SVERIGES GEOLOGISKA UNDERSÖKNING

SERIE C NR 726 AVHANDLINGAR OCH UPPSATSER ÅRSBOK 71 NR 1

MICHAEL B. STEPHENS

STRATIGRAPHY AND RELATIONSHIP BETWEEN FOLDING, METAMORPHISM AND THRUSTING IN THE TÄRNA–BJÖRKVATTNET AREA, NORTHERN SWEDISH CALEDONIDES

WITH 4 PLATES

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ABSTRACT

The stratigraphy, structure and metamorphic history of the area between Stor—Björkvattnet and Tärnaby in the northern Swedish Caledonides are described. Pre-Cambrian igneous rocks and overlying sediments belonging to the Bångfjället Complex are surrounded by the variably metamorphosed sediments and igneous rocks of the Seve-Köli Nappe Complex. The rocks in the Tärna—Björkvattnet area have been divided into four litho-tectonic units:

1. The Bångfjället Complex which consists mainly of Pre-Cambrian quartz syenite unconformably overlain by a thin sequence of arkosic breccia, quartzite and pelite; these rocks are intensely mylonitized at the margins of the complex.
2. The Björkvattnet Unit which occupies the major part of the area and consists of Lower Ordovician(?) to post-Upper Llandovery, low-grade, metasediments and metavolcanites (Köli) on top of garnetiferous schists and amphibolites (Seve).
3. The Lower Laxfjället Unit which consists of gabbro-intruded calcareous phyllites and layered metavolcanites.
4. The Upper Laxfjället Unit which is composed predominantly of aluminous schists and has been correlated with the northern part of the Storfjäll Nappe.

The boundaries between these units are major thrusts and are associated with mylonites.

Multiple deformation phases appear to have affected the Björkvattnet Unit. D1, which has only been recognised in the low-grade rocks, gave rise to a major synclinal repetition of the stratigraphy. The main strain-determining episode occurred during D2 and is a penetrative deformation throughout the unit accounting for the formation of the regional foliation (S2), the various mineral lineations and the prolate-shaped conglomerate pebbles. The D2 phase of deformation contributed to the development of early minor folds which contain S2 as an axial surface structure. The D3 structures are dominated by close to tight, overturned, major and minor folds which trend approximately parallel to the strike of the S2 foliation and the major thrusts, and are associated with the first phase crenulation cleavage (S3) in the low-grade rocks. The D3 structures are particularly well developed adjacent to the major thrusts where the S3 foliation transposes S2. In the high-grade (Seve) rocks the F3 major and minor folds occur without the development of any conspicuous axial surface structure. Significant mylonitization and thrusting are thought to have occurred during D3. The final deformation phase (D4) controls the saddle-shaped outcrop pattern and folds all earlier structures. The major F4 fold is a northerly plunging synform — the Tärna Synform. The later stages of D4 are marked by the development of conjugate kink folds on the steep eastern limb of the Tärna Synform.

The evolution of cleavage development and the folding mechanisms of the F2-F4 structures, including the late-stage kinks, are discussed. The shape and orientation of the early minor folds parallel to the maximum elongation direction in S2 are considered to be controlled essentially by the progressive nature of the D1 and D2 phases and the formation of the regional foliation at a relatively late stage in the D1-D2 deformation history. The deformation sequence is discussed in terms of a model involving progressive strain during emplacement of the allochthonous Björkvattnet Unit on top of the Baltoscandian platform. Critical stages in this model include: 1. the establishment of a tectonic pile during D1 and early D2; 2. collapse of this pile during late D2 leading to penetrative fabric development; 3. substantial lateral transport at the end

of D2 and during D3 associated with the development of mylonite zones and major thrusts; 4. gravity-controlled disturbance of basement and cover due to the superimposition of higher density cover on top of lower density, "granitic" basement (D4).

The metamorphic peak in the Björkvattnet Unit appears to have been established in a relatively static interval after construction of the tectonic pile (D1-early D2) but prior to penetrative fabric development (late D2) related to collapse of that pile. In the higher grade parts of the unit relatively high temperatures and pressures persisted throughout and after D2. The different types of post-S2 structures developed in the low-grade (Köli) and high-grade (Seve) parts of the Björkvattnet Unit, and the minor, post-S2, retrogression phenomena in the transition zone between them are explained in terms of their differing physical properties after D2. This is related to the divergent metamorphic histories of the Köli and Seve parts of the unit, and the important addition of water to the higher level Köli rocks from metamorphic reactions prior to and during late D2.

Structural and metamorphic events in the higher litho-tectonic units on Laxfjället have been tentatively correlated with the deformation sequence in the Björkvattnet Unit. There is no evidence for F1 in the higher litho-tectonic units, whilst folds of type F3 are not present in the Upper Laxfjället Unit. However, as in the Björkvattnet Unit, the peak of metamorphism was established prior to formation of the early, penetrative foliation ($S1_{LLU}$ or $S1_{ULU}$).

The mylonitization of the quartz syenite at the margins of the Bångfjället Complex is associated with thrusting and tight, overturned folding with southerly sense of overturn. The mylonitic structures around the margins of the Bångfjället Complex and in the surrounding metasediments of the Seve-Köli Nappe Complex are described. It appears that the Bångfjället Complex is exposed in the core of a major F3 (Björkvattnet Unit) antiform which during progressive deformation developed mylonite zones and major thrusts parallel to its limbs. The tectonic slicing of the Bångfjället Complex within the Seve-Köli Nappe Complex is related to movement of the Upper Laxfjället Unit over the underlying Seve-Köli sequence either at the end of D2 or during early D3. The initial response to this movement appears to have been the major and minor, overturned folding involving both the allochthonous metasediments and the underlying basement rocks of the Baltic Shield. This was followed by the development of mylonite zones and thrusting parallel to the limbs of the major antiformal structure. Folding appears locally to have continued after thrusting and mylonite formation. All this deformation was completed before the development of the late, northerly plunging folds (F4). It is suggested that the Pre-Cambrian rocks of the Bångfjället Complex are rooted locally beneath the metamorphic allochthon.

GENERAL INTRODUCTION

LOCATION AND TECTONIC SETTING

The Tärna—Björkvattnet area is situated in the county of Västerbotten in northern Sweden, between 65°30′N and 65°45′N latitude and 15° and 15°30′E longitude. The area under consideration occupies the northern and central portions of mapsheet 24F Tärna. A detailed geological map at the scale of 1:50,000 is presented (Plate I).

The work of Törnebohm (1888, 1896) in the Caledonides of the county of Jämtland first introduced the concept of extensive overthrusting within the Scandinavian Caledonides, with superposition of older rocks of variable metamorphic constitution from the west on younger, fossiliferous, generally unmetamorphosed sediments in the east. More recently, attempts have been made to divide the complex rock pile of allochthonous and parautochthonous sheets into discrete tectonic units recognisable along the length of the deformed belt (e.g. Kulling, *in* Gavelin and Kulling 1955; *in* Magnusson et al. 1960; *in* Strand and Kulling 1972, p. 159; Zachrisson 1969, 1973). Present ideas on the spatial arrangement of these units in Västerbotten and the position of the Tärna—Björkvattnet area within the Seve-Köli Nappe Complex, as defined by Zachrisson (1973), are indicated in Fig. 1.

PREVIOUS WORK

The first general survey of the Västerbotten mountains was carried out between 1918 and 1928 and this led to the presentation of a 1:200,000 geological bedrock map (Backlund and Quensel 1929); a description of this map was published later by Quensel (1960). Kulling, who in 1925 discovered an important graptolite fauna and also a fossiliferous limestone horizon near Lake Broken, 27 km SSW from Tärnaby, later published a 1:75,000 map over the Björkvattnet—Virisen area and established a detailed stratigraphy in these low-grade rocks (Kulling 1933). Kulling included in his description the rocks exposed north and northeast of Ängesdal (Fig. 2) and correlated the stratigraphy here with that established in the fossiliferous strata to the south.

Kieft (1952) mapped the area north and northeast of Tärnaby as far as Bångfjället, approximately 15 km northeast of the village (Fig. 2). He described in detail the petrology of the basement rocks and the staurolite schists and amphibolites which lie tectonically above the basement rocks on Laxfjället, just north of Tärnaby. The results of similar petrological investigations in other areas near

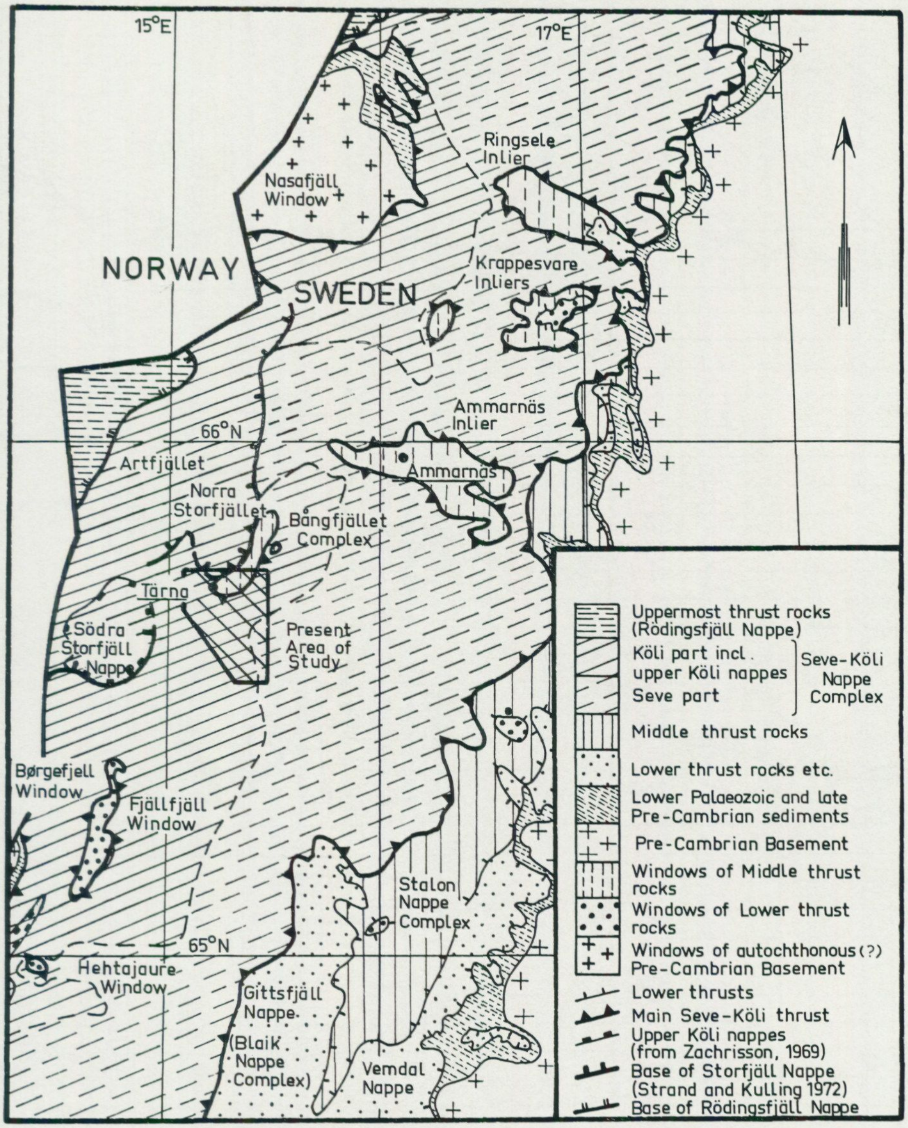


Fig. 1. Regional geological setting of the Tärna – Björkvattnet area (data from the 1:1,000,000 map of Sweden, Kulling, *in* Strand and Kulling 1972 and Zachrisson 1969, 1973).

Tärnaby have been presented by Faddegon (1940) and Murris (1957) (see Fig. 2).

Recent general accounts of the Caledonian geology of Västerbotten include the descriptions accompanying both the 1:400,000 county map of Västerbotten (Kulling, *in* Gavelin and Kulling 1955) and the 1:1,000,000 map of the Pre-

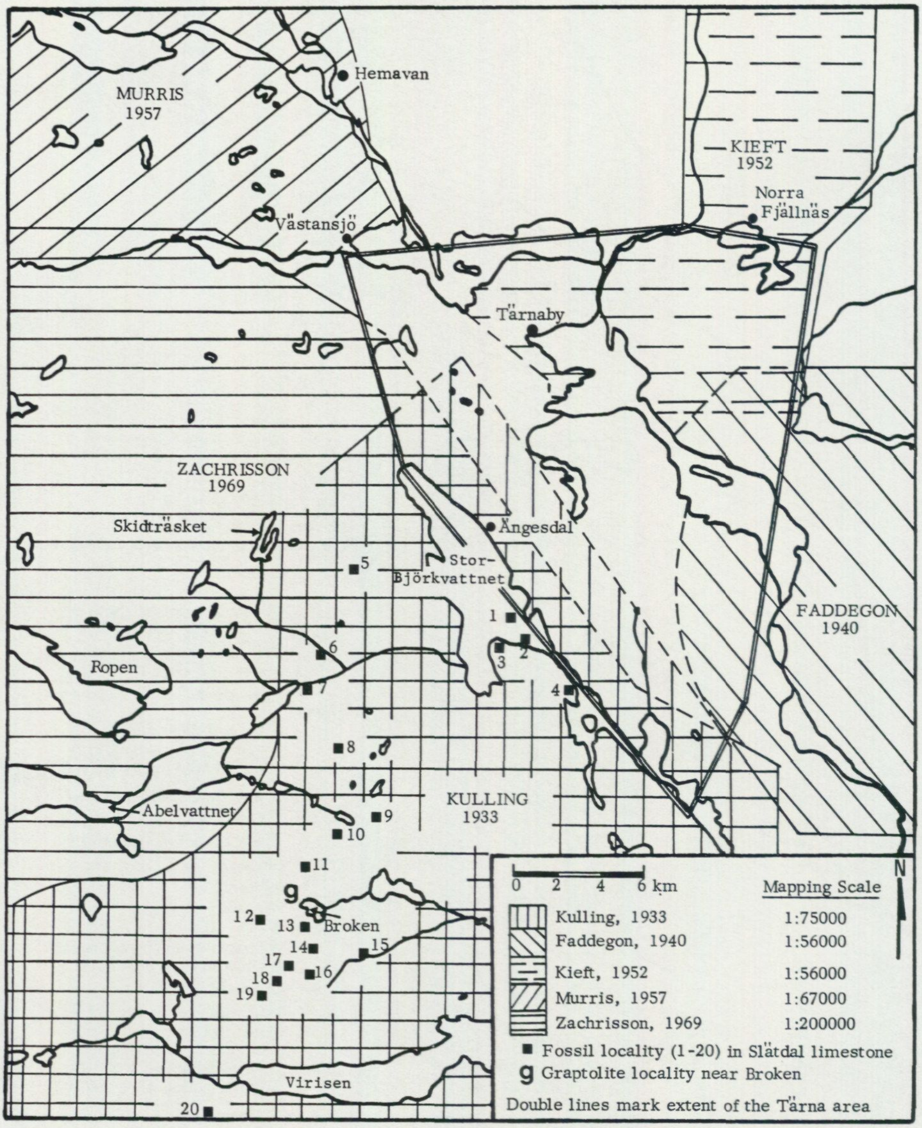


Fig. 2. Location of the Tärna—Björkvattnet area with regard to previous work in adjacent areas.

Quaternary rocks of Sweden (Sveriges Geologiska Undersökning*, 1958 described by Kulling, *in* Magnusson et al. 1960), the summary of stratigraphic and tectonic results of the Ore Prospecting Department's work in the Caledonides during 1962—1968 (Zachrisson 1969) and the synthesis of the Scandinavian Caledonides presented by Kulling (*in* Strand and Kulling 1972).

* Geological Survey of Sweden, henceforth referred to as SGU.

LITHO-TECTONIC CLASSIFICATION

The Tärna—Björkvattnet area consists of (Plate I):

1. The Pre-Cambrian quartz syenite and overlying sediments of the Bångfjället Complex** which outcrop in the northern part of the area, wedge out just to the west of Tärnaby and are surrounded by
2. the variably metamorphosed sediments and igneous rocks of the Seve-Köli Nappe Complex.

The interpretation of the age of the quartz syenite within the Bångfjället Complex as Pre-Cambrian is substantiated by a Rb-Sr, whole-rock, isochron date of 1520 ± 140 m.y. (Stephens and Wilson in prep.).

The metamorphics south, west and east of Tärnaby include both Seve and Köli rocks (see discussion below). The latter lie at the same tectonic level as and contain a similar stratigraphy to the Köli rocks of the type-area (Kulling 1933). These Seve and Köli rocks are referred to here as belonging to the Björkvattnet Unit (Plate I).

The Bångfjället Complex to the northeast and the Björkvattnet Unit to the northwest of Tärnaby are tectonically overlain firstly by a sequence of gabbro-intruded calcareous phyllites and layered metavolcanites (Lower Laxfjället Unit) and later by gabbro-intruded garnet-staurolite schists (Upper Laxfjället Unit) (Plate II). The Laxfjället Units form two separate and, relative to the Björkvattnet Unit, higher litho-tectonic units within the Seve-Köli Nappe Complex.

The boundaries between the litho-tectonic units (Bångfjället Complex, Björkvattnet Unit, Lower and Upper Laxfjället Units) are thrusts and are defined by an intense mylonitization of the rocks adjacent to the thrusts. The major structure which dominates the outcrop pattern and folds these thrusts (Plate 1) is a late, N-plunging synform, referred to here as the Tärna Synform.

The question of subdivision of the Björkvattnet Unit has to be considered in the light of the wider problem regarding usage of the terms "Seve" and "Köli" in Scandinavian literature. These terms, as Zachrisson (1973) points out, are and always have been loosely defined but they fulfil an important role particularly in large-scale regional discussions. Seve is employed as the name for the high-grade schists, gneisses and amphibolites of the lower part of the Seve-Köli Nappe Complex and Köli refers to the overlying, low-grade, sedimentary-volcanic sequence predominantly of Lower Palaeozoic age. When the contact is studied in detail, it does, however, prove difficult to define a more exact boundary. Trouw (1973), in a study of the Marsfjällen area of southwestern Västerbotten, suggests a locally tectonic contact between these two supergroups marked by a predominantly post-metamorphic thrust associated with blasto-mylonites. No such

** Kulling (*in* Strand and Kulling 1972, p. 206) refers to these rocks as the Bångfjället rock complex.

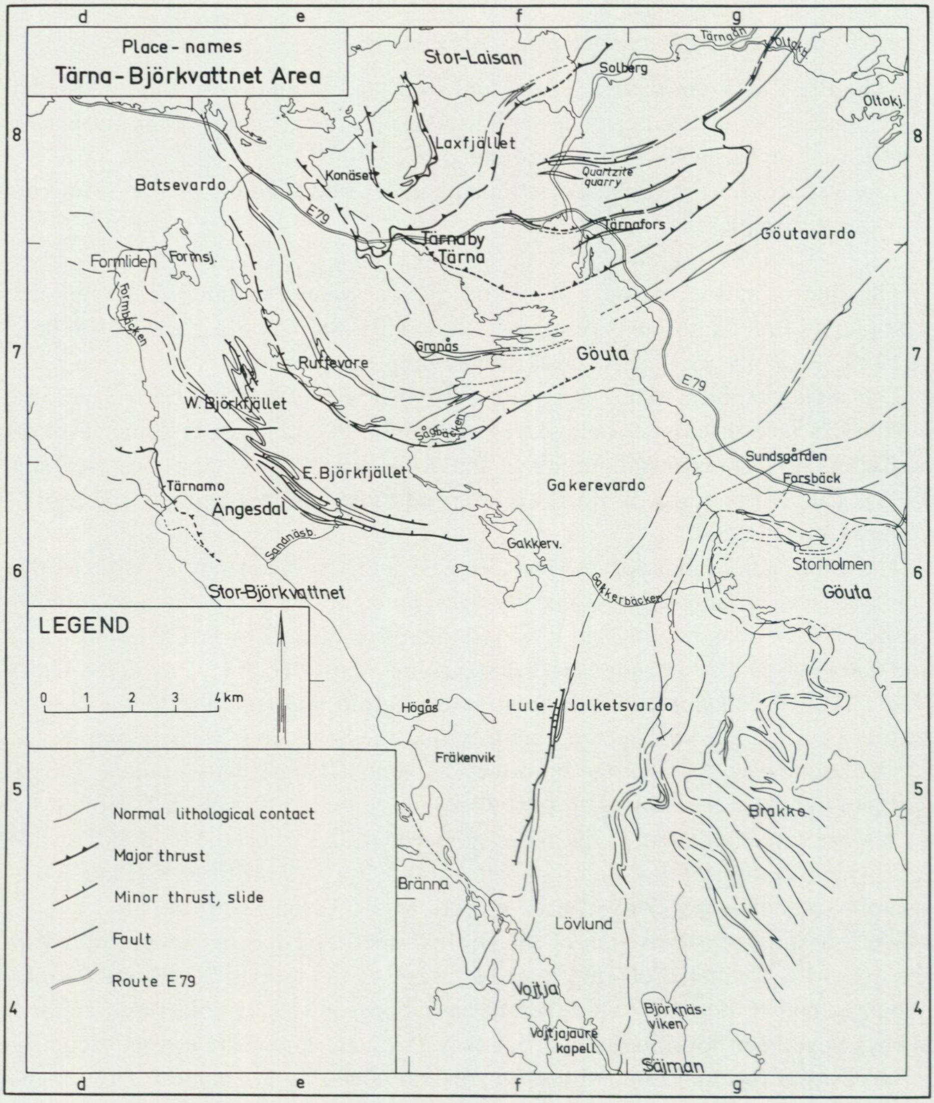


Fig. 3. Place-names referred to in the text.

distinct break has been mapped in the Tärna—Björkvattnet area. Although there are late, retrograde mineral reactions in the transition zone, there is essentially a prograde metamorphic sequence between the low- and high-grade rocks (Stephens 1973). Much of the Björkvattnet Unit lies within the chlorite and biotite zones, and is composed of a well layered sequence of greywackes, quartzites, phyllites and intercalated volcanics. It is within these rocks that the Lower

Palaeozoic stratigraphy established by Kulling (1933) can be recognised. In the southeast of the area, garnetiferous phyllites are intercalated within basic volcanics belonging to the Seima "Series" (Kulling 1958) which is generally accepted as part of the Köli sequence. Higher grade, coarser schists and amphibolites outcrop to the east of this part of the Björkvattnet Unit and also fringe the Bångfjället Complex on its eastern and southern borders.

All localities mentioned in the text are referred to by the nearest topographic feature or settlement name (Fig. 3), and located using the 5 km square grid system adopted for Swedish topographic maps. The numerical and letter characters are found along the map edges (Fig. 3 and Plate I) and the appropriate numerical character is recorded first.

STRATIGRAPHY AND FACIES INTERPRETATION OF THE BJÖRKVATTNET UNIT

INTRODUCTION

Most of Kulling's stratigraphic units (Table 1) have been recognised in the Tärna—Björkvattnet area, and utilized for the stratigraphic nomenclature of the Seve-Köli rocks to the south, west and east of Tärnaby. The "series" of Kulling's nomenclature are equivalent to formations and referred to as such here. No formal revision is presented for those units originally defined by Kulling outside the Tärna—Björkvattnet area, but the Seima "Series", originally defined (Kulling 1958) around Seimajaure (now called Säjman) in the southeast of the area, has been renamed the Seima Formation, and one new formation term — the Forsbäck Formation — has been introduced and is defined here. The informal terms Brakko schists and Tärna schists respectively refer to the garnetiferous schists and amphibolites to the east of the low-grade, sedimentary-volcanic sequence, and the highly deformed schists fringing the Bångfjället Complex.

Following the work of Kulling (1933—1969), Nilsson (1964) and Zachrisson (1964), a regional stratigraphy for the low-grade (Köli) rocks of northwestern Jämtland and southern Västerbotten was published by Zachrisson (1969). The important features of this scheme (Table 1) are as follows:

1. The Blåsjö phyllite (Nilsson 1964), the Lasterfjäll Calcareous Phyllite (Zachrisson 1964) and the Lövfjäll Formation have been correlated, and a complex metavolcanite, quartzite conglomerate and phyllite sequence is considered to have been deposited stratigraphically above the Lövfjäll Formation.
2. The rocks have been divided into three groups, named in ascending stratigraphic order the Tjopasi, Lasterfjäll and Remdalen Groups; the type-area of each of these groups is in the Remdalen area of Västerbotten. The Rem-

TABLE 1. The stratigraphic schemes of Kulling (1933 – 1969), Nilsson (1964) and Zachrisson (1964, 1969).

BJÖRKVATTNET-VIRISEN AREA				
Kulling (1933, 1958, 1969, in Gavelin and Kulling 1955, in Magnusson et al. 1960)				
Viris "Series"	Quartzite/greywacke-quartzite: Quartzite conglom. at base	500-1000 m		SILURIAN
Vesken Formation	Greywacke, shale-boulder conglom.			
Lövfjäll "Series"	Sandy and calc. phyllite, slate and greywacke	1000 m+		
Broken "Series"	Dark, fossil. slate (graptolitic): Calc. qtzite. in lower part	100 m+	Middle/Upper Llandovery	
Slätdal "Series"	Dark slate, fossiliferous lst.	10-50 m	Upper part of Ashgillian	ORDOVICIAN (and possibly older strata)
Vojtja "Series"	Quartzite/quartzite conglom.	10-160 m		
	EROSION BREAK			
Gilliks "Series"	Arkose, polymict conglom., slate: Chloritic schist intercal.	1500 m+		
(Mesket) Seima "Series"	Greenstone lava/agglom: Keratophyre			
Ro "Series"	Basic volcanics: Serpentinite conglom. on peridotite substratum: Qtzite. conglom./serp. breccia	0-200 m TOTAL=3000-4000 m (min.)		

dalen and upper part of the Lasterfjäll Groups, as defined by Zachrisson, are absent in the Björkvattnet Unit.

3. Following the earlier suggestion of Du Rietz (1941), the Viris quartzite (Kulling 1933) is regarded as a coarser facies variation within the Lövfjäll phyllites, rather than a distinctly younger rock unit above the Lövfjäll and Vesken Formations (Kulling 1933, 1969); this variation is thought to occur below the level of gabbro-intruded calcareous phyllite, found exclusively within the Blåsjö phyllite and Lasterfjäll Calcareous Phyllite.

Both textural and compositional terms have been employed in the rock nomenclature. Where composition has been the determining factor, the nomenclature of Wallis et al. (1968) has generally been employed. Mineral qualifying terms such as feldspathic, micaceous etc. are employed where the particular mineral group exceeds 10 % in a modal analysis, but is less than 35 % which is required for it to be represented as a separately defined compositional term (see Wallis et al. 1968).

In the following description of the stratigraphy of the Björkvattnet Unit, the

BLÅSJÖ (NW. JÄMTLAND)	REMDALEN (S.VÄST.)	Zachrisson 1969	
Nilsson 1964	Zachrisson 1964,1969		
Phyllite, quartzite, greenstone and greenschist with Portfjäll conglomerate at base	Phyllite and greenschist with Remdalen Quartzite-Conglomerate Formation at base	REMDALEN GROUP	SILURIAN
Basalt-quartz keratophyre formation	Stekenjokk Quartz-Keratophyre and Lasterfjäll Quartz-Keratophyre bearing Fm.	LASTERFJÄLL GROUP	
Blåsjö phyllite; gabbro-intruded	Lasterfjäll Calcareous Phyllite; gabbro intruded in upper part		
	Bellovare Formation		ORDOVICIAN (and possibly older)
	Tjopasi Formation	TJOPASI GROUP	

different formations are described from bottom to top. Although the stratigraphic relationship between the Brakko and Tärna schists is uncertain, they both stratigraphically underlie the characteristic Köli sediments and volcanics of the upper part of the sequence.

DESCRIPTION OF THE STRATIGRAPHY

BRAKKO SCHISTS

Metasediments including schists and marble pods

The schistosity in these rather monotonous quartz-garnet-(hornblende) schists is defined by aligned flakes of muscovite and a deep olive-green to brown biotite; garnet is deep pink in colour, commonly possesses idioblastic form and, together with biotite, forms conspicuous porphyroblasts. In the more calcareous schists additional phases include blue-green hornblende, with blades commonly 5 cm long, limonite-stained calcite and coarse epidote several millimetres long. The hornblende poikiloblasts are arranged in a garbenschiefer (Spry 1969, p. 269) structure on the main schistosity but they are also deformed and occasionally well oriented in the schistosity.

Minor variations include discontinuous horizons of graphite schist and a single occurrence at Forsbäck (6g) on Route E 79 of a polymictic conglomeratic schist containing scattered pebbles of graphite schist and marble.

A more distinctive schist member occurs discontinuously over Brakko and in the Göuta—Forsbäck area to the north. It consistently overlies the first thick (probably greater than 300 m) amphibolite sheet and proved to be an important marker horizon in the elucidation of the structure of the higher grade area in the east. This schist is conspicuously non-garnetiferous but rich in quartz, epidote-clinozoisite, muscovite, biotite and opaques. A pronounced layering is often present with local concentrations of epidote and opaque minerals alternating with quartz-tourmaline laminae up to 1—2 mm thick; porphyroblasts of calcite, biotite and blue-green hornblende also occur. In the lower part of Gakkerbäcken and on the western shore of Göuta, southwest of Forsbäck (6g), a more homogeneous epidote-plagioclase rock lies above the layered quartz-epidote schist. This rock contains deformed, leucocratic knots up to 0.5 cm across which are composed predominantly of fine-grained plagioclase, white mica and chlorite and are set in a groundmass of epidote group minerals, muscovite and plagioclase. A coarser twinned plagioclase is occasionally present within the knots which appear to be relict phenocrysts; this is substantiated by the local occurrence of deformed pillow structure (Fig. 4). It would appear, therefore, that this more homogeneous lithology represents a metamorphosed lava deposited under submarine conditions.

Discontinuous pods of calcite marble and more continuous layers of massive to faintly foliated epidote-calcite rock (20—25 % epidote-clinozoisite) are present within all the schists. Both contain variable amounts of quartz, muscovite, biotite and less commonly blue-green hornblende up to 2—2.5 cm long. On the southern shore of Storholmen in Göuta (6g), the contact between the marble pods and the schists in which they are enclosed is marked by a reaction zone of extremely coarse-grained calc-silicate rock; this is composed of strongly deformed hornblende blades commonly 10—15 cm long, subidioblastic garnet up to 10 cm across and variable amounts of calcite, epidote, mica minerals, quartz and galena. The accessibility of Si, Al, Fe and Mg from the schists and excess Ca in the originally more continuous limestone layers have permitted the formation of such skarn deposits as rims to the now relic pods of marble.

Metabasites

Strongly lineated epidote amphibolite layers alternate as concordant sheets within the metasediments. Epidote, deep blue-green hornblende and occasional biotite define the mineral lineation conspicuous in hand specimen, whilst plagioclase and leucoxene are invariably present in the groundmass. Mafic layers may alternate with plagioclase-epidote layers up to 10—15 cm thick, whilst more homogeneous amphibolite is occasionally associated with layers studded with strongly deformed acidic lenses up to 1 cm long and 1—2 mm across, which consist of fine-grained plagioclase and subordinate epidote-clinozoisite. These local inhomogeneities are interpreted as recrystallized, more calcic plagioclase phenocrysts.

Discontinuous layers of (garnet)-hornblende garbenschiefer are associated with the amphibolite sheets and often occur at the contact with the adjacent schists where amphibolite and garbenschiefer are closely interlayered. Coarse (commonly 0.8—1 cm long, up to 5 cm long), deep blue-green hornblende poikiloblasts characterize these rocks and are often preserved in a deformed rosette-shaped growth pattern on the poorly developed foliation surface. Other mineral phases which are variably present include garnet, carbonate and coarse (1—2 mm) epidote grains; all these phases are set in a recrystallized quartz-feldspar matrix. A minor variation is towards a more leucocratic type again containing idioblastic garnet and poorly oriented hornblende but with a relatively higher proportion of even-grained plagioclase groundmass; this type weathers to a characteristic buff colour and contrasts sharply with the more basic variety. These garbenschiefer types are readily distinguished from the hornblende schists by the relative paucity of mica minerals and non-schistose structure in hand specimen.

TÄRNA SCHISTS

The schists which lie adjacent to the Bångfjället Complex in Oltokbäcken (8g), on northwestern Göutavardo (8g) and south of Tärnaby (7e—f) are predominantly composed of semi-pelitic units which pass transitionally into more massive psammites and micaceous quartzites with thin pelitic intercalations. Although the dip of the compositional layering and the main foliation is usually steep, these schists tectonically underlie the Bångfjället Complex on northwestern Göutavardo where the dip is 45° or more NW. This unit thins out just west of Tärnaby beneath the major thrust which separates the Björkvattnet and Lower Laxfjället Units (Plate I). Unlike the Brakko schists, bedding is mappable within these rocks and porphyroclasts of various kinds of ”feldspar” are preserved particularly within the more massive layers (Fig. 4). These include lamellar twinned saussuritized plagioclase, strongly altered and untwinned K-feldspar, microcline, microcline perthite, cryptoperthitic feldspar and granitic rock fragments. By contrast, untwinned or simple twinned, unaltered poikiloblasts of albite are also present. These schists, which also contain garnet, muscovite, biotite and zoned epidote porphyroblasts, are characterized by the intense phyllonitization (Knopf 1931; Christie 1960) of the main schistosity. The degree of phyllonitization increases towards the contact with the Bångfjället Complex.

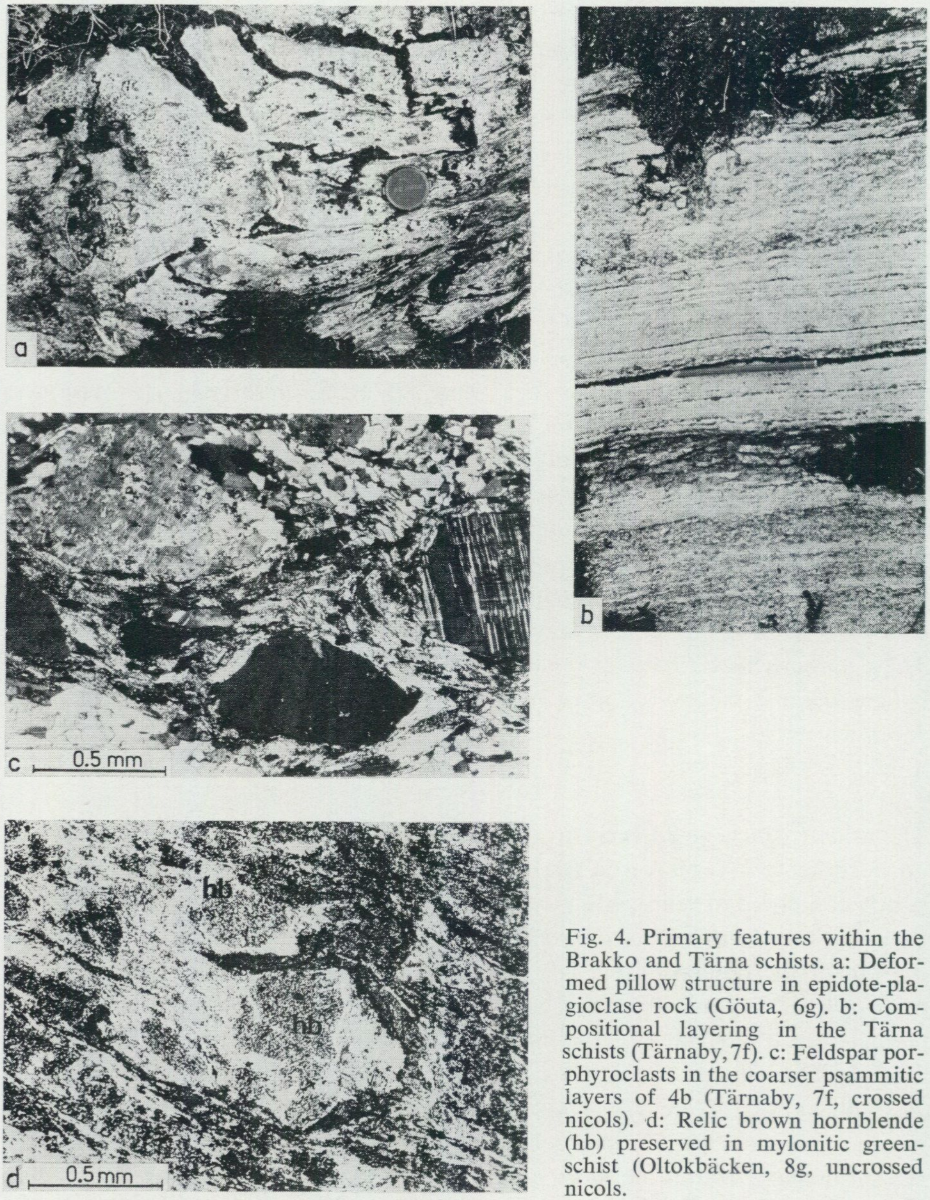


Fig. 4. Primary features within the Brakko and Tärna schists. a: Deformed pillow structure in epidote-plagioclase rock (Göuta, 6g). b: Compositional layering in the Tärna schists (Tärnaby, 7f). c: Feldspar porphyroclasts in the coarser psammitic layers of 4b (Tärnaby, 7f, crossed nicols). d: Relic brown hornblende (hb) preserved in mylonitic greenschist (Oltokbäcken, 8g, uncrossed nicols.

Minor garnetiferous amphibolite layers and a strongly deformed metagabbro lens (NW Göutafjället, 8g), up to 250 m long, also occur within the schists.

Immediately adjacent to the Bångfjället Complex there is a discontinuous development of well layered, mylonitic (Christie 1960) greenschists and feldspar-rich rocks. Along the shore section south of Tärnaby (7f), these rocks are intercalated within the phyllonitic schists described above. The acidic layers are rich in finely recrystallized quartz and feldspar and enclose coarser deformation-twinned plagioclase megacrysts. Finer grained white mica-epidote-leucoxene layers possess a lenticular structure due to the intense deformation. By the intercalation of basic laminae rich in pale blue-green amphibole, epidote-clinozoisite and chlorite, these rocks pass transitionally into more homogeneous greenschists. Coarser relics of a green-brown hornblendic amphibole with optically continuous tremolite overgrowths are preserved within the basic component (Fig. 4). It is suggested that these rocks represent either strongly deformed metavolcanites or a series of deformed and recrystallized gabbroic intrusions whose primary texture has been obliterated during the intense, post-crystallization deformation.

FORSBÄCK FORMATION

The type-area of the Forsbäck Formation is situated along the coast section on the northern shore of Göuta, WSW of Forsbäck, and in the main road sections in the western part of the same settlement (6g). The Forsbäck Formation is approximately 200—300 m thick in its type-area, and consists predominantly of a light grey, garnetiferous phyllite with psammitic composition, containing subordinate intercalations of finer grained, grey phyllite (pelitic composition) with an increased proportion of muscovite, biotite and chlorite; bedding is still preserved in these rocks. The mica and feldspar (plagioclase in the matrix) contents of the psammite both exceed 10 % and the recrystallized matrix encloses relic porphyroclasts of dusky K-feldspar, saussuritized plagioclase, microcline, perthitic feldspar and also granitic rock fragments; occasional bluish quartz and partially recrystallized quartz clasts are also present. In Gakkerbäcken there is at least one thin greenschist horizon, whilst to the south minor graphitic phyllite layers alternate with quartz-biotite phyllites in the upper part of the formation.

The transition beds to the overlying Gilliks Formation are well exposed around the headland on the southwestern shore of the straits connecting the two principal portions of Göuta. They consist of finely interbedded graphitic phyllite and subordinate tuffite (mixed sedimentary-volcanic rock; Zachrisson 1971) and greenschist layers characteristic of the succeeding Gilliks Formation, and sparsely garnetiferous quartz phyllite similar to the remainder of the Forsbäck Formation; garnet appears for the first time in these upper transition beds, whilst quartz, chlorite and plagioclase appear to coexist in the matrix of the phyllites.

Although masked by some lenticularization of the finer grained phyllite layers and retrogression of the metamorphic fabric, these garnetiferous phyllites appear to pass transitionally downwards into the Brakko schists. Absence of carbonate banks, increased feldspar content within the psammitic phyllites and a larger pelite component distinguish the phyllites of the Forsbäck Formation from the Brakko schists.

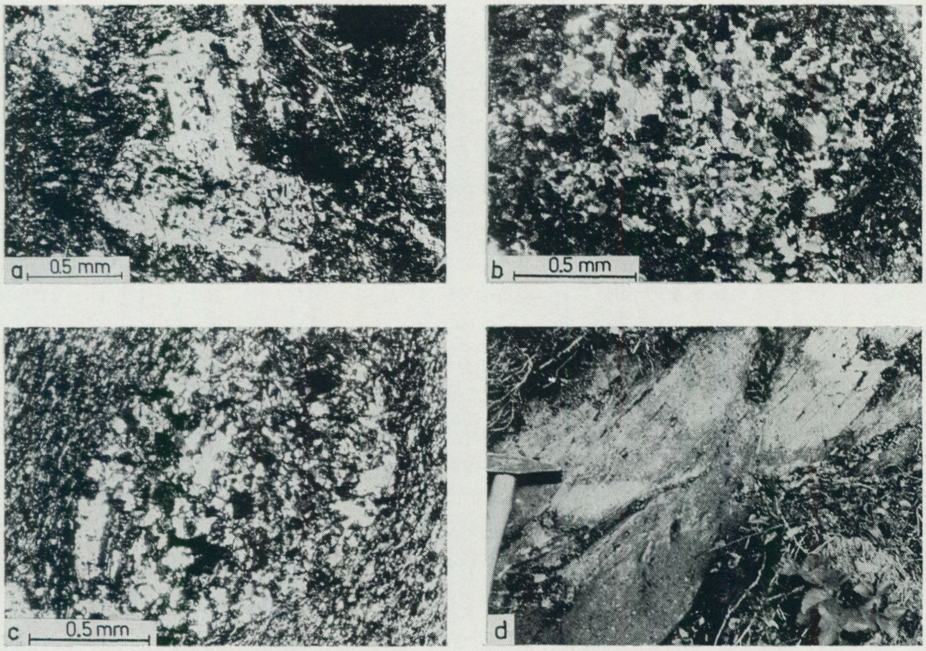
As the Forsbäck Formation is traced southwards from its type-area it thins, and horizons of greenschist and phenocrystic greenstone appear at approximately the same level (Plate I). In the southeast of the area this formation is only represented by thin intercalations of (graphitic) grey, quartz-biotite and garnetiferous phyllites within a thick volcanic sequence (Seima Formation).

SEIMA FORMATION

The Seima Formation has its type-area along the stream sections flowing into the northern and northeastern parts of Säjman (4g). It consists for the major part of massive greenstones which are often amygdaloidal and composed of alternating phenocrystic and non-phenocrystic layers. The amygdales are represented by coarse knots of calcite $\pm$ chlorite. The degree of preservation of the originally more calcic plagioclase phenocrysts depends on the variable metamorphic grade across the strike of these volcanics (Fig. 5). At higher structural levels the matrix is composed of a weakly lineated mass of actinolite, epidote, leucoxene and albite, whilst at deeper levels the greenstones are composed of coarser, sharply zoned amphibole or blue-green hornblende, epidote-clinozoisite and minor biotite. At occasional localitites on the southwestern slopes of Brakko, pillow structure is preserved within these greenstones (Fig. 5). Preliminary investigation of their petrochemistry (Stephens 1973) has shown that, in part, the greenstones have abyssal tholeiite characteristics, indicative of either an active spreading or early stage island arc environment.

The greenstones are associated with well laminated greenschists, particularly in the upper part of the formation. The mafic layers in the greenschists are rich in chlorite, fine-grained actinolite, epidote-clinozoisite, calcite and leucoxene, and alternate with more leucocratic layers containing an increased proportion of quartz and albite. An important basic agglomerate horizon has been mapped just northeast of Björknäsviken (4g), whilst more recrystallized agglomerate layers are also present in the lower part of the formation. The main agglomerate horizon contains fragments of dark greenstone, displaying both phenocrystic and amygdaloidal textures, gabbro and talcose rock set in a green, schistose matrix.

Serpentinite and subordinate metagabbro occur as both major (several hundreds of metres across) and minor (several metres across) lenticular masses *within* the Seima Formation. Visible primary layering is not preserved within these masses and no metamorphic aureole is present in the surrounding rocks. Antigorite, which forms an interlocking mesh framework with fibro-lamellar structure, and anhedral magnesite are the main constituents of the serpentinite. The magnesite is transected by the antigorite fibres and occurs as knots (<1 mm



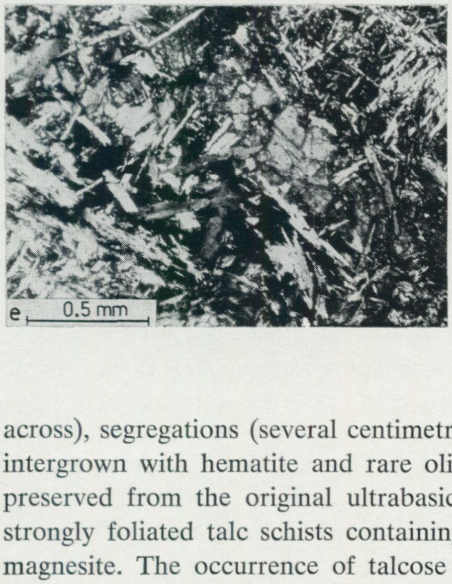


Fig. 5. Primary features within the Seima Formation. a: Relic plagioclase phenocrysts in actinolitic amphibole-epidote matrix (SE of Vojtjajaure kapell, 4f, crossed nicols). b: Epidote-albite knot in epidote-zoned amphibole matrix; original cryst shape poorly preserved (SW of Brakko, 4g, crossed nicols). c: Epidote-plagioclase knot in lineated hornblende + epidote matrix (SW of Brakko, 5g, crossed nicols). d: Pillow structure in greenstone (SW of Brakko, 5g). e: Relic olivine transected by antigorite fibres in serpentinite (SW of Brakko, 5g, crossed nicols).

across), segregations (several centimetres across) and veinlets. Skeletal ilmenite intergrown with hematite and rare olivine relics are the only primary phases preserved from the original ultrabasic rock (Fig. 5). The smaller lenses are strongly foliated talc schists containing subordinate amounts of dolomite and magnesite. The occurrence of talcose rock fragments within the main agglomerate horizon confirms the pre-tectonic character of these ultrabasics and their early hydrothermal alteration to serpentinite and locally talcose rock. They are generally located in the lower part of the Köli and upper part of the Seve sequences throughout the Scandinavian Caledonides (Kulling 1933; Zachrisson 1969; Rui 1972). No serpentinite conglomerate or breccia, commonly associated with serpentinite bodies in other areas (Kulling 1933, *in* Magnusson et al. 1960; Du Rietz 1935; Oftedahl 1969; Trouw 1973), has been mapped in the present study. The enclosure of the predominantly serpentinite lenses *apparently at different levels within* the Seima Formation (Plate I) inhibits the stratigraphic usefulness of Kulling's term Ro "Series" for these rocks.

The Seima Formation has its maximum development around Säjman, where it is probably over 1 to 1.5 km thick. The Seima volcanics appear to thin out southwards (Kulling, *in* Gavelin and Kulling 1955), whilst they are virtually absent around Forsbäck in the northeastern part of the Tärna—Björkvattnet area. As already pointed out, the thinning of the Seima Formation northwards is accompanied by an increase in importance of the metasedimentary intercalations (graphitic grey, quartz-biotite and garnetiferous phyllites) within the formation, such that around Forsbäck metasediments dominate at this level (Forsbäck Formation). The Seima volcanics and associated serpentinites pass upwards into the graphitic phyllites, minor greywacke horizons and tuffaceous volcanics of the lower part of the Gilliks Formation. There is a marked change downwards in metamorphic grade across the strike of the Seima Formation, hornblende and garnet entering for the first time within the volcanics and the minor metasedimentary intercalations respectively. The metabasic rocks in the lower part of the formation are distinctly amphibolitic and the formation is underlain by the garnetiferous Brakko schists. Both the overlying graphitic phyllites etc. belonging to the lower part of the Gilliks Formation and the underlying Brakko schists can be traced northwards into the Forsbäck area. Thus, it appears that the Seima and Forsbäck Formations represent an important facies variation in the lower part of the Köli stratigraphy. This facies change occurs over a few kilometres and illustrates the local significance only of some of the formations.

