correlated with parts of the Middle Köli Nappe Complex (MKNC). The Marko nappe contains mafic rocks whose geochemical affinities suggest derivation from both MORB-like and arc-like sources.

The MKNC was deformed by a series of five superposed Caledonian deformations (D_1-D_5). Only the latest two deformations that affected the MKNC, D_5 & D_4, are recognized in the Storriten and Rombak Complexes. For this reason, the MKNC cannot be tied directly to the underlying domains and therefore to Baltoscandia until D_4 when it was thrust onto the Storriten and Rombak Complexes. The earliest three deformations may have occurred outboard of the lower Paleozoic margin of Baltoscandia. The two earliest deformations (D_1 & D_2) occurred under moderate pressure amphibolite-facies conditions (575° C, 800 MPa) while later deformations occurred under lower P-T conditions. By D_4 the MKNC had cooled to greenschist-facies conditions (T $<$ 450° C, P $<$ 500 MPa).

Most of the strain in the Sitas region was accommodated by large-scale ESE-directed thrusting during D_2 and D_4. D_2 thrusting imbricated the rocks in the Marko nappe and emplaced the Marko nappe onto the Langvatn nappe. D_4 thrusting placed the MKNC onto the Storriten Complex and the Storriten Complex on the Rombak Complex. In addition, it detached the sedimentary cover of the Rombak Complex from its crystalline basement, imbricated the cover and eventually delaminated the uppermost metres to hundreds of metres of the basement forming the Storriten Complex. Below this level, the crystalline basement was not penetratively deformed by Caledonian events.

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Introduction

The Sitas area (Plate 1) is located on the southern margin of the Rombak window, a large fenster of pre-Caledonian basement beneath the Caledonian nappe pile that straddles the Norwegian-Swedish border between 68° and 68° 30' N (Fig. 1). Precambrian granitic gneisses within the Rombak window are locally overlain by the Torneträsk Formation, a very thin Vendian to Lower Cambrian platform sedimentary sequence (Thelander 1982) that correlates with the 'Hyolithus' zone sequence (Kulling 1964) of the Baltic craton. For this reason, the Rombak window can be considered to represent an extension, although perhaps a displaced one (Tilke 1986) of the pre-Caledonian Baltic craton beneath the Caledonian nappe pile. The continental basement of the Rombak window is structurally overlain by an allochthonous terrane of medium-grade metamorphic rocks that are, at least in part, oceanic in origin. The continental basement of the Rombak window was at a mid-crustal depth when this metamorphic allochthon was accreted to it. The Sitas area thus provides an excellent opportunity to study, by direct observation, the geometry of structures that form in response to the accretion of an oceanic terrane at an intermediate crustal level.

This paper summarizes the results of an integrated geological, petrological and structural study that was based on field work carried out during the summers of 1981-1983 and 1985. The study concentrated on: (1) defining a tectonic stratigraphy within the allochthonous Caledonian rocks, (2) determining the nature of the contact between the pre-Caledonian basement and the Caledonian nappes, and (3) determining the pressure-temperature conditions at which the Caledonian nappes were accreted to Scandinavian continental base-

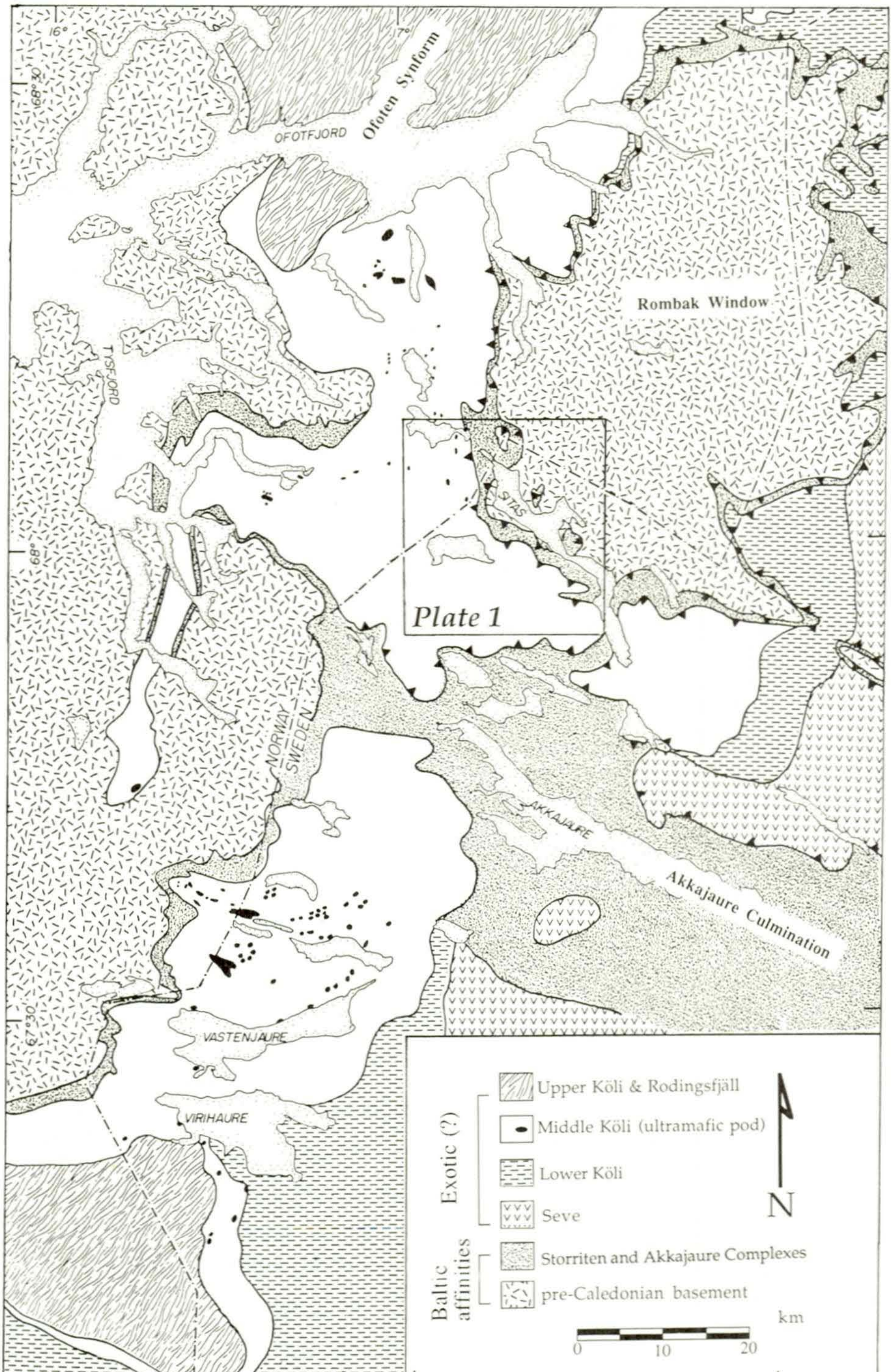

Fig. 1. Simplified tectonic map of the Sitas region. (Compiled from Foslie 1941, 1942, Kautsky 1953, Kulling 1964, 1982, Gustavson 1974, & Hodges 1985).

ment. This report emphasizes the stratigraphic and structural aspects of this study; the metamorphic petrology of the rocks in the study area has been discussed elsewhere (Crowley & Spear 1987, Crowley 1988).

The Sitas region can be divided into 3 tectonic domains separated by two major low-angle post-metamorphic thrust faults, the Frostisen and Mereftasfjell thrusts (Hodges 1985). Within each of the three tectonic domains there is a distinct tectonic stratigraphy. Beneath the Frostisen thrust is the structurally lowest domain, the Rombak Complex composed of the Rombak window granitic gneisses and isolated outliers of the Torneträsk Formation that are in depositional contact with the gneisses. Between the Frostisen and Mereftasfjell thrusts is the Storriten Complex (Hodges 1985) a sliver zone of phyllonites and mylonites derived from lithologies similar to the underlying Rombak Complex. Above the Mereftasfjell thrust, a structurally complex sequence of metasedimentary and meta-igneous rocks crop out. This uppermost domain contains rocks that I correlate with the Middle Köli Nappe Complex (MKNC) of Stephens (1980) and Stephens et al. (1985) exposed in the southern Norrbotten mountains. Within the Sitas area, the MKNC is composed of two thrust nappes, a lower Langvatn nappe and an upper Marko nappe. To the east of the Sitas area, both the Seve and the Lower Köli nappes appear as eastward-thickening wedges between the Storriten Complex and MKNC. The Sitas area is located 15 km west of the westernmost exposures of Lower Köli rocks and 10 km west of the westernmost exposures of Seve rocks.

All of the rocks in the Sitas area were regionally metamorphosed during Caledonian deformation. The metamorphic grade and history of each structural domain are different. The metamorphic grade increases structurally upwards. Caledonian metamorphic conditions in the Rombak Complex never exceeded the middle greenschist facies (biotite grade). Storriten Complex metamorphism reached the upper greenschist facies (garnet grade) and the MKNC was metamorphosed to the amphibolite facies.

Previous work

The Norwegian portion of the study area was initially mapped by Foslie (1941) and published at a scale of 1:100,000. This was included in Gustavson's (1974) 1:250,000 map of the Narvik region. As part of a reconnaissance study of northern Norrbotten county, Kulling (1964) produced the first map of the Swedish portion of the area at a scale of 1:400,000. These maps were valuable guides for this study; however, significant differences exist between my map (Plate 1) and theirs, particularly in areas of difficult access.

This study continues a northwest-southeast transect across the Scandinavian Caledonides from the Lofoten Islands to the Caledonian foreland that was started by Tull (1978), Hakkinen (1977), Bartley (1984) and Hodges (1985). The 20 km transect from the Efjord culmination to the Rombak window, mapped at 1:50,000 by Hodges (1985) is contiguous with the Sitas area. The transect has been continued to the southeast of the Sitas area by Tilke (1986). Part of the Rombak window to the north of this study has been mapped at 1:100,000 (Birkeland 1976). South of the Sitas area, Bjorklund (1985) has mapped and described a thrust complex known locally as the Akkajaure Complex.

Tectonic stratigraphy

Within the Sitas region, 29 mappable lithological units were recognized (Plate 1). All of these units have been multiply deformed and metamorphosed. Contacts between many of the lithological units are clearly tectonic, whereas other contacts are gradational and are assumed to be depositional. Still other contacts are of uncertain origin; these are sharp contacts, but do not appear to be zones of concentrated strain and may be either depositional or tectonic. Although many contacts are tectonic, the lithological units appear in a regular vertical order and form a local tectonic stratigraphy.

