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SER C NR 689 AVHANDLINGAR OCH UPPSATSER ÅRSBOK 67 NR 7

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ABSTRACT

The area studied is a part of the Seve-Köli Nappe Complex. A short lithological description and a tectonic analysis are presented. In the Marsfjällen area three tectonic units, generally dipping gently to the west or north-west, were distinguished. These are from west to east:

1. *The phyllite belt* (Köli), comprising phyllites, conglomerates, carbonate-rich layers and meta-volcanites. These rocks have been subjected to four phases of deformation. The first, F_1, produced a slaty cleavage (S_1) and tight to isoclinal folds. A refolding on several scales took place during F_2 and a crenulation cleavage, at present the main cleavage, was formed in most of the phyllite belt. In some places S_1 was not crenulated but only rotated; here it is labelled $S\,1+2$ since it remained a slaty cleavage. Later phases, F_3 and F_4, locally produced folds. The phyllite belt is a part of the eastern limb of the Ransaren or Eastern Synform (Zachrisson 1969), which in this deformation scheme is an F_3 structure.

   The transition to the next unit, the Seve-Köli contact, is defined in this area by Glass (in preparation) as the break in metamorphic grade between garnet, albite-bearing rocks (Köli) and a retrograded andesine belt (Seve). It is interpreted as an F_2 thrust. An analysis of rotated syn-F_2 garnets, occurring close to this contact, brought about a sense of F_2 shear indicating relative downthrusting of the Köli rocks.

2. *The Svartsjöbäcken Schists and the Marsfjället Gneiss* (Seve). These rocks, with intercalated metabasites, bear the evidence of three deformation phases, of which the second and the third probably correspond to F_2 and F_3 in the phyllites. No F_1 folds could be demonstrated, but remnants of S_1 occur as parallel inclusions in syn-F_1 garnets and as a locally preserved schistosity in large metabasic bodies. The schistosity in the schists, locally of crenulation cleavage type, is interpreted as a second cleavage, S_2. It parallels the axial plane of tight F_2 folds. The schists recrystallized extensively during and after F_2. In the gneisses a schistosity, S_2, and F_2 folds are formed by an intensive, chiefly postcrystalline deformation, resulting in the formation of numerous mylonites or blasto-mylonites. The "transverse" mineral lineation, often mentioned in the literature on the Caledonides (e.g. Kvale 1953), is interpreted as parallel to the X direction of F_2 shear strain, which is close to the shear direction, *a*.

   Below an F_2 blasto-mylonite zone at the base of this unit a jump in metamorphism has been demonstrated by Glass (in prep.). This contact is interpreted as an F_2 thrust. Asymmetrical muscovite "fish" in the surrounding rocks indicate a sense of shear corresponding with overthrusting to the east.

3. *The eastern schist and amphibolite belt* (Seve), comprising schists, metabasites and a few orthogneisses. These rocks have undergone a deformation comparable to the one in the preceding unit, except that the schists are less recrystallized than the Svartsjöbäcken Schists during and after F_2. During F_2 three types of folds, defined on the basis of orientation and style, and mineral lineations in two directions were formed. The lower contact of this unit coincides with the tectonic contact of the metamorphic Seve nappe, lying on top of hardly metamorphosed Eocambrian quartzites and sparagmites. This contact is also interpreted as an F_2 thrust, since it resembles the other thrusts.

In the Eocambrian rocks, underlying the Seve sequence, two deformation phases are recognized, which are interpreted as equivalents of F_2 and F_3.

Ultramafic bodies, which occur abundantly in certain stratigraphic levels, intruded probably in a solid state before or during F_1.

The origin of the crenulation cleavage in the phyllite belt and possible mechanisms for the rotation of synkinematic garnets are discussed in a separate chapter.

CONTENTS

CHAPTER IV

CHAPTER V

CHAPTER VI

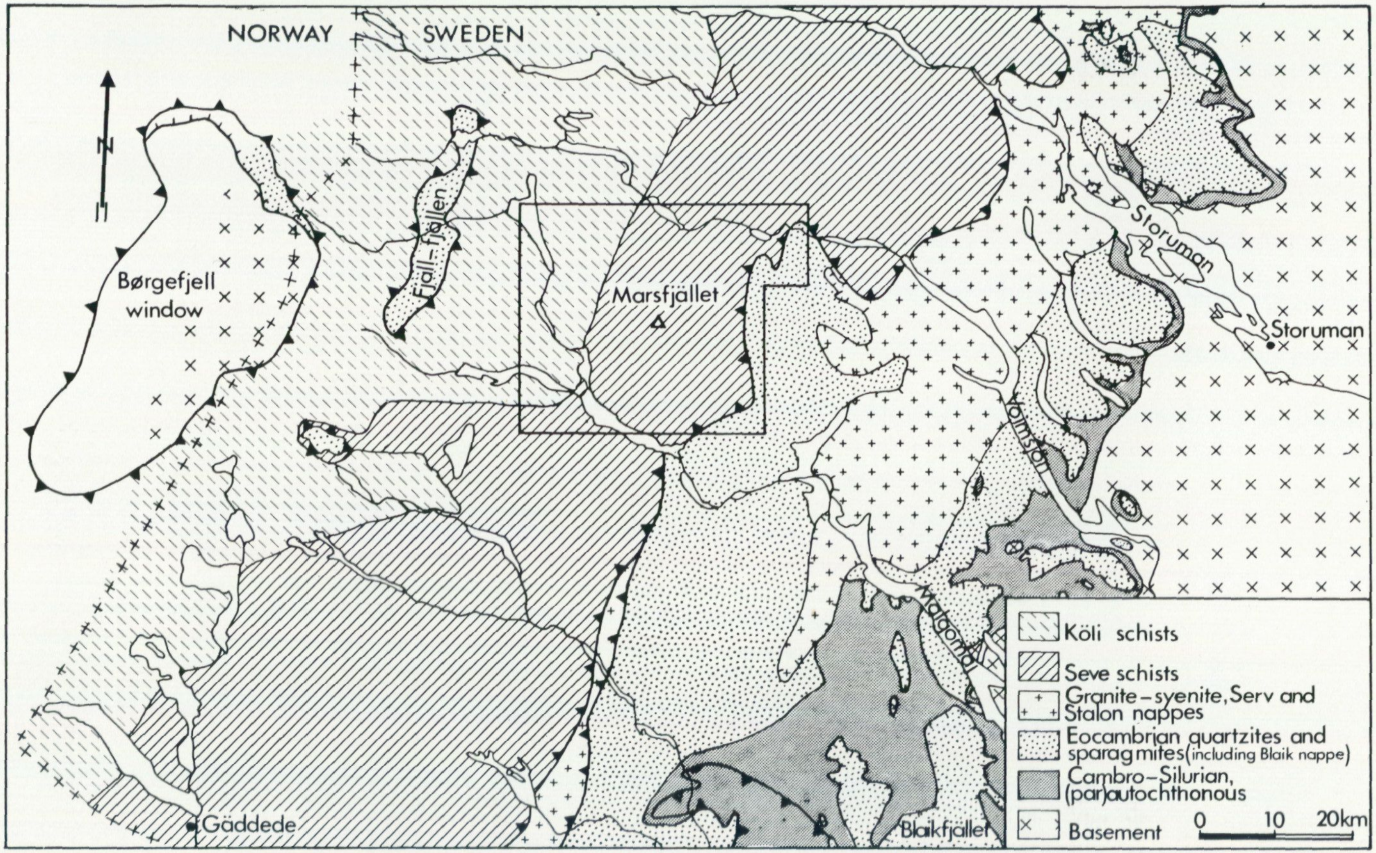

Fig. I-1. Simplified tectonic map of the southwestern part of Västerbotten, after "Karta över Sveriges Berggrund" (1958) with modifications after Zachrisson (1969). The area under consideration is indicated.

CHAPTER I

Introduction

REGIONAL GEOLOGY

The Marsfjällen area, named after Mount Marsfjällen, comprises part of the Caledonides in south-west Västerbotten, Sweden (Fig. I-1). The area is situated in the Seve-Köli Nappe Complex. The Seve sequence consists of gneisses, schists and amphibolites, of unknown age. It is overlain by the Köli sequence, low-grade meta-sediments, mainly phyllites and meta-volcanites of Lower Paleozoic age. The Seve-Köli Nappe Complex overrides the so-called sparagmite- or lower nappes, chiefly containing slightly or non-metamorphosed arkoses and quartzites. These lower nappes, composed of rocks of Eocambrian age, are themselves thrust over sedimentary rocks of Eocambrian and Cambro-Silurian age, which rest upon a basement of pre-Eocambrian igneous and metamorphic rocks (Strand 1961; Strand and Kulling 1972). In general, all the described rock units have slight or moderate dips to the west or north-west.

PREVIOUS WORK

Only the more important and more recent publications will be mentioned here. For a more complete review of previous work the reader is referred to Kulling (1955).

In 1929 Backlund and Quensel published a map on a 1:200,000 scale covering most of the Caledonides in Västerbotten. However, the description followed much later (Quensel 1960). On the detailed map, zones of tectonic importance, such as mylonite zones (hartschiefer), are indicated, but no tectonic interpretations were given. The importance of a quartzite conglomerate (the Voitja conglomerate), which could be mapped throughout the area, was already recognized. It was dated by Kulling as Upper Ordovician, based on fossils in the overlying limestone (Kulling 1933, pp. 167–422).

In 1942 Kulling published the results of extensive work in the Västerbotten mountain range. His map was later combined with that of Gavelin, on a 1:400,000 scale, in the "Berggrundskarta över Västerbottens län". It was published together with a description in 1955. Extensive use was made of Kulling's monograph during the present study. Later compilations by Kulling can be found in: Magnusson et al.: "Description to accompany the map of Pre-Quaternary rocks of Sweden" (1960) and in: "Scandinavian Caledonides" by Strand and Kulling (1972). In "Peridotites, serpentines and soapstones" (1935) Du Rietz not only dealt with ultramafics, but also presented a detailed map of the Gikasjön area on a 1:100,000 scale, which will be discussed in Chapter II. In "The injection metamorphism of the Muruhatten region" (1938) a comparison was made with the Borkafjäll gneisses, which are described in Chapter IV. In a publication in 1941 on the Remdalen ore bodies, Du Rietz included an investigation of the Fättjaure region, accompanied by a map. This map overlaps the present one in the Aunere region. The subject will be dealt with in Chapter II. Michel presented his Ph. D. thesis on "Geology and petrology of the Borkafjäll region" in 1950. His area is almost completely included in the present one. The detailed petrographical descriptions are often referred to.

De Keyzer and Janssen worked in the Fatmomakk and in the Marsfjällen area respectively. In 1952 and 1953 they presented private reports to the University of Amsterdam.

Zachrisson investigated the Köli rocks west of the present area. He presented a detailed stratigraphy which was very helpful during this study (Zachrisson 1964, 1969, 1971).

THE PRESENT STUDY

Under the supervision of Professor H. J. Zwart, a mapping project was started in 1968, both in the Marsfjällen area and in the Trondheim area, near Selbu. During the subsequent years the work was extended in these two areas. New areas, as, e.g., the Gäddede area, were also included. At the moment (1973) some twenty students are involved in this project. The aim of the work in Sweden is not only to prepare detailed maps of the as yet hardly differentiated Seve and Köli rocks, but also to establish the tectonic and metamorphic history of the Seve-Köli Nappe Complex.

The present investigation is based on fieldwork which was carried out during the summers of 1968–1971. Parts of the presented map were prepared by other workers: Glass, Calon, Brandt and Zwart. In this work the tectonics of the entire area are stressed, while Glass's thesis deals with the metamorphic assemblages of the same area.

General rock descriptions and a tectonic analysis are given, based on mesoscopic measurements and microscopic observations. Comments are added on the origin of certain phenomena, such as crenulation cleavage, rotated garnets (Chapter VIII) and asymmetric "fish-shaped" muscovites in mylonites (Chapter IV). No petrofabric analyses are given, as this thesis is meant to be a more general study, leaving detailed investigations to future workers. Petrofabrics of quartz, combined with a study on recrystallization and recovery features, will be published by Glass.

ROCK UNITS

Among Scandinavian workers (e.g. Kulling 1955; Zachrisson 1969) it is common usage to divide the Seve-Köli Nappe Complex into a low-grade Köli part and a high-grade Seve part. The contact between these is locally tectonic and may be accompanied by a jump in metamorphism from, e. g., upper greenschist facies to upper amphibolite facies. The metamorphic isograds are not exactly parallel to the tectonic contact, so that other jumps may occur as well. It is claimed by these workers that several characteristic rock types in the low-grade Köli sequence, such as graphitic phyllite and conglomerate, are not to be found in the higher-grade Seve.

Although this subdivision is obviously very useful in many places, it might lead to confusion in localities where the tectonic contact is not obvious. In these localities the metamorphic grade ought not to be used to divide the various rock units, and the boundary has to be placed "somewhere" below the lowermost characteristic Köli rocks. This problem had to be faced in the present area until, after detailed microscopic observation, Glass (in prep.) established a transition zone, marked by the following features:

1. There is a metamorphic jump from albite-bearing garnet mica schists (Köli) to staurolite-kyanite-bearing garnet mica schists with a basic oligoclase (An_{30}) (Seve).
2. The albite-bearing schists show all the characteristics of a prograde metamorphism whereas the staurolite-kyanite schists are retrograded.
3. The zone is marked by highly deformed but recrystallized "blasto-mylonites".

This zone is taken as the contact between the Köli and the Seve rocks. Apart from this, probably tectonic, contact, an important intra-Seve-thrust has been established (Glass, in prep.). The Seve-Köli rocks in the Marsfjällen area are hence subdivided into three tectonic units. The middle of these is again subdivided into a higher- and a lower-grade part. The tectonic evolution of the resulting four units was investigated separately. The units are from west to east (top to bottom):

1. *The phyllite belt.* This belt belongs to the Köli sequence and is described in Chapter II. It comprises phyllites, quartzites, conglomerates and metavolcanites, probably all of Lower Paleozoic age.
   Following Zachrisson (1969) the belt is subdivided into the Tjopasi Group and the Lasterfjäll Group. The transition to the next unit, the Seve-Köli transition, is described above.

2. The second unit is a formation, the *Svartsjöbäcken Schists* (Chapter III). It consist of schists and amphibolites. Towards the east the schists grade into gneisses.

3. The *Marsfjället Gneiss* (Chapter IV).
   Gneisses of this formation contain the characteristic kyanite-potash feldspar mineral pair. In the metabasic rocks the clinopyroxene-garnet-plagioclase assemblage is common. A mylonite zone separates the gneisses from the next unit. Across this boundary there is a jump in metamorphic grade (Glass, in prep.).

4. *The eastern schist and amphibolite belt* (Chapter V).
   This belt is not given the status of a group, and only a few formations are defined within it, the poor exposure making it impossible to fix clear boundaries. The schists, amphibolites and granite gneisses comprising this belt are metamorphosed under lower to middle amphibolite facies conditions (Glass, in prep.). They appear to grade downwards towards the east, where locally, e.g. south of the present area, an albite-bearing zone constitutes the lowest part of the Seve Nappe.

Chapter VI deals with ultramafic rock bodies. They occur abundantly in the lower-grade rocks.

NOMENCLATURE AND DEFINITIONS

In the petrography grain sizes are subdivided as follows:

5 mm–3 cm	coarse-grained
1 mm–5 mm	medium-grained
1/3 mm–1 mm	fine-grained
1/100 mm–1/3 mm	very fine-grained

The term layering indicates a compositional layering.

For the structures, descriptive terms are used as much as possible, but in the classification of the successive deformation phases an element of interpretation was unavoidable.

The various cleavages are subdivided into three groups (Knill 1960; Rickard 1961; and others).

slaty cleavage – planar fabric penetrative throughout all the rock material (Ramsay 1967, p. 177)

fracture cleavage – planar fabric that is not penetrative on all scales, but forms discrete planar discontinuities (Ramsay 1967; de Sitter 1956).

This term has a genetic implication, but the definition is used descriptively. Another term, such as "non-penetrative cleavage", or "differentiated cleavage" would be better, but the author has no intention of adding another term to the already confusing nomenclature of cleavages.

crenulation cleavage – cleavage that is not penetrative on all scales, formed either by micaceous layers or by sharp breaks. The cleavage planes are separated by thin slices of rock, which contain a crenulated cross-lamination (slaty cleavage) (Rickard 1961; see also Talbot 1964).

In coarser grained rocks (e.g. schist, gneiss, amphibolite) the term schistosity is used instead of cleavage.

All these types grade into one another and no sharp boundaries can be fixed. The two latter types of cleavage are in general accompanied by some sort of tectonic layering which is defined as any compositional layering, except bedding, which is parallel to a tectonic plane. In the description of folds the nomenclature of Ramsay (1967) and Fleuty (1964) is followed. The term *vergence* is used to indicate the direction opposite to the direction of dip of the axial plane of an asymmetrical fold.

After Ramsay (1967, p. 335), the capital letters X, Y, Z are used to indicate the principal finite strains ($X \geq Y \geq Z$), and the letters *a, b, c* are used as kinematic axes for simple shear, *a* is the direction of shear, *b* is perpendicular to *a* in the shear plane, and *c* is perpendicular to *a* and *b*.

The terms irrotational and rotational strain are defined as follows:
If the principal axes of strain before and after deformation are attached to the same material points throughout the strain, the strain is said to be irrotational. In all other strains the principal axes change position with respect to material points during deformation and the strain is known as rotational strain.

The deformation phases are labelled F_1, F_2, etc. in chronological succession; cleavage planes or axial planes and lineations are indicated by the symbols S and L respectively, δ stands for intersections of cleavage planes and bedding, f corresponds to fold axes. These symbols are given the same index as the deformation phase in which the structures corresponding to them are formed. Bedding is referred to as SS.

The orientation of planes is given by the direction of dip and the amount of dip. For instance, 310/15 is the orientation of a plane with a strike N40E and a dip of 15° to the NW. The orientation of lineations is given in the same manner by the direction of plunge and the amount of plunge.

The symbols S_i and S_e are used for internal and external S surfaces in and around porphyroblasts.

With respect to rotated garnets the following symbols are used:

Ω_i : rotation angle of porphyroblast measured between S_i in the centre and the general orientation of S_e.

Ω_e : rotation angle of porphyroblast measured between S_i in the centre and the plane of shear in a model of simple shear.

Ω_c : rotation angle measured between S_i in the centre and S_i in the rim of a crystal.

Ω_s : rotation angle of S_e with respect to the shear plane in a model of simple shear.

γ : amount of simple shear.

α : angle between direction of simple shear, and the longest axis of the strain ellipse or ellipsoid.

In the nomenclature of mylonites the classification given by Spry (1969) is followed (see also Lapworth 1885), except that the field covered by the term "hartschiefer" was equally divided between mylonites (or cataclasites) and blasto-mylonites. Spry defines mylonites as "consisting of strained porphyroclasts embedded in abundant (50 to 90 %) fine-grained to cryptocrystalline matrix; a foliated structure is typical and this distinguishes it from a cataclasite which has an unfoliated matrix". A protomylonite contains less matrix, an ultramylonite more (Spry 1969, p. 229). Since the term "hartschiefer" is not used in this paper, a minor recrystallization of the matrix has to be included in this definition and a blasto-mylonite is then defined as a mylonite with considerable recrystallization of the matrix and/or the porphyroclasts.

Representation of structural data

In order to avoid a tedious description of the orientation of structural data, diagrams were constructed and plotted in various subareas on a tectonic map (Enclosure II). Areas in which these orientations are more or less constant were chosen as subareas.

Contouring of the diagrams. The data are plotted on a Schmidt net (equal area projection) and have been contoured in the cases in which about 30 or more measurements were available. The separate measurements are plotted as well, to facilitate comparison of the various diagrams, except for diagrams with over 200 mesurements.

The diagrams are counted by means of a computer, using a modified version of the programme described by Möckel (1969) and Noble and Eberly (1964). The modification is carried out by Roberti (personal communication), because of the variable number of measurements and the non-statistical method of sampling. Roberti states that since contouring has been carried out with a variable counting circle which measures 100/N per cent of the total area of the diagram (here the hemisphere, which equals 1), the expectation (NA) always equals 1; where N=number of measurements and A=area of the counting circle. The frequency density (d=n/NA) assigned to each counting point can thus immediately be read from the number of measurements counted at that point (n). In order to avoid A becoming too small for large N, the counting circle has been adapted by enlarging the product NA to 2 for N >100. For N >200 NA is enlarged to 4; hence the expectation (NA) becomes 2 and 4 respectively for these diagrams.

In Enclosure II the values of N and A are given with the diagrams.

Diagrams of less than 100 measurements (expectation NA=1) are contoured at 1 % (dashed line), 3 %, 5 %, 7 %, 10 %, 15 %, 20 %, 25 % etc, unless otherwise indicated. If N >100 and NA is for instance 2, contouring was carried out at twice these percentages, for NA=4 at four times the percentages. In some diagrams, where the distributions are apparently concentrated on a girdle or on a partial girdle, this girdle is indicated by a dashed line.

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CHAPTER II

The Phyllite Belt

STRATIGRAPHY AND PETROGRAPHY

Zachrisson (1969) subdivided the Köli rocks into three groups: the Tjopasi Group, the Lasterfjäll Group and the Remdalen Group. The former two were found to be useful mapping units in this area; the latter does not occur. Within the Tjopasi and Lasterfjäll Groups several formations are discerned (Table II-1). The contacts between these formations are conformable.

The map shows a high degree of interpretation, the mapping being hampered by poor exposure and a strong resemblance of rock types between different formations.

Comments on thickness are in general refrained from on account of the large variation and the strong deformation, which is often accompanied by intense intraformational folding. The apparent thickness can in most places be deduced from the map and the sections. The deformation did not obliterate all sedimentary structures, such as, for instance, graded bedding, which was locally used as way-up evidence. The formations distinguished will be described from bottom to top.

The Fatmomakk Formation

The Fatmomakk Formation is a sequence of grey and black phyllites and fine-grained garnet mica schists. The garnet isograd runs obliquely through the formation. It can generally be subdivided into three members which will be described from bottom to top:

1. Fine to very fine-grained light grey phyllites or schists. Locally dark grey and greenish phyllites alternate with the light grey ones. Apart from a few quartzitic beds, the bedding is in general hard to recognize. In this member a greenschist, grading into amphibolite, occurs as well.
   The phyllites and schists contain a varying proportion of quartz (often more than 50 %) and white mica. Albite, chlorite and carbonate are very common, microcline may be found, and porphyroblasts of biotite occur in almost every thin section. Garnet and ilmenite porphyroblasts are also common, and in the darker phyllites sulphides may appear.
   The greenschist consists mainly of chlorite and albite, with a minor proportion of epidote and a pale green amphibole. It changes gradually into an amphibolite consisting of zoned crystals of pale green amphibole with blue green rims in a chlorite and biotite-bearing albite matrix.
2. Black, often graphitic phyllites.
   The very fine-grained graphite may constitute up to 10 % of these rocks. Garnet occurs locally.
3. Well-layered very fine to medium-grained, grey and greenish phyllites. Quartzitic phyllites are abundant in this member; the coarser varieties may grade to microconglomerates, with quartz and plagioclase pebbles. The mineral content is roughly equal to that of the lowermost member. A few thin quartz-feldspar layers and a thin marble horizon occur as well in this member. The marble contains epidote, albite, chlorite and rutile, besides carbonate. The quartz feldspar rocks are very fine-grained, making an accurate determination of the percentage of albite, quartz and perhaps microcline very difficult. Although no phenocrysts were recognized, these rocks are thought to be derived from acid igneous rocks, or "keratophyres". They contain small biotite porphyroblasts and some muscovite. A sandy greenish phyllite contains chlorite, carbonate and large biotite porphyroblasts (up

to 5 mm). Accessory minerals in the Fatmomakk Formation are epidote group minerals, tourmaline and apatite.

An extensive meta-keratophyre, also including metabasic rocks, is exposed north of lake Kultsjön. The position of this layer in the Fatmomakk Formation is uncertain; it is even possible that it may represent a higher part of the stratigraphy (above the Murfjället Formation), as believed by Biermann (pers. comm.) and Zachrisson (1969). In that case the abrubt end of this layer is probably caused by a tectonic contact, running along its northern boundary.

The Murfjället Formation

The Murfjället Formation consists of quartzite and quartzite conglomerate. Ultramafic bodies which are sometimes partly brecciated seem to be concentrated along this stratigraphic level, although they are not limited to it. They will be dealt with separately in Chapter VI.

The variation in thickness of the quartzite and the conglomerate clearly appears on the map. The conglomerate contains pebbles up to 40 cm in diameter, embedded in a quartzitic matrix which, at the transition to the phyllites, might be more pelitic.

The quartzite contains up to 20 % of albite and microcline; some white mica, biotite and chlorite may also occur. Biotite porphyroblasts are common; garnet was found in only one locality. Accessory minerals are epidote and apatite.

The Grundfors Formation

The Grundfors Formation is a sequence of greenish, grey and black phyllites, grey quartzites and quartz-feldspar rocks.

Near Grundfors the formation can be subdidived into three members, from bottom to top:

1. Well-layered black, grey and greenish phyllites. This member is very similar to the upper member of the Fatmomakk Formation.

   Medium-grained beds of 1 mm to several cm alternate with finer-grained beds. Graded bedding is a common phenomenon; many way-up determinations could be made on account of this (Fig. II-1).

   A few quartz-feldspar layers, up to half a metre thick, were found in Daunebäcken. They are composed of small quartz and albite grains and contain chlorite and locally biotite porphyroblasts. They are interpreted as meta-keratophyres (Michel 1950, p. 45 and 46, specimen 346). A phyllite with large clastic feldspars near Grundfors is rich in carbonate. Rounded albite phenocrysts with Karlsbad twins and partly recrystallized large quartz phenocrysts are embedded in a very fine-grained matrix, rich in quartz-feldspar. This rock is thought to be of pyroclastic origin.

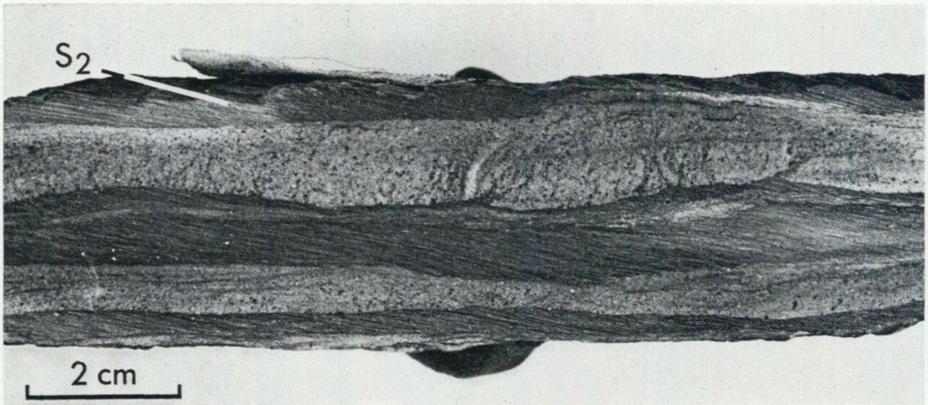


Fig. II-1. Graded bedding in phyllite.

2. Grey quartzite. This member has developed only locally.
3. Greenish to light grey phyllites. The green colour is a result of the chlorite content.

The Graipesvare Formation

A type section of the Graipesvare Formation is exposed in the small river north-east of the dam of lake Ransarn.

A repetition of sequences, ranging from fine to medium-grained light grey quartzitic beds up to very fine-grained black phyllites, is typical of the Graipesvare Formation. Bedding is easy to recognize on a mesoscopic scale. Apart from numerous thin beds of pure quartzite (up to 20 cm thick), a quartzite conglomerate has developed locally in the middle of the formation. This conglomerate has the same field aspects as those in the Murfjället and Ransarn Formations.

North-east of the dam of lake Ransarn the conglomerate is about 30 metres thick, in Vökarbäcken 15 m and in Daunebäcken 5 m; further north it is thicker again.

Some albite occurs in this conglomerate, both in the matrix and in the pebbles. The mica content of the matrix can be up to about 30 %.

The quartzites are composed of clastic quartz and feldspar grains, with minor proportions of microcline and albite. The proportion of feldspar can amount to 20 % of the rock. Porphyroblasts of biotite and ilmenite are still common in this formation.

The Ransarn Formation

The Ransarn Formation consists of quartzite conglomerate, quartzite and marble, or marble-bearing conglomerate. Locally all three occur together, but more often only one or two of these rock types have developed. The quartzite conglomerate contains pebbles, up to about 50 cm in diameter, of quartz, quartzite and gneiss. The matrix is quartzitic. The present thickness near the dam of lake Ransarn is 75 m, in the river south of Stuore Gämo it is about 50 m. On Stuore Gämo the conglomerate is gradational into a microconglomerate, with quartz pebbles in a more pelitic matrix. In the few exposures of the marble-bearing horizon almost pure marble pebbles and/or lenses with some quartz and graphite are, among other pebbles, embedded in a pelitic matrix. The pebbles form only a small percentage of the rock and are normally not in contact with each other (Fig. II-4). In Storbäcken a continuous marble layer, 1 m in thickness, is exposed. A number of crinoid stems were found in the marble.

The Stuore Gämo Formation

The Stuore Gämo Formation is a thick unit of light grey, calcareous and quartzitic phyllites.

Field characteristics are the light grey colour and the homogeneity of these fine to medium-grained phyllites. The weathering of the carbonate-rich layers is also typical (Fig. II-2). The thickness of the beds, varying from 1 cm to several metres, is generally larger than in the other formations.

The percentage of quartz can be up to about 70 %. Besides white mica, carbonate and albite, chlorite is present, both in the form of porphyroblasts and in the matrix. Biotite porphyroblasts have only developed locally (Glass, in prep.). Accessory minerals are: tourmaline, clinozoisite, apatite and iron ore.

Fig. II-2. F_1 fold in Stuore Gämo Formation. S_2 cuts through both limbs of the fold. Small river SW of Näuronjaure.

KULLING (1933,'55,'58) Björkvattnet – Virisen	ZACHRISSON (1964,'69) Southernmost Västerbotten – northern Jämtland		THE PRESENT PAPER Fatmomakk – Borga	DU RIETZ (1935) Gikasjön and Stuore Gämo
	Stekenjokk Quartz–Keratophyre and Lasterfjäll Quartz–Keratophyre bearing Formation	LASTERFJÄLL GROUP		
Viris Quartzite	Lasterfjäll Calcareous Phyllite		Stuore Gämo Formation	Calcareous phyllites, partly quartz–phyllitic
Lövfjäll Phyllite				
Broken Series	Br.S.		Ransarn Formation	Limestone, or phyllites with lenses of limestone and pebbles of quartzite
Slätdal Limestone	Bellovare Formation			
Voitja Conglomerate				Voitja Conglomerate
Gilliks Series	Graphitic phyllites with quartzitic, conglomeratic and limestone horizons	TJOPASI GROUP	Graipesvare Formation	Greywacke quartzite; Microcline quartzite
Seima Series	Volcanic rocks (basic and acid). Grey and graphitic phyllites		Grundfors Formation	Quartzitic phyllites or quartz phyllites
Ro Series; Ro Conglomerates	Serpentinite Conglomerate; Quartzite Conglomerate		Murfjället Formation	Grey phyllites
	Varied sedimentary, tuffitic and volcanic rocks		Fatmomakk Formation	Dark grey or black phyllites

Table II-1. Stratigraphic correlation of the formations in the phyllite belt.

Correlation with stratigraphic columns published previously

The formations described above in part show a strong similarity to stratigraphic columns from related areas. A summary of these stratigraphic columns, together with a tentative correlation, is given in Table II-1.

An obvious correlation is that of the Murfjället Formation to the Ro Conglomerate (Kulling 1958) and to the serpentinite and quartzite conglomerates (Zachrisson 1969). There is also a striking similarity of the Ransarn Formation to the Voitja Conglomerate, including the Slätdal Limestone (Kulling 1955, 1958) or the Bellovare Formation (Zachrisson 1969). Du Rietz (1935, 1941) also considered the rocks which are included here in the Ransarn Formation as Voitja Conglomerate and Slätdal Limestone. However, he considered other conglomerates, that the present writer believes to belong lower in the sequence, also as Voitja conglomerate. The Stuore Gämo Formation clearly corresponds to the Lasterfjäll Calcareous Phyllite (Zachrisson 1969) and the Lövfjäll Series (Kulling 1955). Based on these correlations the other formations can be tentalively correlated to the stratigraphic tables already published with the following differences attracting attention.

1. The greenschists and quartz-feldspar rocks, interpreted as meta-volcanites, from the lower member of the Fatmomakk Formation do not have their equivalent in Kulling's stratigraphy. Zachrisson mentions a similar sequence from the Raukasjö area, northern Jämtland (Zachrisson 1969, p. 12).

2. The volcanic rocks described in the Seima Series (Kulling 1955, 1958) and in the Tjopasi Group (Zachrisson 1964, 1969) only occur in the southern part of this area (Strand and Kulling 1972, p. 224). The thin quartz-feldspar layers in the lower member of the Grundfors Formation are probably the only representatives of these units in the middle and northern part of the area. The almost complete lack of meta-volcanites is thought to have simple depositional reasons, although the wedging out could be influenced by tectonics.
3. The sequence comprising the variegated phyllites, quartzites and the conglomerate of the two upper members of the Graipesvare Formation and of the Grundfors Formation can be correlated with the sequence comprising the graphitic phyllites containing quartzitic, conglomeratic and limestone horizons (Zachrisson 1969), although no limestone horizons occur in this area. The arkoses and polymict conglomerates of the Gilliks Series (Kulling 1955) are quite different (see also Strand and Kulling 1972, p. 224 and Zachrisson 1969, p. 12).
4. The Broken Series (Kulling 1955) has apparently not developed in this area (Zachrisson 1969, p. 13).
5. The microcline quartzite, mapped by Du Rietz (1941) north of Aunere, is not distinguished during this investigation.

Age

Apart from a few crinoid stems in the limestones of the Ransarn Formation, no fossils have been found. According to Kulling (1955), the contact of the Broken Series and the Slätdal Limestone coincides with the boundary between Ordovician and Silurian. If this correlation is correct, the formations of the phyllite belt are probably of Cambro(?)-Ordovician age, except for the Stuore Gämo Formation, which is then Silurian.

Sedimentary environment

The metamorphic rocks described are generally thought to be derived from marine geosynclinal volcano-sedimentary rocks (Kulling 1955, 1958; Du Rietz 1935, 1941; Zachrisson 1969; Michel 1950). The rhythmic sequences of coarser and finer material in the phyllites are characteristic for this type of sediments. The composition is that of an immature sediment; even the quartzites and quartzite conglomerates contain a fair amount of clastic feldspar. This indicates rapid sedimentation, relatively close to the source area. The thin intercalated keratophyres possibly represent tuffs (Zachrisson 1969, p. 8). The greenschists and the thick keratophyres in the south-west of the area could possibly have been lavas and a coarse-grained keratophyre could have been a dike or sill.

The origin of the ultramafic breccia or "conglomerate" will be discussed in Chapter VI.

STRUCTURES

Introduction

On the regional map by Zachrisson (1969) (Fig. II-11) one can see that the phyllite belt is situated in the eastern limb of the Ransaren or Eastern Synform. In Zachrisson's nomenclature this synform is an F_2 structure, since the most conspicuous cleavage plane, which he refers to as S_1, is folded. The present study has revealed that the structure is in fact more complicated. The main cleavage generally appeared to be a crenulation cleavage, product of a second phase of deformation. Therefore the Ransaren Synform is here referred to as a third phase structure. In the eastern limb of this F_3 synform an F_2 antiform occurs, with the crenulation cleavage, S_2, as axial plane cleavage. Furthermore, it is probable that an F_1 synform is folded by this F_2 structure. All in all, the products of four tectonic phases can be distinguished: F_1 produced folds and a slaty cleavage, S_1. During F_2 folds were formed with the most conspicuous regional cleavage plane, S_2, as axial plane cleavage. S_2 is predominantly a crenulation cleavage; it may be accompanied locally by a mineral lineation, $L_{2\,min}$. S_2 is folded in F_3 folds, as, for instance, in the Ransaren Synform. F_4 structures are of the same type as F_3 ones, folding S_2, but with a divergent orientation. Locally they are superimposed upon F_3 folds. These four phases, F_1–F_4, correspond roughly to the four phases distinguished by Zachrisson (1969, 1971), pre-F_1, F_1, F_2 and F_3 respectively.

F_1

Folds. About a dozen folds with the S_2 crenulation cleavage cutting straight through both limbs were distinguished in the field (Fig. II-2). They are apparently older than F_2. The crenulated slaty cleavage (S_1) in the hinges of these folds forms an angle with the bedding of up to 90° (Fig. II-3). Although highly deformed, S_1 appears to be subparallel to the axial planes of these early folds, which are therefore interpreted as F_1 folds. The mapping demonstrated a repetition of certain units, probably due to tight folds, which must predate F_2 as well, since S_2 again runs obliquely through them. Although the relationship between SS and S_1 in the hinges of these large-scale structures is not clear, they are also interpreted as F_1 folds.

Cleavage. The slaty cleavage, S_1, is in many thin sections parallel or subparallel to SS. Although this might partly be due to later deformation, it is thought to illustrate the tightness of F_1 folding. In the field S_1 is rarely to be found as a measurable plane, distinct from SS. Exceptions are west of Murfjället and in the small river SE of Grundfors, where F_2 deformation was relatively weak.

