STRUCTURAL ANALYSIS OF DIFFERENTIATED SCHISTS AND GNEISSES IN THE TAALINTEHDAS AREA, KEMIÖ ISLAND, SOUTHWEST FINLAND

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Structural analysis of the Svecokarelic supracrustal schists and gneisses from around Taalintehdas, SW Kemiö Island, SW Finland, has been carried out and a preliminary interpretation is given of the regional structure of Kemiö Island in terms of the fold generations determined in the Taalintehdas area. The first generation of isoclinal folds is established by means of overprinting relationships in the schists. The second generation is marked by tight to isoclinal folds which further transposed earlier layering and foliation. The dominant N70E trending transposition foliation (S_2) is openly folded by third generation folds with a N35E trending axial plane crenulation cleavage. Parasitic folds abound in the core of large scale F_3 folds; S_3 transposition foliation is developed locally. This distinctive tectonic layering is folded by F_4 chevron folds with locally developed EW axial plane foliation. The F_4 folds are restricted to regions where earlier foliations were oriented NS after the third folding phase. The rocks were metamorphosed under amphibolite facies conditions resulting in the blastesis of sillimanite, andalusite, cordierite and garnet in the schists. The relationship between deformation and metamorphism is discussed.

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Introduction

This paper deals with a structural analysis of a small area of the Svecokarelic Kemiö-Orijärvi belt in SW Finland, previously mapped and extensively described by Eskola (1914). The work is part of a larger programme on the evolution of the Kemiö-Orijärvi belt.[2] Detailed structural analysis of the area around Taalintehdas (Dalsbruk) was undertaken to unravel the complex geological history of this belt. The rocks described are supracrustal mica schists and gneisses with well preserved mesoscopic structures. Being the oldest rock type so far recognized in the Kemiö-Orijärvi belt, the supracrustal rocks are a potential source of the most complete information on the deformation history. First the geological setting, petrography and structures of the schists and gneisses are described, then the microstructures and the relative ages of metamorphic mineral growth are discussed.

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[2] This investigation is part of a petrological project that has been carried out in the Kemiö area since 1974 by staff members and students of the Free University of Amsterdam under the direction of Dr. László Westra.

Finally an attempt is made to synthesize the structure and history of the Taalintehdas area and Kemiö Island.

Geological setting

Kemiö-Orijärvi Belt

The Kemiö-Orijärvi belt is part of the Svecokarelic orogen of Early Proterozoic age. The evolution of the belt has been described in terms of geosynclinal sedimentation of supracrustal rocks (leptites); deformation with concurrent intrusion of »synkinematic» rock suites (the »oligoclase granites» of Eskola (1914), later termed »trondhjemites» and »granodiorites» by Simonen (1960)) including minor gabbroic and peridotitic rocks; continuing deformation and finally emplacement of migmatizing »late kinematic» microcline granites in the closing stages of the orogeny (Simonen, 1971).

Kemiö Island

On Kemiö Island (Fig. 1) the central part of the belt consists of a series of fine-grained supracrustal gneisses with intercalations of amphibolites, and minor amounts of schists, marbles and associated calc-silicate rocks. Elongated bodies of amphibole gneiss, amphibolite and cordierite-anthophyllite rock are widespread, and some of them provide evidence of volcanic origin (pillow lavas, agglomerates). Their mineral parageneses are characteristic of amphibolite facies metamorphism.

The external zones of the belt contain synkinematic infracrustals which are mostly granitic of granodioritic gneisses in the south and tonalitic gneisses in the north. The latter form a series of gabbro-tonalite in relative order of intrusion, with cross-cutting minor granodioritic material. Deformation of supra- and infracrustal rocks was intense and resulted in a generally steep dipping east-west foliation with transposition of layering and intercalated rock bodies parallel to this foliation. Foliations deviating from the general EW trend occur west and north of Taalintehdas and by the dome-shaped structure encirling the islet of Bergö. Large batholiths of non-gneissose late-kinematic microcline granite are encountered north and south of the belt.