GILLIKS FORMATION

The Gilliks Formation is subdivided into a lower part made up predominantly of fine-grained graphitic phyllite, minor carbonate lenses and tuffaceous volcanics, and an upper part composed mainly of quartz- and feldspar-rich greywackes. Important exposures of this formation occur along the straits connecting the two portions of Göuta, west of Sundsgården (6—7g). In this area the Gilliks Formation is greater than 1 km thick.

Lower part

The pyritiferous graphitic phyllites in the lower part of the formation contain various types of carbonate lenses including graphitic limestone, sandy phyllitic limestone and buff-coloured dolomite; these pods are less than a couple of metres across and are often strongly brecciated.

Layered and more homogeneous greenschist, albite-quartz rock and tuffite occur as thin, discontinuous intercalations within the dark phyllites and are interpreted as pyroclastic and mixed sedimentary-volcanic rocks. Thicker and more continuous tuffaceous horizons are, however, present in the Ruffevare—Batsevardo area (8d—e, 7e) and in the southeast around Lövlund and Vojtjajaure kapell (4—5f), adjacent to the Seima Formation.

The Ruffevare greenschist consists of alternating layers rich in pale green chlorite and white mica, and more acidic layers composed of quartz, albite and carbonate; minor limestone pods are also present. This unit forms an important marker horizon in the northwest of the area; it has also been recognised on the eastern limb of the Tärna Synform north of Lövlund, and in strongly attenuated form southeast of Tärnafors (7g) and northwest of Granås (7f). The Lövlund and Vojtjajaure tuffaceous volcanics are up to 200 m thick in places and, although variable in composition, are predominantly composed of layered greenschists with pelitic sedimentary intercalations.

Compact pyrite ore in a 1—2 m thick zone has been mapped along the presently flooded straits connecting Vojtja and Säjman, near the contact between a minor quartz keratophyre tuff and dolomite horizon, and overlying graphitic and grey phyllites. The quartz keratophyre tuff consists of coarser feldspar phenocrysts set in a white mica-chlorite matrix with more even-grained quartz and albite. It passes downwards into coarser feldspar-rich rocks of uncertain agglomeratic/conglomeratic origin.

In the lower part of the Gilliks Formation there are also subordinate greywacke intercalations characteristic of the upper part of the formation, and discontinuous beds of a poorly sorted, polymict conglomerate. The latter contains scattered pebbles of granite and buff coloured dolomite within a brown weathered calcareous matrix, and is found exclusively on the eastern limb of the Tärna Synform. The conglomerate is comparable to the Gilliks polymict conglomerate facies described by Kulling (1933).

Upper part

The upper part of the Gilliks Formation is more homogeneous consisting mainly of greywacke with a variable feldspar and graphite content; there are minor intercalations of graphitic and dark grey phyllite, and predominantly basic tuff. Porphyroclasts of blue and dark quartz, plagioclase, microcline and perthitic feldspar are all variably present; indeed, the feldspar content can be as high as 25 % (Fig. 6). A continental, "granitic" source is indicated.

The greywackes exposed on the eastern limb of the Tärna Synform invariably contain greater than 10 % feldspar; it is here also that the polymict conglomerate facies is found in the lower part of the formation. By contrast, in the Formliden area (8d) the total feldspar content of the greywackes is only a few per cent and

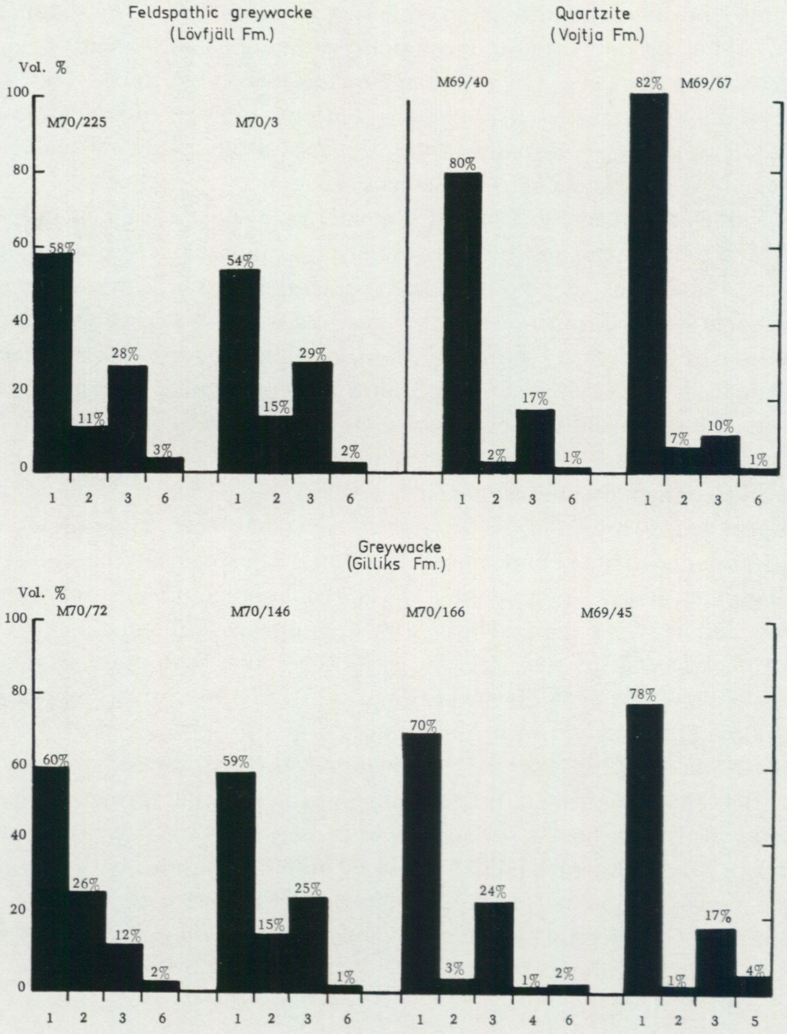


Fig. 6. Representative modal analyses of sandy layers in the low-grade rocks of the Björkvattnet Unit. 1 = Quartz; 2 = Feldspar; 3 = Mica; 4 = Carbonate; 5 = Graphite; 6 = Accessories.

they approach a quartzitic composition (Figs. 6, 7), whilst the polymict conglomerate facies appears to be absent. The matrix in the greywackes is generally composed of recrystallized quartz, white mica and chlorite. A minor amount of disseminated graphite, concentrated particularly around the quartz porphyroclasts, imparts an overall darker colour to the rocks. These greywackes are comparable to the Gilliks greywacke-quartzite facies described by Kulling (1933).

Greywacke layers up to 1—1.5 m thick alternate with laminae and beds of phyllite several centimetres thick. Although the bedding structure is generally plane parallel, load coasts, associated flame structures and also slump structures (Lule-Jalketsvardo, 5f) are occasionally preserved (Fig. 8); the loads are both symmetric and asymmetric in profile, and there is a variable sense of asymmetry between different exposures. Graded bedding is a relatively common feature (Fig. 7).

The uppermost part of the formation is marked on eastern Björkfjället (6—7e), in particular, by thin discontinuous layers of quartz phyllite and rather more massive feldspathic quartzite (Plate II). The phyllite has a distinct buff-coloured appearance, and concentrations of well foliated white mica and chlorite in thin pelitic layers give the rock a finely laminated aspect. At occasional localities there are scattered pebbles of vein quartz and quartzite, up to 2 cm across, within the phyllite.

Both the lower and upper parts of the Gilliks Formation show significant changes in lithofacies. In summary, these include:

1. The variation in the feldspar content of the greywackes.
2. The variable development of the poorly sorted, polymict conglomerate unit.
3. The variation in the type and abundance of the tuffaceous volcanic horizons, particularly in the lower part of the formation.

VOJTJA FORMATION

The Vojtja quartzite conglomerate, which forms the basal part of the Vojtja Formation and Zachrisson's Lasterfjäll Group, consists of deformed quartzite (90—95 % quartz) pebbles set in a micaceous, psammitic to quartzitic matrix. The quartz grains within the pebbles are conspicuously coarser than those in the matrix and often display strongly undulose extinction. On Lule-Jalketsvardo (5f), on the eastern limb of the Tärna Synform, the conglomerate has a lower pebble to matrix ratio and the pebbles appear less deformed (Fig. 9); indeed, subangular to subrounded pebbles predominate. A sharp upper contact with conglomeratic quartzite or homogeneous Vojtja quartzite is preserved northeast of Ruffevare (Fig. 9). The Vojtja quartzite forms massive ridges which outcrop conspicuously near Ruffevare (7e). Subordinate amounts of white mica and rare biotite are present, whilst the feldspar content in the quartzite is generally < 5 %. Minor variations include an increasingly feldspathic (>10 %) and porphyroclastic quartzite on Lule-Jalketsvardo (Fig. 9, compare also the composition of the greywackes in the upper part of the Gilliks Formation on Lule-Jalketsvardo) and a more inhomogeneous quartzite with intercalated pelite laminae on western Björkfjället, where there is also no associated quartzite conglomerate.

Although the bedding structure is commonly plane parallel to wavy in appear-

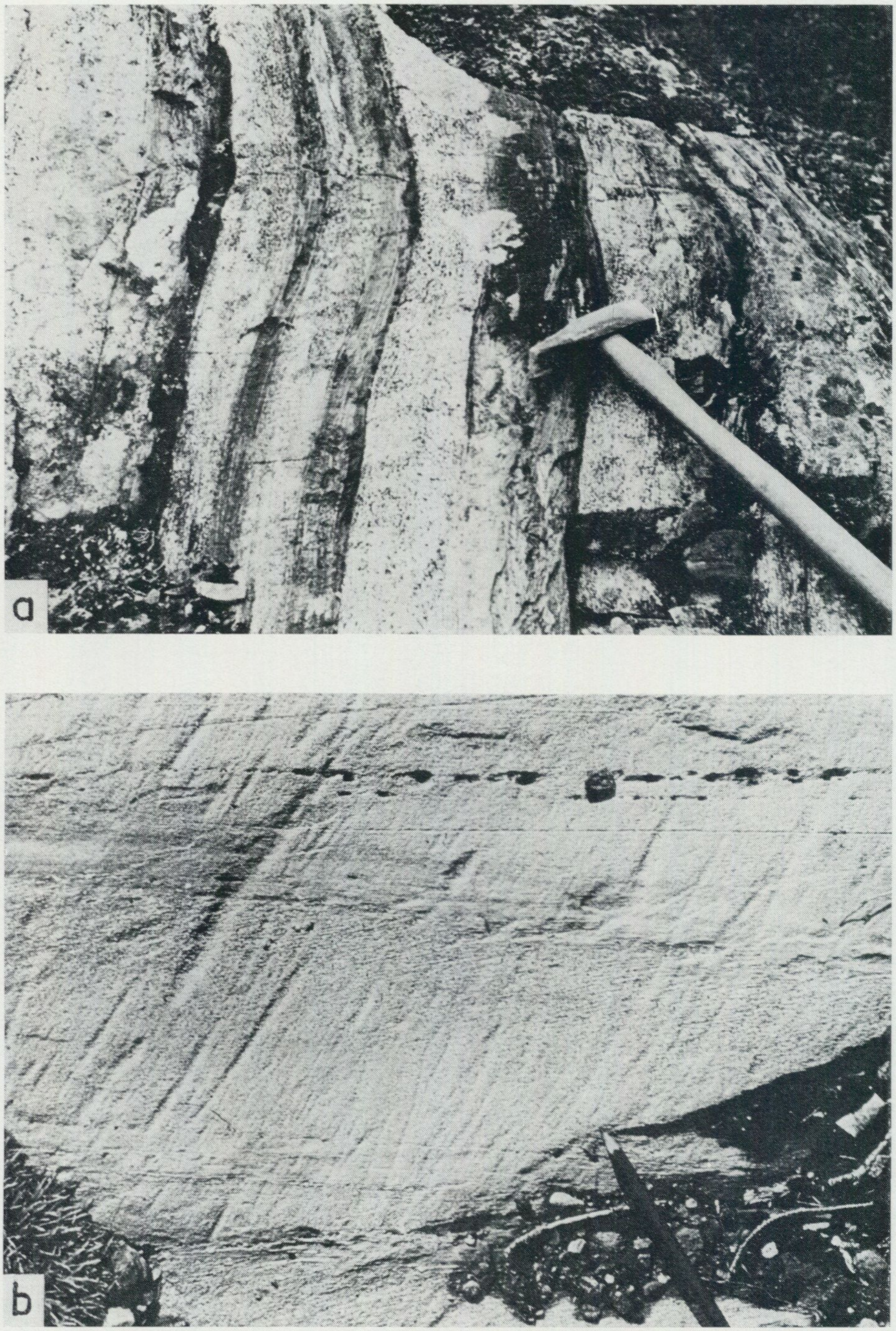


Fig. 7. Primary features within the greywackes of the Gilliks Formation. a: Plane parallel bedding structure (Lule-Jalketsvardo, 5f). b: Graded bedding (Sundsgården, 7g).

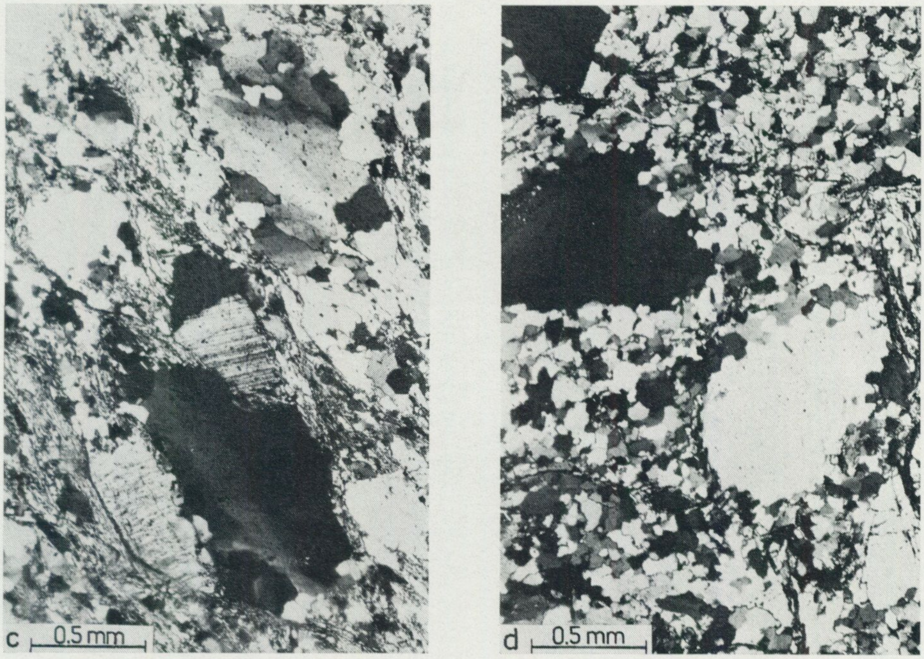


Fig. 7 (continued). Primary features within the greywackes of the Gilliks Formation. c: Feldspathic greywacke — note deformation of quartz and feldspar porphyroclasts and new grains around old, detrital, quartz grains (Lule-Jalketsvardo, 5f, crossed nicols). d: Quartzose greywacke — note development of subgrains and new grains (Formliden, 8d, crossed nicols).

ance, cross bedding is present in the Ruffevare outcrop, and flattened lenticular bedding and possible flaser bedding occur in the more inhomogeneous quartzites on western Björkfjället (Fig. 8).

The Vojtja Formation is discontinuously developed over the Tärna—Björkvattnet area and varies from 0—50 m in thickness.

SLÄTDAL FORMATION

The Slätdal Formation is composed of various types of limestone. In the thickest and most continuous outcrops along the upper part of Formbäcken (7d), light-coloured sandy limestone and stratigraphically younger dark blue-grey graphitic limestone occur as discontinuous pods or more continuous beds within fine-grained calcareous and slightly graphitic phyllites. The graphitic limestone contains >80 % calcite and the disseminated graphite is often concentrated along the calcite-calcite grain boundaries. Near Lule-Jalketsvardo there are lenses of coarsely crystalline calcite, up to 6 mm across, set in a fine-grained calcite-graphite matrix within the overall graphitic limestone; the matrix is strongly cleaved and deflects around the coarser areas which define a distinctly

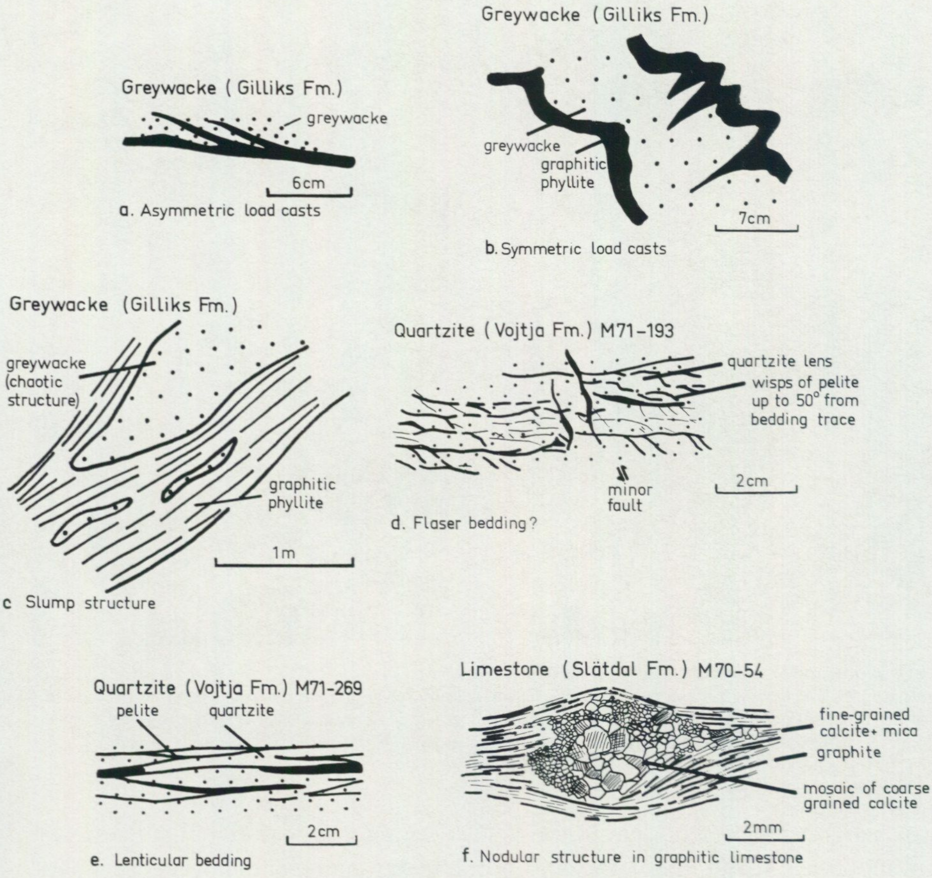


Fig. 8. Sedimentary structures in the Gilliks, Vojtja and Slätdal Formations.

nodular appearance (Fig. 8). Further to the south, outside the Tärna—Björkvattnet area, Kulling (1933) discovered a rich coral fauna in this limestone of Ashgillian age (Kulling, *in* Gavelin and Kulling 1955).

The Slätdal Formation varies from 0—180 m in thickness where it is most extensively developed along Formbäcken.

BROKEN FORMATION

Thin layers of graphitic slate and phyllite, dark quartz phyllite and subordinate calcareous quartzite, with total thickness varying from 0 to approximately 200 m, separate the Slätdal and/or Vojtja Formation from the overlying calcareous phyllites etc. of the Lövfjäll Formation. These lithologies are well exposed along the Sågbäcken stream section (7f) and on Björkfjället. They are corre-

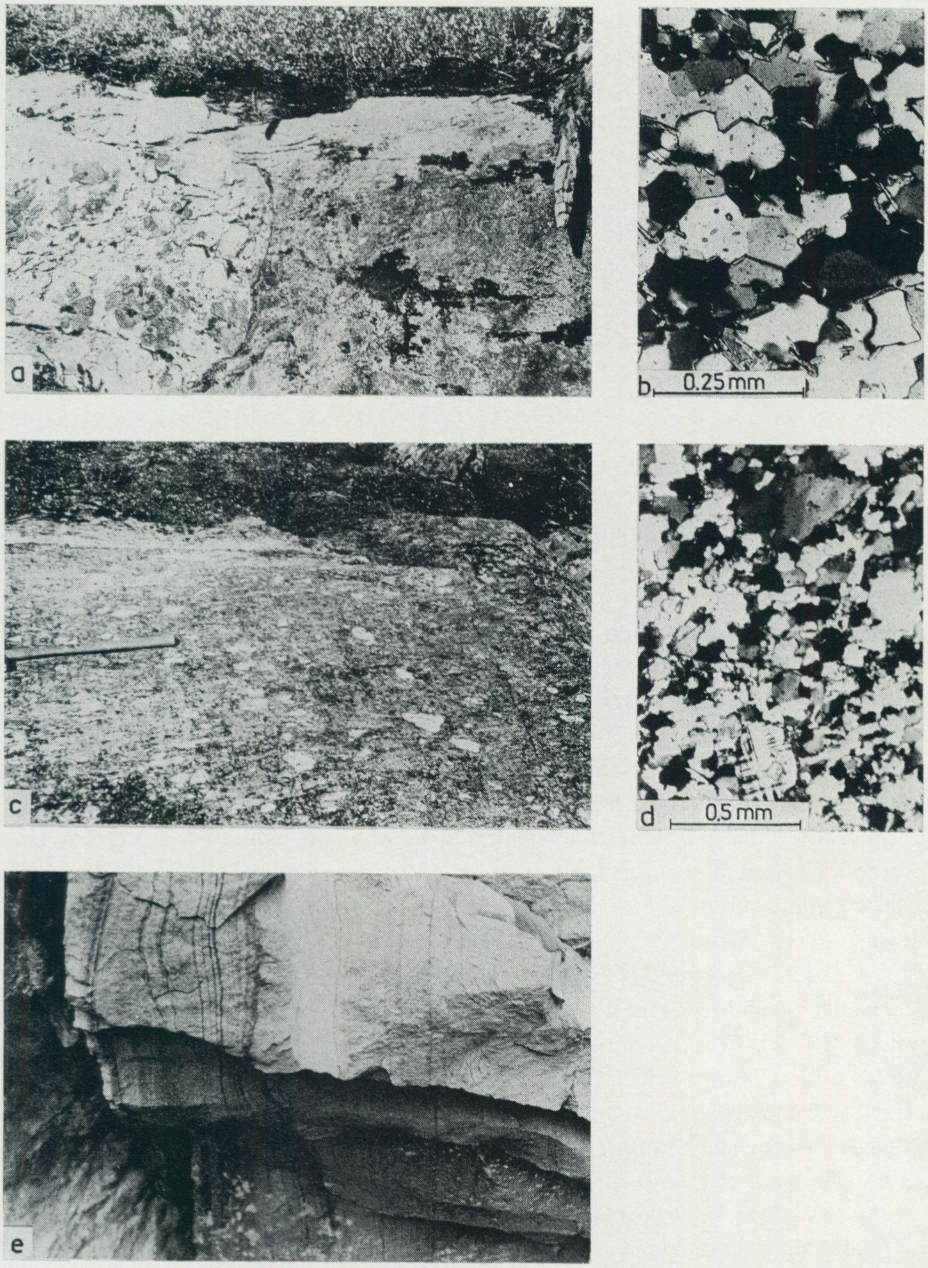


Fig. 9. Primary features within the Vojtja Formation and feldspathic greywacke facies of the Lövfjäll Formation. a: Contact between coarse Vojtja quartzite and Vojtja conglomerate (SE Ruffevare, 7e). b: Vojtja quartzite near Ruffevare — note polygonal grain boundaries (SE Ruffevare, 7e, crossed nicols). c: Matrix-rich Vojtja conglomerate (Lule-Jalketsvardo, 5f). d: Vojtja quartzite with quartz and feldspar clasts and curved/sutured grain boundaries (Lule-Jalketsvardo, 5f, crossed nicols). e: Symmetric load casts in feldspathic greywacke facies of the Lövfjäll Formation (SE of Tärnafors, 7g).

lated with the graptolitic slates found near Lake Broken known to be Middle and Upper Llandovery in age (Kulling 1933, 1969).

The graphitic and quartz phyllites of Björkfjället contain a distinctive layered tuffite horizon (Plate II) which is composed of layers rich in deep green chlorite and white mica alternating with more psammitic laminae (1—5 mm thick); small idioblastic garnets are scattered throughout the rock, whilst the subordinate mica and chlorite in the psammitic layers are concentrated along the quartz-quartz grain boundaries. By an admixture of graphite and mica this type passes transitionally into graphitic phyllite.

LÖVFJÄLL FORMATION

This formation is the youngest exposed within the Björkvattnet Unit. It consists of both finer grained calcareous and grey phyllites, and coarser feldspathic greywackes and phyllitic quartzites.

The strongly foliated quartz-mica-calcite (up to 20 %) matrix in the calcareous phyllites encloses porphyroclastic quartz and feldspar; quartz dominates and old detrital grains often show strongly undulose extinction with new, recrystallized grains at their margins. More pelitic layers are composed of white mica and chlorite, and are preserved as darker gleaming phyllite layers. A volcano-sedimentary source is inferred for these sediments. Bedding thickness varies from 0.5—1 mm thick psammite laminae in pelite to 15 cm thick calcareous phyllite beds alternating with pelite layers 2—3 cm thick. Although the bedding is generally plane parallel, minor slump structures and graded bedding are also present. Graded units in Sandnäsbäcken (6e) and near Högås and Fräkenvik (5f) are approximately 8—10 cm thick, possess sharply defined bases and fine upwards from calcareous phyllite to darker grey phyllite.

The coarser group of rock types is dominated by rather massive feldspathic greywackes which contain abundant quartz and feldspar porphyroclasts set in a weakly foliated quartz-mica matrix; both the mica and feldspar contents exceed 10 %. Less common lithic greywackes also occur with graphitic slate and calcareous quartzite fragments dominating over quartz and feldspar porphyroclasts. The greywackes pass transitionally into more even-grained phyllitic quartzites with >80 % quartz and subordinate calcite, white mica and feldspar. All these rock types contain subordinate intercalations of fine-grained, dark grey phyllite with overall pelitic composition. Much of the detritus appears to have been derived from a sedimentary source. The feldspathic greywacke and phyllitic quartzite beds range in thickness from 1 cm to 0.5 m, although the upper part of this range is more common. Load casts (Fig. 9), associated flame structures and graded bedding are commonly present besides more regular plane parallel bedding structure.

Three areas — Sandnäsbäcken (6e), Lule-Jalketsvardo (5f) and northeast of Ruffevare (7e) — are critical in understanding the relationship between the different rock types within the Lövfjäll Formation. These are now described:

1. The coarser grained feldspathic greywackes and phyllitic quartzites occur entirely within the calcareous phyllites along the Sandnäsbäcken stream section (Plate 1). On the basis of graded bedding evidence, it appears that the calcareous phyllites lie stratigraphically above the coarser grained rock types as well as beneath them as suggested by Kulling (1933).
2. South of Lule-Jalketsvardo peak calcareous phyllites, which lie stratigraphically above the Broken graphitic phyllite, thin out and are apparently replaced northwards by coarser feldspathic greywackes which subsequently overlie the same graphitic phyllite. North of the peak the greywackes also thin out, and are absent on eastern Gakerevardo (6f—g) and southeastern Göutavardo (7g) where calcareous phyllites overlie either Gilliks Formation or discontinuous horizons of Broken Formation and Vojtja Formation (quartzite only). Repetition and faulting of these beds near the peak of Lule-Jalketsvardo (Fig. 10) are associated with folds congruent to the N-plunging Tärna Synform. The occurrence on southeastern Göutavardo of grey and slightly graphitic phyllites which appear to mark a transition from the Lövfjäll Formation down into the underlying Broken Formation suggests that this repetition and faulting are confined to Lule-Jalketsvardo and do not explain the overall relationship between the coarser greywackes and finer calcareous phyllites. The pattern is explained, however, in terms of facies variation where the coarser greywacke facies is developed diachronously in different areas.
3. The extensive occurrence of feldspathic greywacke northeast of Ruffevare (Plate I) suggests that this facies is dominant in that area and developed to the virtual exclusion of the finer grained rocks. Although there is some evidence (discussion later) for tectonic thinning along the contact with the Broken Formation, it is possible that the coarser grained facies is developed down to the base of the Lövfjäll Formation.

Thus, it is concluded that the feldspathic greywackes and phyllitic quartzites represent a coarser facies variation within the finer calcareous phyllites, and that this facies change is diachronous. This is similar to the Viris/Vesken/Lövfjäll relationship in Kulling's (1933) type-area, the Viris/Lövfjäll part of which is discussed by Zachrisson (1969). However, at present, no attempt is made to correlate the coarser greywackes of the Tärna—Björkvattnet area with the coarser facies further south.

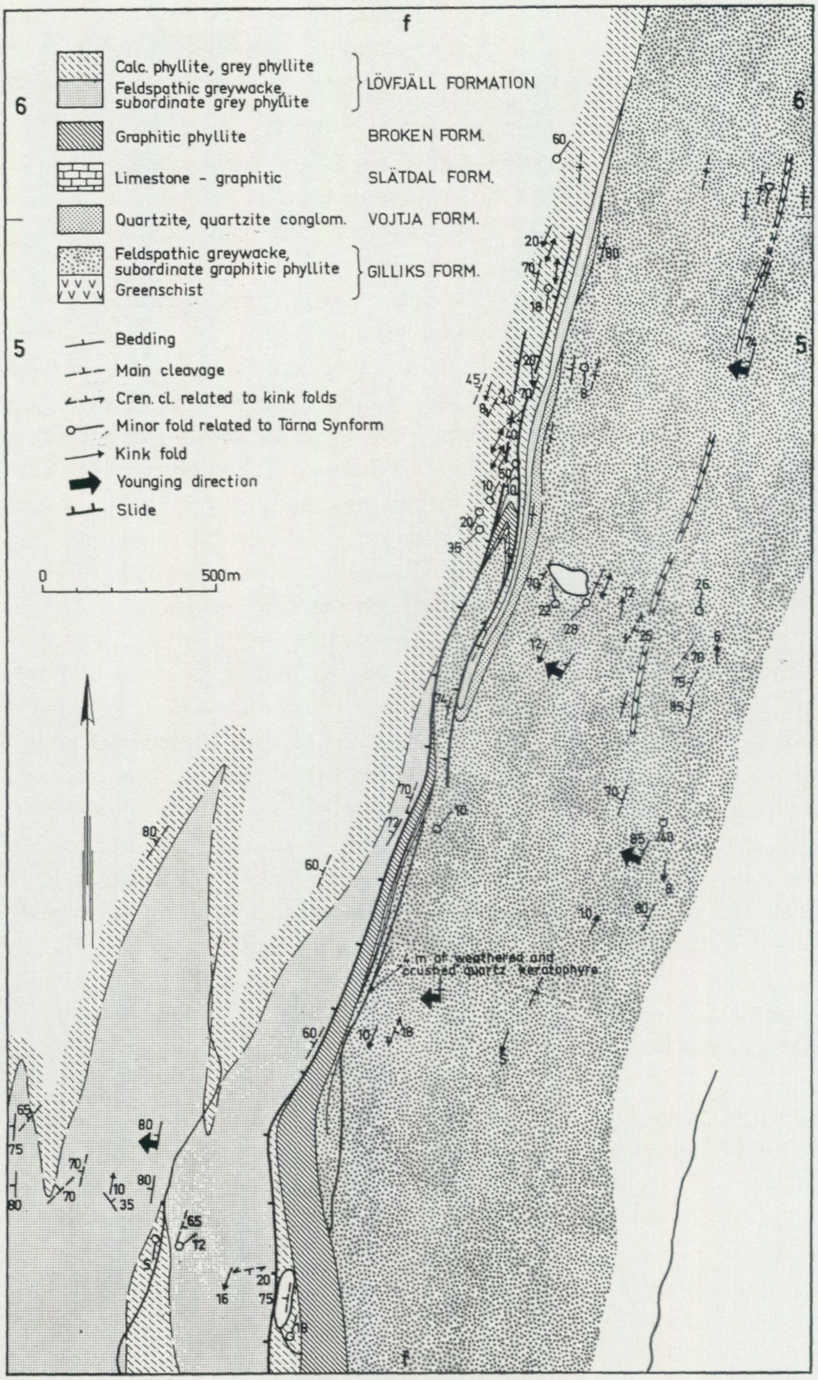


Fig. 10. Geological map of Lule-Jalketsvardo. The main cleavage in the coarser lithologies is the early, regional cleavage; in the phyllitic rocks it is a penetrative crenulation cleavage.

DISCUSSION

DEFINITION OF NORMAL WAY-UP AND INVERTED ZONES

Careful appreciation of the regional consistency and significance of the primary structures on the eastern limb of the Tärna synform indicates a normal way-up sequence. In particular, the calcareous phyllites and coarser greywackes of the Lövfjäll Formation lie stratigraphically above the Vojtja quartzite and quartzite conglomerate which are, in turn, younger than the feldspathic greywackes and graphitic phyllites of the Gilliks Formation. Thus, the framework of Kulling's lithostratigraphy is confirmed in this part of the area.

However, consideration of the vertical disposition of these same formations and also the way-up implications of the primary structures indicate inversion of the sedimentary sequence on the western limb and in the core of the late, N-plunging Tärna Synform. The Lövfjäll phyllites etc. occur in the core and the Gilliks greywackes in the outer arc of the NW-plunging, overturned antiform on western Björkfjället (Plate II); the reverse relationship is apparent in the complementary WNW-plunging synforms on eastern Björkfjället and Ruffevare. Thus, there appears to be a major inversion of the stratigraphy (synclinal) relatively early on in the structural sequence, certainly earlier than the formation of these WNW-plunging structures and later Tärna Synform. The approximate limits of the inverted zone and a summary of the primary way-up data, established from field evidence alone, are shown in Fig. 11. The relationship between this inversion of the stratigraphy and the bedding-early penetrative cleavage intersection is discussed later.

CONTACT RELATIONS BETWEEN THE LASTERFJÄLL AND TJOPASI GROUPS IN THE TÄRNA–BJÖRKVATTNET AREA

On the southern limb of the WNW-trending synform near Ruffevare (Plate I), there appears to be progressive thinning of different units, including particularly the Gilliks greywackes, along a fault associated with the fold (see also Kulling 1933). On the northern limb, however, the Vojtja quartzite conglomerate is in direct contact with a layered greenschist unit which to the northwest appears to be situated in the lower part of the Gilliks Formation; the greywackes characteristic of the upper part of the Gilliks Formation are absent. Near the main path across Ruffevare the contact between the layered greenschist unit (Tjopasi Group) and the Vojtja quartzite conglomerate (base of Lasterfjäll Group) is exposed; although there is some quartz veining, the preservation of layering within the greenschist unit adjacent to the contact and the absence of mylonites, brecciation and discordance are inconsistent with extensive sliding at this level. Thus, the outcrop pattern north of Ruffevare appears to be related either to lack of deposition of the greywacke or to deposition, uplift and then erosion of the greywacke prior to deposition of the quartzite conglomerate.

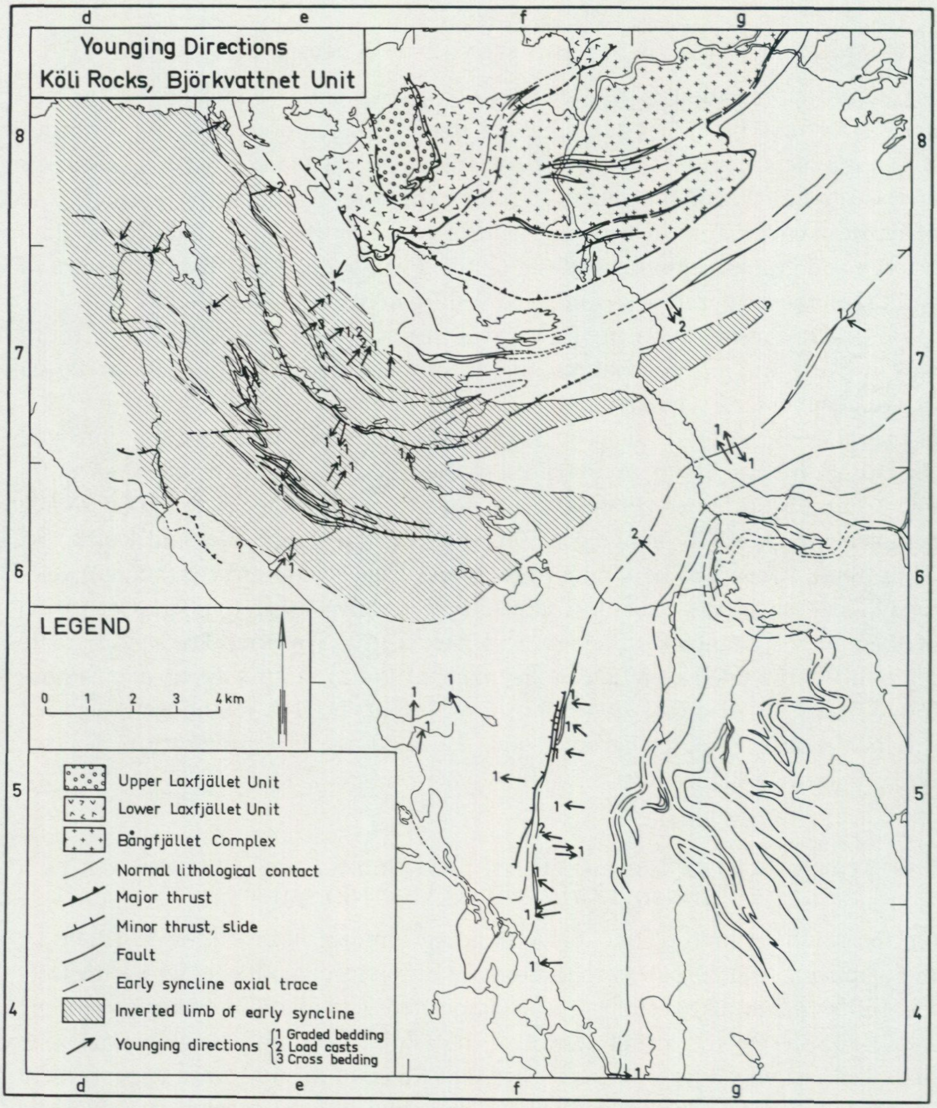


Fig. 11. Summary of younging directions in the Tärna—Björkvattnet area and outline of the inverted zone.

On Lule-Jalketsvardo and near Ruffevare, graphitic limestone, correlated with the Slätdal limestone, lies stratigraphically on top of the quartzite and quartzite conglomerate. In the upper part of Formbäcken, however, there is a transition from Gilliks greywackes to sandy limestone and stratigraphically higher graphitic limestone; the transition beds are composed of several metres of quartz phyllite with thin seams of graphitic, grey phyllite and ribs of sandy limestone. In this section there does not appear to be any break above the Gilliks Formation. It is, therefore, suggested that the absence of Gilliks greywacke immediately north of Ruffevare is more likely due to lack of deposition in this area, and that the Vojtja level represents a significant regression of the sea in late Ordovician times rather than a widespread transgression following significant post-Gilliks Formation tectonic movements.

FACIES INTERPRETATION

Nature and significance of the early Ordovician(?) serpentinite-greenstone association

The low-grade Köli metasediments and volcanics overlie a thick sequence of garnetiferous schists, amphibolites and minor marble horizons (Seve rocks) whose primary character and environment of deposition are obscure due to the extensive grain growth and recrystallization. Petrochemical work on the amphibolites, which are concordant with the main foliation in the interbedded schists, suggests that they are igneous in origin (Stephens 1973), but it is uncertain whether they are sill-like intrusive bodies or extrusive flows. The preservation of pillow structure in the more homogeneous epidote-plagioclase rock suggests that at least this lithology is volcanic in origin.

The greenstone and greenschists of the Seima Formation, which lie above these schists and amphibolites and, except near the base, are rich in chlorite, actinolite and albite, are the lowest, characteristic, Köli rocks. A model for the depositional environment of the Köli has also to take into account the presence of serpentinites within the Seima Formation. The following features are characteristic of the serpentinites in the Tärna—Björkvattnet area:

1. There are few primary minerals preserved from the original ultramafic rock; they are predominantly antigorite-magnesite rocks.
2. They vary in size from a few metres to several hundreds of metres across; the smaller bodies are talcose.
3. They are pod-like in form and occur at different stratigraphic levels within the Seima Formation.
4. The smaller lenses are strongly schistose and the schistosity is concordant with the external foliation.
5. There is no high temperature metamorphic aureole.
6. They are associated with deformed and metamorphosed sediments and volcanics, in particular, occasionally pillowed greenstone.
7. The greenstones are, in part, similar to abyssal tholeiite indicating eruption either at an active spreading site or during an early stage of island arc development.

Recently much interest has centred on the tectonic implications of serpentinites in orogenic belts (see, for example, Bird et al. 1971; Moores 1973) and, in particular, the presence of the ophiolite suite (Conference participants 1972). Although the serpentinites in the Seima Formation are associated with pillowed greenstone and occasional metagabbro lenses, occurrence of serpentinite at different levels within the formation and the absence of the typical ophiolite stratigraphy of ultramafic rock at the base passing up to mafic dykes, pillowed basic lava and deep-water marine sediments (Moores 1973) indicate that they do not strictly belong to an ophiolite suite. Furthermore, unlike the ophiolite complexes of, for example, Greece (Moores 1969), Newfoundland (Dewey and Bird 1971) and Papua-New Guinea (Davies 1968) the serpentinites and greenstones in the Tärna—Björkvattnet area do not form an intrinsically allochthonous sheet, i.e. they are not an isolated tectonic slab within the surrounding volcano-sedimentary sequence (the serpentinites and greenstones do, however, conform to the regionally flat-lying structure of the whole Seve-Köli Nappe Complex). This is supported by the transitional nature of the contact between the Seima Formation (and its stratigraphic equivalent the Forsbäck Formation) and the overlying Gilliks Formation. The serpentinites resemble the intrusive or diapiric bodies described by Chidester and Cady (1972).

The relationship between the foliation both within and around the talc schist bodies suggests that the ultramafics were emplaced in their present position early in the metamorphism-deformation history. Indeed the occurrence of talcose rock fragments within a basic agglomerate horizon near the top of the Seima Formation indicates that emplacement of at least some of the serpentinites was completed within the early Ordovician(?) (Kulling, *in* Magnusson et al. 1960: Yochelson 1963), even before the Gilliks Formation was deposited.

It is suggested that there was an important phase of intrusion of ultramafic bodies during the early Ordovician(?), and that these bodies were emplaced, in the Tärna—Björkvattnet area, into a volcanic sequence of similar age made up, in part, of greenstones with abyssal tholeiite characteristics (Seima Formation). The volcanic and ultramafic rocks overlie a pile of metasediments and basic igneous rocks (Brakko schists belonging to the Seve), the sedimentary part of which, it is tentatively suggested, was deposited in a position marginal to (continental rise/slope?) the Baltoscandian craton. The chemistry of the Seima volcanics (Stephens 1973) and the transitional stratigraphic relationships between the Brakko schists, Forsbäck/Seima and Gilliks Formations indicate that this early Ordovician(?) phase of volcanism and ultramafic intrusion may be interpreted in relation to either the initial expression of island arc development (initial-stage submarine arc of Miyashiro 1974, 1975), or back-arc spreading (Karig 1971) associated with the formation of a relatively small ocean basin close to the margin of the Baltoscandian craton.

The following section is concerned with the remaining Ordovician and Silurian sedimentary history in the Tärna—Björkvattnet area. Post-Seima sedimentation and volcanism appears to have been influenced by both a continental source and the island arc complex itself.

Remaining Ordovician and Silurian sedimentary history

The rapid alternation of coarser and finer grained units, the internal grading of some units, which also contain distinctively sharp bases, and the general plane-parallel nature of the bedding in the Gilliks Formation greywackes are suggestive that these beds were laid down by turbidity currents, and that the sediments were originally turbidites (Bouma 1964). The ideal turbidite sequence defined by Bouma (1962) consisting of five separate divisions is not observed. The graded units appear to be the "middle-absent" type (Walker 1965) which grade from the base upward into the pelitic division without any intermediate lamination or rippling; the presence of load casts and the local development of slump structure conform to the above interpretation. The common occurrence of greywacke beds within the upper part of the Gilliks Formation greater than 30—40 cm and up to 1—1.5 m thick with strongly subordinate pelite intercalations, and the coarseness of the different types of greywacke suggest that they are of proximal type (Walker 1967). Other features of proximal turbidites listed by Walker, such as amalgamation of individual sandstone beds and the common occurrence of scour marks and other erosional features, have not been observed.

The most important variation within the total outcrop of the coarser clastic rocks of the Gilliks Formation appears to be in the feldspar content. The quartzose greywackes of the Formliden area are virtually devoid of feldspar, whilst the greywackes on the eastern limb of the Tärna Synform contain greater than 10 % total feldspar and values up to 25 % occur (the greater feldspar content of these eastern outcrops is even reflected in the succeeding Vojtja quartzite). The polymict conglomerate facies, which is partially composed of granitic pebbles, is notably confined to thin layers within the lower part of the Gilliks Formation, again on the eastern limb of the Tärna Synform, suggesting that there is a common factor between the development of the polymict conglomerate facies and the feldspathic greywackes. Although the source of material for the Gilliks clastics was continental and "granitic", there are considerable differences in the compositional and textural maturity of these sediments, reflecting variable rates of erosion and deposition i.e. variable source stability. The pyroclastic volcanic material, which is prominent in the lower part of the Gilliks Formation and is largely mafic, is thought to have been derived from the island arc complex.

The succeeding lithostratigraphic unit, the Vojtja quartzite/quartzite conglomerate, displays contrasting sedimentological features to the Gilliks Formation; this unit is characterized by generally massive or faint wavy bedding structure with occasional cross bedding and, in the outcrops on western Björkfjället, where no quartzite conglomerate facies is present, lenticular and possible flaser bedding. The discontinuous nature of the Vojtja rocks and the sedimentary structures within them indicate shallow water deposition. Where conglomerate deposition has occurred, uniform turbulence of the water is inferred (beach environment?). Alternation of slack and turbulent water conditions is a prerequisite for the deposition of fine-grained mud within a predominantly quartzitic sequence and the formation of structures such as lenticular and flaser bedding (Reineck and Wunderlich 1968). Such variation in the water turbulence conditions would have been concentrated in either the subtidal or intertidal zones — tidal action being the operative mechanism. The occurrence of these structures on western Björkfjället, where the conglomerate facies is not represented, is considered to support the above interpretation. The stratigraphic relationships between the Vojtja and Gilliks Formations, discussed earlier, suggest that the Vojtja rocks represent a widespread regression of the sea after the deposition of the Gilliks rocks. The occurrence of coral-bearing limestone in the overlying Slätdal Formation (Ashgillian) confirms the relatively shallow water conditions during Upper Ordovician times.

The return to relatively deeper water conditions of sedimentation, heralded by the structureless graphitic phyllites of the Broken Formation (Middle and Upper Llandovery in age), was completed by the time of deposition of the calcareous phyllites and coarser grained greywackes and phyllitic quartzites of the Lövfjäll Formation. The comments made earlier for the Gilliks greywackes apply equally well to the lithologies at this higher level, and a return to turbidite facies is inferred. However, the source of the Lövfjäll sediments appears to have been volcano-sedimentary rather than continental, "granitic". The contrast between the calcareous phyllites and the coarser rock types is marked by an increase in bed thickness (as well as grain size) and the strongly subordinate role of pelite layers in the latter. It is considered possible that the greywackes and phyllitic quartzites were deposited under shallower water conditions (proximal type) than the finely layered phyllites (distal type).