The orientation of F_1 structures can be read from the diagrams of sub-area 6 (Enclosure II). The girdle of SS planes in this area apparently represents F_1 folding, since its axis (230/45) coincides with δ_1 (235/35) and with the F_1 fold

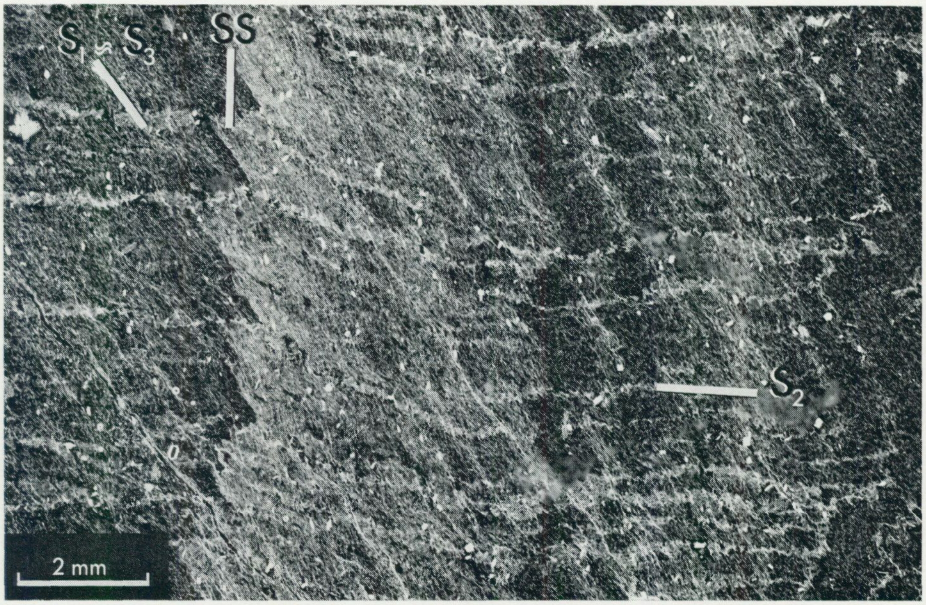


Fig. II-3. Angle between SS and S_1 of approximately 35°. S_1 is crenulated by F_2 and F_3. S_1 is perpendicular to SS; S_3 is parallel to S_1. Hill west of Murfjället (negative print).

axes. The orientation of the maximum of S_1 measurements is close to that of S_2. Their intersections, L_2, have a fairly steep maximum, 305/50, which direction reflects the original dip of S_1 in the direction of L_2. The lack of more data on F_1 structures from other areas makes a reconstruction of their original orientation too uncertain to carry out. The fact that the orientations of S_1 and S_2 are locally close together makes it possible that part of the folds with S_2 as axial plane are in fact further deformed F_1 folds.

In a few localities the longest dimensions of the deformed pebbles are probably situated in S_1, at a clear angle to S_2 (Fig. II-4). Here the strongest deformation of the conglomerate apparently took place during F_1. This observation is supported by Calon (pers. comm.), who has described a large-scale tight F_1 fold in the Ransarn Formation (Voitja conglomerate), on Daunentjakke, just north of the present area. Here the plane containing the longest dimensions of the highly deformed pebbles is folded in F_2 folds. Unfortunately no outcrops were found where suitable measurements could be carried out on deformed pebbles. To give a rough idea of the minimum deformation of F_1+F_2, where F_1 is believed to be dominant, deformed pebbles were measured in two loose blocks. The blocks are situated near the dam of lake Ransarn and are believed to derive from the nearby Ransarn Formation (Fig. II-5). On surfaces A, B and C the longest and shortest dimensions of pebbles are measured. In one block a specific elongation direction of pebbles was visible on surface C. The average

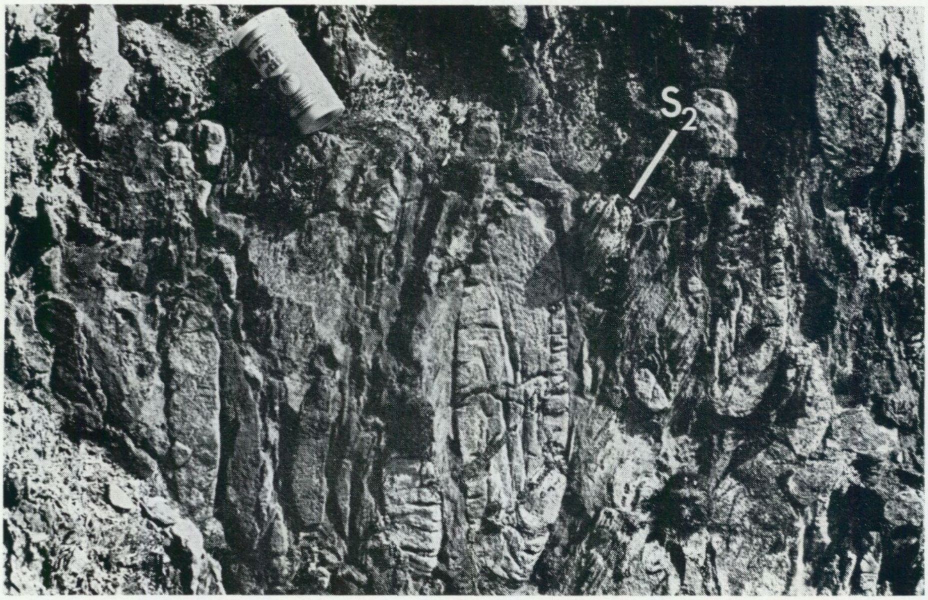


Fig. II-4. Flattened pebbles make an angle of about 35° with S_2. Upper member of Ransarn Formation, Ransarån.

of the axial ratios of 100 measured pebbles in one block is 1:2.8:2.9. In the other block this average is 1:2.5:4.8.

It may be concluded from the foregoing that F_1 is a phase of strong deformation, responsible for deformed pebbles and large-scale tight to isoclinal folding, with a penetrative slaty cleavage as axial plane cleavage. Although the writer is aware of the fact that later deformation (F_2, F_3) certainly modified F_1 folds, in general resulting in "tightening", the conclusion seems nevertheless justified, chiefly based as it is on the relationship between S_1 and SS.

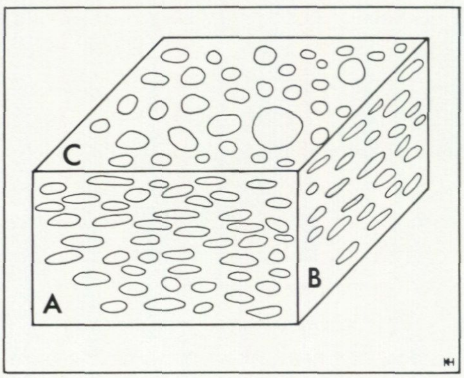


Fig. II-5. Deformed pebbles in a loose block, derived from the Ransarn Formation.

F_2

Folds. Many meso-scale F_2 folds, with S_2 as axial plane, can be observed in the field. They are mainly close or open asymmetrical folds, which by their attitude indicate a large-scale F_2 antiform in the western part of the area, near Näuronjaure and Tjaktsase (Fig. II-6, Enclosure III). Here, especially in the Stuore Gämo Formation, F_2 folds can be seen in almost every outcrop. The asymmetric meso- and micro-scale folds are interpreted as being parasitic folds on the limbs of the large-scale structure. The folds belong to Ramsay's classes 1C, 2 and 3 (Fig. II-7). Towards the east, in the steep limb of the large antiform, they are less abundant and close to the eastern boundary of the phyllite belt, they are very rare. F_2 fold axes are generally parallel to δ_2; they plunge gently to the NNE (Enclosure II).

In most of the sub-areas (Enclosure II) the poles to SS planes are concentrated in a girdle with a maximum and a sub-maximum, representing the long and the short limbs of asymmetric F_2 folds. The angle between maximum and sub-maximum varies between 45° and 120°. This illustrates that most of the folds are close or open. In all the sub-areas there is a maximum of steep SS planes and a sub-maximum of a more horizontal position, but in sub-area 4 the situation is the reverse, because this is the only sub-area in the flat-lying western limb of the big F_2 antiform. Some diagrams show a considerable amount of variation from the expected girdle. This is due to the presence of large ultramafic bodies, which by their contrasting competency caused deviations in the orientation of S planes around them.

Cleavage. Microscopic analysis demonstrated that S_2 is a crenulation cleavage in the western and middle part of the phyllite belt (Fig. II-8). Several variations of this structure are described in the discussion of crenulation cleavage (Chapter VIII).

In the more psammitic beds the cleavage is sometimes of the fracture

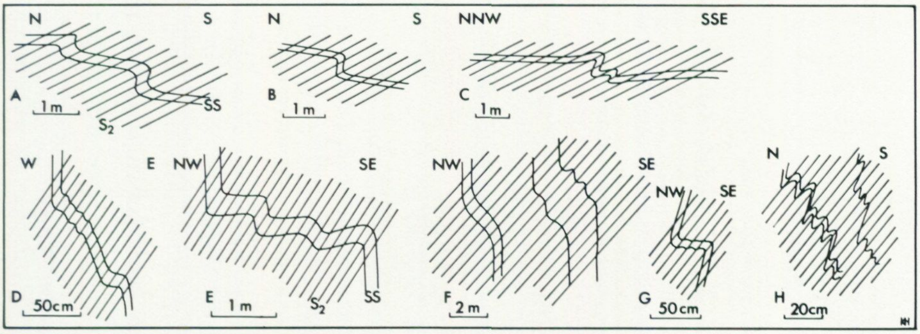


Fig. II-6. Asymmetrical F_2 folds. A, B and C from river SW of Näuronjaure and from western limb of large scale F_2 antiform. D to G from Graipesvare. H from Annasteinenjukke. D to H from eastern limb of F_2 antiform (see also Fig. II-11 and Enclosure III).

Fig. II-7. F_2 folds, closely approximating Ramsay's class 2, in the Grundfors Formation, Storbäcken.

cleavage type and along the eastern border of the belt only a slaty cleavage has generally developed. This cleavage is called S_{1+2} which will be explained in the discussion (p. 33).

The diagrams of S_2 planes show well-defined maxima. Only in the large-scale F_3 structure (sub-areas 9 and 10) has a partial girdle developed. Other deviations are probably due to the ultramafic bodies. The fact that S_2 planes change from a fairly flat-lying position in the west to a steep position in the east is probably due to F_3.

Lineations: Three types of lineations related to F_2 are distinguished: intersection lineations $S_2 \times S_1$ and $S_2 \times SS$, L_2 and δ_2 respectively, and a mineral lineation L_{2min}, defined by an elongated fabric habit of mica and quartz crystals.

L_2 could only be distinguished from δ_2 where S_1 forms a clear angle with SS. In sub-area 6 L_2 plunges WNW (305/50).

δ_2 was both measured in the field and calculated from measurements of SS and S_2. δ_2 is chiefly parallel to F_2 fold axes, its maximum plunges slightly to the NNE. The spread of δ_2 within S_2 is explained by the earlier folding of SS (F_1). If an S_2 plane cuts an F_1 fold, δ_2 bends in S_2 (Fig. II-9). The fact that in most sub-areas δ_2 produces one elongated maximum is considered to be a con-

Fig. II-8. F_2 fold with S_2 crenulation cleavage as axial plane cleavage. Graipesvare Formation, Annasteinenjukke (negative print).

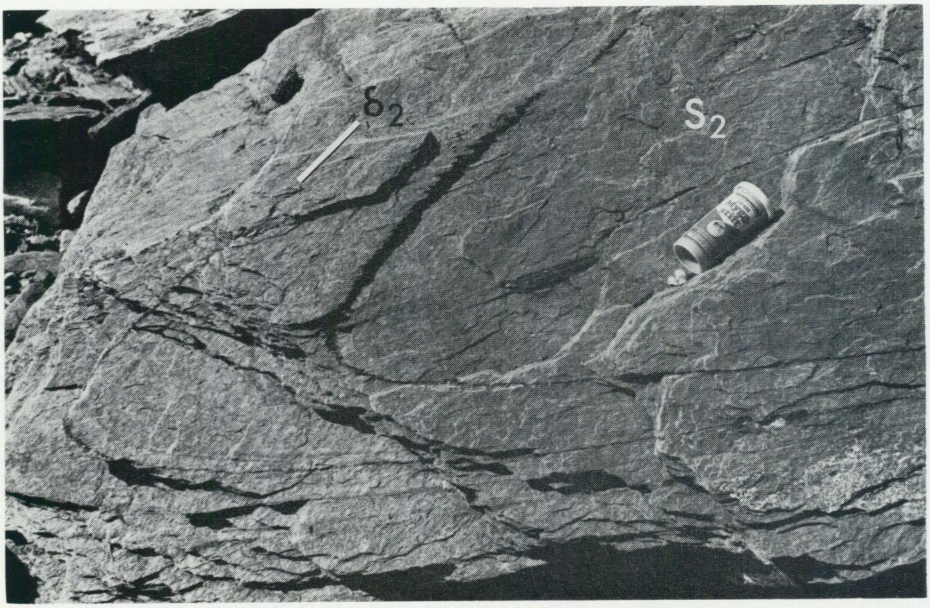


Fig. II-9. Section through F_1 fold, seen on an S_2 plane; outline of fold is parallel to δ_2. Stuore Gämo Formation, small river NW of Gikasjön.

firmation that F_1 folding was tight to isoclinal. In the combined sub-areas 3, 5 and 6 $L_{2\,min}$ has an elongated maximum towards the NW. This direction is about equal to that in sub-areas 13 and 14; it is fairly constant over the entire area.

F_3

As stated before, F_3 is among other things responsible for the formation of the Ransaren Synform. This is a composite structure, with two hinges (Fig. II-11). In one of them, which appears on the map NW of Stornäs, S_2 changes from a steep west-dipping position to a north-dipping one. In this hinge zone numerous micro- and meso-scale F_3 folds have developed. They are open, close or tight folds, locally of the kink- or accordion fold-type (Fig. II-10; Ramsay 1967; de Sitter 1956). Their orientation is related to the major structure; they have steep south-west dipping axial planes and fold axes with a moderate plunge to the NW. The other hinge of the Ransaren Synform is situated NW from Klimpfjäll (Fig. II-11), where S_2 curves around from a N-dipping to a N-S striking, almost vertical position. Here the axial plane, S_3, dips NW and the fold axis plunges NE. Apart from the hinge zone near Stornäs, F_3 folds are not abundant in the phyllite belt. Their orientation is rather variable. The measurements of their axial planes and of their axes are therefore in the diagrams of Enclosure II combined with those of F_4 folds, since it was often impossible

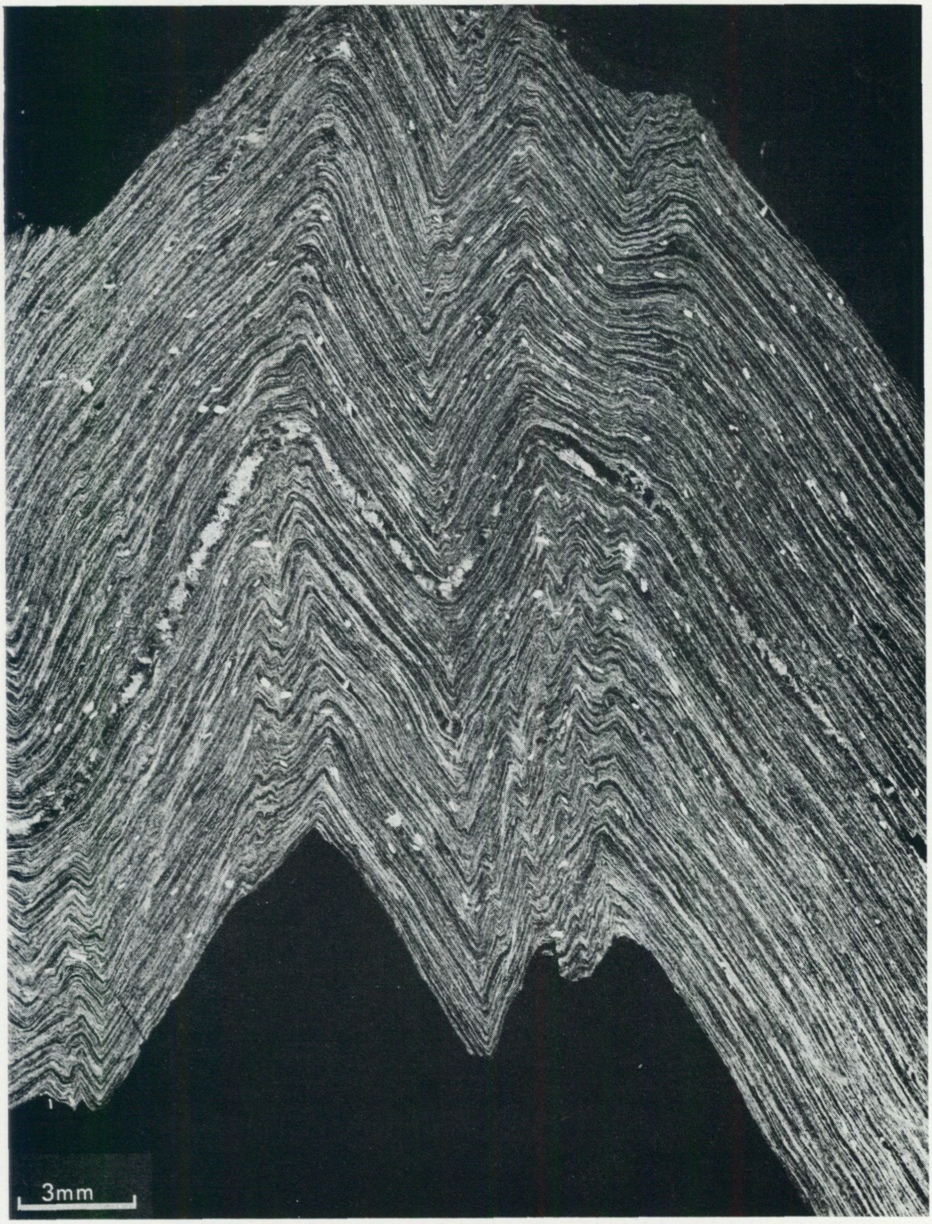


Fig. II-10. F_3 kinkfolds. S_2 is accompanied by a tectonic layering; it curves around small biotite porphyroblasts. Grundfors Formation, Graipesvare (negative print).

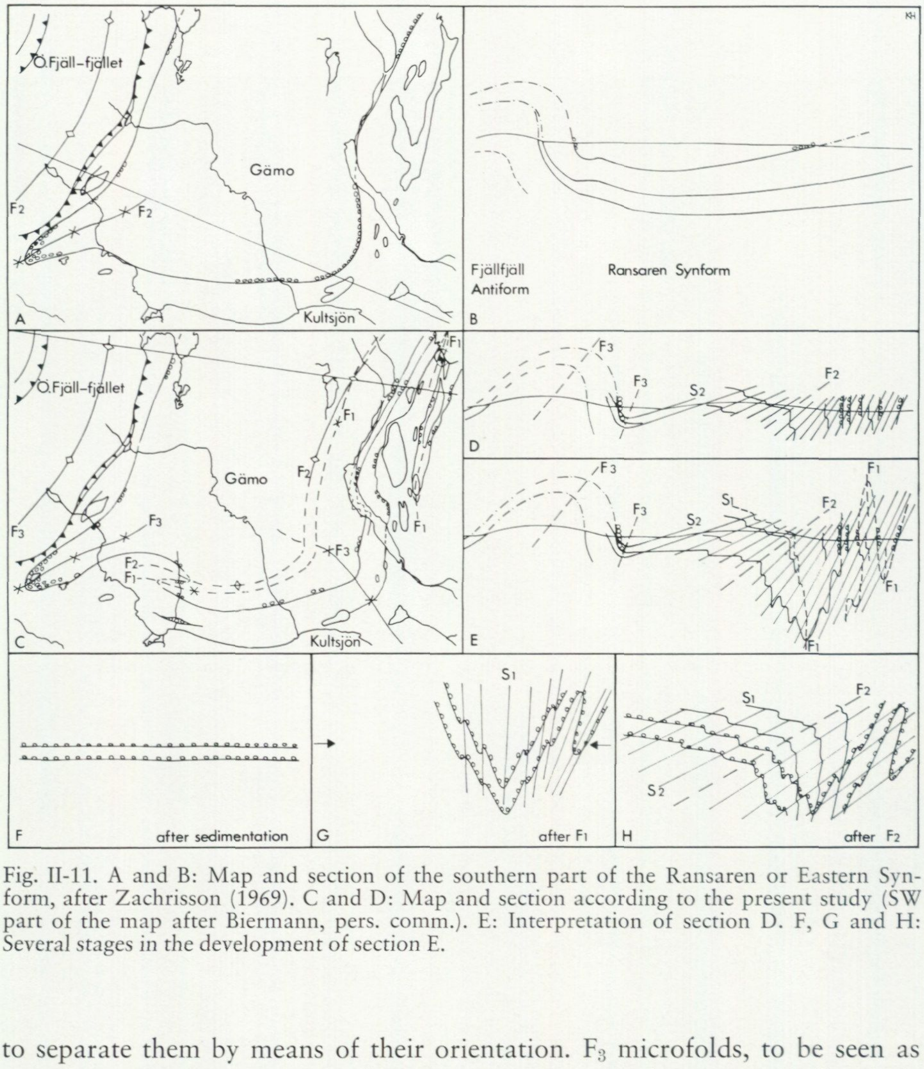


Fig. II-11. A and B: Map and section of the southern part of the Ransaren or Eastern Synform, after Zachrisson (1969). C and D: Map and section according to the present study (SW part of the map after Biermann, pers. comm.). E: Interpretation of section D. F, G and H: Several stages in the development of section E.

to separate them by means of their orientation. F_3 microfolds, to be seen as crenulation lineations on S_2 planes, are common in the entire phyllite belt. They are labelled L_3. In the diagrams they are combined with F_3 fold axes since there is no reason to separate meso- from micro-scale folds.

Along the axial plane of F_3 folds, a new crenulation cleavage has only rarely formed.

The general E–W shortening during F_3, discussed by Zachrisson (1969) is probably the cause of the steep position of S_2 in the eastern part of the phyllite belt. It makes sense that the incompetent phyllites, close to the contact with the more rigid Seve schists, were highly compressed. The same effect can be observed around the rigid ultramafic body of Aunere, where F_3 folds are abundant and very tight.

F_4

In some places F_3 folds or lineations are refolded by a later phase, which is labelled F_4. In general this relationship of superimposition upon F_3 structures is not found. In such cases the F_4 structures can only be distinguished from F_3 phenomena by their diverging orientation. In sub-area 9 (Enclosure II) a subsidiary maximum, belonging to sub-horizontal, slightly north-dipping S planes, represents axial planes of F_4 folds, S_4. This maximum of sub-horizontal S planes exists as well in the diagram of the combined areas 3, 4, 5 and 6. In the diagrams of L_3 and L_4, the maxima of lineations with a slight plunge to the NNE represent F_4 crenulation lineations and fold axes (e.g. sub-area 3).

One of the characteristics of F_3 and F_4 is that the quartz crystals in the fold hinges always have a strong undulose extinction.

METAMORPHISM

This subject will be dealt with in a somewhat stepmotherly fashion to avoid duplication of the work of Glass (in prep.). Only the relations in time between the grain growth of several metamorphic minerals and the deformation phases will be discussed. Ilmenite occurs in the form of elongated, lath-shaped crystals throughout the entire phyllite belt. The crystals are either oriented along S_1 or overprinting S_1 in a random way (Fig. II-12). S_2 curves around them; there-

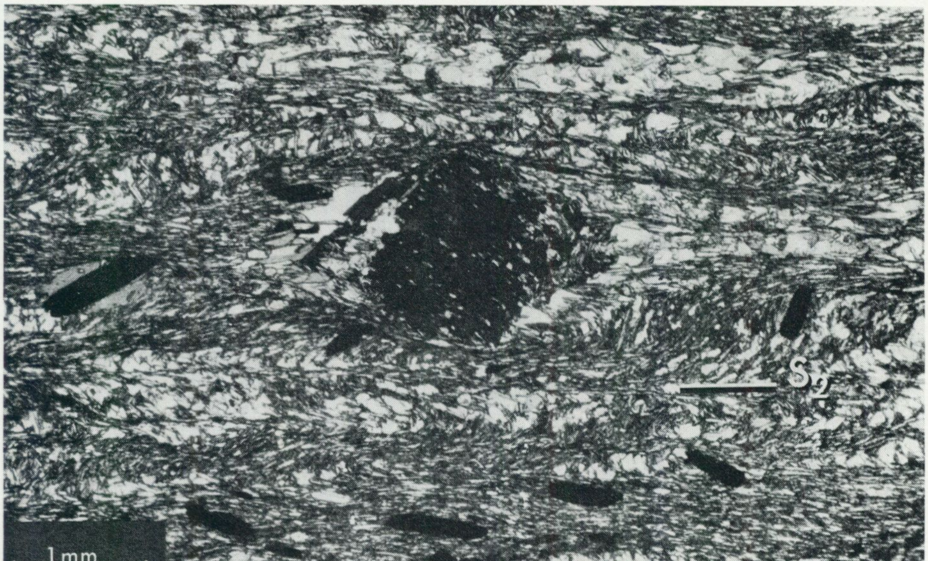


Fig. II-12. Interkinematic biotite porphyroblast with planar S_i fabric, interpreted as S_1. The surrounding crenulation cleavage, S_2, curves around the crystal. Ilmenite crystals are oriented parallel to S_1.

fore F_2 must be postcrystal-growth. The ilmenite crystals are interpreted as late syn-F_1 or interkinematic between F_1 and F_2. Chlorite can be found as a constituent of the matrix, as porphyroblasts or as replacing mineral of biotite and garnet. As a matrix mineral it might be as old as F_1, as the crystals are aligned along S_1. The porphyroblasts are later; some even overgrow F_3 folds and are parallel to S_3, proving that the metamorphic circumstances during F_3 still enabled chlorite to grow.

White mica only occurs as a constituent of the matrix. The very fine-grained crystals are parallel to S_1 or, for mechanical reasons, to S_2. The crystals apparently grew syn-F_1.

The plagioclase, occurring in the phyllite belt, is albite. It is never porphyroblastic.

East of the biotite isograd (Glass, in prep.), porphyroblasts of this mineral are very common. They sometimes contain a planar S_i, interpreted as S_1. The S_2 planes are curved around the more or less randomly oriented crystals (Fig. II-12). They are obviously interkinematic. In a few thin sections earlier biotites are observed, parallel to S_1 and therefore syn-F_1.

Many garnet porphyroblasts have S-shaped inclusion patterns (Figs. II-15, 16B). They grew syn-F_2, as will be shown in the discussion of rotated garnets. Other garnets with straight inclusion patterns might be older, interkinematic ones. Some garnets have late- or post-F_2 rims.

DISCUSSION

The Ransaren Synform and the large-scale F_2 antiform

In Zachrisson's publication of 1969 the outline of the Ransaren synform is indicated both on the map and in a section (Fig. II-11A and B). During the present investigation the situation was found as illustrated in Fig. II-11C and D. The F_2 antiform occurs in the limb of the Ransaren Synform, and according to the S_2-SS relationships the Lasterfjäll Group phyllites underlie the Tjopasi Group, which does not fit in the stratigraphy. It seemed impossible to link the structure to the Ransaren Synform and to understand the positions of the Lasterfjäll and Tjopasi Groups, until the importance of F_1 was recognized. The sketch section in Fig. II-11E shows the interpretation of the structure. To make it more clear, several stages in the development are outlined as well (F, G, H). The big F_1 folds cannot be directly observed, they are pure interpretation. The fact that all F_1 folds found are located near or in the F_1 hinge zones, as indicated on the sketch, supports this interpretation. Other supporting evidence is provided by Biermann (pers. comm.), who mapped the area west of this one. He reported a repetition of the Ransarn Formation (Slätdal Limestone) apparently as the result of pre-F_3 folding (Fig. II-11C). These tight folds, which are also marked on Kulling's map (1955), could very well be the continuation of the F_2 and F_1 structures in the present area.

The slaty cleavage in the eastern part of the phyllite belt

In the eastern part of the phyllite belt the regional cleavage is a slaty cleavage. This can be explained in several ways:

1. The slaty cleavage is S_1; F_2 deformation was absent or too weak to produce a cleavage.
2. The S_1 slaty cleavage was deformed during F_2, but not folded. This can occur if the angle between S_1 and the XY plane of the incremental strain ellipsoid of F_2 does not exceed 45° (Ramsay 1967, p. 92). The resulting slaty cleavage should in such a case be called S_{1+2}.
3. A crenulation cleavage (S_2) was formed, but, as a result of strong progressive deformation and possibly recrystallization, was transposed into a new penetrative slaty cleavage.
4. F_1 deformation was absent and the slaty cleavage is S_2.

Fortunately it could be demonstrated in the areas with crenulation cleavage that biotite porphyroblasts grew interkinematically between F_1 and F_2 (Fig. II-12). The slaty cleavage under discussion is considerably deflected around biotite porphyroblasts (Fig. II-13). Since it is very unlikely that only these specific biotite porphyroblasts, within a transition zone of approx. 100 m, should be pre-F_1, the deflection must have been caused by F_2 and possibility (1) can

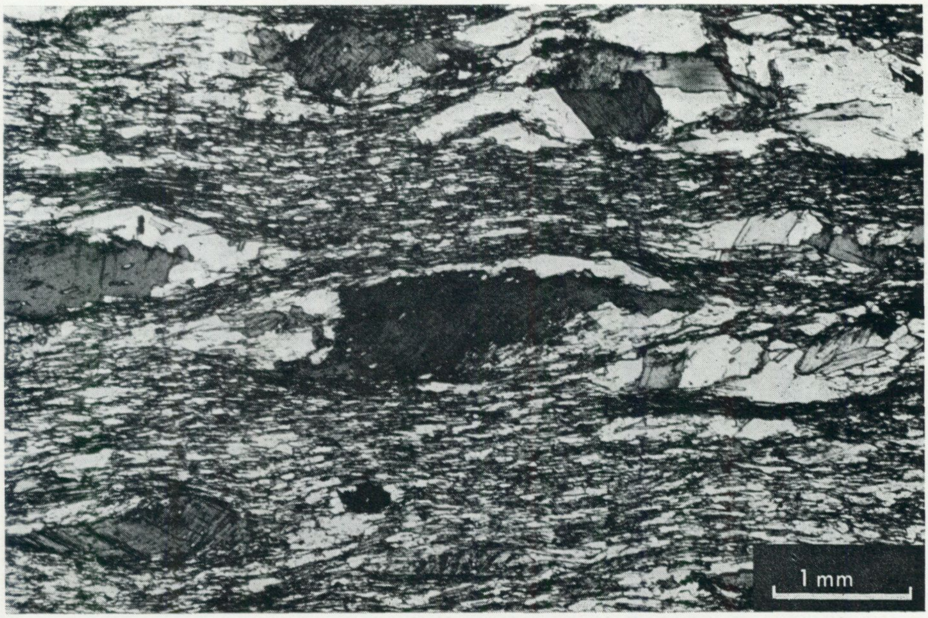


Fig. II-13. Slaty cleavage in the eastern part of the phyllite belt, curving around biotite porphyroblasts with straight or slightly S-shaped inclusions. S_i is continuous with S_e.

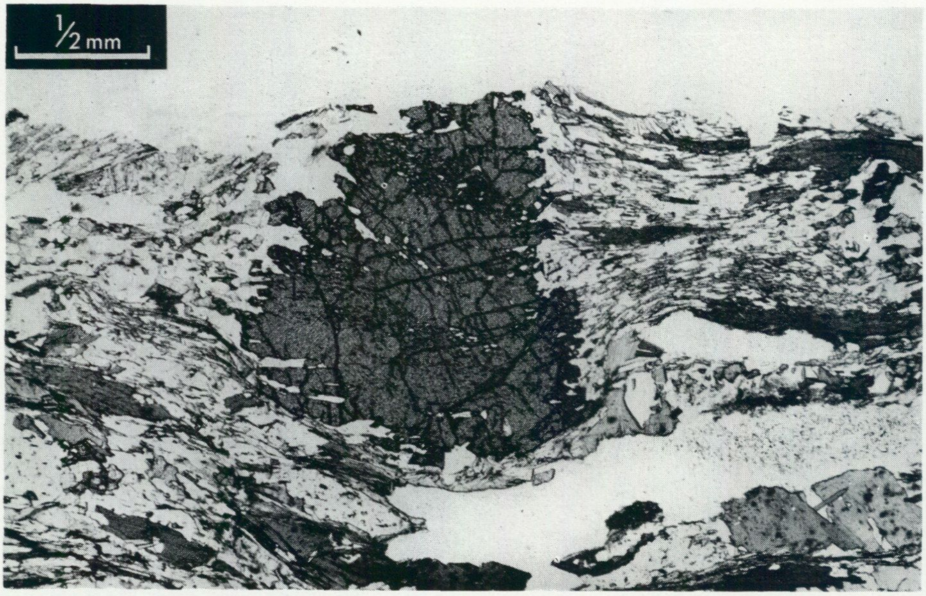


Fig. II-14. Syn-F_2 garnet with continuous S_i–S_e. The slaty cleavage is S_{1+2}. Fatmomakk Formation, near Stornäs.

be rejected. If no biotite porphyroblasts are present the age of the slaty cleavage is more difficult to determine and, especially since in a number of locations only a weak F_2 crenulation cleavage was found, possibility (1) is not everywhere to be excluded. According to the literature, possibility (3) seems rare, but in Rioumajou (Pyrenees; Zwart, pers. comm.) clearcut evidence was found that such a process may happen. Here the history of the new slaty cleavage is documented by helicitic folds in staurolite crystals, which overgrew a crenulation cleavage before it was transposed into the new slaty cleavage. In the present area, at least in a few cases, the transition from S_i to S_e in and around interkinematic biotite porphyroblasts demonstrates by its continuity that a crenulation cleavage was never formed in the vicinity of the biotite and later obliterated (Fig. II-14). Possibility (4) does not seem very likely because of the intensity of the F_1 deformation elsewhere. It can be concluded that possibility (2) is the most probable, and that the slaty cleavage in the eastern part of the phyllite belt is S_{1+2}.

From the sections in Fig. II-11F, G and H it appears that large scale F_1 structures were probably present before F_2. It makes sense that the eastern limb of the eastern F_1 synform took up a position in the F_2 stress field different from the rest of the folded sequence, which resulted only in a rotation of the slaty cleavage in this limb, whereas it was crenulated in the rest of the area.

Rotated garnets

Introduction. A number of garnets in the schists from the Fatmomakk Formation have S-shaped inclusion patterns, which indicate a rotation relative to the surrounding cleavage plane during growth (Figs. II-15, 16B). This cleavage plane is a slaty cleavage (S_{1+2}) and the S_i is in general continuous with S_e (Fig. II-14). In some cases the rotation apparently continued after growth, producing an angle but no disruption between the outcoming S_i and S_e (Fig. II-15). The size of the inclusions is comparable to that of the matrix crystals. In conformity with the interkinematic biotites, the garnets are interpreted as syn-F_2. Some have late- or post-F_2 rims.

Mesurement of rotation angles. Powell and Treagus (1967, 1970) demonstrated, with the aid of a three-dimensional model, which shapes of inclusion patterns are to be expected in more or less round "rotated" crystals.

In their model the assumption is made that the rotation axis is a straight line, which, as pointed out by Wilson (1971), need not be the case. In this investigation an attempt was nevertheless made to locate straight rotation axes with the aid of the (| |), ( ), | |, or)(patterns, as predicted by Powell and Treagus, for sections parallel to the rotation axis (Figs. II-16A; III-8; V-6). Rosenfeld's (1970) method was followed of searching for the orientation of the rotation axes by preparing thin sections parallel to the schistosity.

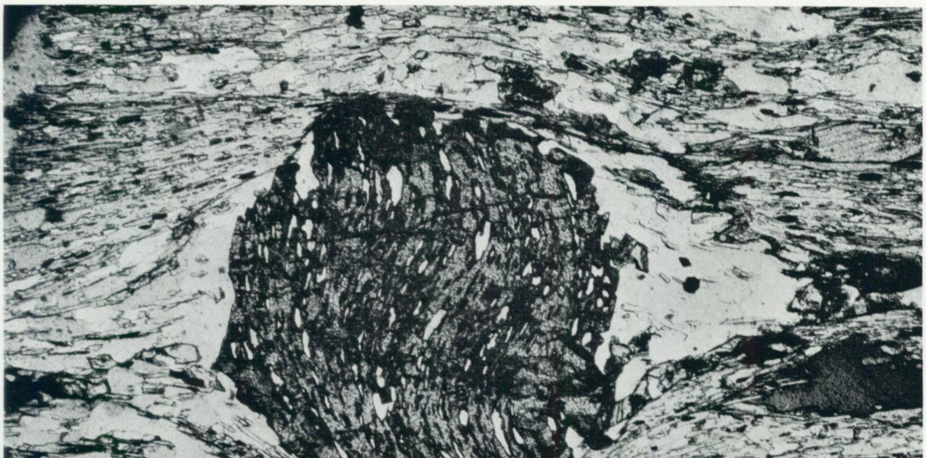
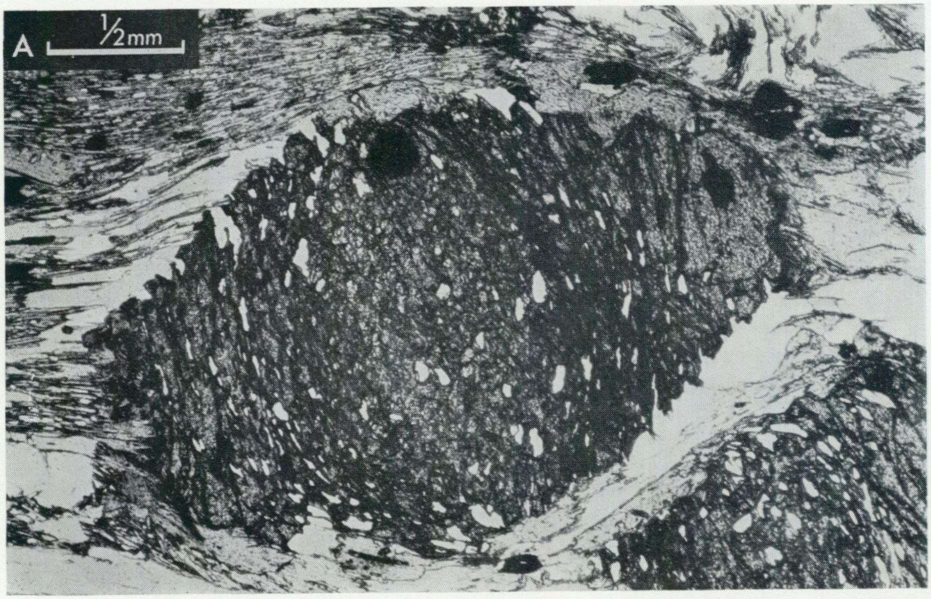


Fig. II-15. Syn-F_2 garnet, as in Fig. II-14, but S_i makes an angle with S_e of about 45°.