Taalintehdas area

The southern part of the area around Taalintehdas is composed of schists separated by a sharp lithological boundary from the Masugnsträsk gneisses in the north (Fig. 2). Intricately foldel quartz veinlets are common in both schists and gneisses. Pegmatite and aplite veins and dykes also occur. These veins and dykes are boudinaged and folded, but some cross-cut older structures and are apparently younger. Pegmatites are more abundant in the schists. The zonal pegmatites consist of microcline, plagioclase, biotite, muscovite, quartz, and occasionally beryl.

The *Taalintehdas schists* are layered (10—30 cm scale), dark grey or brownish schists with gneissic intercalations up to 20 cm thick. The schists are fine grained (0.1—0.3 mm) and consist essentially of quartz, muscovite, biotite and oligoclase. Cordierite porphyroblasts (or pseudomorphs) and fibrolitic sillimanite are common and occasionally andalusite porphyroblasts are observed. Garnet porphyroblasts are restricted to a layer of biotite schist which occurs in the sequence.

The *Masugnsträsk gneisses* are dominantly fine-grained (ca. 0.1 mm) quartz-feldspar gneisses with isolated, intricately folded patches of medium-grained (ca. 1 mm) biotite schist and amphibolite intercalations up to 10 cm thick. The constituent minerals of the gneisses are quartz, microcline, oligoclase,