The rocks of the Lövfjäll Formation are the youngest exposed within the Björkvattnet Unit. Climactic polyphase folding, metamorphism and subsequent thrusting occurred during the Silurian (Roberts 1971a), probably within the Wenlock (Wilson et al. 1973; Gee and Wilson 1974).

CONCLUSIONS

The stratigraphy of the Björkvattnet unit is summarized in Fig. 12, and compared with the scheme of Kulling (1933—1969) and the regional group terminology of Zachrisson (1969). The data is arranged in four sections chosen to illustrate the facies variations which have been described and interpreted earlier. Sections 1 and 2 are situated on the inverted limb of the early, overturned, major syncline which has been defined on the basis of a major repetition in the stratigraphy; this particularly accounts for the reverse distribution of the rocks of the Lövfjäll and Gilliks Formations in the later, WNW-plunging folds on Björkfjället. Sections 3 and 4 are situated on the normal way-up limb of this early structure.

The following comments refer to the application of the lithostratigraphic terminology of Kulling (1933—1969) in the Tärna—Björkvattnet area:

1. The Lövfjäll, Broken (Middle and Upper Llandovery in age), Slätdal (Ashgillian age), Vojtja and Gilliks "Series" proved useful lithostratigraphic units of formation status and were employed, as such, in the present stratigraphic scheme.
2. The Lövfjäll Formation consists of both a finer calcareous phyllite and a coarser greywacke/phyllitic quartzite facies. The latter appears to be developed diachronously within the finer calcareous phyllites. This is similar to the relationship between coarser and finer lithofacies at this stratigraphic level in Kulling's (1933) type-area (see Zachrisson 1969). However, exact correlation of the greywackes and phyllitic quartzites of the Tärna—Björkvattnet area with the coarser facies further south is not attempted here.
3. The Seima "Series", defined by Kulling (1958) for the volcanics exposed around Seimajaure (now called Säjman) in the southeast of the area, has been renamed the Seima Formation according to the procedures outlined in Harland et al. (1972). Serpentinites occur at different stratigraphic levels within the Seima Formation and serpentinite conglomerate has not been recognised; Kulling's Ro "Series" has, therefore, not been used as a mappable unit within the Tärna—Björkvattnet area.
4. One new formation — the Forsbäck Formation — has been recognised in this study. This is composed of various garnetiferous quartz and grey phyllites which appear to be the equivalent in the northeastern part of the area to the Seima volcanics.
5. The informal terms Brakko schists and Tärna schists have been introduced for the undifferentiated higher grade rocks beneath the Forsbäck and Seima Formations in the east of the area, and the lithologies which rim the Bångfjället Complex near Tärnaby respectively. The former are probably equivalent to the Svartsjöbäcken Schists in the Marsfjällen area (Trouw 1973), whilst the latter include the "quartzite-garnet-mica-schist series" of Kieft (1952).

From the facies interpretation of the stratigraphy, the following model is suggested for the sedimentation history of the Björkvattnet Unit:

1. Deposition of a thick pile of sedimentary material associated with basic igneous activity of uncertain age and origin marginal to (continental rise/slope?) the Baltoscandian craton; these rocks now constitute the Brakko

schists, form part of the Seve sequence and the sediments, at least, are probably pre-early Ordovician in age. The primary characters of the schists and amphibolites are obscure due to the degree of metamorphism.

2. Intrusion of ultramafic bodies during the early Ordovician into a basic volcanic sequence of similar age composed, in part, of greenstones with abyssal tholeiite characteristics (Seima Formation). This igneous activity is thought to be related to the initial expression of island arc development or the opening of a back-arc basin between the Baltoscandian craton and the island arc itself.
3. Deposition of predominantly turbidite sequences during the remainder of the Ordovician and Silurian, interrupted only during the late Ordovician by a significant regression of the sea which resulted in the deposition of shallow water quartzite conglomerate, cross-bedded quartzite and limestone. The turbidites within the Ordovician Gilliks Formation appear to have been deposited relatively near to a continental, "granitic" source, but show varied facies relationships related primarily to the stability of the particular source area. The turbidites within the Silurian Lövfjäll Formation were derived from a volcano-sedimentary source and display facies variation possibly controlled by deposition under different water depth conditions.
4. Climactic polyphase folding, metamorphism and thrusting of the whole sequence occurred soon after deposition of the post-Upper Llandovery Lövfjäll Formation, probably within the Wenlock.

Consideration of the spatial distribution of the different facies within the Ordovician/Silurian Köli rocks of the Björkvattnet Unit (Figs. 12 and 13) suggests the following:

1. Deposition northeast of Ruffevare (section 1, Fig. 12) is thought to have occurred close to a high within the depositional basin. This is suggested from the lack of deposition of the Gilliks turbidites and the development of a clean, beach-type(?) conglomerate facies in the late Ordovician; the Silurian turbidites are conspicuously the shallower water facies type in this area.
2. The Formbäcken section (section 2, Fig. 12) is thought to represent deposition offshore from section 1, sedimentation being influenced by a relatively stable source supplying little feldspar debris. The increased maturity of the Gilliks Formation greywackes, the continuity of sequence at the Slätdal—Vojtja level, the thick development of Slätdal limestone and the later deposition of possibly deeper water (distal) turbidites (Lövfjäll Formation) are the important characteristics of this section.
3. Sedimentation in the Vojtja/Lule-Jalketsvardo area (section 4, Fig. 12) was influenced by its much closer proximity to a more unstable, continental, "granitic" source which provided abundant feldspar detritus. This is indicated

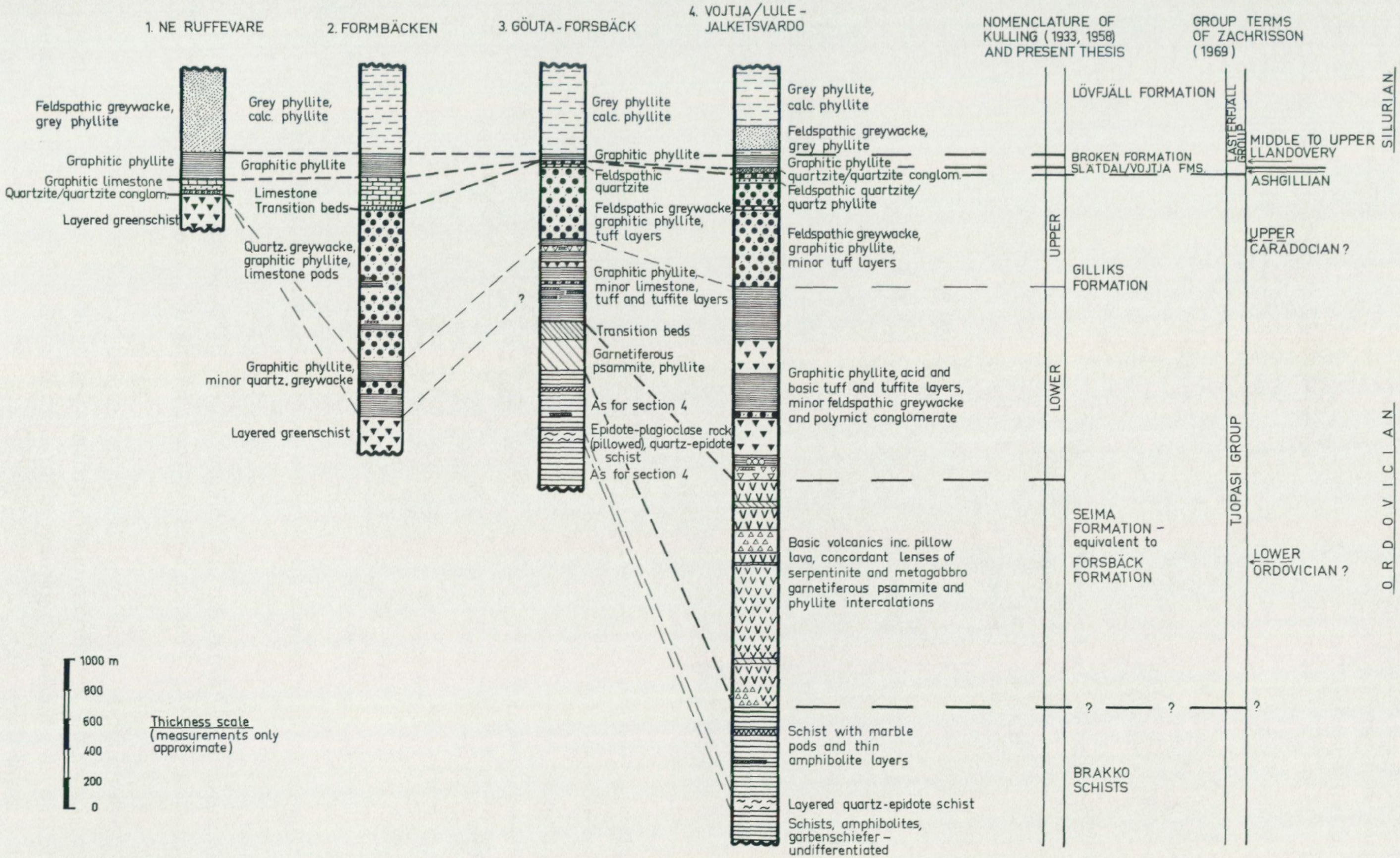


Fig. 12. Summary of the stratigraphy within the Björkvattnet Unit.

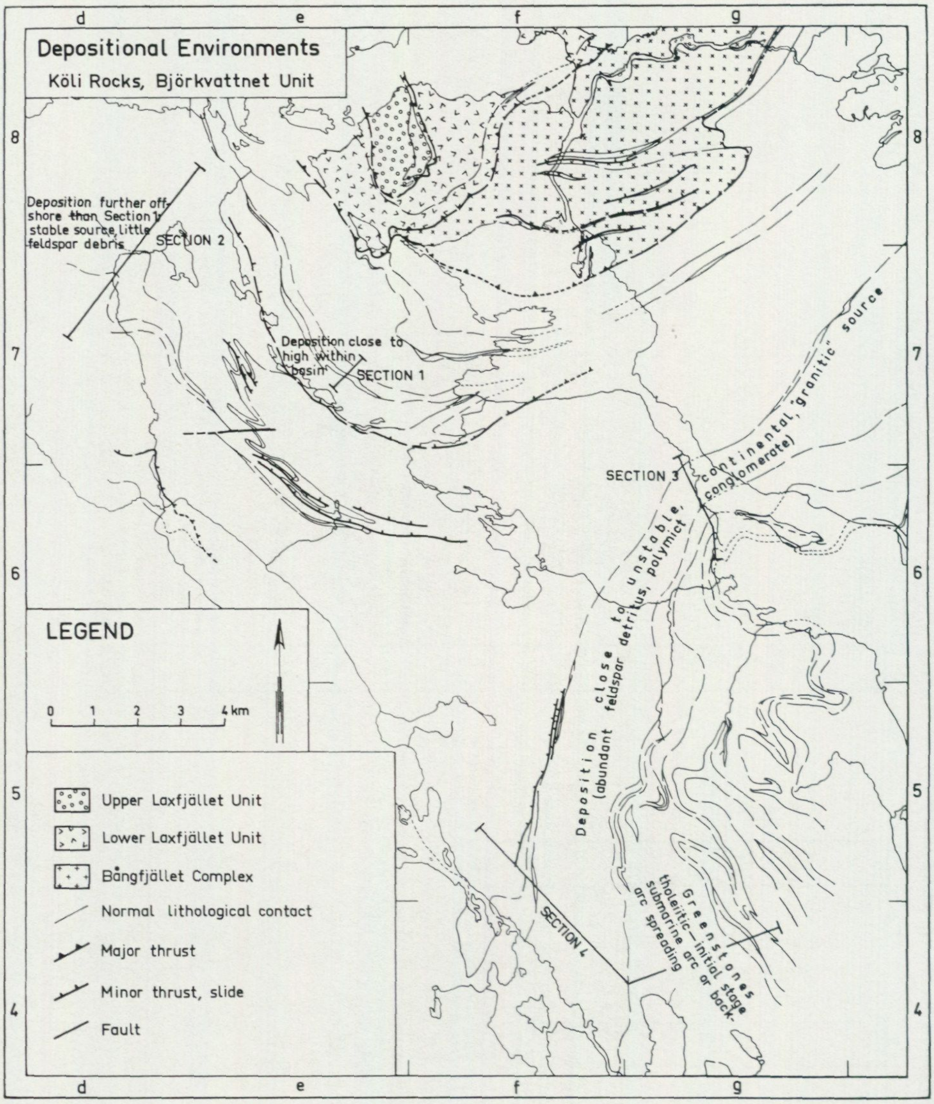


Fig. 13. Depositional environments within the low-grade rocks of the Björkvattnet Unit.

from the compositional and textural immaturity of the sediments in the Gilliks Formation, and the presence above of more feldspathic Vojtja quartzite.

The predominantly mafic pyroclastic and mixed sedimentary-volcanic rocks interbedded within the Ordovician/Silurian clastics are thought to have been derived from an island arc complex on the opposite side of the basin to the continental, granitic basement.

It is suggested that the above is consistent with the mounting evidence for initial contraction of a proto-Atlantic ocean (Wilson 1966) and development of consuming plate margins and possibly back-arc oceanic basins in latest Cambrian/ early Ordovician times (Dewey 1969, 1971; Bird et al. 1971; Gale and Roberts 1972). On a plate-tectonic model, the Wenlock deformation, metamorphism and thrusting resulted from the eventual closure of the proto-Atlantic ocean (McKerrow and Ziegler 1972) and the subsequent collision of the Baltoscandian and Greenlandian margins in this part of the Caledonides.

STRUCTURAL EVOLUTION OF THE BJÖRKVATTNET UNIT

INTRODUCTION

Zachrisson (1969) summarized the geology of a 150 km strip of northern Jämtland—southern Västerbotten; this included the first comprehensive account of the major and minor structures of this part of the Swedish Caledonides. Zachrisson emphasized the influence of polyphase deformation, but only in the low-grade (Köli) rocks where at least two important phases of folding were recognised. The first phase, referred to as F1, produced "isoclinal, generally minor folds and a strong schistosity subparallel to bedding" with a NW—SE to E—W axial orientation. The second phase (F2) produced more open structures with a generally NE—SW or NNE—SSW orientation (i.e. parallel to the trend of the orogenic belt as a whole) and steeply dipping axial surfaces. According to Zachrisson, this later phase deformed the main schistosity and associated low angle thrusts and also accounted for the formation of four major folds which dominate the outcrop pattern of the whole area (Fig. 14); these are from west to east, the Riksgränsen Antiform, the Western Synform (see also Nilsson 1964; Zachrisson 1964), the Fjällfjäll Antiform and the Eastern Synform. It is supposedly within the cores of the late antiforms that autochthonous(?) basement (Børgefjell Window) or allochthonous Lower thrust rocks (Fjällfjäll and Hehtajaure Windows) are found (Figs. 1, 14). Zachrisson also noted chevron-type folds (F3) with flat-lying axial surfaces exclusively within the Western Synform; these folds deform the F2 axial surface cleavage and are consistently overturned down-dip.

To the west and northwest the rocks have been transported in a series of nappe structures which were emplaced early in the structural/metamorphic history (Rutland and Nicholson 1965; Nicholson and Rutland 1969). This contrasts sharply with the intense post-metamorphic thrusting observed at the leading edges of the different allochthonous units, particularly the more easterly ones (Kulling, *in* Gavelin and Kulling 1955). Recently, however, in the Björkvattnet—Virisen area, Gee (unpublished SGU material reported in Zachrisson 1969) has

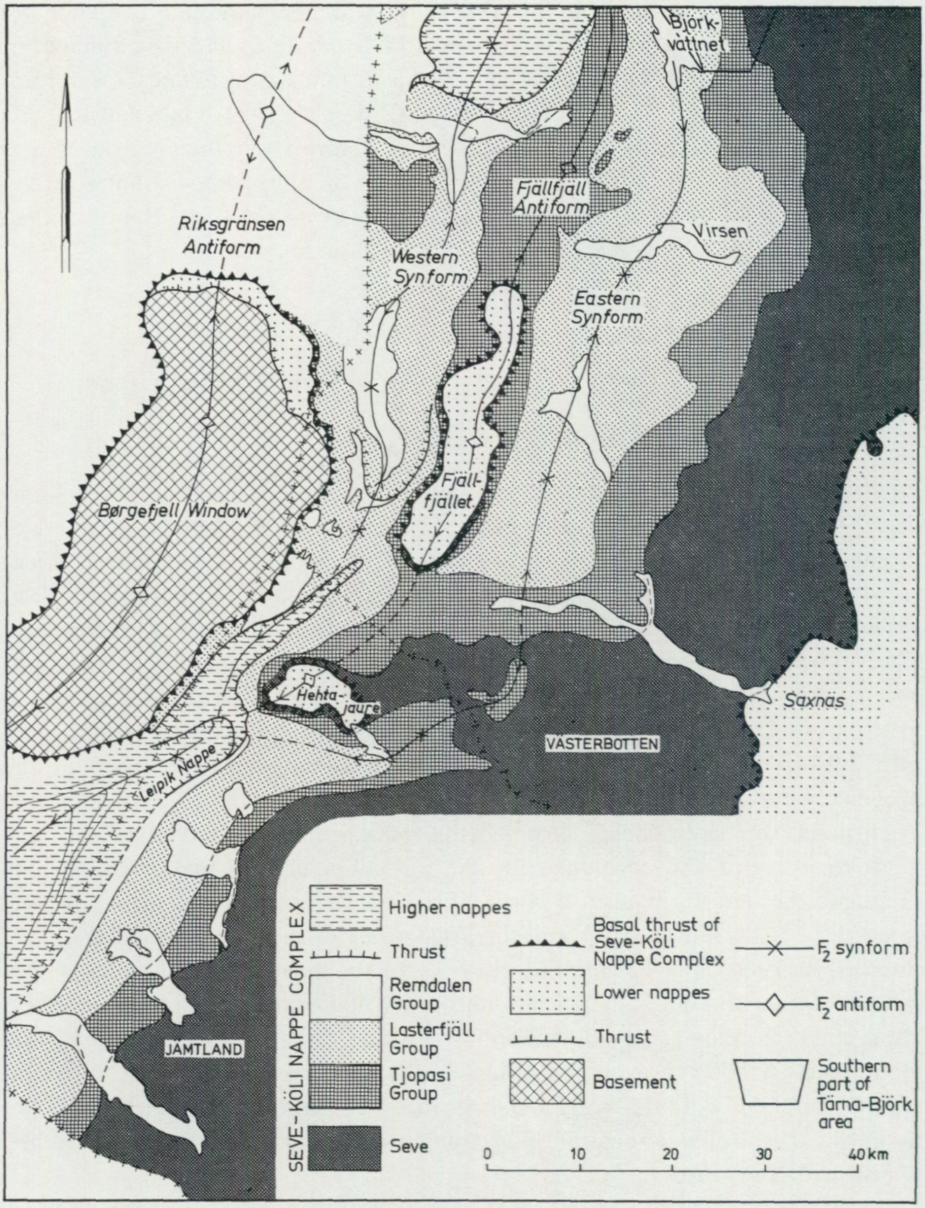


Fig. 14. Summary of major folds in northern Jämtland – southern Västerbotten according to Zachrisson (1969).

confirmed the presence of an early, westwards closing, recumbent syncline near Lake Väsken with the Viris quartzite preserved in its core (see also Kulling 1933). The main schistosity (Zachrisson's S1) cuts across the axial surface of this major fold and associated minor folds, and appears to have formed at a later stage; Zachrisson (1969) referred to these folds as pre-F1. Thus, there is evidence for the presence of early, large-scale, recumbent folds within the Seve-Köli Nappe Complex.

Description of the major and minor structures within the Björkvattnet Unit is presented here in the form of a structural analysis of representative subareas chosen on the grounds of relative structural and metamorphic homogeneity. The mylonitic structures within the Tärna schists, immediately adjacent to the Bångfjället Complex, are described and discussed later. The terminology of Fleuty (1964a) is used to describe particular fold characteristics such as tightness and fold attitude, whilst the term slide is used according to the definition of Fleuty (1964b) who states that is is "a fault formed in close connection with folding, which is broadly conformable with a major geometric feature (either fold limb or axial surface) of the structure, and which is accompanied by thinning and/or excision of members of the rock-succession affected by the folding".

STRUCTURAL ELEMENTS

All the planar and linear elements which have been measured are defined in Table 2. In the low-grade rocks the early, regional cleavage (S_{cl}) is generally parallel or subparallel to bedding (S_b); exceptions to this relationship occur in the hinge zones of certain minor folds where S_{cl} forms an axial surface cleavage to the fold. The presence of an earlier, pervasive surface appears to be a prerequisite for the development of the secondary crenulation cleavages (Rickard 1961). There are at least two separate phases of crenulation cleavage development associated with the deformation of S_{cl} and these are referred to as S^1_{cc}, S^2_{cc} etc. These crenulation cleavages are locally very penetrative and, particularly in the finer phyllites, S_{cl} may be transposed into the later foliation.

In the high-grade rocks bedding is only occasionally preserved. The dominant planar structure is a pervasive schistosity (S_s) which is particularly prominent in the garnetiferous schists but is not so obvious in the amphibolites where an associated linear structure is present. Axial surfaces of minor folds and tight crenulations which deforms S_s were also measured. Again at least two separate phases are present, but only in the later phase is there any indication of secondary cleavage development similar to the crenulation cleavages of the lower grade rocks.

The lineations in both the low- and high-grade rocks have been divided into two groups. Group 1 includes the S_b—S_{cl} intersection lineations, abundant minor folds and various crenulation lineations in the low-grade rocks and a pronounced

TABLE 2. Summary of structural elements in the Björkvattnet Unit.

		Low-grade (Köli) rocks	Symbol	High-grade (Seve) rocks	Symbol
Planar elements		1. Bedding	S_b	1. Bedding	S_b
		2. Regional phyllitic cleavage and axial surfaces of associated minor folds	S_{cl}	2. Regional schistosity and axial surfaces of associated minor folds	S_s
		2. Crenulation cleavages later than and locally transposing S_{cl}, and axial surfaces of associated minor folds	S^1_{cc}, S^2_{cc} etc.	3. Axial surfaces of minor folds which deform S_s (includes late crenulation cleavage)	-
Linear elements	GROUP 1	1. S_{cl} on S_b and S_b on S_{cl} intersection lineations: the latter is a compositional lineation pitching down the S_{cl} surfaces 2. Minor fold axes and crenulation lineations including mullions in NW of area		1. Quartz rodding 2. Minor fold axes including late crenulation of S_s	
	GROUP 2	1. Penetrative mica and quartz lineation on S_{cl} surfaces particularly in the quartzites 2. Long axes of deformed pebbles (Vojtja conglomerate), agglomerate fragments and amygdales/phenocrysts (Seima Formation)		1. Mineral lineations: (a) Penetrative mica and slender hornblende lineation in S_s (b) Elongation of garnet, biotite and hornblende porphyroblasts in S_s 2. Long axes of deformed feldspar phenocrysts in the amphibolites	

quartz rodding lineation and minor fold axes in the high-grade rocks. The bulk of the minor folds in both the low- and high-grade rocks deform S_{cl} and S_s respectively, although there are occasional folds with S_{cl} or S_s as an axial surface structure. The quartz rodding lineation is especially well developed in the garnetiferous schists where it lies in S_s; it plunges down the axes of minor folds related to S_s, and the rodding lineation is deformed with S_s in subsequent phases of folding.

The group 2 lineations include the various mineral lineations contained in S_{cl} and S_s and the long axes of deformed ellipsoidal objects. In the low-grade rocks the mineral lineation is particularly prominent in the Vojtja quartzite where it is defined by a preferred dimensional orientation of quartz and mica on the S_{cl} surfaces. The long axes of deformed pebbles in the associated Vojtja conglomerate parallel this mineral lineation. The present shape of the pebbles is, in general, related to the strong deformation associated with the formation of S_{cl} (Stephens 1975). The mineral lineations in the high-grade rocks include both a penetrative preferred orientation of undeformed muscovite, biotite and slender hornblende prisms on the S_s surfaces as well as a dimensional orientation of coarser garnet, biotite and hornblende porphyroblasts. The coarse biotite and hornblende porphyroblasts show evidence of deformation in S_s. The biotites are usually lenticular in shape in sections normal to S_s, whilst the hornblendes, which often lie on S_s in an apparent rosette arrangement, are bent and often boudinaged and show undulose extinction. The lineation defined by these coarser porphyroblasts is parallel to the penetrative mica and slender hornblende lineation. The other type of Group 2 lineation in the high-grade rocks is the elongation of recrystallized feldspar phenocrysts in the amphibolites.

ANALYSIS OF REPRESENTATIVE SUBAREAS

SUBAREA 1: BJÖRKFJÄLLET – RUFFEVARE – GÖUTAVARDO

Major structures

The structure of western Björkfjället (7e) is dominated by an overturned antiform which is followed to the southeast by the complementary synformal complex of eastern Björkfjället (Plates I—III). On a major scale these Björkfjället folds deform S_{cl} as well as S_b (Plate III, 1a and b) and it is inferred that they plunge at low angles predominantly towards the northwest. The sense of overturn of these folds is SW to S, although locally more upright folds are present in the complex closure zones of both the antiform and its complementary synform. Minor slides strike parallel to the axial surface traces of the Björkfjället folds and account for apparent discontinuities in the sequence (Plate II). As noted above, the calcareous phyllites belonging to the Silurian Lövfjäll Formation are disposed in the core of the West Björkfjället Antiform and the greywackes of the upper part of the Gilliks Formation (Ordovician) rim this structure; the reverse relationship exists in the complementary East Björkfjället Synform. Thus, these folds were superimposed on an already inverted sequence.

On the basis of way-up evidence from sedimentary structures, the extent of the inverted limb of this early, recumbent, major fold has been indicated on Fig. 11. It proved impossible to establish over the whole area whether or not the succession consistently faced upwards on the regional cleavage (S_{cl}) and thus whether the early isoclinal folding which accounts for the major inversion was synchronous with or occurred prior to the formation of S_{cl}. However, in the Sandnäsbäcken stream section (6e) limited structural data indicate the following (Fig. 15):

1. In the lowest part of the stream the beds are consistently right way-up but upstream they are inverted and appear to remain so until the east Björkfjället ridge; consequently, a NW-trending synclinal trace is defined in the Sandnäsbäcken stream section.
2. There is a general increase in the dip of both S_b and S_{cl} up the stream section.
3. The S_b-S_{cl} relationship remains *constant* across the synclinal trace viz. S_{cl} is slightly steeper than S_b (Fig. 15).

The above evidence from Sandnäsbäcken suggests that the major synclinal repetition of the stratigraphy was established prior to the formation of S_{cl}.

A similar structure is found northeast of Björkfjället, around Ruffevare (7e). In this area, there is an important repetition of the stratigraphy across the main ridge where a layered greenschist belonging to the Gilliks Formation forms the core of the repeated sequence (Plate I). This unit closes eastwards so that on the opposite shore of Göuta it is absent. The zones east of Ruffevare and on the

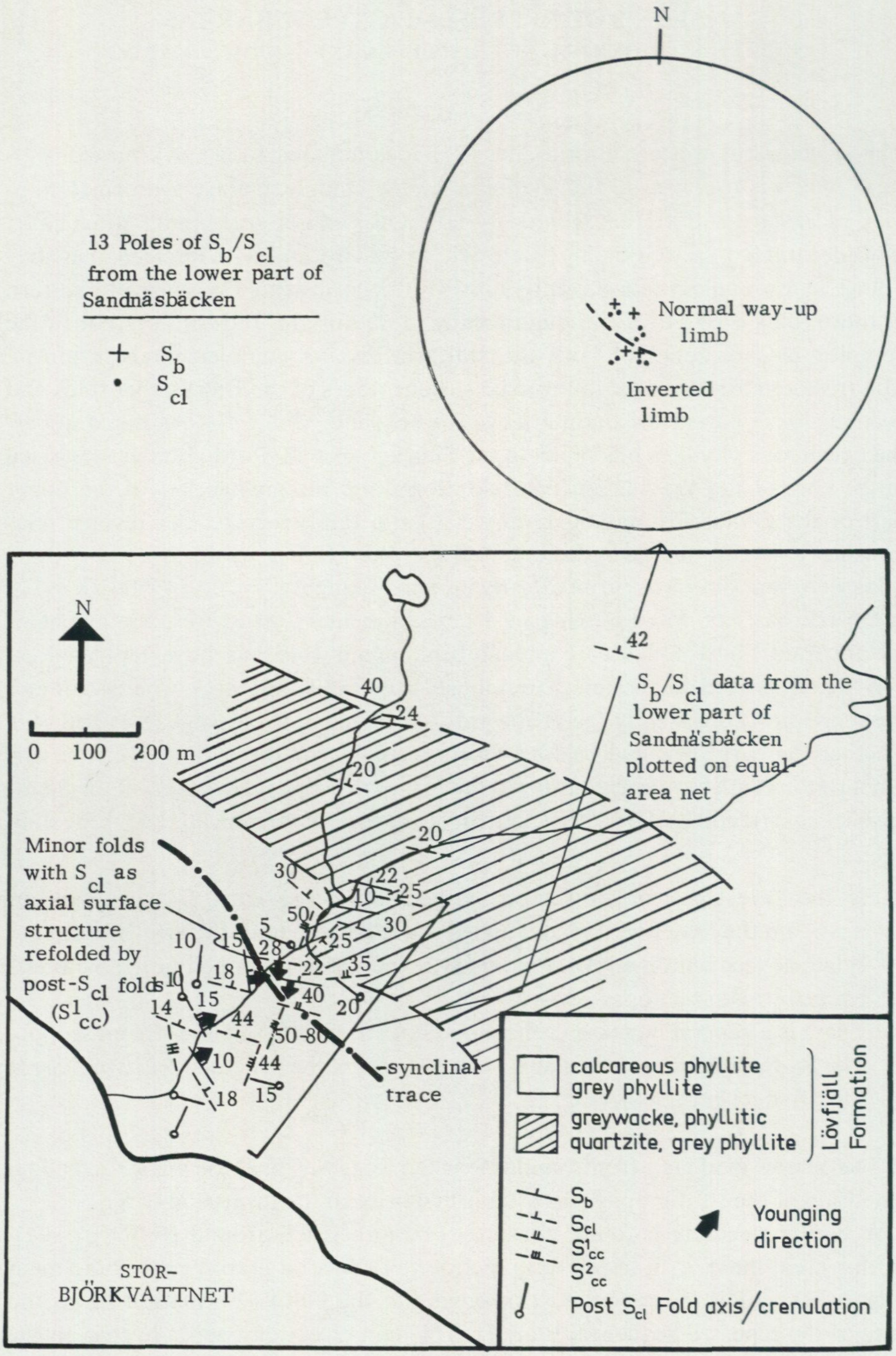


Fig. 15. Structural data from Sandnäsbäcken.

opposite shore of Göuta are dominated by minor structures congruent to a W-plunging, post-S_{cl} synform, hereafter referred to as the Ruffevare Synform (Plate III). Once again older rocks are disposed in the core of the synform which is probably overturned southwards (the dips on the northern limb are variable and the sheet dip difficult to determine) and clearly complements the post-S_{cl} Björkfjället folds.

Locally important slides occur on both limbs of the Ruffevare Synform. On the southern limb, where there is progressive excision of different stratigraphic units (Plate I and Kulling 1933), the discontinuity is subparallel to the limb of the synform and is marked topographically by a system of lakes in a depression between the main Ruffevare ridge and the slopes leading up to Björkfjället. On the northern limb the graphitic phyllites belonging to the Broken Formation thin out towards the northwest, and east of Batsevardo (8e) the contact between the greywacke facies of the Lövfjäll Formation and the layered greenschist of the Gilliks Formation is marked by a few metres of lenticular and quartz veined, pyritiferous, grey and graphitic phyllite.

On the northern limb of the Ruffevare Synform the penetrative crenulation cleavage associated with this fold (S^1_{cc}) dips north at a shallower angle than S_b/S_{cl}. To the north the stratigraphy is again descended near Granås (7f) where an attenuated sequence of graphitic phyllite, volcanics and subordinate Gilliks greywacke occurs (Plate I). However, the relationship between S_b/S_{cl} and S^1_{cc} remains constant. This suggests that the synclinal repetition between Ruffevare and Granås is earlier than the formation of the Ruffevare Synform and that this synform is situated on the inverted limb of the earlier fold.

The swing in strike of both S_{cl} and S_b from NW—SE on west Björkfjället to WNW—ESE on east Björkfjället continues eastwards onto Gakerevardo and Göutavardo where E—W and NE—SW regional strike is observed. This open, NNE-plunging synformal structure (the Tärna Synform) also deforms the axial traces of the Björkfjället and Ruffevare folds and produces the characteristic, gently refolded structure of subarea 1. The plot of S_b/S_{cl} poles from the whole subarea (Plate III, 1c) indicates the two post-S_{cl} axial trends; the NE—SW π-circle defines the earlier NW axial trend and the essentially S-overturned character of this folding (there is, however, a subordinate maximum indicating SW dips related particularly to more upright folding northeast of Ruffevare), whilst the spread of poles related to more steeply dipping, northeast attitudes for S_b/S_{cl} reflects the later, open folding by the Tärna Synform.

Considering the whole subarea, then, the early inversion of the stratigraphy accounts for the outcrop pattern in which rocks of the Tjopasi Group occur both in the core as well as around the outer arc of the later Ruffevare—Björkfjället Synform Complex (see Plate III). Kieft (1952) defined the axial trace of an "overturned syncline" predominantly on and north of Göutavardo; this structure is equivalent to the northeast extension of the post-S_{cl} Ruffevare—

Björkfjället Synform Complex. Both the early syncline axial surface trace and the Ruffevare—Björkfjället Synform Complex are deformed in the latest major structure — the open, NNE-plunging Tärna Synform.

Minor structures

The earliest minor structures include:

1. occasional S_b-S_{cl} intersection lineations and syn-S_{cl} minor fold axes and
2. the penetrative mineral and deformed pebble lineations (Fig. 16).

These lineations are closely related to the formation of S_{cl}. The minor folds (Fig. 17), particularly prominent in the Sågbäcken stream section (7f) and along Route E79 near Tärnafors (7g) and Sundsgården (7g), are tight to isoclinal in shape and S_{cl} forms an axial surface structure. The penetrative mineral lineation in the Vojtja quartzite is thought to define the direction of maximum elongation in S_{cl}; this is substantiated by the deformation in the Vojtja quartzite conglomerate (Stephens 1975). The pebbles are strongly deformed with prolate ellipsoids predominating and the statistical pebble lineation, which is considered to define the maximum elongation in S_{cl}, is parallel to the mineral lineation direction in the associated quartzite. The orientation of both lineation types 1 and 2 is, however, very similar (Plate III, 1d). On the steep eastern limb of the late Tärna synform (domain IV, Plate III, 1d) the plunge of these lineations increases to 65—70°, elsewhere (domains I to III) the plunge is generally up to 30° in a NW to W direction.

The post-S_{cl} minor structures reflect the major, post-cleavage, refolding relationship described above. The most abundant structures are minor folds and subparallel crenulation lineations which are congruent to the Ruffevare—Björkfjället Synform Complex and which deform S_{cl}, related minor folds and the S_b—S_{cl} intersection lineations. The folds are close to tight and often strongly asymmetric in shape (Figs. 17, 18). Commonly associated with these structures is a crenulation cleavage ($S^1{}_{cc}$) which, particularly in the phyllites in the vicinity of the major fold hinge zones, is highly penetrative and transposes S_b/S_{cl}. Although there is some evidence for the development of zones of contact strain (Ramberg 1961) in pelite adjacent to psammite layers, where $S^1{}_{cc}$ converges downwards towards the synformal areas, $S^1{}_{cc}$ is generally subparallel to the axial surfaces of the associated folds. The orientation of $S^1{}_{cc}$ reflects the general SW—S sense of overturn of the Björkfjället folds and their deformation by the later Tärna Synform (Plate III, 1e).

Transecting the Björkfjället and Ruffevare folds are subordinate minor structures, including a weakly developed $S^2{}_{cc}$ crenulation cleavage and lineation. Minor folds associated with this late phase of deformation are open (Fig. 17); these minor structures have been tentatively correlated with the formation of the

Fig. 16. Prolate-shaped quartzite pebbles in the Vojtja conglomerate (Ruffevare, 7e).

late Tärna Synform. S^2_{cc} on the western limb of the synform decreases in dip away from the axial zone and progressively changes strike from N—S to NE and ENE (Plate III, 1f).

For the purpose of demonstrating the reorientation of earlier lineations in the later Tärna Synform, the lineation data have been divided into that related to S_{cl}, the Ruffevare—Björkfjället Synform Complex and the Tärna Synform (summarized in Plate III, 1d—f); the first two sets of data have been further separated into partial subareas or domains within the later Tärna Synform so as to enhance their structural homogeneity. The following points are apparent (Table 3):

1. The vector means of the lineations associated with the Ruffevare—Björkfjället Synform Complex in domains I and II correspond closely to the respective fold axes deduced from deformed S_b/S_{cl} data (Plate III, 1a, b).
2. The two sets of earlier linear structures are approximately coaxial (angle between vector means $<20°$) on the gently dipping western limb of the Tärna Synform.
3. The distributions of the earlier sets of lineation data about the Tärna Synform axis diverge as the eastern limb of the synform is approached. This suggests that their pre-Tärna Synform orientations were not coaxial.
4. The angles between the latest phase and the two earlier phases change systematically as the earlier lineations are traced from the western limb (angles $= 46—51°$) via the hinge zone (angles $= 85—87°$) to the eastern limb (angles $= 77—108°$) of the Tärna Synform.

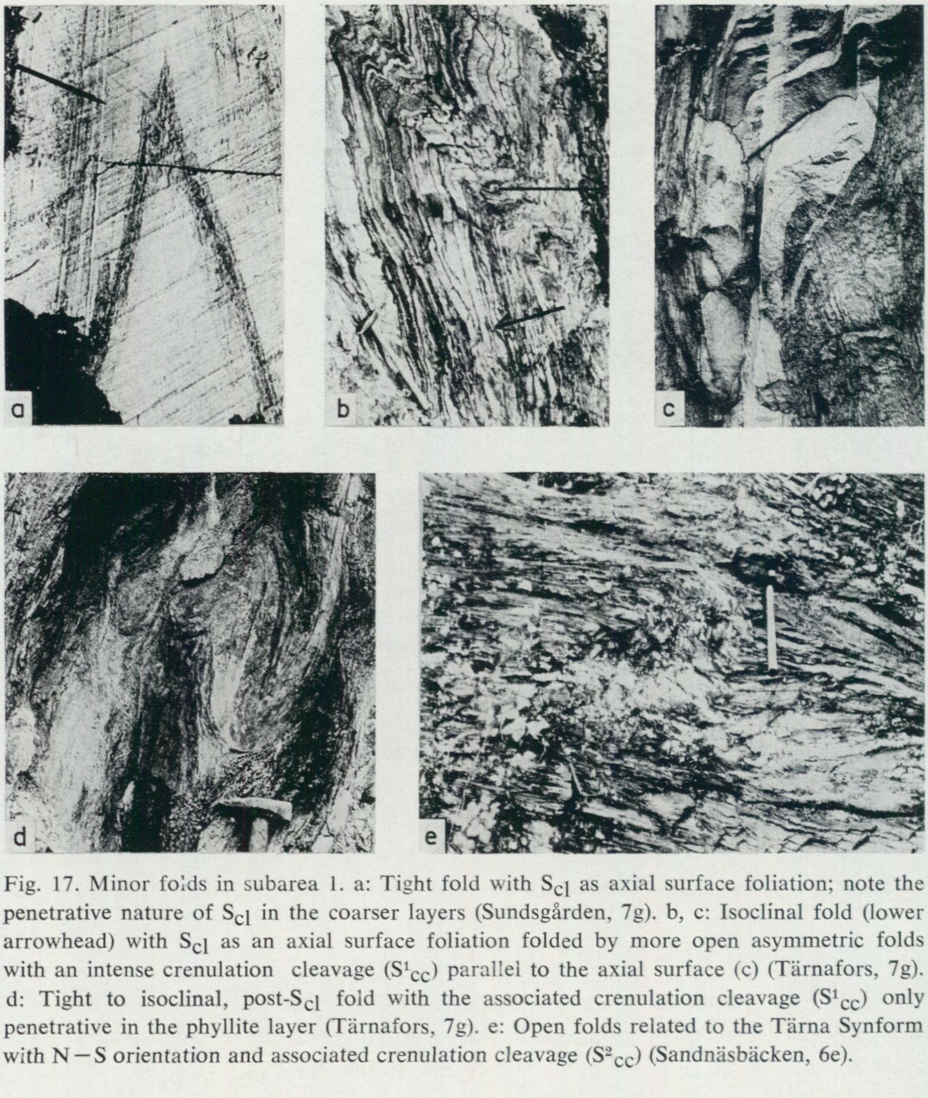


Fig. 17. Minor folds in subarea 1. a: Tight fold with S_{cl} as axial surface foliation; note the penetrative nature of S_{cl} in the coarser layers (Sundsgården, 7g). b, c: Isoclinal fold (lower arrowhead) with S_{cl} as an axial surface foliation folded by more open asymmetric folds with an intense crenulation cleavage (S^1_{cc}) parallel to the axial surface (c) (Tärnafors, 7g). d: Tight to isoclinal, post-S_{cl} fold with the associated crenulation cleavage (S^1_{cc}) only penetrative in the phyllite layer (Tärnafors, 7g). e: Open folds related to the Tärna Synform with N – S orientation and associated crenulation cleavage (S^2_{cc}) (Sandnäsbäcken, 6e).

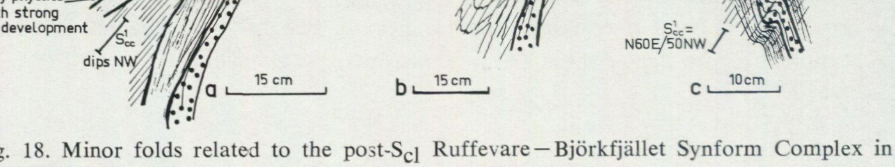


Fig. 18. Minor folds related to the post-S_{cl} Ruffevare – Björkfjället Synform Complex in subarea 1. a: Tärnafors, 7g. b: Tärnafors, 7g. c: Granås, 7f.

TABLE 3. Deformation of earlier lineations by the late Tärna Synform in subarea 1 — summary of angular relationships.

Partial Subarea	Vector Mean				$L_{cl} \wedge L_B$	$L_{cl} \wedge L_{TS}$	$L_B \wedge L_{TS}$
	L_{cl}	N	L_B	N			
W. Björkfjället	N30W/12	32	N42W/21	29	15°	46°	51°
E. Björkfjället	N56W/32	17	N54W/18	35	14°	57°	62°
Gakkervattnet (Hinge zone of Tärna Synform)	N82W/17	22	N84W/18	37	3°	85°	87°
Gakerevardo	S60W/60	7	S65W/17	45	42°	77°	108°

L_{cl} = lineations associated with S_{cl} (MEAN)

L_B = lineations associated with the Björkfjället folds (MEAN)

L_{TS} = lineations associated with the Tärna Synform (MEAN)

$L_{cl} \wedge L_B$ = angle between L_{cl} and L_B etc.

N = No. of measurements

5. The later, post-S_{cl} minor folds plunge at variable angles in a NNE direction (Plate III, 1f); the relatively high axial direction stability of these late structures is explained by the high angle that the Tärna Synform axial surface makes with the surfaces which it deforms (S_b, S_{cl} and S^1_{cc}).
6. Since the western limb of the Tärna Synform appears to be the short, rotated limb of the structure, the original orientation of the lineations associated with the Ruffevare—Björkfjället Synform complex was possibly closer to NE—SW.

SUBAREAS 2 AND 3: LULE-JALKETSVARDO — LÖVLUND AND FORMLIDEN

Major structures

The post-S_{cl} deformation in subarea 1 is dominated by NW- to SW-plunging, major and minor folds which are overturned southwards and are gently folded by an open, NNE-plunging, major synform (the Tärna Synform). The earlier, overturned structures fold an already inverted stratigraphy related to an important phase of isoclinal, recumbent folding (synclinal) developed probably prior to the formation of S_{cl}. Subareas 2 and 3 (Plate III), however, are dominated on both a major and minor scale by structures which correspond to the late, open folding mentioned above. The major structure in the Lule-Jalketsvardo—Lövlund subarea is the continuation of the Tärna Synform, whilst the Formliden subarea includes part of the complementary antiform (Formliden Antiform). These close, major folds possess a tighter profile than the gentle to open part of the Tärna Synform in subarea 1. Plots of S_b/S_{cl} poles for subareas 2 and 3 demonstrate their northerly plunge and steep axial surface dip (Plate III, 2a and 3a). The Tärna Synform is a conspicuously asymmetric fold with a steep easterly limb.

The stratigraphy within subarea 2 is regionally the right way-up being situated on the normal limb of the early, recumbent syncline; the beds, thus, young upwards in the later Tärna Synform. However, along Formbäcken (subarea 3), on the inverted limb of the early syncline, the beds young progressively southwards towards the eroded core of the late antiform. The stratigraphy is descended again at the downstream end of Formbäcken (7d) where greywackes of the Gilliks Formation occur (Plate I and Kulling 1933). Calcareous phyllites of the Lövfjäll Formation both underlie and overlie the greywackes with a consistent "northerly" dip. It would appear, then, that the Gilliks rocks close beneath Stor-Björkvattnet in a fold overturned towards the southwest (Plate I). Local sliding has occurred at least on the northern limb where several units are missing, and the basal Lövfjäll phyllites show faintly phyllonitic microstructure in the micaceous layers and zones of strongly recrystallized quartz in the coarser, siliceous layers. This pre-Formliden Antiform fold is geometrically similar to the Björkfjället folds defined in subarea 1, and is tentatively correlated with the phase of folding which produced the Björkfjället folds.

Minor structures

Although around Formliden there are lineations associated with S_{cl} (NW-plunging) and minor folds which deform S_{cl} on WNW axes (Plate III, 3b), subareas 2 and 3 are dominated by minor folds and crenulation lineations congruent to the Tärna Synform and Formliden Antiform respectively (Plate III, 2b and 3c).

The minor structures related to the Tärna Synform in subarea 2 are gently plunging, NNE- to SSW-trending, often disharmonic minor folds (Fig. 19) which occasionally show zones of contact strain in pelite adjacent to psammite layers. Similar structures congruent to the Formliden Antiform in Subarea 3 are close, N- and NNW-plunging minor folds (including bedding mullions), particularly well exposed in WNW-trending joint sections and also in road cuttings along Route E79 (8d—e). Such minor folds in both subareas are associated with an intense crenulation lineation on the S_{cl} surfaces, N- to NNE-trending axial surfaces, penetrative crenulation cleavage development (S^2_{cc}) and shear discontinuities on their limbs. S^2_{cc} is particularly penetrative in the finer phyllite layers where it may transpose the earlier foliations (S_b/S_{cl}), but is an obvious secondary crenulation cleavage in the coarser layers (Fig. 20).