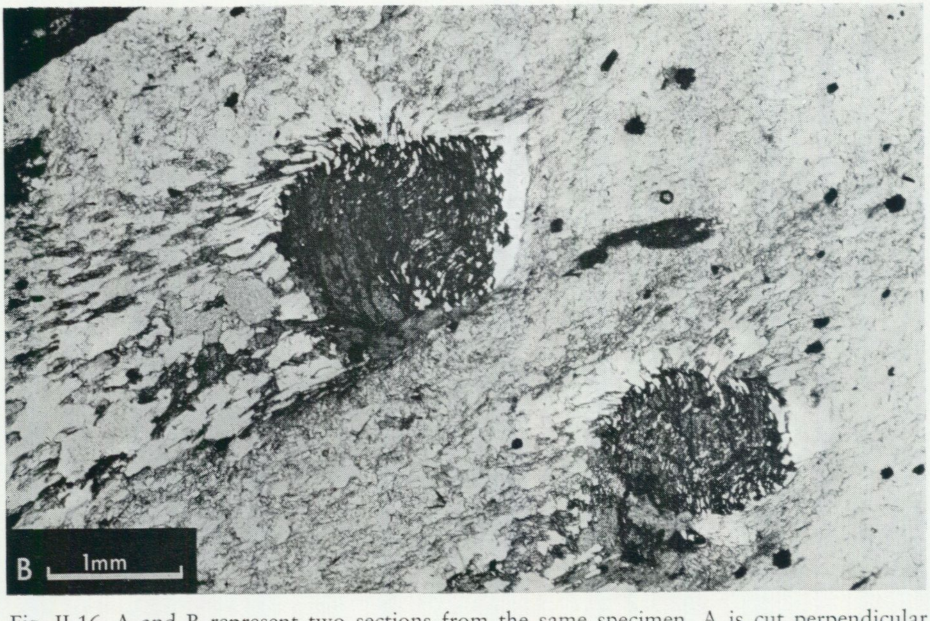


Fig. II-16. A and B represent two sections from the same specimen. A is cut perpendicular to S_{1+2} and parallel to L^2_{min}. The ( ) pattern in the garnet indicates that the section is parallel to the garnet rotation axis.

B is cut parallel to S_{1+2} and shows S-shaped inclusion patterns. The) (pattern in the surrounding material indicates the orientation of the postcrystalline F_3 rotation axis.

Eleven specimens contained sufficiently clear S-patterns to be analyzed. In ten of these recognizable || or)(patterns were observed and in thin sections perpendicular to the rotation axes rotation angles Ω_i were measured with reference to the general orientation of the S-plane.

In the eleventh specimen ( ) patterns were found in a section perpendicular to S (Fig. II-16). Strongly undulose quartzes around the garnets and throughout the thin section indicate that F_3 deformation was responsible for the rotation of the garnet rotation axis through 70° out the slaty cleavage plane. However, the internal rotation angle Ω_c measured perpendicularly to the axis provides information on the amount of F_2 deformation.

Six from the ten Ω_i values measured are between 30° and 120°; the other three are: 130°, 170° and 180°. One Ω_c value is 135°. This is a biassed estimator of Ω_i, overestimating it up to about 30° (Wilson 1971). The rotation axes have a N-S or NE-SW orientation and all the rotations are anticlockwise, looking north (Fig. II-17).

With these data and assuming that the conclusions reached in the discussion of the various possible mechanisms which could be responsible for the rotational textures (Chapter VIII) are correct, the following results could be calculated.

1. The amount of simple shear γ, during F_2 in the schists from the Fatmomakk Formation was at least 5–6.2.
2. The flattening of S_{1+2} around the garnets (see Chapter VIII) is in the order of 40–70 %. The shortening of Z, calculated for $\gamma=5$ with an initial $\gamma=0.4$ is 94 %. This means that considerable deformation preceded the nucleation of garnet, which is in accordance with F_1 deformation (see Chapter VIII for further explanation).
3. The sense of shear during F_2 was, as far as the eleven specimens are representative, anticlockwise, looking north.
4. The specimen deformed by F_3 indicates a minimum γ of 2.4 and the same sense of shear during F_3.

Folding mechanisms

The significant later deformation makes comment on F_1 folding mechanisms hazardous. On the other hand, the numerous F_2 folds provide valuable material to speculate upon F_2 folding mechanisms. The folds belong to classes 1C, 2 (never ideal) and 3 (Ramsay 1967, p. 364) (Figs. II-6, 7, 8); most of them are close to class 2.

Ramsay (1962, 1967) discussed extensively the mechanisms possibly responsible for such "similar" folds. He concludes that these folds are best explained as the result of initial buckling, followed by homogeneous deformation (see also De Sitter 1956; Flinn 1962; Ramberg 1964 and Milnes 1971). In this mechanism the variety in F_2 folds, from class 1C to 3, can be

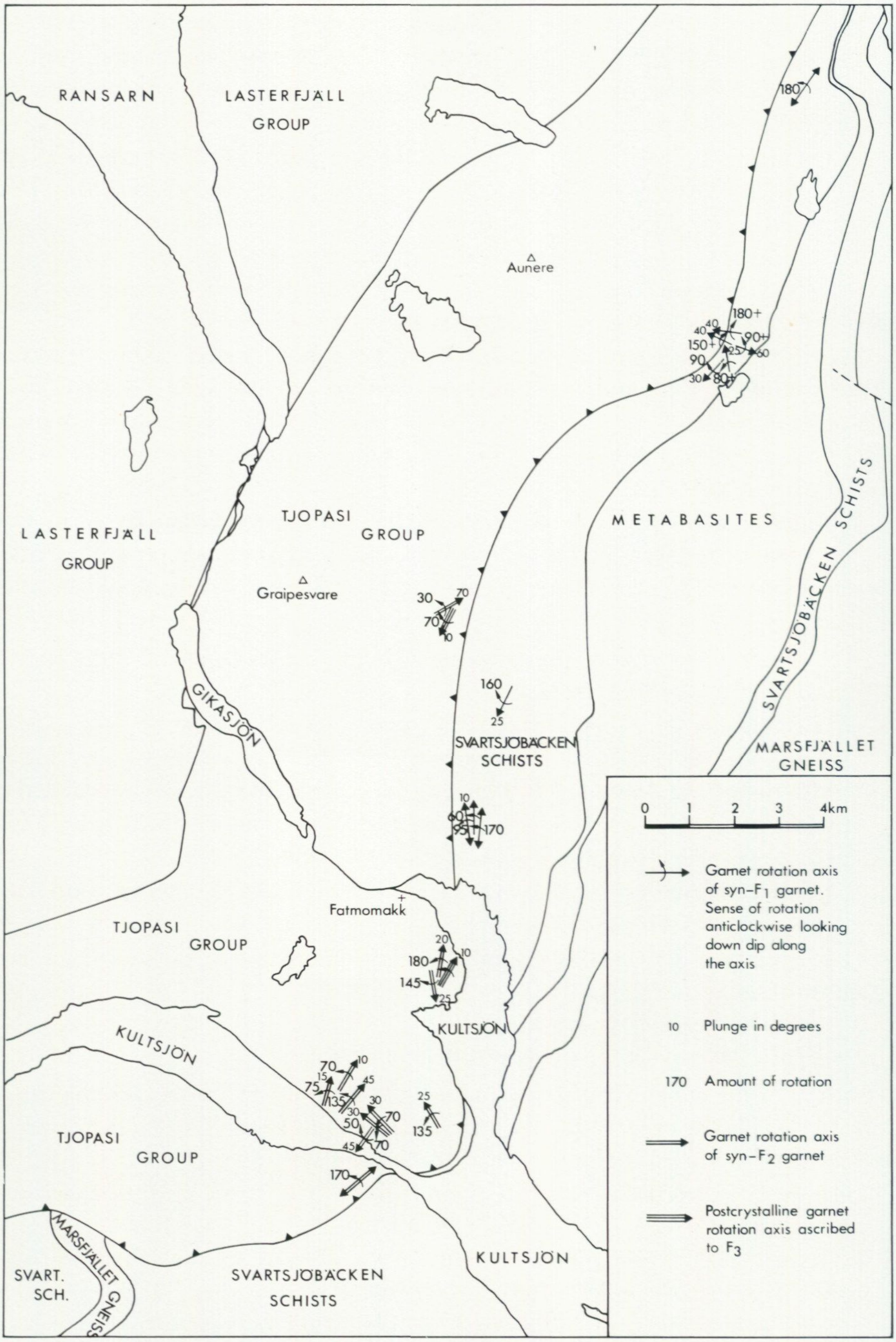


Fig. II-17. Map showing the western part of the area. F_2 garnet rotation axes, with the sense of rotation, are indicated in the schists from the Fatmomakk Formation. In the Svartsjöbäcken Schists F_1 rotation axes are marked (see Chapter III).

explained as the result of variations in the relative importance of the buckling in the total deformation, caused by variations in viscosity contrast and the thickness of layers. It is argued in the discussion on crenulation cleavage (Chapter VIII) that no slip or shear need take place parallel to the crenulation cleavage planes, but that buckling (or kinking) of the slaty cleavage combined with solution transfer can explain the facts observed. The combination of the two mechanisms leads to buckling on two scales, one of the bedding planes and one of the slaty cleavage planes. The result is a rather irregular outline of the bedding plane folds, often interpreted by slip on cleavage planes, but in fact caused by small parasitic folds in the slaty cleavage.

The old problem of the reason of slip or shear in opposite senses on surfaces of maximum flattening, i.e. axial plane cleavage, in so-called "passive flow" or passive slip folds (Donath and Parker 1964) or in "slip folds" (Whitten 1966) no longer exists in such a model. F_3 folds are probably formed according to the same mechanism, but the relatively higher importance of buckle and kink shapes indicates relatively less homogeneous deformation probably as a result of a less ductile state of the rock (see also Dewey 1965). This is in accordance with the lower metamorphic circumstances during F_3. The gentle F_4 folds with subhorizontal axial planes developed preferably in phyllites with a well-developed steep cleavage. They could very well have originated as a result of gravity forces.

CONCLUSIONS

1) During a first phase of deformation (F_1) a slaty cleavage (S_1) was formed, associated with tight folds. Deformed pebbles in conglomerates show that the deformation was, at least locally, intense. Towards the end of F_1 ilmenite and very locally biotite porphyroblasts began to grow. The original orientation of F_1 structures could not be determined, but the present SW-dipping position of axes suggests an originally SW–NE trend.
2) In an interkinematic stage, between F_1 and F_2, most of the biotite and ilmenite porphyroblasts grew.
3) During a second phase of deformation (F_2), S_1 was deformed. A crenulation cleavage (S_2), a slaty cleavage (S_{1+2}) and locally a fracture cleavage (S_2) were formed during this phase. F_2 produced numerous close to open folds probably by the mechanism of initial buckling followed by homogeneous deformation. In the schists of the Fatmomakk Formation, garnets grew and with the aid of curved inclusion patterns the sense and minimum value of the simple shear component in the homogeneous deformation during and after garnet growth could be determined: $\gamma_{min} = 5–6.2$. The sense of rotation is, looking north, anticlockwise (Fig. II-17).

   It is believed that S_2 and L_{2min} together form an LS fabric as described by Flinn (1965). According to this interpretation, L_{2min} is parallel to the X axis of the F_2 strain ellipsoid and S_2 is parallel to the XY plane. It follows from the diagrams that the maximum of F_2 fold axes (15/30) is in between the general X and Y directions of F_2 strain.

The general orientation of F_2 principal strains is according to this interpretation as follows: X approx. 320/55, Y approx. 220/5, Z approx. 125/35, except from the southwestern part of the area, where important F_3 folding took place.

4) During a third deformation phase (F_3) folding took place on a macro-scale, and locally on a meso- or micro-scale. A new crenulation cleavage (S_3) only developed in a few places. The folds are mainly formed by buckling and kinking of S_2 and SS. The structures generally have a N–S trend (Zachrisson 1964, 1969), but numerous deviations occur. In the SW part of the area the principal F_3 axial trend is approx. 330/30; S_3 is here approx. 230/80.
5) A fourth deformation phase (F_4) locally produced gentle or open folds with subhorizontal axial planes and approx. N–S axes or crenulation lineations. The phase is of minor importance and could very well be the result of gravity forces.
6) The zone of so-called "Kärvskiffrar" (porhyroblastic phyllites or schists) often mentioned in Scandinavian literature (e.g. Magnusson et al. 1960) is in the present area explained as a zone of phyllites and fine-grained schists in which the metamorphism was high enough to permit porphyroblasts to grow (e.g. biotite, hornblende, garnet), but where later metamorphism and deformation were not sufficiently important to destroy the grain size contrast, either by coarsening of the matrix, or by deformation and recrystallization of the porphyroblasts.

CHAPTER III

The Svartsjöbäcken Schists

INTRODUCTION

The Svartsjöbäcken Schists comprise garnet mica schists and metabasic rocks. These two rock types will be dealt with separately, as they show different results of the tectonic phases. The nature of the transition from the phyllite belt to this unit is described in Chapter I (p. 10). The garnet mica schists within and below the transition zone are highly deformed. The well-developed schistosity is strongly deflected around garnets and the few visible folds are isoclinal. Deeper in the succession the metabasic rocks become more abundant and the rocks show less traces of intense deformation. The transition to the next unit, the Marsfjället Gneiss, is the metamorphic transition from schist to gneiss, marked by the "potash feldspar-in" isograd (Glass, in prep.) and the fact that kyanite becomes ubiquitous. The nature of this metamorphic transition is discussed in detail by Glass (in prep.).

SCHISTS

General aspects

The well-developed schistosity gives the schists a pronounced planar appearance in the field. They are quartz-rich, fairly homogeneous schists, in which bedding is in general not recognizable. Garnet porphyroblasts with an average grain size of about one cm are abundant. They are often idioblastic. In a few localities the schists contain hornblende porphyroblasts ("Garbenschiefer"). Michel (1950, pp. 50–55) described part of these schists as "the Western-garnet-mica-schist-series".

Petrography

The main constituents of the schists are quartz and muscovite. Other common minerals are: biotite, garnet, epidote group minerals and plagioclase. The plagioclase is zoned with cores of An_{30}–An_{50} and rims going down to about An_{20}. Chlorite crystals locally crosscut the schistosity; the mineral replaces biotite and garnet. Epidote group minerals are often zoned, some have an orthite core. Kyanite and staurolite occur locally. Kyanite often has a muscovite replacement corona. The hornblende in the "Garbenschiefer" is bluish green. Accessory minerals are: tourmaline (very common), apatite, carbonate, ilmenite, hematite, sphene, rutile and zircon.

Structures

Microscopic study revealed that the well-developed schistosity is generally a transposition fabric. The muscovite crystals are locally arranged in regular microfolds, resembling a coarse crenulation cleavage (Fig. III-1). Fine parallel inclusions in the centre of many garnets indicate that there was a slaty cleavage in the rock at the time when these crystals began to grow (Fig. III-8). S_i of these garnets is never continuous with S_e, it is interpreted as S_1; the main schistosity, S_e, as S_2. The mineral lineation accompanying the schistosity is then $L_{2\,min}$ and folds having S_2 as axial plane are F_2 folds. Structures in which S_2 is deformed are ascribed to F_3.

F_1. Only the slaty cleavage, S_1, included in garnet crystals and locally preserved in the structures resembling crenulation cleavage, documents this phase.

F_2. F_2 folds, folding quartz veins, bedding or fine irregular quartz-rich layers, are common (Figs. III-2, 3). In the upper part they are difficult to locate on account of their tightness and the homogeneity of the rocks. Lower in the succession they are tight to close and show better in the field. Their orientation is rather irregular. The few fold axes measured, however, show

Fig. III-1. Coarse crenulation cleavage in the Svartsjöbäcken Schists, Tjäpsvardo.

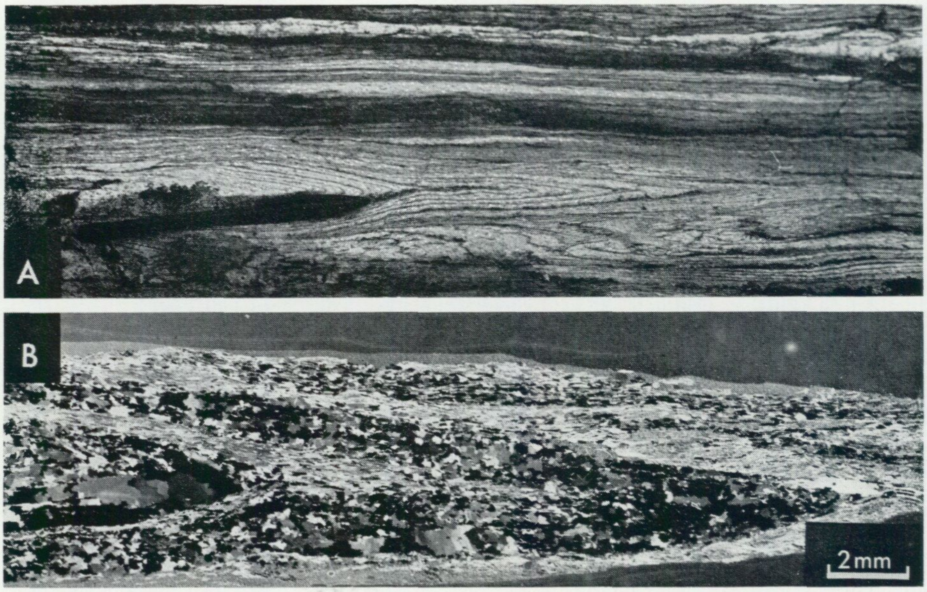


Fig. III-2. Tight to isoclinal F_2 folds. A: in the field. B: under the microscope (negative print), Svartsjöbäcken.

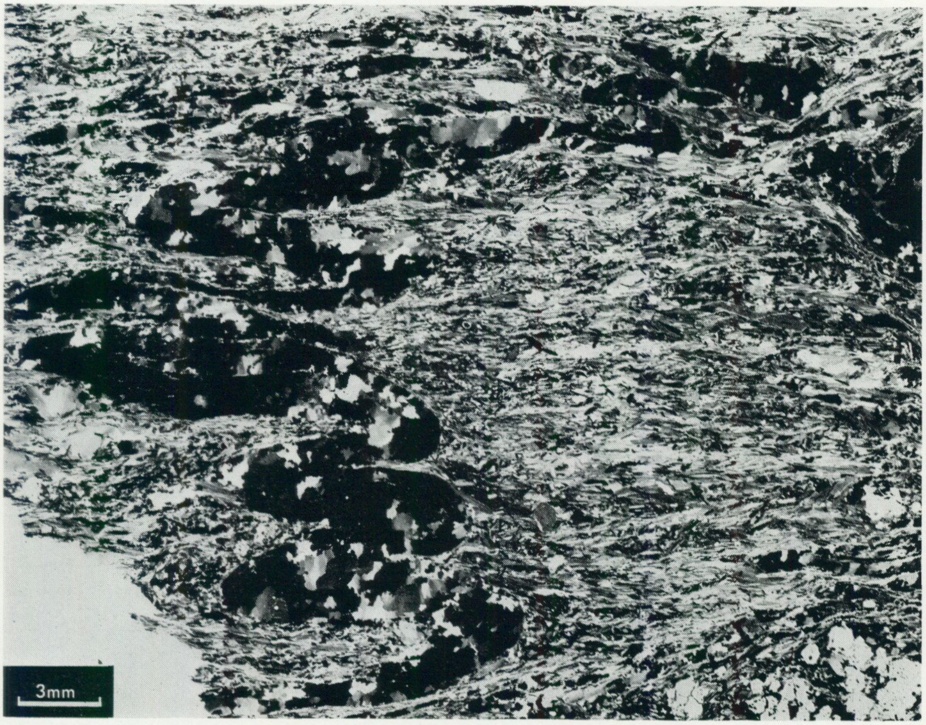


Fig. III-3. F_2 folds in quartz vein, with typical S_2 transposition fabric (negative print, crossed nicols).

a tendency to concentrate around L_{2min} (Enclosure II). Except for the structures described, resembling crenulation cleavage, S_2-planes can be normal schistosity planes, consisting of parallel micas. The most common, however, especially in the upper part of the formation, is an intermediate cleavage type: micas, almost free of strain, form a kind of irregular interlace and are only statistically parallel to the S-plane (Fig. III-3).

The quartz grains are generally free of strain; they form elongated aggregates and sometimes show resemblances to granulite textures. The orientation of S_2-planes is fairly constant. They dip slightly to moderately to the W or NW. L_{2min} is not everywhere developed, locally it is, however, very pronounced and in a few places it even domniates the S_2 structure. The lineation is formed by trains of crystals and by fabric habit of individual minerals. The orientation is mainly down dip, plunging W to NW.

F_3. Only locally S_2 is folded, mainly in open folds. It can be observed in thin sections that these folds always have strongly deformed quartzes in the hinge (Glass, in prep.), often with deformation lamellae. The deformed quartzes indicate local F_3 deformation without folding. In a few places F_3

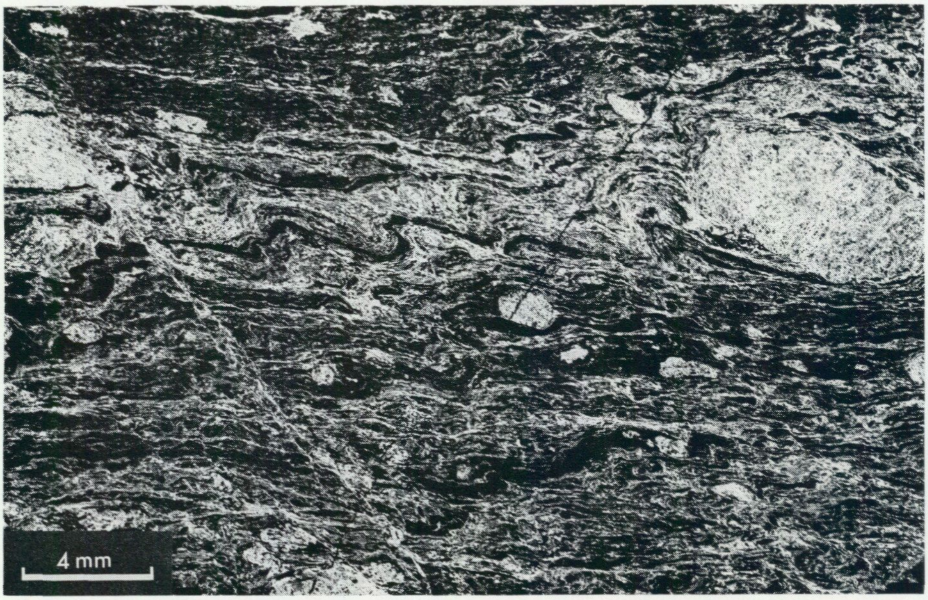


Fig. III-4. F_3 mylonite (negative print).

mylonite zones are encountered (Fig. III-4). They are characterized by the fact that no visible recrystallization took place.

From the diagrams (Enclosure II) it can be seen that the orientation of S_3, though rather irregular, does not deviate much from that of S_2. The fold axes dip NW in sub-area 11. In sub-areas 13 and 14 they are more irregular.

METABASIC ROCKS

Introduction

Due to the poor exposure it could not be established whether the metabasic rocks form elongated bodies or continuous layers with a varying thickness. The mapping suggests that both possibilities occur.

The thickest continuous amphibolite is exposed in Djupbäcken; it is at least 300 metres thick.

A high epidote content is very characteristic of these rocks. The mineral, together with plagioclase, is concentrated in thin layers (<2 mm thick), which alternate with layers poor in epidote-plagioclase (up to 1 cm thick), probably as a result of metamorphic segregation. Thicker layers, up to several metres in thickness, rich in epidote-plagioclase and sometimes carbonate, also occur. Locally the amount of epidote-plagioclase substantially exceeds that of hornblende. These thicker layers, which are not so well differentiated on a small scale, could very well be remnants of bedding. Another characteristic of these rocks is the scarcity of garnet. It was only found in a few hornblenditic parts. In Rodinsbäcken a level is exposed with hornblenditic bodies (up to 3 m Ø), embedded in a mylonitic schist, rich in biotite and garnet and surrounded by amphibolite. These bodies consist purely of hornblende, actinolite, garnet and ore minerals.

On Saletjålt and Klingere some scattered exposures of entirely fresh crystalline rock occur. This rock type contains biotite, plagioclase, garnet and a variable proportion of hornblende. These minerals have a random fabric, they apparently grew under static conditions. In an amphibolite boudin near Gitshobben, the hornblende also has a random fabric.

The map shows a layer on Fieteres in which carbonate and epidote are the predominant minerals.

Michel (1950, pp. 82–87) described the northern part of the metabasic rocks under the name "the Western-amphibolite-series". His observations can in general be agreed with. For a detailed description of rock types the reader is referred to Michel (1950).

Petrography

The metabasic rocks are in general fine to medium crystalline. Coarse crystals occur locally, both in hornblenditic parts and in parts rich in epidote-plagioclase.

The main constituents are: bluish green hornblende, epidote, normally zoned plagioclase (An_{30}–An_{20}), quartz, biotite, carbonate and opaques (ilmenite-hematite).

Sphene is a common accessory. Chlorite is sometimes formed as a secondary mineral, concentrated along joints. White mica was found in some carbonate-rich rocks. The rocks in general contain too much epidote, plagioclase, quartz, biotite and/or carbonate to be called amphibolites.

As no original textures are preserved it is very difficult to say anything about the nature of the source material. However, the interbedding with schists of sedimentary origin, and the composition make a tuffitic or other volcanic origin of at least part of them probable.

Michel arrives at the same conclusion, namely that "the Western-amphibolite-series" consists chiefly of metamorphosed extrusive basic igneous and tuffaceous rocks. According to him, metamorphosed intrusive basic igneous rocks might be also present, whereas argillaceous sedimentary rocks form local intercalations (Michel 1950, p. 87).

Structures

Apart from a plano-linear hornblende fabric, abundant folds occur. They are classified in three groups:

1. A compositional layering, probably bedding, is folded. If the hornblende

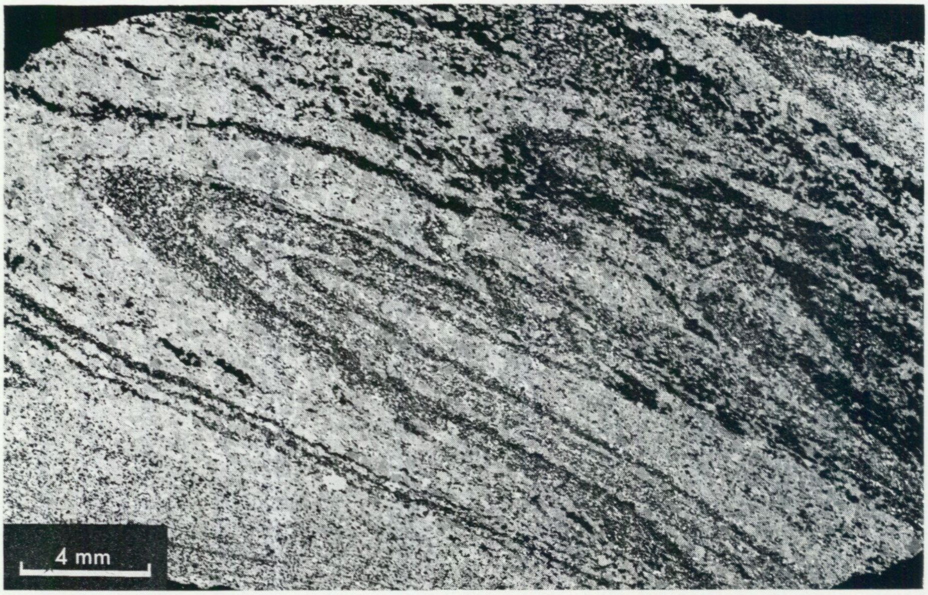


Fig. III-5. F_1 or F_2 fold (group 1) in epidote-rich amphibolite near Blerikstugan. No clear planar hornblende fabric is present (negative print).

crystals form a planar fabric, they are oriented in the axial plane. If a linear hornblende fabric occurs, it is parallel to the fold axis (Fig. III-5).

2. A hornblende fabric, often accompanied by a compositional layering, which is probably the result of metamorphic segregation, is folded. If quartz is present, the crystals are apparently recrystallized since they have no undulose extinction, not even in the fold hinges. Sometimes recrystallized hornblende crystals are oriented in the axial plane. The fold axes are parallel or, in one case, perpendicular to the mineral lineation of hornblende (Fig. III-6).
3. As group 2, but in this group do the quartz crystals have a strong undulose extinction. If no quartz is present, folds of this group cannot be distinguished from those of group 2. In a few of such cases the overprinting of group 2 folds by these ones allows their identification. The fold axes are sometimes parallel to the mineral lineation, sometimes they are not (Fig. III-6).

The plano-linear hornblende fabric is formed by a preferred orientation of hornblende crystals. The lineation is defined as the preferred orientation of *c*-axes, or longest dimensions of the crystals. The cleavage or schistosity plane is defined as the plane in which the hornblende crystals have their longest dimensions. This may coincide with the (100) plane, but not necessarily. All transitions occur from a purely linear fabric to a plano-linear fabric to a purely planar fabric.

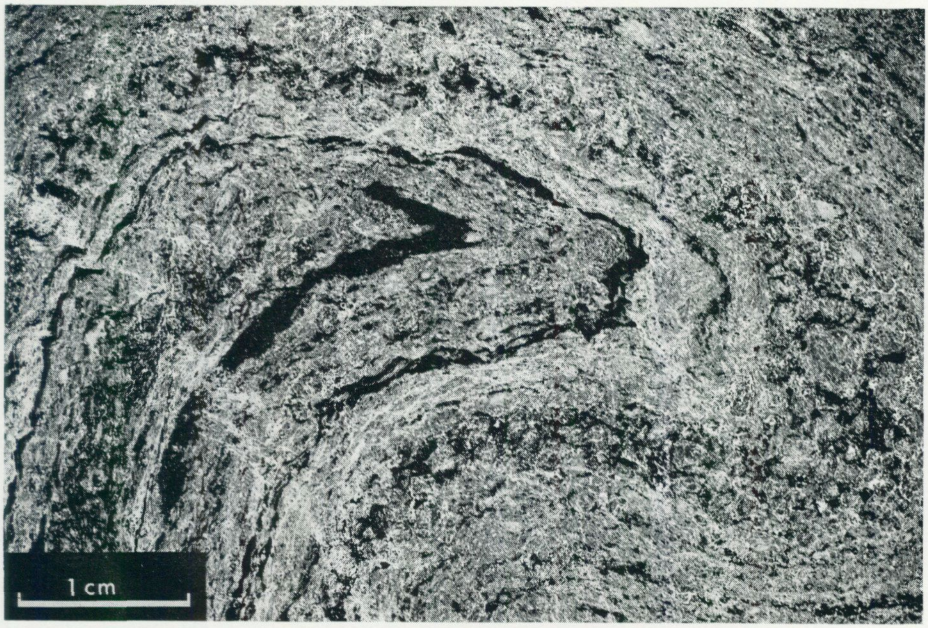


Fig. III-6. F_3 fold (group 3) overprinting F_2 fold (group 2). Note the folded fabric in the F_2 fold (negative print).

Based on these observations the following tentative history can be given of the tectonic events in the amphibolites.

A: During a first deformation phase (F_1) a planar and probably linear fabric was formed (S_1). The hornblende crystals grew during this phase and a metamorphic segregation took place.
Folds of the original compositional layering, probably bedding, were possibly formed. They can, however, not be assigned to this deformation with certainty since the folds of group I could just as well have formed during F_2.

B: A second deformation, F_2, folded the metamorphic segregation layers and S_1 into tight folds. It can be demonstrated that quartz and, in some occasions, hornblende recrystallized during F_2. Where no folds are formed S_1 and L_1 were probably deformed homogeneously to assume a position related to the finite strain ellipsoid after F_2. The fact that most F_2 folds have their axes parallel to the mineral lineation makes it probable that this lineation was formed during F_2, either mechanically or as a result of recrystallization.

C: During a third deformation, F_3, only local folding took place. In some localities F_2 folds are refolded. No minerals recrystallized during this deformation, which therefore probably took place under low-grade conditions (undulose quartzes). The $L_{2\,min}$ was locally used again as fold axis during F_3. In such cases the (100) planes of hornblende crystals follow the fold.

The folds of group 1 were in this scheme formed during F_1 or F_2, those of group 2 during F_2 and those of group 3 during F_3.

It seems logical to correlate these deformation phases with those of the surrounding schists. The more so since they fit very well, as far as the metamorphic circumstances and the orientation are concerned.

DISCUSSION

Rotated garnets

In this formation garnets with S-shaped inclusions are predominantly surrounded by a transposition schistosity. Only in a few cases is a slaty cleavage preserved. Most of the garnets have small inclusions in the centre, interpreted as representing the S_1 slaty cleavage. Towards the rim the inclusions become coarser, until in the rim itself they are comparable to, or even coarser than the matrix crystals (Fig. III-7). Apart from the cases in which a slaty cleavage still exists, S_i is not continuous with S_e. The textures described are interpreted as follows:

1. After the formation of a slaty cleavage during F_1, garnets nucleated and were rotated during growth.
2. During and probably after F_1 the rocks coarsened. This is documented in the rims of the garnets.

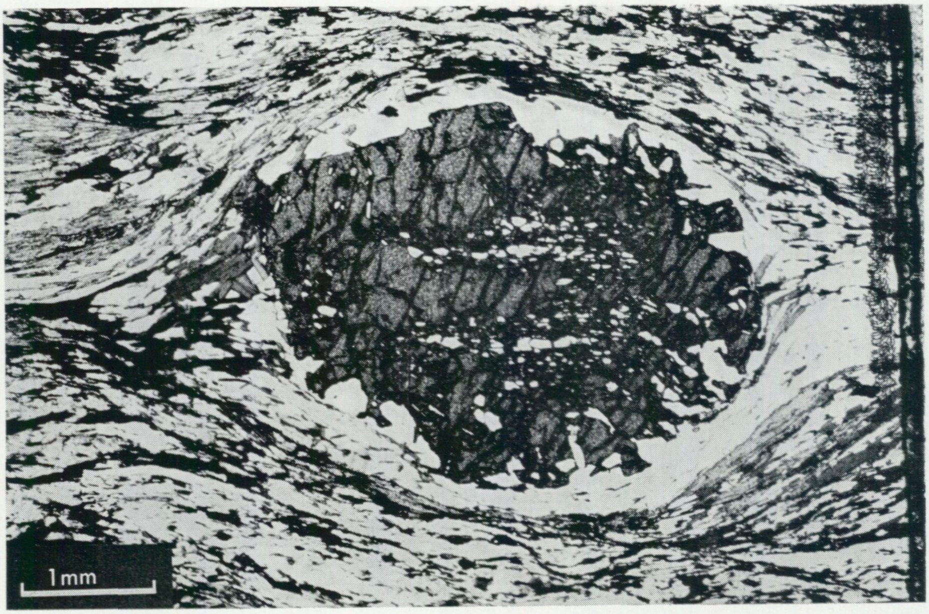


Fig. III-7. Syn-F_1 garnet with small inclusions in the centre (S_1) and larger ones in the rim, reflecting a coarsening of the matrix during growth. S_i is not continuous with S_e (S_2).

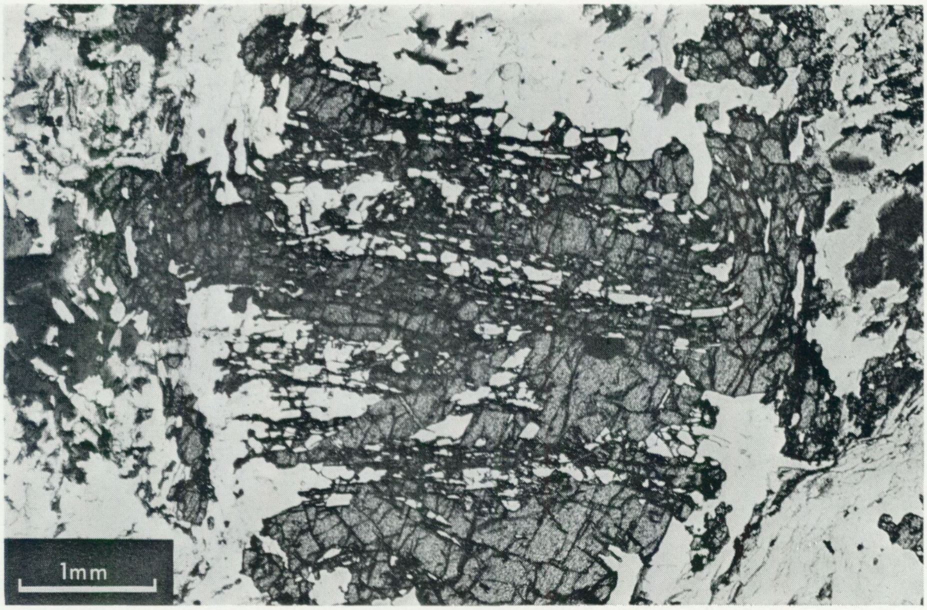


Fig. III-8.) (pattern in a syn-F_1 garnet cut parallel to the rotation axis. Section parallel to S_2.

3. During F_2 the schistosity was deformed into a transposition schistosity. The grain size was reduced. Recrystallization of quartz and mica took place and locally small new garnets and rims around old ones grew. In a few places the slaty cleavage remained intact to form an S_{1+2} plane.