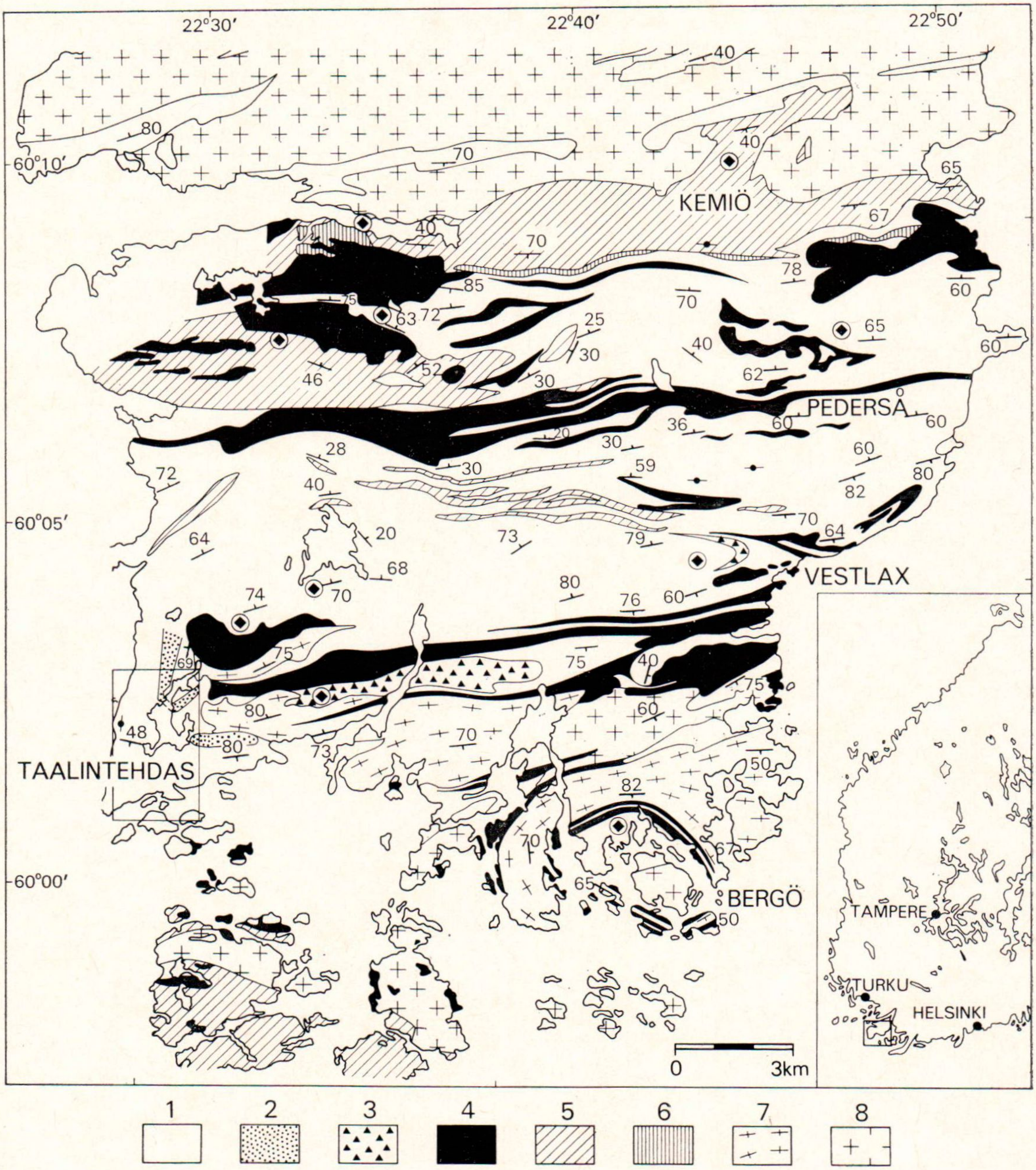
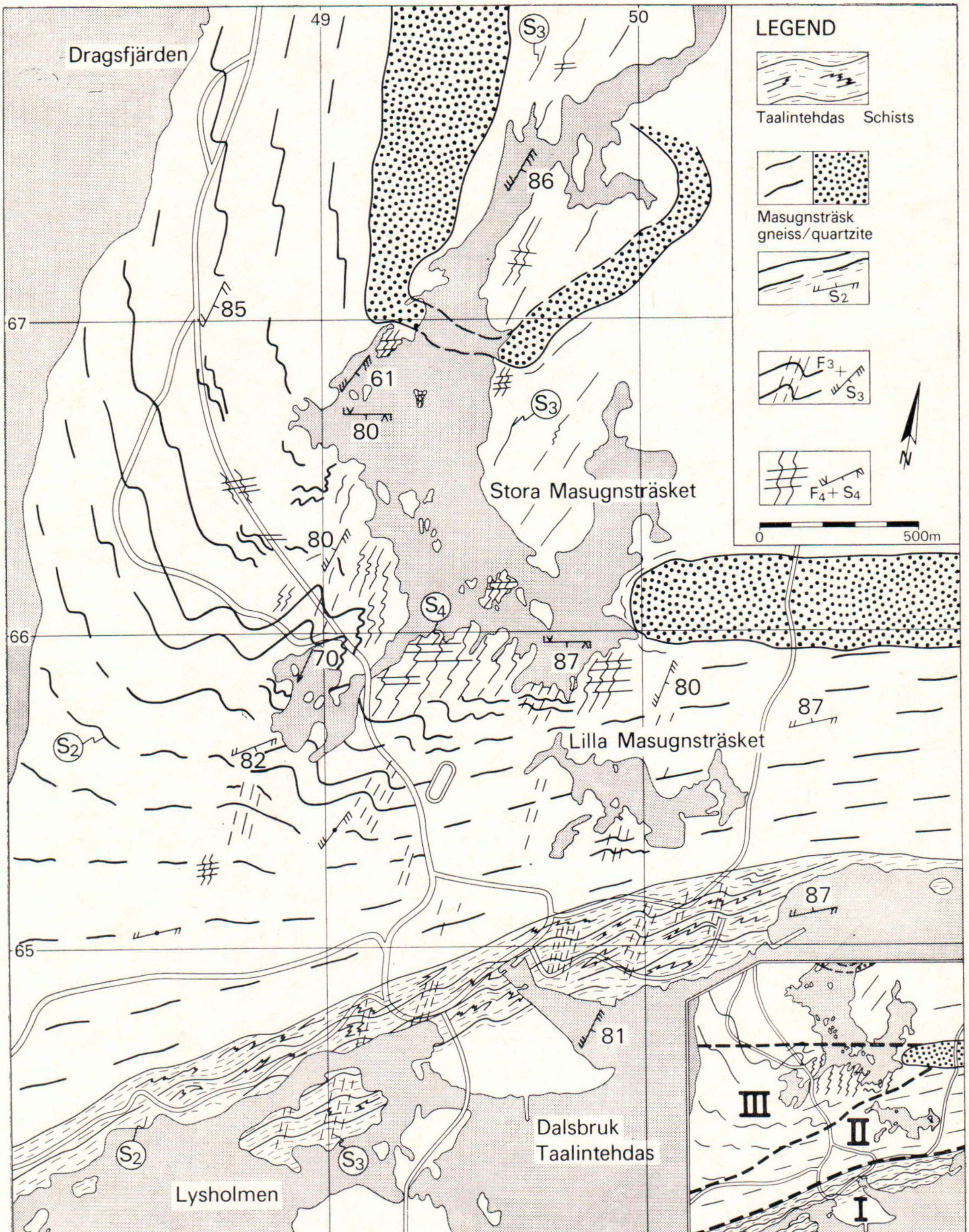