A set of NW/WNW, steeply plunging crenulation lineations is also developed in subarea 2 (Plate III, 2b). This set is particularly conspicuous in the phyllites of the Lövfjäll Formation on Lule-Jalketsvardo but there is no clear

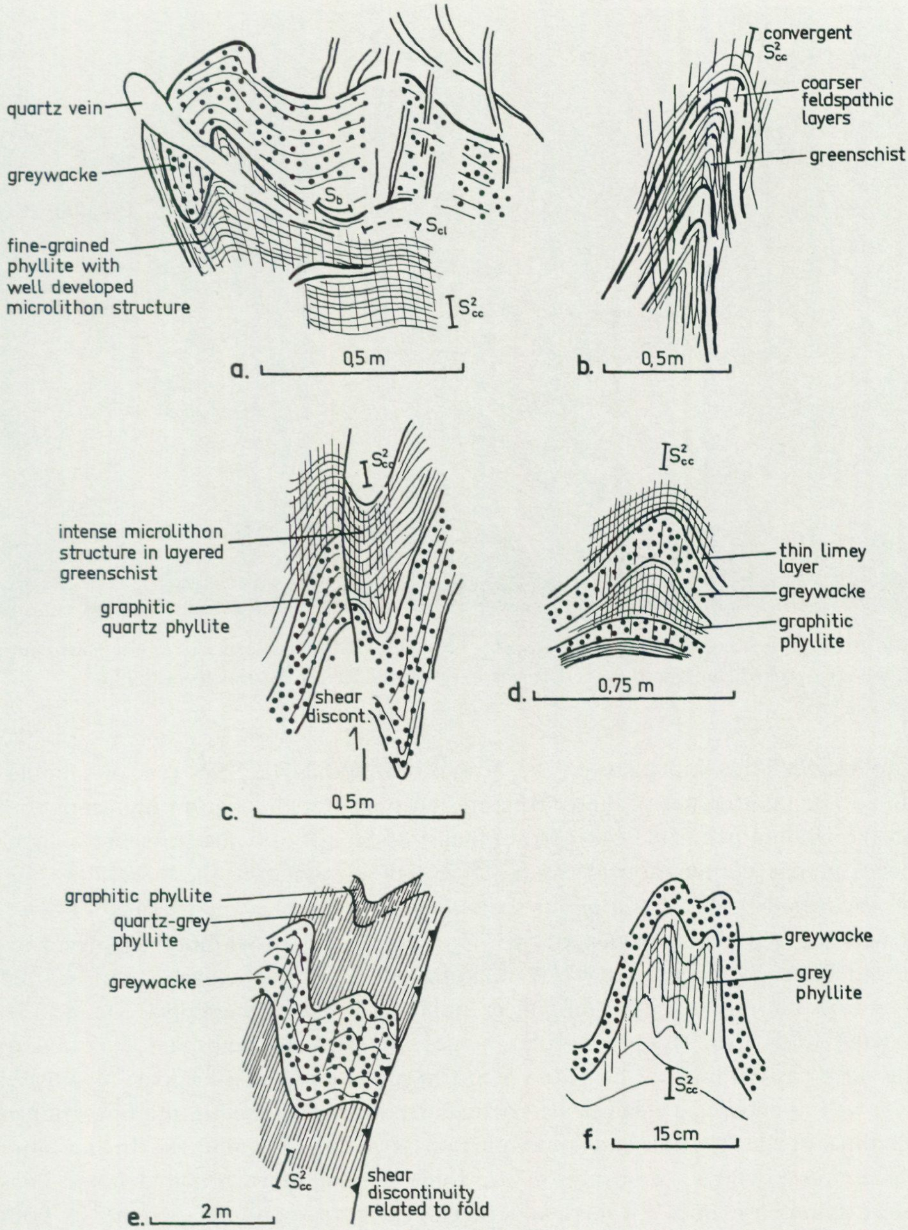


Fig. 19. Minor folds in subareas 2 and 3. a: Early, syn-S_{cl} isocline deformed by folds congruent to the northerly plunging Tärna Synform – note the intense transposition of S_{cl} in S^2_{cc} in the phyllitic layer (Lule-Jalketsvardo, 5f). b–e: Close minor folds related to the Tärna Synform – note the intense crenulation cleavage (S^2_{cc}) in the finer grained layers, the disharmonic nature of fold (d) and the presence of limb discontinuities (Lövlund/Lule-Jalketsvardo, 4–5f). f: Disharmonic, post-S_{cl} fold in the core of the Formliden Antiform, (NE Batsevardo, 8e).

Fig. 20. Transposition of S_{cl} parallel to S^2_{cc} in phyllite; S^2_{cc} is a distinct crenulate structure in the greywacke layer where S_{cl} is deformed around the fold (Lule-Jalketsvardo, 5f).

refolding relationship between the NW/WNW and NNE/SSW sets of crenulations. The interference of these structures accounts for the major outcrop pattern on the Bränna peninsula (4e—f), southeast of Stor-Björkvattnet, where a significant plunge culmination of the N—S axes is present (see the 1:200,000 map of Zachrisson 1969). On a minor scale there is a complex interference of N—S structures, including the penetrative S^2_{cc} cleavage here within the hinge zone of the Tärna Synform, and WNW-trending structures particularly characterized by a cross-cutting, steeply N-dipping crenulation cleavage (Stigh pers. comm.).

Kink folds with sharply angular hinges and straight limbs are confined to the steep eastern limb of the Tärna Synform (includes the eastern part of subarea 1 as well as subarea 2). These are particularly well developed in the fine-grained phyllites of the Lövfjäll and lower part of the Gilliks Formations. In the more calcareous phyllites, for example, the folds possess more rounded hinge lines (near symmetric buckle folds) and there is a corresponding increase in both their wavelength and amplitude (Fig. 21).

The deformation of both S^2_{cc} and the WNW-plunging crenulation lineation by these kink folds suggest that they post-date all the other sets of structures in subarea 2. Locally a crenulation cleavage is developed parallel to the axial surface of the kinks which generally plunge at low angles NNE—SSW (Plate III, 2c). However, their axial orientation is closely controlled by the strike of S_{cl}, the more NE- to SW-trending kink axes near Forsbäck and on southeastern Göutavardo reflecting the gentle swing in strike of S_{cl} towards a NE—SW direction from subarea 2 to the eastern part of subarea 1. The kinks are dominated by a set of folds with gently E-dipping axial surfaces (S_{kink}) which overturn down-dip; however, a set of kink folds with more N-dipping axial surfaces (S'_{kink}) is also developed. Rarely these two sets occur together and combine to form a conjugate relationship (Johnson 1956; Ramsay 1962a) in which the obtuse angle between the two axial surfaces apparently faces the direction of maximum shortening (see also Anderson 1964; Roberts 1971b). Earlier micas appear to be reoriented parallel to the axial surfaces of the kink folds which also contain occasional quartz veins; minor faults are present parallel to the axial surfaces of those kink folds with down-dip sense of overturn (S_{kink}).

SUBAREA 4: BRAKKO – GÖUTA – FORSBÄCK

Major structures

The structure of Brakko is interpreted as a series of tight to isoclinal major folds (Brakko Fold Complex) which deform the pervasive schistosity (S_s) and are now overturned to the northeast (Plates I and III). The axial trend of the major folds is indicated in Plate III (4a) which also shows the dominant NW strike and gentle to moderate (<45—50°), SW dip of S_s. The most important structure is an antiform which closes on the western side of Brakko (Plates I and III). This is deformed by a later, more gentle and upright fold (antiformal arch) again with an approximate WNW trend. The later, antiformal arch accounts for the swing in strike of the beds from NW (dipping SW) to NNE (dipping W) on the west side of Brakko.

In the Göuta—Forsbäck part of the subarea the dominant structure is a late, post-S_s, upright antiform — the Göuta Antiform — which gently folds S_s, apparently plunges WNW (Plate III, 4b), dies out rapidly at higher tectonic levels and belongs to the same phase as the antiformal arch referred to above. Major, overturned folds of the type which characterize Brakko have not been detected.

There is a rather rapid decline in the propagation of the tight, overturned folds on Brakko towards the characteristic Köli rocks to the southwest. Furthermore, near the top of the Brakko schists and in the Forsbäck Formation, in the Göuta—Forsbäck part of the subarea, there is retrogression of the metamorphic assemblages and local phyllonitization of S_s (Fig. 22). It is suggested, therefore, that there is a broad zone of post-S_{cl}/S_s adjustment in the rocks above the Brakko schists (Forsbäck Formation in the north, base of Seima Formation in the south) which, particularly in the north, is associated with retrogression and local phyllonitization phenomena. As discussed later, this is thought to be related to the

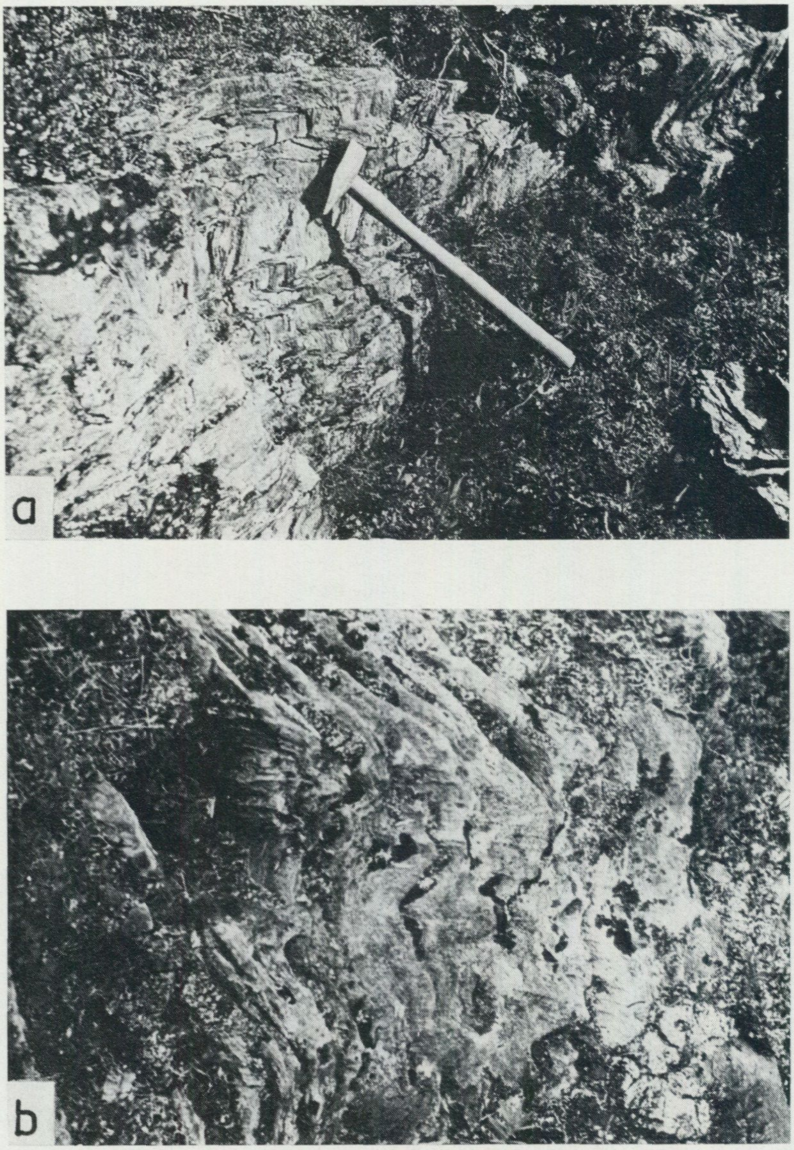


Fig. 21. Kink folds on the eastern limb of the Tärna Synform. Folds show a down-dip sense of overturn. Lithology in (a) is fine-grained phyllite; in (b) coarser grained calcareous phyllite.

Fig. 21 (continued). Kink folds on the eastern limb of the Tärna Synform. Conjugate arrangements. The kinks in (c) deform a complex transposition foliation (S_{cl} in $S^2{}_{cc}$). Lithology – fine grained phyllite.

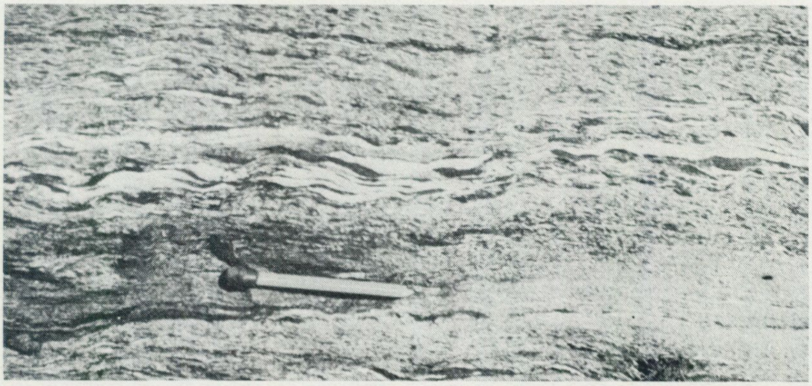


Fig. 22. Lenticularization of pelite layers within the garnetiferous psammites of the Forsbäck Formation (Forsbäck, 6g).

development of broadly similar structural sequences but rather different metamorphic histories — particularly post-S_{cl}/S_S — in the Köli and Seve rocks. The question of how much movement has occurred at this level is more difficult to answer. However, the same sequence from garnetiferous schists and amphibolites to either a serpentinite-greenstone association or garnet-bearing quartz and grey phyllites has been mapped from Göutavardo (Tärna) in the north to Fättjaure in the south (Stigh pers. comm.) — a strike length of approximately 50 km. This suggests to the author that only minor movement is feasible at this level. Thus, it is inferred that the prograde metamorphic sequence between the Seima/Forsbäck Formations and Brakko schists (Stephens 1973) is the important relationship and that this prograde sequence is locally disturbed by only minor post-metamorphic movement and associated retrogression and phyllonitization effects.

Minor structures

The earliest minor structures are linear fabric elements related to S_S and include the dominant quartz rodding in the schists, occasional isoclinal minor folds strongly transposed in S_S, mineral lineations of various types (Table 2) and the oriented long axes of deformed and recrystallized phenocrysts in the amphibolites (Fig. 23); the last two types of lineation are considered to represent the maximum elongation direction in S_S. The strongly transposed isoclines possess a poorly defined profile with strongly sheared limbs and occasionally detached fold hinges (Fig. 23). All these lineations generally plunge 30—40° to the west (Plate III, 4c).

Post-S_S minor structures dominate the subarea. On Brakko the most common minor folds are tight to isoclinal in shape, overturned and often show limb discontinuities (Fig. 24). No obvious crenulation cleavage or axial surface structure is associated with these folds. Like the major folds to which they relate (Brakko Fold Complex), plunge angles are generally $<40°$ towards the NW—WNW and axial surfaces dip SW and W (Plate III, 4d). These minor folds are deformed on a large scale in the later Göuta Antiform, so that easterly trends and N-dipping axial surfaces are encountered on the northern limb of the antiform in the Göuta—Forsbäck area.

Later post-S_S folds are open to close and upright in shape, often show good refolding relationships in relation to the earlier structures, are related to the Göuta Antiform and, thus, dominate in the northern part of the subarea, and may be associated with a weak crenulation cleavage situated along the axial surfaces of microfolds in S_S in the more pelitic and fine-grained amphibolitic layers (Fig. 24). Orientation data suggest the existence of a NE-trending (dipping NW) and a NW- to WNW-trending (variable dips NE) set of axial surfaces (Plate III, 4e) the time relationship between which is uncertain. The pattern is

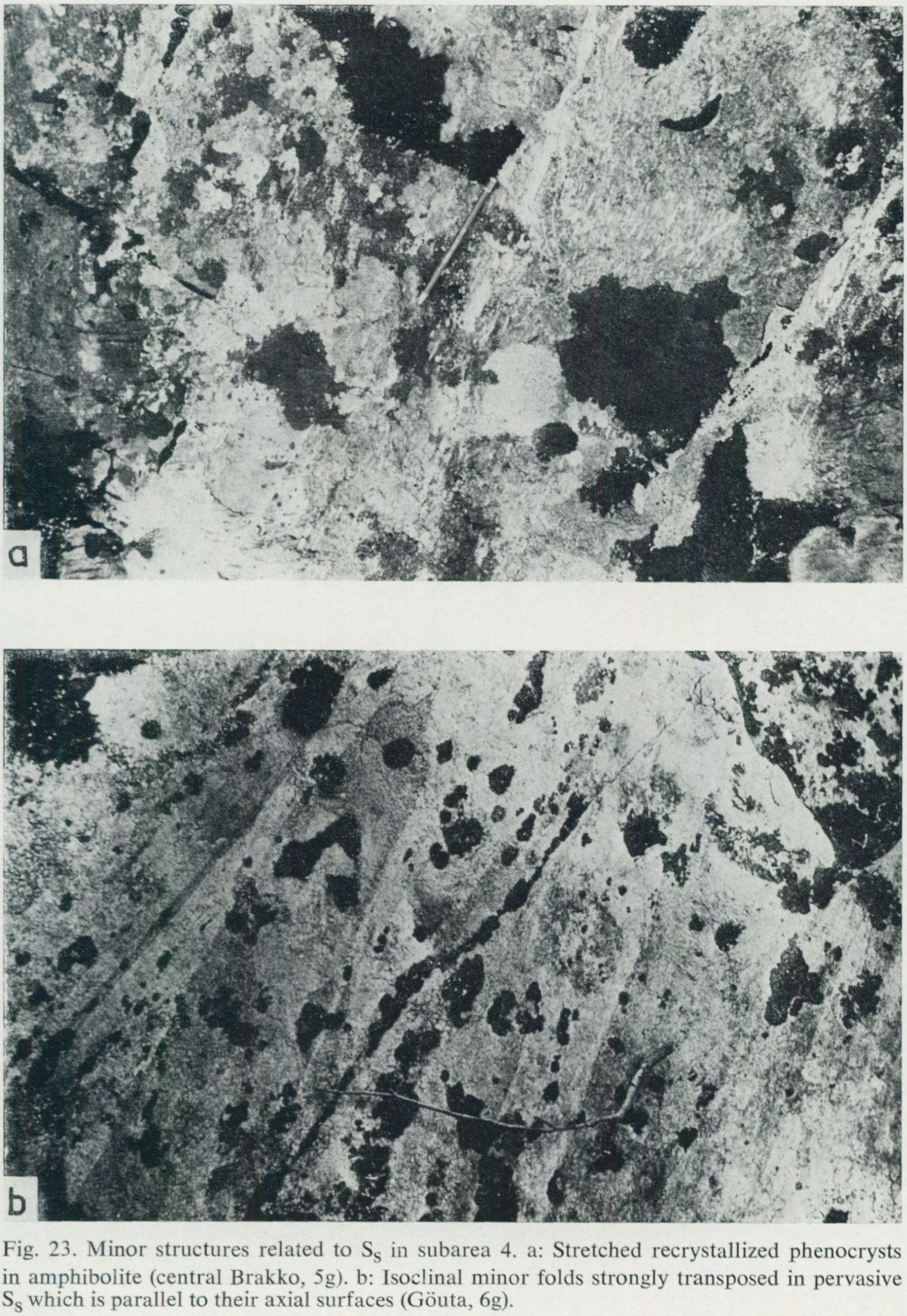


Fig. 23. Minor structures related to S_S in subarea 4. a: Stretched recrystallized phenocrysts in amphibolite (central Brakko, 5g). b: Isoclinal minor folds strongly transposed in pervasive S_S which is parallel to their axial surfaces (Göuta, 6g).

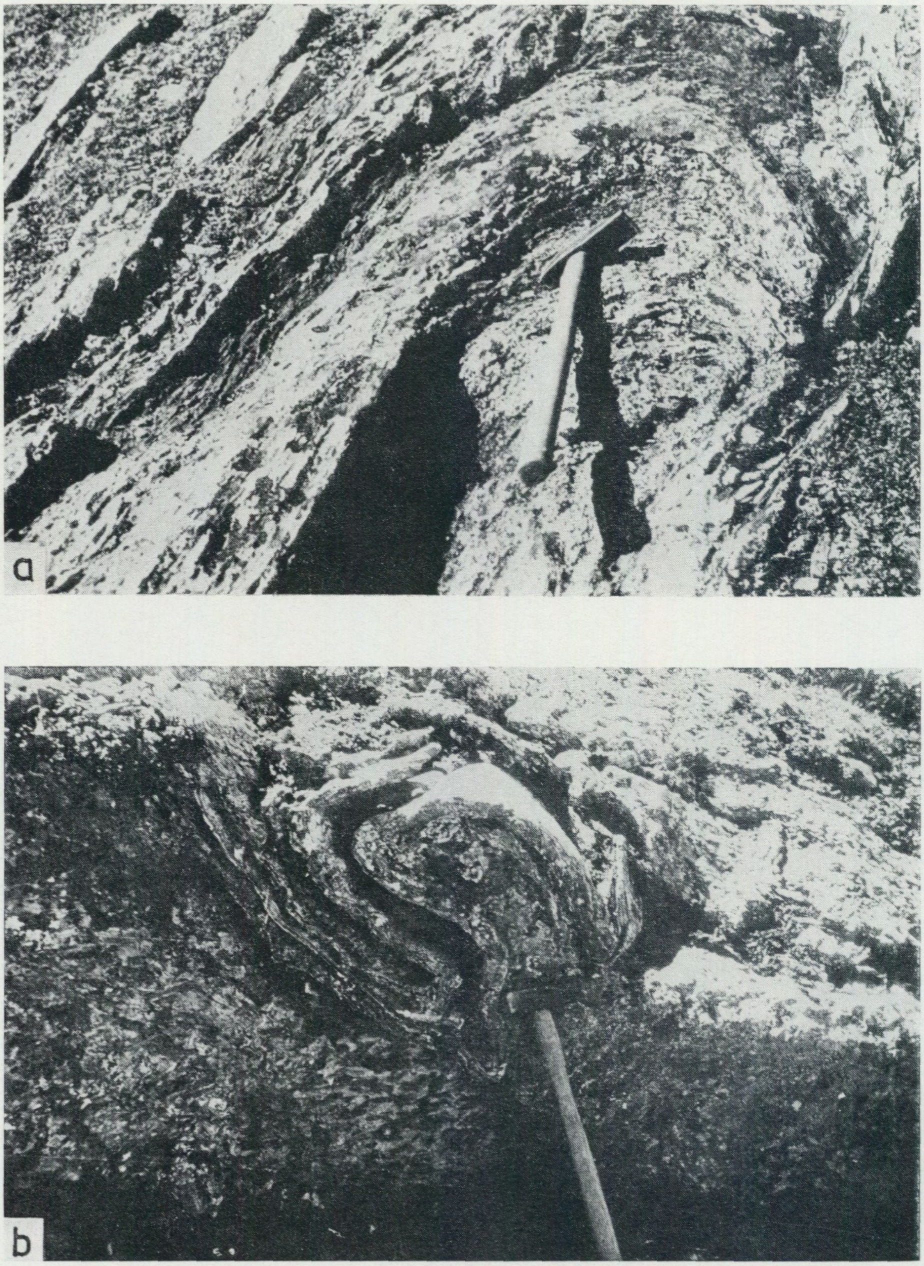


Fig. 24. Post-S_S minor folds in subarea 4. a: Tight to isoclinal fold plunging gently NW — note the development of a limb discontinuity and the absence of any later axial surface foliation (central Brakko, 5g). b: Post-S_S, isoclinal fold deformed by later open to gentle fold (central Brakko, 5g).

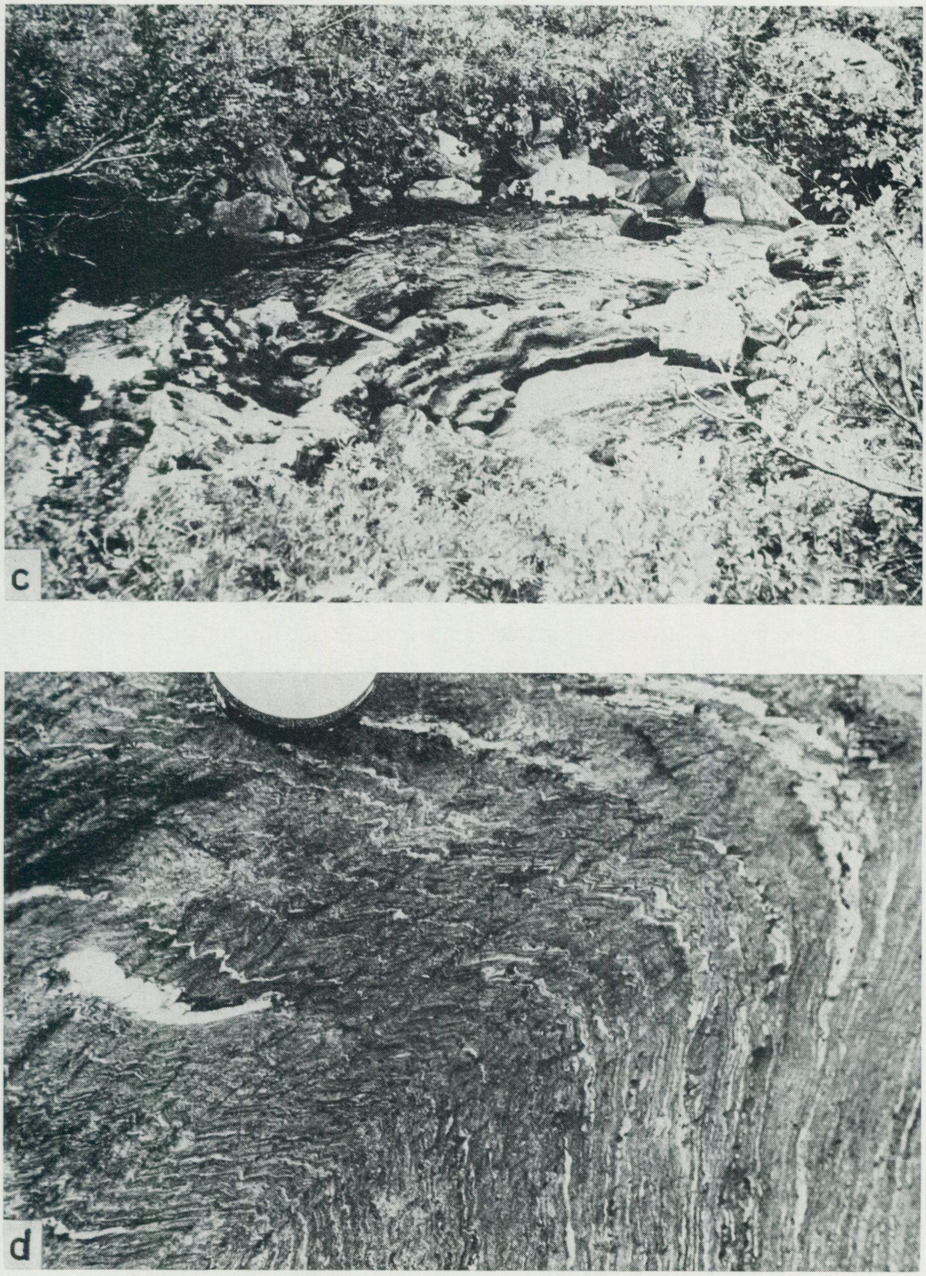


Fig. 24 (continued). Post-S_S minor folds in subarea 4. c: Gentle minor folds plunging WNW; these structures are related to the Göuta Antiform (west Brakko, 5g). d: Development of crenulation cleavage in late, open folds associated with the Göuta Antiform (Göuta, 6g).

very similar to that in subarea 2 where, however, the NNE-trending folds dominate over a set of crenulations plunging NW/WNW. The general cross-cutting combination of the Tärna Synform and Göuta Antiform in the eastern part of the area accounts for the overall saddle-shaped outcrop pattern. The occurrence of other gentle NW- and WNW- trending folds crossing the general N—S to NE—SW structures (Zachrisson 1969, 1971) has been recorded in other areas in this part of the Caledonides (e.g. Murris 1957). Trouw (1973) has suggested that the Ransaren Synform (southerly continuation of the Eastern Synform) is a composite structure with cross-cutting NE and NW components. Furthermore, a similar late fold pattern has been described by Henley (1970) from the Norwegian Sulitjelma region.

SYNTHESIS OF THE STRUCTURAL SEQUENCE

The earliest deformation (D1) recognised in the low-grade rocks is an important phase of recumbent, isoclinal folding which deforms the primary compositional layering (S_b=S0) and accounts for the synclinal repetition of the stratigraphy. The exact geometry of this major fold is obscure on account of the local absence of closure zones and related minor structures. The analysis in subarea 1 demands that this phase pre-dates the Ruffevare—Björkfjället Synform Complex and suggests that it formed prior to the establishment of the early, regional cleavage (S_{cl}). The latter inference is strengthened when the so-called pre-F1 folds of Zachrisson (1969) are considered — in particular the large, pre-F1, recumbent syncline near Lake Väsken. It is possible that the early syncline defined in the present area is the northerly continuation of this recumbent structure; since Zachrisson correlated his F1 phase with the formation of the early, regional schistosity, then the pre-cleavage interpretation for the early syncline in the Björkvattnet Unit is supported. S_{cl} is, therefore, referred to as S2 here.

There is a distinct pattern of deformation in both the low- and higher grade rocks associated with the formation of the early, regional foliation. The trend of minor folds broadly related to S_{cl}/S_s and the maximum elongation direction in S_{cl}/S_s are parallel to each other and plunge in a westerly direction throughout the Björkvattnet Unit. Since the coarse-grained schistosity (S_s) in the higher grade rocks can be traced without discordance or evidence for later transposition into the early cleavage (S_{cl}) of the low-grade rocks, it is concluded that an intense phase of relatively more homogeneous strain affected both groups of rock simultaneously. Thus, S_s is also referred to as S2 in the structural sequence and all minor folds related to S_{cl}/S_s are referred to as F2. These D2 structures are correlated with F1 of Zachrisson (1969) and F1 of Trouw (1973).

S2 and its associated, pervasive lineations are folded by overturned, major and minor folds which plunge NW to SW in the Köli rocks (original orientation possibly closer to NE—SW) and NW—WNW to ENE in the Seve rocks i.e. approximately parallel to the strike of S2 (the spread in orientation of both these folds and the lineations associated with S2 is related to deformation by the later Tärna Synform and Göuta Antiform). They are associated with the development of the first phase crenulation cleavage (S^1_{cc}), which in subarea 1 is locally penetrative and transposes S0/S2, and, as will be discussed later, significant mylonite formation and thrusting. Besides the locally important slides associated with the overturned Ruffevare—Björkfjället Synform Complex there are also:

1. Post-metamorphic mylonitization of the Tärna schists in a steeply dipping zone which becomes more prominent nearer to the thrust contact with the Bångfjället Complex.
2. Relatively less important phyllonitization and retrogression phenomena near the top of the Brakko schists and in the Forsbäck Formation around Göuta—Forsbäck.

The consistent folding relationship between the overturned folds related to the first phase crenulation cleavage and S2 supports the view that the overturned folds and their associated minor structures form a separate phase of deformation (D3); the overturned folds and associated axial surface crenulation cleavage are, thus, referred to as F3 and S3 respectively. These structures appear similar to the F2 structures described by Trouw (1973).

The final deformation phase (D4) contrasts with the earlier phases by the general development of open to gentle folds with steep axial surfaces, associated with crenulation cleavages (including S^2_{cc}) and the restricted development of kinks particularly in the more phyllitic lithologies. However, in subareas 2 and 3 both major and minor folds are close and the northerly trending crenulation cleavage (S^2_{cc}) is locally penetrative and transposes S0/S2. Since there is an uncertain time relationship between the northerly plunging Tärna Synform and Formliden Antiform and the WNW-plunging Göuta Antiform, they are referred to here as $F4_a$ and $F4_b$ respectively.

The kink folds ($F4_{kink}$) are developed exclusively on the steep eastern limb of the Tärna Synform. The kinks are generally overturned down the steep dip but occasionally are found in a conjugate relationship together with the incipient development of a pair of axial surface crenulation cleavages ($S4_{kink}$). Although the kinks deform the northerly trending crenulation cleavage ($S^2_{cc} = S4_a$) and the crenulation associated with the NW- to WNW-trending, $F4_b$ folds, there appears to be a significant relationship between their development and the formation of the steepened eastern limb of the Tärna Synform. For this reason the kinks are also referred to as belonging broadly to the D4 phase. The later structures (F4 and $F4_{kink}$) in the Björkvattnet Unit are tentatively correlated with F2 and F3 of Zachrisson (1969), and F3 and F4 of Trouw (1973).

AXIAL SURFACE FOLIATION CHARACTERISTICS AND EVOLUTION

CHARACTER AND FORMATION OF S2

S2 in the low-grade rocks is generally defined by an alignment of the trace of the (001) planes of phyllosilicates, including white mica, biotite and chlorite (Fig. 25a). In the more pelitic lithologies S2 is penetrative down to at least the scale of individual phyllosilicate grains. In the quartz-rich phyllites and greywackes the S2 foliation is anastomosing, the thin mica seams deflecting around coarse porphyroclasts of quartz and feldspar. The quartz clasts are flat and elongate in S2 (Fig. 25b), show conspicuously undulose extinction and occasional deformation lamellae, and are sometimes surrounded by fine new grains (Fig. 25c). There is a similar old grain-new grain relationship in the purer quartzites where two extremes of microstructure appear to be present (see Fig. 9):

1. Sutured and irregular quartz-quartz grain boundaries with undulose extinction inside the old quartz grains and smaller, new, unstrained grains surronding the old grains; in some specimens it is difficult to distinguish old and new grains although grain boundaries remain irregular.
2. Gently curved or straight grain boundaries. This microstructure is tending towards a truly polygonal-shaped quartz aggregate. The grain size is noticeably larger than in those quartzites showing both old and new grains (Fig. 9).

On a larger scale S2 is concomitant with the deformation of lithified quartzite pebbles and a penetrative quartz and mica lineation in the quartzites and greywackes. Strain analysis on the prolate-shaped pebbles indicates that the D2 finite strain ellipsoid is probably constrictional in the quartzite conglomerate with k values (Flinn 1962) ranging from 6 to 19 and that the strain is high (Stephens 1975). Prismatic actinolite/tremolite in the more basic metavolcanics is also strongly oriented parallel to the mineral lineation. The rocks are metamorphosed in the chlorite and biotite zones of the Barrovian type series.

S2 in the low-grade rocks passes transitionally downwards into the coarse-grained mica schistosity of the Brakko schists (Fig. 25d); as already noted the L-S fabric (Flinn 1965) is similar in both the low- and high-grade rocks, the slender hornblendes in the amphibolites defining a penetrative lineation on the S2 surface. These hornblendes contrast with the deformed hornblende poikiloblasts in the garbenschiefer and calcareous schists which show undulose extinction, are boudinaged parallel to the penetrative mineral lineation and are also bent (Fig. 26). The deformed hornblendes, together with some coarse garnet and biotite poikiloblasts, define a dimensional mineral orientation parallel to the penetrative hornblende lineation. The peak of metamorphism was established *prior* to the formation of S2 (Stephens 1973 and see discussion later) as indicated by the growth relationships of biotite, garnet and hornblende poikiloblasts.

However, similar grade conditions appear to have persisted during the development of the S2 fabric in the Brakko schists.

It is suggested that important mechanisms which may be responsible for the formation of S2 include intracrystalline slip, recovery and recrystallization processes in the more quartz-rich rocks and syntectonic grain growth in the pelitic and basic rocks.

The presence of undulose extinction, deformation lamellae and new grain development in and around old, detrital quartz grains within the more quartz-rich rocks suggests the importance here of intracrystalline slip, recovery and recrystallization (Wilson 1973). Both microstructures within the purer quartzites are related to the S2 foliation i.e. one phase of deformation. The extent to which intracrystalline deformation, recovery and recrystallization processes are operative varies in different parts of the sequence. The microstructure with irregular grain boundaries is dominated by intracrystalline slip processes in the old grains and the development of recovered and subsequently recrystallized new grains. The quartzites with gently curved or straight crystal boundaries and increased grain size are dominated by recrystallized new grains which have approached an equilibrium state, defined by unstrained polygonal grains. This equilibrium state is achieved by grain boundary migration of the new quartz grains (Kretz 1966).

The relative homogeneity of the S2 fabric, the absence of intermediate mica orientations (excepting the obvious pre-S2 biotite porphyroblasts in the high-grade schists) and the metamorphic grade, which is consistent with the growth of white mica, chlorite and biotite in reactions involving clay minerals, quartz and feldspar (and subsequently white mica and chlorite), suggest the importance of syntectonic growth of new micas in the pelitic rocks. It is uncertain how much mechanical rotation of inequidimensional particles by bulk finite strain and mass transfer of material preceded this grain growth in an early stage of S2 development, and was subsequently obliterated during the growth of new micas at a late stage.

In the amphibolites and hornblende-rich rocks syntectonic grain growth and to a lesser extent mechanical rotation of pre-existing mineral constituents are thought to be important. It is suggested that the slender hornblende crystals in the amphibolites grew during D2, defining a penetrative lineation on the S2 surface, whilst the coarse hornblende poikiloblasts appear to have grown prior to the formation of S2 and to have been deformed and rotated into S2 parallel to the penetrative hornblende lineation. This mineral lineation is parallel to the oriented long axes of deformed and recrystallized plagioclase phenocrysts in the amphibolites which are considered to represent the maximum elongation direction in S2.

Thus, it is suggested that the S2 foliation developed by different mechanisms depending on rock type and presumably the variable physical conditions during

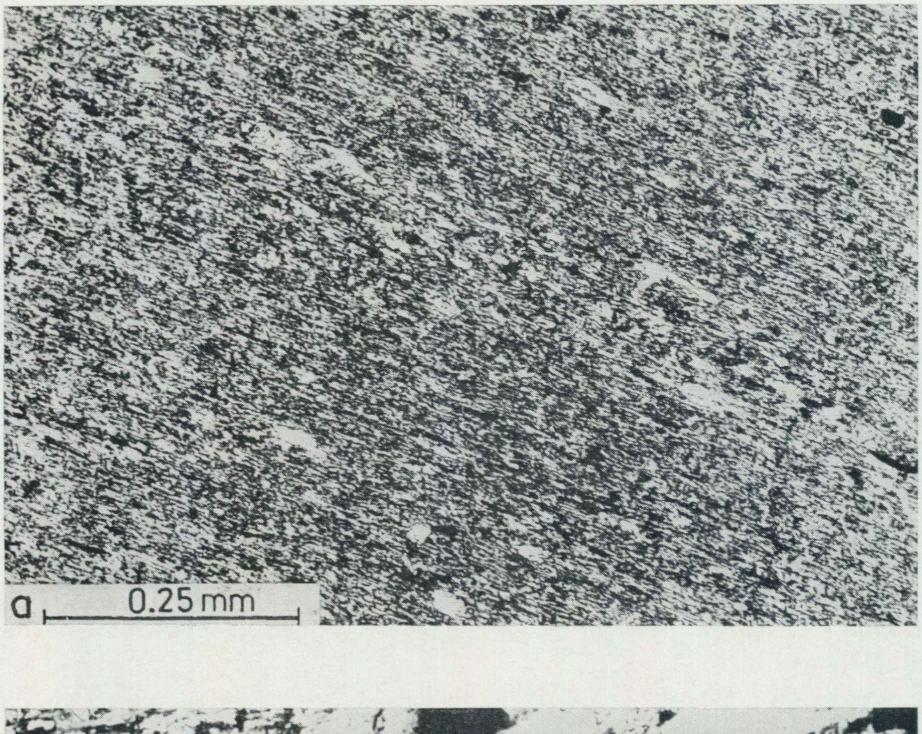
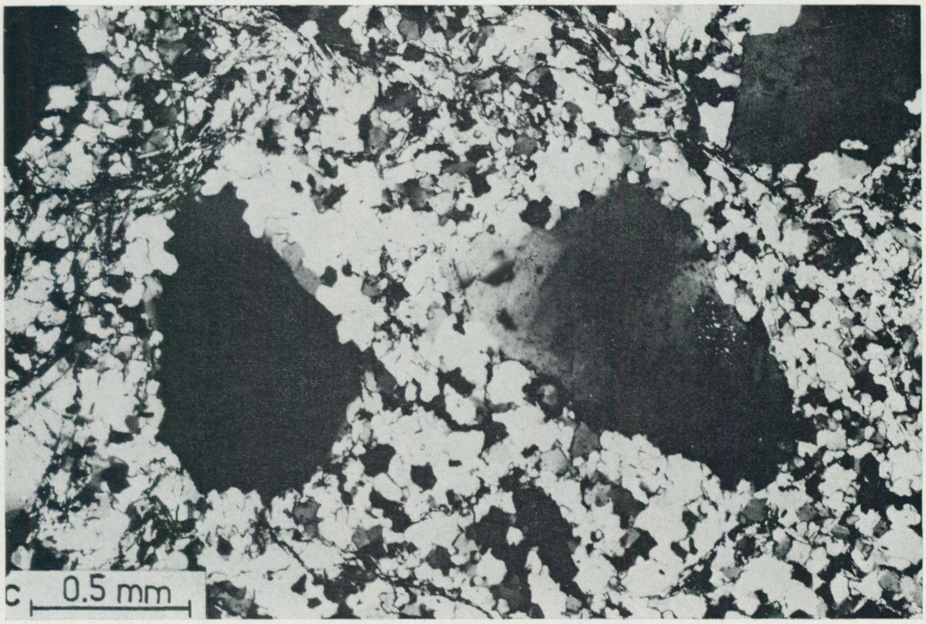


Fig. 25. Summary of S2 characteristics. a: Penetrative S2 in grey phyllite; section cut perpendicular to the cleavage (E. Björkfjället, 6e, uncrossed nicols). b: Deformed detrital quartz with deformation lamellae and undulose extinction enclosed by S2 foliation in calcareous phyllite (Sandnäsbäcken, 6e, crossed nicols).



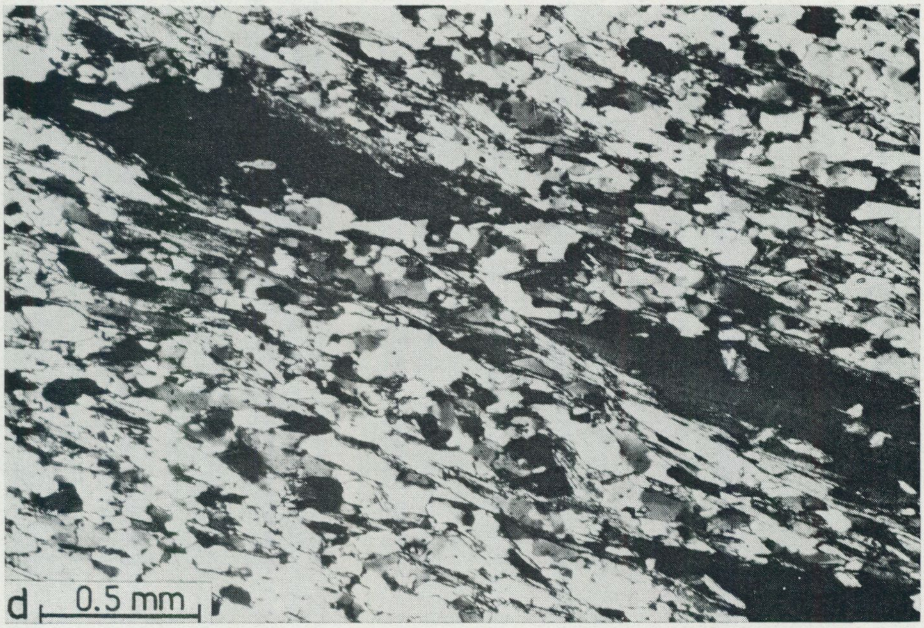


Fig. 25 (continued). Summary of S2 characteristics. c: New grain development around old detrital quartz grains (Formliden, 8d, crossed nicols). d: Coarse-grained muscovites in S2 enclosing pre-S2 biotite porphyroblasts (Brakko, 5g, crossed nicols).

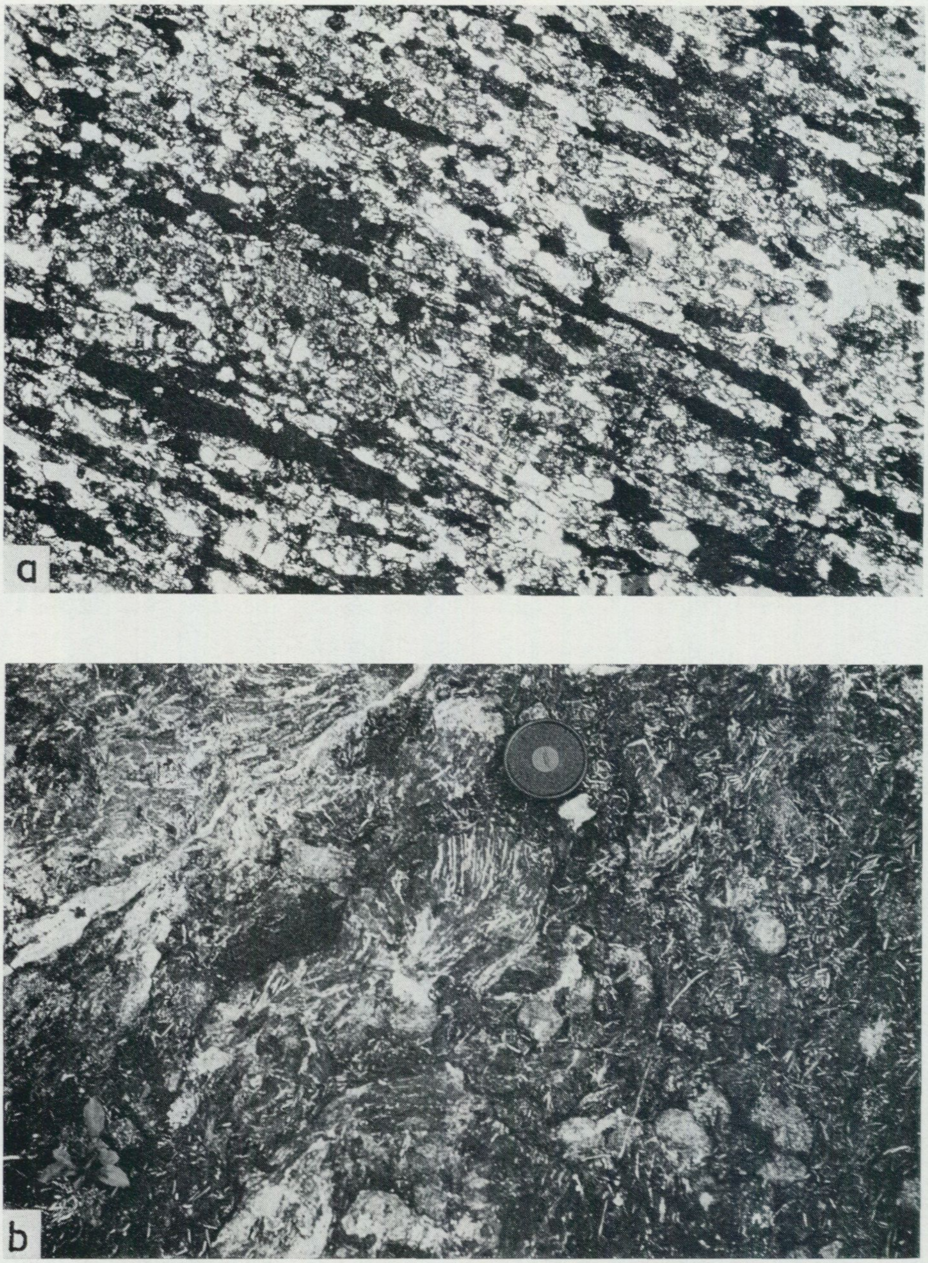


Fig. 26. Hornblende-S2 relationships in the metabasites. a: Well oriented unstrained hornblendes in amphibolite cut parallel to the mineral lineation in S2 (Brakko, 5g, crossed nicols). b: Deformed, coarser hornblendes in skarn rock located between schist and relict marble pod (Göuta, 6g).