There was little reason to search for F_1 rotation axes in thin sections parallel to S_2, but as no better method was available it was done all the same. The rather amazing result was that in eight of the twelve specimens searched the g.r.a. (garnet rotation axes) were indeed situated in S_2 (Fig. III-8). There is no point in measuring the rotation with respect to S_2, as can be demonstrated by means of a simple diagram (Fig. III-9). The internal rotation angles Ω_c were, however, measured, to obtain information on the shear during F_1. Ω_c was found to be

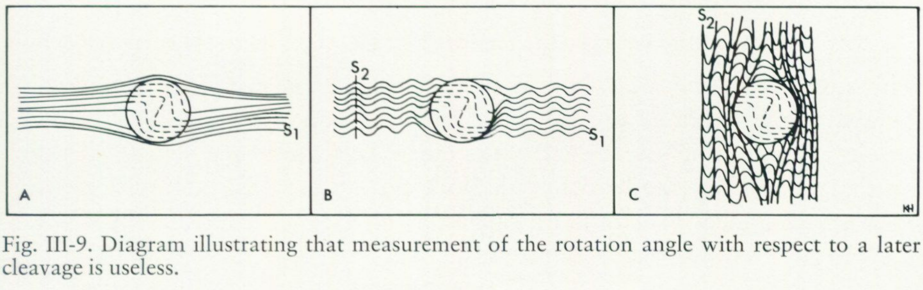


Fig. III-9. Diagram illustrating that measurement of the rotation angle with respect to a later cleavage is useless.

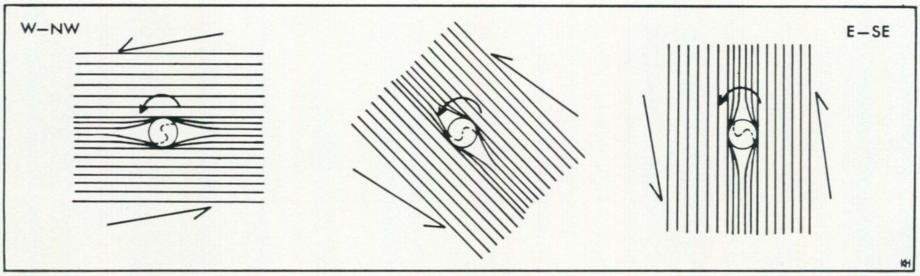


Fig. III-10. Possible orientations of S_1 before F_2. The intermediate situations are, of course, possible as well.

about 90°. It is a biassed estimator of Ω_i up to about 30° too high (Wilson 1971); Ω_i in turn is a biassed estimator of Ω_e up to about 30° too low (see Chapter VIII), so that $\Omega_e = 90° \pm 30°$. This results in a minimum shear during F_1 of 3.1 ± 1 ($\Omega_e = |{}^1/_2\,\gamma|$, Chapter VIII). The rotation axes now plunge slightly N to NE and all the rotations are, looking in that direction, anticlockwise (Fig. II-17). Although F_2 might have changed the attitude of the g.r.a. it has probably not changed the sense of rotation. The angle between S_1 and the present position of S_2 was probably larger than 45°, otherwise no crenulation cleavage or transposition fabric would have formed. The orientation of S_1 before F_2 was therefore probably as outlined in Fig. III-10. In one specimen with a slaty cleavage S_{1+2}, the same sense of rotation is observed. Ω_i of 95° here represents both F_1 and F_2. A minimum shear γ of 3.2 ± 1 does not give much information on F_2, because the strong deflections of S_2 (the shortening is about 90 %) indicate a strong deformation as it is. Other samples indicate the same sense of rotation by little drag folds and asymmetric deflection patterns of S_2. One exception of a small syn-F_2 garnet, showing a clockwise rotation of about 25° confirms the rule.

The conclusion seems justified that during F_2 the sense of shear was anticlockwise looking N to NE, just like during F_1. This conclusion tallies with the results from the schists of the Fatmomakk Formation on the sense of shear during F_2.

The transition zone between the phyllite belt and the Svartsjöbäcken Schists

This transition zone, described in detail by Glass (in prep.), is characterized by a strongly deformed upper part of the Svartsjöbäcken Schists and a small gap in metamorphic assemblages. It is apparently a tectonic contact although no sharp boundary can be detected. Since the intense deformation dates from F_2, it seems logical to date the thrusting as an F_2 phenomenon. According to the porphyroblasts, the metamorphism was still high during F_2 (Glass, in prep.). If the thrusting took place before F_2, any metamorphic break would probably have been effaced. If, on the other hand, the thrusting postdates F_2, a low-grade mylonite zone is to be expected, which is not found.

The strain analysis on the base of rotated garnets revealed that the sense of F_2 shear on both sides of the transition zone is anticlockwise, looking north. F_1 and F_3 shear in the Svartsjöbäcken Schists have probably the same sense. The thrust is therefore interpreted as a relative downthrust, which is in accordance with the fact that lower grade rocks are lying, with a gap, on higher grade ones.

Correlation of deformation phases

F_2 in the Svartsjöbäcken Schists has many aspects in common with F_2 in the phyllite belt:

1. It is responsible for the regional cleavage or schistosity.
2. The metamorphic circumstances were relatively high; recrystallization of quartz, mica and locally of hornblende can be demonstrated.
3. Their S-planes are parallel, and parallel to the thrust contact.

It is therefore concluded that they can be correlated with each other. The correlation of F_1 in the Svartsjöbäcken Schists and F_1 in the phyllites is uncertain since one of the two units could have been transported, during F_2, from another area with a different tectonic history.

F_3 in the Svartsjöbäcken Schists probably stands for both F_3 and F_4 in the phyllites. The spread in fold axes and axial planes supports this idea. It can, however, not be proved since no superimposition could be demonstrated.

Folding mechanisms

As discussed in Chapter II, the plano-linear fabric in the schists is interpreted as an LS fabric such as described by Flinn (1965). This means that S_2 is oriented along the XY plane, and $L_{2\,min}$ along the X direction of the F_2 strain ellipsoid. F_2 fold axes are rather irregularly distributed in the XY plane (S_2), but show a tendency to concentrate around L_{2min}, or the X axis of the strain ellipsoid (Enclosure II, sub-areas 13, 14). According to Flinn (1962), this is to be expected if these fold axes are passively deformed during homogeneous deformation. In the phyllites the F_2 fold axes are concentrated in between X and Y, probably as in these lower metamorphic rocks the anisotropy during deformation was higher. Another reason for the difference may be that the homogeneous deformation in the schists was stronger. As indicated by the tightness of the folds

(Flinn 1962) the general principle of initial buckling, followed by homogeneous deformation, as discussed before, also seems valid, the only difference with the phyllites being that the homogeneous deformation was more pronounced in the schists.

CONCLUSIONS

The events in the Svartsjöbäcken Schists after sedimentation may be summarized as follows:

1. During F_1 a slaty cleavage was formed.
   Garnets began to grow and were rotated by F_1 in an anticlockwise direction looking N. In the basic rocks a hornblende fabric was formed, possibly associated with folding.
2. Interkinematic or late syn-F_1 coarsening to schist.
3. During F_2 folding of S_1 and SS took place, associated with the formation of a penetrative plano-linear fabric. The schistosity is partly of the slaty cleavage type (S_{1+2}) and partly of the crenulation cleavage type (S_2). In general it is however an intermediate transposition schistosity. The orientation of S_2 is in general about 290/30 with $L_{2\,min}$ approx. down dip. Most of the minerals which are deformed during this phase recrystallized during or after their deformation.
   A relative downthrust took place from E to W of the phyllite belt with respect to the Svartsjöbäcken Schists, as documented by the sense of rotation of syn-F_2 garnets in the schists from the Fatmomakk Formation.
4. Local folding and minor thrusting took place during F_3, under low metamorphic conditions. The orientation of F_3 phenomena is irregular.

CHAPTER IV

The Marsfjället Gneiss

INTRODUCTION

The Marsfjället Gneiss comprises pelitic and quartzo-feldspatic gneisses and metabasic rocks. The gneisses are characterized by the mineral pair potash feldspar-kyanite. The metabasic rocks often contain the mineral association hornblende-plagioclase-garnet-clinopyroxene. Locally the rocks of this formation are migmatized. The transition from the Svartsjöbäcken Schists into this formation is described in the preceding chapter. The lower boundary of the Marsfjället Gneiss is situated below a blastomylonite zone (exposed on Gakkangaise and Risfjället) or below a complex belt with several mylonitic levels (exposed on Borkafjället). Below this boundary the schists and amphibolites of the eastern schist and amphibolite belt occur. The boundary obviously has tectonic significance, since a considerable metamorphic jump can be demonstrated (Glass, in prep.).

A section of Borkafjället is described separately, since the gneisses on this mountain differ somewhat from those occuring on Marsfjället. The rocks from Borkafjället have been described by Du Rietz (1938) and Michel (1950).

GNEISSES

Field aspects

West of the transition into the Svartsjöbäcken Schists the gneisses gradually grow coarser. In the fine- to medium-grained, mainly quartzo-feldspatic gneisses, close to the boundary, a schistosity is usually well developed, often accompanied by a mineral lineation and sometimes by a fine layering. The latter is formed by an alternation of biotite-rich dark layers, up to about 1 mm thick, and layers of about 1 cm, rich in quartz and feldspar. As the gneisses grow coarser, both the schistosity and the layering become less well defined. The layering grades into veinlets and patches, sometimes very irregular, and locally the schistosity is vague. In places melanocratic (biotite-garnet-kyanite) layers, up to 2–3 cm thick, alternate with leucocratic (quartz–feldspar) layers with a thickness of up to 5 cm. These layers are irregular.

Particularly in the middle and lower part of the formation leucosomes occur, i.e. bodies or lenses consisting principally of quartz and feldspar. They apparently originated before the last important deformation, since the bodies are deformed into lense shapes. They can be subdivided as follows:

1. Intruded pegmatitic veins and layers primarily composed of quartz and plagioclase. Coarse muscovite may be present.
2. Leucosomes which are believed to have formed in situ, because of the melanocratic rims around them. They are bodies from a few centimetres across up to lenses of about 1×3 m. The composition is: plagioclase + potash feldspar + quartz ± garnet.

Larger leucocratic bodies, up to 10×50 m in size, also occur. They are very rich in potash feldspar and contain garnet. They possibly reflect a different original composition. For more detailed descriptions the reader is referred to Du Rietz (1938), Michel (1950) and Glass (in prep.).

Petrography

Quartz is in general abundant. The crystals range from coarse, with a strong undulose extinction, to very fine-grained, often undulose as well. The small crystals apparently grew at the expense of the large ones; locally these new ones grew so large that they can hardly be distinguished from the old ones (Hobbs 1968). Perthitic or microperthitic potash feldspar occurs in medium to coarse-grained, often strained crystals. Both microcline and orthoclase are common. Plagioclase is also medium to coarse-grained; it is often full of sericite as well as fine kyanite needles. The anorthite content ranges from An_{30}–An_{50} with rims down to An_{20}. The crystals are sometimes surrounded by a rim of very small, apparently recrystallized new grains. Kyanite, in general, constitutes only a few percent of the rock. It may, however, locally amount to as much as 20 %. The grain size is variable, particularly as a result of replacement by muscovite, shown by coronas. The crystals are often bent or folded.

Garnet is ubiquitous, the crystals are chiefly poikiloblastic. Some are elongated; a few crystals have straight inclusion patterns. The red, or dark brown, biotite is mainly fine-grained as a result of the deformation. Muscovite is never abundant; the crystals are often secondary; in the lower part of the formation, on Gakkangaise and Marsfjället, muscovite is completely lacking. Accessory minerals are: epidote group minerals, scapolite, apatite, sphene, rutile, zircon, carbonate, tourmaline and secondary chlorite.

It may be concluded from the foregoing that a strong postcrystalline deformation, accompanied by retrograde metamorphism caused undulose extinction in many minerals (e.g. quartz, feldspar, kyanite, biotite etc.), reduction of grain size (e.g. quartz, plagioclase, biotite) and irregular grain boundaries.

METABASIC ROCKS

General aspects

In the middle part of the formation numerous metabasic intercalations form layers, lenses, boudins and a few crosscutting dykes, ranging in thickness from one to several hundreds of metres. On Borkafjället the metabasic rocks occur throughout the formation. Several types could be distinguished by the following mineral associations (Glass, in prep.):

1) hornblende + plagioclase
2) hornblende + plagioclase + garnet
3) hornblende + plagioclase + garnet + clinopyroxene + quartz
4) hornblende + plagioclase + clinopyroxene
5) hornblende + plagioclase + garnet + clinopyroxene + scapolite + quartz
6) orthopyroxene + hornblende + plagioclase + quartz.

The first five of these are common; the sixth is very rare. The rocks are medium to coarse crystalline with local leucocratic veins and patches rich in plagioclase (up to about 25 cm). These patches are often surrounded by dark hornblende-rich rims, indicating a metamorphic segregation origin. Big garnets are often conspicuous, they tend to be concentrated, together with clinopyroxene, along the leucocratic parts. The schistosity is in general very weak or not developed at all. A vague segregation layering, marked by parallel leucrocratic veins and fine layers, occurs generally. A fine interlayering of basic and acid rocks locally suggests a sedimentary origin.

Petrography

Hornblende is abundant in almost every sample; it is a green, brownish green or brown variety. In a few cases the crystals are surrounded by a rim of colourless amphibole (anthophyllite).

The light green diopside is locally surrounded by a rim of green hornblende. Normally zoned plagioclases are either approx. An_{55} or approx. An_{80} in the core. In the rims the composition falls to An_{40} and An_{60} respectively. Scapolite is a common accessory mineral, often replacing plagioclase. Other accessory minerals are: epidote group minerals, sphene and rutile. Biotite occurs locally, as well as a trace of muscovite, chlorite and carbonate. Some quartz is in general present. In one specimen kyanite was found. Although retrogression and postcrystalline deformation also affected these rocks, the influence is much weaker than in the gneisses and often hardly visible. The polygonal, approximately straight grain boundaries, produced during growth, are chiefly preserved.

A SECTION OF AINANTJAKKE AND ORTSEN

A number of specimens, taken by Glass, along a section line on Stöken, Ainantjakke and Ortsen, represent different rock types, which will presently be surveyed from E to W. Unfortunately these differences are not sufficiently consistent throughout the area to be mapped. The rocks along the section are subdivided into three units: the lower (1), the middle (2) and the upper unit (3).

1. At the base of the lower unit (Ortsen, Gakkangaise) feldspar-free restites, often rich in quartz, occur abundantly in the gneisses (see also the blasto-mylonite zone on Risfjället). On top of these, perthite, kyanite, garnet gneisses and granulites are present. They are locally more or less layered. Near the top of this unit pink gneisses, rich in potash feldspar, are ubiquitous. Muscovite is lacking in this unit. There are no quartz-feldspar mobilisates associated with the restites. They may have possibly risen to higher levels.
2. The metabasic rocks described above are confined to the middle unit (Ortsen). The gneisses in this part are quartzo-feldspatic; most of them are banded gneisses. Some plagioclase-rich varieties, lacking any potash feldspar, locally contain orthopyroxene. Coarse crystalline quartz-feldspar mobilisates also occur in this unit.
3. In the upper unit (Ainantjakke) the gneisses are strongly planar and linear. The grain size is less than in the other gneisses. This might be due to a stronger deformation or a smaller grain size to begin with. Rocks free of muscovite do not occur, but the mineral is mainly secondary. Some blasto-mylonites occur, rich in biotite and scapolite.

3a On Stöken, still higher in the succession, a number of deformed mica-rich gneisses is present, with highly bent micas and kyanite.

THE BLASTO-MYLONITE ZONE ON RISFJÄLLET AND GAKKANGAISE

The blasto-mylonite zone on Gakkangaise is about 100 metres thick. The zone predominantly comprises mylonitized gneisses, derived from the overlying sequence. In the lower part mylonitic garnet mica schists derived from the underlying rocks are also present.

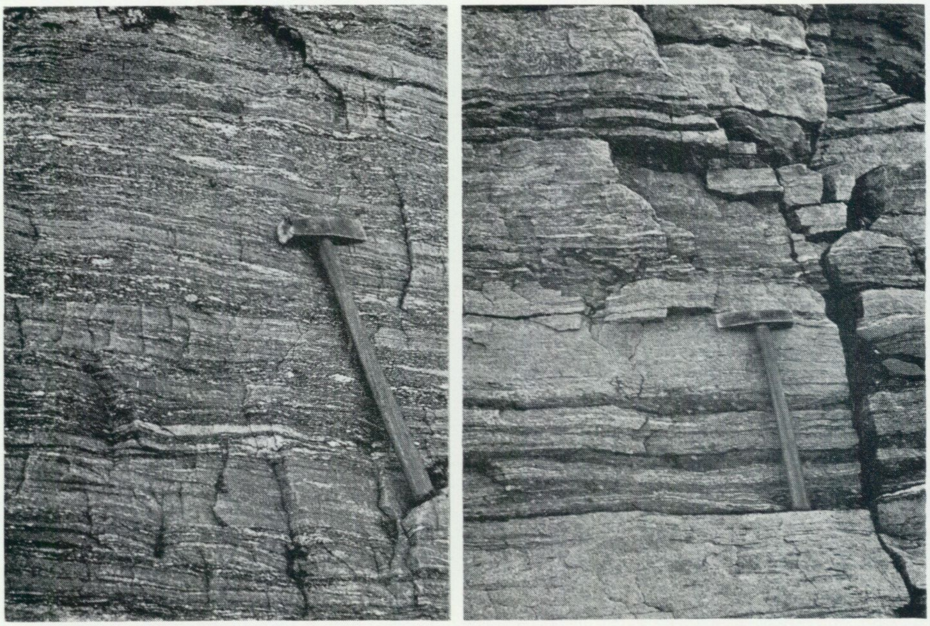


Fig. IV-1. Layering in blasto-mylonite zone, Risfjället. Note the difference in grainsize in the various layers.

The zone is not exposed on Marsfjället. On Risfjället the thickness is about 250 metres (Fig. IV-1). Melanocratic restites, composed of biotite, garnet, kyanite and some quartz occur locally. Quartzitic rocks, with the same minerals, lacking feldspar, are quite common. They may also be restites from very quartzitic source rocks. Towards the south this zone becomes considerably thinner (a few metres on Ropentjakke), and towards the north, between Risfjället and Offerkullen, it is cut off as an imbricate zone by a thrust or fault (Enclosure III). On Borkafjället it is represented by a thick complicated zone of mixed rocks.

The interpretation of the thrust, associated with this zone, as an up-, or overthrust is not only based on the observation that high grade rocks lie on top of low grade ones, but also on a strain analysis on the base of asymmetrical "fish-shaped" muscovites (see Discussion of S_2 planes and mylonitization and Chapter V).

A SECTION OF BORKAFJÄLLET

On Mount Borkafjället the following rock units were distinguished from E–W (Michel 1950); the numbers are indicated on the map (Enclosure I).

1) Fine- to medium-grained amphibolite, rich in epidote and poor in plagioclase, with few garnets.

   Varieties are:
   - coarse hornblendites
   - massive garnet hornblende rocks, rich in sulphides, often surrounded by biotite garnet rocks
   - rocks rich in biotite, scapolite and quartz
   - rocks with the mineral assemblage: quartz, epidote, plagioclase, hornblende, in order of abundance.

Schistosity is poorly developed; a coarse layering occurs in places, defined by a variation in grain size and/or mineral composition.
Petrography: Both bluish and brownish green hornblende are present. The plagioclase has an oligoclase composition. Other minerals are: epidote group minerals (sometimes with orthite core), biotite, garnet, quartz and scapolite.
Accessories are: carbonate, sulphides, sphene and rutile. Apart from the occurrence of scapolite, this unit fits very well in the eastern schist and amphibolite belt (Chapter VI).

2) This unit comprises gneisses, schists and a minor quantity of basic rocks. A major part of the unit is mylonitized and forms strongly planar and linear gneisses.
Different rock types are:
- potash feldspar, kyanite, muscovite gneiss
- potash feldspar, muscovite gneiss (no kyanite)
- scapolite, muscovite, plagioclase schist (no potash feldspar)
- rocks with the assemblage: hornblende, garnet, clinopyroxene, epidote
- garnet, oligoclase, epidote-bearing amphibolite
- biotite, garnet-bearing hornblendite.

The metabasic rocks form thin lenses in this mixed mylonite zone.
The first high-grade rocks occur in this part.

3) Fine-grained basic amphibolite with much epidote, some diopside and locally garnet (plagiopyrigarnite); scapolite is common. The plagioclase in this unit is approx. An_{40}.

4) This unit comprises many different rock types:
- layered potash feldspar, kyanite, muscovite gneiss (very micaceous, occurring on the northern slope of Borkafjället).

– potash feldspar-bearing quartz, plagioclase gneiss (the plagioclase is An_{32}) – plagioclase, quartz gneiss + biotite and garnet ± green hornblende ± colourless hornblende – abundant amphibolite lenses	rocks occuring on the crest of Borkafjället
– plagioclase, garnet, kyanite, biotite rock + trace quartz – garnet biotite rock (restite) – garnet restite – quartz, garnet, biotite rock (restite?) – hornblende ± some garnet (restite?)	rocks occuring on the south-east side of Borkafjället

In general the rocks contain little, but sometimes a good deal of scapolite.

5) Basic, medium-crystalline amphibolite, in places rich in garnet.
Plagiopyrigarnites occur as well as garnet-bearing amphibolites, lacking clinopyroxene.

6) A unit consisting of mixed rock types, all more or less strongly mylonitized. The zone is about 100–200 m. thick, and cuts off several amphibolite layers. Some rock types are:
- kyanite, potash feldspar gneiss + muscovite
- quartz, plagioclase rock + kyanite and scapolite ± hornblende
- quartz, garnet, biotite rock ± plagioclase
- quartz-rich rock bearing green hornblende porphyroclasts + scapolite ± colourless hornblende
- potash feldspar, scapolite, garnet, diopside, green hornblende, quartz, plagioclase rock
- quartz, scapolite, garnet, diopside, epidote rock.

Above this unit the rocks are in general like the Marsfjället Gneiss elsewhere, except for a few deviations:
1. There are few mobilisates or real migmatised gneisses.
2. Most gneisses are quartz, plagioclase, garnet gneisses, containing little or no potash feldspar.
3. Gneisses containing much kyanite, biotite and garnet are common on the west slope of Borkafjället.
4. There is more variation in rock types, for instance the following types occur:
   - potash feldspar, scapolite gneiss
   - kyanite, scapolite gneiss
   - kyanite, muscovite, quartz gneiss, lacking feldspar
   - mica-rich gneiss, lacking kyanite and potash feldspar
   - carbonate-rich rock
   - scapolite, biotite, quartz rock.

The boundary between the Marsfjället Gneiss and the eastern schist and amphibolite belt is drawn at the contact of units (1) and (2), since the lowermost high-grade gneisses occur in unit (2).

STRUCTURES

Introduction

The fissility or schistosity planes are, as microscopic study brought to light, the result of postcrystalline deformation (Glass, in prep.). In the metabasites, which are locally to a lesser extend affected by this deformation, an older schistosity plane, defined by preferred orientation of hornblende can be observed. This plane can be correlated with the planar metamorphic segregation layers. It is labelled S_1 and ascribed to the oldest recognized deformation phase, F_1. The postcrystalline deformation phase is then F_2, and the deformation which distorts the S_2 planes is called F_3.

F_1

S_1 is found almost exclusively in thick amphibolite bodies. It is best preserved in the eastern part of the large metabasic body on Risfjället. A gneissic intercalation, with a selectively better developed S_2, illustrates the angular relationship between S_1 and S_2 (Fig. IV-2). The surrounding amphibolite has a steep S_1 schistosity, parallel to the compositional layering. Other gneiss intercalations contain a more or less random fabric (Fig. IV-6). This is explained by recrystallization after F_1. These intercalations were apparently protected by the amphibolite during F_2, since S_2 has hardly developed. The fact that metamorphic segregation took place along S_1 planes suggests that the metamorphism during F_1 was close to its climax. S_1 planes, where recognized, dip steeply to the east (Enclosure II, sub-area 16). In the localities where S_1 is still visible, a few folds were found with S_1 as axial plane. The segregation layering, previously associated with F_1, is folded in these folds. They probably originated late during F_1, after the segregation layering along S_1 was formed. The fact that

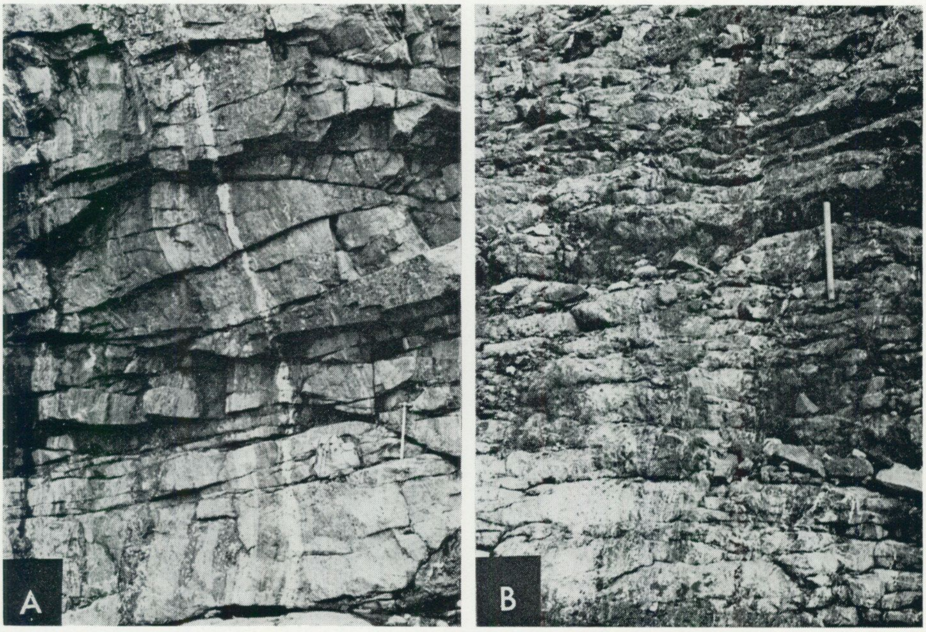


Fig. IV-2. Development of S_2 in amphibolite on Risfjället. A: S_1 is parallel to the compositional layering (SS), S_2 appears as fractures. B: as in A, but S_2 has developed better in these, more gneissic layers.

no more F_1 folds were detected is probably the result of obliteration by the important syn- and postdating metamorphism and deformation.

F_2

Folds of the planar metamorphic segregation layering, which was essentially formed during F_1, have the S_2 plane as axial plane and are called F_2 folds (Figs. IV-3, 4). As these folds are tight to isoclinal, the metamorphic segregation layering, parallel to S_1, is in general also parallel to S_2. Asymmetrical folds, both with vergences to the SW and to the NE, are common. Fold axes of F_2 folds are hard to measure; only few are plotted in the diagrams (Enclosure II). Most of these axes are parallel to the local L_{2min}, plunging W-NW. A few folds are found, folding the S_2 mylonitic layering. According to the recrystallized quartz grains (see below, under F_3) these are still F_2 folds. They have the same orientation as other F_2 folds (Fig. IV-5).

Several stages are found between slightly deformed rocks and blasto-mylonites, both in the gneisses and in the metabasic rocks. S_2 planes and the associated lineation L_{2min}, defined by trains of fine crystals (e.g. quartz and biotite), can therefore be observed in their various stages of development (Figs. IV-6, 7, 8). In little deformed rocks a conjugate set of planes can be observed. In somewhat more deformed rocks, one strongly undulating plane, curving around

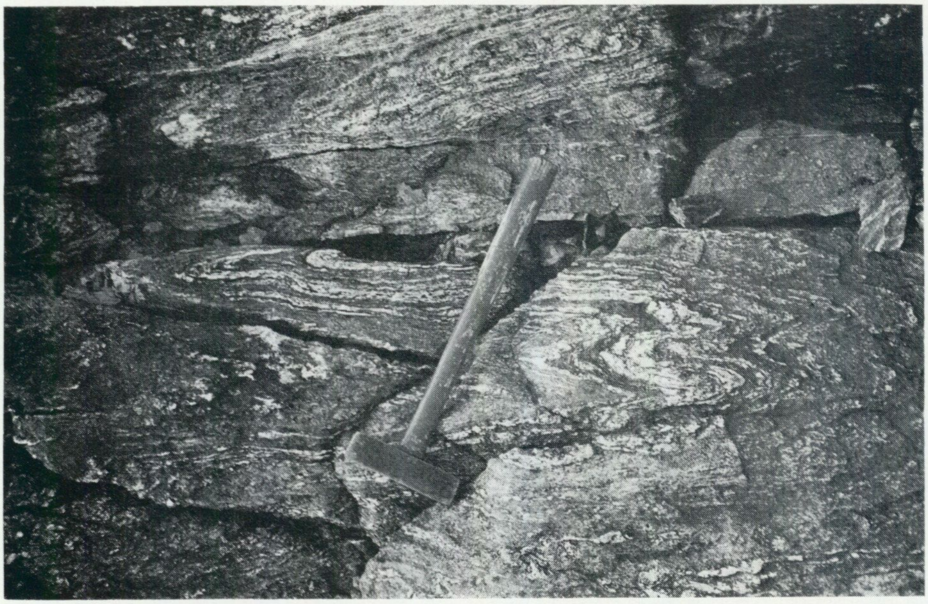


Fig. IV-3. F_2 folds in gneiss. Note that no real axial planar fabric has developed.

Fig. IV-4. Thin section of F_2 fold (negative print).

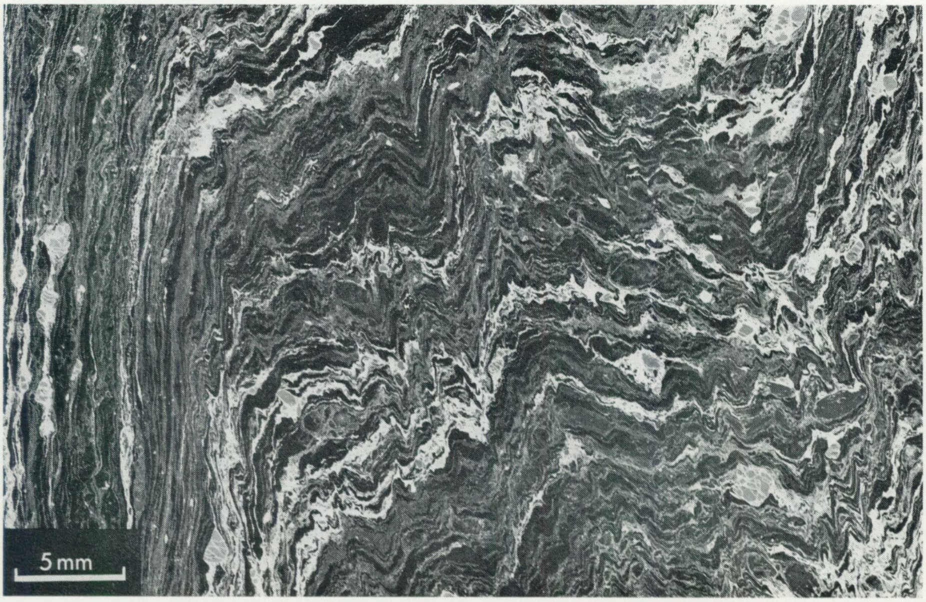


Fig. IV-5. F_2 fold in mylonitic layer, Borkafjället.

porphyroclasts, has developed. Only in the highly deformed rocks, or in originally finer-grained rocks, does S_2 have a more planar appearance. The orientation of S_2 can be read from the diagrams (Enclosure II). On Stöken and Ainantjakke the planes do not dip W to NW, as they do elsewhere, but SW. A possible explanation is the fact that the entire formation wedges out towards the south.

F_3

In some places S_2 is folded in F_3 folds. These folds are open to close and often asymmetrical. Quartz crystals in the fold hinges are deformed under low grade conditions, as may be seen from the strong undulose extinctions, accompanied by deformation lamellae (Glass, in prep.). In one locality on Borkafjället a mylonite occurs without visible recrystallization (Fig. IV-9). It is probably a product of F_3 deformation. Most asymmetrical F_3 folds in the blasto-mylonite zone have equal orientations: N–S axes and a vergence to E or SE (Fig. IV-10). This confirms the general picture of east-west shortening during F_3 (Zachrisson 1969). The orientation of other scattered folds is more irregular. The fact that F_3 folds in the blastomylonite zone are more regular than elsewhere is probably caused by the better developed schistosity of these rocks; that they are more abundant may be the result of concentrated F_3 deformation along this boundary between two large rigid masses. The steep attitude of the schistosity (S_2) along this boundary may very well have the same reason.

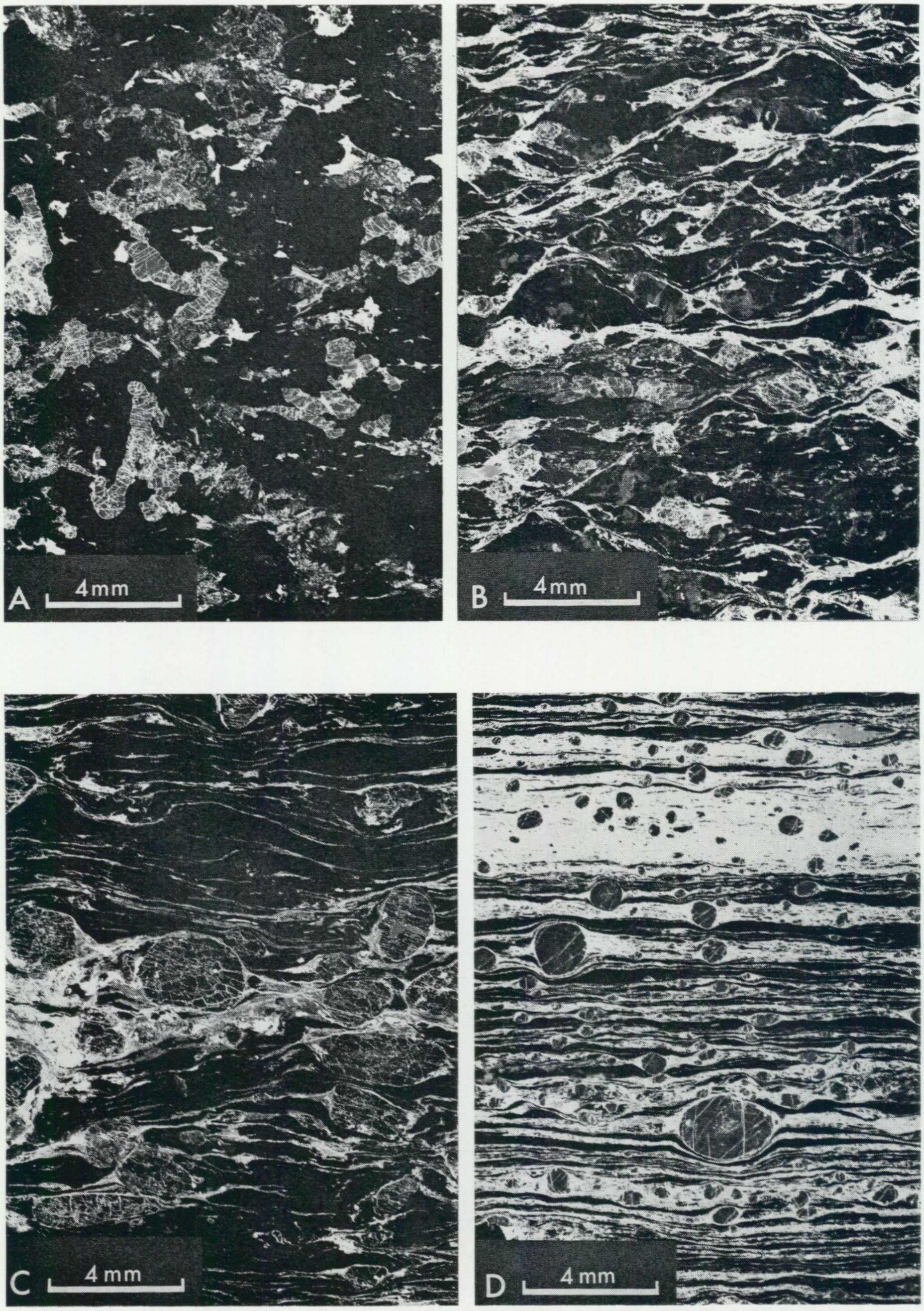


Fig. IV-6, 7. Development of S_2 in gneiss. A. Slightly deformed gneiss, "protected" by amphibolite, Risfjället. B: Gneiss, deformed by F_2. Two composite S-planes have developed. C: F_2 mylonitic gneiss. D: Strongly deformed F_2 mylonite.

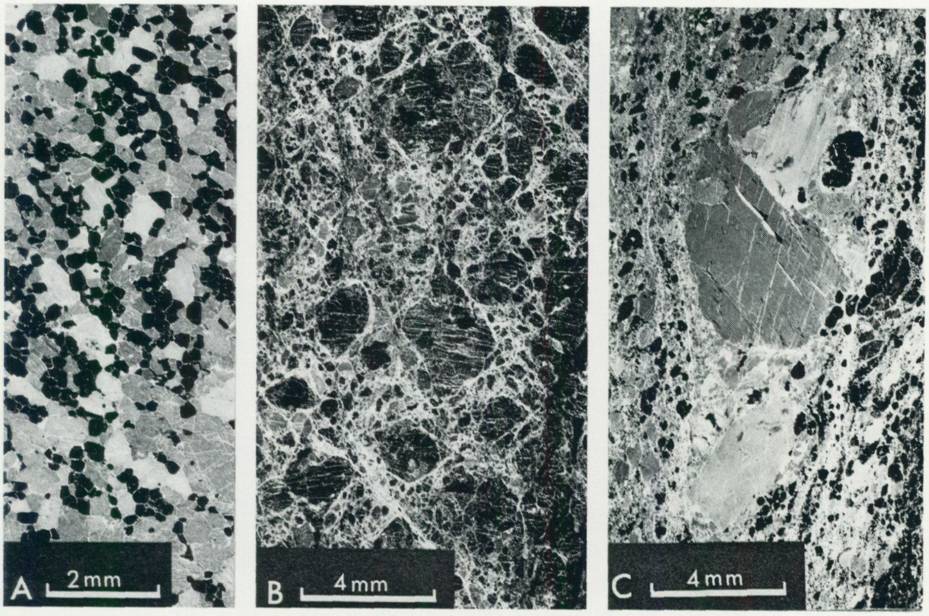


Fig. IV-8. Development of S_2 in amphibolite. A: Hardly deformed amphibolite. B: Deformed amphibolite with two composite S-planes. C: Mylonitic amphibolite.