Fig. 1. Simplified geological map of Kemiö Island: 1. metapelite and metapsammite; 2. quartzite; 3. pillow basalt and pyroclastics; 4. metagabbro and amphibolite; 5. tonalitic gneiss and amphibole gneiss; 6. granodioritic gneiss and augengneiss; 7. granitic and granodioritic gneiss; 8. microcline granite. Locations where structural analysis has been carried out are indicated by a black square.

biotite and muscovite. Garnet porphyroblasts are encountered in the biotite schists. The essential constituents of the amphibolite are hornblende, quartz, and sericitized plagioclase. Quartzite bodies, 250 m wide in outcrop, are included in the gneiss series.


LEGEND
Taalintehdas Schists
Masugnsträsk gneiss/quartzite
S_2
F_3 + S_3
F_4 + S_4
0 500m
Dragsfjärden
Stora Masugnsträsket
Lilla Masugnsträsket
Dalsbruk Taalintehdas
Lysholmen
49
50
67
66
65
III
II
I

Structural analysis [1]

With the aid of overprinting relationships, four fold generations can be distinguished in the supracrustal rocks around Taalintehdas. In addition to local analysis of refolded fold structures, form surface maps were prepared to gain insight into the geometrical relationships of the mesoscopic structures and to facilitate macroscopic interpretation (Hobbs et al., 1976). The area can be divided into three structural domains, distinguished on the basis of the dominating structural elements (Figs. 2 and 10).

Taalintehdas schists

The most prominent structural feature in the Taalintehdas schists is a pervasive N70E striking foliation defined by transposed layering and preferred orientation of micas and other minerals. Intrafolial folds associated with this foliation are isoclinal and are refolded by tight to isoclinal asymmetric F_2 folds related to the pervasive schistosity, S_2 (Fig. 3). Locally this transposition foliation has been deformed into open folds, and a new crenulation cleavage (S_3) with a N35E strike is present. The form surface map of an exposure in the schists shows an example of the mesoscopic structures observed (Fig. 4). Since fold axes, mineral lineations and foliations are subvertical (Fig. 10), the map is a reasonable approximation to a structural profile (Figs. 2 and 4).

Fig. 2. Interpretation of macroscopic structure of the Taalintehdas area. The large scale fold in the centre of the map is an F_3 fold. Inset shows structural domains: I. Taalintehdas schists; II. Masungnsträsk gneisses characterized by predominance of S_2 transposition foliation; III. Masungnsträsk gneisses with dominant F_3 parasitic folding and widespread occurrence of F_4 structures.