The stratigraphy of this region was first described by Foslie (1941). Foslie did not recognize the presence of tectonic contacts within the Caledonian nappe pile and considered the Rombak window basement granitoids to be Caledonian age intrusive rocks. Kulling (1964) provided the first detailed descriptions of the Vendian to Upper Cambrian stratigraphy within the Rombak and Storriten Complexes. In his discussions of the stratigraphy within the MKNC, Kulling relied heavily on the work of Foslie (1941, 1942) but did recognize the tectonic nature of the contact between the basement granitoids and the Caledonian metasediments and of some of the contacts within the nappe pile. Gustavson (1966, 1972) proposed the first detailed tectonic stratigraphy for the high-grade Caledonian metasediments at this latitude. This tectonic stratigraphy was modified and subdivided into a local tectonostratigraphy by Hodges (1985). Although this study has relied heavily on Hodges' interpretations, several major differences exist between the tectonic stratigraphy presented below (Fig. 2) and that of Hodges (1985). The tectonostratigraphic nomenclature presented below is informal and is presented to facilitate reference to previously unnamed packages of rocks.

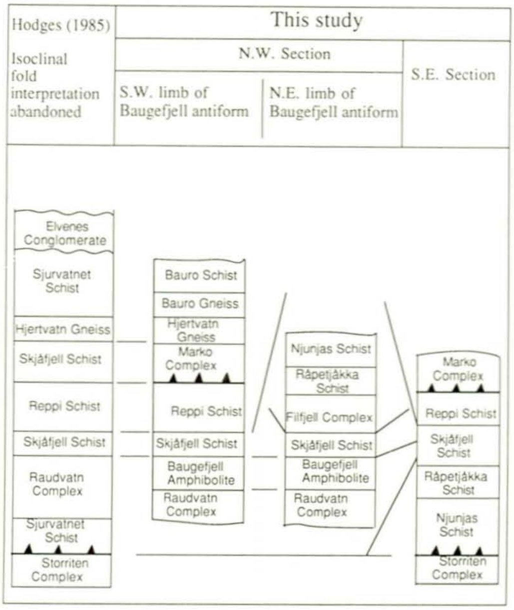

Fig. 2. Tectonic stratigraphy of the Sitas region. The tectonic stacking of lithological units in three parts of the study area is compared with that of the adjacent Efjord-Rombak transect of Hodges (1985).
Correlations with the Efjord region are not possible unless Hodges' isoclinal fold is abandoned.

Rombak Complex

The Rombak Complex is composed of both the basement gneisses that are exposed in the Rombak window and the sedimentary cover that is in depositional contact with the gneisses. It is dominated by the 1780±85 Ma (Gunner 1981) Rombak Granite-Gneiss, first described by Vogt (1942). The Rombak Granite-Gneiss is lithologically identical to and continuous with the Skjomen granite of Foslie (1941) and Kulling's (1964) Sjangeli granite. The granite-gneiss is a foliated, coarse-grained, porphyritic biotite granite with accessory apatite, magnetite and fluorite. Xenoliths and pendants of a supracrustal sequence composed of fine-grained amphibolite, amphibolitic schist, feldspathic quartzite, and muscovite-biotite schist are common within the gneiss. This lithological assemblage is called here the Skåddåive Assemblage, after the mountain Skåddåive where a large pendant of these lithologies crops out.

On the north slope of the mountain Råpetjåkka, the Rombak Granite-Gneiss is unconformably overlain by a thin (< 10 m) sedimentary sequence. This sequence includes quartz-pebble conglomerate, quartzite, and muscovite-rich phyllite which correlates with the Vendian lower sandstone and lower siltstone members of the Torneträsk Formation of Thelander (1982).

Storriten Complex The Storriten Complex is a structurally complex zone of tectonic slices in which thin, variably mylonitized sheets of granitic gneiss and micaceous quartzite are interleaved in a matrix of phyllonitic schist. Low-strain zones in slices of granitic gneiss and quartzite are lithologically similar to the Rombak Granite-Gneiss and basal quartzite of the Torneträsk Formation. In fact, the basal unconformity and quartz-pebble conglomerate of the Torneträsk Formation is preserved in several thrust slices. The Storriten matrix phyllonite is a heterogeneous muscovite + biotite ± garnet ± chlorite schist. It is commonly graphitic and in places calcareous. Very graphitic beds in the matrix schist are lithologically similar to the Middle to Upper Cambrian Alum Shale Formation. The calcareous beds in the Storriten matrix are of uncertain derivation as none are present in autochthonous sections described by Kulling (1964), Thelander (1982), or Tilke (1986) and perhaps suggest that rocks younger than those in the autochthonous sections (i.e. Ordovician) are contained in the Storriten Complex.

The Storriten Complex was mapped as the Middle Thrust Rocks by Kulling (1964) and is broadly correlative with other Middle Allochthon (Gee & Zachrisson 1979) rocks in the Norrbotten mountains. It differs from Kulling's typical Middle Allochthon in that it is dominated by metasedimentary rocks rather than granitic mylonites. Just to the south of the Sitas area, the proportion of granitic mylonite in the Storriten Complex increases and the proportion of metasediment decreases. Kautsky (1953), Bjorklund (1985) and B.C. Burchfiel (pers. comm. 1988) have mapped and described this granite mylonite complex as the Akkajaure Complex. On the western side of the Rombak window, Gustavson (1966, 1974) called similar rocks the Storfjell Group. At this latitude, the Storriten Complex, Middle Allochthon, Storfjell Group and Akkajaure Complex all represent the same tectonic level, the Storriten Complex being distinguished from the others by being dominated by metasedimentary rocks. These correlations are essentially similar to those proposed by Tull et al. (1985).

The abundance of metasedimentary rocks within the Storriten Complex may superficially suggest correlation of at least parts of the Storriten Complex with the sediment dominated Lower Thrust Rocks of Kulling (1964) or Lower Allochthon (Gee & Zachrisson 1979). However, it is my conviction that the Storriten Complex represents a single structural level which due to the presence of granitic mylonites is best correlated with the Middle Allochthon. Low-angle thrust faults within the Storriten Complex ramp upwards from a sole thrust, the Frostisen thrust to a roof thrust, the Mereftasfjell thrust to form a duplex structure (Plate 1 & Fig. 8). Some of these faults carry granitic basement in their hanging walls and cut across the entire Storriten Complex. This structural geometry requires the metasediment dominated parts of the Storriten Complex that superficially resemble the Lower Allochthon to sit both structurally above and below those parts of it which are more typical of the Middle Allochthon. I feel that this geometry effectively precludes the Storriten Complex from being a combination of the Lower and Middle Allochthons. Furthermore, the structurally lowest granitic mylonites in the Storriten Complex crop out on the southeast slope of the mountain Njunjas and appear to correlate with mylonites of the Akkajaure Complex (Bjorklund 1985) approximately 6 km to the south, supporting correlation of the Storriten Complex with the Middle Allochthon.

Middle Köli Nappe Complex

Within the Sitas area, the Middle Köli Nappe Complex (MKNC) is composed of eleven informal lithotectonic units. Fossils have not been found in any of these units; thus, their age is unknown but is assumed to be early Paleozoic. Primary structures are generally too poorly preserved to determine facing direction; therefore, the relative ages of the units in the MKNC stratigraphy are also uncertain.

The nomenclature used here is based on that of Hodges (1985). However, a major difference exists between this tectonic stratigraphy and that of Hodges. Lithological similarities above and below a calcareous schist, the Reppi Schist prompted Hodges (1985) and Tull et al. (1985) to postulate the existence of a major isoclinal fold with the Reppi Schist in its core. Although his mapping suggests that this fold should continue into the Sitas area, it does not. Furthermore, lithological correlations from the Sitas area to the contiguous Efjord-Rombak transect of Hodges are difficult if the lithotectonic sequence is isoclinally folded. Correlation is more straightforward if the lithologic symmetry about the Reppi Schist noted by Hodges (1985) is merely apparent and the Reppi Schist-cored isoclinal fold interpretation is abandoned. The tectonic stratigraphy presented below and in Fig. 2 abandons the isoclinal fold interpretation.

In this area, the MKNC can be divided into two thrust nappes, the structurally lower Langvatn nappe and the structurally higher Marko nappe. Each nappe has a unique tectonic stratigraphy and early metamorphic history. The two nappes are separated by an unnamed thrust fault across which they were juxtaposed subsequent to the metamorphic peak of rocks in the Marko nappe and at approximately the time of metamorphic peak of the Langvatn nappe (Crowley & Spear 1987).

Langvatn nappe

The Langvatn nappe is composed of a rather monotonous sequence of quartz-rich garnet two-mica schist and psammite. It has been informally divided into seven mappable lithotectonic units that are laterally continuous at least on a scale of kilometres to tens of kilometres. From the base upwards these units are: the Raudvatn Complex (Hodges 1985), Baugefjell Amphibolite, Skjåfjell Schist (Hodges 1985), Filfjell Complex, Njunjas Schist, Råpetjåkka Schist, and Reppi Schist (Foslie 1941).

These units are recognized by: (1) differences in schist composition, and (2) distinctive but volumetrically minor associated lithologies. The contacts between these units are parallel to the regional mica foliation (S_1) and are probably both depositional and tectonic.

The Raudvatn Complex is the structurally lowest unit in the Langvatn nappe that crops out in this area. It was informally named for exposures along the lake Raudvatn, 15 km north of Sitas (Hodges 1985). It is a tectonic melange composed of lenses of light-grey calcite marble, amphibolite, and ultramafic rocks in a matrix of flaggy weathering, garnet two-mica schist and calc-psammite. The tectonic lenses range in size from pods less than 1 m in diameter to lenses that are more than 400 m long. The amphibolites are hornblende-plagioclase-epidote amphibolites derived from both fine-grained (volcanic?) and coarse-grained (intrusive?) protoliths. Igneous textures (euhedral clinopyroxene pseudomorphs) are locally preserved in the ultramafic lenses but the igneous mineralogy has been altered to an amphibolite-facies serpentine + actinolite + forsterite ± talc assemblage.

The Baugefjell Amphibolite is a thin (0-50 m thick) but laterally persistent fine-grained, hornblende-plagioclase-epidote-sphene amphibolite. It is informally named here after the mountain Baugefjell on which large exposures of the amphibolite crop out. Fine-grained amphibolite is commonly interbedded with orange-weathering calcite marbles. The uniform fine grain size, intercalation of calcareous sediments, and the local preservation of plagioclase microlite pseudomorphs suggest derivation of the amphibolite from a volcanic protolith.

The Skjåfjell Schist is the 'typical' garnet two-mica schist of the Langvatn nappe. It was informally named by Hodges (1985) for exposures on the mountain Skjåfjell, 20 km NW of Sitas. It is silvery-grey, muscovite-rich and rhythmically layered on a scale of centimetres to 10's of centimetres. Graded bedding is locally preserved. Thin ($<$ 1 cm) quartz stringers and euhedral almandine garnet (1 mm to 5 cm) are conspicuous on the weathered surface. Interbeds of quartzitic schist, hornblende schist, calcite marble, and small lenses of amphibolite are present but are rare.

The Filfjell Complex is a unit of tectonic melange composed of lenses of calcite marble, amphibolite, and ultramafic rocks in a matrix of garnet two-mica schist and psammite. It is named here for the mountain Filfjell on which it is best exposed. The Filfjell matrix schist is lithologically similar to the Skjåfjell Schist. Ultramafic pods within the matrix are commonly altered to zoned bodies with serpentinite cores surrounded by a rind of talc-magnesite schist. Smaller ultramafic bodies are commonly entirely altered to talc-magnesite schist. The magnesite generally forms distinctive large (1-5 cm) euhedral poikioblasts.