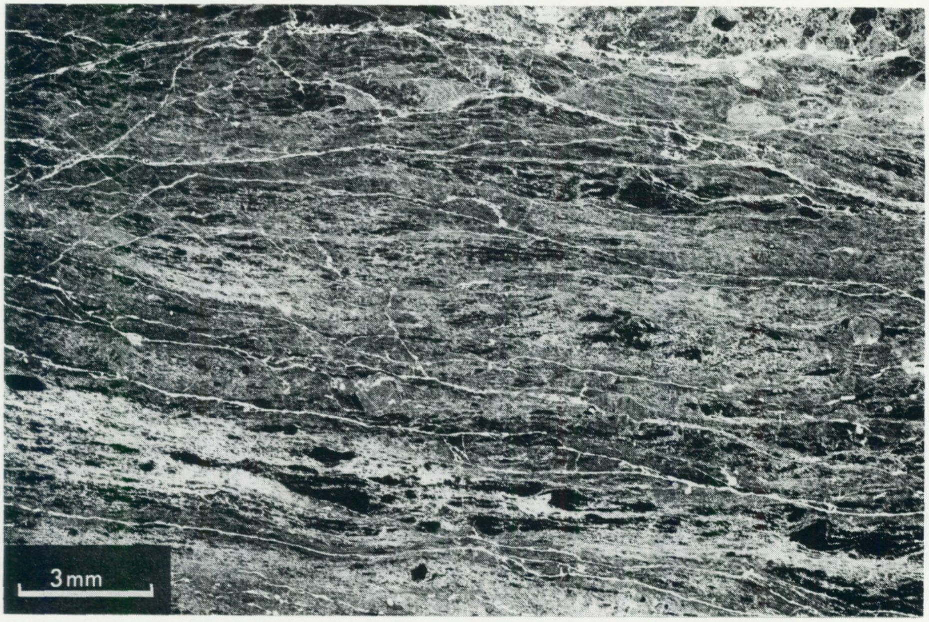


Fig. IV-9. F_3 mylonite in gneiss, Borkafjället. No visible recrystallization took place.

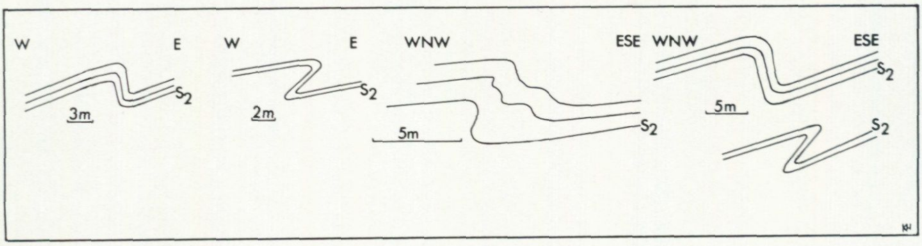


Fig. IV-10. Asymmetrical F_3 folds in blastomylonite zone, Risfjället.

DISCUSSION

Correlation of deformation phases

Comparing the tectonic history of the Marsfjället Gneiss and the Svartsjöbäcken Schists, there seems to be little doubt as to the correlation of the three deformation phases in both units. The S_2 planes grade into one another and the metamorphic circumstances during the phases fit reasonably well. The only deviation in this correlation is that the schists recrystallized fairly well during and after F_2, whereas the gneisses recrystallized to a much lesser extent in the same period (Glass, in prep.).

S_2 planes and mylonitization

Johnson (1967) stressed the meaning of the lamination in mylonites (in the present case S_2), and states the following:

> "it is apparently a plane of extension, as becomes directly clear from the shape of brecciated grains. They form lens shaped or ovoid aggregates, always elongated along the mylonite layering. If a lineation is present it is normally defined as the longest extension of these aggregates, streaks etc, in the layering itself."

Better indications of the orientation of the strain ellipsoid are hardly to be expected. The mylonitic layering is therefore interpreted as being parallel to the XY plane, the lineation to the X direction of strain. As Johnson states and as follows from the fact that (blasto)-mylonite is a descriptive term, the apparently strong deformation could have been either rotational or irrotational. One may imagine that in a succession of different strata certain layers with specific mechanical properties were deformed much more than others. These layers may show all the characteristics of mylonites, but they do not represent any thrust, fault or other strong rotational deformation.

Mylonite zones are, however, often associated with thrusts or faults. A rotational strain in these zones seems to be the more probable, but in places a more complex relationship between the mylonites and the thrusting was demonstrated.

Christie (1963), for example, distinguished between primary mylonites (blasto-mylonites) and secondary mylonites in the Moine thrust. The question which of these has to be correlated with the main displacements is still under discussion (Johnson 1965; Christie 1965). The joint occurrence of mylonites and thrusts is apparently in itself no proof of contemporaneous formation.

Good indicators of rotational deformation in mylonite zones are rotated garnets (if Ω_i >45°, see Chapter VIII), if it can be demonstrated that they grew during mylonitization. Asymmetrical folds are dangerous in this respect since they do not necessarily indicate a rotational deformation on a larger scale. The blasto-mylonite zone in the Marsfjället Gneiss separates high grade gneisses from low grade schists. There appears to be little doubt that the gneisses are thrust over the low grade rocks. No secondary mylonites, as, for instance, described by Christie (1963) from the Moine thrust, are found along well-exposed sections. It follows that the thrusting probably predates the low grade deformation of F_3 (Zachrisson 1969). On the other hand, the thrusting was later than the climax of metamorphism, as appears from the metamorphic jump. This climax was probably reached shortly after F_1 (Glass, in prep.). The most logical conclusion is that important thrusting took place during F_2 and that the blastomylonites were formed during thrusting. In other words: the deformation responsible for the blastomylonites was a rotational strain. Rotational strain can be regarded as simple shear accompanied by pure shear. A model of homogeneous simple shear is therefore used to illustrate the circumstances under which S_2 was formed (Fig. IV-11).

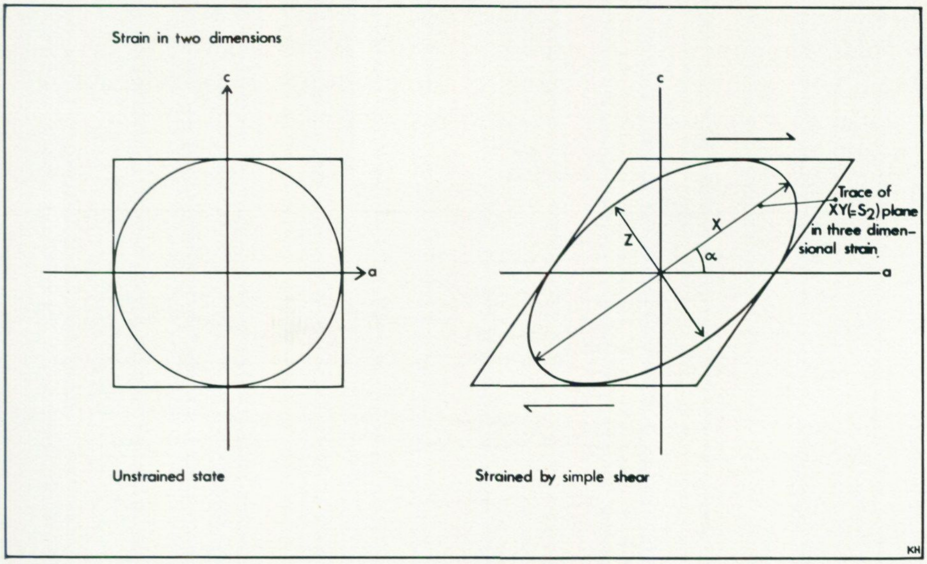


Fig. IV-11. Diagram illustrating simple shear in which S_2 is believed to be parallel to the XY plane.

It appears that S_2, believed to be parallel to the XY plane of finite strain, forms an angle α with the shear plane *ab,* depending on the amount of shear γ (Ramsay and Graham 1970). In thrust zones, where γ is high, α is expected to be close to zero.

A conspicuous feature of mylonitic zones is often the heterogeneity of deformation, reflected by thin, highly mylonitic layers, alternating with less mylonitic ones. Porphyroclasts, less deformed than their surroundings, provide a heterogeneity on a smaller scale. Sometimes this heterogeneity in the amount of shear is reflected in differences in α in different domains (Fig. IV-12). In the relatively less deformed domains α is larger.

From such structures the shear direction can be derived. In addition small asymmetrical folds might indicate the sense of shear as well, but, as stated before, these folds alone are not sufficient. In a number of cases "fish-shaped" muscovite porphyroclasts show the same relationship between their basal cleavage and the surrounding S-plane (Fig. IV-13). They are interpreted in the same way and also used to determine the sense of shear. Eisbacher (1971) used these markers for the same purpose, but he did not give an explanation of the origin. The results are given, together with those from the eastern schists, in Chapter V (p. 80).

From the model of simple shear it can be understood that asymmetries in a texture become weaker with increasing deformations, since the "flattening" perpendicular to S, or XY, is very strong for high γ values (see Chapter VIII). In such cases the monoclinicity of the fabric becomes less pronounced. The fabric approaches an orthorhombic one and this may be the reason why many mylonites possess nearly orthorhombic fabrics.

Folds: Basic concept is that folds, especially in higher metamorphic rocks which were relatively homogeneous during deformation, may represent only a minor part of the total strain and that the orientation of the fold axes is

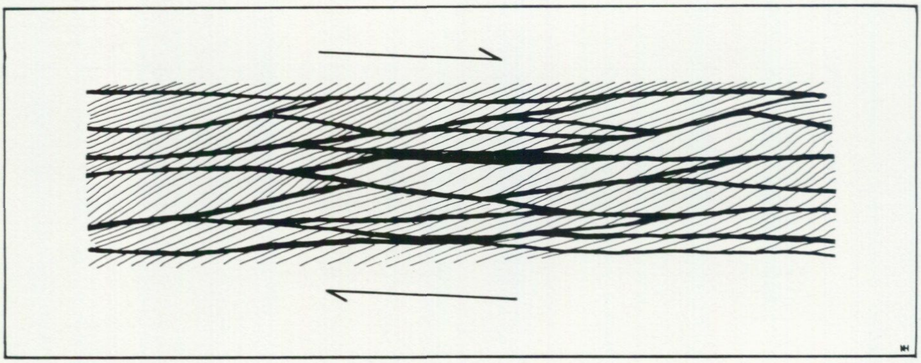


Fig. IV-12. Diagram showing that the difference in deformation of various zones is reflected by a variation of the angle α between the S-plane and the shearplane.

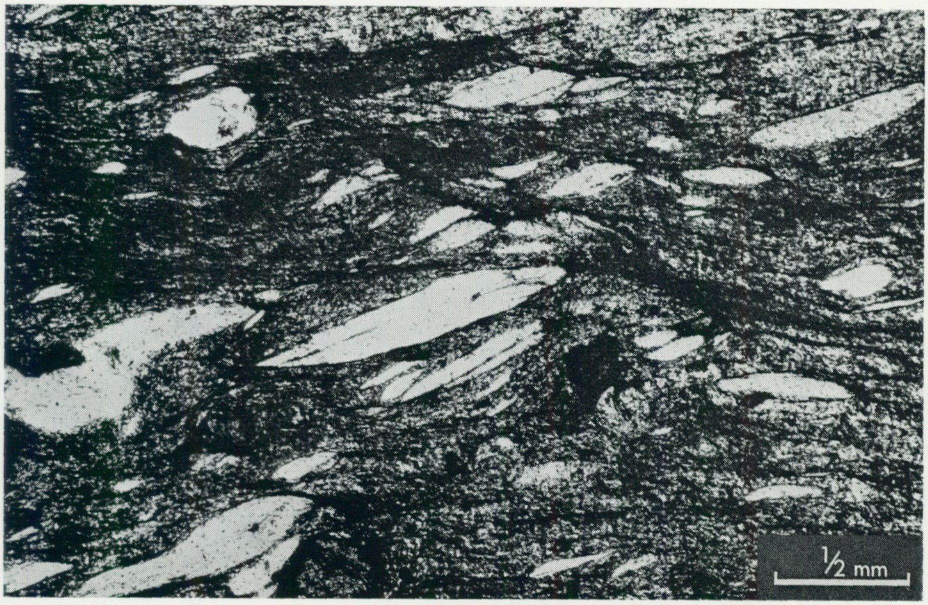


Fig. IV-13. "Fish-shaped" muscovite crystals; these are used as indicators for the sense of shear.

not necessarily related to a principal direction of strain. During a deformation, responsible for mylonites, folds may be formed of older layering or cleavage, or of the mylonitic layering itself. The latter may, for instance, occur if, under the influence of local heterogeneities (e.g. porphyroclasts), the mylonitic plane is rotated out of the XY position and becomes shortened. This is always a local effect which might be reflected in the so-called intrafolial folds. As soon as folds have formed they will be passively rotated during progessive deformation until their axis is close to the X direction of strain and hence to the mineral lineation. Especially in the strongly deformed mylonites this process, described by Flinn (1962), seems to provide a valid explanation for the fact that most folds in such zones have their axis oriented parallel to the mineral lineation (Christie 1963; Kvale 1953; Lindström 1955). However, in several mylonitic belts, also outside the present area, it is observed that such folds have their axis *exactly* parallel to the mineral lineation. It seems strange that folds are rarely detected "on their way", in particular since a considerable amount of strain is required for a near complete rotation and because the folds originate only during the deformation and thus are rotated only during the remaining deformation.

The author wishes to put forward another mechanism, based on heterogeneous deformation. It is very common for the strain to be heterogeneous in a plane perpendicular to S, i.e. that the amount of simple shear in different layers is variable, since this is at least partly the reason why the mylonitic layering exists at all.

It seems logical that the amount of pure shear within these various layers is also variable, meaning that certain layers are more flattened than others. This will have little effect in the shear direction (*a*), since the layers are extended in this direction as it is, due to the simple shear, but in the Y direction (*b*) local extension and constriction would be the result, possibly producing intrafolial folds with an axis parallel to X.

CONCLUSIONS

1. The gneisses are paragneisses, principally derived from felsic rocks (rounded zircons, presence of aluminium silicates, local layering).
2. The metabasites probably derived both from intrusive and extrusive basic rocks. The extrusives are syn-sedimentary, the intrusives are pre or syn-F_1 since they have the same tectonic and metamorphic history as the gneisses. A few basic dikes are younger since they crosscut the S_1 segregation layering. They are, however, metamorphosed and deformed so that their intrusion probably took place in the interkinematic stage between F_1 and F_2.
3. During a first recognized phase of deformation, F_1, an S fabric and metamorphic segregation layering were formed.
4. The F_1 fabric is in most places obscured by recrystallization (Glass, in prep.).
5. During F_2 intense postcrystalline deformation took place throughout the formation, leaving only thick amphibolites relatively unaffected. The gneisses were thrust in easterly direction over the eastern schists and a blastomylonite zone was formed at the contact. The mineral lineation L_{2min} is interpreted as the X direction of F_2 strain, which is close to the direction of simple shear, *a*. The metamorphic circumstances during F_2 were still within the amphibolite facies (Glass, in prep.).
6. During a later deformation phase, F_3, minor folding and thrusting took place locally, under low grade conditions. The orientation of F_3 structures again indicates an E-W shortening.

CHAPTER V

The eastern schist and amphibolite belt

INTRODUCTION

This belt is not given the status of a formal lithostratigraphic unit for the following reasons:

1. There is a considerable variety of rock types with little continuity either to the north or to the south.
2. The lack of exposure in relatively large areas made the determination of boundaries between different units impossible.

The belt is bounded on either side by tectonic contacts. The lower boundary is the thrust contact of the metamorphosed Seve nappe on the very low grade Eocambrian quartzites and sparagmites. The upper boundary is the blastomylonite zone described in Chapter IV.

The belt is subdivided into schists and metabasites; more or less isolated exposures in the eastern part are dealt with separately.

SCHISTS

General aspects

The most common type of schist is:

1. Schist rich in muscovite and plagioclase. Conspicuous, medium to coarse-grained muscovite flakes and locally developed plagioclase "augen" are characteristic of this schist.

Other types of schists are:

2. Hornblende biotite schist (well exposed between Vallentjälte and Saksenvardo).
3. Quartz feldspar schist, often fine-grained. This type of schist is often associated with amphibolite layers.
4. Carbonate-rich schist (exposed in the river Vojmån).

Minor amphibolite layers, lenses and boudins occur abundantly in all schists. An alternation of types 1 to 3 is well exposed on Offerkullen. Original bedding is in general no longer visible, it is only preserved in the carbonate-rich schist.

The schistosity is rather weak, and cannot be compared with the characteristic schistosity of the Svartsjöbäcken Schists.

Folds are common, particularly between Borkafjället and Saksenvardo and on Offerkullen. It is not merely bedding, but rather an old schistosity which is folded into coarse crenulations. Only quartz veins produce nice fold patterns on a small scale.

Part of this unit is described by Michel as "the Eastern-garnet-mica-schist-series". He states that the schists contain much less garnet and on the whole less mica than the rocks of "the Western-garnet-mica-schist-series" (the Svarsjöbäcken Schists), and that the plagioclase content seems in general to be larger (Michel 1950, p. 97). These statements are confirmed by the present investigation, apart from the comparison of the mica contents. The scarcity of biotite in the eastern schists is obviously compensated for by the abundance of muscovite.

Petrography

1. In the schists rich in muscovite and plagioclase numerous muscovite grains have a "fish-shaped", undulose appearance (Fig. IV-13). The porphyroblastic plagioclase, with an oligoclase composition, may constitute up to 50 % of the rock. Some grains have myrmekitic textures, others are full of musco-

vite inclusions. Coarse deformed grains (old) are locally surrounded by small strain-free grains (new). Quartz also occurs in old and new grains; the old grains are rare, the new grains locally form polygonal aggregates. Coarse garnets have many fine-grained inclusions, sometimes arranged in planar patterns. The rims, which are free of inclusions, probably grew contemporaneously with small idiomorphic crystals. The scarce biotite is predominantly fine-grained. Epidote group minerals occur abundantly in many rocks as fine, often zoned grains; in other rocks they are hardly found. The carbonate proportion is variable too. East of Gakkangaise a garnet, kyanite, staurolite schist, lacking any plagioclase, is exposed. Kyanite occurs in several other rocks, e.g. a kyanite, plagioclase gneiss, very rich in plagioclase, on the eastern slope of Mount Risfjället. Some staurolite with muscovite coronas was detected on Offerkullen.

Primary and secondary chlorite occurs locally in minor quantities. Accessory minerals are apatite, sulphides, rutile and zircon.

2. In the hornblende biotite schists, medium-grained biotite, as well as plagioclase, is very common. All transitions to biotite-bearing amphibolites occur. West of Saksenvardo, some of these schists contain kyanite. The coexistence of hornblende and kyanite seems to be an equilibrium association.
3. The quartz-feldspar schists are in general fine to very fine-grained. The best exposures are on Offerkullen, in Rissjöbäcken, and on Henriksfjället. They locally contain medium to coarse plagioclase crystals, which might represent old phenocrysts. The composition and the interlayering and association with metabasic rocks suggest a volcanic origin (keratophyres) of these rocks.

Structures

Microscopic analysis revealed that, with few exceptions, all the eastern schists suffered a strong postcrystalline deformation, leaving the muscovites as contorted undulose "fish" and the plagioclases as porphyroclasts in a fine to very fine-grained matrix of recrystallized quartz, fragments of biotite, plagioclase and other minerals (Fig. V-2).

The few rocks which apparently escaped or almost escaped this deformation show a medium to coarse-grained gneissose texture, with more or less randomly oriented muscovite flakes and large, often undulose old quartz grains (Fig. V-1). The medium to coarse-grained state of these rocks and the more or less random fabric explain why the folds, formed during the postcrystalline deformation, are so badly outlined. There was simply no good planar feature to be folded (Fig. V-2).

The early history of the schists can be deduced from the inclusions in garnets. Some garnets have straight or almost straight patterns of small inclusions, indicating a late or post-F_1 growth over a fine slaty cleavage, S_1. Apparently the

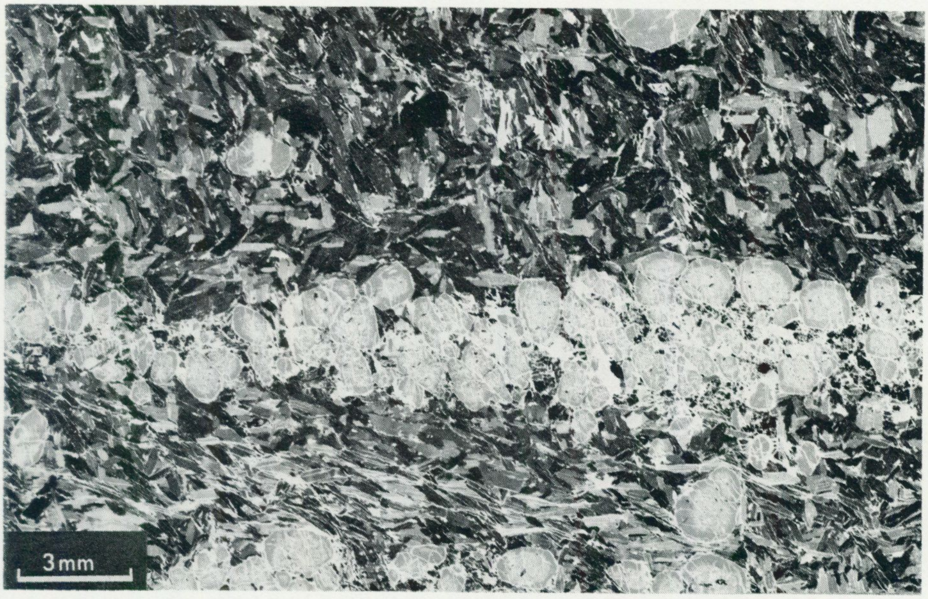


Fig. V-1. Garnet mica-schist with more or less random muscovite fabric. Little affected by post-crystalline deformation (F_2). Eastern slope of Gakkangaise (negative print).

Fig. V-2. Garnet mica-schist, intensely deformed by F_2. The F_2 folds are badly outlined since there was no good planar feature to be folded. Girifjället (negative print).

grain size only coarsened after F_1, giving rise to a badly oriented fabric. The postcrystalline deformation, F_2, postdated the coarsening and reduced the grain size of quartz and other minerals, as in the Marsfjället Gneiss. F_2 deformation was at any rate not homogeneous on a large scale, since all varieties occur, from non-deformed gneisses to blastomylonites. As in other units, F_2 folds are characterized by recrystallized quartz in the hinges. Most of their axes plunge to the W or NW (see, e.g., Enclosure II, sub-areas 24 and 26).

In accordance with the position of their axial planes, F_2 folds can be subdivided into two types:

1. Gentle or open folds with subvertical axial planes, dipping NE or SW. Examples of this type are exposed on Offerkullen.
2. Tight to isoclinal folds with axial planes dipping 30°–40° NW.

No evidence of deformation later than F_2 has been found in the schists.

METABASIC ROCKS

Field aspects

Within the eastern schist and amphibolite belt metabasites occur abundantly. They form large elongated bodies and/or continuous layers with variable thicknesses (Girifjället, Såttan, etc). Apart from these masses, numerous amphibolite boudins and/or small separated bodies and layers occur in the schists. Transitions between schists and amphibolites are found as well. Many amphibolites show a conspicuous layering, even in the middle of thick bodies (Fig. V-3). Layers rich in quartz-feldspar, of variable thickness (0.5 mm – several metres), alternate in an irregular manner with hornblende-rich layers. The asymmetry of these layers, an abrupt change on one side often being accompanied by a gradational one on the other, and their variable thickness, make a sedimentary origin more probable then metamorphic segregation. Other field aspects are the generally medium grain size, the common occurrence of garnet and the local presence of carbonate-rich lenses (Girifjället).

Petrography

In the amphibolites on Girifjället the following minerals occur abundantly: bluish green hornblende, plagioclase, epidote, quartz, garnet, carbonate, sulphide. Biotite, chlorite and some muscovite occur in minor proportions. Accessory minerals are: sphene, rutile and apatite.

The hornblende crystals often show irregular grain boundaries as a result of postcrystalline deformation. Many plagioclase crystals are strongly zoned (Glass, in prep.) and deformed. Fine to medium-grained undulose old crystals are surrounded by finer-grained new ones as a result of recrystallization. Quartz shows the same relationships; here the new grains are in general predominant. Epidote group minerals are rich in iron, but zoisitic varieties occur as well. The grains are chiefly equigranular; some are slightly elongated in the

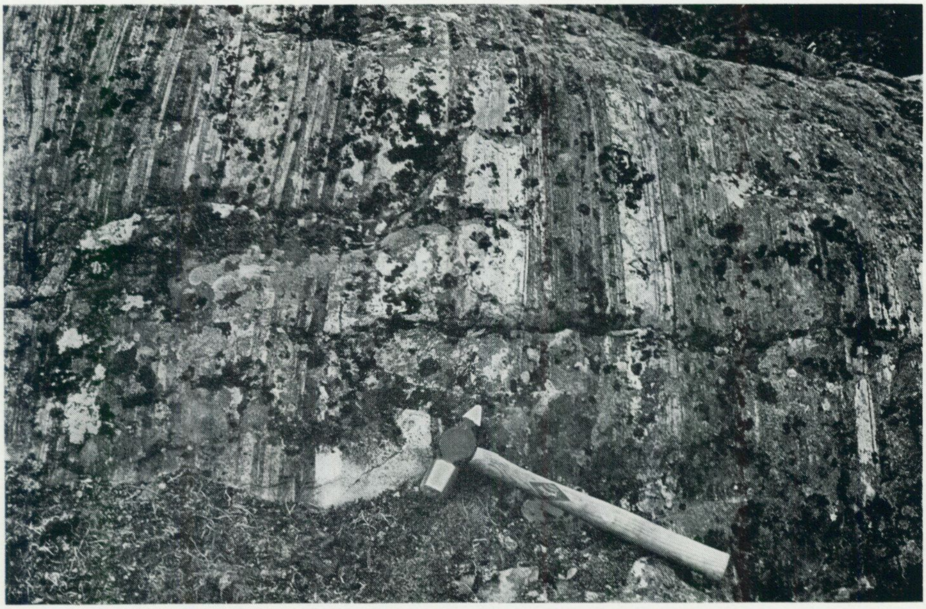


Fig. V-3. Well layered rock, consisting of amphibolite and quartz-feldspar-rich layers. Offerkullen.

schistosity. Some garnets contain S-shaped inclusion patterns; in many others the abundant quartz and carbonate inclusions are irregularly arranged.

Chlorite is probably secondary.

On Vallegietje the epidote group mineral is a grey zoisite. It occurs abundantly unlike garnet which is only found in one specimen. Carbonate does not occur here at all; in one locality some diopside is found.

The Såttan and Marsliden amphibolites are petrographically comparable to the one on Girifjället. Only garnet and carbonate occur less abundantly.

Structures

As in the amphibolites from the Svartsjöbäcken Schists, a plano-linear hornblende fabric is usually present. The folds are again subdivided into three groups:

1. The compositional layers or quartz veins are folded. The hornblende fabric is parallel to the axial plane. The fold axes are parallel to the mineral lineation (Fig. V-4).
2. The hornblende fabric is folded. Quartz and sometimes plagioclase is recrystallized. Most of the fold axes are parallel to the mineral lineation.
3. The hornblende fabric is folded. Hornblende crystals are broken, quartz and plagioclase crystals have undulose extinctions. The fold axes are not parallel to the mineral lineation.

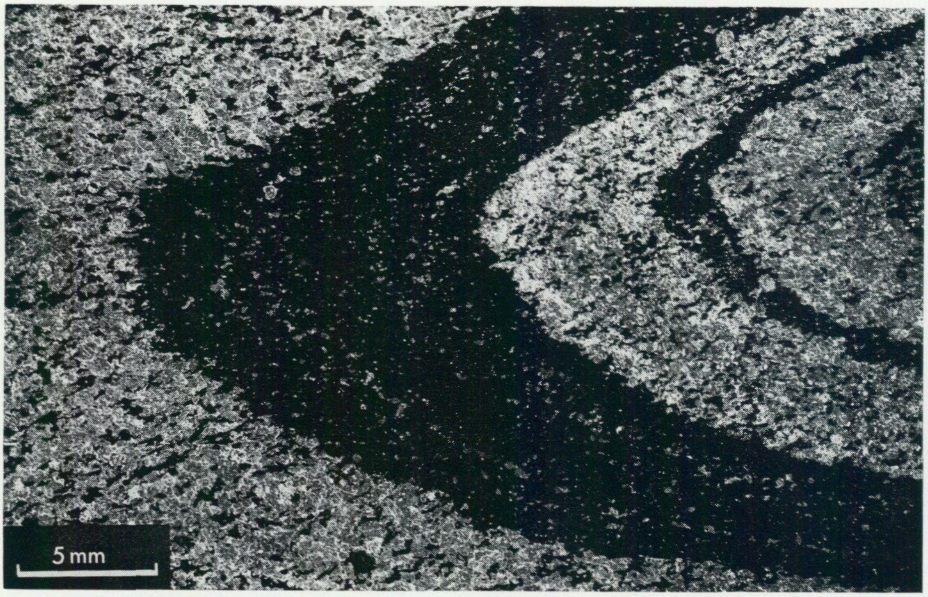


Fig. V-4. F_1 fold (group 1) in amphibolite from the eastern schist and amphibolite belt. Vinevare, south of the present area (negative print).

Only four folds of group 1 are detected. One is very tight to isoclinal, the other three are open to close. They are ascribed to F_1, which is also responsible for a penetrative plano-linear hornblende fabric (S_1 L_1).

A second, postcrystalline deformation, F_2, formed folds of group 2 or modified the S_1L_1 fabric to an S_{1+2} L_{1+2} fabric. Quartz and locally plagioclase recrystallized, but no recrystallization of hornblende could be demonstrated. In places mylonitic or blastomylonitic layers were formed. Particularly thin intercalated schist layers between thick amphibolite masses (e.g. Girifjället) show intense F_2 deformation (Fig. V-2). In the diagrams of sub-area 25 (Enclosure II, Girifjället) one can see that the F_2 fold axes and the mineral lineations roughly coincide. Both diagrams show a maximum to the NW and a sub-maximum to the NNE. A comparable picture appears from the diagrams of Såttan (sub-area 22); here the F_2 fold axes coincide almost uniformly with the NE–SW sub-maximum of L_{1+2}. The meaning of these two directions will be discussed below.

F_3 only very locally (e.g. Vallegietje) produced some open folds. It seems a very unimportant phase in this part of the area.

There is little difficulty in correlating these three deformation phases with those in the schists. The only difference is that no traces of F_3 are found in the schists.

ROCKS FROM SCATTERED EXPOSURES IN THE EASTERN SCHIST AND AMPHIBOLITE BELT

The Seve rocks which occur east of the thick amphibolite units within the eastern schist and amphibolite belt are poorly exposed. They will be discussed according to the locality of outcrop from north to south.

The Dikanäs Schists

East of Henriksfjället garnet mica-schists, metabasites and quartz-feldspar schists are exposed. They lie on top of the lower grade sparagmites and quartzites, separated by a mylonite zone. Calon (pers. comm.) called these rocks the Dikanäs Schists (see also Kulling 1955).

Calon subdivided the formation, from bottom to top, into three members:

1) Rocks which are interpreted as meta-volcanites.
   These rocks chiefly consist of fine to very fine-grained quartz-feldspar schists with a variable content of muscovite, biotite and epidote. Medium to coarse-grained plagioclase crystals embedded in the matrix may represent former phenocrysts. A biotite-rich amphibolitic rock is supposed to represent a more basic variety in these meta-volcanites.
2) Biotite, garnet schists, which resemble the schists from the Fatmomakk Formation; medium-grained biotite and garnet porphyroblasts are embedded in a fine to very fine-grained muscovite-rich matrix. No cataclastic textures indicate a former coarser crystallinity, as in the other eastern schists. The schists contain a considerable quantity of primary and secondary chlorite. Carbonate is often concentrated in veins and fissures. The low percentage of quartz would not allow a comparison to the schists from the Fatmomakk Formation. Garnet apparently grew in two generations, inclusion-rich cores of large crystals are often surrounded by idiomorphic, inclusion-free rims. Small idiomorphic crystals belonging to the second generation locally occur in abundance. The large garnets sometimes have numerous tiny graphite inclusions. Ilmenite porphyroblasts occur locally.
3) Metabasites showing quite a variety of mineralogical content.
   Apart from biotite-bearing layers, the often well-layered amphibolitic rocks contain garnet-biotite, plagioclase-epidote (or zoisite) and/or carbonate-rich layers. These various layers range in thickness from less than one cm to several dm. Chlorite is a common constituent. The rocks of this member are also interpreted as meta-volcanites. The layering is thought to represent original bedding.

The original stratigraphic position of these members is uncertain.

Deformation in the Dikanäs Schists. The pronounced cleavage is locally a crenulation cleavage, which is curved around garnet and biotite porphyroblasts. The biotites, as the muscovites in the schists previously described, are undulose "fish" in an intensely deformed matrix, in which the quartz did not recrystallize, as distinct from the other schists. Only locally did some recovery of strongly deformed quartz grains take place. The postcrystalline deformation phase responsible for the crenulation cleavage is obviously F_2.

The only evidence of F_1 are the crenulated cleavage and parallel inclusions in garnet crystals (S_1). Some garnets show beautiful S-shaped inclusions, indicating a para-crystalline rotation of up to 450°. S_i is completely detached from S_e (Figs. V-5, 6). According to the argumentation in Chapter VIII, these garnets must have grown during F_1. This means that during F_1 shear movements in the order of $\gamma = 15$ ($\Omega_e = |{}^1/_2\, \gamma|$) took place locally. Since the rocks were heterogeneously de-

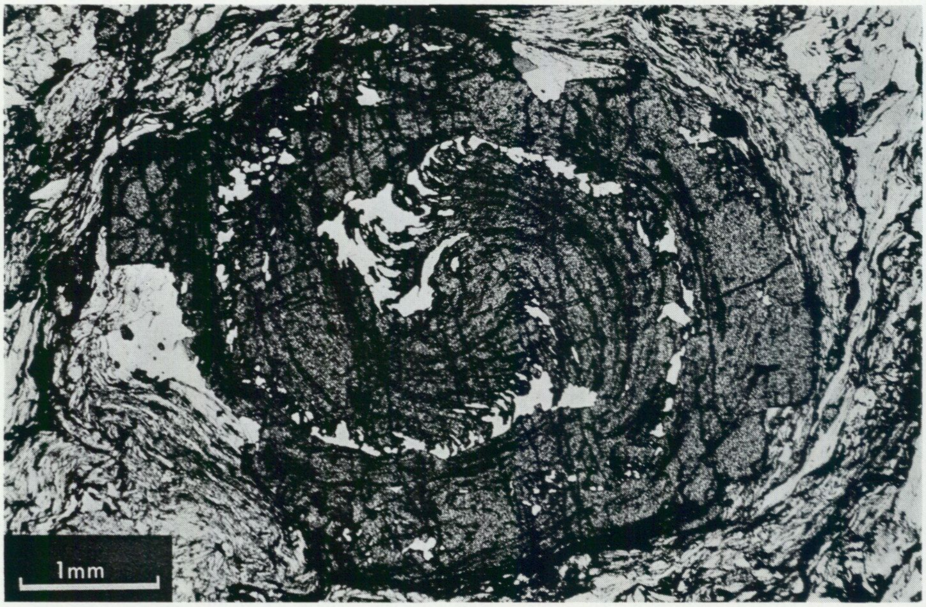


Fig. V-5. Syn-F_1 garnet from the Dikanäs Schists. Ω_c is about 450°. S_i is completely detached from S_e (S_2). Thin section kindly lent by T. Calon. Bergsjöån, near Stennäs.

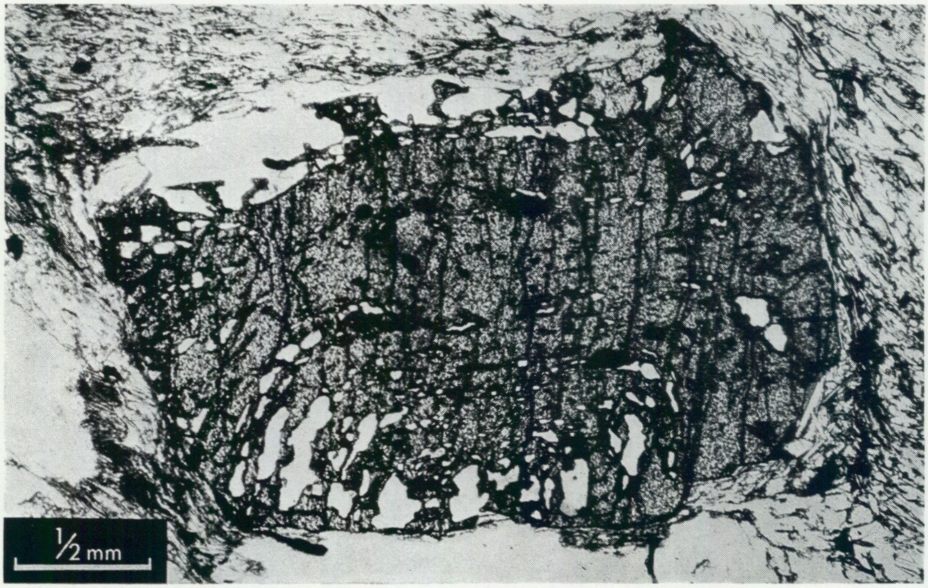


Fig. V-6.)(pattern in rotated syn-F_1 garnet from the Dikanäs Schists. Section parallel to garnet rotation axis, which is oblique to S_2. Along the road to Dikanäs, east of the present area.

formed during F_2, the orientation of the rotation axes is now irregular and the sense of shear during F_1 could not be determined. Some specimens hardly affected by F_2 show that the orientation of the pre-existing minerals was close to random, others show a well-defined F_1 fabric (S_1), generally parallel to bedding.