F_1 structures

The existence of F_1 folds is evident only where they are overprinted by F_2 folds. Where such overprinting structures are absent it is impossible to distinquish F_1 and F_2 folds, since they have the same orientation and essentially the same style. F_1 folds, however, are generally somewhat tighter, but this might be true only where they are refolded. The amplitude of F_1 folds rarely exceeds 50 cm (Figs. 3 and 5). Because of intense later deformation it has not been possible to study the relationship of S_1 and S_0 (presumed initial layering). The original appearance of S_1 foliation is obscured by F_2 structures. S_1 can be recognized as a regular schistosity defined by preferred orientation of mica or as a crenulation cleavage in the hinges of some open F_2 folds. The foliation folded by F_2 commonly constitutes irregular differentiated layering (quartz rich / mica rich). Numbers of the quartz veinlets which invaded the schists show F_1 folding. These observations suggest that F_1 deformation may have been associated with metamorphic differentiation.

F_2 structures

Tight or isoclinal F_2 folds (cm—m scale) with predominantly dextral vergence are widespread in the schists. Their hinges are sharp or rounded. The layering folded by F_2 folds is discontinuous, and rootless intrafolial

[1] Terminology of axial plane foliations in this paper is in accordance with that used by Hobbs et al. (1976), Chapter 5. Differentiated crenulation cleavage is similar to the layered crenulation cleavage used by Marlow and Etheridge (1977). The terms S_1, S_2, S_3, ... refer to the axial plane schistosity of fold generations F_1, F_2, F_3, ... These terms have no genetic implication. Thus, earlier foliations transposed parallel to the S_2 axial plane schistosity are also referred to as S_2, and not as S_{2+1} or S_{2+1+0}, since it is generally impossible in the field to distinguish between newly developed axial plane cleavages and older foliations transposed parallel to them.

Fig. 3. Overprinting relationships in quartz-mica schist. Rootless, isoclinal fold is interpreted as an F_1 structure because of the vergence of surrounding F_2 folds. S_3 crenulation (parallel to the pencil) trends NNE. Location 65.23—49.98 (see grid in Fig. 2).

folds are common. Layering is seen on all scales to be transposed parallel to the S_2 axial plane schistosity. A penetrative steeply plunging mineral lineation (L_2) is approximately parallel to B_2 fold axes. The nature of the microfabric is strongly dependent on the relative amounts of quartz and mica. The quartz-rich material exhibits regular mica schistosity or differentiated crenulation cleavage with widely spaced thin M-(mica)-domains separated by thick QF (quartz-feldspar)-domains (cf. Marlow and Etheridge, 1977). The M films are irregularly distributed in the rock and some end abruptly. The mica orientation in the M domains is subparallel to S_2, whereas oblique or normal orientations prevail in the QF-domains (Fig. 6). In rocks with subequal amounts of quartz and mica the M domains of the differentiated crenulation cleavages are thick and irregularly distributed anastomosing domains composed of muscovite with (001) cleavage planes subparallel to S_2, and minor amounts of biotite with irregularly oriented (001) cleavage planes. In mica-rich rocks kinking of mica aggregates with some local differentiation into mica-rich and quartz-rich zones is common.