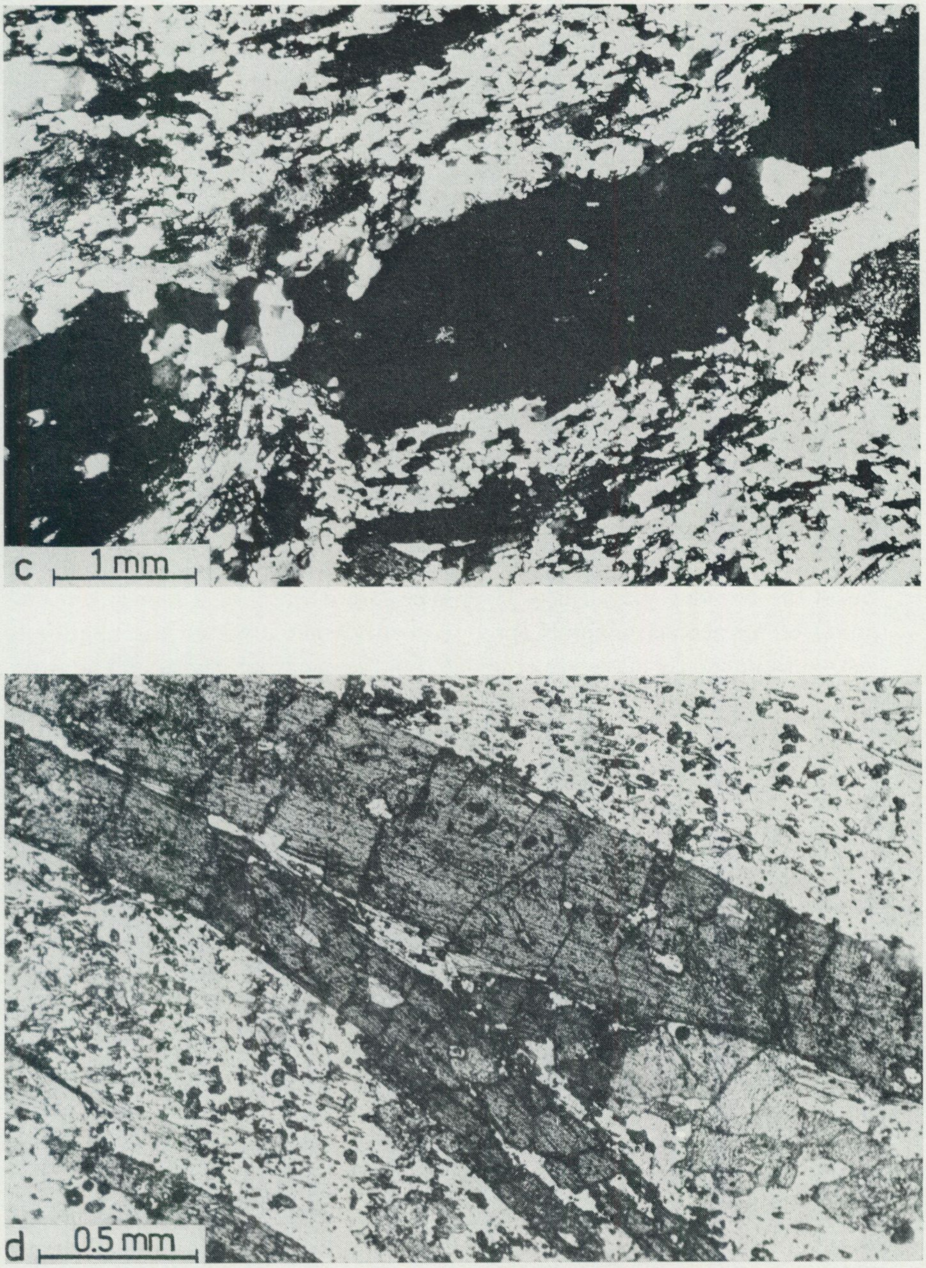


Fig. 26 (continued). Hornblende-S2 relationships in the metabasites. c: Boudined hornblende in feldspathic garnet-hornblende garbenschiefer (SW Brakko, 4g, uncrossed nicols). d: Strongly curved and deformed hornblende in garbenschiefer layer near the base of the Seima Formation (SW Brakko, 4g, uncrossed nicols).

D2 at different levels and positions in the sequence. Intracrystalline deformation, recovery and recrystallization mechanisms appear important in the more quartz-rich rocks, whilst syntectonic grain growth mechanisms are significant in the pelitic and basic rocks. These processes appear to be operative in response to an intense, relatively more homogeneous strain close to the peak of metamorphism.

CHARACTER AND FORMATION OF THE CRENULATION CLEAVAGES ($S3$ AND $S4_a$)

Unlike S2, the S3 and $S4_a$ cleavages are conspicuously semi-penetrative and strongly heterogeneous. The following summarizes their principal characteristics:

1. They either appear as a series of narrow dark lines running along the middle of the strongly attenuated short limbs of minor asymmetric crenulations or are broader zones also associated with the short limbs of asymmetric microfolds and both limbs of symmetric crenulations (Fig. 27, a, b, c). The narrow dark lines are composed of fine-grained opaque minerals and/or graphite, whilst the broader zones (mica-layer type) are defined by a higher concentration of micaceous minerals (predominantly rotated S2 micas) relative to quartz. Relic quartz clasts are attenuated in the mica-rich layers and approximate more closely the pre-crenulation cleavage shape in the relatively quartz-rich areas between the micaceous layers (Fig. 27c). Both the narrow trains of opaque minerals and the micaceous layers are extremely irregular. Both crenulation cleavage types grade into one another and often terminate at the intersection with psammitic laminae or where the microfolds become conspicuously more symmetric in profile.
2. The cleavages described above grade into a differentiated crenulation cleavage (see also Williams 1972) consisting of layers alternately rich in phyllosilicates and phyllosilicates plus quartz (Fig. 27d). The mica and chlorite in the relatively quartz-free domains are strongly oriented at an angle of up to 30° with the length of the domain. The chlorite-rich layers within the differentiated S3 crenulation cleavage in the tuffites of the Broken Formation (eastern Björkfjället) are themselves microfolded; this gives rise to a new crenulation cleavage parallel to the microfold axial surfaces which are up to 40° from the S3 trace (Fig. 28a).
3. Often in pelite layers the crenulation cleavage is highly penetrative, although there are still isolated lenses where there is an increased amount of quartz and the earlier S2 orientation is preserved i.e. the S2 foliation is transposed into the later surface. However, the mica-rich layers are broad enough to give the fabric in the pelite an "imposed slaty cleavage" appearance, particularly in hand specimen (Fig. 27e).
4. The crenulation cleavages often fail to penetrate the psammitic layers which

are simply buckled in a series of microfolds of larger wavelength (by a factor of 10 or above) than the crenulation within the alternating pelite layers, When the S3 or $S4_a$ cleavage does penetrate these layers, it parallels the axial surfaces of the microfolds as a normal crenulation cleavage (see Fig. 20).

5. Thin psammite laminae etc. are *apparently* offset along the crenulation cleavage surfaces to produce a typical microlithon structure (de Sitter 1954). The dark opaque zones which define the crenulation cleavage surfaces are thicker where the apparent displacements are larger (Fig. 27f). This feature has also been noted in connection with crenulation cleavage by Williams (1972) and Trouw (1973).
6. Neither of the generations of crenulation cleavage belong to a conjugate arrangement of secondary cleavage surfaces.
7. The S3 and $S4_a$ crenulation cleavages are approximately axial planar to associated minor folds. They occasionally show distinctive fanning arrangements which vary with lithology defining, in particular, locally divergent fans in pelitic layers (Fig. 28b). In all the fanning arrangements studied on individual folds, neither finite neutral points (Ramsay 1967, p. 416) nor associated arcuate hinge cleavage (Mukhopadhyay 1965; Roberts 1971c) have been observed.
8. The crenulation cleavages are almost exclusively developed in the low-grade Köli rocks. The F3 folds in the Seve rocks do not have any associated axial surface structure. A weakly developed crenulation cleavage is only present in the schists and amphibolites in the core of the $F4_b$ Göuta Antiform.

A number of workers, including Voll (1960), Plessman (1964), Nicholson (1966), Durney (1972) and Williams (1972), have emphasized the high mobility of quartz and calcite in the evolution of compositional layering parallel to crenulation cleavage. In particular, Nicholson (1966) suggested that quartz migrates away from the highly strained short limbs towards the less strained long limbs of asymmetric folds, and away from both limbs towards the hinge areas of symmetric folds.

The experiments of Means and Williams (1972) produced *conjugate* fault-like and crenulation cleavage discontinuities in specimens compressed *normal* to an already established foliation. They concluded that transfer of material was a contributory mechanism in the cleavage differentiation process. They also established that in specimens deformed wet at relatively lower strain rates, formation of crenulation cleavage was favoured over faulting. Other experiments, where layered specimens were deformed wet and loaded parallel to the foliation (Williams and Means 1971), led to the formation of a crude axial planar crenulation cleavage in the pelitic layers. The presence or absence of water as a pore fluid was critical in all the experiments and they suggested that the transport mechanism was transfer of material in aqueous solution.

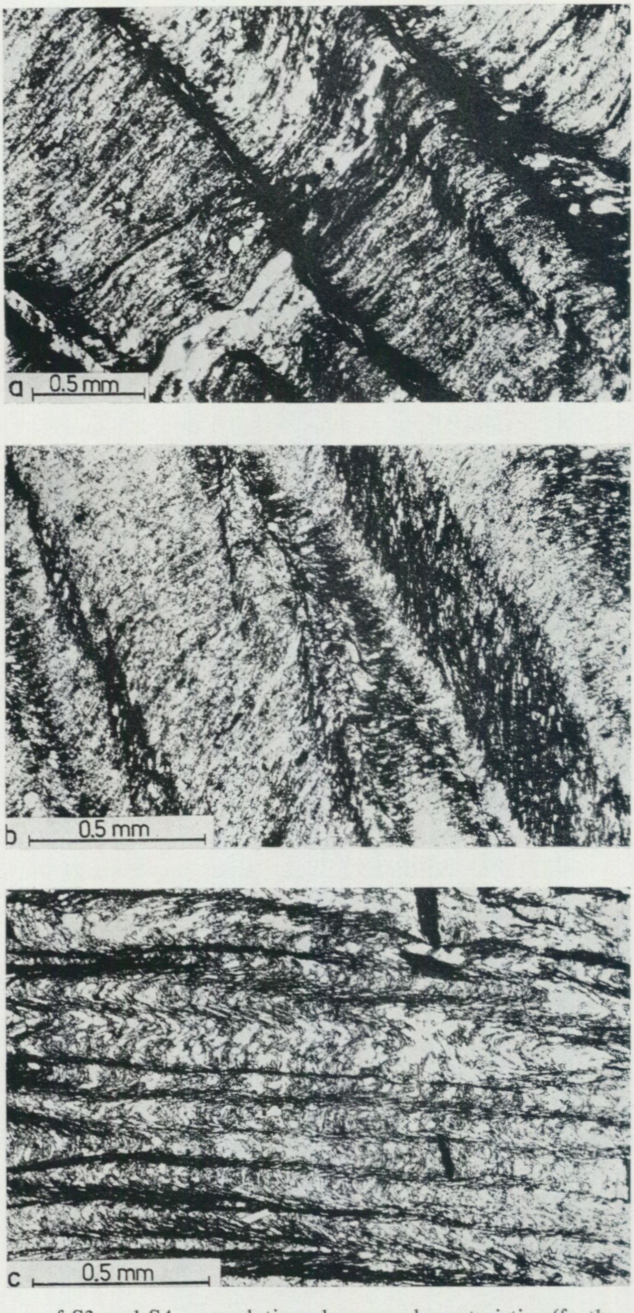


Fig. 27. Summary of S3 and $S4_a$ crenulation cleavage characteristics (further explanation in text). a: S3 crenulation cleavage; asymmetric crenulations (E. Björkfjället, 6e crossed nicols). b: S3 crenulation cleavage; mica layer/asymmetric crenulation type (E. Björkfjället, 6e, crossed nicols). c: S3 crenulation cleavage; symmetric crenulation type (Granås, 7f, uncrossed nicols).

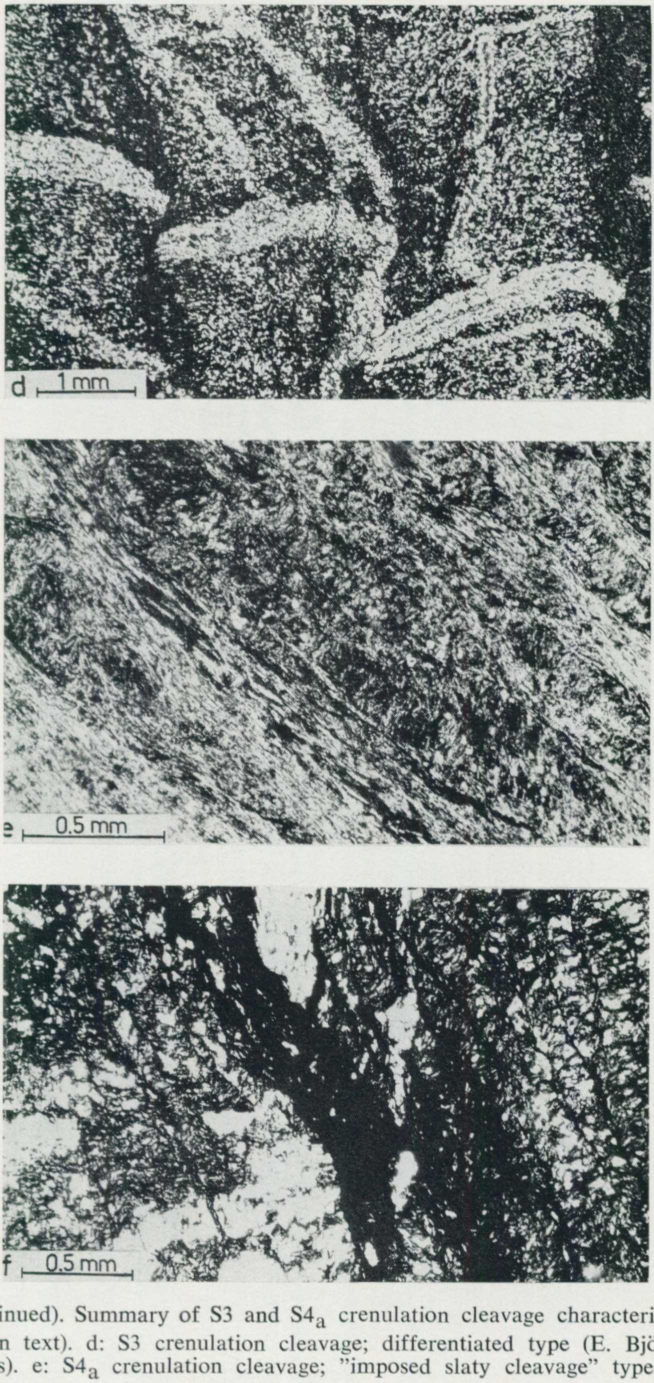


Fig. 27 (continued). Summary of S3 and $S4_a$ crenulation cleavage characteristics (further explanation in text). d: S3 crenulation cleavage; differentiated type (E. Björkfjället, 6e, crossed nicols). e: $S4_a$ crenulation cleavage; "imposed slaty cleavage" type (Vojtjajaure kapell, 4f, crossed nicols). f: S3 crenulation cleavage (E. Björkfjället, 6e, crossed nicols).

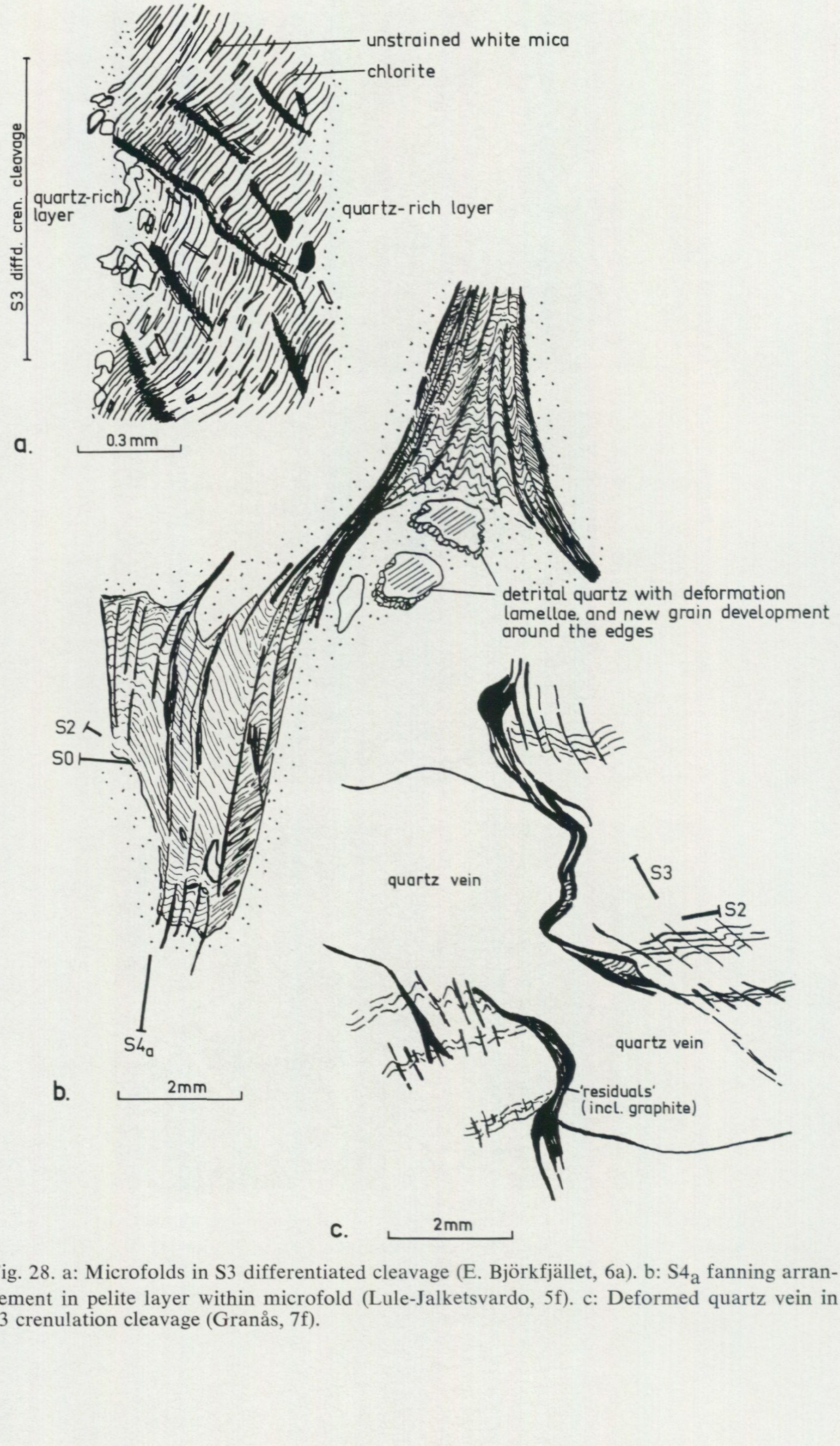


Fig. 28. a: Microfolds in S3 differentiated cleavage (E. Björkfjället, 6a). b: $S4_a$ fanning arrangement in pelite layer within microfold (Lule-Jalketsvardo, 5f). c: Deformed quartz vein in S3 crenulation cleavage (Granås, 7f).

Depletion of quartz in the mica-rich zones (remaining quartz grains strung out and highly irregular in shape) is consistent with the compositional variation in the S3 and $S4_a$ crenulation cleavages of the present study (Fig. 27c). In Fig. 28c one of the better developed, irregular, crenulation cleavage surfaces (S3), marked by a concentration of graphite, failed to penetrate and instead deflected along the contacts of a folded quartz vein (pre-D3) within the layered quartz-graphite phyllite. It is precisely within this region that the quartz vein is reduced in thickness. Again it is suggested that there has been redistribution of silica during formation of the S3 cleavage. It is possible, then, that there has been no slip at all on individual crenulation cleavage surfaces, and their fault-like character (Point 5 above) is purely apparent due to the actual removal of material in the cleavage zones (see also Trouw 1973).

It is considered that mechanical rotation of the S2 phyllosilicates and subsequent transfer of material — particularly the more mobile phases, quartz and calcite — from microfold limb areas by either grain boundary diffusion or solution transport processes are the dominant mechanisms in the formation of the S3 and $S4_a$ crenulation cleavages of the low-grade rocks; new grain growth plays only a minor role. Favourable orientations for transfer of material appear to be governed by the angle the quartz-mica interfaces make with the compressive stresses; interfaces at high angles to the compressive stresses appear more favourable sites for the removal of material. This is consistent with the location of the cleavage zones along the short limbs of asymmetric microfolds and along both limbs of symmetric microfolds. It is thought that the presence of a pore fluid is crucial — either in promoting the rate of grain boundary diffusion (Spry 1969, p. 14) or acting as a solvent in solution-redeposition reactions. Presumably the rocks were relatively consolidated by the time of formation of S2 and it has been noted earlier that the metamorphism reached its peak prior to and during S2. It is thought unlikely, therefore, that the fluid required for the above processes to operate was predominantly connate in origin. However, the reactions which form the principal phyllosilicates that to a large extent define the S2 cleavage and schistosity do release important quantities of water (Holland and Lambert 1969). General migration upwards of this water coupled with low to medium strain rates would provide the optimum conditions for the development of later, variably differentiated crenulation cleavages, associated with minor symmetric and asymmetric buckle folds, in the structurally higher levels of the sequence.

FOLDING MECHANISMS OF THE F2, F3 AND F4 STRUCTURES

ANALYSIS OF LAYER THICKNESS VARIATION IN F2, F3 AND $F4_a$ FOLDS

F2

F2 folds are uncommon, invariably tight to isoclinal in shape with occasional shear discontinuities on their limbs, approximate closely the geometry of similar folds (Ramsay 1962b) and contain S2 as a penetrative axial surface foliation. Their axial orientation is parallel to the maximum elongation direction in S2. Dip isogon (Elliott 1965) data and the corresponding T'_α v α (Ramsay 1962b, 1967, p. 366 and Plate IV, 1) plots for three such folds are summarized in Plate IV, 2a—c. The composition of the various layers is indicated on Plate IV. The characteristic features of the T'_α v α plots are as follows:

1. The maximum dip (α) on the fold limbs is commonly 80—90° and is invariably >70°.
2. The relatively incompetent layers either plot close to the locus $T'_\alpha = 1$ (Class 2 of Ramsay 1967, p. 365) or lie in the field indicating *lower* values of T'_α on the limb of the fold (Class 3).
3. The relatively competent layers partially resemble the above in that they lie close to the locus $T'_\alpha = 1$ but at relatively high values of α they deflect into the Class 1C field.

The tendency towards the Class 1C type of folded surface in the relatively competent layers (and thus *higher* values of T'_α on the limbs of the fold), together with the contrast in detailed geometry in layers of different competence and the overall approximation to the similar fold (Class 2) model are the principal features of these minor structures.

F3

The overturned F3 major structures exert an important control on the present disposition of the lithologies and associated minor folds are particularly abundant near the major F3 hinge zones. These folds deform S0 and S2 and are associated with the first phase crenulation cleavage (S3), which locally transposes the earlier foliation, and significant thrusting and mylonite formation. Measurements on relatively more competent and incompetent layers (Plate IV, 3a—e) in minor folds deforming both low-grade and higher grade rocks were carried out. The main results are as follows:

1. The maximum angle of dip (α) on the fold limbs rarely exceeds 80° and is commonly in the range 60—80°.
2. The relatively incompetent layers plot close to the locus $T'_\alpha = 1$ (Class 2); however, they strictly display Class 1C or compound Class 1C-3 characteristics with T'_α generally increasing sharply near α_{max} on the fold limbs.
3. The more competent layers fall into the Class 1C field, the locus T'_α v α actually lying nearer to the curve $T'_\alpha = \sec \alpha$ (parallel fold model, Class 1B) than $T'_\alpha = 1$.
4. Within the limited scope of the present analysis, there is no obvious difference between the fold shape properties of the F3 structures in the low- and higher grade rocks.

Compared with the F2 folds, the most important difference is in the relatively competent layers; their shape in the F3 structures lies closer to the parallel rather than the similar fold geometry. There is a more pronounced influence of Class 1C geometry in the incompetent layers, although, like the F2 structures, they still approximate to the Class 2 type.

$F4_a$

The northerly trending Tärna Synform and complementary Formliden Antiform ($F4_a$) and the WNW-trending Göuta Antiform ($F4_b$), with relatively steep axial surfaces, control the present saddle-shaped outcrop pattern. The shape of individual layers in these folds changes rapidly along the plunge of the particular major structure (Plate I). There is an acme of development of associated minor folds in the axial zones of the major folds and particularly in the tightened and thickened zones in the southern part of the Tärna Synform and the northern part of the Formliden Antiform. Here, also, there is local transposition of S0/S2 parallel to the second phase crenulation cleavage ($S4_a$), of similar mechanical significance to the earlier S3 crenulation cleavage; the $F4_a$ minor folds, which are generally open elsewhere, become close in shape and possess well-defined zones of contact strain (Ramberg 1961) around relatively more "viscous" layers. Results from the investigation of the shape of individual folded layers in such tightened folds from the two areas mentioned above and also in two open folds ($F4_a$) from Laxfjället are now summarized (Plate IV, 4a—g):

1. α_{max} on the limbs of the tighter folds is approximately 70°, whilst in the open folds from Laxfjället it is 50°.
2. The layer thickness variation in the relatively incompetent units is clearly distinct from that in the more competent layers; the T'_α v α plots in the former either lie in the Class 3 field or plot along an extreme Class 1C locus.
3. The competent layers display very regular plots on the diagram; the loci generally lie in the Class 1C field but commonly approximate to the Class 1B parallel fold model.

The $F4_a$ minor folds are somewhat similar to the F3 structures as far as the detailed geometry of the folded layers is concerned. However, there is a definite resemblance of the shape of certain competent layers to the parallel fold model.

DEFORMED LINEATION PATTERNS IN F3 AND F4 FOLDS

The departure of the shape of F3 and $F4_a$ folds from a strictly parallel form is indicated from the loci of deformed L2 and L2/F3 lineations in certain F3 and $F4_a$ folds respectively (Fig. 29 a, c; see also Table 3). In particular, the deformed L2 lineation loci in the F3 folds (Fig. 29a) approximate to a partial great circle.

Ramsay (1960) has emphasized the importance of planar deformed lineation loci for recognising ideal similar folds but approximately the same pattern can be formed in similar-type folds with buckling and later, strong simple shear components (Ramsay 1967, p. 482). It appears from Fig. 29a that the shearing direction plunges NNE at approximately 30°. Assuming that the orientation of the F3 folds prior to folding by the Tärna Synform was closer to NE—SW, then the original orientation of the shearing direction was approximately NW—SE.

By contrast, the deformed lineation loci in the $F4_b$ and $F4_{kink}$ folds maintain a relatively constant angle with the later fold axis (Fig. 29b), indicating that the $F4_b$ and $F4_{kink}$ folds closely approximate the parallel form.

MECHANISM OF FORMATION OF F2 FOLDS

Whereas in the F3 and $F4_a$ folds certain aspects of parallel and similar folding are variably preserved, the F2 structures resemble the ideal similar fold geometry not only as regards constancy of profile shape and thinning of individual layers but also in the detailed shape of the folded surfaces. However, there is a clear departure from the ideal geometry in the more competent layers, the T'_α v α locus deflecting significantly into the Class 1C field at high values of α. This pattern has been recorded for naturally occurring similar-type folds in other areas (Ramsay 1967, p. 432; Anhauesser 1969; Tobisch and Glover 1971; Hudleston 1973) and has led to the hypothesis that they result by strong, progressive deformation of initially open buckle folds (Flinn 1962; Ramsay 1962b; Hageskov 1972; Hudleston 1973).

In this regard, de Sitter (1958) investigated the fold shape of parallel flexural folds and suggested that there is a limit to shortening by the process of buckling alone. For simple flexural folds, where the outer layers slip over the inner ones towards the hinge zones, this limit occurs when shortening of the original length of the bed reaches 36 %. Assuming the fold does not then rupture, further compressive strain is thought to be absorbed by flattening (relatively more homogeneous strain) of the original buckle fold and the fold shape is modified from the parallel form (Class 1B) to the so-called flattened parallel form with characteristic Class 1C shape (Ramsay 1962b).

Since the regional foliation (S2) and lineation fabric is associated with the F2 folds, it is concluded that there is a close relationship between the most intense relatively homogeneous strain, which accounts for the S2 fabric, and those folds which approximate most closely to the similar fold model. It is suggested, then, that the relatively more homogeneous D2 strain succeeded the initiation of probably open, buckle-type folds associated with a more inhomogeneous stage in the deformation. Buckling, flattening and minor shearing then occurred simultaneously (Hudleston 1973) and, during progressive deformation, the shape and more significantly, the orientation of the early folds may have been modified.

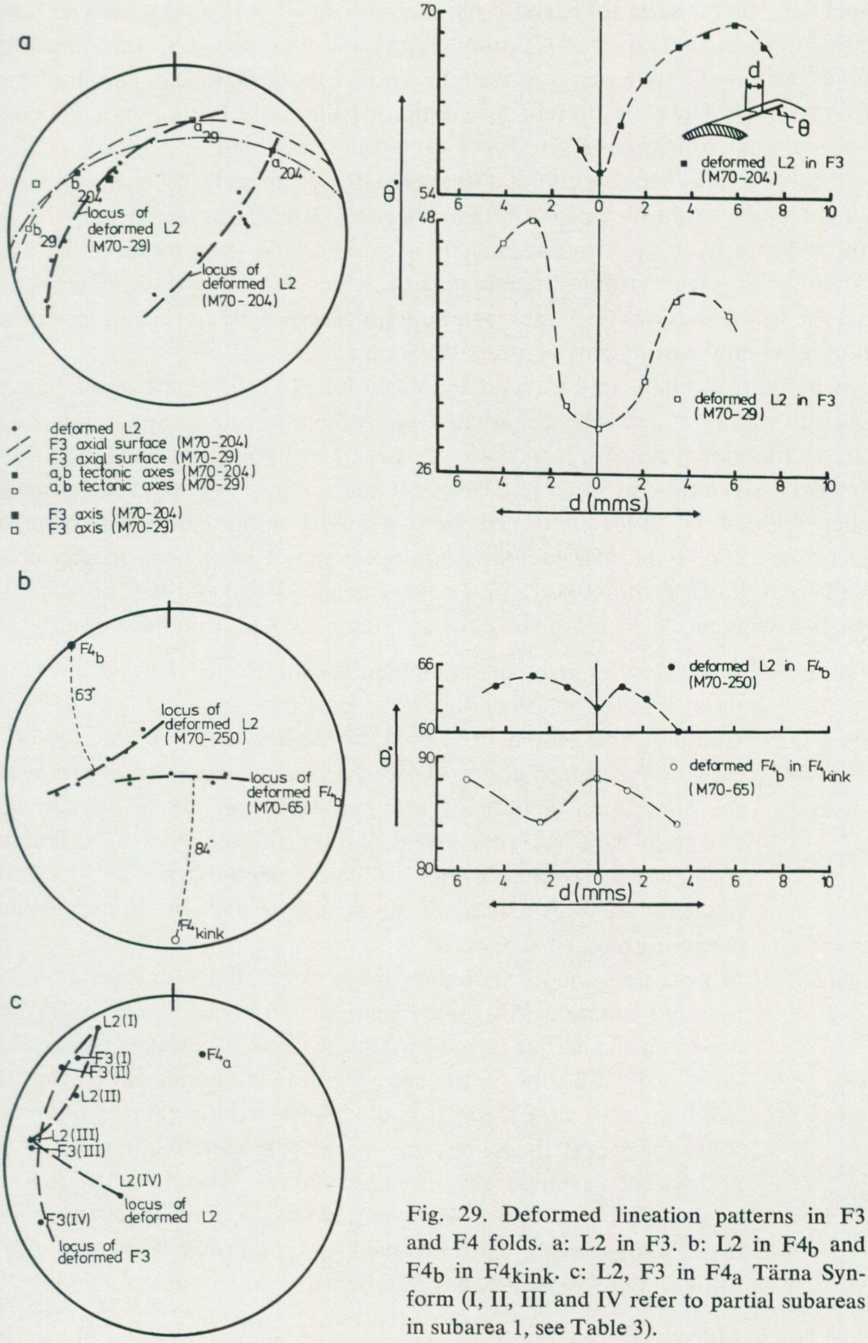


Fig. 29. Deformed lineation patterns in F3 and F4 folds. a: L2 in F3. b: L2 in $F4_b$ and $F4_b$ in $F4_{kink}$. c: L2, F3 in $F4_a$ Tärna Synform (I, II, III and IV refer to partial subareas in subarea 1, see Table 3).

Isoclinal, similar-style folds could have been formed, and fold axes which originally lay at an angle to the maximum elongation direction in S2 could have been rotated within the fold axial surface into parallelism with the elongation direction (Flinn 1965). Thus the present orientation of the early folds is not necessarily their original orientation. The finite D2 strain, recorded and measured in the deformed conglomerate pebbles (Stephens 1975), appears to be constrictional but the early strain history is obscure. It is tentatively suggested that the folds initiated in a strongly constrictional situation and were then modified by a later flattening or plane strain situation during which S2 formed. It is considered that the whole deformation was essentially progressive, being related to increased shortening on the principal compressive strain axis.

It is inferred, then, that F1 and the early buckle fold phase (more inhomogeneous strain) of D2 may be broadly equivalent to one another, this folding having initiated prior to formation of the S2 foliation and lineation fabric. However, since S2 cuts obliquely from one limb, across the axial surface, to the other limb of F1 folds described from adjacent areas (pre-F1 structures of Zachrisson 1969), the incremental strain axes cannot have been strictly coaxial in all areas during protracted F1-F2 folding (see also Ramsay 1965; Powell 1974). The following model is put forward for progressive F1-F2 fold development.

D1: Early development of recumbent, buckle folds (F1) with fold axes parallel to the length of the orogenic belt.

Early D2: Continued development of buckle-type folds (F2). Where the D1-D2 progressive deformation followed a coaxial incremental strain path, the orientation of both F1 and early D2 folds would be coincident and their axial surfaces would parallel the incipient S2 orientation. In areas where some rotation of incremental strain axes occurred, the early D2 folds would deviate in axial surface orientation from the earlier formed F1 structures.

Late D2: More homogeneous flattening strain associated with formation of the regional foliation (S2) and lineation fabric. S2 would be superimposed on the earlier formed folds such that it parallels the axial surfaces of F1-F2 folds in the coaxial strain situation, but transects the axial surfaces of F1 folds in the non-coaxial strain situation (any folds formed at the same time as the regional foliation would also possess S2 as an axial surface structure). This late D2 flattening strain, combined with shearing parallel to the limbs, could have modified the shape of the earlier formed folds and rotated the axes of those folds whose axial surface is parallel to S2 as discussed above.

The coaxial strain situation is thought to apply more closely in the Tärna—Björkvattnet area where the early minor folds all have S2 as an axial surface structure and where, consequently, it is not possible to distinguish F1 from F2 minor structures. However, the evidence from Sandnäsbacken does suggest that S2 cuts across, albeit at a small angle, the limbs of the major F1 syncline and, thus, the incremental strain axes are strictly non-coaxial. The high degree of proximity between the early minor fold axes and the maximum elongation direction in S2 is consistent with the presumed constrictional nature and high strain values of the D2 finite strain (see general discussion by Sanderson 1973).

The non-coaxial strain situation is thought to apply to those areas mentioned by Zachrisson (1969) where F1 and F2 minor folds are discordant; the former are transected by the regional cleavage, while the latter have the regional cleavage as an axial surface structure.

The development of early minor folds, then, is considered to be controlled by two important factors:

1. The progressive nature of the deformation from D1 to D2 and
2. The formation of the regional foliation at a relatively late stage in the D1-D2 deformation history, such that it is superimposed on the earlier formed folds and transects minor fold axial surfaces where the incremental strain axes have followed a distinctly non-coaxial strain path.

Since S2 in the model above is initiated late in the D1-D2 deformation history, the relative orientation of S2 and the D2 finite strain ellipsoid will largely depend on the earlier D1-D2 strain history. If the strain path is coaxial then S2 may well be oriented perpendicular to the direction of greatest finite (D2) shortening in the rock. However, if the strain is non-coaxial, i.e. if the axes of the incremental strain ellipsoid do not parallel the finite strain ellipsoid axes at the time of foliation development (late D2), then S2 will be oriented in some oblique orientation to the direction of greatest finite (D2) shortening. It would appear that both these relationships between the S2 foliation and the D2 finite strain ellipsoid may exist in the Västerbotten Caledonides.

MECHANISM OF FORMATION OF F3 AND F4 FOLDS

Dip isogon analysis of the F3 and particularly the $F4_a$ folds demonstrates the common Class IC shape and occasional preservation of nearly parallel form in the folded competent layers. From the previous discussion it may be inferred that these folds initiated by a buckling instability within the competent layers but their shape has subsequently been modified. In the minor folds analysed it would appear that the degree of modification is less in the $F4_a$ compared to the F3 folds. However, both sets of folds show substantially less modification than the F2 folds discussed earlier.

The deformed lineation data suggest a strong simple shear component in the formation of the F3 folds. Since the F3 folds deviate quite considerably from the ideal similar fold geometry, however, the buckling and superimposed strong simple shear model is preferred as a mechanism for their formation. This hypothesis gains some support from the broad association of F3 folds with post-metamorphic mylonitization and thrusting phenomena, the mylonite and thrust zones being interpreted as zones of exceptionally high shear strain. Since the pre-Tärna Synform orientation of these folds was possibly closer to NE—SW with easterly sense of overturn (i.e. axial orientation along rather than across the trend of the orogenic belt) and, thus, broadly similar to the original orientation of the F1-F2 folds in the model adopted above, initial fold axis and axial surface orientations appear to have been relatively static from D1 to the end of D3.

The trend towards an increased influence of parallel folding characteristics observed in the $F4_a$ folds, is completed in the $F4_b$ and $F4_{kink}$ structures which closely approximate the parallel fold model. It is inferred that there has been little or no modification of the shape of the original buckles by either simultaneous or later flattening and/or shearing deformation.

ANALYSIS OF THE LATE-STAGE KINK FOLDS ($F4_{kink}$)

The kink folds ($F4_{kink}$) are developed exclusively on the steep eastern limb of the $F4_a$ Tärna Synform. A detailed geometrical analysis of the kinks to estimate the stress field orientation which led to their development is presented here.

The average attitude of the two sets of axial surfaces ($S4_{kink}$ and $S4'_{kink}$) have been plotted in Fig. 30 by determining the vector means of the two clusters of axial surface data. The great circle to the concentration maximum of the S0/S2 data in subarea 2 (see Plate III, 2a), which is approximately the average orientation of S0/S2 on the eastern limb of the Tärna Synform, has also been plotted.

The average trends of the conjugate pair of fold axes ($F4_{kink}$ and $F4'_{kink}$) have been constructed from the S0/S2-$S4_{kink}$ and S0/S2-$S4'_{kink}$ intersections respectively. The symmetry of these folds is close to orthorhombic. Thus, by employing the method outlined by Ramsay (1962a) and noting that the obtuse angle between the two axial surfaces faces the direction of maximum shortening, the principal stress axes have been located. The following is inferred regarding particularly the symmetry and stress field orientation of the kinks:

1. The relationship between the obtuse angle defined by the conjugate pair of axial surfaces and the direction of maximum shortening is similar to that recorded in the experiments of, for example, Paterson and Weiss (1966). However, this angle is approximately 140° in the present study and somewhat higher than the experimentally determined values of 120°.
2. The maximum principal stress axis (σ_1) lies very close to the foliation which it deforms but strictly these structures possess an overall monoclinic symmetry. σ_1 plunges steeply and is transmitted approximately down the steeply dipping S0/S2 foliation.

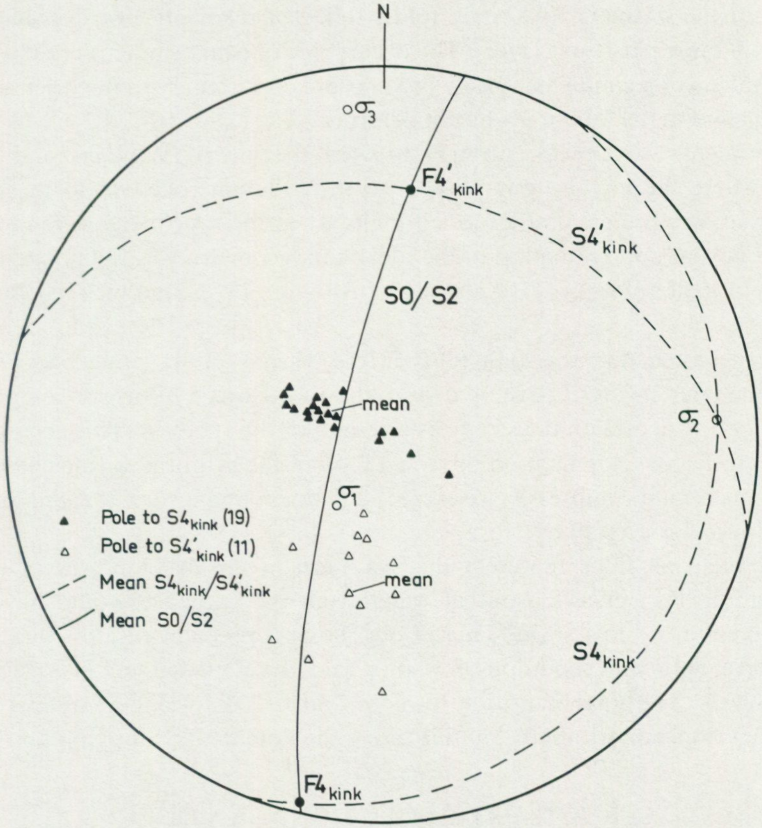


Fig. 30. Stress field orientation of the kink folds on the eastern limb of the Tärna Synform.

3. The kink axial surfaces formed oblique to σ_1 and the direction of maximum compressive strain. It is, however, uncertain that they developed along the surfaces of maximum shearing stress in a primary, conjugate shear system (simple shear model of Dewey 1965) since again it is the obtuse angle which faces the direction of maximum shortening (see also Ramsay 1967, p. 452—453 and Roberts 1971b).

Cobbold et al. (1971), from experiments with plasticine models, have stressed the importance of both the degree of anisotropy and the angle between the compression direction and the initial layering in the development of a variety of *related* structures including sinusoidal buckle folds, kink bands, conjugate kink folds and boudins. The anisotropy is related to variations in the stress/strain properties in particular directions rather than differences in the rheological properties (competence contrast) of the different layers. The near symmetric sinusoidal folds in, for example, the more calcareous phyllites, which are obviously related to the ordinary kink style folds, reflect a relatively low degree of anisotropy in these particular layers. The occurrence of near symmetric folds and the two members of a conjugate system of kinks are consistent with the orientation of σ_1 being close to the foliation which it deforms.

Experiments on already strongly foliated specimens (Williams and Means 1971), where the loading was parallel to the foliation (relevant to the present field situation), preferentially formed kinks when the specimens were deformed dry; the closest approximation to the ideal kink geometry was obtained in specimens deformed at low (1×10^3 compared to $1—3 \times 10^4$ pound inch $^{-2}$) confining pressure.

It is suggested that the kink folds in the Tärna—Björkvattnet area formed during the later stages of development of the $F4_a$ Tärna Synform as a gravitational collapse structure on the steep eastern limb of the synform. This accords with their exclusive development on the steep limb of the synform, a calculated near-vertical maximum compressive stress (σ_1) and the dominance of the kink member with a down-dip sense of overturn.

Since the kink folds represent the last stage in the development of the D4 deformation, the squeezing out of water from the sequence would have been at a maximum by this stage. Thus, from the experimental results cited above, there were optimum conditions for kink fold development in the structurally higher levels. The final transition to more "brittle" deformation style is shown by the development of minor faulting along the dominant, E-dipping set of axial surfaces.

STRUCTURAL EVOLUTION — A MODEL

The occurrence of various linear and planar structures associated with minor folds and the refolding relationships, on both a major and minor scale, provide the evidence for a definite time sequence in the development of the structures within the Björkvattnet Unit. However, consideration of the F1-F2 fold model and the orientation and mechanisms of formation of the F3 (and F4) folds suggest that the deformation was not episodic but progressive in character; the development of a particular foliation and/or refolding relationship are markers along the continuous deformation path. Any generalized model for the structural evolution of the Björkvattnet Unit must take into account this time sequence and yet make sense in terms of a progressive strain interpretation; it should also seek to explain the rather varied fold orientations. The following model is suggested from the interpretations of the structural sequence and the folding mechanisms presented in earlier sections:

Stage I (D1-Early D2)	Development of a pile of recumbent folds which face eastwards and trend parallel to the orogenic belt; a constrictional strain situation with extension on the limbs of the folds at a high angle to the trend of the orogenic belt.

Stage II (Late D2) — Collapse of the tectonic pile leading to considerable modification of the shape of the earlier formed folds and the development of the regional foliation and lineation fabric; a flattening or plane strain situation related to increased shortening on the principal compressive strain axis (due to progressive collapse of the tectonic pile) and with the principal elongation direction remaining at a high angle to the trend of the orogenic belt.

Stage III (Latest D2-D3) — Development of mylonite zones and major thrusts at a late stage in the collapse of the tectonic pile associated with continued, easterly overturned folding and the development of locally intense crenulation cleavage at higher structural levels. This is the stage of substantial laterial transport involving emplacement of the tectonic pile over the "granitic" basement of the Baltoscandian platform as well as overturned folding and thrusting of basement and cover together. The strain during this stage involved a large simple shear component oriented at high angles to the trend of the orogenic belt. The shear strain was inhomogeneously distributed throughout the tectonic pile, the maximum shear strain being presumably located in the more important mylonite and thrust zones (see also Zwart 1974).

Stage IV (D4) — Development of late antiforms and synforms with steep axial surfaces and locally strong crenulation cleavage, possibly related to gravitational disturbances due to the tectonic superimposition of higher density cover on top of lower density, "granitic" basement.

Stage V (Late D4) — Gravitational collapse of the steepened limb of the late Tärna Synform at higher structural levels, producing the kink folds with flat-lying axial surfaces and largely down-dip sense of overturn.

Besides reorientation of earlier fold axes and axial surfaces in later folds, particularly the Stage IV antiforms and synforms, variation in the orientation of fold axes and/or axial surfaces is thought to be a result of one or more of the following processes:

1. Non-coaxial strain paths during Stage I.
2. Rotation during Stage II of fold axes lying within the regional foliation towards the maximum elongation direction.
3. Possible rotation of fold axes during Stage III towards the maximum elongation direction associated with the simple shear deformation; rotation of this type would be localized in the areas of maximum shear strain i.e. the more important mylonite and thrust zones.

In summary, the structural sequence reflects the progressive deformation associated with emplacement of the allochthonous rocks of the Björkvattnet Unit (Seve-Köli Nappe Complex), from the fold nappe stage through the collapse and

TABLE 4. Summary of the deformation sequence in the Björkvattnet Unit and tentative correlation with the schemes of Zachrisson (1969) and Trouw (1973).

		Low-grade (Köli) Rocks (Subareas 1-3)	High-grade (Seve) Rocks (Subarea 4)
Gravity-controlled disturbance of basement/cover	D4 S^2_{cc} (northerly orientation) = $S4_a$	Development of late kink folds	
		Development of major open to gentle folds. Intense conc. of minor structures in certain restricted zones inc. S^2_{cc} (locally transposes S_{cl}).	Development of late antiformal arch and related minor folds. Weak development of crenulation cleavage at this stage only
		Kink folds ($F4_{kink}$) follow S2 trend on gently dipping axial surfaces. Minor folds plunge gently N-S on steep axial surfaces ($F4_a$). Rare, steep WNW-plunging crenulation ($F4_b$).	Minor folds generally plunge WNW but subordinate N-NE plunging folds also present; steep axial surfaces.
Thrusting	D3 S^1_{cc} (Low-grade only) = S3	Intense deformation of S_{cl} etc. in certain zones - continuous development of both major and minor folds and important thrusts. Development of S^1_{cc} (locally transposes S_{cl}).	Deformation of S_s etc. by tight to isoclinal major and minor folds with associated slides.
		Minor folds plunge gently and trend approximately parallel to S2 strike (original orientation = NE-SW?). Major folds are overturned to the SW-S (original orientation = SE?).	Minor folds plunge gently and trend approximately parallel to S2 strike. Major folds are overturned to the NE-E.
Collapse of pile	D2 $S_s = S_{cl}$ = S2	Development of regional phyllitic cleavage (S_{cl}) and penetrative lineations inc. pebble elongation. Associated with occasional minor folds.	Development of regional schistosity (S_s), penetrative mineral lineations and rodding. Associated with occasional minor folds.
Tectonic pile established		Lineations plunge NW-WNW.	Lineations plunge WNW.
	D1	Early synclinal repetition of stratigraphy.	?
		Uncertain orientation.	

plastic deformation developing into thrust nappe stage to the final post-emplacement stage, which involved gravity-controlled disturbance of both basement and allochthonous cover related to newly established density contrasts. Much of the internal strain within the allochthonous cover rocks studied here occurred prior to the thrusting stage. Furthermore, the reasons for the development of those initial Stage I folds remain obscure.