A distinct F_3 deformation is evident in several thin sections. Either the S_2 plane is gently folded or some irregular fractures, often combined with local crenulations, crosscut the S_2 planes.

Krutberget

The Krutberg forms a "half klippe" of Seve rocks, resting on the Eocambrian quartzites and sparagmites.

A section of this mountain, from top to bottom is as follows:

1) Ultramafic bodies, accompanied by a metagabbro. They are hardly deformed. The clinopyroxene in the metagabbro has rims of hornblende.
2) Mylonitized diabase porphyry, several metres thick, with plagioclase-epidote patches, probably representing former plagioclase phenocrysts.
3) Granite gneiss, about 20 m thick, with a strong linear fabric (L=288/10).
4) Amphibolitic mylonite, approx. 25 m thick, with irregular, detached folds of quartz-rich layers. The quartz in the fold hinges is recrystallized.
5) Quartzite, belonging to the Eocambrian quartzites and sparagmites.

Girifjället and Rutetje

North on Girifjället, just below the thick amphibolite, an exposure of mylonitized, microcline-bearing mica-schist outcrops. This meta-arkose may be an equivalent of the Meta-arkose Formation, as described by v.d. Harst (1956) in the Borga valley and by Brandt (pers. comm.) near Saxnäs.

On Rutetje an approx. 10 m thick layer of granite gneiss is exposed, intercalated in amphibolites; it is composed of microcline, plagioclase, quartz and some dark green to brown hornblende. The layer is strongly deformed by postcrystalline deformation (F_2).

Vallegietje

On Vallegietje a layer of granite gneiss, several metres in thickness, with about the same composition as on Rutetje, occurs below a thick amphibolite. It rests upon plagioclase-rich schists, which alternate with metabasic rocks. The schists contain biotite, musovite, epidote, garnet and, in one specimen, microcline. Some of the metabasites show conspicuous patches of plagioclase and zoisite, which probably represent altered plagioclase phenocrysts. The deformation is comparable to the one in the other eastern schists and amphibolites.

Garsbäcken

In Garsbäcken, directly below a huge ultramafic body, an approx. 10 metres thick zone containing several rock types outcrops. These rock types are:

1. Very fine-grained carbonate-rich phyllite, with biotite.
2. Fine-grained muscovite-garnet schist.
3. Mylonitized garnet-bearing amphibolite.

In the carbonate-rich phyllites isoclinal, asymmetrical folds have developed, with NW dipping axes and a vergence to the SW (Fig. V-7). The axial plane cleavage of these folds, which is parallel to the general cleavage in the surrounding rocks, is a

Fig. V-7. F_2 fold in carbonate-rich layer, exposed in Garsbäcken (negative print).

crenulation cleavage. This makes a correlation of these folds with F_2 in the eastern belt very probable. A weak overprinted crenulation is ascribed to F_3. In the fine-grained muscovite-garnet schists, S_2 is crenulated by F_3 as well. Underneath this zone, amphibolites with variable F_2 deformation are exposed. In the lower part of these amphibolites ther ocks did hardly suffer any postcrystalline deformation. Fresh quartz and feldspar grains show approximately straight polygonal grain boundaries. Apart from hornblende, albite, biotite and garnet, these rocks contain carbonate, epidote group minerals, quartz, sphene and apatite. The F_1 fabric present in these rocks forms both a clear S-plane (S_1) and a strong mineral lineation ($L_{1\,min}$) in NW direction (294/8).

Below this unit hardly deformed microcline-bearing granites, with dark green amphiboles, resemble the described granite gneisses on Vallegietje and Rutetje.

The Grytsjö Schists

Near Grytsjö some carbonate-bearing garnet-mica schists are exposed. The biotite content of these schists rich in muscovite and plagioclase is somewhat higher than in most of the schists previously described. However, biotite forms no porphyroblasts, as in the Dikanäs Schists.

Two of the three specimens examined show an undulating S_2 plane with "fish-shaped" micas, well recrystallized quartz and small pre-F_2 garnets. The irregularity of the undulations and the fresh quartz grains indicate that they were not formed by a later deformation, but that S_2 was formed in this non-planar manner.

The third specimen is quite different. Fresh micas and large undulose quartz grains indicate a weak F_2 deformation, without recrystallization. It is interesting to note that the hardly deformed S_1 fabric bends slightly around the garnet porphyroblasts, proving a syn-F_1 growth of this mineral. The garnets, though rich in graphite inclusions (like in the Dikanäs Schists), unfortunately show no regular S_i patterns.

Below these rocks, just overlying the Eocambrian quartzites and sparagmites, a strongly sheared granite occurs, with large microcline porphyroclasts. It contains some green-dark green hornblende, plagioclase, quartz and secondary carbonate.

EOCAMBRIAN QUARTZITES AND SPARAGMITES, UNDERLYING THE EASTERN SCHIST AND AMPHIBOLITE BELT

General aspects

These rocks are more or less beyond the scope of this study; no mapping of different rock units was carried out, but a general reconnaissance proved the following.

Three main rock types occur:

1. Quartzites (Ström Quartzite, Kulling 1955).
   These are chiefly pure, coarse-grained quartzites, sometimes containing a minor proportion of carbonate and feldspar.
   As the percentage of feldspar increases they grade into arkoses.
2. Arkoses. A variable amount of plagioclase and microcline accompanies the quartz in these rocks. They are fine to coarse-grained. Carbonate occurs locally. As the proportion of fine-grained matrix increases they grade into slates.
3. Slates. These are very fine-grained slates with a variable amount of coarser quartz, plagioclase or epidote grains. Chlorite and carbonate may be present.

Structures

In most of the rocks a slaty cleavage, S_1, has developed. In the slates it is defined by parallel mica flakes, in the more quartzitic and arkosic rocks by flat, deformed quartz grains.

F_1, responsible for this slaty cleavage, was not very homogeneous on a large scale, since some rocks have a much better developed S_1 plane than others. Folds of the bedding, with S_1 as axial plane cleavage, occur locally and are interpreted as F_1 folds.

In general S_1 is at a small angle to SS; in some cases it is parallel. A later deformation, F_2, produced irregular crenulations, (micro) faults and folds of S_1 (Fig. V-8). In the more quartzitic parts this deformation was not penetrative, as it caused only small faults. In finer-grained material the faults often grade into crenulated zones. Most F_2 folds are open folds, some are tight (e.g. near the dam E of Saxnäs). They are asymmetrical folds with NE-SW axes and with a vergence to the SE.

DISCUSSION

Comparison of deformation phases

It may be concluded from the foregoing that three deformation phases, F_1, F_2 and F_3, were active within the eastern belt. These three phases and the grade of metamorphism at which they took place are very similar to those in the

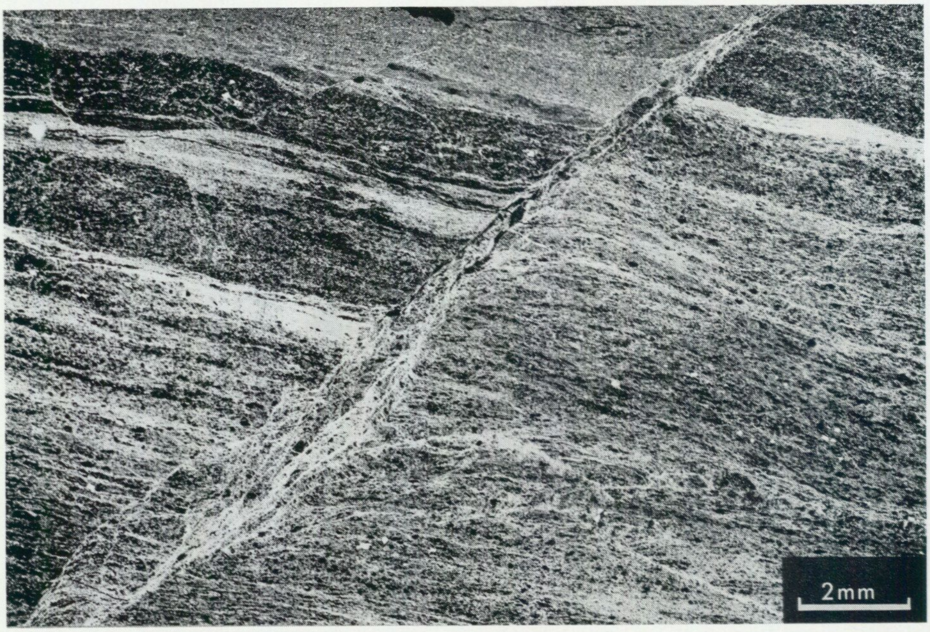


Fig. V-8. F_2 microfault in a slate from the Eocambrian sparagmites. F_2 in these rocks corresponds to F_3 in the Seve rocks. Koksikkammen (negative print).

Marsfjället Gneiss. However, the correlation of F_1 in the Marsfjället Gneiss with F_1 in the eastern belt is uncertain since the units were only brought together during F_2.

The Eocambrian quartzites and sparagmites show evidence only of two deformation phases. Starting at the end of the history, which both Seve and underlying rocks probably had in common, it appears that F_2 in the quartzites and sparagmites corresponds fairly well to F_3 in the Seve. The same irregular crenulations, grading locally into small faults, occur in both units. It makes sense to correlate the important F_2 in the Seve with the main phase in the quartzites and sparagmites F_1.

The mylonite zone, separating the Seve rocks from the underlying ones was probably formed during F_2 (Seve), as it is not significantly different to all the other F_2 mylonites, producing a gap in metamorphic assemblages and being folded by F_3. This indicates that probably a substantial part of the overthrusting of the Seve over the underlying Eocambrian rocks took place during F_2.

Folding mechanisms

During F_2 the overthrusting of the Marsfjället Gneiss over the eastern belt, and of the entire Seve nappe over the Eocambrian quartzites and sparagmites, from WNW to ESE probably took place. From 8 shear determinations out of

13, based on asymmetrical muscovite "fish" (see Chapter IV – Discussion), a corresponding sense of shear, clockwise looking north, within the eastern schists was inferred. Two indicate the reverse and three a direction NE-SW. Nevertheless, the conclusion seems justified that the general deformation during F_2 was a rotational strain, approximately represented by clockwise simple shear looking N-NE, accompanied by pure shear.

During F_2 three types of folds and mineral lineations in two directions were formed in the eastern belt (Fig. V-9).

1. Folds of type 1 are open or gentle, large scale folds with NW-SE (or E-W) axes, parallel to the mineral lineation. They have steep to vertical axial planes.
2. Folds of type 2 are mainly small-scale, tight to isoclinal "intrafolial" folds with NW-SE (or E-W) axes, parallel to the mineral lineation. They have flat-lying axial planes parallel to S_2.
3. Folds of type 3 are predominantly large-scale, open to close folds. They have NE-SW axes, parallel to a less pronounced mineral lineation, and NW dipping axial planes.

Folds of type 1 occur on Offerkullen and between Girifjället and Rutetje. They can be explained using two models developed by Cloos. Cloos (1946, pp. 26–29) ascribes these types of folds to two different parallel to *a*-models and defines *a* as the direction of principal movement. In this context it is important not to confuse movement direction with the X axis of the strain ellipsoid (Ramsay 1967, Fig. 6.45, 1969, p. 53; Ramsay calls the movement direction displacement or translation; see also Ramberg 1962). In the model of simple shear the X axis, after some deformation, forms a small angle α to the movement or shear direction *a*, within the *ac* plane.

The first of Cloos's models is the following. He states that:

"the large undulations which show on the maps in the sinuous arrangement of strata – – – are not necessarily folds accompanied by shortening, nor the approach of two points toward each other . . .

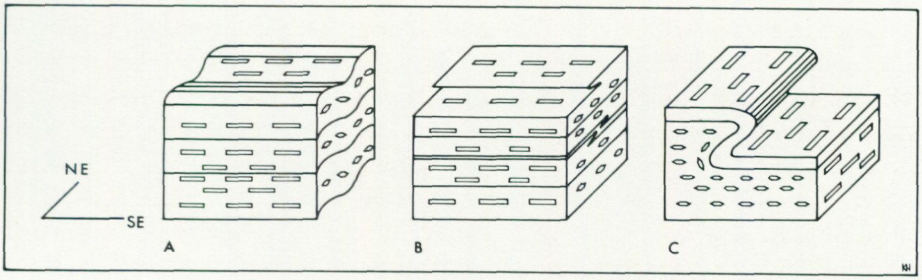


Fig. V-9. Three types of F_2 folds in the eastern schist and amphibolite belt. Hornblende crystals are drawn to illustrate the mineral lineation, but type 1 and type 2 folds occur in the schists as well. For further explanation see text.

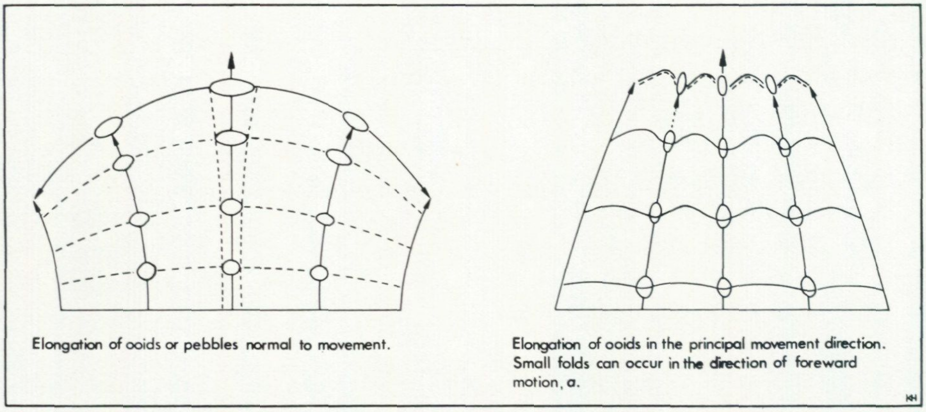


Fig. V-10. A possible explanation for folds with their axis parallel to *a*, after Cloos (1946).

Obviously, thrust planes will be highly uneven and these corrugation axes are not fold axes, but large scale scour marks and striation."

Although "scour marks and striation" suggest brittle deformation on single planes, which in the present case obviously did not take place, the general principle of undulating shear planes seems valid.

In addition Cloos provided a model of "converging movement" in which "small folds can occur in the direction of forward motion" (Cloos 1946).

In Fig. V-10 the elongation of ooids (X) is parallel to the "principal movement", *a*. This coincidence does not invalidate the principle that minor constrictions perpendicular to the shear direction can produce folds of type 1. The mineral lineation, indicating X, is in both models parallel or almost parallel (α) to the fold axes, which are parallel to *a*.

Folds of type 2 are discussed in Chapter IV (Discussion of S_2 planes and mylonitization).

Type 3 folds are the "normal" or "classical" asymmetrical folds, to be expected in a model of simple shear, their fold axes parallel to Y, and with a vergence in the shear direction. These folds occur on Såttan and Girifjället, in thick amphibolite layers in which the general shear stress was probably not released by passive internal deformation, but where the layers reacted in a more active way by buckling. The intense large-scale folding in these parts illustrates this. The divergent internal deformation in these folds is reflected in the orientation of a NE-SW mineral lineation parallel to the fold axes. This lineation is of local importance; as an indicator of the X direction of strain it is obviously not in accordance with the overall strain picture. It is hard to determine whether the mineral lineation in these rocks is at right angles to X or whether the local X direction of strain is at right angles to the general X direction. The results of van Zuuren's work (1969) are of interest in this respect. He came to the conclusion that:

> "a pronounced preferred orientation of hornblende *c*-axes parallel to the fold axis or lineation is a common feature. The occurrence of a sub-maximum, perpendicular to the fold axis in S, has often been observed. On the other hand, the reverse situation, i. e. a pronounced maximum in A and a poorly developed one in B, is a comparative rarity".

A and B are axes of fabric symmetry, apparently based on the fold axis (B). Although this system is based on fold axes and not on principal strain directions it does not appear abnormal that a sub-maximum of hornblende *c*-axes occurs, perpendicular to the main maximum. It may be possible that this sub-maximum, which is probably oriented in Y, predominates locally.

CONCLUSIONS

1. The eastern schist and amphibolite belt consits of schists, metabasites and orthogneisses. The schists are derived from pelitic sediments, the metabasites partly from intrusive and partly from extrusive rocks. The extrusives are locally intercalated with acid extrusive rocks. The orthogneisses, occurring near the lower contact of the belt, may very well be deformed basement gneisses.
2. During a first phase of deformation, F_1, a penetrative plano-linear fabric is formed. Locally strong shear movements (in the Dikanäs Schists γ is up to 12) occurred. Some schists, or phyllites at the time, were fine-grained (e.g. Dikanäs Schists), others (most of the eastern schists) coarsened towards the end of F_1.
3. Further coarsening after F_1 in the schists led to (almost) random fabrics in medium-grained schists and gneisses.
4. During F_2 strong shear movements, clockwise looking N-NE, took place, resulting in the large thrust of Seve rocks over the Eocambrian quartzites and sparagmites and of the Marsfjället Gneiss over the eastern schist and amphibolite belt. During this postcrystalline deformation the grain size in the medium-grained schists was reduced. In the amphibolites the F_1 fabric was folded or modified to an F_{1+2} fabric. Three types of folds and mineral lineations in two directions were formed. F_2 is equivalent to F_1 in the Eocambrian quartzites and sparagmites, where a penetrative planar fabric, S_1, was formed.
5. A third phase of deformation, F_3, took place under low grade conditions. The deformation is of minor importance. In the coarser-grained rocks no traces were detected, except for a few folds in amphibolites. In the finer-grained schists (e.g. Dikanäs Schists) undulations of S_2, fractures and irregular crenulations were formed. F_3 is equivalent to F_2 in the Eocambrian

rocks, where the same structures as in the fine-grained schists are formed. Asymmetrical F_2 folds in these rocks have an E-SE vergence. They reflect an E-W shortening as elsewhere in the area.

CHAPTER VI

Ultramafic bodies

INTRODUCTION

As elsewhere in the Scandinavian Caledonides, ultramafic bodies occur abundantly within certain stratigraphic units (Zachrisson 1969; Kulling 1955; Du Rietz 1935 i.a.). In the present area about hundred of these bodies are mapped (Enclosure I). Although the contacts are rarely exposed, they appear to be more or less elongated, lens-shaped bodies, often arranged in rows so as to indicate boundinage of originally larger bodies or sheets. The lenses are mainly concordant with the predominant cleavage, S_2, which curves around them. The petrography of a large number of ultramafic bodies occurring in the present area is described in detail by Du Rietz (1935), Michel (1950), Janssen (1953) and de Keyzer (1952). In this chapter the structures of the bodies will be stressed. As the ultramafics in the four major stratigraphic units show different aspects they are dealt with separately.

THE PHYLLITE BELT

General aspects

Most of the ultramafic bodies in the present area are situated in the phyllite belt (approx. 60). Within this belt their occurrence is limited, with a few exceptions, to the Fatmomakk and Murfjället Formations.

The bodies in this belt are mainly carbonate-bearing serpentinites with minor proportions of tremolite, talc, ore minerals such as chromite and magnetite, and locally chlorite. The percentage of carbonate locally predominates over that of serpentine (e.g. in the bodies north of Aunere). A few talc phyllites are exposed in Aunebäcken. The original olivine and pyroxene are only preserved locally, in particular in the middle of huge bodies such as the one on Aunere. Even here the alteration into serpentine is very irregular as can be illustrated by two specimens from the top of Aunere which contain 10 % and 90 % of olivine respectively. Also in the smaller bodies east and north-east of Aunere high olivine or pyroxene percentages occur locally. One odd body, situated higher in the stratigraphy, in the upper member of the Graipesvare Formation, is described in detail by Michel (1950). It is exposed in Daunebäcken and contains rocks of several compositions.

De Keyzer (1952) described the few exposed contacts with the host rock. It appears that, up to a few metres away from the contact, the phyllites are very rich in minerals from the tremolite-actinolite group and in (clino)-zoisite.

Fig. VI-1. Compositional layering (S_o) in ultramafic body. Small body east of Aunere.

Structures

In several localities (e.g. the centre of Aunere and one of the small bodies east of Aunere) a conspicuous compositional layering can be observed in the field (Fig. VI-1). The layers less resistant to weathering are almost pure dunites, with approx. 90 % of olivine and up to 10 % of serpentine. The ones more resistant to weathering are strangely enough much richer in serpentine (70–80 %); apart from some 10 % of relic olivine they contain up to 10 % of pyroxene or tremolite. This layering generally runs straight through the bodies and does not seem to bear any relationship to the shape of the body or the surrounding cleavage. It rather resembles a remnant layering from the time of consolidation and is therefore called S_o.

The appearance of most of the bodies is massive, the surrounding cleavage only intruding the outer rim as a coarse irregular, often anastomosing, fracture cleavage. On the scale of a thin section the cleavage planes appear to be zones with a smaller grain size or with more or less parallel serpentine flakes. They alternate with irregular lenses or fragments of unoriented serpentine or carbonate material.

In many localities the ultramafics have the appearence of a breccia, sometimes grading into a kind of "conglomerate" (Fig. VI-2). The habit of these rocks was carefully described by Du Rietz (1935, Graipesvare) and by Michel (1950, Aunere and surroundings). The author wishes to draw attention to

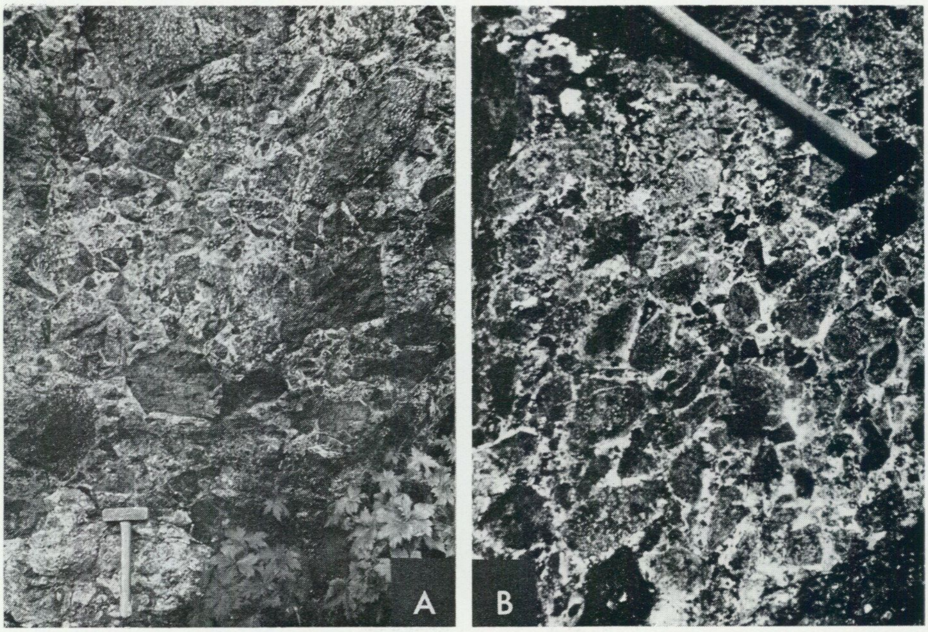


Fig. VI-2. Breccia (A) and "conglomerate" (B) in ultramafic rock bodies. A, south of Murfjället; B, north of Graipesvare.

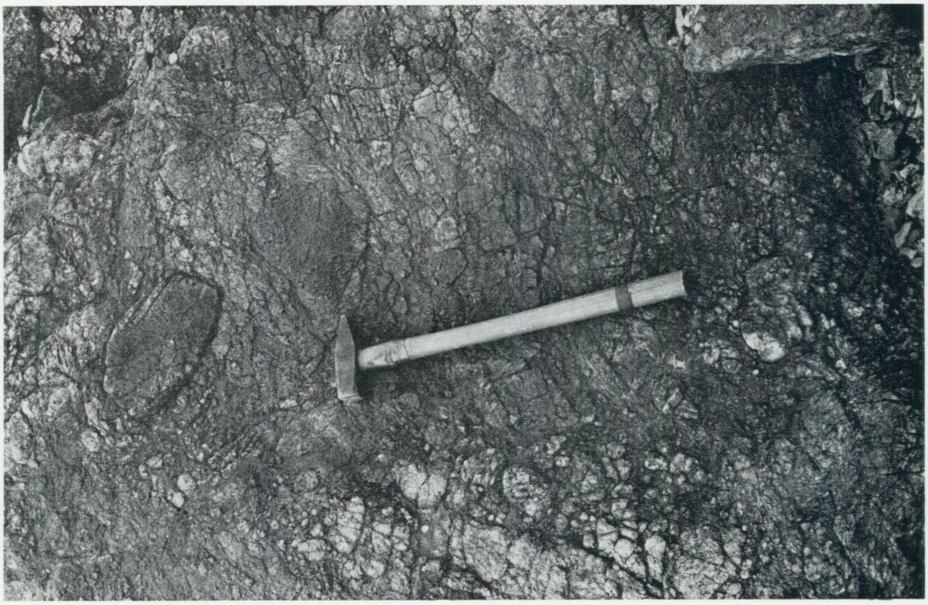


Fig. VI-3. Breccia in huge ultramafic body on Aunere, about 500 m. from the rim.

several facts concerning these breccias and "conglomerates": 1. Their occurrence is irregularly distributed over the bodies. They are even observed in central parts of the Aunere mass (Fig. VI-3). 2. The difference between fragments and matrix can only be expressed in terms of grain size or by a difference in the degree of alteration (e.g. from serpentine to carbonate). 3. The transition from brecciated types to massive types in always gradual. The origin of the breccias and "conglomerates" will be discussed below. In order to illustrate the deviations in orientation of S_2 around the Aunere mass, a detail map and section of this mountain are given in Fig. VI-4.

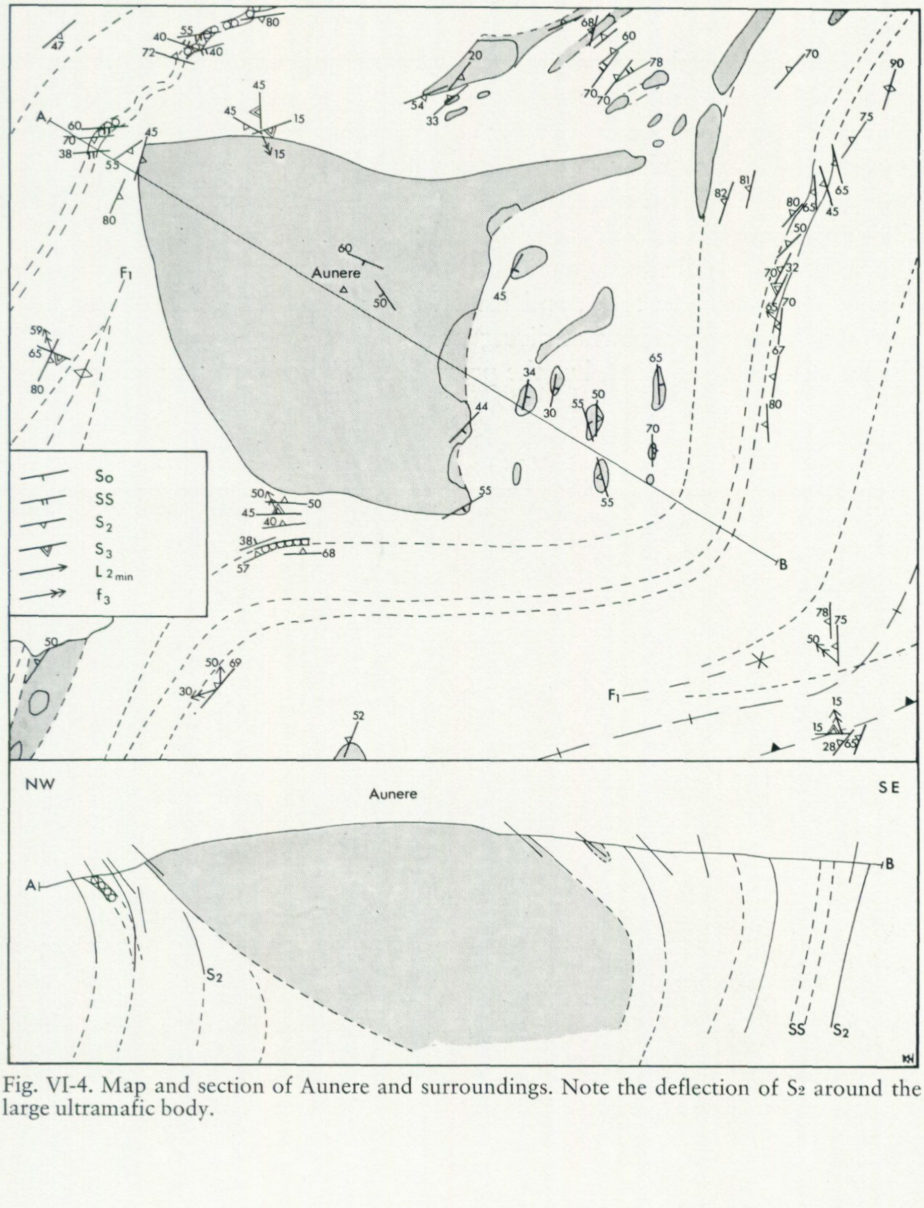


Fig. VI-4. Map and section of Aunere and surroundings. Note the deflection of S_2 around the large ultramafic body.

THE SVARTSJÖBÄCKEN SCHISTS

As previously noted by Du Rietz and Michel, the ultramafic bodies in the schists are in general less altered than those in the phyllite belt. In ten out of thirteen specimens examined, olivine still constitutes between 10 % and 60 % of the rock; in the other three the mineral is completely altered. Minor proportions of pyroxene are often preserved as well. Serpentine may constitute up to 90 % of the rock, tremolite up to 30 %. Common accessories are talc, carbonate and ore minerals.

The contacts with the surrounding rocks, mainly amphibolites, are only exposed in a few places. The amphibolite along these contacts is highly chloritized.

As distinct from the bodies in the phyllite belt, those in this formation are never brecciated but generally possess a well-developed cleavage (Fig. VI-5). This cleavage is locally a crenulation cleavage (Fig. VI-6), obviously a product of a deformed slaty cleavage. It is often highly folded to kink or accordion folds (Fig. VI-5). In the elongated body west of Saletjålt two sets of these late folds are superimposed over each other.

The cleavage is obviously equivalent to S_2 in the surrounding rocks, the deformed slaty cleavage to S_1, and the folds to F_3 and F_4 folds, although the orientation may be somewhat divergent (F_4 is not separately distinguished in the Svartsjöbäcken Schists, but the present occurrence seems a good proof of

Fig. VI-5. F_3 kinkfolds in ultramafic body west of Saletjålt. Note the well developed S_2.

Fig. VI-6. Crenulation cleavage, S_2, in ultramafic body. Small body north of Häbberskullen.

its existence). In order to illustrate the variations in orientation detail maps of three bodies, with the various structural data plotted, are given in Fig. VI-7.

In the localities where a crenulation cleavage has developed the serpentine flakes are oriented along S_1. This means that these flakes grew pre- or syn-F_1. Various tremolite crystals grew pre-, syn- and post-F_2, as can be demonstrated by means of overprinting criteria.

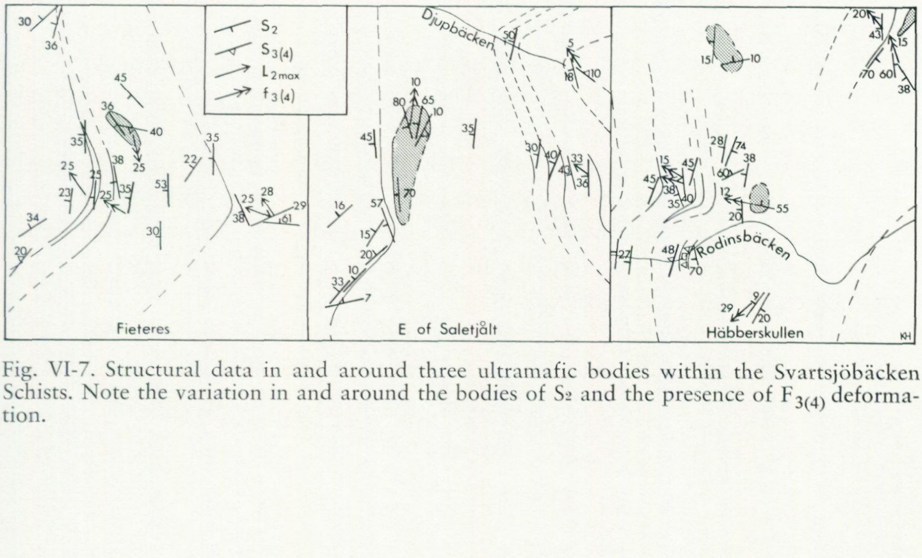


Fig. VI-7. Structural data in and around three ultramafic bodies within the Svartsjöbäcken Schists. Note the variation in and around the bodies of S_2 and the presence of $F_{3(4)}$ deformation.

THE MARSFJÄLLET GNEISS

Only few very small bodies are detected in the gneiss. They are mainly dunites, hardly altered, the proportion of serpentine not exceeding 10 %. In general no schistosity is visible; only in the body north on Risfjället is a fold in the compositional layering accompanied by a vague axial plane schistosity.

THE EASTERN SCHIST AND AMPHIBOLITE BELT

The metamorphic grade of this belt is comparable to that of the Svartsjöbäcken Schists as are the ultramafic bodies. However, in this belt they are much larger and probably for that reason less altered. Many bodies are dunitic in composition, but pyroxene-bearing peridotites are found as well. The cleavage is in general poorly developed, probably also as a result of the size of the bodies. A detailed description of one of these bodies is given by Calon (pers. comm.).

DISCUSSION

As stated before, most of the ultramafic bodies in the phyllite belt are concentrated in the Fatmomakk and Murfjället Formations. This is in accordance with the general view that, apart from a few exceptions, they occur in the older part of the Köli sequence (Zachrisson 1969; Kulling 1933, 1960; Strand 1960). The explanation that this concentration is caused by early intrusion of the bodies, followed by erosion and sedimentation of monomict serpentine conglomerates, seems unlikely to the present author (Kulling 1933, 1960; Zachrisson 1969). Michel (1950) extensively discussed the origin of the breccias and "conglomerates"; he concluded that the ones in the northern part of the present area have a tectonic origin. The present author agrees with this opinion and wishes to endorse Michel's statement by adding the following arguments to those put forward by him.

1. The transition from brecciated and conglomeratic ultramafics to massive varieties is always gradual. An unconformity of conglomerate on massive ultramafic rock, as is to be expected of a sediment, is not observed.
2. The breccias occur irregularly distributed throughout the bodies, even close to the centre of the huge body on Aunere. This is hard to understand if the origin is sedimentary.
3. The association with the quartzite conglomerate is not convincing in the present area. Transitions or contacts between the two have not been found.
4. The breccias and conglomeratic rocks occur in many bodies in the phyllite belt. They are definitely not bound to one stratigraphic level, as should be expected of such a peculiar sediment.
5. The important F_1 and F_2 deformation in the phyllite belt also attacked the ultramafic bodies, as may be seen from their lens shape, boudinage etc. No penetrative cleavage or schistosity is, however, formed. This and the

transitions between the coarse fracture cleavage and the breccias suggest that the deformation in these bodies was of a heterogeneous, brittle type. It seems likely that during such a deformation breccias or conglomeratic structures were formed or intensified. In the Svartsjöbäcken Schists, however, where the possibly more intensive deformation took place under higher metamorphic circumstances, a penetrative cleavage was formed instead of the heterogeneous breccias.

According to this latter point on the origin of the breccias and "conglomerates", the views of Avé Lallemant (1969) are of interest. He described a lherzolite body in the Pyrenees which is heavily brecciated along the margins and along irregularly distributed zones throughout the entire body. The breccias show a striking resemblance to those from the present area, as could be established by the present author during a recent visit.

Avé Lallemant interpreted the breccias as a result of the intrusion of the lherzolite as a rigid body. This manner of intrusion is demonstrated to be likely for many other ultramafic bodies (Möckel 1969; De Roever 1957), on the basis of the arguments that they often possess older structures than their host rock and that they do not have a contact aureole. These two arguments seem to be valid in the present area as well, the compositional layering being an older structure. It can therefore not be excluded that the breccias originated during the intrusion of the bodies. Later deformation phases should in that case only have intensified these structures and locally they should have modified them into pseudo-conglomerates.

The deviations in orientation of structures in the ultramafic bodies, as illustrated in Figs. VI-4 and VI-7, are probably best understood by comparing the bodies with porphyroblasts. The contrasting mechanical properties of the bodies and their surroundings caused initial deviations in the principal directions of strain and differences in rotation during progressive deformation, probably resulting in the present deviations.

CONCLUSIONS

1. The initial intrusion of the ultramafic bodies must be pre- or syn-F_1, since the serpentinization is pre- or syn-F_1.
2. The locally preserved compositional layering, not related to the shape of the bodies, and the lack of contact aureoles make intrusion as rigid bodies probable.
3. The different metamorphic circumstances are probably the main reason for the fact that the ultramafic bodies in the Svartsjöbäcken Schists contain a penetrative cleavage whereas those in the phyllite belt are massive or brecciated with only locally a coarse fracture cleavage.

4. The ultramafic breccias and "conglomerates", occurring in the phyllite belt, do not have a sedimentary but a tectonic origin. They are either formed by brittle F_1 and F_2 deformation or during the intrusion of the bodies in a rigid state. In the latter case the deformation phases should only have played an intensifying or modifying role.