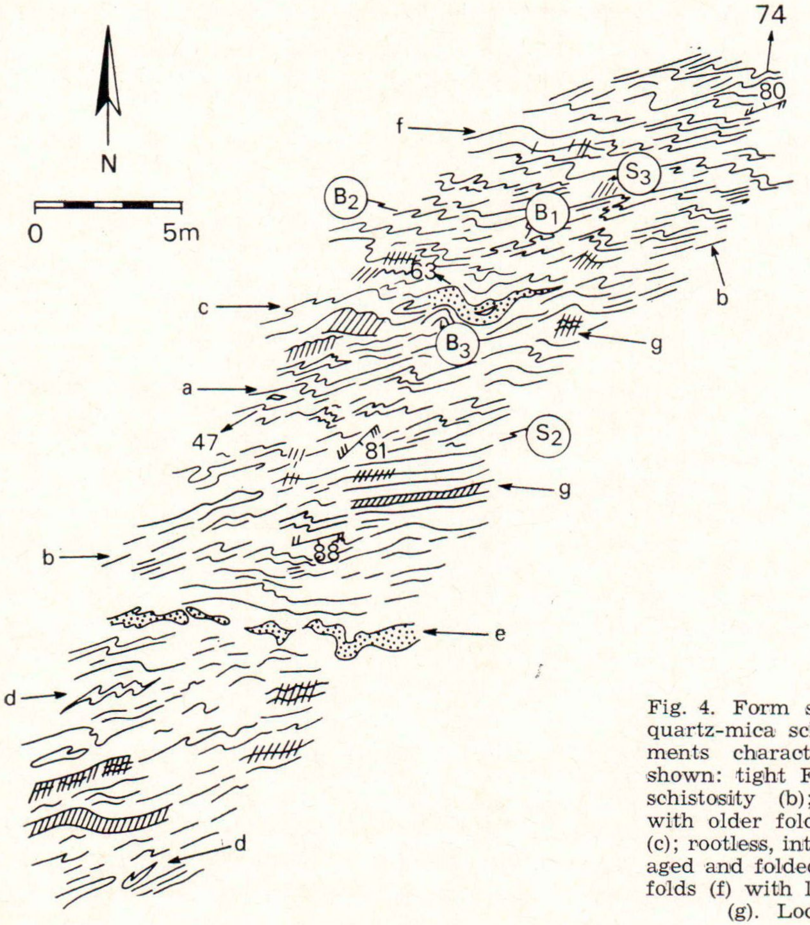


Fig. 4. Form surface map of exposure in quartz-mica schist. Several structural elements characteristic of the schists are shown: tight F_2 folds (a) with penetrative schistosity (b); overprinting relationships with older folds are visible in their limbs (c); rootless, intrafolial folds (d) and boudinaged and folded gneissic layers (e); open F_3 folds (f) with local S_3 crenulation cleavage (g). Location 64.99—49.78.

F_3 structures

The imprint of F_3 folding on the schists is localized on mesoscopic and on microscopic scale. In the field open dextral folds with wavelengths of the order of metres occur at regular intervals of 50—100 m. In many places is the only F_3 phenomenon observed crenulation cleavage, which may or may not be associated with mesoscopic folding. Fold hinges in quartz-rcih rocks are rounded, whereas micaceous rocks show kink bands and, less frequently, chevron folds.

On microscopic scale F_3 deformation has formed open crenulations and local crenulation cleavage with differentiation into M and QF domains (Fig. 7; cf. Marlow and Etheridge, 1977, Fig. 2). In places where S_2 is a differentiated crenulation cleavage, S_3 tends to develop as a differentiated crenulation cleavage of the same type. Where both cleavages make a small angle it is difficult to distinguish between them.

F_4 structures

Open folds, kink folds, kink bands, and crenulations deforming F_2 and F_3 structures occur in restricted areas in the schists. Kink structures are common on thin section scale, and are restricted to thin mica films. Folds

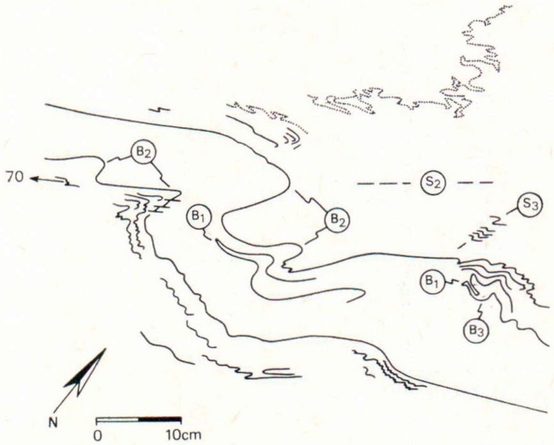


Fig. 5. Mesoscopic overprinting relationships in quartz-mica schist and folded quartz veinlets (dotted lines). Location 64.50—65.18 (E coast of Lysholmen).

and kinks seem to occur only in foliated micaceous materials striking close to north-south. These structures have been correlated

Fig. 6. S_2 differentiated crenulation cleavage with quartz-rich and mica-rich domains. The crenulated S_1 mica cleavage is visible in the quartz-rich domains. (D522a; location 64.50—65.18, same as Figure 4).

with the fourth folding phase established in the Masugnsträsk Gneisses to the north.