Kulling (1964) mapped, but did not name a hornblende schist or garbenscheifer unit. It is informally named here the Råpetjåkka Schist after the mountain Råpetjåkka on which large hornblende schist outcrops are found. The Råpetjåkka Schist is a thick ($>$1m) bedded, calcareous, hornblende ± garnet schist that is interbedded with garnet 2-mica schist. The coarse-grained hornblende commonly makes garbenscheifer sprays on foliation surfaces. Thin, orange-weathering, calcite marble interbeds are widespread.

The Njunjas Schist is a heterogeneous sequence of garnet-free calcareous psammites and garnet two-mica schists. Interbeds of orange-weathering calcite marble and graphitic garnet schist are common. The lithological variation of the Njunjas Schist is its most distinctive characteristic. It is informally named here for exposures on the mountain Njunjas.

The Reppi Schist of Foslie (1941) is the most distinctive unit in the Langvatn nappe. It is a dark-brown weathering calcareous schist that forms flaggy outcrops due to the alternation of more and less calcareous bands a few to tens of centimetres thick. Garnet porphyroblasts and white clinozoisite laths are conspicuous on weathered foliation surfaces. At the southeast end of the Sitas area, on the north slope of the mountain Marko, a hornblende-in, clinozoisite-out isograd crosses the Reppi Schist. To the southeast, on the low-grade(?) side of this isograd, the Reppi Schist looks quite different. There, it is no longer dark-brown colored, but is a dark-green weathering hornblende schist. The alternation of more and less calcareous bands persists in the hornblende-bearing Reppi Schist, but it no longer produces the distinctive flaggy outcrops. The hornblende-bearing Reppi schist could be confused for the Råpetjåkka Schist except that the Reppi Schist is generally more calcareous and thinner bedded than the Råpetjåkka Schist.

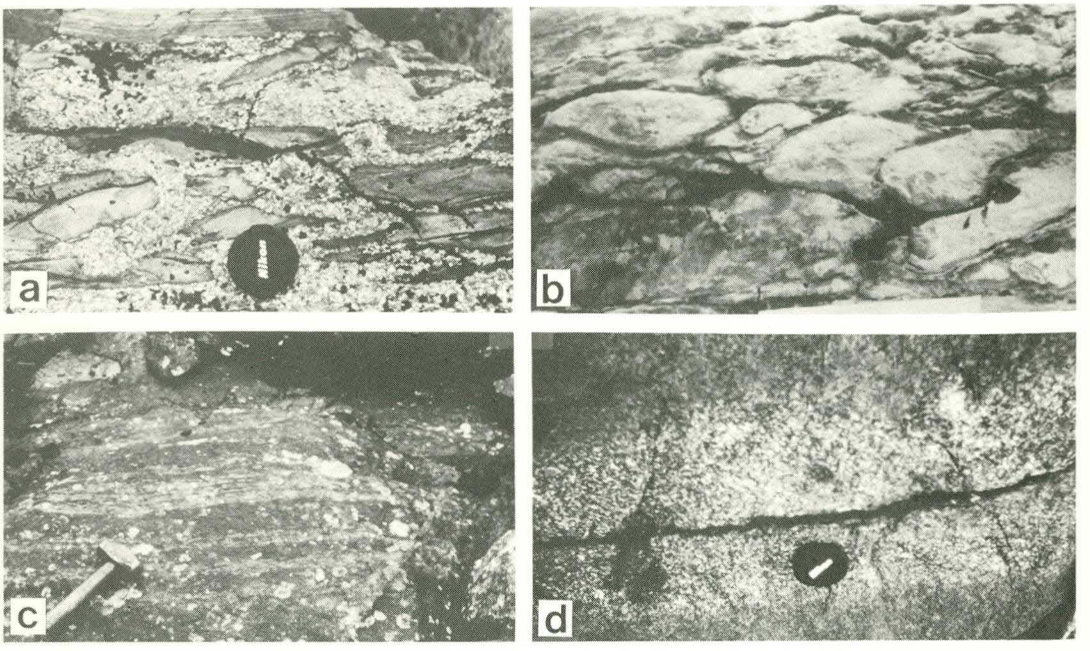

Fig. 3. Marko Complex lithologies
a) Elongate calcareous schist xenoliths in migmatitic gneiss.
b) Pillows in fine-grained amphibolite.
c) Primary mineralogical foliation in layered gabbro.
d) Alteration of igneous mineralogy to a greenschist-facies assemblage along fractures in layered gabbro.

Marko nappe

The Marko nappe has been informally divided into 4 lithotectonic units: the Marko Complex, Hjertevatn Gneiss, Bavro Gneiss, and the Bavro Schist. The bulk of the Marko nappe is made up of the Marko Complex, a structurally complicated unit composed of six distinct mappable lithologies: (1) garnet two-mica schist, (2) migmatitic gneiss, (3) calcareous schist, (4) amphibolite, (5) felsite, and (6) gabbro. Within the Marko Complex, these lithological units are repeated and truncated along low-angle faults (Fig. 9 & Plate 2, section F-F'). Strain is concentrated along these contacts and lithological layering (relict bedding?) is greatly attenuated along the contacts between different lithological units. This structural complexity and a lack of fossils or facing indicators has precluded the establishment of a stratigraphic sequence within the Marko Complex.

The Marko Complex garnet schist is a grey-weathering, quartz-rich, garnet two-mica schist. Hornblende is often present, but it rarely forms a coarse garbenscheifer texture. Small (cm to 10's of cm) amphibolite enclaves and thin amphibolitic dikes (< 1 m thick) that are locally discordant with lithological layering are common in the schist. The migmatitic gneiss is a coarse-grained biotite two-feldspar gneiss. It is commonly garnetiferous. Elongate psammitic to amphibolitic xenoliths (restite?) are present in most exposures of the gneiss (Fig. 3a). The calcareous schist unit is a thinly banded (1 to 10 cm) plagioclase-epidote-clinozoisite-biotite schist. Fine-grained (< 1mm) hornblende is everywhere present, but is not obvious in the field. The felsite unit is a fine-grained, light green-weathering, plagioclase and epidote-rich rock. It is presumed to be volcanic in origin, but no flow, pyroclastic or pillow structures are preserved.

A variety of amphibolitic rocks are contained in the Marko Complex. The mineralogy of the amphibolites is typically hornblende + plagioclase + epidote + sphene ± clinozoisite ± zoisite ± quartz. They vary from fine- to coarse-grained and are commonly strongly lineated. Locally, pillows are preserved and elongate structures reminiscent of stretched pillows are very common in the fine-grained amphibolites (Fig. 3b). Symmetrical chilled margins are preserved on some small bodies of amphibolite indicating an intrusive origin for some of the amphibolites. A large amphibolite sheet which crops out on the NE shore of the lake Bavrojavri (Plate 1) is composed of coarse-grained amphibolite that has been intruded by multiple generations of fine-grained amphibolite dikes.

A tectonic slice of layered olivine two-pyroxene gabbro crops out on the south slope of the mountain Marko. Cyclical variations in the olivine/pyroxene ratio and the planar alignment of plagioclase laths produce a mineralogical (cumulate?) foliation in this gabbro (Fig. 3c). Assuming that this foliation represents cumulate layering and thus approximates a paleohorizontal surface, the gabbro represents a tectonic slice derived from a mafic intrusive that was at least 2.5 km thick. Throughout much of the body, the igneous plagioclase + clinopyroxene ± olivine ± orthopyroxene mineralogy has been preserved. Locally, along joints that are perpendicular to the cumulate layering, the igneous mineralogy has been altered to a greenschist-facies albite-actinolite assemblage (Fig 3d). The mineralogical foliation in the gabbro is at a high angle to both the contact with the surrounding metasedimentary rocks and the mica foliation in those rocks. Within approximately 10 m of its contact with the metasedimentary rocks, the gabbro foliation becomes transposed so that it is approximately parallel to the contact. Along this contact, the gabbro has developed an amphibolite-facies hornblende-plagioclase mineral assemblage and is commonly invaded by felsic material believed to be migmatite derived from the surrounding metasedimentary rocks.

The Hjertevatn Gneiss of Hodges (1985) structurally overlies the Marko Complex. The gneiss is a plagioclase-rich banded gneiss composed of alternating tonalitic and amphibolitic layers. Thick bands (tens of m) of garnet schist are interleaved with the gneiss. Hodges (1985) suggested that the Hjertevatn Gneiss was a migmatite derived by partial melting of the surrounding metasedimentary units. A Rb-Sr isotopic study of thin slabs (‹ 1 cm) cut from two samples of the gneiss (Table 1) suggests that the gneiss represents a sea-water altered juvenile igneous rock. On a Rb-Sr isochron diagram (Fig. 4) the data produce a linear array with a negative slope. This array is interpreted to have been formed by two-component mixing of a juvenile magma ($^{87}Rb/^{86}Sr = 0.708$ $^{87}Sr/^{86}Sr_0 = 0.7040$) with Rb and Sr derived from sea-water ($^{87}Rb/^{86}Sr = 0.041$ $^{87}Sr/^{86}Sr_0 = 0.7085$).

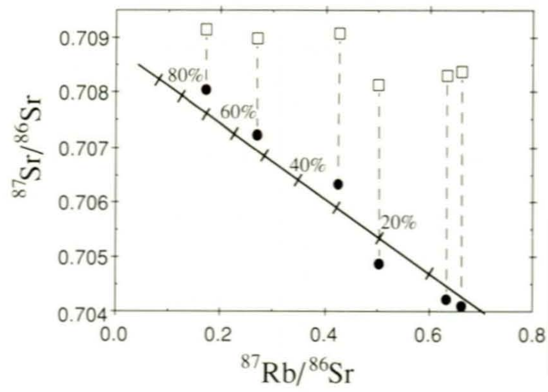

Fig. 4. Rb-Sr isochron diagram of thin slabs from the Hjertevatn Gneiss. Squares are present day values, circles are 450 Ma values. Present day data forms an array with a negative slope. This array cannot be interpreted in terms of an age of formation, but could represent a mixing line between a low $^{87}Sr/^{86}Sr$, high Rb/Sr component (juvenile magma) and a high $^{87}Sr/^{86}Sr$, low Rb/Sr component (sea-water). The mixing line is contoured in terms of percent of sea-water component in the mixture.

The Bavro Gneiss is a distinctive, resistant, dark-grey weathering, pelitic gneiss that is informally named here for exposures on the north shore of Bavrojarvi. Kyanite and small (1 mm), round, inclusion-free, almandine garnets are abundant in the gneiss. Small (cm) kyanite-bearing quartzo-feldspathic (migmatitic?) enclaves are common.

The Bavro Schist is a brownish-weathering two-mica schist that is informally named here for exposures on the north shore of the lake Bavrojavri. It is a very garnetiferous pelitic schist.