CHAPTER VII

The results considered in the context of the Seve-Köli Nappe Complex

The conclusions from the preceding chapters are summarised in a synoptic diagram (Enclosure IV). Some items from this table will be shortly discussed with respect to current opinions on the Seve-Köli Nappe Complex. In the area investigated the Seve-Köli Nappe Complex has been subdivided into three major tectonic units. Most workers on a more regional scale subdivided the Seve-Köli Nappe Complex into two parts, the Seve and the Köli sequences. It is hard to judge to what extent the intra-Seve thrust is of regional importance. Directly south from the mapped area the Marsfjället Gneiss is truncated, probably by the underlying thrust (Brandt, pers. comm.). Further south and westwards the gneiss reappears locally, always at least on one side truncated by a tectonic contact. On Mount Borgafjällen a tectonic contact between rocks equivalent to the Svartsjöbäcken Schists and rocks equivalent to the eastern schist and amphibolite belt plays still an important role in the Seve tectonics (v.d. Harst 1956; Stevens, pers. comm.; v. Bever Donker, pers. comm.). To what extent this tectonic contact can be traced further along the Seve nappe is as yet undetermined. As can be seen on Enclosure IV, four deformation phases are recognized in the phyllites, three in the central and eastern units, and two in the Eocambrian quartzites and sparagmites. It is unsure whether the first of these phases, F_1, which did not affect the Eocambrian rocks, took place at the same time in all three major units, since the emplacement of these units presumably only took place during F_2. In the phyllites F_1 can be dated as being post lower Silurian, since the lower Silurian sediments are affected by this phase. It seems probable that F_1 deformed all Seve-Köli rocks more or less contemporaneous and that this phase is responsible for the "downfolding" of the sediments, bringing them to deep-seated areas, where they were metamorphosed under high pressure conditions syn- and post-F_1. F_1 produced tight folds, accompanied by a slaty cleavage, S_1, in the phyllites and a penetrative fabric in the higher grade rocks. Although earlier thrusting is not to be excluded, the emplacement of the nappes took place during F_2, accompanied by important deformation, resulting in folding and mylonitization. F_2 apparantly postdates the climax of metamorphism, since it is, particularly in the high grade rocks, a post-crystalline deformation. During F_2 the slaty cleavage, S_1, in the phyllite belt was generally deformed to a crenulation cleavage. Locally the deformation resulted only in rotation of S_1, to form S_{1+2}.

During later phases, F_3 and F_4, less important deformation took place under low grade conditions. The rocks from the phyllite belt show the most effects from these phases. In the schists and gneisses only scattered F_3 folds are found.

These four deformation phases correspond roughly to the ones mentioned by Zachrisson (1969). His pre-F_1 corresponds to F_1 as defined here; his F_1 can be correlated with F_2, his F_2 with F_3 and his F_3 with F_4.

Henley (1970) labelled deformation phases in the Sulitjelma region D_1 to D_4 (see also Wilson 1972). These phases show some similarity with F_1 to F_4 in the Marsfjällen area; for instance also in Sulitjelma the schistosity is seen as a result from the first two phases.

According to Glass (in prep.) the metamorphism in the eastern belt is inverted. Since it is unsure whether the eastern belt is composed of one thrust sheet or of several ones, it does not necessarily follow that the whole eastern belt is inverted.

The thrusting of the gneisses over the eastern schist and amphibolite belt was in easterly direction; this was confirmed by analyses of asymmetrical muscovite "fish". Strain analyses from rotated garnets brought to light that F_2 shear along the contact between the Seve and the Köli rocks was in opposite sense. Even F_1 shear in the Svartsjöbäcken Schists was probably opposite to the general "west over east" picture (Fig. VII-1). This anomalous shear direction is explained by assuming that the more rigid Seve rocks were pushed from below in easterly direction, whereas the less competent Köli phyllites, riding on the back of the Seve, lagged relatively behind. It is in this respect of interest that Wilson (1972), who analysed rotated garnets in the Sulitjelma region, found the normal "west over east" sense of shear there.

Many authors discussed the meaning of the so-called transverse lineation, common in the Caledonian mountain chain (Kvale 1953; Lindström 1955; Strand and Kulling 1972; Christie 1963). These lineations, which are mainly mineral lineations, are interpreted in this thesis as being parallel to the longest axis of the strain ellipsoid (X). In a model of simple shear X is close to parallel to the shear direction *a,* which is, of course, more or less perpendicular to the trend of the mountain chain. The origin of folds with their axis parallel to this lineation is discussed (Chapter IV, Discussion of S_2 planes and mylonitization).

In Fig. VII-1 a tentative reconstruction is given to illustrate in what way the nappes are believed to have travelled to their present setting. The nappes are thought to have formed by a process of plastic imbrication (Nicholson and Rutland 1969), as is strongly suggested by the fact that they wedge out westwards (Zachrisson 1969; Nicholson and Rutland 1969). The author is well aware of the fact that the present state of knowledge does not fully justify a reconstruction like this, but it might have value to stimulate a discussion.

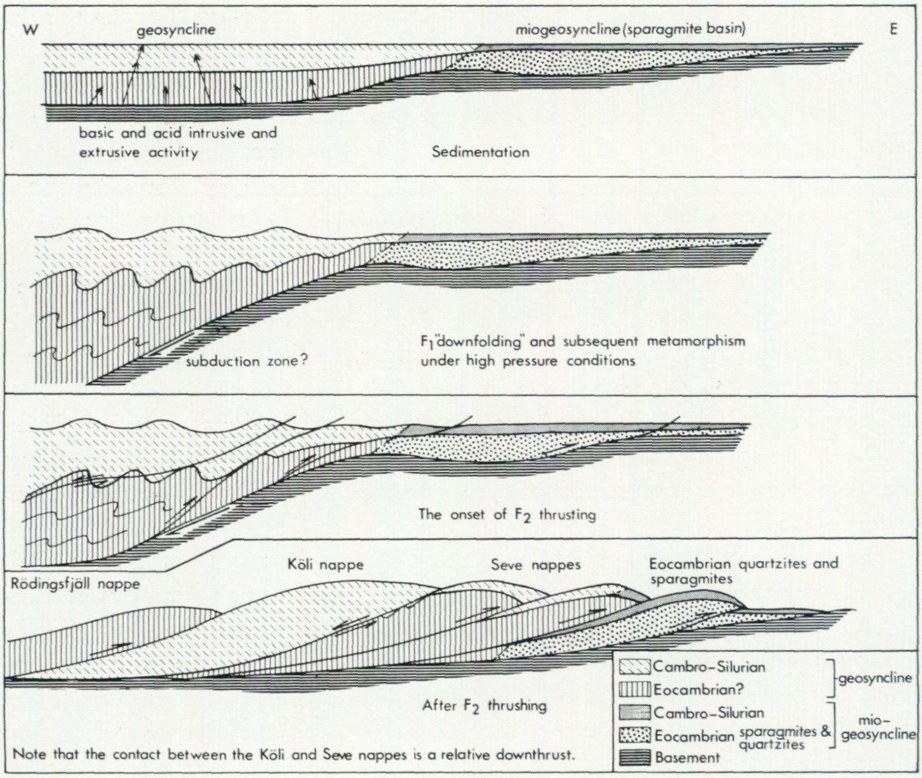


Fig. VII-1. Tentative reconstruction of the tectonic evolution of the Seve-Köli Nappe Complex.

CHAPTER VIII

General discussion of crenulation cleavage

Careful observation of crenulation cleavage (S_2) in many thin sections from the phyllite belt proved the existence of several distinct types:

1. The S_1 slaty cleavage is gently buckled into symmetrical microfolds; no real crenulation cleavage occurs, but S_2 can be defined as the axial plane of the microfolds (Fig. VIII-1A). This case is interpreted as an incipient stage in the formation of crenulation cleavage. In the fine-grained phyllites it is very rare, in the coarser ones it is more common but still only occurs locally.
2. The S_1 slaty cleavage is buckled, generally into asymmetrical microfolds. A distinct crenulation cleavage is defined by irregular opaque zones, which appear in a thin section as dark lines in the middle of the short limbs of the microfolds, parallel to the axial planes (Figs. VIII-1B, 2; II-8). These dark zones are always much thinner than the zones between them. This is the most common type in slates and fine-grained phyllites. It may occur in

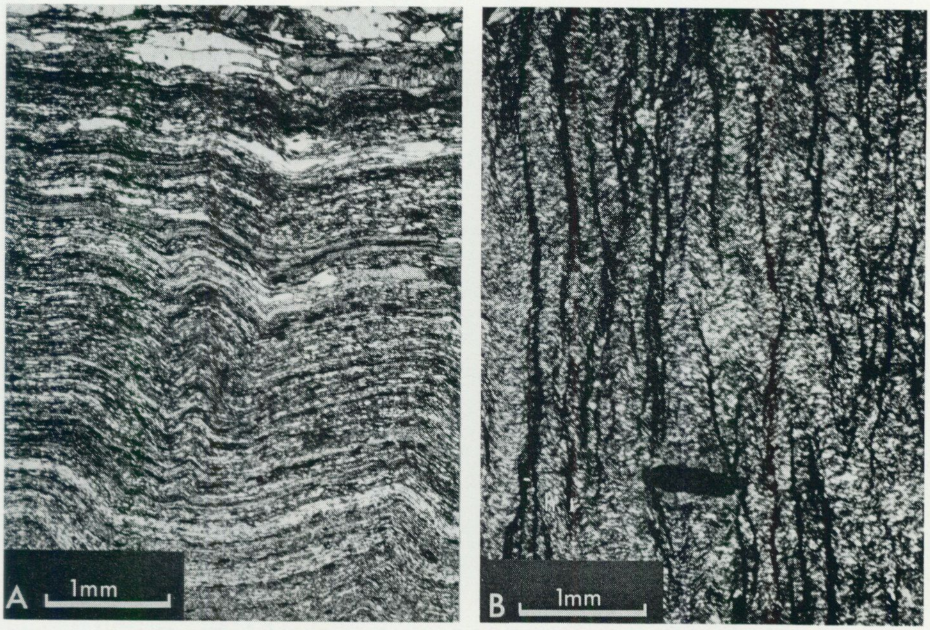


Fig. VIII-1. A, crenulated slaty cleavage; no real crenulation cleavage has developed yet (type 1). B, crenulation cleavage (type 2).

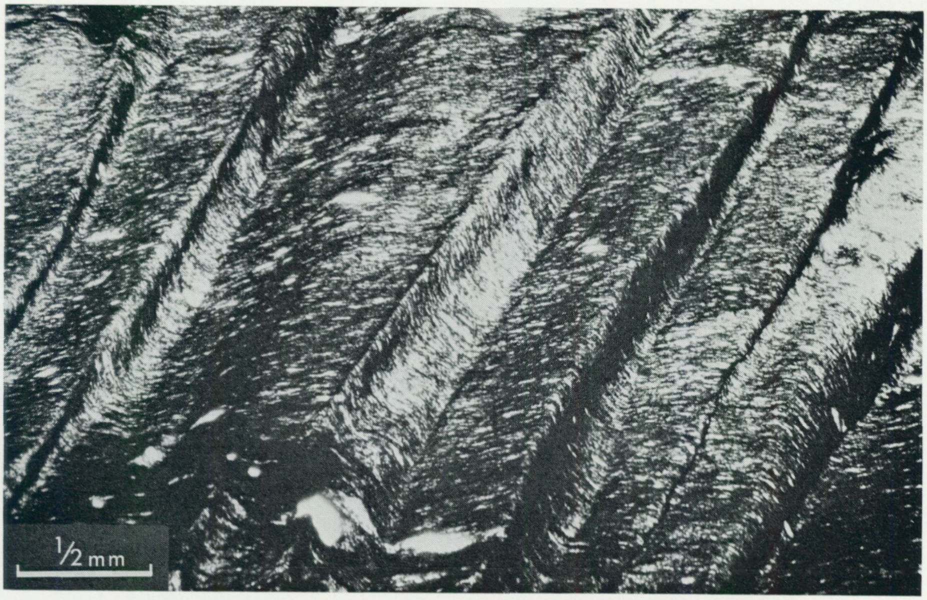


Fig. VIII-2. Crenulation cleavage, consisting of dark zones, situated in the middle of the short limbs of asymmetrical microfolds.

combination with type 1, in which case this type is principally confined to limbs of larger folds, whereas type 1 occurs in the hinges.

3. The S_1 slaty cleavage is kinked. Distinct "crenulation cleavage lines", as described in type 2, occur in the hinges of the kinks. This type is found in a few localities in the lower grade part of the phyllite belt.
4. The S_1 slaty cleavage is buckled into asymmetrical folds. One limb, generally the "flexure-like" short one, is richer in mica and/or opaque minerals than the other limb. The micas in this limb are slightly oblique with respect to the axial plane of the folds (S_2) (Fig. VIII-3A). The mica-rich layers can be thinner, equal in thickness, or even thicker than the layers relatively rich in quartz. This type is common in the coarser-grained phyllites and in the fine-grained schists.

All these different types may grade into one another or into slaty- and fracture cleavage (e.g. Fig. VIII-3B). This, as well as the fact that S_2 is a pervasive S-plane, parallel to the axial plane of F_2 folds, throughout the phyllite belt, suggests that S_2 has a similar relationship to the strain as a slaty cleavage. Many workers (Ramsay 1967, p. 180; Dieterich 1969; Talbot 1964; Flinn 1965) agree that slaty cleavage develops along the XY plane of finite strain. Others (e.g. Williams 1972) are not convinced by the supporting evidence that this has to be always the case. If this opinion is however correct, as believed by the

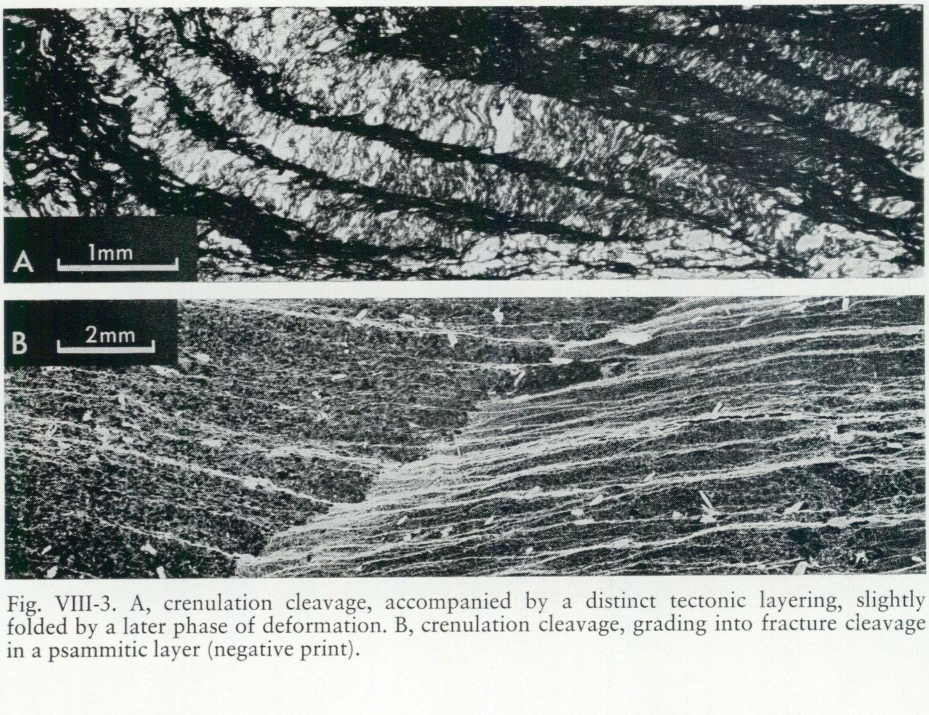


Fig. VIII-3. A, crenulation cleavage, accompanied by a distinct tectonic layering, slightly folded by a later phase of deformation. B, crenulation cleavage, grading into fracture cleavage in a psammitic layer (negative print).

author, and if S_2 really has a similar relationship to the strain as a slaty cleavage, S_2 should be parallel to the XY plane of the finite F_2 strain.

Much work has been published concerning the mechanisms which might be responsible for the formation of crenulation cleavage. Among the more recent Hoeppener (1956), Knill (1960), Rickard (1961), Wilson (1961), Talbot (1964), Dewey (1965), Plessman (1964), Ramsay (1967, p. 390), Talbot and Hobbs (1968), Williams (1972) and Durney (1972) should be mentioned. Several of these (e.g. Hoeppener, Knill, Wilson) consider the crenulation cleavage planes essentially as rotating slip planes. The name "strain-slip cleavage" speaks for itself. Others (e.g. Rickard, Talbot, Dewey, Ramsay, Talbot and Hobbs) criticize this genetic name, though they still explain the phenomenon at least partly by slip on S_2 or shear parallel to S_2. Rickard considers the slip on S_2 planes as a secondary phenomenon, controlled by the microfolds (Rickard 1961, pp. 328–329). Wilson (1961) gives an excellent review of the literature on this subject; according to him:

> "Strain-slip cleavage, like fracture cleavage, is the result of the failure of the rock along discrete, closely spaced parallel surfaces."

Talbot (1964, p. 1042) states:

> "Crenulation cleavage is pictured as a kinking or microfolding followed by some shearing parallel to one of the limbs."

Dewey (1965, p. 485, 486) described "pelitic strain bands" as one of four kink band types:

> ". . . pelitic strain bands are generally small scale structures forming a fine crenulation cleavage. Sericite or muscovite commonly forms a felt at the margins or strain bands co-planar with the axial surface, and slip may occur on these felts to produce kink planes."

Ramsay (1967) states:

> "In coarsely crystalline rocks the shear zones may be up to 3 cm apart, and the micas are folded into coarse fold pleats between these zones to form what is sometimes called a crenulation cleavage or herringbone cleavage. The planes of slip are commonly the loci of recrystallization of the rock material during the production of the folds."

Finally Talbot and Hobbs (1968, p. 584):

> "Bedding is observed crossing the layering and commonly has been differentially displaced along the cleavage planes."

It appears that displacements along apparent microfaults, or flexure-like structures are used as evidence for slip or shear on the crenulation cleavage planes. If these planes are however parallel to the XY plane of finite strain, as believed by the author, such slip or shear cannot take place and the apparent displacements have to be explained in another way.

Many of the previously mentioned workers (e.g. Plessman, Williams, Durney) also drew attention to the mechanism of chemical migration due to solution, resulting in compositional layering parallel to the crenulation cleavage. In the area described here the S_2 crenulation cleavage planes locally displace quartz veins or thin beds in a fault-like manner. The author is however convinced that the apparent displacement can be explained adequately by initial buckling or kinking, followed by solution transfer of soluble minerals such as quartz and calcite from the limb zones of the microfolds. Slip or shear parallel to S_2 need not take place in such a model, in which the S_2 planes are comparable to structures such as stylolites in limestones (Plessman 1964; Durney 1972). The arguments in favour of the above statement are as follows:

1. Apart from type 1, where no real crenulation cleavage exists, the crenulation cleavage is defined by dark layers, however thin they may be, with a diverging composition. The origin of this compositional layering can only be explained in terms of solution transfer.
2. The morphology of the S_2 planes (i.e. dark layers) is often very irregular (Fig. VIII-4). This irregularity fits much better in the stylolite concept than in the concept of slip planes.
3. The apparent displacement of quartz veins and beds by S_2 planes can be explained by solution transfer of quartz and/or carbonate from flexure-like fold limbs. In such limbs all transitions are found, from areas of no solution effects to well developed differentiation (Fig. VIII-5). The fact that larger

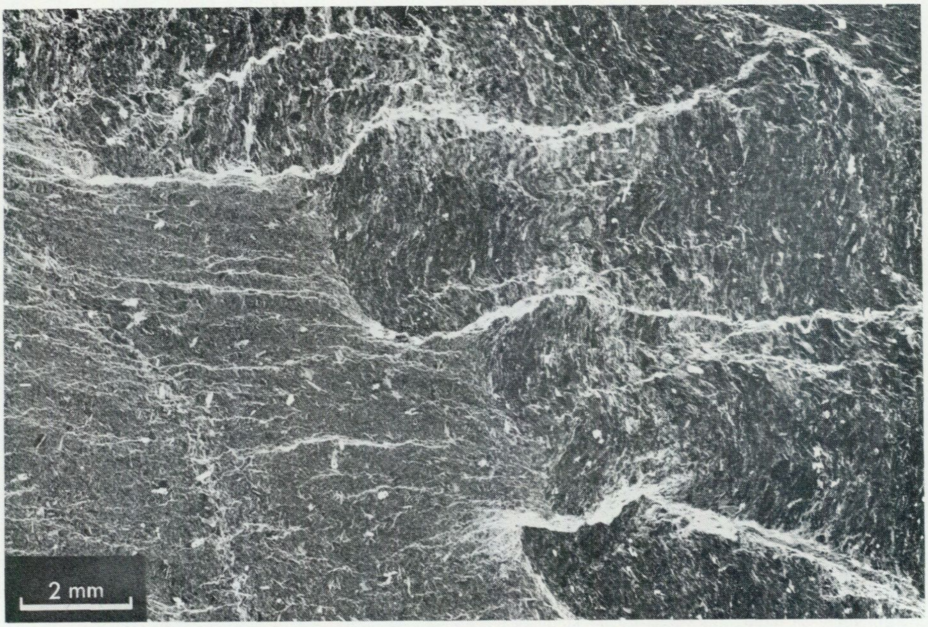


Fig. VIII-4. Irregular crenulation cleavage planes (S_2) with apparent displacement of SS. The S_2 planes bear more resemblance to stylolites than to microfaults (negative print).

Fig. VIII-5. Crenulation cleavage. Note different degree of differentiation in the fold limbs.

displacements are in general accompanied by thicker dark layers illustrates this (see also Williams 1972).

4. All details of the four types of crenulation cleavage described can be explained by the model. Type 1 represents the stage after buckling; no solution transfer took place as yet. Type 3 represents kinking, where solution transfer took place along the "broken" fold hinges, leaving a residue of dark material. Type 2 and type 4 are the normal types, where solution transfer of quartz and/or carbonate took place after buckling. Growth of mica in the mica-rich layers is improbable, because the micas have the same size and appearance as those in the quartz-rich layers (see also Williams 1972). It is more likely that the quartz migrated (Plessman 1964; Dewey 1965; Williams 1972; Durney 1972). Where they can be compared with relatively little deformed rocks (e.g. fold hinges), the mica-rich or opaque layers are always much darker, whereas the quartz-rich layers are about equal in composition. It follows that the quartz migrated probably away from the dark layers and precipitated in veins or pressure shadows (Plessman 1964; Williams 1972; Durney 1972). The numerous quartz veins occurring locally confirm this conclusion.

Thus the observed fault-like displacements are interpreted as being apparent rather than real. This interpretation fits the idea that S_2 is parallel to the XY plane of finite F_2 strain.

The question remains as to why solution transfer is mainly restricted to those limbs of microfolds which are most nearly parallel to the axial planes. Williams (1972) explained this by assuming:

"... movement zones that would become "channel ways" for the fluid migrating through the rock. Quartz would then be dissolved from the walls of these "channel ways"."

Although this idea seems likely for the more brittle deformed rocks and for the more ductile deformations as soon as the crenulation cleavage planes have formed, it seems difficult to locate these "channel ways" in the middle of fold limbs at an initial stage.

Another explanation is based on the important role of pressure-solution at quartz-mica contacts (Voll 1960, p. 534). It appears that quartz is apt to dissolve along these contacts more than on other contacts. The position of these contacts in the stress field is of course different after buckling (Fig. VIII-6). In the limbs of the microfolds or, in asymmetrical microfolds, in those limbs which are most nearly parallel to the axial plane (B), a larger proportion of quartz-mica contacts is in a position at a high angle to the main stress direction, assuming that this direction is close to perpendicular to the axial plane of the microfolds. According to the fact that pressure-solution occurs mainly on surfaces statistically perpendicular to the main stress direction (Durney 1972) pressure-solution is clearly favoured in those parts of the fold limbs where the quartz-mica contacts and hence the micas are most nearly parallel to the axial plane. The more quartz is dissolved the better the

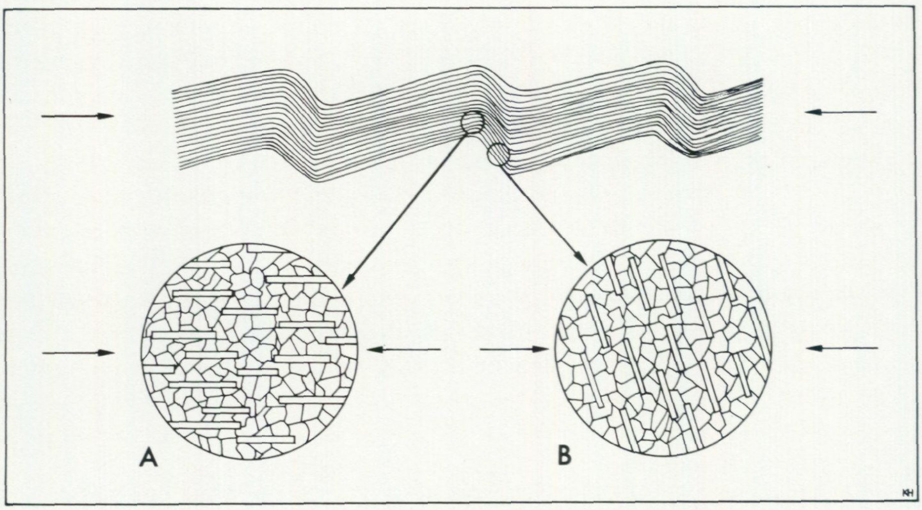


Fig. VIII-6. Diagram illustrating that after an initial stage of buckling the orientation of quartz-mica contacts in hinges (A) and limbs (B) of microfolds is such that pressure-solution is favoured in the hinges. See text for further explanation.

micas become perpendicular to the main stress. The process will thus be accelerated, especially as soon as the "channel ways" begin to work. This may explain why the pressure solution appears to be restricted to such narrow zones.

The described process of preferred solution transfer from certain domains might also play a prominent role in the development of fracture- and possibly even of slaty cleavage (Dewey 1965). In the area studied crenulation cleavage in pelitic layers often grades into fracture cleavage in psammitic layers (Fig. VIII-3B) and even within the same lithology both structures are grading into one another. As also recognized by previous workers (e.g. Wilson 1961), the phenomena are so closely allied that a common mechanism seems highly probable.

Discussion of rotational textures in garnet crystals

To explain rotational textures in garnets, several mechanisms are proposed in the literature. These mechanisms can be roughly subdivided into two categories:

1. mechanisms explaining the rotation of planes during progressive or superimposed deformation, with respect to fixed axes, assuming that the garnets remain stationary or rotate at a different rate (e.g. Ramsay 1962; Wilson 1971).
2. mechanisms explaining the rotation of the garnets themselves, for instance under the influence of shear movements with respect to the same fixed axes (e.g. Spry 1963; Cox 1969; Rosenfeld 1970).

Ramsay (1962, pp. 322, 323) gives a model of pure shear deformation in which S-planes are rotated around stationary garnets (Fig. VIII-7). He claims that rotations of up to 90° can be explained in this way.

The S-planes defining S_i and S_e are in general slaty cleavage planes. This has two important implications:

1. Slaty cleavage is believed to form parallel to the XY plane of finite strain, which is a stationary plane in pure shear deformation. Therefore the mechanism proposed by Ramsay apparently only explains rotations caused by a superimposed deformation, since the rotating slaty cleavage cannot be formed as a result of the same pure shear deformation.
2. If the cleavage in Ramsay's model has been rotated more than 45°, the cleavage planes must have been shortened in the early stages of deformation (Fig. VIII-7, see also Ramsay 1962, Figs. 15, 16, and de Sitter 1956, p. 282). Shortening of slaty cleavage will, in the opinion of the author, in general result in crenulation and/or crenulation cleavage. Under these circumstances helicitic folds are to be expected in the garnets rather than regular S-patterns.

Fig. VIII-7

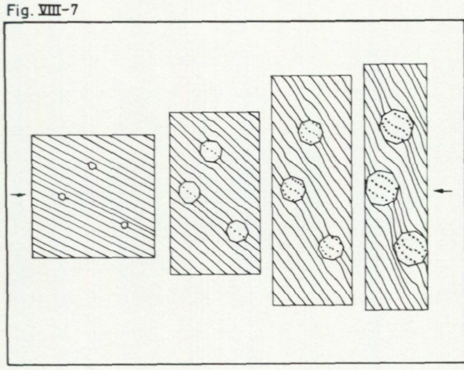


Fig. VIII-9

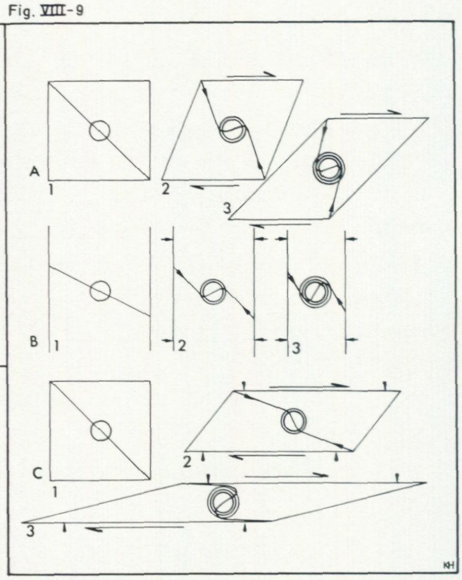


Fig. VIII-7. The development of curved inclusion trails within syntectonic minerals in rocks deformed by flattening, after Ramsay (1962).

Fig. VIII-9. Mechanisms of rotation, after Powell and Treagus (1970). A, simple shear inducing buckling. B, buckling in an oblique stress field. C, buckling during rotational homogeneous strain. In each case the diagonal thick line represents a rock layer of relatively low ductility which responds to stress by buckling. The length of the layer remains constant throughout the deformation. The arrows on the layers represent directions of resolved compressive stress while those outside each diagram indicate the orientations of external stress systems. The circles show the progressive growth of an incompressible crystal. 1, 2, and 3 represent successive stages in the deformation. Rotation of both the layering and the crystal occurs in each case.

Ramsay (1962) suggested that rotation angles larger than 90° may be produced by successive non-parallel flattenings or by strong inhomogeneous local shear. The first of these mechanisms is further dealt with by Wilson (1971) under the name of rotational strain. He demonstrates how rotations of more than 90° can be attained during a process of rotational strain (Fig. VIII-8). Unfortunately in this respect he only considers the rotation of passive planes (e.g. plane M), whereas he himself states elsewhere that a slaty cleavage plane during subsequent deformation is not likely to be passive. It is therefore interesting to consider the behaviour of a slaty cleavage plane in Wilsons model, which remains parallel to the XY plane of finite strain. In the case of one progressive deformation, in which the slaty cleavage is both formed and rotated, the cleavage should begin to form in C (Fig. VIII-8, plane S), parallel to the X direction of incremental strain. After the rigid body rotation in D the cleavage should be deformed in E to a position related to the finite strain ellipsoid (in the two dimensional model: ellipse). In Wilsons model, in which the garnets rotate with the rigid body rotation and remain stationary during strain, this would mean that after a strong deformation with a rigid body rotation of 50° the garnet rotation with respect to the surrounding cleavage plane should be approx. 30° (Fig. VIII-8).

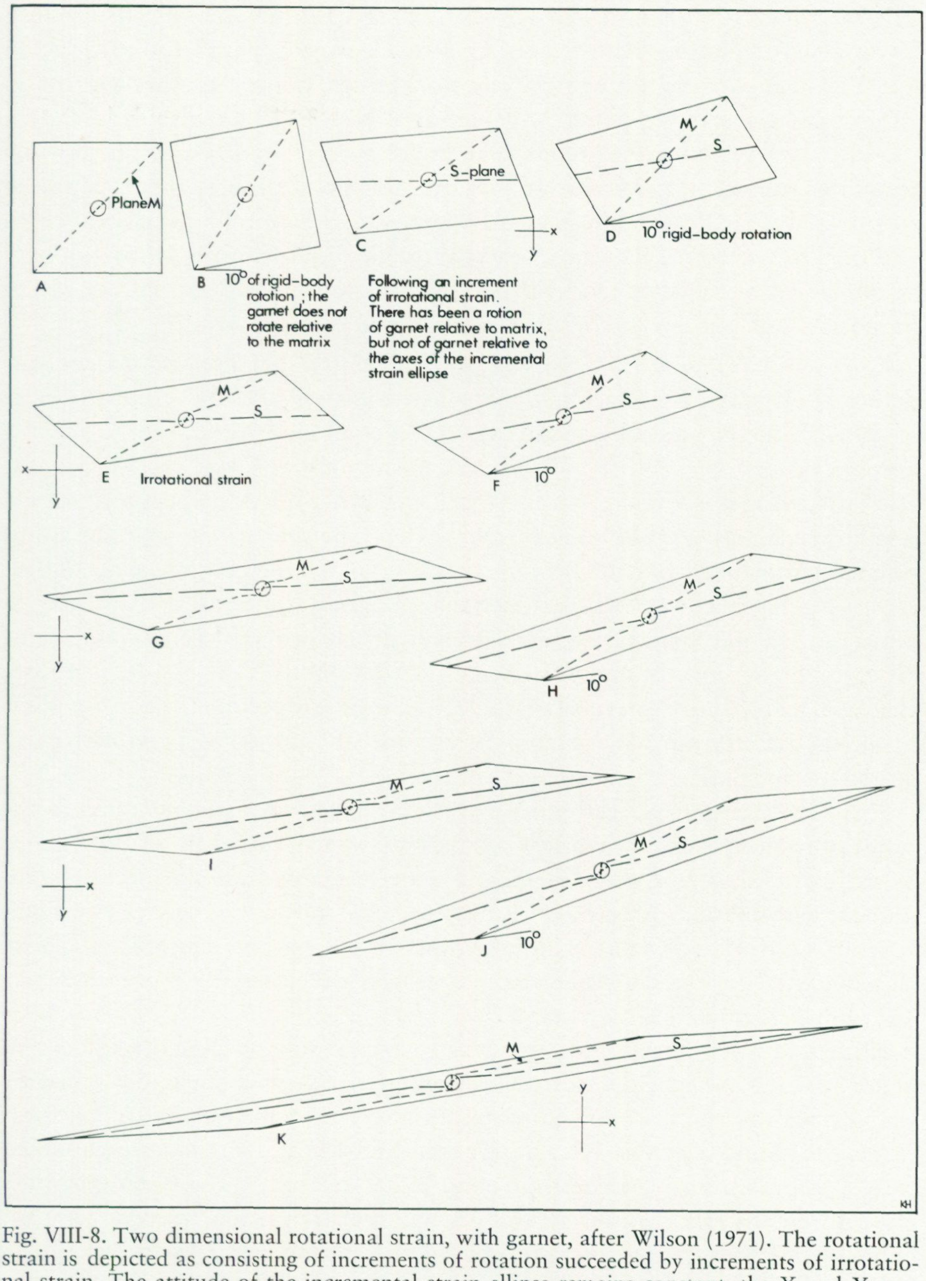


Fig. VIII-8. Two dimensional rotational strain, with garnet, after Wilson (1971). The rotational strain is depicted as consisting of increments of rotation succeeded by increments of irrotational strain. The attitude of the incremental strain ellipse remains constant, the X and Y axes being indicated. The figure indicates the amount of relative rotation between a plane "M" and the garnet.
Plane "S" has been added by the author; it remains parallel to the X direction of finite strain. For further explanation see text.

If the rotational deformation is superimposed and has for instance such a favourable orientation that the existing slaty cleavage is parallel to plane M in Fig. VIII-8, higher rotation angles can be attained. Since the cleavage rotates faster than the passive plane M (Ramsay 1967, 3.11, see also below), the rotation angles will be even higher than the ones given by Wilson. The favourable orientations to which this applies are, however, in three dimensions not very probable, as Wilson himself demonstrated and as will be discussed below.

It may be concluded that larger rotation angles, up to approx. 100° can only be explained by Wilsons model if the garnets grow during superimposed deformation with a special favourable orientation.

Powell and Treagus (1970, pp. 812, 813) suggested that a bulk homogeneous deformation may cause buckling of the active matrix layering (S) at the sides of growing incompressible crystals so that a combined rotation of S and the crystals occurs (Fig. VIII-9). This mechanism only explains rotation in the case of superimposed deformation, since an older S-plane is buckled. A process in which garnet crystals cause, and at the same time overgrow microfolds, often accompanied by crenulation cleavage, should in the opinion of the author rather produce helicitic folds or irregular inclusion patterns than regular S-patterns. The mechanism can by no means give a general explanation of the observed facts, since in many cases with a continuous S_i–S_e the S_e is a slaty cleavage, believed to be parallel to the XY plane of finite strain. This is the plane of maximum streching and cannot be a plane of shortening as would be required for buckling.

Spry (1963) and Cox (1969) explain the S-shaped inclusion patterns as the result of rotating garnets in a flowing matrix. In their model the slaty cleavage or schistosity acts as a shear plane and hence remains stationary during progressive deformation. Wilson (1971, pp. 248, 249) showed, by means of evidence from the work of Ramsay and Flinn, that this is not possible, since the slaty cleavage will be rotated by the strain out of its parallelism with the shear plane. This will be further discussed below.

Rosenfeld (1970) proposed a model in which a spherical rigid body is embedded in a homogeneous isotropic medium. If this medium is deformed by simple shear the body rotates through an angle Ω_e with respect to the shear plane. He found a simple relationship between Ω_e and the amount of shear $\gamma : \Omega_e = |{}^1/_2\,\gamma|$, both theoretically and experimentally. This relationship is quite different to the formula given by Schmidt (1918, p. 301): $\Omega = \gamma$, which is based on the model of a sphere rotating without slippage between two plates, underestimating γ by a factor $^1/_2$ according to Rosenfeld (1970, pp. 15, 16).

Rosenfeld takes into account that the reference plane S (slaty cleavage) need not coincide with the shear plane and can therefore rotate as well. He gives a simple relationship for the case in which both rotations take place around one axis: $\Omega_e = \Omega_s + \Omega_i$, in which Ω_s is the rotation angle of the S plane with respect to the shear plane, measured in the same direction as Ω_i, which stands for the rotation of the garnet with respect to the S-plane.

Rosenfeld states that his example of simple shear may be accompanied by pure shear to represent a more general rotational deformation, without changing the mechanism of garnet rotation by shear movements.