CONCLUSIONS

The following outlines the more important inferences from the above (see Table 4):

1. The sequence of deformation is similar in the low- and high-grade parts of the Björkvattnet Unit.
2. The earliest deformation (D1) involved the major synclinal repetition of the stratigraphy. This accounts for the occurrence of the same stratigraphic units within the core as around the rim of the later, overturned Ruffevare—Björkfjället Synform Complex; no F1 minor folds, however, have been distinguished in the area. The D1 phase here is correlated with the pre-F1 structures of Zachrisson (1969).

Zachrisson, 1969	Trouw, 1973
F3 - Kink folds on flat-lying axial surfaces.	F4 - Local folding, crenulations on sub-horizontal axial surfaces.
F2 - Development of major antiforms and synforms. NNE-SSW or NE-SW orientation and steep axial surfaces.	F3 - Folding on several scales. Steep axial surfaces with variable orientation.
Associated with F1 or younger = local and possibly more regional low angle thrusts.	F2 - Folding; formation of crenulation cleavage, mineral lineation and locally a mylonitic foliation. Folds plunge gently NNE to NW. Mineral lineation plunges NW to WNW.
F1 - Development of main schistosity (S1). General NW-SE or E-W orientation for related folds.	F1 - Tight folding and formation of penetrative foliation (and lineation in metabasites); deformation of conglomerates.
PRE-F1	

3. Later deformation involved firstly a phase of relatively more homogeneous strain (D2) which gave rise to the regional, penetrative foliation (S2) and lineation fabric including the intense NW—WNW pebble stretching; minor, isoclinal folds — the earliest minor folds recognised in the area — are associated with the S2 foliation but the axial orientation of these folds (Table 4) is parallel to the main elongation direction in S2 and at a high angle to the trend of the orogenic belt.
4. The D2 structures are folded by an important phase of major and minor folding which plunges gently and approximately follows the strike of S2 in both the low- and high-grade rocks (original orientation possibly closer to NE—SW). The major folds are overturned towards the S to SW in the low-grade rocks and towards the NE in the high-grade rocks (difference due to later deformation), and are accompanied by locally intense crenulation cleavage (S3) development in the low-grade rocks. This earlier, post-S2 phase (D3) is also associated with significant thrusting and mylonite formation.
5. The last phase (D4) is composite and accounts for the present outcrop pattern. Upright, northerly ($F4_a$) and cross-cutting WNW-plunging ($F4_b$) major structures control the overall saddle-shaped outcrop pattern; abundant minor folds and a locally penetrative crenulation cleavage

($S4_a$) are associated with the northerly structure. Late kinks and symmetrical buckle folds with gently dipping axial surfaces are developed on the steep eastern limb of the $F4_a$ Tärna Synform. The maximum compressive stress axis during the formation of the kinks appears to have been transmitted down the steep foliation and the overall symmetry of these structures is close to orthorhombic.

6. It is suggested that the regional foliation (S2) in both the low- and high-grade rocks developed by different mechanisms depending on rock type and position in the sequence. Intracrystalline deformation, recovery and recrystallization mechanisms appear important in the more quartz-rich rocks, whilst syntectonic grain growth mechanisms are important in the pelitic and basic rocks.
7. The later S3 and $S4_a$ crenulation cleavages are semi-penetrative, compositionally heterogeneous and grade into an "imposed slaty cleavage". It is considered that the dominant mechanisms involved in their formation are rotation of the S2 phyllosilicates accompanied by redistribution of particularly quartz by fluid assisted grain boundary diffusion or solution transport processes. It is suggested that migration of water released from the reactions which form the phyllosilicates in S2, coupled with low to medium strain rates, provided the best conditions for crenulation cleavage development in the structurally higher levels of the sequence.
8. A morphological study of minor folds suggests the progressive decrease from D2 to D4 in the importance of flattening and shearing strains relative to buckling in the fold development. The F2 minor folds, therefore, depart most acutely from a parallel fold geometry but approximate closely to the similar fold model.
9. It is suggested that the F1 and F2 folds are broadly equivalent to one another, being related to a progressive deformation sequence from D1 to D2. It is considered that the late D2, relatively more homogeneous strain, which succeeded initiation of the F1-F2 folding, not only modified the shape of the folds but also could have rotated fold axes lying within S2 from a direction closer to the intermediate axis of the D2 strain ellipsoid into parallelism with the maximum elongation direction in S2. Formation of the S2 foliation as an event during the development of the folding implies that the relationship between S2, the axial surfaces of the early (F1 and F2) folds and the direction of greatest finite (D2) shortening in the rock depends on the nature of the D1-D2 progressive strain path. A coaxial strain path leads to superimposition of the S2 cleavage parallel to the axial surfaces of the earlier formed folds with consequent shape and orientation modification as described above; in this situation S2 may well be perpendicular to the direction of greatest finite (D2) shortening. A coaxial strain path is thought to apply more closely in the Tärna—Björk-

vattnet area where S2 is an axial surface structure to the early minor folds. Furthermore, the overall constrictional nature of the D2 finite strain ellipsoid and the high strain values in the Tärna—Björkvattnet area are consistent with the development here of early fold axes oblique to the trend of the orogenic belt. A non-coaxial strain path leads to both transection by the S2 cleavage of some of the early fold cores (F1) and an axial surface relationship with others (F2); in this situation it would be possible to distinguish F1 and F2 folds and the S2 cleavage would be oriented oblique to the direction of greatest finite (D2) shortening.

10. The preservation of zones of contact strain (Ramberg 1961) and cleavage fans in the F3 and F4 structures confirm the increased influence of buckling mechanisms in their development. The near-planar deformed lineation patterns in F3 folds, however, suggest an important component of simple shear in their development.
11. The occurrence of late kink folds etc. on the steep eastern limb of the Tärna Synform is thought to be related to the gravitational collapse of this limb; this is in agreement with the near-vertical σ_1 orientation. The kinks show only parallel fold characteristics. It is suggested that the kink folds formed under relatively dry conditions at low confining pressure.
12. The structural sequence and folding mechanisms are consistent with a model involving progressive deformation during emplacement of the allochthonous Björkvattnet Unit rocks on top of the Baltoscandian platform. The model involves an initial fold nappe stage (D1-D2), an intermediate stage involving collapse, plastic deformation and subsequent thrust nappe tectonics (D2-D3), and a final stage involving gravity-controlled disturbance of basement and cover as an adjustment to newly established density contrasts (D4).

METAMORPHISM-DEFORMATION RELATIONSHIPS IN THE BJÖRKVATTNET UNIT

INTRODUCTION

The morphological nature of the different cleavages (S2, S3 and $S4_a$) in the low-grade, Köli rocks and the regional schistosity in the high-grade, Seve rocks has already been discussed. As far as mineral growth and recrystallization are concerned, the formation of new quartz grains and the growth of mica, chlorite and amphibole are important factors in the formation of the regional, penetrative foliation (S2) but are of minor importance during formation of the later crenulation cleavages.

Data on the relationship between poikiloblast growth and the deformation sequence, as defined by the tectonic foliations and associated minor folds, are presented here. The main poikiloblasts include biotite, garnet, hornblende, albite and muscovite which occur in the higher grade rocks of the Björkvattnet Unit. The distribution of metamorphic zones throughout the Tärna—Björkvattnet area is summarized in Fig. 31.

POIKILOBLAST-S2 RELATIONSHIPS

BIOTITE

Besides the growth of biotite parallel to S2, coarse biotite porphyroblasts, commonly 2—4 mm long and varying in colour from deep bottle-green to brown, occur within the Brakko and Tärna schists. In some specimens these porphyroblasts are randomly oriented but generally they are lenticular in shape and are contained in S2, being crudely aligned parallel to the L2 mica lineation. They are enclosed by muscovite and finer biotite which define the S2 schistosity (see Fig. 25d). Late-stage biotite also replaces hornblende particularly in the layered quartz-epidote schists and in the impure epidote-rich marbles.

GARNET

Garnet is commonly several millimetres across but in rare skarn-type assemblages varies up to 10 cm across. Garnet shapes are variable; in the more psammitic layers they are commonly xenoblastic, whilst in the more mica-rich layers the boundaries are sharper giving rise to subidioblastic shapes. Occasionally the garnets are surrounded by leucocratic, quartz-rich rims and/or larger quartz grains which appear to be replacing the garnet peripherally. The growth relationships between the garnets and S2 are clarified by a study of the internal inclusion trail (S_i) within the garnet and its relationship to the external schistosity trace (S_e). S_i is most commonly composed of dimensionally oriented quartz grains, whilst S_e is generally equivalent to S2. Six different geometries are now described; types 1, 2 and 6 are by far the most common;

1. These garnets (Fig. 32a) contain quartz inclusions but no distinct inclusion trail. They are fractured and elongate parallel to the L2 mineral lineation with quartz eyes developed in this direction. There is a sharp deflection of the S2 micas around the garnets with paucity of mica in the zones adjacent to the garnet in the direction of the L2 lineation. The garnets do not transect the micas of the S2 schistosity in what would be a variably deflected mica fabric, which, according to Misch (1971), suggests the combined effects of constant-volume replacement and pushing-out of the pre-existing S-surfaces under post-schistosity growth conditions. Instead, the microstructure appears to be

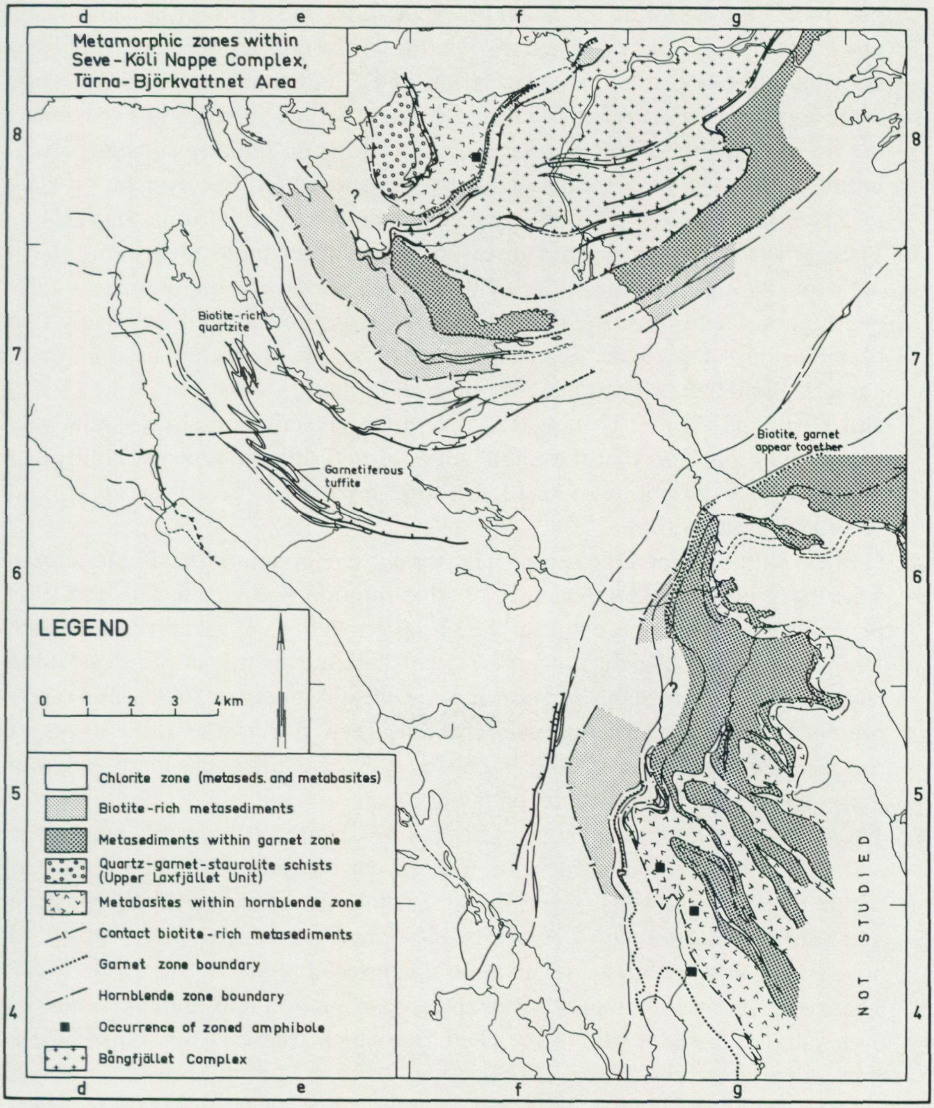


Fig. 31. Summary of metamorphic zones in the Björkvattnet and Laxfjället Units.

the result of deflection of S2 around pre-existing porphyroblasts (Zwart 1962). Another important feature is the smaller grain size of the quartz inclusions within the porphyroblast relative to the quartz outside, where the principal fabric coarsening was probably syn-S2.

2. These garnets display the same deformation and mica deflection characteristics as above but the quartz inclusions, which are again finer-grained than the matrix quartz, are strongly oriented (Fig. 32b, c). In all sections nor-

mal to S2 the S_i fabric is planar (straight line trace in section) and often discordant to S2. It is suggested that these crystals grew under relatively static conditions prior to the formation of S2 and that the inclusion fabric represents an earlier, relict foliation of uncertain origin. In some specimens the direction of the S_i trace is variable. Assuming an initially constant orientation of S_i, variable amounts of relative rotation, therefore, appear to have occurred between the matrix and the garnet during later deformation (D2).

3. The garnets in this group are rare. They display a radial inclusion pattern with four to six radiating zones of the garnet filled with fine-grained quartz and epidote inclusions, the sectors between these zones being relatively free of inclusions (Fig. 32d); S2 is once again totally deflected around these garnets. Similar patterns have also been described by Rast and Sturt (1957) and Henley (1970). It is suggested that these crystals grew under relatively static conditions and that the radial zones are crystallographically controlled and represent directions of rapid dendritic growth in the crystal (Rast and Sturt 1957; Rast 1965).
4. Garnets with S-shaped inclusion trails are more common in the Tärna schists. They have grown syntectonically, relative rotation between matrix and porphyroblast occurring here during the actual growth of the garnet (Zwart 1960, 1962; Spry 1963).)(-inclusion patterns, which define the axis of paracrystalline rotation within the plane of section (Powell and Treagus 1970), are rarely present (Fig. 33a). These garnets probably grew during the early stages of formation of the complex S2-S3 schistosity which forms the main foliation in the Tärna schists (see discussion later).
5. There are abundant, small (generally 0.1—0.2 mm; <0.7 mm), idioblastic and non-poikiloblastic garnets in thin layers of garnet-hornblende gneiss within the Brakko schists (Fig. 33b). These garnets cut across the S2 foliation and contrast sharply with the garnets described previously. They appear to have grown slowly during a relatively static, post-S2 phase.
6. These garnets display apparently two stages of growth and are characterized by a subidioblastic or idioblastic clear rim which transects S2. Where these rims are present on type 1 garnet cores, there is a similarity in the mica deflection pattern to that described by Misch (1971) for his entirely post-tectonic porphyroblasts. However, the recognition here of a distinct idioblastic rim allows the alternative explanation that the early growth stage was pre-schistosity with later total mica deflection around the garnet during the formation of S2, and then continued post-S2 growth so that the garnet now partially transects the schistosity. The occurrence of the same pattern of mica deflection in garnet poikiloblasts where subidioblastic rims are present on cores with straight and discordant trails (Fig. 33c) supports the above interpretation.

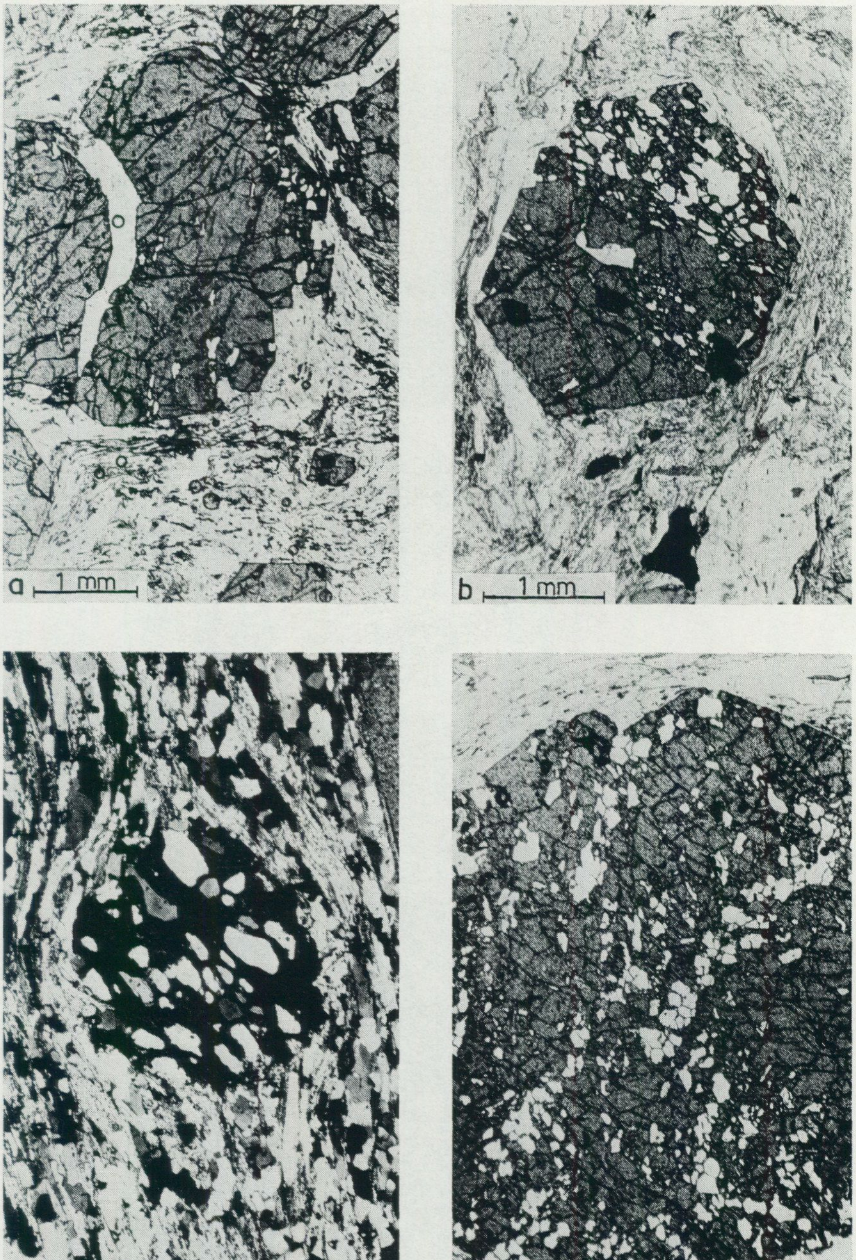


Fig. 32. Pre-S2 garnet poikiloblasts. a: Type 1 (central Brakko, 5g, uncrossed nicols). b: Type 2 (Gakkerbäcken, 6g, uncrossed nicols). c: Type 2 (Gakkerbäcken, 6g, crossed nicols). d: Type 3 (Brakko, 5g, uncrossed nicols).

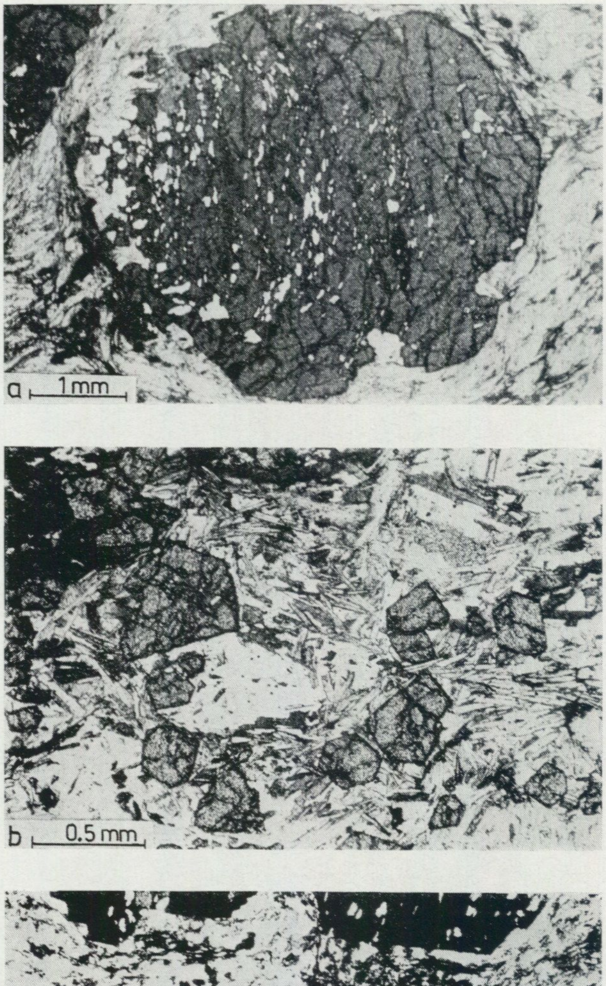


Fig. 33. Syntectonic and post-S2 garnet poikiloblasts. a: Type 4; axis of paracrystalline rotation contained in the plane of the section (NW Göutavardo, 8g, uncrossed nicols). b: Type 5 (eastern Brakko, 5g, uncrossed nicols). c: Type 6; idioblastic garnets with straight and discordant quartz inclusion trail and partial mica deflection, but with non-poikiloblastic rim which transects S2 – initial growth pre-S2, continued growth post-S2 (Gakkerbäcken, 6g, crossed nicols).

HORNBLENDE

The coarse hornblende porphyroblasts, which are often several centimetres long in the quartz-garnet-hornblende schists and hornblende garbenschiefer, are often fractured and boudined in S2, and the schistosity commonly deflects around these crystals (see Fig. 26). The hornblendes are generally arranged in rosette shapes on the S2 surfaces, although some porphyroblasts lie across the S2 foliation. The rosettes are elongated and the coarse hornblendes lie roughly parallel to the penetrative L2 mineral lineation. Occasional hornblendes contain planar quartz inclusion trails (straight line trace in section) which are discordant to S2 (Fig. 34a). More commonly the poikiloblastic core zones of longitudinally oriented crystals contain no distinct inclusion trail and are surrounded by non-poikiloblastic rims (Fig. 34b). The above features suggest that the coarse hornblende porphyroblasts grew prior to the formation of S2, and that they were deformed and rotated, during the D2 deformation, into the regional schistosity.

ALBITE

Poikiloblastic, lenticular albite is restricted to the Tärna schists. The trace of the quartz inclusion fabric in the untwinned albite poikiloblasts is straight, either approximately parallel to the trace of the external foliation (complex S2-S3 schistosity) or markedly discordant to it (Fig. 35). There is clear deflection of this foliation around the albite poikiloblasts. Kieft (1952) also reported albite poikiloblasts with S-shaped inclusion trails.

MUSCOVITE

Porphyroblastic muscovite occurs as highly deformed lenticular grains within the complex S2-S3 schistosity of the Tärna schists, whilst randomly arranged, coarse (1—2 mm across) muscovites cut across the hornblendes in the garbenschiefer rocks (Brakko schists).

RETROGRESSIVE MINERAL REACTIONS

There is a variable development of post-S2 retrogressive mineral reactions which, particularly in the schists rimming the Bångfjället Complex, are associated with lenticularization and transposition of the S2 schistosity during D3. Besides development of retrogressive effects in the Tärna schists, there also appears to be a concentration of these reactions in the Göuta—Forsbäck area, in the zone where there is a relatively rapid increase in metamorphic grade.

The common reactions include the chloritization of the biotite/muscovite schistosity, the development of anhedral masses of chlorite, sericite and calcite

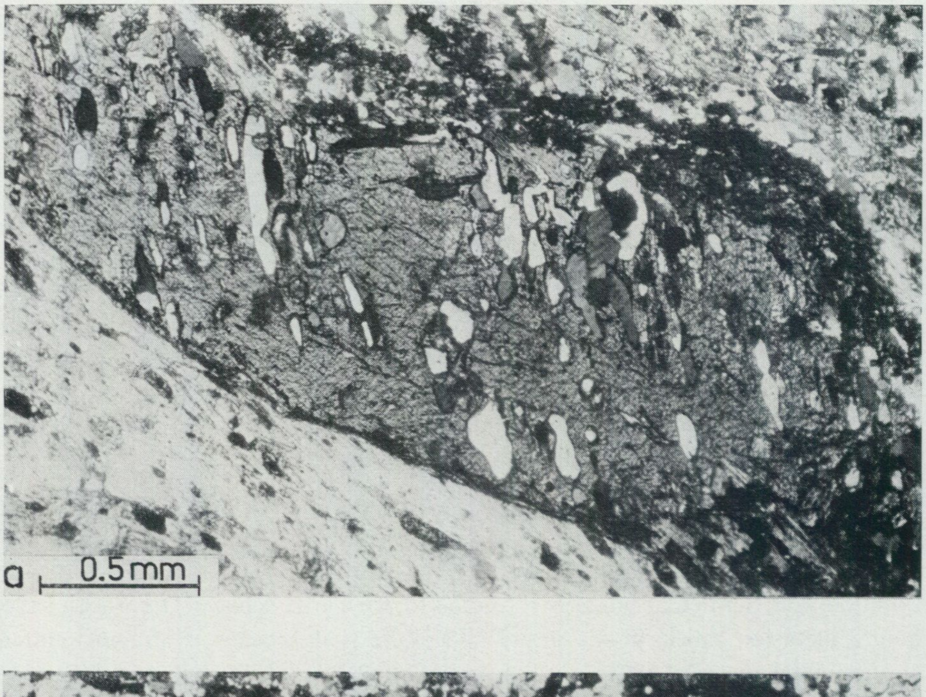


Fig. 34. Hornblende inclusion fabrics. a: Straight and discordant S_i trail with respect to the S2 schistosity (Gakkerbäcken, 6g, crossed nicols). b: Typical fabric with random inclusion pattern in centre of blast and non-poikiloblastic rim (SW Brakko, 4g, crossed nicols).

Fig. 35. Pre-S2 albite poikiloblast with straight and discordant S_i. Note the strongly phyllonitic nature of S_e (= S2/S3) (NW Göutavardo, 8g, crossed nicols).

either peripheral to or throughout garnet porphyroblasts and the breakdown of hornblende to chlorite, biotite, sericite and epidote which are present in varying proportions.

DISCUSSION AND CONCLUSIONS

The data above are summarized in Fig. 36 which shows the relationship between mineral growth and recrystallization and the different deformation phases in the Björkvattnet Unit.

D1-D2 METAMORPHISM-DEFORMATION HISTORY

In the Seve schists a metamorphic peak appears to have been established after the formation of a fine-grained quartz fabric but prior to the formation of the regional, penetrative foliation i.e. post the quartz inclusion fabric but pre-S2. It is suggested that the piling up of major fold nappes in the younger Köli sequence during D1 and early D2 may have contributed to the establishment of high-grade metamorphism of the lower Seve sequence prior to collapse and major transport of the tectonic pile during late D2 and D3; thus the metamorphism reached its peak prior to late D2 and probably after D1-early D2.

The main strain-determining phase (late D2) throughout the Björkvattnet Unit followed establishment of the peak of metamorphism. However, growth of muscovite, biotite, hornblende and garnet during formation of S2 in the schists and amphibolites indicates that relatively high-grade, metamorphic conditions persisted throughout the whole of D2 in these rocks.

D3-D4 METAMORPHISM-DEFORMATION HISTORY

Significant mylonite formation and thrusting are thought to have occurred during D3, at a late stage in the collapse of the tectonic pile. The metamorphic data show that, in the structurally lower parts of the Björkvattnet Unit (Seve rocks), relatively high temperatures persisted after D2. The later garnets (types 5 and 6) are conspicuously smaller and non-poikiloblastic, and occur either as rims on earlier crystals or as new grains. There was, however, a relatively rapid fall-off in metamorphic grade at higher levels (Köli rocks) in the unit. This is indicated by only limited evidence for post-S2 recrystallization of quartz and growth of chlorite and white mica (no biotite). As stated earlier, the dominant mechanism in the formation of the later cleavages in the Köli rocks appears to have been rotation of S2 micas accompanied by redistribution of the more mobile phases such as silica. It is suggested that persistence after D2 of relatively high-grade metamorphic conditions in the Seve rocks effected low ductility contrasts and deformation still proceeded there at relatively low strain rates. This contrasts with tectonically higher levels in the Björkvattnet Unit where the metamorphic peak had waned. It is considered that the D3 (and D4) deformation in the Köli rocks proceeded under relatively faster strain rate conditions and that there was addition of water to the system from chemical reactions during and after D2.

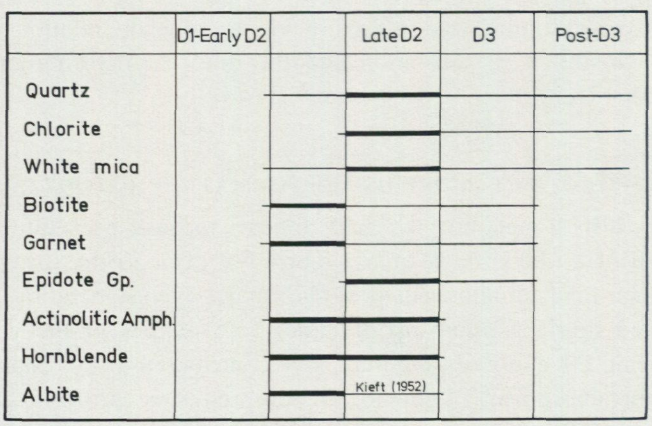


Fig. 36. Summary of grain growth/recrystallization and deformation relationships in the Björkvattnet Unit.

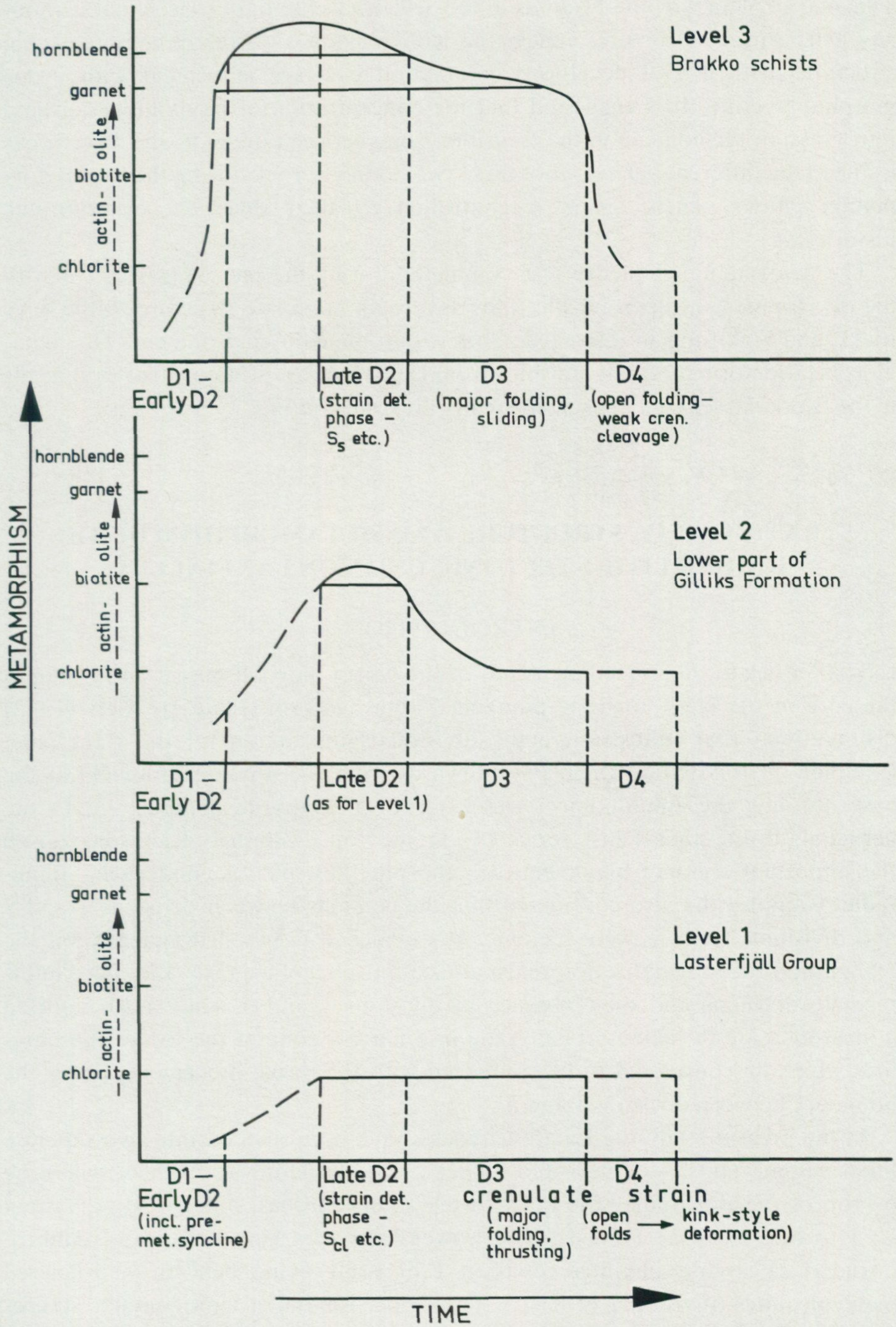


Fig. 37. Variation of metamorphic grade with respect to time and tectonic level within the Björkvattnet Unit.

This may explain why the F3 folds in the Seve rocks are tight to isoclinal without any axial surface structure, and in the Köli rocks are distinctly crenulate-style structures with a well developed crenulation cleavage associated with metamorphic layering. It is suggested that the concentration of phyllonitization and retrogression phenomena in the transition zone between the Köli and Seve rocks is due to the different behaviour of these two bodies of rock during the later deformation phases which, again, is controlled by their divergent metamorphic histories.

The deformation sequence was completed during the later stages of D4 with the development of open buckle folds and weak crenulate structure in the Seve rocks, and kink folds in restricted zones within the Köli environment. The metamorphism-deformation relationships from D1 to D4 for different structural levels in the Björkvattnet Unit are schematically illustrated in Fig. 37.

STRATIGRAPHY, STRUCTURE AND METAMORPHISM OF THE HIGHER LITHO-TECTONIC UNITS ON LAXFJÄLLET

INTRODUCTION

Laxfjället lies to the immediate south of the Norra Storfjäll massif and occupies the core of the late, northerly plunging Tärna Synform (Plate I). Kieft (1952) distinguished four lithostratigraphic divisions lying on top of the Bångfjället Complex on Laxfjället, which pass upwards from the phyllite series (1) at the base, through the amphibolite series (2) and gneiss-norite complex (3) to the garnet-staurolite-mica-schist series (4) at the top. Although Kieft recognised the important tectonic break between the phyllites and the underlying Bångfjället Complex, he also considered that the contacts between divisions 1 and 2 and divisions 3 and 4 were tectonic. His evidence is based principally on the recognition of migmatization/feldspathization phenomena in the two intermediate divisions and their absence in divisions 1 and 4. Kieft also reported a local increase of metamorphic grade in a narrow zone at the top of the phyllites which he considered to be connected with the thrust movements along the boundary between divisions 1 and 2.

In the present study the Laxfjället rocks have been divided into two separate litho-tectonic units — Lower and Upper Laxfjället Units — both of which lie on top of the Bångfjället Complex to the south and east and the Björkvattnet Unit to the west (see Plate I). The Lower Laxfjället Unit consists of gabbro-intruded calcareous phyllites (division 1 of Kieft) lying beneath well-layered metavolcanites (division 2 of Kieft). The Upper Laxfjället Unit contains staurolite schists with an important gabbro mass (the Laxfjället gabbro), feldspar-rich rocks of uncertain origin and a thin dolomitic marble horizon near the base

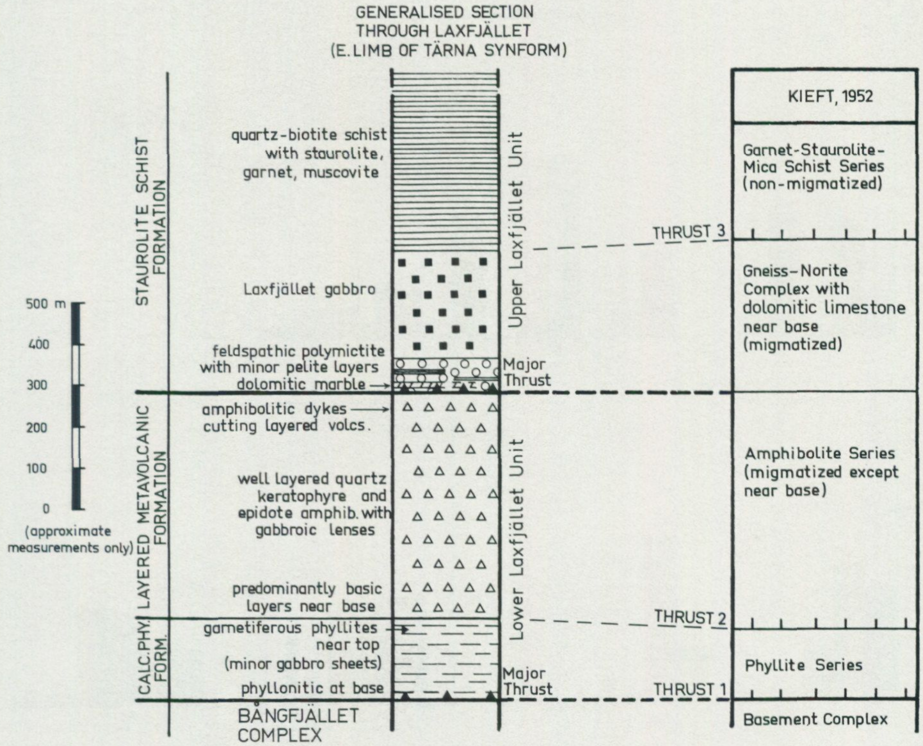


Fig. 38. Litho-tectonic units and stratigraphy on Laxfjället.

(essentially divisions 3 and 4 of Kieft). Metamorphic grade increases upwards throughout the Lower Laxfjället Unit (distribution of metamorphic zones indicated on Fig. 31), whilst the rocks of the Upper Laxfjället Unit were metamorphosed under amphibolite facies conditions. The boundary between the schists and the layered metavolcanites is considered to be a zone of intense deformation and to mark an important tectonic break in the Laxfjället sequence.

The tectonic contacts beneath the gabbro-intruded phyllites and the staurolite schists preclude the establishment of any formal stratigraphic nomenclature for these rocks. Thus, the informal names Calcareous Phyllite Formation and Layered Metavolcanic Formation (Lower Laxfjället Unit) and Staurolite Schist Formation (Upper Laxfjället Unit) are employed for the mappable rock associations on Laxfjället. A generalized type-section for these litho-tectonic units and formations is defined beneath the main ski lift from Tärnaby village (west) to the top of Laxfjället (8e to 8f). This is summarized in Fig. 38 which also includes a comparison with Kieft's interpretation of the stratigraphy. It is emphasized that the true stratigraphic order is unknown, and, thus, Fig. 38 only shows the structural sequence within the Lower and Upper Laxfjället Units.

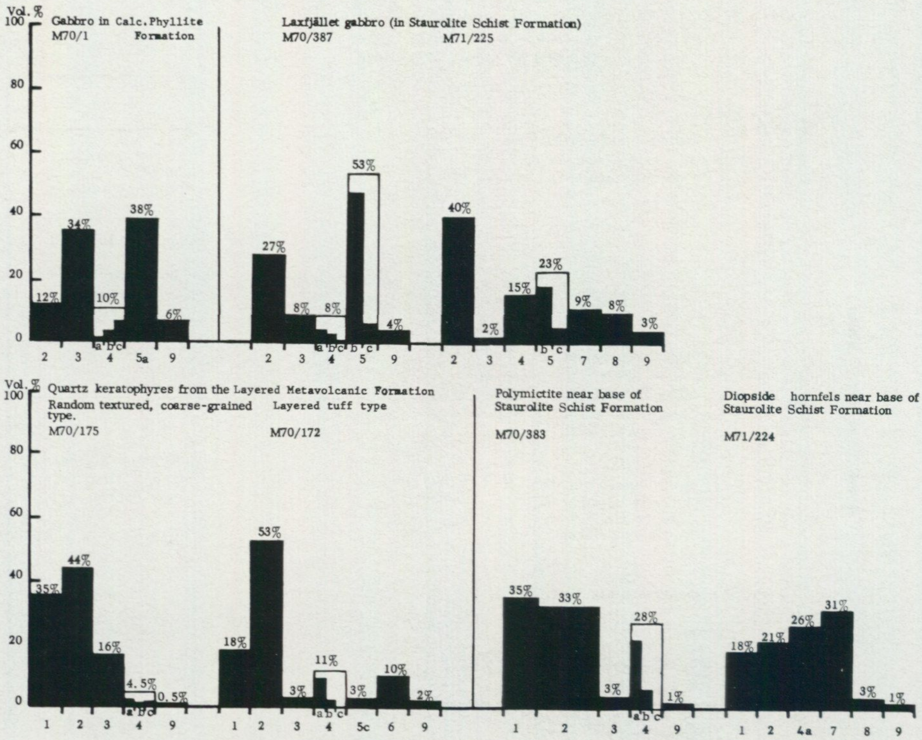


Fig. 39. Representative modal analyses of the main lithologies within the Lower and Upper Laxfjället Units. 1 = Quartz; 2 = Feldspar; 3 = Epidote Group; 4 = Phyllosilicate, a = biotite, b = muscovite, c = chlorite; 5 = Amphibole, a = pale green hornblende (ig.), b = brown-green hornblende (ig.), c = blue-green hornblende; 6 = Garnet; 7 = Clinopyroxene; 8 = Opaques; 9 = Accessories inc. sphene, sericite, apatite and quartz in the gabbros.

LOWER LAXFJÄLLET UNIT — STRATIGRAPHY, STRUCTURE AND METAMORPHISM

STRATIGRAPHY

Calcareous Phyllite Formation

The less deformed and recrystallized parts of this formation are composed of closely alternating laminae and beds of subordinate, fine-grained pelite and coarser grained calcareous phyllite, with relic feldspar porphyroclasts preserved in a quartz-calcite-mica matrix. The plane parallel bedding structure is on a fine scale, the pelite component occurring as laminae and the coarser layers commonly 1—2 cm thick.

Metamorphic grade increases upwards and only the basal beds contain quartz-white mica-chlorite assemblages. Biotite is common at higher levels and abundant garnet, often localized in the more pelitic laminae, epidote-clinozoisite and more rarely blue-green hornblende are present near the top of the formation (see

TABLE 5. Summary of the petrography of the deformed and variably metamorphosed gabbros in the Lower and Upper Laxfjället Units.

	STAGE 1 Amphibole (Primary)	STAGE II Amphibole (Metamorphic)	Pyroxene	Type of Feldspar
Laxfjället Gabbro	Brown-green to brown hornblende; commonly present in ophitic and subophitic relationship with original plagioclase feldspar.	Various modifications of Stage I amphibole inc:- 1) recrystallization of old grains - mass of fine-grained, brown-green to brown hornblende, 2) augening of the initial random texture to produce a clear linear fabric - particularly obvious near the edge of the intrusion, 3) new grain growth both as rims on Stage I amphibole and as mass of needle-like prisms of pale green amphibole (actinolitic?)	Near colourless, zoned clinopyroxene occurs as relics within brown-green hornblende (Stage I) and with Stage II amphibole rims etc.	Ranges from An_{46} to An_{56} i.e. Andesine/ Labradorite range. Alteration to mass of zoisite-clinozoisite, untwinned plag. feldspar and sericite.
Minor gabbros in Layered Metavolc. Fm.	Strongly lenticularized, coarser blades and X-sections of a pale green-brown to deeper brown/ green hornblende.	Blue-green hornblende present either as rims on Stage I hornblende or as lineated mass of new amphibole.	———	?
Gabbro lenses in Calcareous Phyllite Formation	Isolated and broken blades of very pale-green to distinctly blue-green hornblende. Relic ophitic texture rarely preserved.	Slender needle-like prisms of secondary pale blue-green to blue-green actinolitic amphibole occurring particularly near the edges of the Stage I hornblende. Strongly schistose types at edges of the intrusions dominated by lineated mass of very pale green tremolitic/ actinolitic amphibole.	———	An_{35-36} i.e. Andesinic. Predominantly replaced by abundant albite and epidote group minerals.

also Kieft 1952, p. 23—24). Towards the base retrogressive conversion to calcareous phyllonite with tectonically controlled lenticular structure is superimposed upon the low-grade metamorphic fabric.

Lens-shaped masses of metagabbro occur as concordant sheets within the Calcareous Phyllite Formation. The most important body is 1.5 km long and 0.5 km wide at its maximum extent and is situated to the west of Tärnaby. Petrographic details and a representative modal analysis are summarized in Table 5 and Fig. 39 respectively. The amphibole is divided into two types; blades of a near colourless to pale blue-green hornblende are commonly surrounded at their edges by a fine-grained aggregate of needle-like actinolitic amphibole (and pale green chlorite). Correlation of these types with a primary igneous amphibole and a later metamorphic amphibole respectively is tentatively suggested in Table 5. The rocks near the contact with the surrounding phyllites are fine-grained greenschists with a strong foliation (oriented actinolite, chlorite and epidote) which corresponds to the early, penetrative cleavage in the phyllites; relics of a coarser grained, pale green amphibole are rarely preserved here. The deformation and metamorphism of these gabbroic bodies is consistent with an intrusion age prior to formation of the early foliation.

Layered Metavolcanic Formation

This formation consists of a close alternation of three lithologies:

1. Fine- and even-grained albite-quartz rocks (feldspathitic composition — see Wallis et al. 1968 and Fig. 39) with rare, subhedral megacrysts (0.5 mm across) of plagioclase feldspar interpreted as relic phenocrysts.

2. Coarser layers with a higher proportion of lenticularized and saussuritized plagioclase megacrysts (An_{6-8} and ranging up to 2—3 mm in size) in a subordinate albite-quartz groundmass; lithotypes 1 and 2 contain minor amounts of oriented biotite and muscovite, which define a weak schistosity, as well as scattered garnet, epidote and blue-green hornblende.
3. Epidote amphibolite layers, with blue-green hornblende as the metamorphic amphibole, which occasionally contain stretched, leucocratic aggregates of epidote-clinozoisite and untwinned plagioclase interpreted as recrystallized, originally more calcic feldspar phenocrysts.

The high albite content and the phenocrystic texture suggest that the feldspathitic rocks are igneous in origin (quartz keratophyre). The regular interstratification of these rocks with more basic epidote amphibolite and thin biotite-rich layers suggests a predominantly water-lain tuff succession.

More homogeneous pods of coarser grained acidic and metagabbroic material are also present and are interpreted as subvolcanic, minor intrusions. The former are strongly lineated (streaky mineral lineation) and either aphyric or porphyritic. The basic lenses display similar structural and compositional features to the larger bodies in the Calcareous Phyllite Formation. However, the earlier amphibole here is a green-brown hornblende which is not only enclosed as relics in an oriented blue-green hornblende/epidote-zoisite aggregate, but also contains a sharply defined rim of the same blue-green hornblende.

On the subordinate peak of Laxfjället (8f), near the top of the formation, amphibolite dykes, generally < 0.5 m wide, cut the layering (primary?) within the host quartz keratophyre/epidote amphibolite sequence (Fig. 40a). These dykes contain coarse, euhedral, feldspar phenocrysts (up to 1 cm long) towards the centre and possess aphyric, chilled margins. The contact with the surrounding layered sequence is sharp, and irregular offshoots of basic rock from the main dyke(s) enclose the feldspathitic material of the earlier, layered sequence (Fig. 40a). The dykes are deformed by minor structures related to the late Tärna Synform and often possess well foliated margins, the foliation being continuous with the early, penetrative schistosity in the surrounding rocks (Fig. 40b); pre-schistosity intrusion is inferred.