Table 1: Rb and Sr analyses from thin slabs cut from two Hjertevatn Gneiss samples.

	GT-24A	GT-24D	GT-29A	GT-29C	GT-29D	GT-29E
	Present-day values					
Rb (ppm)	44.2	45.9	30.6	40.2	13.9	35.6
Sr (ppm)	201.0	199.8	324.9	230.3	234.4	240.7
$^{87}Rb/^{86}Sr$	0.635	0.664	0.273	0.505	0.172	0.427
$^{87}Sr/^{86}Sr$	0.70833	0.70838	0.70899	0.70814	0.70916	0.7091
	450 Ma values					
$^{87}Rb/^{86}Sr_0$	0.635	0.663	0.273	0.504	0.172	0.427
$^{87}Sr/^{86}Sr_0$	0.70426	0.70413	0.70724	0.70491	0.70806	0.70636

Regional Correlations

Structural complications and the lack of persistent, distinctive lithologies within the MKNC make regional correlations difficult. However, several of the units mapped in this area are sufficiently similar to ones mapped in Ofoten (Foslie 1941, Gustavson 1974, Hodges 1985, Steltenpohl 1987) and southern Norrbotten (Foslie 1942, Kautsky 1953, Kulling 1982, Sundblad 1986, pers. comm. 1987) to propose correlations. These correlations are somewhat tenuous as much of the Sitas tectonostratigraphy cannot be traced either to the north across the Ofoten synform or to the south across the Akkajaure culmination.

Both the Langvatn and the Marko nappes continue into the Efjord region studied by Hodges (1985) and are part of Gustavson's (1966) Narvik Group. Gustavson (1972, 1974) placed an unnamed thrust fault within the Narvik Group, at the base of the Reppi Schist. Hodges (1985) and Tull et al. (1985) disputed the existence of this thrust arguing that the Reppi Schist cropped out in the core of an isoclinal fold. Continuing the Sitas tectonostratigraphy to the northwest into the Efjord region suggests that the thrust that separates the Langvatn and Marko nappes also continues into the Efjord region as units from both the Langvatn and Marko nappes are recognized by Hodges (1985). Within the Sitas area, the fault between the two nappes is at the top of the Reppi Schist. Therefore, I would continue this fault to the northwest and place a fault at the top of the Reppi Schist within the Efjord region.

Differences in the metamorphic petrology between the Langvatn and Marko nappes further support the continuation of both nappes and the thrust that separates them into the Efjord region. Marko nappe garnets have unzoned cores and sharply zoned rims (Fe/Mg increases markedly in the outer 100μ) while the Langvatn nappe garnets have cores that are more strongly zoned than the rims (Crowley & Spear 1987). Garnet zonation profiles were analyzed from Efjord samples by Hodges & Royden (1984). Most of their garnets came from schists that were structurally above the Reppi Schist. All of these profiles are similar to the Marko garnet profiles. One sample (79-18A) was collected from a schist below the Reppi Schist, and its zonation profile is similar to Langvatn garnet profiles. This is further evidence that a major fault within the MKNC continues from the Sitas area into the Efjord region and that the Reppi Schist does not sit in the core of an isoclinal fold.

The Langvatn nappe thins rapidly to the north, pinching out between Skjomen and Rombakfjord. The Marko nappe also thins to the north, but probably persists north of Rombakfjord and continues to the northwest as far as the Skånland area studied by Steltenpohl (1987) where he recognized similar rocks as an unnamed allochthon immediately beneath the Evenes Group.

To the south, the Sitas MKNC is eroded from over the Akkajaure culmination. Where it reappears on the south side of the culmination, the MKNC broadly correlates with rocks from Kautsky's (1953) Vasten and Salo nappes. Melanges similar to those seen in the Langvatn nappe are present in both the Vasten and Salo nappes. However, the Langvatn nappe tectonostratigraphy is not recognized in the Vasten and Salo nappes, nor are the muscovite-rich garnet schists that dominate the Langvatn nappe abundant in either the Vasten or the Salo nappes (Crowley, unpublished mapping 1985). Foslie (1942) mapped a calcareous schist within Kautsky's Vasten nappe that he correlated with the Reppi Schist. This calcareous schist is a thick-bedded micaceous calc-phyllite that is unlike the thin-bedded Reppi Schist seen in the Sitas area both in terms of bulk composition and sedimentary structures. Therefore, I feel that Foslie's (1942) correlation is not justified and that little if any of the Langvatn nappe tectonostratigraphy continues to the south of the Akkajaure culmination.

Lithologies similar to those of the Marko nappe are relatively common within the Vasten and Salo nappes in the Stipok region studied by Sundblad (1986). The matrix schists and mafic dike-intruded schists of the Marko Complex are quite similar to schists in the Allak Complex (Stephens et al. 1984, Sundblad 1986) that crops out on the mountain Allak 40 km to the southwest. However, the Allak Complex contains two generations of mafic dikes, the later of which cuts across the regional foliation. Only the earliest phase of dikes, which pre-date the development of the regional foliation, is recognized in the Sitas region. The Hjertevatn gneiss is lithologically very similar to the Middle Köli quartz keratophyre mapped by Kulling (1982) and Sundblad (1986). This correlation is consistent with the interpretation

<table>
<tr><th></th><th>D_0</th><th>D_1</th><th>D_2</th><th>D_3</th><th>D_4</th><th>D_5</th></tr>
<tr><td>Marko nappe</td><td>Tectonic stacking (?)</td><td>Penetrative mica foliation
Amphibolite facies metamorphism
Isoclinal folding</td><td rowspan="2">ESE directed thrust faulting

Stretching lineation

Amphibolite facies metamorphism

Isoclinal folding</td><td rowspan="2">NW trending inclined cross folding</td><td rowspan="4">ESE directed thrust faulting

Isoclinal folding

Mylonitic and phyllonitic foliation

Stretching lineation

Greenschist facies metamorphism</td><td rowspan="4">NE trending NW vergent back folding</td></tr>
<tr><td>Langvatn nappe</td><td>Melange formation

Tectonic stacking</td><td>Penetrative mica foliation
Amphibolite facies metamorphism
Isoclinal folding</td></tr>
<tr><td>Storriten Complex</td><td colspan="4">Greenschist-facies metamorphism
Penetrative foliation
Isoclinal folding</td></tr>
<tr><td>Rombak Complex</td><td colspan="4">No structures formed</td></tr>
</table>

Fig. 5. Summary of Caledonian structural events in the Sitas region.

based on Rb/Sr isotopic data that the Hjertevatn gneiss protolith originating as a seawater altered igneous rock. A similar explanation was proposed by Stephens (1980) to explain chemical variations in Middle Köli volcanic rocks in the Stekenjokk region. The Bavro gneiss is lithologically similar to a garnet schist mapped by K. Sundblad (pers. comm. 1987) and included in undifferentiated schists of his Stipok terrane (Sundblad 1986). The remarkable similarities of individual lithologies in both the Sitas and Stipok sections argues strongly for correlation. However, these units are not in the same sequence in the two sections and many of the units of the Stipok terrane (Sundblad 1986) do not continue north into the Sitas region. Unexplained structural complications exist in the MKNC between the well studied Sitas and Stipok regions.

Structural evolution

The rocks in the Sitas region underwent a sequence of superposed Caledonian deformations (D_0 to D_5) that include at least 3 episodes of low- angle faulting and 5 phases of folding (Fig. 5). One and possibly two phases of deformation have affected the crystalline Rombak Complex in the Precambrian prior to the deposition of the Torneträsk Formation. All of the Caledonian deformations (D_0 to D_5) affect the rocks of the MKNC, while only the youngest episode of thrusting (D_4) and the two youngest phases of folding (D_4 & D_5) affect the Storriten and Rombak Complexes. A phase of deformation has affected the Storriten Complex prior to D_4; however, this deformation cannot be correlated with any of the events recognised in the MKNC. All of the deformations except D_1 produced map-scale structures, while all except D_0 produced regionally extensive petrofabrics.

Precambrian deformation

Away from the Frostisen thrust, the Rombak gneiss is weakly to strongly foliated with foliation-parallel mylonite zones developed locally.

This foliation has a very consistent orientation, dipping steeply to the northwest (Fig. 7a) and is at a high angle to the subhorizontal Frostisen thrust. The supracrustals in the Skåddåive pendant (Plate 1) are folded into tight to isoclinal, upright to inclined folds with axial surfaces and an axial planar cleavage that parallels the foliation in the Rombak Gneiss. This foliation is truncated by the basal unconformity of the Torneträsk formation indicating that it was formed prior to Caledonian deformation (Fig. 6).

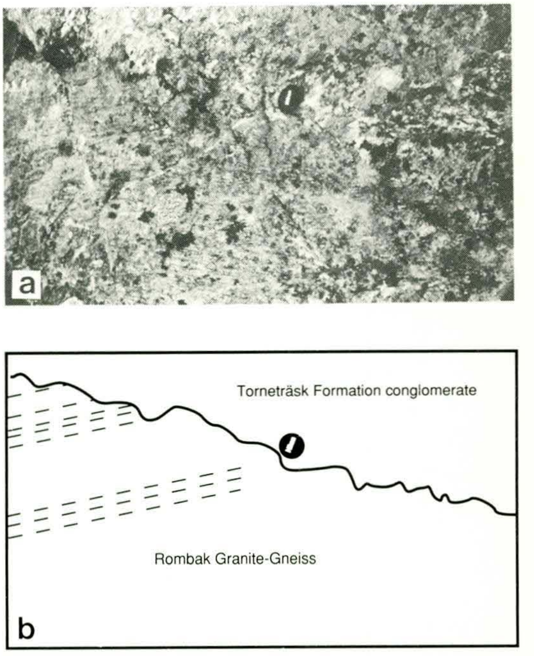

Fig. 6. Torneträsk Formation basal uncomformity truncating foliation in Rombak Granite-Gneiss).
a) Photograph
a) Line drawing from photograph. Dashed lines showing pre-Caledonian foliation.

D_0 - Melange formation

The first Caledonian deformation (termed here D_0) to affect the rocks in the Sitas area is seen only in the Langvatn nappe. It produced the zones of tectonic melange (the Raudvatn and Filfjell Complexes) found in the Langvatn nappe. During D_0, ultramafic, mafic and carbonate rocks, lithologies typical of the oceanic lithosphere, were chaotically intermixed with the quartz-rich clastic sedimentary rocks that now constitute the matrix schists. D_0 is a rather cryptic deformational phase because D_0 fabrics have been annealed by subsequent amphibolite-facies metamorphism and transposed by subsequent deformations. The only record that remains of D_0 is the tectonic stacking and slicing within the Raudvatn and Filfjell Complexes. Lithological contacts within the melange zones and between the melange zones and the surrounding schists are considered to be D_0 faults. Other contacts within the Langvatn nappe may also be D_0 faults. All lithological contacts within the Langvatn nappe, including those interpreted to be D_0 faults are everywhere parallel to the S_1 regional foliation and are locally folded by F_1 folds. This demonstrates that the D_0 tectonic mixing predated the development of D_1 petrofabrics.