The mechanism proposed by Rosenfeld can explain high rotation angles if the shear strain γ is sufficiently large.

Some workers (Peacey 1961; Langheinrich 1964; Zwart and Oele 1966) stressed the fact that porphyroblasts in different fold limbs may be rotated in opposite sense (Fig. VIII-10).

In the massive of Rocroi (Zwart and Oele 1966) low grade metamorphic slates with magnetite porphyroblasts are deformed. From the curved strain shadows around the porphyroblasts the sense of rotation and hence the sense of shear could be determined. The inferred opposite sense of shear in alternating fold limbs is explained by the buckling of competent beds as the dominant deformation mechanism.

Langheinrich (1964) too, reported an opposite sense of rotation in different fold limbs, from slates in the Harz. The fact that the rotation in these slates is

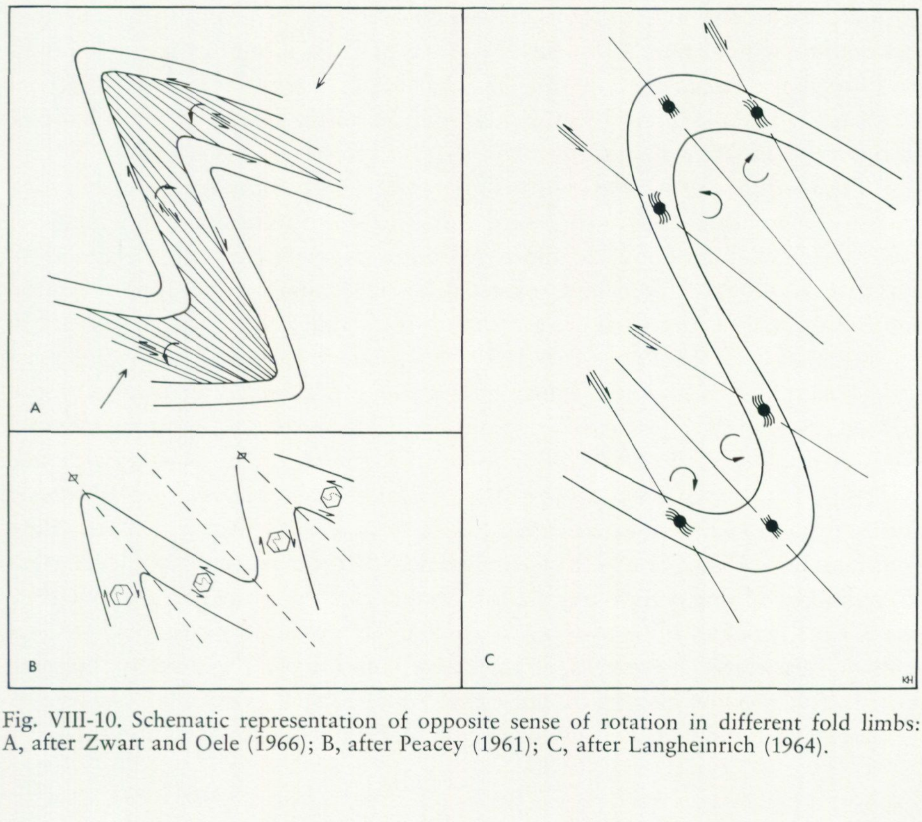


Fig. VIII-10. Schematic representation of opposite sense of rotation in different fold limbs: A, after Zwart and Oele (1966); B, after Peacey (1961); C, after Langheinrich (1964).

the reverse of that in the slates of Rocroi is explained by shear movements on the fanning cleavage planes (Fig. VIII-10; see also Zwart and Oele 1966, p. 73). This kind of shear movements is, however, unlikely to occur during folding, as is argued elsewhere (Chapter II, Discussion of folding mechanisms; see also Ramsay 1962). It seems more probable that buckling was the dominant deformation mechanism in this case too.

Peacey (1961) described a thin section of a schist in which the garnets are rotated in opposite senses. She interpreted these rotations as due to the fact that the garnets are situated in different limbs of early folds. In this schist the S_i of the garnets is not continuous with S_e as a result of recrystallisation. The garnet rotation axes are not determined and the picture is disturbed by later folding. These facts make her conclusion rather doubtful.

In a higher grade metamorphic region (garnet mica schists), where homogeneous deformation is believed to be relatively more important than the initial buckling, such rotations in opposite senses are not very likely to occur.

A few details of the mechanisms mentioned will be discussed below:

What happens to slaty cleavage during progressive and/or superimposed deformation?

If the slaty cleavage or schistosity in a rock is really a penetrative fabric down to the scale of the individual minerals, without any layering, the cleavage is believed to remain parallel to the XY plane of finite strain during progressive and/or superimposed deformation (Ramsay 1967, p. 180; Flinn 1965; Dieterich 1969 and others). The problem is then reduced to the question: what happens to the XY plane of finite strain?

Ramsay (1967, p. 92, Fig. 3–30) showed that both during progressive rotational deformation and during superimposed deformation the XY plane of finite strain changes position, different from a passively deformed plane as for instance a colour marked layer (Fig. VIII-11). During a superimposed deformation of say F_2, the F_1 finite strain ellipsoid will be deformed into a new ellipsoid (Ramsay 1967, pp. 165–167). This new ellipsoid will change in shape from increment to increment until at the end of F_2 the finite strain ellipsoid, related to F_1 plus F_2, is reached. During this process the XY plane related to F_1 is rotated to the XY plane related to F_1 plus F_2. The S_1 slaty cleavage, which is, if not shortened during the rotation, still intact and probably even more pronounced, should be called S_{1+2}, since it is the result of both deformations. If S_1 is shortened during its rotation it seems more probable that it is crenulated and that a new crenulation cleavage (S_2) is formed, probably along the XY plane of F_2 strain.

In nature slaty cleavage is often accompanied by some kind of layering, however vague this may be. In such cases the problem arises that during progressive rotational deformation, or during superimposed deformation the layer-

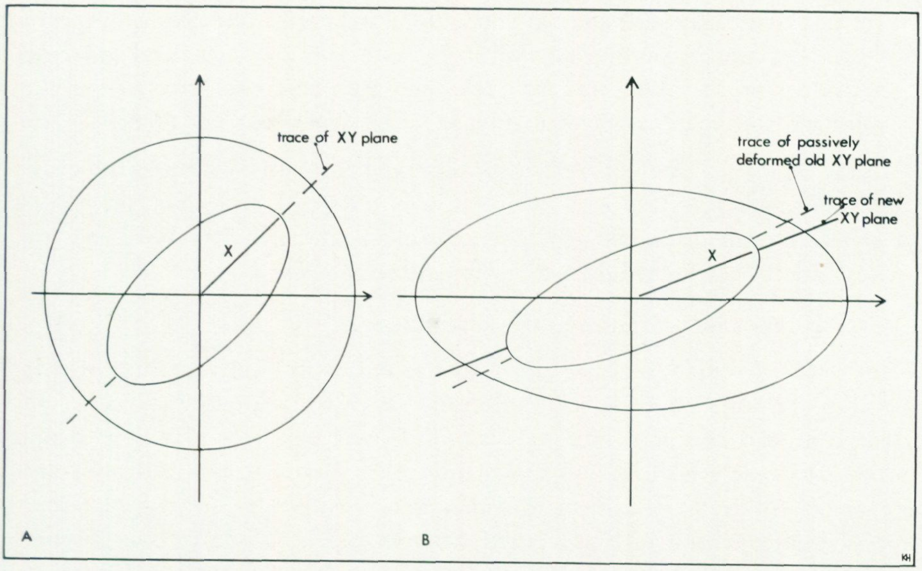


Fig. VIII-11. The homogeneous deformation of an ellipse; A, before deformation; B, after deformation. Note that the new long axis has an orientation different from the passively deformed old one, after Ramsay (1967).

ing is deformed passively, whereas the cleavage should follow the new XY plane. Since the angle which should be the result of these different rotations is rarely observed in natural samples, it may be that in such cases the cleavage does not manage to remain always parallel to the XY plane of the finite strain ellipsoid.

The importance of straight rotation axes

Wilson was the first to point out the significance of straight rotation axes (Wilson 1971, pp. 250, 251).

The concept is clear: a garnet, or the surrounding S-plane can rotate from increment to increment of deformation around different axes or around one single axis (see p. 35). Only in the last case will a garnet produce something like a straight rotation axis, in the first case this axis is a curved line. If both the garnet and the surrounding S-plane rotate, then they have to rotate around the same axis to produce a garnet with a straight rotation axis. Wilson determined under which circumstances S-planes rotate around straight axes during progressive and/or superimposed deformation. To plot the change in attitude of a plane during three dimensional progressive deformation Flinn (1962, p. 394) introduced the concept of a "structural movement path", the locus of the pole to the plane plotted on a stereographic projection. Wilson (1971, p. 251) states:

"If a structural movement path lies along a great circle of the projection then the plane is rotating around a fixed axis. If the structural movement path does not lie along a great circle then the axis of rotation of the plane will vary from increment to increment."

In Fig. VIII-12 a number of structural movement paths is plotted for passive planes during certain types of strain (after Flinn 1962). In four different cases of strain Wilson discusses which structural movement paths lie along great circles and hence which planes rotate around fixed axes.

Case 1 stands for irrotational strain with $0 < k < \infty$, ($k = \frac{a-1}{b-1}$, $a = \frac{X}{Y}$, $b = \frac{Y}{Z}$ Flinn 1962). All planes, either active or passive, with their pole in a principal plane of the superimposed strain ellipsoid, have a structural movement path along a great circle according to Wilson. This is true, but planes with poles in the XY plane are shortened and therefore, as discussed above, irrelevant to the garnet rotations. It is clear that this case is only relevant as far as superimposed deformations are concerned, otherwise the S-plane will not rotate at all.

Case 2 represents irrotational strain in which $k=0$ or $k=\infty$. All planes move around fixed axes, but deformations with these k-values are very exceptional.

In case 3 rotational strain is discussed, in which there is parallelism of the axis of rigid body rotation and one of the axes of the finite and incremental strain ellipsoids. Planes orientated so as to rotate around the same axis as the rigid body rotation, obviously have a fixed rotation axis (case 3a). According to Wilson, planes with poles in another principal plane will also rotate around straight axes (case 3b). This holds for a passive plane, but not for a slaty cleavage plane, which after every increment of strain assumes the new position of XY in the finite strain ellipsoid.

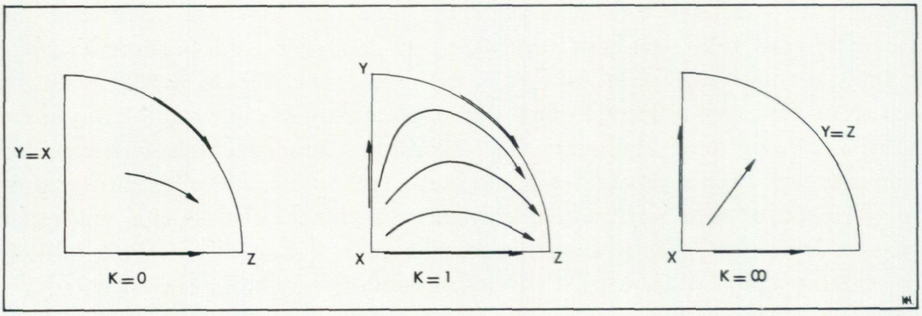


Fig. VIII-12. Structural movement paths of poles of planes for $k=0$, 1, and ∞, after Flinn (1962).

Case 4 deals with simple shear. According to Wilson, all planes will rotate around their straight intersection with the shear plane, but this again holds only for passive planes. Only S-planes which are perpendicular to the *ac* plane of shear will rotate around a straight axis (*b*).

As case 2 is very exceptional and case 4 is reduced to a special case of case 3a, only case 1 and case 3a remain relevant to the problem: case 1 for superimposed and case 3a for progressive and superimposed deformation.

What happens to slaty cleavage during simple shear?

It may be concluded from the foregoing that simple shear or a more general rotational deformation, in which the tectonic *b* axis and the rigid body rotation axis are parallel to the garnet rotation axis, is the most probable mechanism for explaining garnet rotations. Irrotational strain can only produce small relative rotations (up to 45°) in the accidental case that an old S-plane has a special position in a superimposed strain ellipsoid (case 1). Rosenfeld (1970) gives all the details of the rotation of garnets in the mechanism proposed.

Since the principles of simple shear also apply for the more general rotational deformations under conditions mentioned (see also Rosenfeld 1970), it seems relevant to investigate what happens to a slaty cleavage plane during simple shear. As above, it is again assumed in this discussion that slaty cleavage remains parallel to the XY plane of finite strain. In Fig. VIII-13 the change in position of the XY-plane is illustrated for progressive and various cases of superimposed simple shear. A illustrates progressive simple shear in which a slaty cleavage is formed and rotated. α is the angle between the shearplane and the cleavage (XY). B, C and D illustrate how an old slaty cleavage can be reoriented during superimposed simple shear. Only cases in which the old cleavage is perpendicular to *ac* are relevant, so that the ellipses drawn in the undeformed state are sections through the old strain ellipsoids perpendicular to XY and containing Z. The short axis of these ellipses is Z; the long one, X' ($X \geq X' \geq Y$), indicates the slaty cleavage. Cases in which $\alpha > 90°$ are not considered because of the shortening involved and probably allied crenulation of the slaty cleavage.

In case A, which may be considered the general case for garnet rotations during one progressive deformation, it is assumed that a slaty cleavage begins to develop after some 30 % of shortening (Ramsay 1967, p. 180; Cloos 1947). This is a fairly rough statement; little is known about the subject and the percentage will vary considerably in different rocks. If this statement is correct it would mean that a cleavage is formed at an angle α of approx. 40° to the shear plane, after a shear γ of approx. 0.4 (Ramsay 1967, 3.10). During progressive deformation this cleavage can rotate (Ω_s) up to 40° in the same sense as the garnets, but in general the rotation during garnet growth will be less, since the garnet growth need not overlap the entire deformation and since infinite

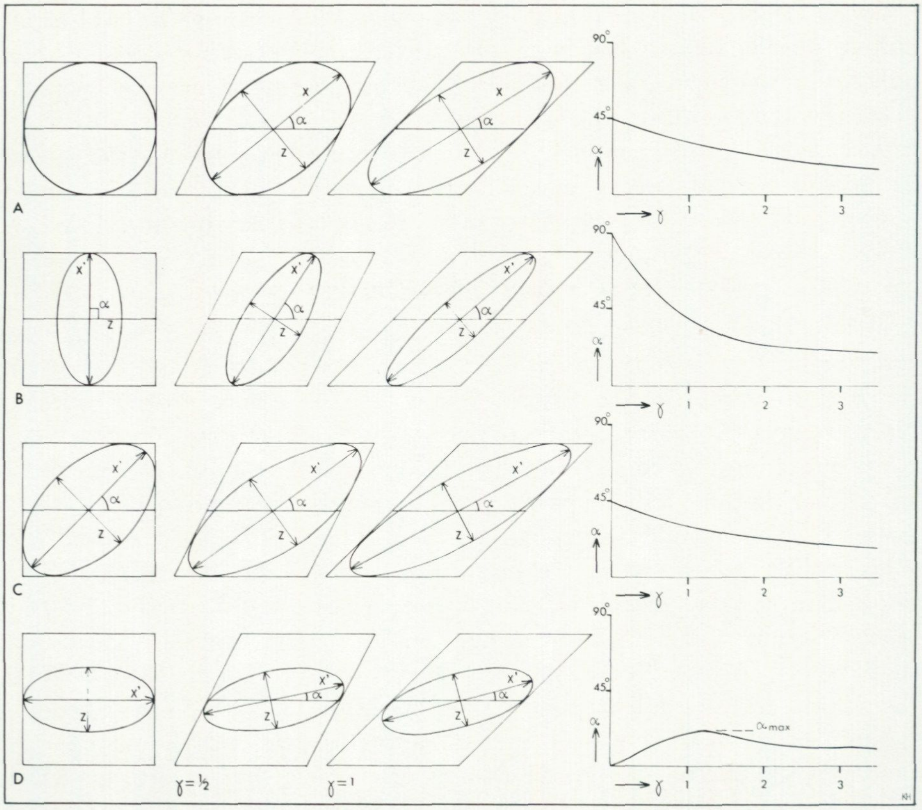


Fig. VIII-13. Two-dimensional representation of progressive simple shear in undeformed and in previously deformed rocks. A, simple shear of undeformed rock; X is the trace of the XY plane of finite strain. α is the angle between this plane and the shear plane. B to D, simple shear of previously deformed rocks. The ellipses drawn in the undeformed state represent sections perpendicular to XY through the finite strain ellipsoid belonging to earlier deformation. X' indicates the trace of XY (the slaty cleavage). α is the angle between XY and the shearplane.

deformation is required to rotate the S-plane through 40°, the maximum Ω_s value possible. However, it seems probable that in many natural cases the rotation angle Ω_i measured is up to approx. 30° too small, according to Rosenfelds: $\Omega_e = \Omega_s + \Omega_i$. The exact value of Ω_s can probably not be established in natural examples.

Cases B, C, and D of superimposed simple shear seem to be more accidental because of the special position the second stress field has to occupy. Nevertheless, it is interesting to see what happens to the angle α between S_1 and the shear plane. In all cases between C and D the graph first shows a slight increase of α, followed by a decrease. The value of α_{max} in case D depends on the shape of the inherited strain ellipsoid; the flatter this ellipsoid, the lower the value of α_{max}.

The graphs of α show that the theoretical maximum rotation of the cleavage possible is $+90^\circ$ (case B) and -45° (case D) if clockwise rotations are counted positive. This conclusion is based on the assumption that slaty cleavage remains always parallel to the XY plane of finite strain, which, as discussed above, bears theoretical problems as soon as the slaty cleavage is accompanied by a material layering. It is therefore unsure whether a slaty cleavage really rotates in an anticlockwise sense in cases like case D. The criticism of mechanisms based on shear along cleavage planes remains nevertheless valid, since it was shown by Ramsay and Graham (1970) that in shear belts a slaty cleavage forms parallel to XY planes and not parallel to the shear plane.

It appears from the graphs A–D that the length of Z decreases during the simple shear deformation. This means that the S-planes are compressed around the garnets, which must result in deflection patterns (Ramsay 1967, 3.10). It is interesting to compare the measured flattening from these deflection patterns with the maximum percentage of shortening calculated from the value of simple shear belonging to the rotation. A fairly rough method was applied for measuring the flattening during and after growth from a deflection pattern (Fig. VIII-14). The method is, however, only used for comparison, not for an exact quantitative analysis.

Diameter *d* of a garnet is compared with the distance *e* between two cleavage planes that just curve around the garnet. The flattening is given by $\frac{d-e}{d}$; if *d* equals 100 the flattening is thus expressed as a percentage.

If the initial and final value of the shear γ are known, the "shortening" of Z can be calculated by means of a formula given by Ramsay (1967, p. 85, 3–67) in which λ_2 stands for Z:

$$\lambda_2 = \frac{\gamma^2 + 2 - \gamma(\gamma^2 + 4)^{1/2}}{2}$$

Ω_e can be estimated by adding 30° or less to the measured rotation angle Ω_i, dependent on the amount of this angle. From $\Omega_c = |\,^1/_2\,\gamma|$ the value of shear

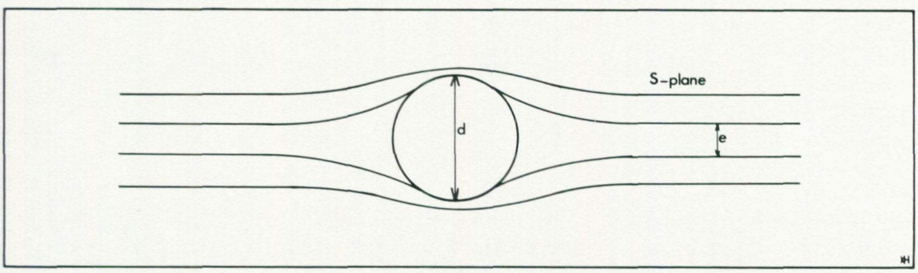


Fig. VIII-14. Diagram illustrating that, if layering or cleavage (S) curves around synkinematic garnets, a rough estimation of the flattening perpendicular to S can be obtained by comparing diameter *d* of a garnet with the distance *e* between two S planes that just curve around that garnet.

γ between nucleation of the garnet and the end of the deformation can then be computed. This value of γ gives, however, not the initial and final values desired, since it is not known how much shear preceded the nucleation. The higher this preceding shear is, the lower is the shortening percentage of Z during the same deformation. The maximum value for the shortening of Z can therefore be calculated by substituting zero for the initial γ. It was argued previously that only after a shear γ of approx. 0.4 a slaty cleavage begins to develop. Since there is generally an S_i present in the center of rotated garnets, proving the presence of a cleavage at the time of nucleation, the value 0.4 seems a better substitute for the minimum initial γ.

If the measured value of the flattening is higher than the calculated shortening of Z, an extra pure shear deformation must be responsible at least for the surplus. If, on the other hand, the value measured is lower, the value substituted for the initial γ is apparently wrong and should be higher. In specific cases this minimum value for the initial γ can be even calculated from the difference.

An example may clarify this procedure:

Suppose a measured rotation angle Ω_i is 120°, and Ω_s is estimated to be 25°.

From $\Omega_e = \Omega_i + \Omega_s$ follows that $\Omega_e = 120° + 25° = 145° \approx 2.5$ radians. Substitution in $\Omega_e = |{}^1/_2\, \gamma|$ gives: $\gamma =$ approx. 5.

Let the initial γ be 0.4, then the finite γ is 5.4.

Substitution of these values for γ in:

$$Z = \frac{\gamma^2 + 2 - \gamma(\gamma^2 + 4)^{1/2}}{2}$$

gives successive values for Z of 0.7 and 0.04, or a shortening of 94 %. If the measured value for the shortening is for example 98 %, there must at least have been an extra pure shear deformation responsible for the 4 % difference. If, on the other hand, the measured value is, for example, 60 %, the initial γ must have been higher than 0.4 or other deformation must have preceded the nucleation of garnet.

CONCLUSIONS

1. If straight rotation axes can be demonstrated, a mechanism of simple shear, either or not accompanied by pure shear, in which rotation of S-planes and garnets takes place around the tectonic *b* axis, is the most probable mechanism for explaining rotational fabrics in garnet crystals.

2. In this process not only the garnets rotate, but also the surrounding slaty cleavage. The rotation of the garnets with respect to the shear plane can be calculated by means of: $\Omega_e = \Omega_i + \Omega_s$.

3. The amount of simple shear during and after garnet growth can be computed by means of the formula: $\Omega_e = |{}^1/_2\, \gamma|$.

4. Comparison of measurements of the flattening, perpendicular to the cleavage, with calculations of the shortening of Z from the shear γ, can give information on the minimum value of extra pure shear or on the minimum amount of deformation preceding the nucleation of garnet.

REFERENCES

GFF = Geologiska Föreningens i Stockholm Förhandlingar
SGU = Sveriges Geologiska Undersökning

AVÉ LALLEMANT, H. G., 1969: Structural and petrofabric analysis of an "Alpine-type" peridotite: the lherzolite of the French Pyrenees. Leidse Geol. Med., v. 42, p. 1–57.

BACKLUND, H. G. and QUENSEL, P. D., 1929: Karta över berggrunden inom Västerbottens fjällområde. SGU Ca 21.

CHRISTIE, J. M., 1963: The Moine thrust zone in the Assynt region, northwest Scotland. Univ. California Pubs. Geol. Sci., v. 40, no. 6, p. 345–440.

– 1965: Moine thrust: a reply. J. Geol., v. 73, p. 677–681.

CLOOS, E., 1946: Lineation, a critical revue and annotated bibliography. Geol. Soc. Am. Mem., 18, reprinted with supplement, 1957.

– 1947: Oolite deformation in South Mountain Fold Maryland. Geol. Soc. Am. Bull., v. 58, p. 843–917.

COX, F. C., 1969: Inclusions in garnet: discussion and suggested mechanisms of growth for syntectonic garnets. Geol. Mag., v. 106, p. 57–62.

DEWEY, J. F., 1965: Nature and origin of kink bands. Tectonophysics, v. 1, p. 459–494.

DIETERICH, J. H., 1969: Origin of cleavage in folded rocks. Am. J. Sci., v. 267, p. 155–165.

DONATH, F. A. and PARKER, R. B., 1964: Folds and folding. Geol. Soc. Am. Bull., v. 75, p. 45–62.

DU RIETZ, T, 1935: Peridotites, serpentines and soapstones of northern Sweden. GFF 57, p. 133–260.

– 1938: The injection metamorphism of the Muruhatten region. SGU C 416.

– 1941: Nyare undersökningar inom Remdalens malmtrakt och dess omgivningar. SGU C 439.

DURNEY, D. W., 1972: Solution-transfer, an important geological deformation mechanism. Nature, v. 235, p. 315–317.

EISBACHER, G. H., 1970: Deformation mechanics of mylonitic rocks and fractured granites in Cobequid Mountains, Nova Scotia, Canada. Geol. Soc. Am. Bull., v. 81, p. 2009–2020.

FLEUTY, M. J., 1964: The description of folds. Proc. Geol. Ass., v. 75, p. 461–493.

FLINN, D., 1962: On folding during three-dimensional progressive deformation. Quart. J. Geol. Soc., p. 385–433.

– 1965: On the symmetry principle and the deformation ellipsoid. Geol. Mag., v. 102, p. 36–46.

GLASS, J., (in prep.): Some observations on the petrology and metamorphic history of the Marsfjällen area, Caledonides of Västerbotten, Sweden.

VAN DER HARST, T. H., 1956: Geology and petrology of the Borga region, Västerbotten, Sweden. Diss. Amsterdam.

HENLEY, K. J., 1970: The structural and metamorphic history of the Sulitjelma region, Norway, with special reference to the nappe hypothesis. Norsk Geol. Tidsk., v. 50, nr. 2.

HOBBS, B. E., 1968: Recrystallization of single crystals of quartz. Tectonophysics 6, p. 353–401.

HOEPPENER, R., 1956: Zum Problem der Bruchbildung, Schieferung und Faltung. Geol. Rundschau, Bd. 45, Hf. 2, p. 247–283.

JANSSEN, F. L., 1953: unpublished report. Amsterdam.

JOHNSON, M. R. W., 1967: Mylonite zones and mylonite banding. Nature, v. 213.1, p. 246–247.

– 1965: The Moine thrust: a discussion. J. Geol., v. 73, p. 672–676.

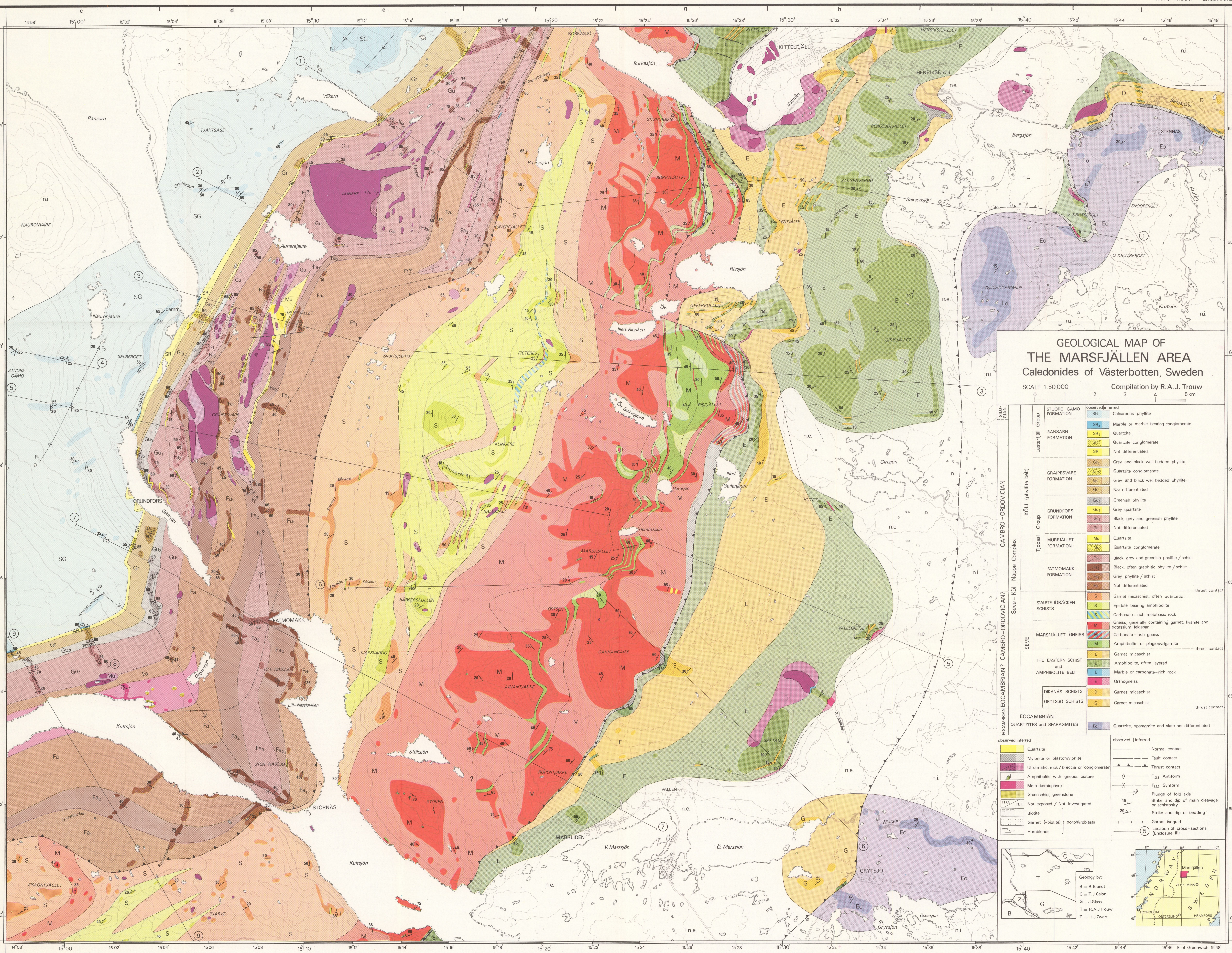
Karta över Sveriges Berggrund, 1958, skala 1:1,000,000. SGU Ba 16.

DE KEYZER F., 1952: unpublished report. Amsterdam.

KNILL, J. L., 1960: A classification of cleavages, with special references to the Craignish district of the Scottish Highlands. Internat. Geol. Cong., 21st Copenhagen 1960, Rept. pt. 18, p. 317–325.

Kulling, O., 1933: Bergbyggnaden inom Björkvattnet–Virisen-området i Västerbottensfjällens centrala del. GFF 55, p. 167–422.
– 1942: Grunddragen av fjällkedjerandens bergbyggnad inom Västerbottens län. SGU C 445.
– 1955: Den kaledoniska fjällkedjans berggrund inom Västerbottens län. Part 2 in Beskrivning till berggrundskarta över Västerbottens län. SGU Ca 37, p. 101–296.
– 1958: Om begreppen Voitjakonglomerat, Rokonglomerat och Mesketlava i Västerbottensfjällen. GFF 80.
Kvale, A., 1953: Linear structures and their relation to movements in the Caledonides of Scandinavia and Scotland. Geol. Soc. London Quart. Jour., v. 109, p. 51–73.
Langheinrich, G., 1964: Vergleichende Untersuchungen über das Verhältnis der Schieferung zur Faltung unter Berücksichtigung des Stockwerkproblems. N. Jb. Geol. Paläont., Abh. 120, p. 41–80.
Lapworth, C., 1885: The Highland controversy in British geology. Nature, v. 32, p. 558.
Lindström, M., 1955: A tectonic study of Mt. Nuolja, Swedish Lapland. GFF 77, H. 4, p. 557–566.
Magnusson, N. H., Thorslund, P., Brotzen, F., Asklund, B. and Kulling, O., 1960: Description to accompany the map of the Pre-Quaternary rocks of Sweden. SGU Ba 16.
Michel, H., 1950: Geology and petrology of the Borkafjäll region. Diss. Amsterdam.
Milnes, A. G., 1971: A model for analysing the strain history of folded competent layers in deeper parts of orogenic belts. Eclogae geol. Helv., v. 64–2, p. 335–342.
Möckel, J. R., 1969: Structural petrology of the garnet-peridotite of Alpe Arami. Leidse Geol. Med., v. 42, p. 61–130.
Nicholson, R. and Rutland, R. W. R., 1969: A section across the Norwegian Caledonides; Bodø to Sulitjelma. Norges Geol. Unders., no. 260, 86 p.
Noble, D. C. and Eberly, S. W., 1964: A digital computer procedure for preparing beta diagrams. Am. Jour. Sci., v. 262, p. 1124–1129.
Peacey, J. S., 1961: Rolled garnets from Morar, Inverness-shire. Geol. Mag., v. 98, p. 77–80.
Plessman, W., 1964: Gesteinslösung, ein Hauptfaktor beim Schieferungsprozess. Geol. Mitt., v. 4, p. 69–82.
Powell, D. and Treagus, J. E., 1967: On the geometry of S-shaped inclusion trails in garnet porphyroblasts. Mineral. Mag., v. 36, p. 453–456.
– 1970: Rotated fabrics in metamorphic minerals. Mineral. Mag., v. 37, p. 801–814.
Quensel, P., 1960: Beskrivning till karta över berggrunden inom Västerbottens fjällområde. SGU Ba 21.
Ramberg. H., 1964: Selective buckling of composite layers with contrasted rheological properties, a theory for simultaneous formation of several orders of folds. Tectonophysics, v. 1, p. 307–341.
– 1966: The Scandinavian Caledonides as studied by centrifuged dynamic models. Bull. Geol. Inst. Univ. Uppsala, v. 43–1.
Ramsay, J. G., 1962: The geometry and mechanics of formation of "similar" type folds. J. Geol., v. 70, p. 309–327.
– 1967: Folding and Fracturing of Rocks. McGraw-Hill, New York, N.Y., 568 p.
– and Graham, R. H., 1970: Strain variation in shear belts. Can. J. Earth Sci., v. 7, p. 786–813.
Rickard, M. J., 1961: A note on cleavages in crenulated rocks. Geol. Mag., v. 98, p. 324–332.
de Roever, W. P., 1957: Sind die Alpinotypen Peridotitmassen vielleicht tektonisch verfrachtete Bruchstücke der Peridotitschale? Geol. Rundschau, v. 46, p. 137–146.
Rosenfeld, J. L., 1970: Rotated garnets in metamorphic rocks. Geol. Soc. Am., Spec. Paper 129.
Schmidt, W., 1918: Bewegungsspuren in Porphyroblasten Kristalliner Schiefer. S. B. Akad. Wiss. Wien, Abt. 1, v. 127, p. 293–310.
de Sitter, L. U., 1956: Structural Geology, 2nd ed., 1964, McGraw-Hill, New York, N.Y., 551 p.
Spry, A., 1963: The origin and significance of snowball structure in garnet. J. Petrol., v. 4, p. 211–222.
– 1969: Metamorphic textures. Pergamon Press, London, 350 p.
Strand, T., 1961: The Scandinavian Caledonides. A review. Am. Jour. Sci., v. 259, p. 161–172.
– and Kulling, O., 1972: Scandinavian Caledonides. Wiley-Interscience, London, New York.
Talbot, J. L., 1965: Crenulation cleavage in the Hunsrückschiefer of the Middle Mozelle region. Geol. Rundschau, v. 54, p. 1026–1043.
– and Hobbs, B. E., 1968: The relationship of metamorphic differentiation to other structural features at three localities. J. Geol., v. 76, p. 581–587.

VOLL, G., 1960: New work on petrofabrics. Liverpool Manchester Geol. Jour., v. 2, p. 503–567.
WHITTEN, E. H. T., 1966: Structural Geology of folded rocks. Rand-McNally & Co, Chicago, Illinois, 663 p.
WILLIAMS, P. F., 1972: Development of metamorphic layering and cleavage in low grade metamorphic rocks at Bermagui, Australia. Am. Jour. Sci., v. 272, p. 1–47.
WILSON, G., 1961: The tectonic significance of small scale structures and their importance to the geologist in the field. Ann. Soc. Geol. Belg., v. 84, p. 423–548.
WILSON, M. R., 1971: On syntectonic porphyroblast growth. Tectonophysics, v. 11, p. 239–260.
– 1972: Strain determination using rotational porphyroblasts, Sulitjelma, Norway. J. Geol., v. 80, p. 421–431.
ZACHRISSON, E., 1964: The Remdalen Syncline. SGU C 596.
– 1969: Caledonian geology of northern Jämtland–southern Västerbotten. SGU C 644.
– 1971: The structural setting of the Stekenjokk ore bodies, Central Swedish Caledonides. Econ. Geol. v. 66, p. 641–652.
VAN ZUUREN, A., 1969: Structural petrology of an area near Santiago de Compostella (NW Spain). Leidse Geol. Med., v. 45, p. 1–71.
ZWART, H. J. and OELE, J. A., 1966: Rotated magnetite crystals from the Rocroi Massif (Ardennes). Geol. en Mijnb., v. 45, p. 70–74.



För spridning godkänd i Rikets allmänna kartverk 1973.02.27

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CROSS-SECTIONS through THE MARSFJÄLLEN AREA

CALEDONIDES OF VÄSTERBOTTEN, SWEDEN

by R.A.J. TROUW

CROSS-SECTIONS 1,3,5,7,9 : SCALE 1:50,000

CROSS-SECTIONS 2,4,5A,8 : SCALE 1:10,000

- Lasterfjäll Group
- Tjopasi Group (general)
- Quartzite / quartzite conglomerate
- Black, graphitic phyllite

Köli sequence (phyllite belt)

- Svartsjöbäcken Schists
- Marsfjället Gneiss

Seve sequence

- Schist from the eastern schist and amphibolite belt
- Orthogneiss

Seve sequence

- Eocambrian quartzite and sparagmite
- Marble or cabonate-rich rock
- Amphibolite or other metabasite
- Ultramafic rock

- Normal contact
- Thrust or fault contact
- $S_{1,2}$ Cleavage

The location of cross-sections 1–9 is indicated on the map (Enclosure I)

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