The Layered Metavolcanic Formation thins out markedly on the western limb of the Tärna Synform and northwest of Stor-Laisan its thickness is only a few tens of metres (Strömgård, pers. comm.). The lower contact with the Calcareous Phyllite Formation, well exposed on both shores of Stor-Laisan (east), is transitional. The albite + blue-green hornblende + epidote assemblage of the basic rocks within the Layered Metavolcanic Formation is consistent with the metamorphic grade indicated by the garnetiferous phyllites at the top of the underlying Calcareous Phyllite Formation. Phyllonitization and retrogression of the metamorphic fabric within the metavolcanites are, however, concentrated

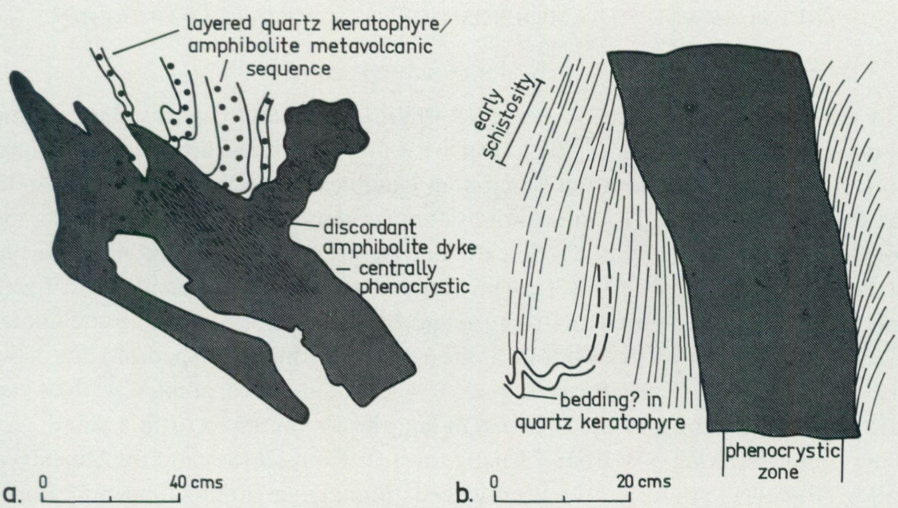


Fig. 40. Amphibolite dyke — metavolcanite relationships. In b. the early schistosity deflects towards the dyke-metavolcanite contact and penetrates the dyke along its margins.

near the *upper* contact with the Staurolite schist Formation (Upper Laxfjället Unit). Thus, the attenuation of the Layered Metavolcanic Formation is thought to be related to the thrusting of the Upper Laxfjället Unit on top of the metavolcanites.

Kieft (1952) considered that the bulk of his ”amphibolite series” (Layered Metavolcanic Formation) was migmatized and ”that the main process leading to the formation of this migmatite complex was a metasomatic replacement of the amphibolites by the acid component”. Kieft stated that this process involved the introduction of considerable amounts of Na and Si and that it occurred later than the intrusion of the ”diabase” dykes. However, it can be stated that:

1. The mineral association in, particularly, the amphibolites does not indicate the high temperatures required for the processes envisaged by Kieft.
2. The preservation of relic phenocrystic textures in the feldspathitic rocks (”the acid component”) and their regular interstratification with epidote amphibolites do not substantiate the hypothesis that the former are migmatitic.
3. The so-called ”diabase” dykes are discordant to the layering between the feldspathitic rocks and the epidote amphibolites.

Except for the minor intrusions, a primary volcanic (pyroclastic?) origin is favoured here for the well layered rock units within this formation.

STRUCTURE AND METAMORPHISM-DEFORMATION RELATIONSHIPS

Major structures

The phyllonitic base of the Calcareous Phyllite Formation marks the tectonic boundary of the Lower Laxfjället Unit with the underlying litho-tectonic units, both where it lies on top of the Bångfjället Complex and where it oversteps onto the Björkvattnet Unit. The increased mylonitization of the phyllites towards the contact with the Bångfjället Complex is particularly well illustrated near the top of the east Tärnaby ski run (8f). The contact is marked by abundant stringers and veins of quartz, lenticular structure and by dark glassy veinlets in the immediately underlying rocks (pseudotachylite veins in strongly crushed phyllite have also been reported by Kieft 1952, p. 25). The Bångfjället Complex wedges out just west of Tärnaby and the two enveloping thrusts merge to form a single tectonic contact striking NW from Tärnaby towards Konäset (8e) and the Umeälven valley (Plate I). There is also pronounced thinning, mylonitization and retrogression of the Layered Metavolcanic Formation just beneath the thrust contact with the Upper Laxfjället Unit.

The thrusts are folded around the NNW-plunging synform which dominates the major structure of Laxfjället (Plate III, 5a), and which is a northward continuation of the late Tärna Synform defined within the Björkvattnet Unit.

Minor structures

The early, penetrative foliation ($S1_{LLU}$) is defined by oriented, fine-grained white mica and chlorite in much of the Calcareous Phyllite Formation, but is more schistose in appearance at higher structural levels in the Lower Laxfjället Unit; this is consistent with the increase in metamorphic grade upwards throughout the unit. Various types of lineation are contained within $S1_{LLU}$ including a mineral lineation in the quartz keratophyres, a penetrative amphibole lineation in the metabasites and elongation of recrystallized phenocrysts in the epidote amphibolites. These lineations are considered to represent the direction of maximum elongation in $S1_{LLU}$. However, this foliation also forms an axial surface structure to minor, tight to isoclinal folds ($F1_{LLU}$) (Fig. 41a, b) which are associated with a quartz rodding and define a second type of lineation broadly related to $S1_{LLU}$. Both lineation types plunge NW at angles less than 45° (Plate III, 5b).

The locally high strain zone at the base of the Calcareous Phyllite Formation is characterized by the development of intense lenticular structure, associated with isoclinal and occasional intra-folial folds in which $S1_{LLU}$ is intensely transposed into the axial surface orientation ($S2_{LLU}$). These folds, often with N-dipping minor thrusts subparallel to their limbs, are, thus, referred to here as $F2_{LLU}$.

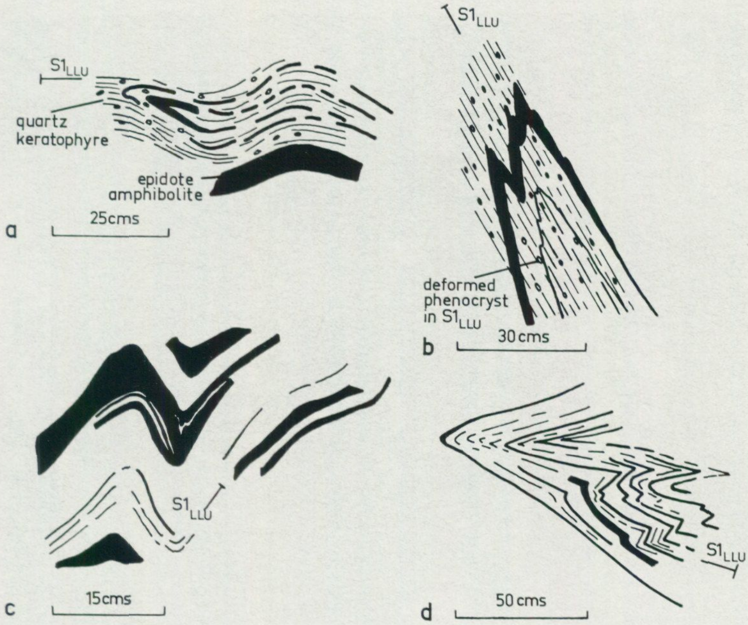


Fig. 41. Minor folds in the Lower Laxfjället Unit. a: $F1_{LLU}$ isocline modified by more open $F3_{LLU}$ folds related to the Tärna Synform. Section parallel to the $F1_{LLU}$ fold axis reveals hornblende prisms parallel to the fold axis. b: $F1_{LLU}$ isocline. c: Open and upright $F3_{LLU}$ folds which deform $S1_{LLU}$ and are related to the Tärna Synform. d: Late chevron folds.

The most conspicuous minor folds (and associated crenulation lineation) in the Lower Laxfjället Unit fold $S1_{LLU}$ and are congruent to the late Tärna Synform (Fig. 41c). Particularly near the hinge zone, these minor folds are open to close and upright in shape with steeply dipping axial surfaces and generally NNW-plunging axes (Plate III, 5c). Since the Tärna Synform folds the major thrusts and is, therefore, probably later than the $F2_{LLU}$ structures, these folds are referred to as $F3_{LLU}$.

The final deformation event is the local development, on the eastern limb of the Tärna Synform, of chevron folds with gentle, W-dipping axial surfaces and W- to WSW-plunging axes (Fig. 41d and Plate III, 5c). These folds deform the earlier rodding lineation and occasional, multiple hinges are also present. They are considered to be a minor, late-stage development in the structural evolution of the Lower Laxfjället Unit.

Timing of metamorphism

The inverted distribution of metamorphic zones within the Lower Laxfjället Unit is indicated in Fig. 31. Oriented white mica and chlorite define $S1_{LLU}$ in the lower part of the Calcareous Phyllite Formation, whilst biotite coexists

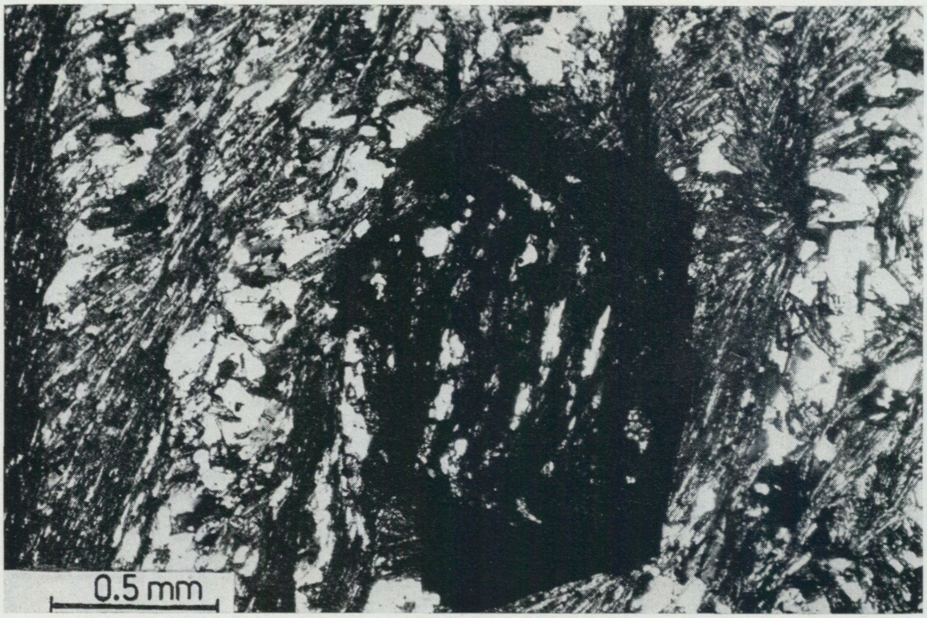


Fig. 42. Syn-S1$_{LLU}$ garnet with clear, post-tectonic, idioblastic rim. The S1$_{LLU}$ micas are reoriented parallel to the later S2$_{LLU}$ crenulation cleavage which only partially deflects around the garnet blast (south shore of Stor-Laisan, 8f, crossed nicols).

with these phyllosilicates in S1$_{LLU}$ towards the top of the formation and is present in the overlying, quartz keratophyres. Strongly oriented, slender needles of blue-green hornblende define the S1$_{LLU}$ mineral lineation in the epidote amphibolites.

Porphyroblasts of garnet and hornblende occur near the top of the Calcareous Phyllite Formation and in the overlying metavolcanites. The main foliation in the garnetiferous phyllites at the eastern end of Stor-Laisan (8f) is a differentiated crenulation cleavage (S2$_{LLU}$) which deforms the phyllosilicates of S1$_{LLU}$. The garnets here display poikiloblastic cores with S-shaped tracks and non-poikiloblastic, idioblastic rims which not only transect the S1$_{LLU}$ micas but also cut across the S2$_{LLU}$ differentiated crenulation cleavage (Fig. 42); the quartz inclusions are finer and more elongate than the matrix quartz. It is inferred that their initial growth stage was syntectonic with respect to S1$_{LLU}$ and later growth continued after formation of the S2$_{LLU}$ crenulation cleavage. Elsewhere in the phyllites and quartz keratophyres there are several occurrences of pre-S1$_{LLU}$ garnets, with deflection of S1$_{LLU}$ around the garnet and development of quartz eyes, and small (0.1—0.3 mm), idioblastic and non-poikiloblastic, post-S1$_{LLU}$ garnets. Hornblende porphyroblasts are either deformed and rotated into the S1$_{LLU}$ foliation or show faintly S-shaped inclusion trails continuous with the external foliation ($S1_{LLU}$). Thus, growth of porphyroblastic hornblende took place prior to and during $S1_{LLU}$ development.

It is concluded that growth of garnet and hornblende initiated prior to $S1_{LLU}$ and, together with phyllosilicate grain growth, continued during the development of $S1_{LLU}$. Garnet continued to be stable, probably until after formation of the $S2_{LLU}$ crenulation cleavage.

UPPER LAXFJÄLLET UNIT — STRATIGRAPHY AND METAMORPHISM-DEFORMATION RELATIONSHIPS

STRATIGRAPHY – STAUROLITE SCHIST FORMATION

The bulk of this formation consists of aluminous quartz-biotite schist (± graphite) with coarser poikiloblasts of staurolite, garnet and muscovite. The schistosity is defined by oriented flakes of biotite, muscovite and minor amounts of clinozoisite. In the lower part of the formation, however, the rocks are more variable and are strongly deformed; this level is dominated by a coarse-grained, schistose, feldspathic polymictite (no principal compositional group of minerals exceeds 35 % in a modal analysis; see Wallis et al. 1968) which is associated with a discontinuous dolomite- and diopside-rich marble horizon. A large gabbro body (the Laxfjället gabbro) occurs near the base of the formation. The basal lithologies are now described and their petrological and structural significance discussed in more detail.

Feldspathic polymictite

The coarser component is made up of various types of feldspar (Figs. 39 and 43a—c) including in relative order of importance:

1. Plagioclase which is commonly zoned and which displays simple lamellar (albite and pericline) and complex combined carlsbad and lamellar twinning (Gorai 1951); the composition varies from oligoclase to sodic andesine, and many of the twins are mechanical being related to the strong deformation that these rocks have suffered.
2. Zoned, untwinned and simply twinned (carlsbad) feldspar.
3. Mixed Na-K feldspar including zoned string perthite, microcline and microcline perthite.

Composite grains are also an important constituent of the total "feldspar" content (Fig. 43b). The coarser feldspars range in size from 1—2 mm to 1—2 cm across, are commonly smooth-edged with a distinctive augen structure and often show slight saussuritization. The feldspars contain no inclusions of the matrix but the biotite schistosity totally encloses them (Fig. 43a). It appears, then, that

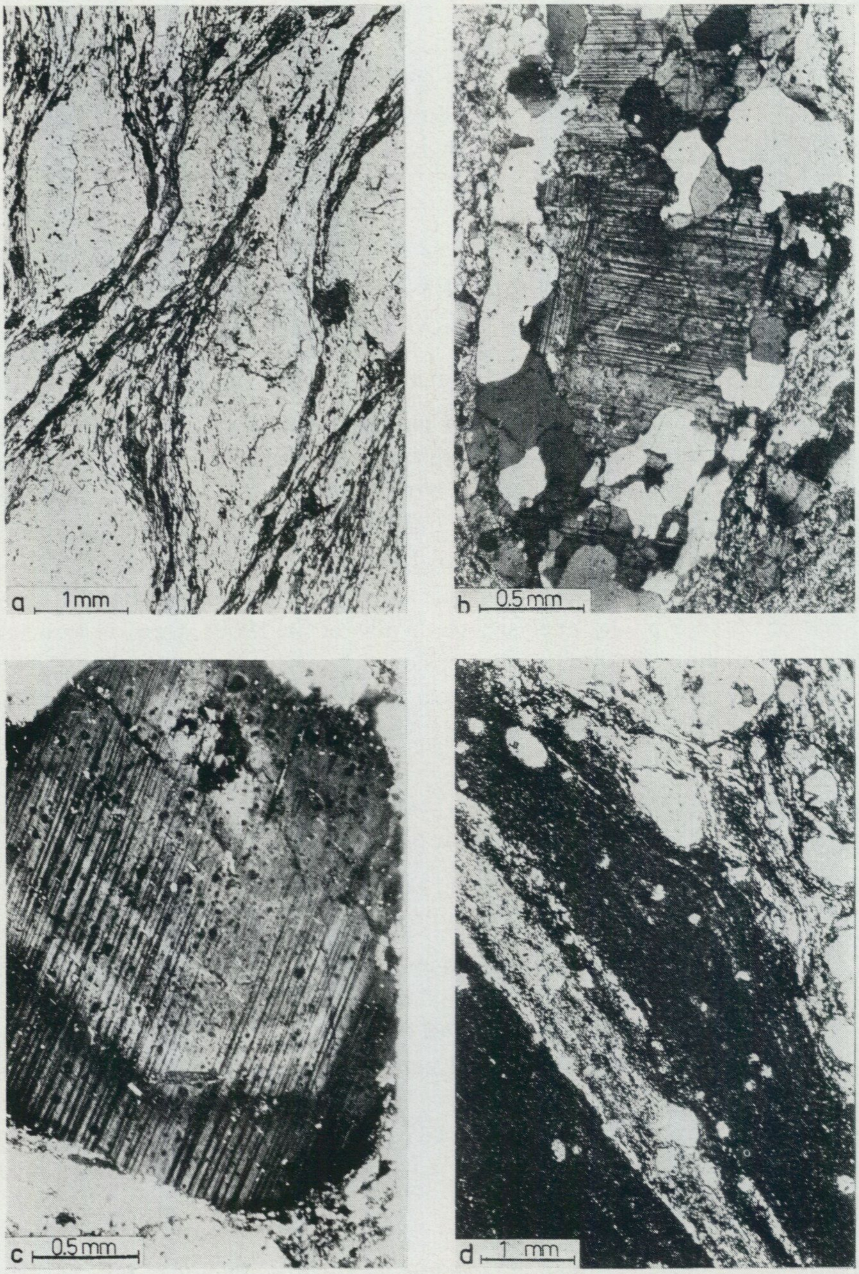


Fig. 43. Characteristics of the feldspathic polymictite (base of Staurolite Schist Formation). a: Biotite schistosity deflects around deformed pre-schistosity, feldspar grains (Laxfjället peak, 8f, uncrossed nicols). b: Composite quartz + feldspar grain (Laxfjället peak, 8f, crossed nicols). c: Zoned and mechanically twinned plagioclase grain (SW Laxfjället, 8e, crossed nicols). d: Probable mylonitic layering in strongly deformed polymictite (eastern Laxfjället, 8f, uncrossed nicols).

they grew prior to formation of the schistosity and were deformed at a later stage. The matrix consists of recrystallized quartz, biotite and subordinate muscovite, epidote and clinozoisite; there are also small garnet (1 mm maximum) and zoned hornblende porphyroblasts.

The polymictite member displays different types of sharply defined compositional layering including (Fig. 43d):

1. Alternations of coarser feldspar-rich layers and fine-grained pelitic laminae.
2. Alternations of coarser and finer grained feldspar-rich layers.
3. Minor amphibolite layers within the dominant feldspathic polymictite.

The origin of this layering is uncertain. However, the polymictite has a mylonitic appearance and is thought to occupy a zone of intense deformation at the base of the Upper Laxfjället Unit.

The pelite layers resemble the overlying aluminous schists and, by a decrease in the number and size of the feldspar crysts, the feldspathic polymictites pass transitionally upwards into these schists.

Dolomite- and diopside-rich marble

Discontinuous dolomite marble occurs near the base of the formation within the feldspathic polymictite zone. On the southern shore of Stor-Laisan (E limb of Tärna Synform) it is a pure dolomite marble with only minor chlorite-rich laminae. This changes facies slightly to a more impure type on eastern Laxfjället consisting of dolomite, calcite and tremolite in a granoblastic aggregate with minor amounts of diopside. On the southern slopes of Laxfjället, however, well layered pelite and diopside marble occur at the same stratigraphic level. The leucocratic layers consist of a granoblastic to weakly oriented aggregate of diopside, brown-green hornblende, clinozoisite, An-rich plagioclase (An_{60}) and variable amounts of quartz and poorly oriented biotite (Fig. 39); the texture is distinctly hornfelsic.

It appears that these layered diopside-rich rocks represent the impure extreme of a progressive, southerly change in facies of the carbonate horizon towards a more siliceous type. The increased content of diopside and brown-green hornblende is consistent with the extensive development of the Laxfjället gabbro and its associated thermal aureole on the southern side of Laxfjället (Winkler 1967, chapter 4).

The Laxfjället gabbro

This basic intrusion forms the main summit of Laxfjället and is predominantly a single concordant body with a maximum width of 400 m. In the better preserved parts it is composed of an interlocking framework of reddish-brown biotite, brown-green to brown hornblende, which rarely encloses plagioclase in an

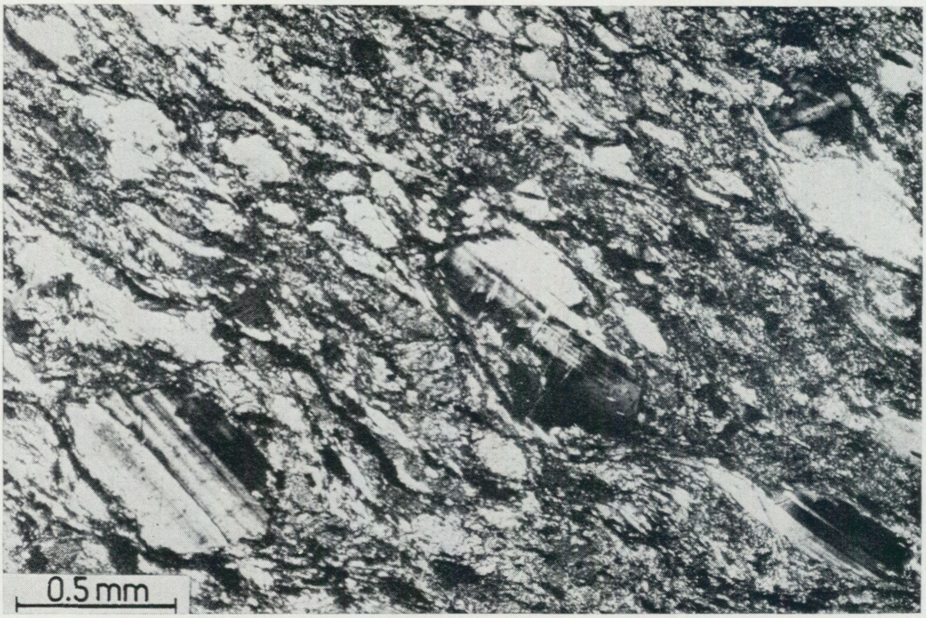


Fig. 44. Lineated flaser gabbro with deformed plagioclase crysts at border of Laxfjället gabbro (eastern Laxfjället, 8f, crossed nicols).

ophitic texture, and lath-like plagioclase which varies in composition from andesine to sodic labradorite. The relations between the different amphibole types and clinopyroxene, which occurs in minor amounts, are summarized in Table 5. The secondary amphibole, together with chlorite and (clino)-zoisite, increase in importance in the more strongly lineated and recrystallized types near the margins of the body. Here also the original feldspar shape is obscured as distinctive lenticular microstructure predominates (Fig. 44).

The deformed character of the concordant gabbroic bodies in the Lower Laxfjället Unit and the nature of the later, metamorphic amphibole in these rocks are consistent with their intrusion prior to formation of the early foliation. Although the Laxfjället gabbro is conspicuously more homogeneous and primary textures are better preserved, this is probably controlled by the actual size of the body since flaser gabbro again predominates near its borders. Furthermore, in the more deformed diopside hornfelses associated with this intrusion, the diopside porphyroblasts are totally enveloped by a fine-grained, biotite-rich foliation (Fig. 45) which is equivalent to the main schistosity. The deformed granoblastic microstructure suggests, therefore, that the Laxfjället gabbro was emplaced prior to formation of the schistosity.

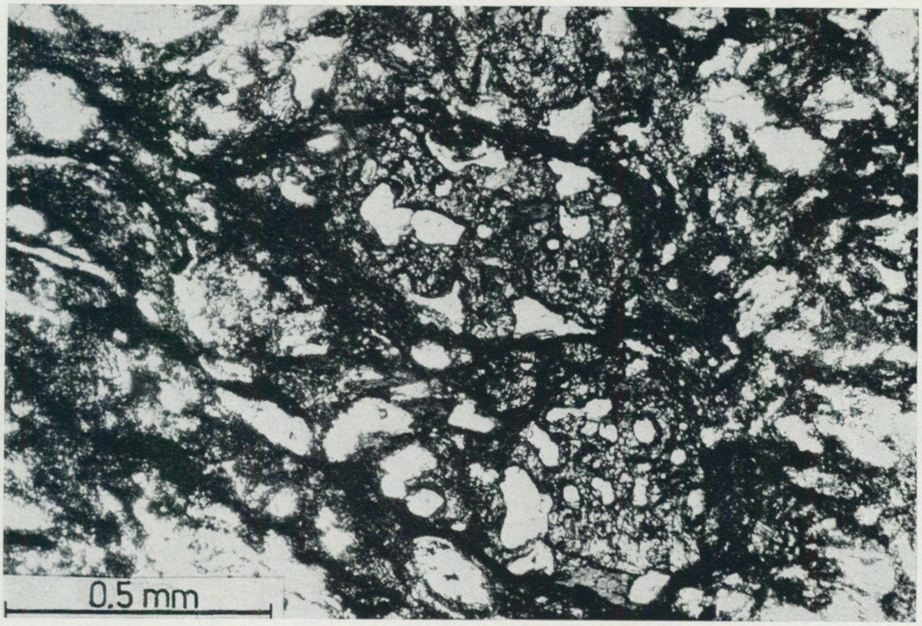


Fig. 45. Deformed diopside marble; diopside poikiloblasts are elongate parallel to the biotite lineation which corresponds to the main schistosity in the adjacent schists (Laxfjället peak, 8f, uncrossed nicols).

Genetic implications

Kieft (1952) compared the genesis of the so-called "gneiss-norite complex" (approximately equivalent to the lower, gabbro-intruded part of the Staurolite Schist Formation) with the "fragment-bearing gneisses" described from the northern part of Norra Storfjället by Aleva (1950) and Mulder (1951). He concluded that, after the basic intrusion, the rocks were highly metamorphosed and a considerable introduction of alkalies took place. The introduction of Na was inferred from the occurrence of rather sodic plagioclase "porphyroblasts" in the "gneisses" (equivalent to the schistose feldspathic polymictite here) and the migmatization of the underlying amphibolites. Kieft considered that the only relics of the original schists were "granulite-like rocks" (equivalent to the hornfelsic diopside marble here) which are now enclosed in the metasomatically altered "gneisses".

The quite different compositions of the feldspars at the base of the schists compared with the underlying amphibolites and feldspathites indicate that the migmatization hypothesis cannot be applied to both units as Kieft envisaged. Reinterpretation of the petrological significance of the amphibolitic and feldspathic rocks beneath the Staurolite Schist Formation has already been discussed; in particular, a volcanic origin was argued for these well layered units. The coarse feldspars in the polymictite horizon either relate directly to the intrusion of the Laxfjället gabbro as a contact effect or do, indeed, represent a limited migmatization of the schists. The local occurrence of irregular muscovite pegmatite veins up to 1.5 m across, which intrude both the gabbro complex and the schists, and the extensive lateral but strongly confined vertical extent of the deformed, feldspar-rich rocks support the latter interpretation. These rocks have been mapped north of Stor-Laisan on both limbs of the Tärna Synform (Strömgård, pers. comm.), where apparently no basic intrusions are associated, and they are restricted to the lower part of the same formation. Kieft's "granulite-like rocks" represent a thermally metamorphosed, siliceous carbonate horizon, stratigraphically equivalent to the dolomite marble.

METAMORPHISM-DEFORMATION RELATIONSHIPS

The Staurolite Schist Formation forms the core of the Tärna Synform and the zone of intense deformation at the base of the unit is folded in the synform. The formation is dominated by a penetrative biotite-muscovite schistosity ($S1_{ULU}$) which contains an approximately NW-plunging mineral lineation (Plate III, 5b). $S1_{ULU}$ is deformed by open to gentle folds with upright axial surfaces which plunge mostly to the north (Plate III, 5c), being congruent to the Tärna Synform, but along the south shore of Stor-Laisan (8e) plunge westerly on steep, N-dipping axial surfaces. Eyed interference structures between early, tight to isoclinal folds, probably related to $S1_{ULU}$ ($F1_{ULU}$), and the later, upright, northerly plunging folds occur near the subordinate of the two Laxfjället peaks (8f). Since there is no clear refolding relationship between the two sets of post-$S1_{ULU}$ folds, they are both referred to as $F2_{ULU}$ structures. There is a notable absence in the Upper Laxfjället Unit of tight to isoclinal, post-schistosity folds with gentle, N-dipping axial surfaces and transposition characteristics, equivalent to F2 of the Lower Laxfjället Unit.

The $S1_{ULU}$ schistosity deflects around the "sponge-like" staurolite and several of the garnet poikiloblasts. Furthermore, the quartz inclusions are distinctly finer grained than the matrix quartz. Thus, it appears that the metamorphism within the Upper Laxfjället Unit had reached amphibolite facies conditions (Winkler 1967, p. 106) prior to formation of $S1_{ULU}$. After $S1_{ULU}$ the metamorphism was sufficiently high to allow the growth of coarse (1—2 mm across) muscovite blades in the layered diopside hornfels and idioblastic garnet in the aluminous schists. The post-$S1_{ULU}$ character of these blasts is shown by their cross-cutting relationship to the schistosity. Near the base of the formation staurolite porphyroblasts are retrogressed to a fine-grained muscovite aggregate.

TECTONIC IMPLICATIONS

The less deformed part of the Calcareous Phyllite Formation, with its characteristic intrusions of metagabbro, and the overlying Layered Metavolcanic Formation, containing the relatively thick development of quartz keratophyre (acid volcanics), resemble the rocks in the proposed upper part of the Lasterfjäll Group, described from the Remdalen area in the Western Synform (Zachrisson 1969). According to Zachrisson, these rocks lie stratigraphically above the major part of a calcareous phyllite sequence (Lasterfjäll Calcareous Phyllite) which is correlated with the Lövfjäll phyllites to the east. West of Tärnaby, however, the gabbro-intruded calcareous phyllites lie in a separate litho-tectonic unit above the typical Lövfjäll calcareous phyllites and greywackes (Lower Laxfjället and Björkvattnet Units respectively). The regional significance of this tectonic break awaits further detailed investigations.

The schists and dolomite marble of the Upper Laxfjället Unit, which rest with tectonic contact on top of the phyllites and metavolcanites, continue northwards onto Norra Storfjället and resemble the high-grade aluminous schists on the eastern side of Södra Storfjället. The Upper Laxfjället Unit is, thus, tentatively correlated with the northern part of the Storfjäll Nappe (Kulling 1964 and Fig. 1) which according to Kulling *(in* Strand and Kulling 1972, p. 161) ”lies between the typical Seve-Köli rocks in the southeast to east and the Rödingsfjäll nappe . . . in the northwest to west”. Kulling, however, placed the lower boundary of the Storfjäll Nappe in the Tärnaby area at the base of the metavolcanites on Laxfjället (see Fig. 188 *in* Strand and Kulling 1972). As stated earlier, there is no evidence that this level is a major tectonic break. The lower boundary of the Storfjäll Nappe is situated in this study in the zone of intense deformation between the metavolcanites of the Lower Laxfjället Unit and the staurolite schists of the Upper Laxfjället Unit.

The inversion of metamorphic zones within the Laxfjället rocks may be explained in two ways:

1. The rocks of the Upper Laxfjället Unit (Storfjäll Nappe?) were thrust under relatively high-grade metamorphic conditions (syn- or late-metamorphic but pre-$F2_{ULU}$) on top of the low-grade phyllites and metavolcanites; this effected a local increase in metamorphic grade in those units lying immediately beneath the Upper Laxfjället Unit.
2. The metamorphism originally increased with depth in the normal manner but late- or post-metamorphic (but pre-$F2_{ULU}$) thrusting, involving relative movement between the Upper and Lower Laxfjället Units, effected overturned folding and inversion of metamorphic zones in the lower unit.

In the first case the metamorphism of the Lower Laxfjället Unit converged towards that of the Upper Laxfjället Unit due to thrusting of the latter over the former during metamorphism. In the second case, a ”normal” metamorphic zone

TABLE 6. Tentative correlation of structural events and porphyroblast growth between the Björkvattnet, Lower Laxfjället and Upper Laxfjället Units.

BJÖRKVATTNET UNIT			LOWER LAXFJÄLLET UNIT		
D4	$F4_{kink}$ - Late kink folds (flat-lying axial surfaces) $F4_a/F4_b$ - Major, open to gentle antiforms and synforms; locally intense crenulation cleavage and minor folding (steep axial surfaces)	GARNET, MUSCOVITE	$D3_{LLU}$	Late chevron folds (flat-lying axial surfaces) Open to close minor folding (steep axial surfaces)	GARNET
D3	Major and minor folding; Locally intense crenulation cleavage; Significant mylonitization and thrusting	ALBITE	$D2_{LLU}$	Localization of isoclinal and occasional intra-folial folds at base of unit; transposition foliation, crenulation cleavage	AMPHIBOLE
D2	Early, regional foliation; penetrative lineations incl. deformed pebble lineation Tight to isoclinal minor folds	AMPHIBOLE, BIOTITE	$D1_{LLU}$	Early, penetrative foliation and lineation fabric Tight to isoclinal minor folds	
D1	Early, synclinal repetition				

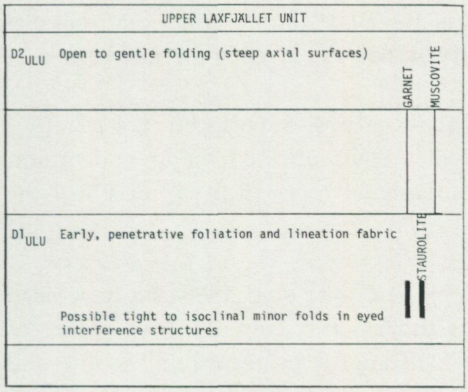
distribution was disturbed due to late- or post-metamorphic deformation. Cogent arguments in favour of the second explanation include the following:

1. Retrogressive effects in the basal schists of the Upper Laxfjället Unit and phyllonitization of $S1_{LLU}$ at the top of the Lower Laxfjället Unit indicate late- or post-metamorphic movement between the two units. However, garnet, hornblende and biotite were stable prior to (and during) foliation development in the Lower Laxfjället Unit i.e. before thrusting occurred.
2. Post-metamorphic, overturned folding is also prominent in the underlying Björkvattnet Unit and, on Göutavardo, accounts for the inversion of metamorphic zones (Plate I and Fig. 31). However, such folding appears to be absent in the Upper Laxfjället Unit. These observations support the hypothesis that movement of the Upper Laxfjället Unit (Storfjäll Nappe?) effected overturned folding and local inversion of the metamorphic zones in the underlying rocks.

CORRELATION OF STRUCTURAL-METAMORPHIC EVENTS BETWEEN THE LAXFJÄLLET AND BJÖRKVATTNET UNITS

LOWER LAXFJÄLLET AND BJÖRKVATTNET UNITS

The development of penetrative, NW-plunging lineations, which probably represent the direction of maximum elongation in $S1_{LLU}$, parallel to syn-early schistosity, tight to isoclinal, minor folds ($F1_{LLU}$) is the same as the D2 structural pattern in the Björkvattnet Unit. Furthermore, the deformation of these structures by folds with N-dipping axial surfaces and transposition characteristics close to the thrust contact with the Bångfjället Complex is comparable to the D2-D3 sequence in the lower litho-tectonic unit. $S1_{LLU}$ is, thus, tentatively correlated



with S2 and $S2_{LLU}$ with S3. Comparison of metamorphism-deformation relationships between the two units supports the above correlation; the metamorphism reached its peak prior to development of the early, penetrative foliation in both units ($S1_{LLU}$ in the Lower Laxfjället Unit; S2 in the Björkvattnet Unit) and garnet continued to grow during and after its formation. $F3_{LLU}$ and the late kink structures of the Lower Laxfjället Unit are more confidently correlated with $F4_a$ and $F4_{kink}$ of the Björkvattnet Unit i.e. the structures related to the development of the late Tärna Synform. The correlations above are summarized in Table 6.

UPPER LAXFJÄLLET AND BJÖRKVATTNET UNITS

Tentative correlation of $S1_{ULU}$ with S2 of the Björkvattnet Unit is suggested in Table 6. $S1_{ULU}$, like S2, contains the penetrative, westerly plunging mineral lineation and formed after the main phase of poikiloblast growth (staurolite and garnet in the case of the Upper Laxfjället Unit). The northerly plunging $F2_{ULU}$ folds, which are related to the Tärna Synform, are correlated with $F4_a$, whilst the upright, W-plunging, post-$S1_{ULU}$ folds are probably equivalent to $F4_b$ of the Björkvattnet Unit (Table 6).

CONCLUSIONS

1. The rocks on Laxfjället belong to two separate litho-tectonic units (Lower and Upper Laxfjället Units), both of which lie structurally above the Björkvattnet Unit.
2. The gabbro-intruded calcareous phyllites and metavolcanites of the Lower Laxfjället Unit resemble the upper part of the Lasterfjäll Group, described

from the Remdalen area in the Western Synform. The gabbro-intruded calcareous phyllites lie with tectonic contact on top of the Lövfjäll Formation calcareous phyllites and greywackes of the Björkvattnet Unit. The rocks of the Upper Laxfjället Unit are tentatively correlated with the northern part of Kulling's (1964) Storfjäll Nappe.

3. Late- or post-metamorphic thrusting (late- or post-$S1_{ULU}$ but pre-$F2_{ULU}$) of the Upper over the Lower Laxfjället Units effected major and minor, post-early schistosity folding in the underlying rocks ($F2_{LLU}$ or F3 of the Björkvattnet Unit) associated with local inversion of the metamorphic isograds.
4. The D1, D2 and D3 structural events in the Lower Laxfjället Unit have been correlated with D2, D3 and D4 of the Björkvattnet Unit. The two phases of deformation in the Upper Laxfjället Unit appear to correspond to D2 and D4 of the Björkvattnet Unit.
5. In both units the metamorphism reached its peak prior to formation of the early, penetrative foliation. Metamorphic conditions enabled garnet, mica and hornblende to grow during formation of the early foliation, and garnet and muscovite to be stable after its formation. The relationship between metamorphism and deformation in the higher tectonic units on Laxfjället is, thus, similar to that in the Björkvattnet Unit.

NATURE OF THE BÅNGFJÄLLET COMPLEX AND THE DEFORMATION ASSOCIATED WITH THE BOUNDARY THRUSTS

INTRODUCTION

The large masses of granite gneiss and coarse grained granite on Nasafjäll and Børgefjell, to the north and southwest of Tärnaby respectively (Fig. 1), are exposed as windows beneath the Seve-Köli Nappe Complex and higher tectonic units (Nicholson and Rutland 1969; Zachrisson 1964, 1969). Geochronological work (Priem et al. 1967, 1968; Wilson and Nicholson 1973) on these windows indicates their Pre-Cambrian age and a similarity in age pattern to that of the Baltic Shield rocks to the east. However, there is still debate whether these windows are completely basal and autochthonous in character (see discussion in Wilson and Nicholson 1973).

A similar tectonic explanation has been put forward for the occurrence of lower thrust units beneath the Seve-Köli Nappe Complex, including the Ammarnäs Inlier, the Bångfjället Complex (Kulling *in* Gavelin and Kulling 1955, *in* Magnusson et al. 1960, *in* Strand and Kulling 1972) and the Fjällfjäll Window (Zachrisson 1969) (Fig. 1). These authors suggest that the lower tectonic units are exposed in the eroded cores of late antiforms beneath the metamorphic nappe complex and that these antiforms deform the basal Seve-Köli thrust. According to Zachrisson (1969), the quartzites and arkoses of the Fjällfjäll Window are exposed in the core of a late, upright, F2 antiform (Fjällfjäll Antiform) with NNE—SSW to NE—SW trend. The F2 structures of Zachrisson (1969) have been correlated here with the $F4_a$ major and minor folds of the Tärna—Björkvattnet area.

Only the stratigraphic and structural relations in and immediately adjacent to the southern part of the Bångfjället Complex are discussed in the present study. The Bångfjället Complex rarely exceeds 4—5 km across but extends for approximately 20 km along strike. The igneous rocks of the complex have been compared to the Baltic Shield rocks in the foreland to the east of the Scandinavian Caledonides (Kieft 1952), and dating work has confirmed their Pre-Cambrian age (Stephens and Wilson in prep.).

STRATIGRAPHY AND PETROGRAPHY OF THE BÅNGFJÄLLET COMPLEX

Kieft (1952) divided the Bångfjället Complex into three divisions in the following stratigraphic order:

1. Igneous rocks including quartz syenite, gabbro and subordinate granodiorite.
2. Low-metamorphic, highly feldspathic sediments — predominantly gneisses which surround the igneous rocks of the complex.
3. Quartzite series.

The igneous part of the complex (division 1 of Kieft 1952) is homogeneous and predominantly quartz syenitic in composition; less important gabbro bodies also occur. However, the most significant variation recognised here is the progressive microstructural modification of the igneous part of the complex towards the contacts with the overlying Calcareous Phyllite Formation and underlying Tärna schists. Although there are thin slices of quartzite, sometimes highly feldspathic, and layered feldspathic quartzite/pelite within the igneous complex (Plate I), it is argued below that Kieft's division 2 represents the marginal, foliated and progressively more deformed products (including flaser gneiss and mylonite) of the relatively more homogeneous, central part of the complex.

The sedimentary part — referred to here as the Oltokken Formation (equivalent to Kieft's "Quartzite series") — mantles the igneous rocks along the eastern edge of the complex. This formation has been mapped from Oltokbäcken (8g) to northwestern Göutavardo (8g) where it is folded and sliced into the igneous rocks of the complex (Plate I). The quartzite has been extensively quarried approximately 1.5 km north of the intersection of the Norra Fjällnäs road and Route E79 (Fig. 3). Although the top of the formation is tectonic, the excellent preservation of the quartzites etc. in and around the quarry has permitted a study of its internal stratigraphy; the type-area is thus located here.

QUARTZ SYENITE AND ASSOCIATED ROCKS

The most conspicuous feature of the Bångfjället Complex is the preservation of a more or less central, homogeneous part composed of coarse-grained, quartz syenite with a random, granular texture and anhedral crystal boundaries; there is a transition to a more granitic type in some specimens with quartz content $> 10\ \%$. Due to recrystallization at the peripheries of mineral grains, there is a weak pseudoporphyritic microstructure. The coarser phases include microcline, perthite, orthoclase, less common plagioclase (albite-oligoclase) and micrographic intergrowths of quartz and feldspar. These show minor deformation effects including kinking and misorientation of individual grains, undulose extinction and mechanical twinning in the plagioclase. The mafic minerals are concentrated as isolated aggregates (Fig. 46) often 0.5 cm across and are dominated by randomly oriented sheaves of undulose olive-green biotite, epidote and rare hornblende. Accessories include apatite and sphene.

Finer grained varieties also occur within the complex but they show a tendency towards a more quartz-rich end-member. On northwestern Göutavardo (8g) apophyses of such micrograrite penetrate the coarse, quartz syenite. Thus, these finer grained types may represent a slightly more acidic, late-stage development during the total emplacement of the predominantly quartz syenitic intrusion.

Besides the severe saussuritization of the coarser feldspars, there is a variable amount of secondary white mica and chlorite associated with the mafic knots. The recrystallization of quartz in the groundmass is accompanied by the growth of secondary, pellucid albite rims on original, strongly saussuritized, feldspar crysts. Radial clusters of undeformed, deep yellow-brown stilpnomelane are occasionally present.

BASIC INTRUSIVE ROCKS

Although the exact size and shape of the coarse-grained gabbros within the central part of the complex are uncertain, they do appear to be strongly subordinate to the quartz syenite.

The less deformed varieties are composed of an interlocking aggregate of augite, commonly intergrown with hornblende and with a poorly preserved, ophitic relationship to the plagioclase, green-brown hornblende and strongly saussuritized plagioclase; there is a variable amount of biotite, which forms embayed margins adjacent to augite and hornblende, and the mafic minerals are often concentrated into melanocratic knots. In the more deformed and recrystal-

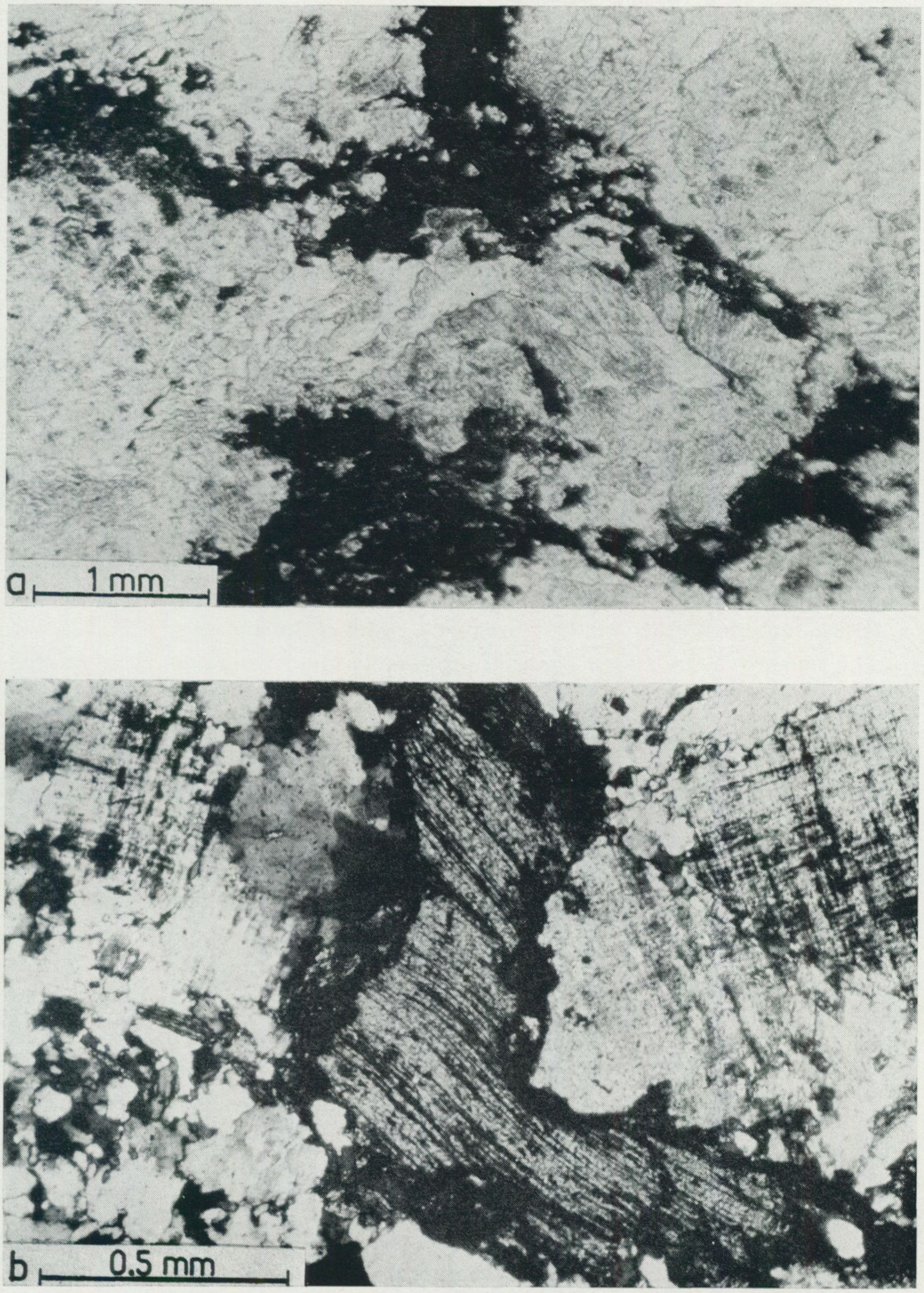


Fig. 46. Petrographic details of the relatively undeformed quartz syenite. a: Note random texture and concentration of mafic minerals into melanocratic knots (west of the bridge over the Tärnaån near Tärnaby, 8f, uncrossed nicols). b: Deformed sheaves of olive-green biotite in quartz syenite (NW Göutavardo, 8g, crossed nicols).

lized types, there is no clinopyroxene and the coarser hornblende relics possess very pale blue-green actinolite and chlorite as irregular patches within and as rims around the individual grains. The foliated fine-grained matrix is composed of oriented chlorite and white mica as well as leucoxene and epidote.