D_1 - Isoclinal folding and development of the regional mica foliation

The first penetrative Caledonian fabrics were produced during D_1. The effects of D_1 are recognized in both the Langvatn and the Marko nappes. These include small scale (cm to m scale) isoclinal folds and a penetrative mica foliation (S_1). No map-scale structures were produced by D_1. S_1 is approximately parallel to lithological layering and was deformed by all subsequent deformations. The attitude of S_1 varies widely across the Sitas area (Fig. 7b). Changes in the attitude of S_1 were used to define structures produced by later deformations. Transposition of S_1 to parallel S_4 has produced a weak southwest-dipping maximum in the attitude of S_1 while F_5 folding has formed a very weak girdle about a southwest trending axis (Fig. 7b). D_1 appears to be roughly synchronous with the metamorphic peak of rocks in the MKNC as both S_1 micas and metamorphic porphyroblasts are deformed by all subsequent deformations.

Preliminary $^{39}Ar/^{40}Ar$ hornblende geochronology (Tilke 1986) and a detailed petrological study (Crowley & Spear 1987) suggest that the metamorphism of the MKNC in the Sitas area was diachronous, with the Marko nappe reaching its thermal peak as much as 50 Ma before the Langvatn nappe. This strongly suggests that different deformations were responsible for the development of penetrative synmetamorphic foliations (S_1) within the Langvatn and Marko nappes. However, due to the transposition of S_1 in subsequent deforma-

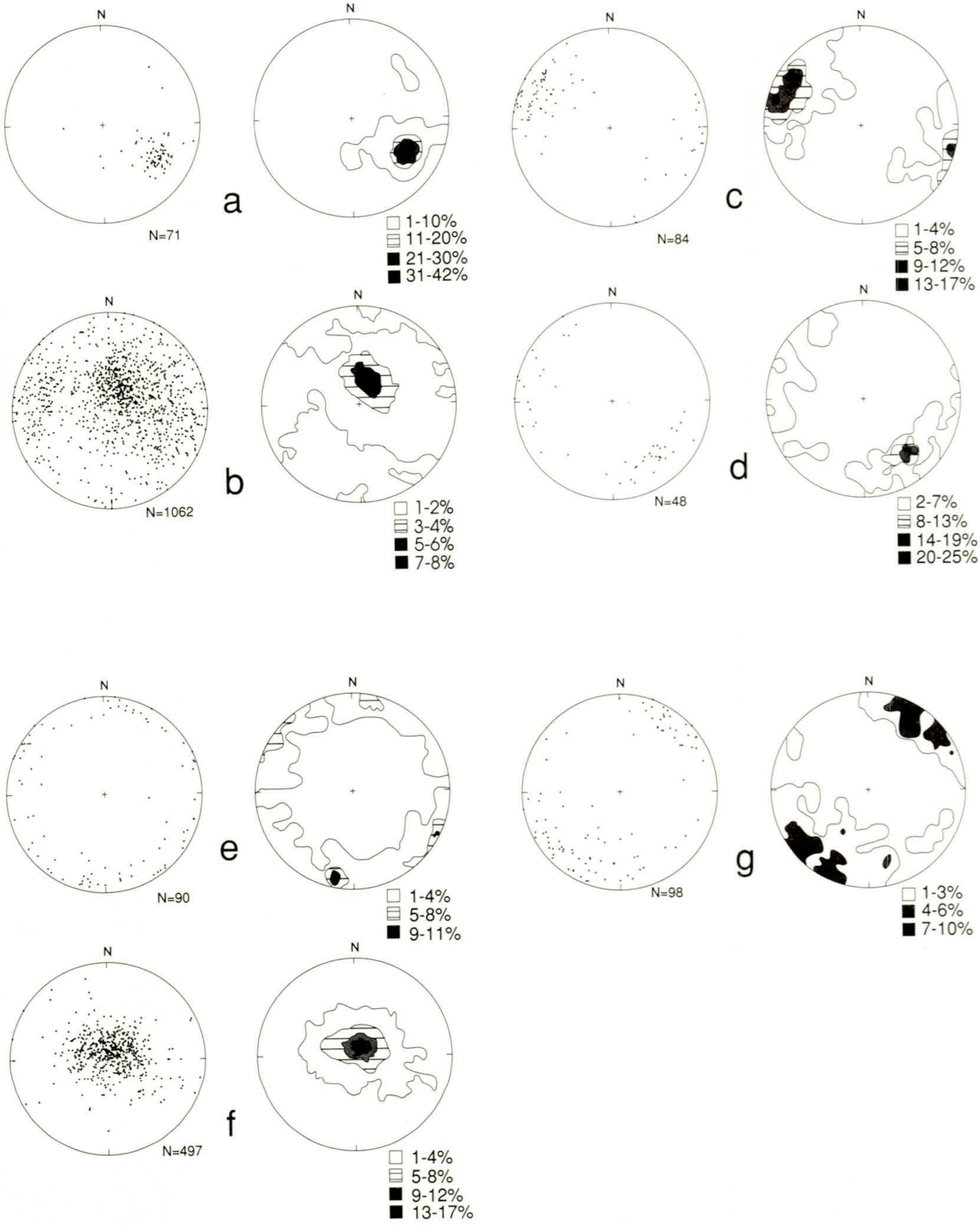

Fig. 7. Equal-area stereoplots (see text for discussion). Contours: % points/1% area.
a: Precambrian foliation in the Rombak Gneiss.
b: S_1 mica foliation in the MKNC.
c: L_2 lineations in the Marko nappe.
d: F_3 fold axes in the MKNC.
e: F_4 fold axes.
f: S_4 phyllonitic cleavage.
g: F_5 fold axes.

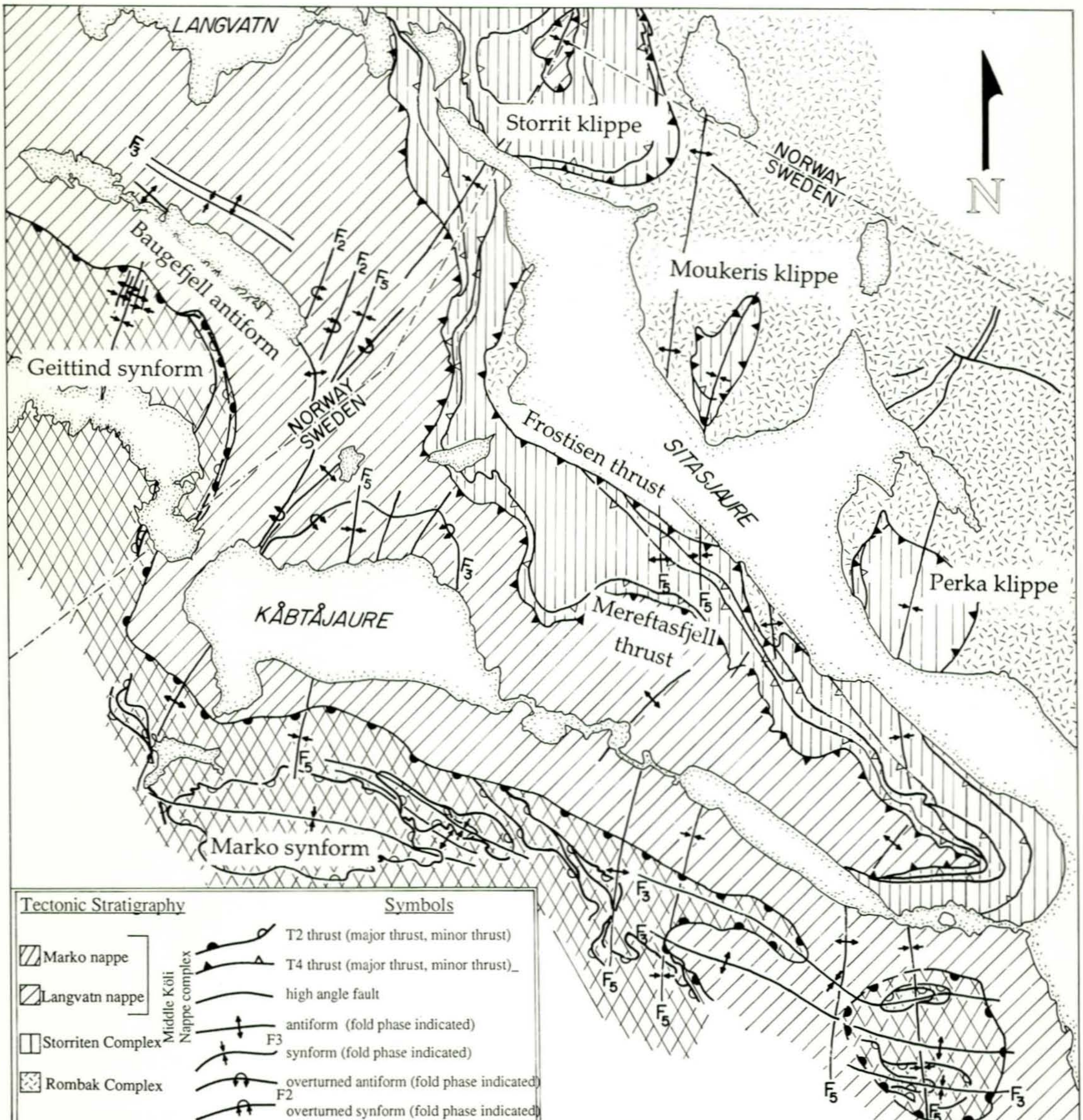

Fig. 8. Major Caledonian structures in the Sitas region.

tions, it was not possible to differentiate between the penetrative mica foliations of the two nappes so they have been grouped together here.

D_2 - Thrusting and isoclinal folding

The Marko nappe was thrust onto the Langvatn nappe during D_2. However, the two nappes responded differently to D_2 deformation. Within the Marko nappe, small-scale thrust faults developed at a low angle to lithological layering producing the lithological repetitions that characterize the Marko Complex (Fig. 9). Competent lithologies, particularly the migmatitic gneiss developed a mylonitic foliation near these thrusts. Less competent lithologies, particularly the calcareous schist were attenuated within a few tens of metres of the faults (Fig. 10a). Lithological layering (bed-

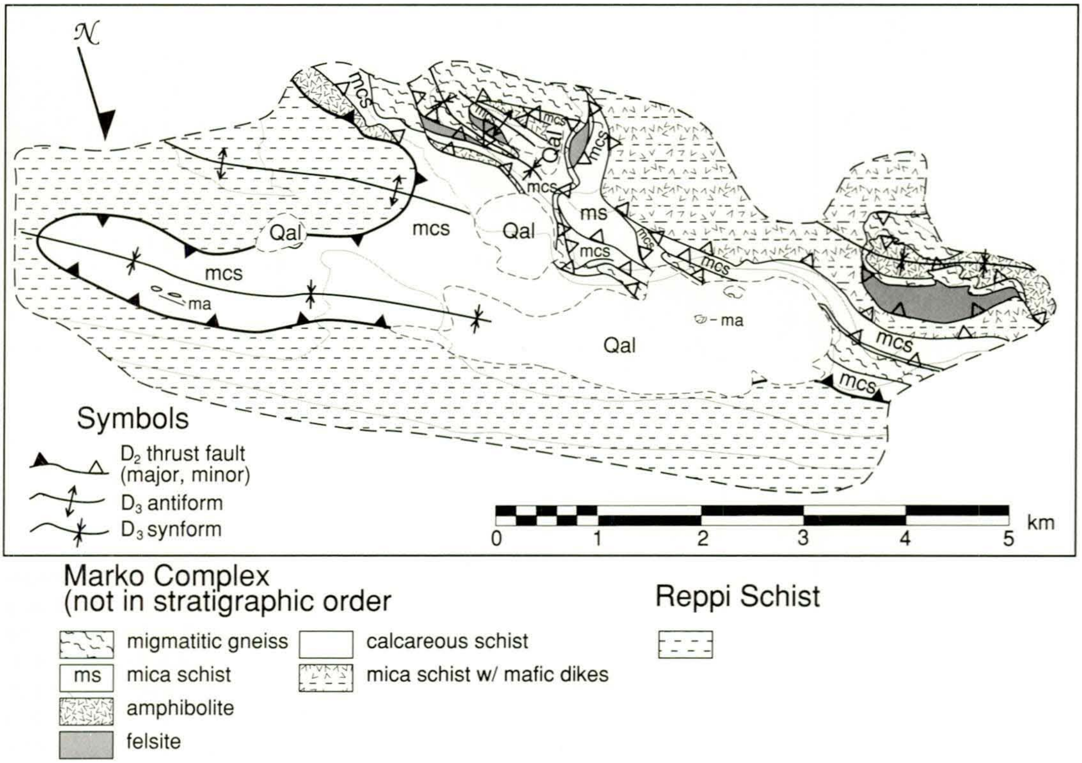

Fig. 9. Detailed geological map of the north slope of the mountain Marko showing D_2 lithological imbrications within the Marko Complex.
The faults bounding the imbricates all sole into a decollement at the top of the Reppi Schist.

ding?) in these units was greatly thinned near the thrusts.

The top of the Reppi Schist served as a decollement horizon for this thrusting. Many of the thrusts within the Marko nappe are listric, merging into the decollement at the top of the Reppi Schist (Fig. 9 & Plate 2, Section F- F'). There is no tectonic mixing between the Reppi Schist and any of the Marko nappe lithologies; therefore, none of these thrusts can continue down through the Reppi Schist. The thrust faults locally deform the S_1 foliation in the Marko Complex, but are themselves folded by F_3 folds. This geometry indicates that the Marko nappe was emplaced during D_2.

Small-scale D_2 thrusts have not been recognized in the Langvatn nappe. Instead, there are isoclinal folds that fold the S_1 mica foliation. These folds range in wavelength from centimetres to 100's of metres.

The direction of tectonic transport during D_2 is inferred to have been towards the ESE. D_2 transport is parallel to shallowly plunging, WNW and ESE trending L_2 mineral and stretching lineations (Fig. 7c). Small-scale thrusts within the Marko nappe all ramp upwards towards the ESE (Fig. 9), suggesting that the transport is ESE directed. This is consistent with the vergence direction deduced from the asymmetry of pegmatite boudins near the thrust (Fig. 10b).

The displacement on the thrust is estimated to be greater than 25 km on the basis of the structural overlap of the Marko nappe on the Langvatn nappe measured parallel to the ESE transport direction. The amount of shortening accommodated on the minor faults within the Marko nappe cannot be estimated without knowing the original stratigraphy. However, attempts have been made to balance and palinspastically restore cross-section F-F' (Plate 2) assuming different stratigraphies for the Marko Complex, each required there to be greater than 100% shortening within the Marko nappe.

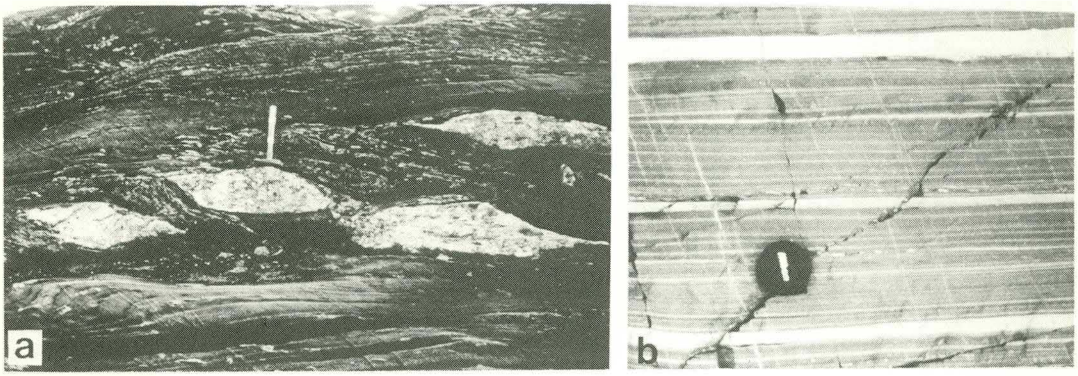

Fig. 10. D_2 structures in the Marko Complex.
a) Attenuated bedding in calcareous schist near a D_2 thrust fault.
b) Asymmetrical pegmatite boudins in calcareous schist indicating top to the ESE transport during D_2.

D_3 - NW-trending folds

D_3 produced open to tight upright to inclined, NW-SE trending folds (Fig. 7d). The scale of F_3 folds ranges from minor folds which are centimetres in wavelength to map-scale folds kilometres in wavelength (e.g. the Baugefjell antiform, Plate 2, section A-A'). F_3 folds deform the S_1 foliation, F_2 folds and the thrust separating the Marko and Langvatn nappes (Plate 1). F_3 folds tend to be NE vergent suggesting tectonic transport to the NE during D_3.

D_4 - Thrusting, imbrication and isoclinal folding

The present tectonic stacking of the Sitas area was formed during D_4 when the Frostisen and Mereftasfjell thrusts were active. D_4 thrusting placed the MKNC onto continental basement equivalent to the Rombak Complex, imbricating the basement and its depositional cover to form the Storriten Complex. D_4 is the first deformation to affect all of the structural domains in the Sitas area.

The Mereftasfjell thrust truncates lithological units within both the MKNC and the Storriten Complex while the Frostisen thrust truncates units within the Storriten and Rombak Complexes. D_4 thrusting produced isoclinal F_4 folds and a phyllonitic spaced cleavage (S_4) that are ubiquitous within the Storriten Complex schists. In addition, D_4 produced a mylonitic foliation in the granite and quartzite slices in the Storriten Complex. F_4 folds and an S_4 spaced cleavage are present in the MKNC in the hangingwall of the Mereftasfjell thrust; however, the folds and cleavage become less intense moving structurally upwards and die out a few hundred metres above the thrust. In the footwall of the Frostisen thrust, a zone of ultramylonite less than one metre thick formed in the Rombak Gneiss basement. Below this mylonitic zone, the basement is not penetratively deformed; the effects of D_4 are limited to the development of anastomosing shear zones that vary in width from less than 1 cm to 0.5 m wide. The spacing between these shear zones increases downward from the Frostisen thrust until they die out between 100 and 200 metres below the thrust.

The Storriten Complex is a D_4 duplex with the Frostisen thrust as its floor and the Mereftasfjell thrust as its roof. This duplex system appears to have transported the MKNC en masse as there is no tectonic mixing between the MKNC and the Storriten Complex. Faults within the Storriten Complex ramp consistently upwards towards the ESE (Fig. 8) suggesting ESE-directed transport. This is consistent with the orientation of stretching lineations in Storriten granitic and quartzitic mylonites. The orientations of F_4 fold axes within the Storriten Complex vary widely, but are generally shallowly plunging and roughly cluster around NE-SW and NW-SE trends (Fig. 7e). F_4 isoclinal folds probably formed with NE-SW-trending axes, at a high angle to the ESE transport direction and were subsequently rotated towards the transport direction. The wide scatter in fold orientations seen in Fig. 7e is likely due to varying degrees of transposition.

F_4 folds deform a pre-D_4 foliation in the Storriten schists. This foliation is cut and transposed by S_4. The pre-D_4 foliation is a synmetamorphic foliation that can be traced as an inclusion foliation in the cores of helicitic garnets and followed into the matrix foliation. This requires the Storriten Complex to have been deformed and metamorphosed prior to D_4. However, the style of deformation and the timing of deformation relative to porphyroblast growth is different from that of the pre-D_4 deformations recognized in the MKNC and therefore cannot be correlated with any of the deformational events that affected the MKNC.

Many of the basement imbricates within the Storriten Complex are bounded by faults that ramp from a sole thrust, the Frostisen thrust, up to a roof thrust, the Mereftasfjell thrust (Fig. 8). Therefore, the displacement on the Frostisen-Mereftasfjell thrust system must be equal to or greater than the total displacement on all of these smaller faults. By summing minimum displacements on each imbricate within the Sitas area, the displacement on the Frostisen-Mereftasfjell thrust system is estimated to be greater than 58 km. This is greater than the minimum displacement estimated on the basis of the structural overlap of the MKNC on the Rombak Complex.

Within the Sitas area, D_4 thrust faulting appears to post-date NW-trending F_3 folding. The limbs of the F_3 Baugefjell antiform are transposed into sub- parallelism with the Mereftasfjell thrust (Plate 2, section B-B'). S_4 dips gently to the SW across the entire region and does not appear to have been folded by NW-trending folds (Fig. 7f). However, on a regional scale, the Sitas area appears to sit on the NE limb of an open NW-trending synform that folds both the Frostisen and the Mereftasfjell thrusts (Fig. 1) . This suggests the existence of a post-D_4 phase of NW-trending folds that is not well represented within the Sitas area.

D_5 - Upright NE-trending folds

D_5 produced small- to large-scale upright folds with shallowly-plunging NE-SW-trending axes (Fig. 7g) and steep SE-dipping axial surfaces. F_5 folds deform all pre-existing structures and form the prominent map-scale NE-trending antiforms and synforms (Plate 2, sections D-D" and E-E'). F_5 folds tend to be concentric and open folds. A spaced cleavage commonly formed in the hinges of F_5 folds.

Regional metamorphism

Mineral assemblages within the Sitas area appear to reflect one major phase of prograde Caledonian metamorphism. However, both petrological (Crowley & Spear 1987) and preliminary geochronological (Tilke 1986, pers. comm. 1986) evidence indicate that this metamorphism is diachronous and may extend over a period of approximately 50 Ma. Movement on the post-metamorphic Frostisen and Mereftasfjell thrusts has inverted the metamorphic sequence so that the highest grade rocks are now found at the highest structural levels. The Rombak Gneiss and outliers of the Torneträsk Formation exposed on the southern side of the Rombak window have been metamorphosed to middle greenschist facies (biotite grade). Structurally upwards in the Storriten Complex, the metamorphic grade increases to the upper greenschist facies. The mineral assemblage garnet + biotite ± chlorite + muscovite + quartz + plagioclase is stable. Upwards into the MKNC, the metamorphic grade increases into the amphibolite facies. Pelitic and psammitic rocks generally contain the mineral assemblage garnet + biotite + muscovite + quartz + plagioclase ± hornblende and the mineral assemblage hornblende + plagioclase + epidote + sphene is common in mafic rocks. The grade of metamorphism also increases going structurally upwards from the Langvatn to the Marko nappe. The pelitic mineral assemblage garnet + biotite + muscovite + kyanite + plagioclase is present in some Marko nappe schists. In addition, quartzo-feldspathic enclaves interpreted to be in-situ partial melts are present in Marko nappe rocks.