OLTOKKEN FORMATION

Character in the type-area (quartzite quarry)

The Oltokken Formation has been divided into four parts which, on the basis of cross- and graded bedding evidence, have been arranged in the following stratigraphic order:

1. The basal beds consist of a coarse-grained (3—5 mm on average), poorly sorted micro-breccia/conglomerate of arkosic composition. There are abundant fragments of quartz, showing deformed old grain-new grain relationships, and K-feldspar including microcline and microcline perthite; the feldspars appear fractured but their original clastic shapes are preserved. The matrix is an aggregate of recrystallized quartz and feldspar. This basal member lies adjacent to the quartz syenite immediately south of the quarry.
2. The coarse-grained arkose passes transitionally upwards into an even-grained white quartzite with variable feldspar content (c. 10 %) and intercalated pelite; cross-bedding occurs in the quartzites in the southern face of the quarry (Fig. 47).
3. The quartzite passes upwards to a finely layered succession of the same feldspathic quartzite, grey phyllite (pelitic composition) and green-grey siltstone. The fine-grained phyllites are composed of low-grade mineral assemblages including quartz + fine-grained white mica ± minor biotite. The grey phyllite layers progressively increase in importance towards the top of the layered sequence, beds approximately 1 m thick being common.
4. The youngest beds in the quarry consist of a thickly bedded succession of dark, occasionally graded quartzites and subordinate, fine-grained pelites. The dark quartzites are dominated by porphyroclasts of quartz, commonly blue or dark in colour, microcline, orthoclase and microcline perthite which show the same microstructures as the coarser fragments in the basal arkose. The dark colour of these sediments is due to disseminated graphite.

The fine-grained pelite intercalations throughout the whole formation are well cleaved. The dominant foliation in outcrop is a crenulation cleavage which deforms an earlier penetrative surface of uncertain origin defined by dimensionally oriented quartz and mica.

The whole formation has a minimum thickness of 130 m in the type-area but elsewhere it is either strongly attenuated or completely absent.

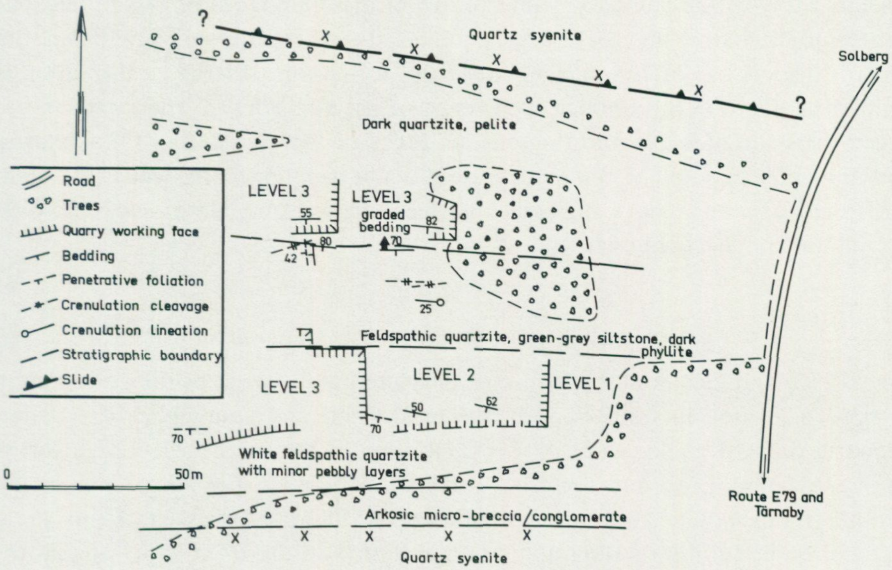


Fig. 47. Sketch map of the eastern part of the quarry in the Oltokken Formation.

Character outside the type-area and relationship to the igneous rocks of the complex

On Göutavardo (8g) and in Oltokbäcken (8g) there are attenuated occurrences of different parts of the succession described above, the layered unit being conspicuously represented on Göutavardo. However, the basal arkose is also exposed in Oltokbäcken, just upstream from the bridge, where it again lies immediately adjacent to the igneous part of the complex, All the rocks in this strip are, however, more deformed than their counterparts in the quarry section (Fig. 48); the coarser quartz clasts are either completely recrystallized or possess a strong development of subgrain boundaries with intense undulose extinction in the old grains, the feldspar clasts are kinked and misoriented with new grains along narrow zones within the old grains and the matrix is also strongly recrystallized. A penetrative foliation in the subordinate pelite laminae is deformed into a later crenulation cleavage which is subparallel to the zones of intense deformation in the more competent layers.

Tectonic slices of sedimentary material are also present within the igneous complex to the south of the main exposures of the Oltokken Formation (Plate I). Lithologically they resemble the massive, feldspathic quartzite and layered quartzite/pelite sediments found in the middle of the Oltokken Formation in the type-area.

The stratigraphically younger age of the Oltokken Formation relative to the igneous rocks of the complex is inferred from the abundance of clastic feldspar fragments, including microcline perthite, in the quartzites etc. and from the general disposition of lithofacies within the formation. As far as the latter is concerned, there is the critical occurrence of a basal arkose. The contacts between the quartzites etc. and the igneous rocks of the complex in Oltokbäcken and immediately south of the quarry are thought to be primary, whilst the main thrusting appears to have occurred between the variably deformed Oltokken Formation and the Tärna schists.

Comparison with the stratigraphy of the Caledonian front

Gee (1972) and Gee et al. (1974) have presented an account of the stratigraphy of the Tåsjö area in northwestern Västernorrland and southern Västerbotten which is situated geologically in the front zone of the main Caledonian thrust sheets. A comparison and tentative correlation between the succession in the Oltokken Formation and the stratigraphy in the controversial Tåsjö area, together with Kulling's stratigraphic nomenclature in the front zone, are shown in Fig. 49. The basal Risbäck Group and the tillite are not represented in the Bångfjället Complex, the lowermost part of the Oltokken Formation being characterized by an arkosic micro-breccia/conglomerate. The remainder of the succession appears similar to the quartzite-shale succession of the late Pre-Cambrian to Cambrian Gärdsjön Formation.

DEFORMATION ASSOCIATED WITH THE BOUNDARY THRUSTS TO THE BÅNGFJÄLLET COMPLEX

BÅNGFJÄLLET COMPLEX

Although there are narrow restricted shear zones within the more homogeneous part of the Bångfjället Complex, the most important modification of the random, granular texture of the quartz syenite occurs particularly near the margins of the complex, i.e. towards the boundary thrusts. The process may be summarized by the progressive development of a foliated structure (mylonitic foliation) enclosing relic feldspar porphyroclasts, by the relative increase in the ratio of recrystallized matrix to coarser grained porphyroclasts and by the progressive decrease in overall grain size and darkening in rock colour. Stilpnomelane, white mica, chlorite and epidote are conspicuous in this group of highly deformed rocks. The progressive sequence, relatively undeformed quartz syenite to flaser gneiss to mylonite, is discussed in terms of four rather arbitrarily defined stages:

1. The development of pseudoporphyritic microstructure (Fig. 50a); this often occurs within the main mass of relatively undeformed quartz syenite and has already been described.

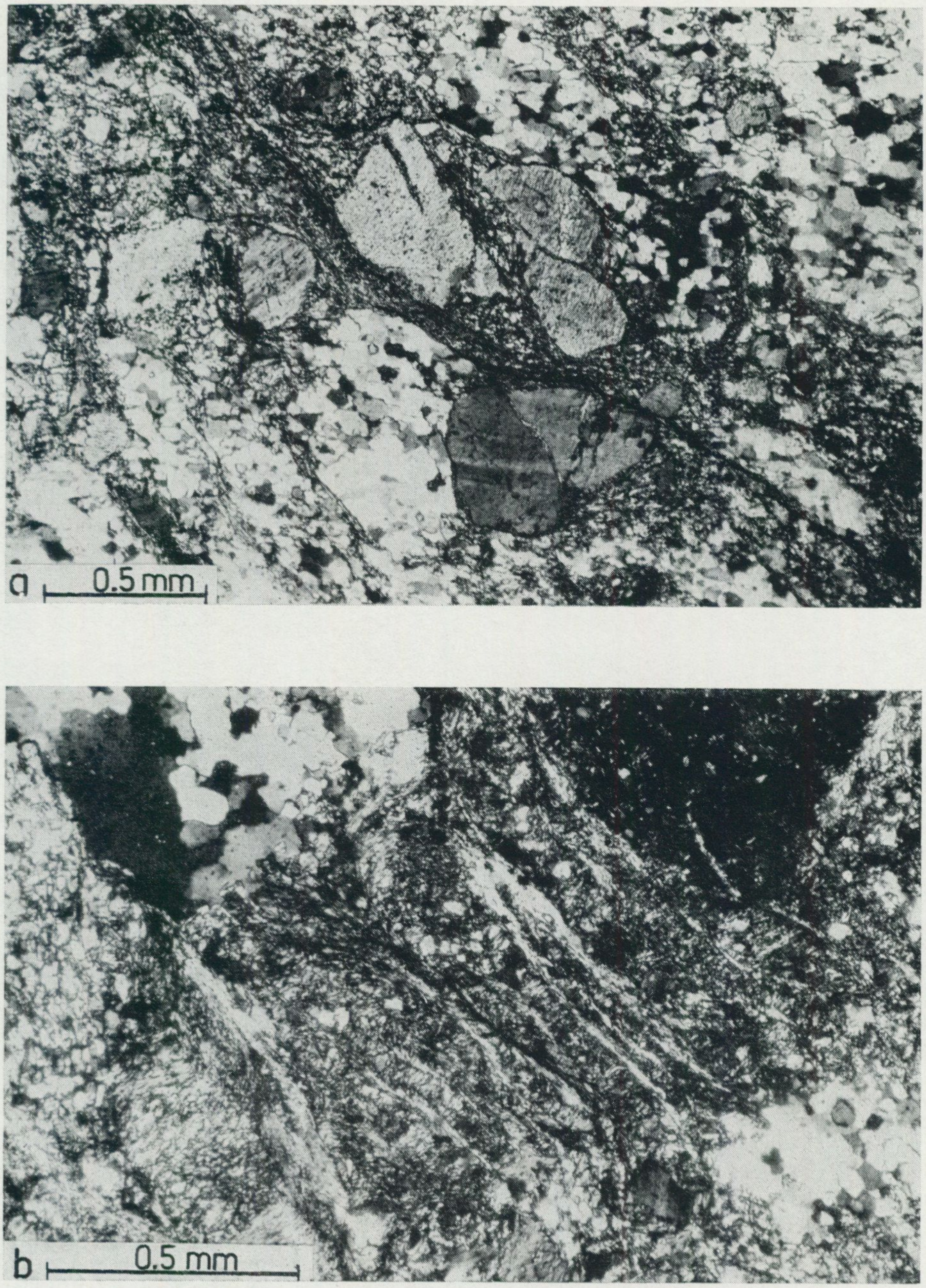


Fig. 48. Deformation characteristics in the arkosic micro-conglomerate adjacent to the quartz syenite in Oltokbäcken. a: Note development of narrow deformation zones (NW—SE across photograph), fine-grained recrystallized matrix and deformation and recrystallization of the quartz and feldspar clasts (Oltokbäcken, 8g, crossed nicols). b: Deformation of the early cleavage in a pelite lens parallel to later crenulation cleavage (Oltokbäcken, 8g, crossed nicols).

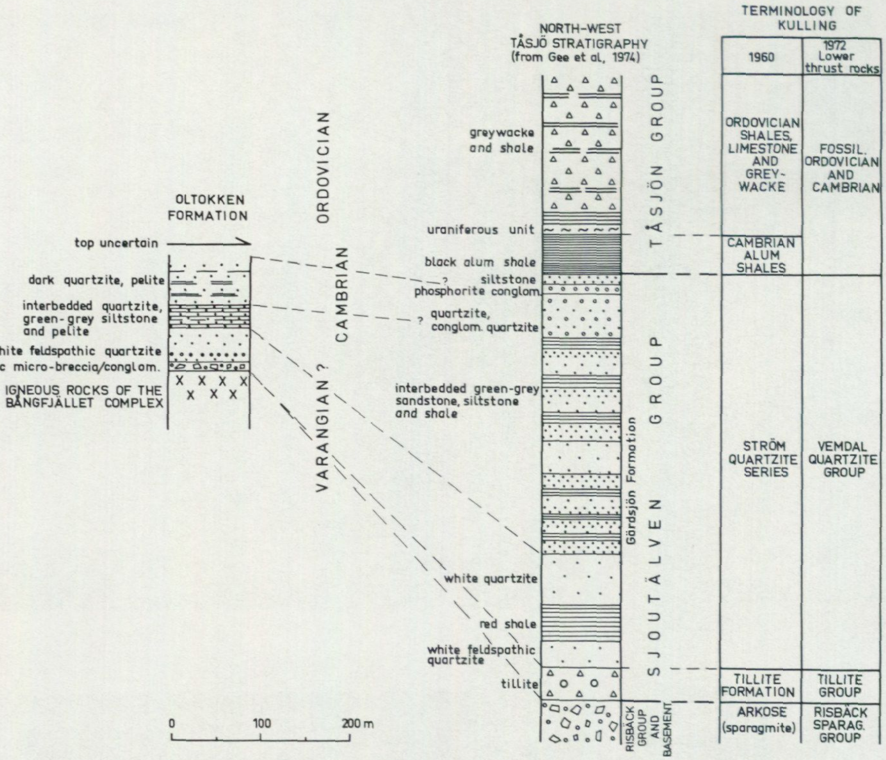


Fig. 49. Comparison of the stratigraphy in the Oltokken Formation with that in the Eastern Caledonian front.

2. The random, granular texture is modified into a distinctly foliated microstructure with schlieren of biotite and epidote enclosing severely misoriented and faintly elongate fragments of relic feldspar (Fig. 50b). The latter is again dominated by K-feldspar and perthite. Deformation twin planes in rare plagioclase grains are either lensed and discontinuous or step faulted. Although the coherence of the feldspar grains is broadly retained, any original quartz is completely recrystallized. Finally, the matrix/strained porphyroclast ratio is notably increased.
3. The foliation is now the dominant feature of the rock and some coarser

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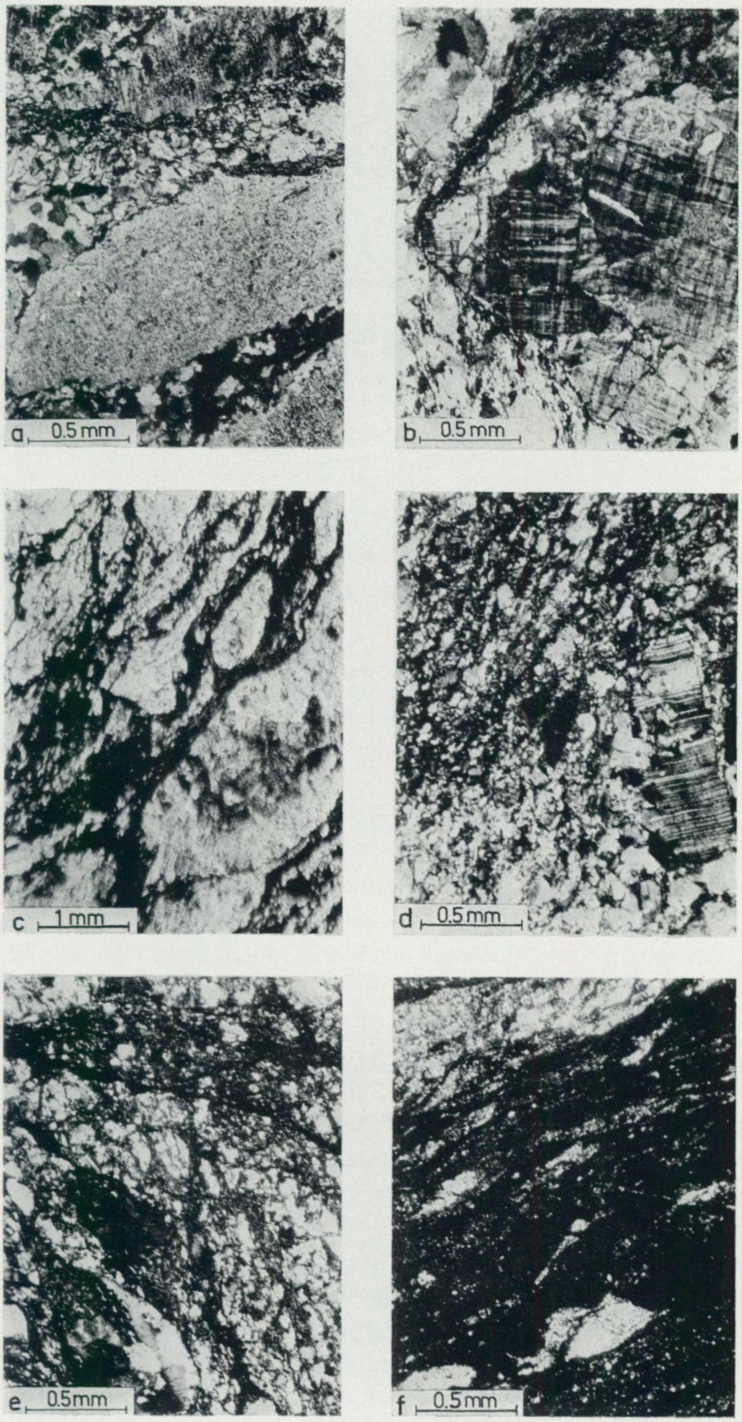
Fig. 50. Deformation of the quartz syenite towards the margins of the Bångfjället Complex. a: Stage I — development of pseudoporphyritic microstructure; note misorientation (kinking) of coarse feldspar crysts and the development of new grains around deformed old grains (west of the bridge over the Tärnaån near Tärnaby, 8f, crossed nicols). b: Stage II (bridge over Oltokbäcken, 8g, crossed nicols). c: Stage III — flaser or augen gneiss microstructure (Tärnaby, 8e, uncrossed nicols). d: Stage IV — mylonite; note mechanical twins in the relic plagioclase cryst (Tärnaby, 8f, crossed nicols). e: Stage IV — mylonite; anastomosing mylonitic foliation encloses relic feldspar crysts and minor, discordant, pseudotachylite veins (upper part of figure) are present (Tärnaby, 8f, crossed nicols). f: Discordant pseudotachylite vein in mylonite (Tärnaby, 8f, crossed nicols).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e 0.5 mm
f 0.5 mm

Fig. 51. Relic, coarse, brown-green hornblende (hb) in schistose white mica-actinolite-epidote groundmass (Tärnaby, 8f, crossed nicols).

feldspar grains are so recrystallized that identification of their original character is difficult. White mica is oriented along the foliation, whilst the relic feldspar porphyroclasts are augen-shaped. The fine-grained quartz-feldspar matrix occupies a greater proportion of the rock which is now a typical flaser gneiss (Tyrrell 1926; Christie 1960) (Fig. 50c; compare Fig. 46).

4. The final stage is characterized by the development of a dark, fine-grained rock with a strongly foliated appearance as above, but with only scattered relic feldspar porphyroclasts and with greater than 50 %, fine-grained recrystallized matrix (Fig. 50d, e). Particularly on the thrust contacts, these mylonites (Lapworth 1885; Christie 1960) are often transected by apparently later, discordant veins of dark, amorphous material — hyalomylonite or pseudotachylite (Higgins 1971) — containing relic, crushed and angular fragments of the host rock (Fig. 50f). Other cross-cutting veins are composed of dark green chlorite, epidote and fine-grained, deep brown-yellow stilpnomelane.

Around the more deformed edges of the complex there are also occurrences of fine-grained greenschist within the main mass of flaser gneiss. These more basic rocks are composed of strongly oriented actinolite, chlorite and epidote-zoisite as well as variable amounts of white mica, biotite, leucoxene and feldspar; coarser relics of a green-brown hornblende with a near colourless amphibole rim in optical continuity are preserved in the foliated groundmass (Fig. 51). It is considered that these are deformed and recrystallized relics of the more gabbroic lithotype.

Kieft (1952) considered that the gneisses which surround the truly igneous part of the complex, and which correspond to the flaser gneiss and mylonite described here, are low-grade, highly feldspathic sediments equivalent to the sparagmites (Risbäck Group, see Fig. 49) in the Caledonian front zone, to the east. However, it is apparent from the above that there is a microstructural transition between the flaser gneiss (and mylonite) and the homogeneous, quartz syenite. Other arguments in favour of the gneisses being the deformed derivatives of the quartz syenite include the following:

1. In Oltokbäcken (8g), where the Oltokken Formation mantles the complex and is itself strongly deformed (see Fig. 48), there is virtually no occurrence of flaser gneiss. The homogeneous quartz syenite, protected by the quartzite from the main thrust, is preserved virtually right up to the contact with the basal arkose and quartzites.
2. The most conspicuous development of flaser gneiss and mylonite is around Tärnaby. This is consistent with the wedging out of the Bångfjället Complex, the mergence of the enveloping thrusts and the corresponding greater deformation in this area.

The effect of Caledonian metamorphism on the igneous part of the complex is uncertain, but since stilpnomelane is present as both well preserved radial clusters and in vein form in the more deformed rocks, it is probable that this mineral, at least, is related to the Caledonian movements. According to Winkler (1967), stilpnomelane is restricted to a high-P, low-T environment and commonly confined to the chlorite zone of the classical Barrovian sequence. Murris (1957) has also described similar occurrences of this mineral in association with thrust zones in the Gieravardo—Jofjället region to the northwest of the Tärna—Björkvattnet area. The low-grade metamorphism of the igneous rocks is consistent with the low-grade mineral assemblages observed in the pelite layers within the Oltokken Formation.

SURROUNDING METASEDIMENTS OF THE SEVE-KÖLI NAPPE COMPLEX

Calcareous Phyllite Formation, Lower Laxfjället Unit

The early, phyllitic cleavage ($S1_{LLU}$) in the pelites towards the base of the formation is deformed in a variety of structures which are related to the intense deformation near to the thrust contact between the Lower Laxfjället Unit and the Bångfjället Complex; the rocks near the base of the Calcareous Phyllite Formation are, indeed, phyllonitic.

The most conspicuous feature in the less deformed types is a micro-boudinage structure (Fig. 52a) which is considered to be an early development of the characteristic fish-scale microstructure (Roper 1972) of the more deformed phyllonites (see also the Tärna schists below). The $S1_{LLU}$ phyllosilicates are deformed in the micro-boudinage structure into lenticular aggregates, the boundaries of which appear to be zones of shearing; this is accompanied by recrystallization of the quartz-rich matrix.

Strongly attenuated, isoclinal, minor folds (Fig. 52b), in which $S1_{LLU}$ is deformed and N-dipping shear discontinuities are developed parallel to the limbs, occur near to the thrust contact. These folds are still intact, whilst others, of more uncertain origin, are the intra-folial type (Fig. 52c) where the early cleavage is deformed in a rootless fold which is enclosed between the enveloping foliae of a crenulate-style, transposition foliation ($S2_{LLU}$). In the mylonitic phyllites, where these intra-folial structures are well developed, the fish-scale microstructure is intense and perhaps related to the attenuated limbs of isoclinal folds.

Significantly in the Calcareous Phyllite Formation the early cleavage, which is deformed by the later deformation, is a fine-grained, quartz-white mica-chlorite assemblage. There is no evidence for a retrogressed, early schistose fabric with associated growth of biotite and garnet porphyroblasts. This situation is quite distinct from that in the Tärna schists which rim the Bångfjället Complex on its southern and eastern margins.

Tärna schists, Björkvattnet Unit

Although the sheet dip (Ramsay 1967, p. 351) of S0/S2 is steep in Oltokbäcken (8g) and south of Tärnaby (7f), the Tärna schists lie structurally beneath the Bångfjället Complex on northwestern Göutavardo (8f—g). Mylonitic green-schists and feldspar-rich rocks are discontinuously developed along the contact between the Bångfjället Complex and the garnetiferous schists proper. West of Tärnaby, where the Bångfjället Complex wedges out and the boundary thrusts merge to form a single thrust contact separating the Lower Laxfjället and Björkvattnet Units, there is subsequent thinning out of stratigraphic units, including the Tärna schists, beneath this contact.

Isoclinal minor folds, with S2 as an axial surface structure and a quartz rodding lineation parallel to the fold axis, are preserved within the Tärna schists, away from the contact with the Bångfjället Complex. However, the most conspicuous minor folds deform S2 on N-dipping axial surfaces and plunge NW to SW approximately parallel to the strike of the major thrusts (Plate III, 6b). These minor folds become tight to isoclinal or intra-folial in shape near to the contact with the Bångfjället Complex and the main foliation is apparently com-

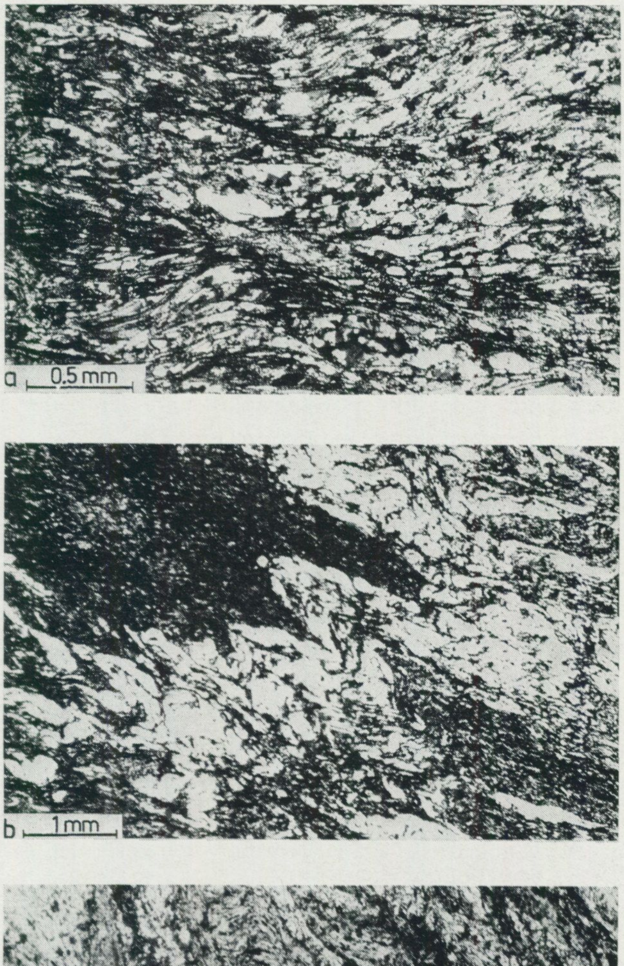

Fig. 52. Deformation associated with the thrust contact beneath the Calcareous Phyllite Formation. a: Incipient phyllonitic structure in relatively undeformed, biotite-rich, calcareous phyllite (north of Tärnaby, 8f, crossed nicols). b: Isoclinal, post-$S1_{LLU}$ folds in calcareous phyllonite (road from Tärnaby to Konäset, 8e, uncrossed nicols). c: Intra-folial structure enclosed within the $S2_{LLU}$ transposition foliation in calcareous phyllonite (Tärnaby 8e, uncrossed nicols).

posite (Fig. 53a). However, in certain layers within the mylonitic greenschists, it is clear that an earlier schistosity is transposed parallel to a later crenulate-style foliation (Fig. 53a, b) which forms either an axial surface structure to isoclinal folds or an enclosing surface to intra-folial folds. The orientation of the post-S2 folds is consistent with their refolding in the $F4_a$ Tärna Synform (Plate III, compare 1e and 6b). Thus, they are thought to belong to the F3 phase and the complex foliation within the schists is considered to be the S2 surface highly transposed in the S3 axial surface orientation.

The characteristic feature of the Tärna schists is, however, the phyllonitization and retrogression of the coarse metamorphic fabric towards the contact with the Bångfjället Complex. The following microstructural relationships are preserved:

1. Isolated muscovite and albite porphyroblasts are lenticular in shape and the former show severe undulose extinction (Fig. 53c).
2. Lenticular aggregates (fish-scale microstructure) of fine-grained muscovite display partial to complete polygonization and represent either partially or completely recovered single crystals or deformed lenses of muscovite (Fig. 53d). Again in the less deformed types the incipient development of this structure appears to be micro-boudinage and shearing of the early schistosity (S2). The quartz-feldspar matrix displays coarser deformed areas with ribbon structure but is dominated by strain-free, recrystallized material with straight and sharp grain boundaries and relatively even grain size.
3. Conspicuous ribbon structure is present in the more quartzitic layers.
4. Garnet porphyroblasts are fractured and appear flat and elongate in the complex mylonitic foliation. A similar structure has been described by Dalziel and Bailey (1968) and Ross (1973) for other such deformed garnets in mylonitic rocks.
5. The garnet and biotite porphyroblasts are severely retrogressed.

POST-THRUSTING DEFORMATION

On Göutavardo the phyllonite zone within the Tärna schists and the thrust contact between the schists and the Bångfjället Complex are folded by a complimentary set of major, close to tight folds (Plates I and III). These folds deform the mylonitic foliation in the igneous rocks of the complex and the main foliation (S2 or transposed S2) in the schists. Furthermore, in the quartzite quarry, in the

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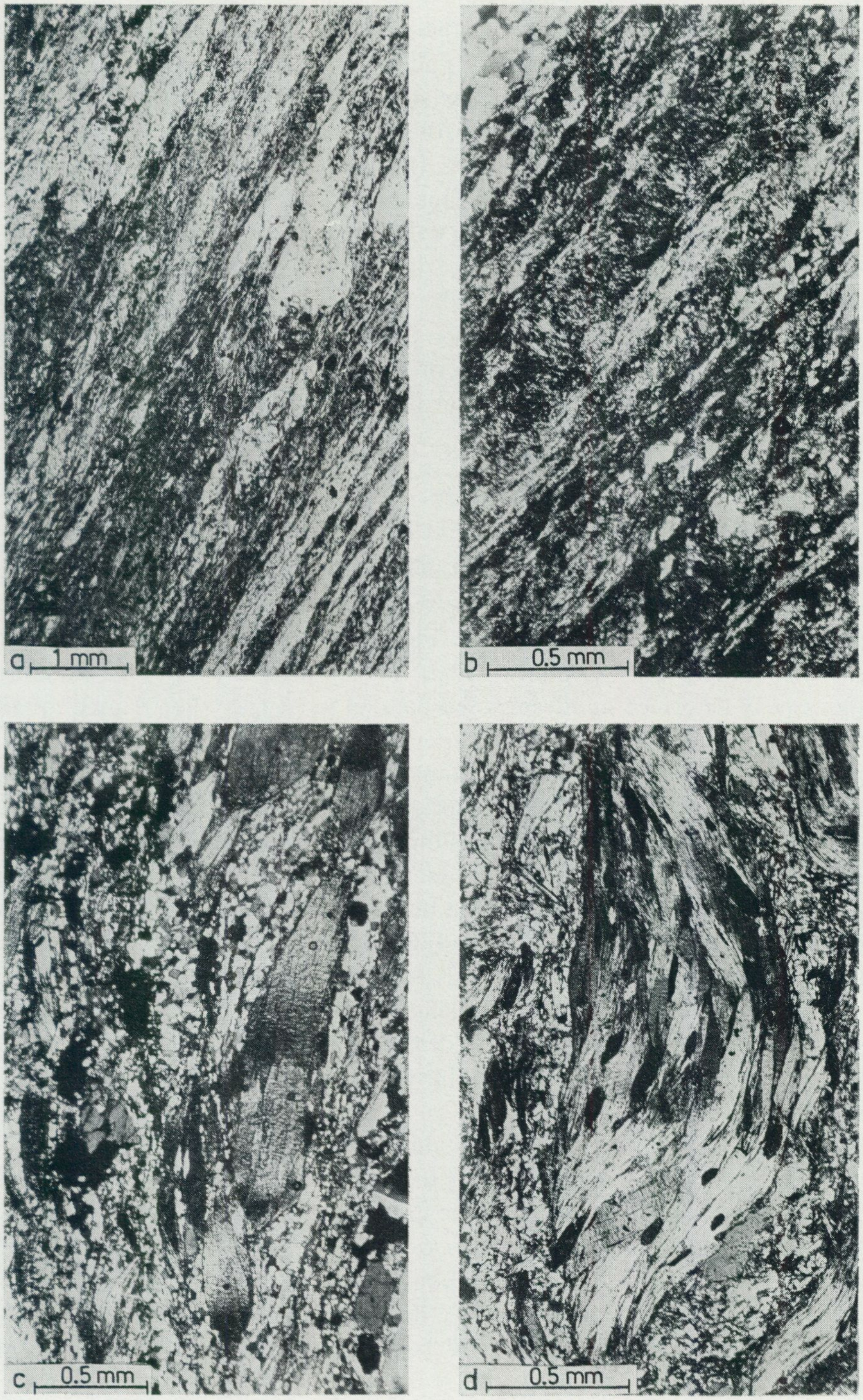
Fig. 53. Deformation associated with the thrust contact between the Bångfjället Complex and the Tärna schists. a: Relic intra-folial hinges in mylonitic greenschist; main foliation is S2 transposed into S3 (Oltokbäcken, 8g, uncrossed nicols). b: Growth of white mica and chlorite parallel to S3 which transposes the S2 micas in a crenulate-style foliation (Oltokbäcken, 8g crossed nicols). c: Intense phyllonitic (fish-scale) microstructure; the muscovite porphyroblasts are strongly deformed (south of Tärnaby, 7f, crossed nicols). d: Lenticular aggregate of smaller, polygonized crystals of muscovite; the muscovites here have partially recovered their pre-strain optical properties (south of Tärnaby, 7f, crossed nicols).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core of the main antiform, both the bedding and the penetrative cleavage are deformed and the main foliation in the pelites is a crenulation cleavage. The Göutavardo folds (Plate III) appear to plunge WSW to SW and to be related to relatively steep N- to NW-dipping axial surfaces (Plate III, 6a and 7). The deduced major fold orientation is consistent with the orientation of minor folds within the Bångfjället Complex (Plate III, 7) and F3 folds within the Tärna schists (Plate III, 6b). Although the Göutavardo folds appear to belong to the F3 phase, they are incongruent to the larger F3 structure to the SE (Ruffevare—Björkfjället Synform Complex).

All the above structures, including the axial surfaces of the Göutavardo folds (Plate III), are folded around the late ($F4_a$ of the Björkvattnet Unit) Tärna Synform. Indeed, the outcrop pattern of the Bångfjället Complex is controlled by the late synform, being situated on its eastern limb and deformed around its nose east of Tärnaby (Plate I).

SUMMARY OF METAMORPHISM-DEFORMATION RELATIONSHIPS

The microstructural relationships in the Tärna schists indicate that the retrogression of the schistose fabric and the associated mylonite formation (phyllonites) are post-metamorphic. The foliation evidence from all the metasediments adjacent to the Bångfjället Complex indicates that the S2 ($S1_{LLU}$) foliation is progressively deformed and transposed parallel to S3 ($S2_{LLU}$). This transposition foliation is parallel to the axial surfaces of isoclinal folds (and related intrafolial structures) which plunge approximately parallel to the strike of the major thrusts. The transposition foliation and associated minor folds are particularly well developed where the Bångfjället Complex wedges out around the village of Tärnaby. The main character of the metasediments adjacent to the complex, however, is the lenticular or fish-scale microstructure, the early form of which appears to be post-S2 ($S1_{LLU}$) micro-boudinage, probably related to deformation on the limbs of the isoclinal folds. The transposition foliation in the surrounding metasediments to the Bångfjället Complex and the mylonitic foliation developed within the complex are folded around the late Tärna Synform. The mylonites, therefore, are thought to have formed during the D3 deformation phase of the Björkvattnet Unit.

The low-grade metamorphic condition of the Bångfjället Complex and the discordance of the boundary thrusts to the metamorphic isograds (Fig. 31) indicate the post-metamorphic nature of the thrusting. Since the boundary thrusts are folded around the late Tärna Synform, the main thrusting episode probably also occurred during D3 ($D2_{LLU}$). This accords with the F3 folds in the Björkvattnet Unit to the south of Tärnaby which have important slides on their limbs, are overturned to the S and SW and have the same axial orientation as the conspicuous minor folds in the vicinity of the boundary thrusts.

From the above, it appears that the mylonitization within the Bångfjället Complex and surrounding metasediments, the F3 ($F2_{LLU}$) folding, and the thrusting are all closely related both in time and space. Assuming that the Göutavardo folds are late F3 structures, the evidence on Göutavardo suggests that the folding locally continued after formation of the mylonites and major thrusting.

DISCUSSION

TECTONIC SETTING OF THE BÅNGFJÄLLET COMPLEX

There is a remarkable contrast between the metamorphic state of the rocks in the Bångfjället Complex and that in the schists which are immediately adjacent to the basal thrust and which belong to the Seve-Köli Nappe Complex. The low-grade nature of the basement rocks is inconsistent with the hypothesis that they were intimately involved with the present higher grade cover to form the core of a pre- or syn-metamorphic, anticlinal fold nappe. A study of the structural features both within and immediately adjacent to the complex indicates the severe post-metamorphic deformation associated with thrusting during D3 ($D2_{LLU}$).

Perhaps one of the more interesting phenomena concerning the geometry of the Bångfjället Complex is its position in the late Tärna Synform. It is within related, N- and S-plunging antiforms that the Børgefjell and so-called Fjällfjäll Windows have been situated (Fig. 14). The rocks of the Bångfjället Complex occur, however, on the eastern limb and, together with the boundary thrusts, are deformed around the hinge zone of the northerly plunging Tärna Synform. The Bångfjället Complex is not, therefore, exposed in the eroded core of a late, N-plunging antiform.

The following points are considered important in any discussion of the tectonic setting of the Bångfjället Complex:

1. The Calcareous Phyllite Formation above the Bångfjället Complex is distinct, both as regards composition and the nature of the pre/syn-S2 ($S1_{LLU}$) metamorphic fabric, from the predominantly underlying Tärna schists.
2. The Layered Metavolcanic Formation, which lies with transitional contact above the Calcareous Phyllite Formation on Laxfjället, is absent beneath the Bångfjället Complex.
3. West of Tärnaby, where the Bångfjället Complex wedges out, the boundary thrusts are not folded around the closure of the complex, rather they merge to form a single tectonic contact striking NW towards the Umeälven valley (Plate I).

The different sequences above and beneath the Bångfjället Complex are also inconsistent with the hypothesis that the rocks of the complex are exposed in a simple, late antiform. The structural relationships around Tärnaby suggest a more complex situation than that already described for the occurrence elsewhere of lower tectonic units within the Seve-Köli Nappe Complex (Kulling, *in* Gavelin and Kulling 1955, *in* Magnusson et al. 1960; Zachrisson 1969).

It is suggested that the closure of the Bångfjället Complex just west of Tärnaby is a relic, F3 ($F2_{LLU}$), antiformal hinge complementary to the Ruffevare—Björkfjället Synform Complex to the south and southeast. This is consistent with the concentration of F3 ($F2_{LLU}$) minor folds and the strong transposition of S2 ($S1_{LLU}$) into S3 ($S2_{LLU}$) in the area around Tärnaby village. Furthermore, the southerly sense of overturn of this antiform is the same as the F3 folds in the Ruffevare—Björkfjället Synform Complex. It is considered that the boundary thrusts to the Bångfjället Complex developed as major slides parallel to the limbs of the antiform during fold development, such that now the Bångfjället Complex is a displaced tectonic slice of Pre-Cambrian igneous rocks together with their original autochthonous cover, within the metamorphic nappe complex. The Bångfjället Complex is tentatively rooted in the autochthonous basement rocks of the Baltic shield which underlie the metamorphic allochthon in this part of the Västerbotten Caledonides. This hypothesis is consistent with the divergent metasedimentary sequences on either side of the Bångfjället Complex and the mergence of the boundary thrusts to form a single tectonic break northwest of Tärnaby. Thus, the boundary thrusts are not necessarily the folded portions of the Seve-Köli thrust at the base of the Seve-Köli Nappe Complex. Instead they appear to control the "sliced fold' or imbricated nature of the basement rocks which have been displaced upwards into the metamorphic allochthon, relative movement occurring between basement and metamorphosed cover rocks along both the upper and lower tectonic contacts. The postulated relationship between the Bångfjället Complex and the metamorphic allochthon is illustrated in the main cross-section (Plate I, A—C).

TIMING OF THRUSTING

The model above implies, of course, that thrusting of the Seve-Köli Nappe Complex over the basement rocks preceded overturned folding and slicing of basement and metamorphosed cover together. Data on the timing of low-angle thrusting in the Seve-Köli Nappe Complex is sparse. Zachrisson (1969) stated that local and possibly more regional thrusts developed during or slightly younger than F1 but prior to F2 (late- or post-F2, pre-F4 using the terminology for the Björkvattnet Unit in this study). Trouw (1973) believed that the thrust contact between the Köli and Seve rocks, the intra-Seve thrusts and the basal Seve-Köli thrust were F2 in age (F3 in this study). Furthermore from the data presented earlier, relative movement on a low-angle thrust surface occurred between the

Upper and Lower Laxfjället Units during late- or post-$S1_{ULU}$ but pre-$F2_{ULU}$ (late- or post-S2, pre-F4 using the Björkvattnet Unit terminology). In summary, it appears that the Seve-Köli Nappe Complex was thrust into position on top of the Baltoscandian platform at the end of D2 or early during D3. Thus, D3 folding and slicing of basement and cover closely followed emplacement of the metamorphic allochthon over the platform.

HYPOTHESIS FOR THE TECTONIC EMPLACEMENT OF THE BÅNGFJÄLLET COMPLEX

It is considered that two factors are important in understanding the tectonic emplacement of the Bångfjället Complex:

1. The complex is situated just beneath the Upper Laxfjället Unit (Storfjäll Nappe?).
2. There is a close temporal and spatial relationship between the development of mylonite zones around the Bångfjället Complex, the F3 ($F2_{LLU}$) folding and the major thrusting; in particular the axial orientation and sense of overturn of the folds is consistent with the presumably NW—WNW to SE—ESE movement on the major thrust at the base of the Upper Laxfjället Unit.

The following model is suggested for the tectonic emplacement of the Bångfjället Complex during D3 ($D2_{LLU}$):

1. At the end of D2($D1_{LLU}$) or early during D3($D2_{LLU}$) the Upper Laxfjället Unit and the underlying Seve-Köli sequence were transported eastwards, on separate low-angle thrusts, over the underlying crystalline basement with its thin veneer of Late Pre-Cambrian to Lower Palaeozoic sediments.
2. Major and minor, F3 ($F2_{LLU}$) folding developed in the lower Seve-Köli sequence beneath the Upper Laxfjället Unit. This folding was related to the relative movement between the Upper Laxfjället Unit and the lower Seve-Köli sequence. The folding involved basement participation, the basement rocks coring the major antiformal structure. The mylonites adjacent to and within the Bångfjället Complex are thought to have developed during the folding along zones of intense deformation associated with the F3 ($F2_{LLU}$) antiform. Most of the microstructures within the different mylonites indicate a ductile mode of deformation together with recovery/recrystallization processes (Bell and Etheridge 1973) which took place under relatively low strain-rate conditions; a ductile mode is also suggested from the folding.
3. During a later stage there was a change-over to a high strain-rate cataclastic mode of deformation. This deformation involved thrusting of the basement rocks into their present position within the metamorphic allochthon, with

the resultant tectonic slicing of the lower Seve-Köli sequence into the separate Björkvattnet and Lower Laxfjället Units. The cataclastic deformation accounts for the concentration of pseudotachylite veins at the thrust contacts. Movement occurred along the zones of intense deformation associated with the formation of the mylonites.

It is possible that the change-over from a ductile (folding-mylonite formation) to cataclastic (thrusting) mode of deformation did not simply occur in a single stage, rather the two modes of deformation developed simultaneously and the thrusting took place as a series of discrete steps in the overall deformation. It has been suggested that the intense deformation or increased ductility in mylonite zones may be due to the movement and concentration of fluids (Beach 1973). The change-over at any particular stage in the deformation from a ductile to cataclastic mode may be controlled by a build up of excess fluid pressures along mylonite zones. Thrusting of the basement slice might then initiate according to the mechanism of Hubbert and Rubey (1959) on attainment of some critical pore fluid pressure.

It is thought that an important contributor to these fluids might be the water released from the metamorphic reactions in the Seve rocks (see discussion earlier). For example, the water may be trapped along the basement-cover contact on the southeastern limb of the antiform, where the basement rocks are locally overturned on top of the Seve-Köli sequence. The resulting weakness along this contact would localize subsequent deformation there giving rise to the formation of mylonites in a zone of intense deformation on a limb of the antiform. The movement of the Upper Laxfjället Unit over the underlying Seve-Köli sequence would not only initiate the F3 ($F2_{LLU}$) folding but would also increase the overburden pressure in the rocks lying beneath the higher thrust unit. The relative increase in overburden pressure would be transferred to the pore pressure in the underlying rocks. In zones where the fluids cannot escape (e.g. the overturned limb of the antiform along the lower basement-cover mylonite zone), the pore pressure may approach the overburden pressure and effect thrust faulting (Hubbert and Rubey 1959).

CONCLUSIONS

1. Consideration of the internal stratigraphy of the Bångfjället Complex has shown that an alternating series of feldspathic quartzites and pelites with a basal, arkosic micro-breccia/conglomerate (Oltokken Formation) unconformably overlies a relatively homogeneous igneous rock sequence of Pre-Cambrian age which is intensely mylonitized at the margins of the complex.

The stratigraphy of the Oltokken Formation is thought to be similar to the Late Pre-Cambrian to Cambrian Gärdsjön Formation in the autochthonous and parautochthonous front zone below the metamorphic nappes to the east.

2. There is a close temporal and spatial relationship between development of the mylonites adjacent to and within the Bångfjället Complex, the major thrusting at the margins of the complex and tight to isoclinal folding with southerly sense of overturn and fold axes approximately parallel to the strike of the major thrusts. It appears that all these structures occurred during D3 of the deformation sequence defined in the Björkvattnet Unit. They are post-metamorphic and are all refolded by the late Tärna Synform. Most of the microstructures within the mylonites imply a ductile mode of deformation. Cataclasis related to the thrusting is evident from the presence of pseudotachylites virtually adjacent to the thrust contacts.
3. It is suggested that the Bångfjället Complex is exposed in the core of a major F3 ($F2_{LLU}$) antiform which, during progressive deformation, developed mylonite zones and subsequently major thrusts parallel to its limbs. The Bångfjället Complex is now exposed as a tectonic slice within the Seve-Köli Nappe Complex.
4. It is considered that the Bångfjället Complex is rooted locally in the autochthonous basement rocks of the Baltic Shield which underlie the allochthon i.e. the boundary thrusts to the Bångfjället Complex pass down into the underlying autochthon. The Bångfjället Complex is not necessarily part of a flat-lying thrust sheet (Middle thrust rocks *in* Strand and Kulling 1972, p. 206) between the Seve-Köli Nappe Complex and the autochthon rooted at some unknown distance to the west.
5. The position and tectonic emplacement of the Bångfjället Complex is thought to be related to movement of the Upper Laxfjället Unit (Storfjäll Nappe?) over the underlying Seve-Köli sequence at the end of D2 or during early D3. Major and minor overturned folding (F3) beneath the Upper Laxfjället Unit, involving both the cover rocks and the underlying basement, appears to have been the initial response to this movement.
6. It is suggested that mylonitization along the limbs of the major basement antiform and subsequent thrusting are related to fluid movement and concentration along the basement-cover contact, particularly on the overturned limb of the antiform.
7. The change-over from a ductile (mylonite formation, folding) to cataclastic (thrusting) mode of deformation is thought to be related to the build up of high fluid pressures along the mylonite zones. This is a direct consequence of the increase in overburden pressure due to thrusting of the Upper Laxfjället Unit on top of the underlying Seve-Köli sequence.

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NGU = Norges Geologiske Undersøkelse
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