The P-T conditions of metamorphism within the Storriten Complex and MKNC have been estimated by garnet-biotite geothermometry (Ferry & Spear 1978) and garnet-biotite-plagioclase-muscovite geobarometry (Hodges & Crowley 1985). In calculating P-T, all phases except biotite were considered to be nonideal solutions using solution models described by Hodges & Crowley (1985). A more thorough description of the metamorphic conditions is given by Crowley & Spear (1987).

Rocks from the Storriten Complex equilibrated at approximately 450° C and 600 MPa (Fig. 11). The scatter in calculated P-T does not exceed estimated analytical error (± 20°C; ± 50 MPa). The calculated P-T of equilibration of the MKNC scatters about 575°C and 800 MPa. The scatter in calculated P-T exceeds analytical error, particularly for Marko nappe samples. The calculated temperature for the Marko nappe is significantly lower than that at which: (1) the staurolite-out assemblage garnet + biotite + kyanite is stable (Piggage & Greenwood 1982) and (2) migmatitic partial melts appear (Thompson 1982). Inasmuch as both the staurolite-out assemblage and migmatitic enclaves are present in pelitic rocks in the Marko nappe this constrains the peak P-T conditions of the Marko nappe to be 625-775°C at a pressure greater than 700 MPa (Crowley & Spear 1987). There is no evidence that rocks from the Langvatn nappe ever experienced P-T conditions in excess of those calculated by thermobarometry and therefore did not undergo this pre-D_2 high-temperature heating. The P-T calculated for the MKNC by thermobarometry (575°C and 800 MPa) is interpreted to reflect the conditions at which the two MKNC nappes were juxtaposed during D_2.

Rocks from both the MKNC and Storriten Complexes had cooled to middle-greenschist facies conditions by D_4 time, when the Frostisen and Mereftasfjell thrusts were active. Throughout the Storriten Complex and in the MKNC near the Mereftasfjell thrust, garnets have been partially to completely chloritized and a greenschist facies S_4 muscovite + biotite ± chlorite cleavage developed.

During D_4, the Rombak Complex reached its middle greenschist facies metamorphic peak, with the development of a spaced muscovite + biotite ± chlorite ± epidote cleavage.

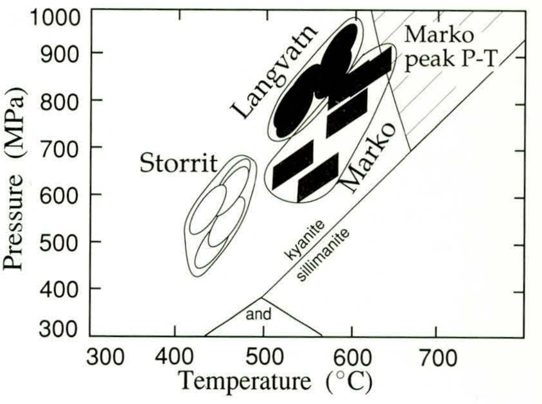

Fig. 11. P-T of metamorphism of the Storriten Complex and MKNC determined by garnet-biotite-muscovite-plagioclase thermobarometry. The symbol size approximates the analytical uncertainty in the P-T determinations. The peak P-T of the Marko nappe is estimated to be higher than that determined by thermobarometry on the basis of the stability of the assemblage garnet + biotite+ kyanite + muscovite +quartz. (After Crowley & Spear 1987, Crowley 1988)

Geochemistry of mafic rocks from the Marko nappe

The Marko Complex contains a diverse range of mafic rocks that include: fine-grained pillowed amphibolites, fine- to coarse-grained amphibolite dikes, coarse-grained amphibolite and a layered olivine-pyroxene gabbro. In general, the mafic rocks crop out as fault-bounded slices; thus, it is difficult to use field relations to determine the genetic relationships between different mafic bodies. In order to determine, (1) whether all of the mafic rocks in the Marko Complex were derived from either a single or from multiple volcanic/plutonic sources, and (2) the tectonic nature of these sources, samples from the Marko gabbro, a mafic dike that intrudes the gabbro, and fine-grained pillowed amphibolite have been analyzed for major and trace elements (Table 2 and Figs. 12, 13). Major elements were analyzed by XRF at the University of Massachusetts, Amherst. Trace elements were analyzed by XRF at the University of Massachusetts (Sr, Zr, Ba, Ni, Cr, Co, Sc, Zn, V, Nb) and by INAA (REE, Cs, Rb, Th, U, Ta, Hf) at the Massachusetts Institute of Technology. The data presented here represent too small a data set from which to draw conclusions about the origin of the Marko Complex; however, it is sufficiently high-quality data to conclude that all of the mafic rocks cannot be cogenetic.

All of the gabbro samples are relatively low in SiO_2, high in MgO, and have low incompatible and high compatible trace element concentrations. In general, there are good positive linear correlations between incompatible trace elements. The dike sample tends to plot near the extension of the correlation lines at higher incompatible element concentrations. Plots of incompatible elements against compatible ele-

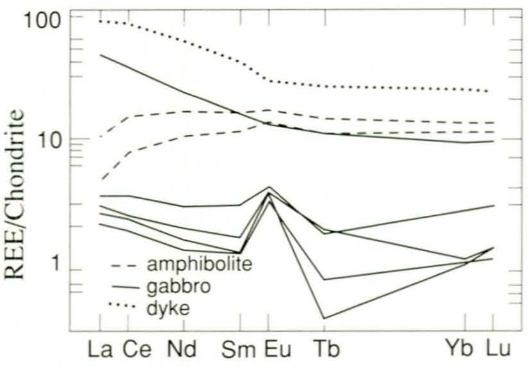

Fig. 12.Chondrite-normalized REE plot. The Marko gabbro and its feeder dike have light REE-enriched patterns while the pillowed amphibolites have light REE-depleted patterns.

Table 2: Major and trace element analyses of mafic rocks in the Marko Complex.

	MT-3 amph.	MT-6 dike	MT-11 evolved gabbro	MT-16 gabbro	MT-17 gabbro	MT-18 gabbro	MT-22 gabbro	MT-23 amph.
				Major Elements (wt%)				
SiO_2	45.90	55.86	56.03	47.41	42.59	45.15	50.35	43.47
TiO_2	1.01	2.39	0.62	0.13	0.11	0.12	1.32	0.31
Al_2O_3	20.11	13.97	14.70	20.77	13.61	14.72	15.98	7.33
Fe_2O_3*	11.22	11.86	6.49	5.72	9.22	11.28	9.51	12.68
MnO	0.25	0.19	0.13	0.09	0.15	0.18	0.17	0.20
MgO	4.27	3.01	7.09	12.80	20.92	19.57	7.79	28.99
CaO	14.35	6.36	10.33	11.35	7.81	7.31	11.85	5.21
Na_2O	2.51	3.49	2.23	1.87	1.40	1.58	1.99	0.86
K_2O	0.21	1.49	0.54	0.08	0.07	0.12	0.20	0.09
P_2O_5	0.07	0.27	0.06	0.01	0.01	0.01	0.11	0.01
Total	99.97	99.05	98.27	100.24	95.90	100.05	99.36	99.19
				Trace elements (ppm)				
Cs	–	0.8	1.0	–	–	–	–	–
Rb	–	49.0	29.0	–	–	–	–	–
Ba	1.5	279.1	200.0	29.2	10.4	20.0	7.5	23.3
Th	–	7.4	3.1	–	0.1	–	0.2	–
U	–	2.0	0.9	–	–	–	–	–
Ta	–	0.8	0.3	–	–	–	0.2	–
Nb	0.8	11.2	4.5	–	1.0	0.6	1.0	0.3
Hf	1.6	6.8	2.9	0.1	0.1	0.1	2.4	0.3
Zr	55.5	270.3	101.3	4.2	3.7	4.8	95.0	10.0
Sr	164.7	177.4	204.4	226.2	175.4	245.9	151.4	99.6
La	1.40	25.74	13.66	0.89	0.69	0.81	3.26	1.09
Ce	6.55	66.40	29.10	1.89	1.59	1.85	12.30	2.86
Nd	6.37	35.20	13.90	1.19	0.85	0.97	9.70	1.81
Sm	2.28	7.86	3.22	0.33	0.26	0.26	3.16	0.60
Eu	1.03	2.14	0.98	0.35	0.25	0.29	1.21	0.31
Tb	0.68	1.22	0.53	0.10	0.04	0.02	0.68	0.09
Yb	2.40	5.11	2.06	0.24	0.23	0.22	2.79	0.59
Lu	0.37	0.74	0.31	0.05	0.04	0.05	0.42	0.10
Ni	185	10	41	278	579	325	63	1108
Cr	272	3	199	184	256	86	326	1444
Co	69	29	30	51	82	94	34	–
Sc	44	30	34	10	10	8	39	–
V	225	269	130	30	26	23	227	77
Zn	97	126	59	38	59	73	73	93

ments do not produce linear correlations, but rather are concave upwards and asymptotic to both the axes. Again the dike sample plots along an extension of the graph at higher incompatible element concentration. All of the gabbro samples have low REE contents, light REE-enriched patterns and positive Eu anomalies (Fig. 12). The dike sample has much higher REE concentrations (20-100 times chondrites) and a light REE-enriched pattern that parallels the pattern of the gabbro, except that it lacks a Eu anomaly.

The mineralogical layering, plagioclase foliation, low incompatible element concentrations, and positive Eu anomalies all suggest that the gabbro formed by crystal accumulation. This means that the analyzed samples do not represent compositions that were once liquids. However, the dike that cuts the gabbro may be cogenetic with the gabbro and represent a liquid similar to the one from which the gabbro crystallized. Numerical simulation of fractional crystallization (Crowley 1985) using trace element partition coefficients (Drake & Weill 1975, Frey et al. 1978) supports this contention and indicates that the gabbro crystallized from a light REE-enriched magma with incompatible element abundances greater than 10 times chondrites that was chemically similar to the dike.

The fine-grained pillowed amphibolite samples have distinctly different chemistries from the gabbro. They tend to have lower compatible and higher incompatible element concentrations. The REE patterns are light REE-depleted and have no Eu anomalies. The light REE depleted pillowed amphibolites cannot be related to the gabbro by either fractional crystallization or partial melting. The light REE-enriched gabbro and light REE-depleted amphibolites must have come from geochemically distinct sources.

The tectonic setting of these sources can be determined by comparison with modern mantle-derived mafic rocks of known origin (Fig. 13). On a 'spider' diagram in which incompatible trace elements are arranged from left to right in order of increasing compatibility in mafic magmas (Wood et al. 1979, Tarney et al. 1980) MORB, OIB and arc basalts have distinctly different trace-element patterns. MORB has a relatively smooth pattern that increases in concentration from left to right, OIB tends to first increase and then decrease, and arcs tend to have an overall negative

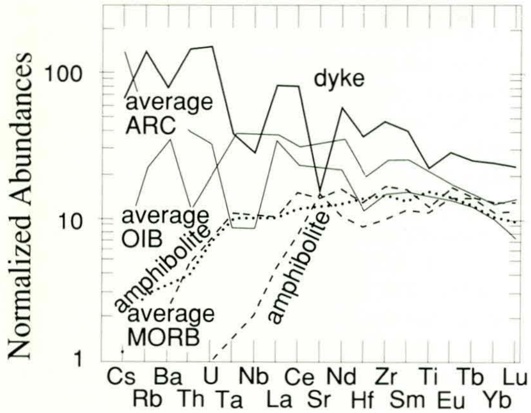

Fig. 13.Chondrite-normalized incompatible trace element patterns for mafic liquids in the Marko complex. Average MORB, OIB and arc shown for comparison. The Marko gabbro has the large negative Ta & Nb anomaly that characterizes arc rocks while the pillowed amphibolites have patterns similar to MORB. Average MORB from Wood et al. (1979, 1980). Average OIB from BSVP (1981) and Leeman et al. (1980) except Cs, Rb, Ba, and La which are calculated using average K concentrations and average element ratios in Morris & Hart (1983). Average arcs from BSVP except Ta which was calculated by comparison with BSVP Nb data.

slope with a large negative Ta (Nb) and smaller negative Hf anomaly. The Marko gabbro has an overall negative slope and a large negative Ta (and Nb) anomaly and a smaller negative Hf anomaly suggesting derivation from an arc-like source. The pillowed amphibolites have smooth patterns with overall positive slopes suggesting derivation from a MORB-like source.

Discussion

Tectonic Setting of the MKNC

The MKNC is composed of a sequence of clastic metasedimentary units that have been stripped from their crystalline basement. The basement upon which their protoliths were deposited may have been either sialic or oceanic, but is interpreted to have been in a fore-arc position in proximity to both a continental mass and an active subduction margin. The Langvatn nappe contains distinct zones of tectonic melange (Raudvatn and Filfjell Complexes) with lenses of mafic and ultramafic rocks in matrix of quartz-rich psammite and schist. The composition of the lenses (i.e. mafic and ultramafic rocks) indicates derivation from an oceanic rather than continental source. Although the lenses are small, lithologies representative of much of the oceanic crustal and upper mantle section are present. The high quartz and mica content of the matrix schists and psammites indicates a high quartz and feldspar content of the matrix sediments prior to metamorphism. This suggests that the sediments were derived from a continental source and therefore proximity to a continental mass during deposition. Cherty rocks, characteristic of pelagic environments are not present in matrix of the melanges.

The tectonic juxtaposition of continentally derived clastic rocks and oceanic crystalline rocks similar to that seen in the Langvatn nappe occurs in modern fore-arc accretionary complexes such as those found on Nias (Moore & Karig 1980), Barbados (Speed & Larue 1982), or Taiwan (Page & Suppe 1981). The tectonic lenses in modern fore-arc complexes range in size from less than one metre to ophiolite slabs greater than tens of kilometres long, but lenses approximately the size of those in the Langvatn nappe (m to 0.5 km) are common.

The stacking of the Langvatn nappe lithologies and formation of the Langvatn melanges is interpreted to have occurred in a fore-arc position during the earliest deformation (D_0) that affected rocks in the Sitas region. The relationship between the plate margin responsible for this deformation and the paleomargin of Baltoscandia is unclear. There is no direct evidence that the margin of Baltoscandia was involved in this deformation. However, since melanges similar to those in the Langvatn nappe crop out over a wide region this boundary may be laterally extensive and mark an important zone along which oceanic material was consumed in the Caledonides.

The Marko nappe contains quartz and feldspar-rich metaclastic rocks that were derived from a continental source, and mafic igneous rocks with geochemical affinities to both MORB and arcs. The diversity of tectonic environments represented by these rocks cannot be easily understood in terms of modern tectonic environments. This structural complexity is here interpreted to represent the juxtaposition of rocks that formed in a variety of tectonic environments. It is suggested that this juxtaposition also occurred in a fore-arc environment. Therefore, both the Langvatn and the Marko nappes contain fore-arc rocks. The two nappes may have been derived from the same fore-arc; however, the metamorphic petrology of rocks from the two nappes indicates a separate thermal evolution for each nappe until they were juxtaposed by D_2 thrusting. Therefore, we cannot tie the two nappes together until after D_2 thrusting.

Structural Evolution During Terrane Accretion

The deformation of the continental basement in the Sitas region occurred at a mid-crustal depth (18-25 km) as a response to the accretion of one or more exotic terranes during the closure of the Iapetus ocean basin. Although the lithologies involved are different, the style of deformation seen in the Sitas area is quite similar to that seen in other orogens at much shallower levels (e.g. Bally et al. 1966, Boyer & Elliott 1982). Most of the large-scale deformation in the Sitas area is accommodated by movement along large-scale low-angle faults with folds and penetrative ductile deformation forming second-order, albeit spectacular structures.

At shallower levels, accretion-related deformation is commonly thin-skinned, with a decollement forming in the cover above a crystalline basement (e.g. Bally et al 1966, Elliott & Johnson 1980). Workers in the Scandinavian Caledonides have argued over whether Caledonian deformation was thin-skinned (e.g. Gee 1975, Gee et al. 1978, Gee & Zachrisson 1979) or basement involved (e.g. Gee 1975, Hodges et al. 1982, Tilke 1986, Gee 1986). Within the Sitas area, D_4 deformation was basement involved. However, this deformation had a thin-skinned-like geometry due to the presence of a decollement at a shallow level in the basement.

The basement slices in the Storriten complex are thin (1 m to 0.5 km) and are commonly in contact with or spatially associated with quartzites that are the depositional cover of the basement gneiss. In addition, within the Rombak Complex, the basal part of the Torneträsk Formation is locally preserved in depositional contact on the Rombak Gneiss. Together, these two observations suggest that the basement slices in the Storriten Complex detached from a relatively shallow level (< 0.5 km) below the basement-cover contact. This implies that (at least for 20 km across strike) a mechanical contrast existed at a shallow level in the basement that acted as the mechanical equivalent of a decollement horizon in a sedimentary sequence. The source of this mechanical contrast is enigmatic, but may have been controlled by the availability of downward-propagating fluids from the dehydrating sedimentary cover (Bartley 1982).

There is little Caledonian deformation within the basement below the Frostisen thrust. Below the thrust, there is a thin (<1m thick) mylonite zone and for another 100-200 m there are thin anastomosing shear zones. Therefore, beneath the 'basement decollement', at least within the Sitas region, the basement acted as a relatively rigid block. The structures that formed during the accretion of the MKNC are basement involved, but are geometrically and kinematically similar to structures formed at shallower levels by thin-skinned deformation.

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Geological Map of the Sitas Region

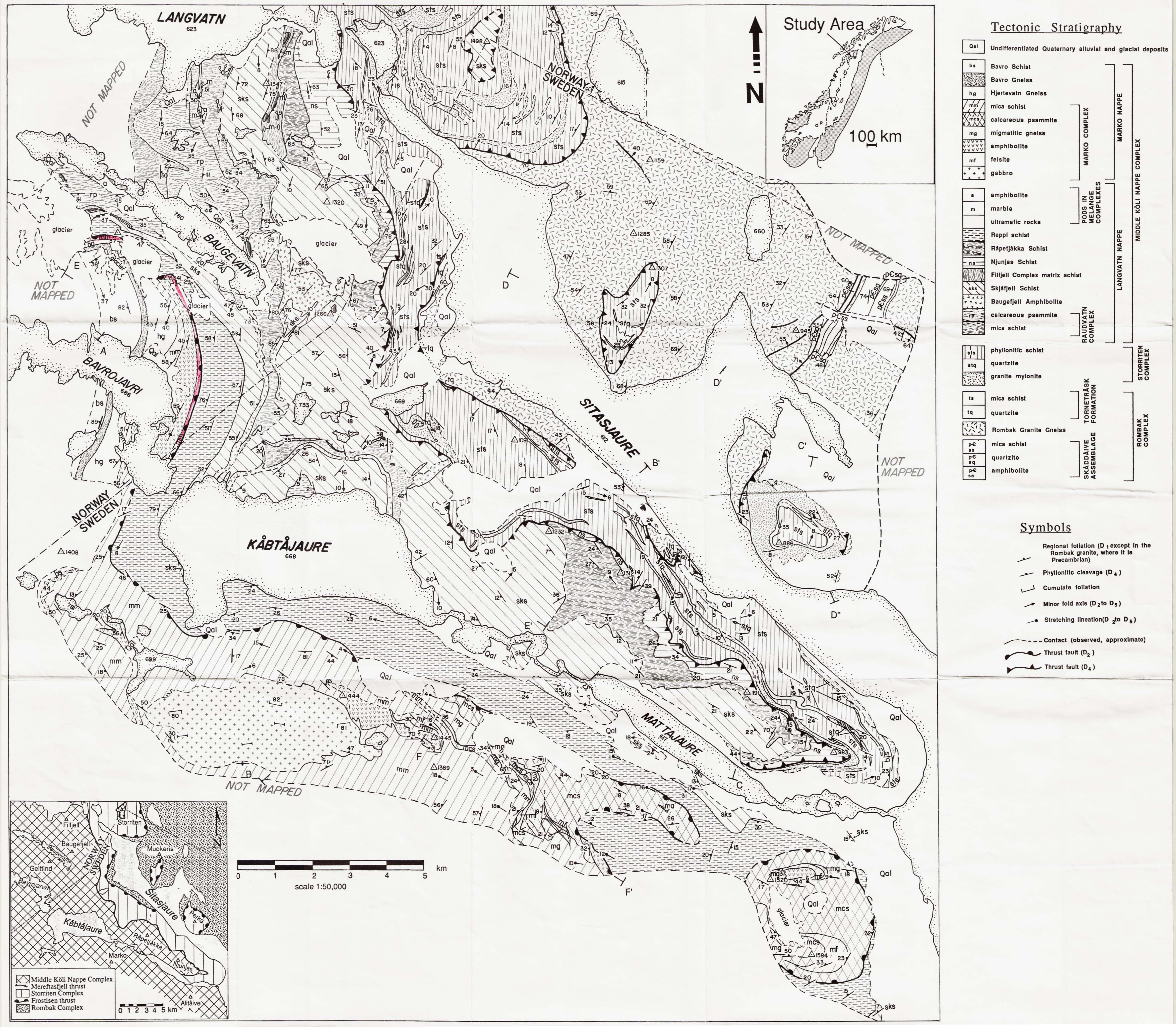

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Plate 2: Geological Cross Sections of the Sitas Region

symbols as in Plate 1