In vertical outcrops a horizontal crenulation lineation is observed in the S_2 foliation plane. It crenulates the L_2 mineral lineation, but it could not be correlated conclusively with F_3, F_4 or any other deformation event.

Masugnsträsk gneisses

The Masugnsträsk gneisses have been divided into two structural domains: a southerly domain, (domain II in Fig. 2) characterized by a pervasive foliation that is correlated with the S_2 in the Taalintehdas schists, and a northerly domain (domain III in Fig. 2) in which F_3 and F_4 folds are prominent.

Southerly domain: F_1—F_4 structures

Rootless isoclinal folds are scarce in the gneisses. In places fold-like structures, which may be F_1 or F_2, are defined by lithological contacts such as streaks and intercalations of biotite schist and boudinaged amphibolite. Layering is generally transposed parallel to S_2. The S_2 foliation is defined partly by irregular differentiated layering with thin M films separated by thick QF domains with local remnants of crenulations. More regular differentiated S_2 layering with penetrative mica films has developed locally. Some outcrops exhibit a weak F_3 crenulation and in very rare cases F_3 folds and an S_3 crenulation cleavage can be detected. A few F_4 kink bands may occur in such localities.

Northerly domain: F_1—F_3 structures

The northerly domain is characterized by the pronounced development of F_3 and F_4 structures. Generally three foliations (S_2, S_3,

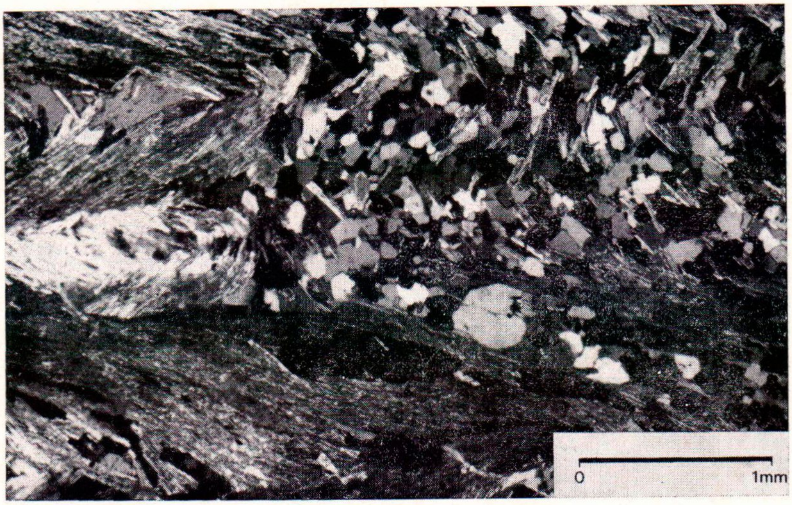


Fig. 7. F_3 microstructures in quartz-mica schist. Mica aggregates are kinked, whereas a crenulation has developed in the quartz-rich domain. Note the serrated kinkband boundaries in the mica and the strain-free nature of both mica and quartz. Crossed nicols. (F501c; location 65.08—49.79)

S_4) can be observed. Notwithstanding the imprints of later deformations, S_2 can be recognized on microscopic scale as somewhat differentiated, penetrative schistosity, regular biotite schistosity, differentiated layering, or foliation marked by elongate aggregates and streaks of biotite.

F_3 folds are open to tight and are related to the NS trend of the foliation which occurs on a macroscopic scale to the west and northwest of the Taalintehdas area (Fig. 2). The S_3 is coarse crenulation cleavage in rocks where the S_2 is of the same type. In other gneisses, differentiated schistosity is defined by biotite and some muscovite, generally with elongated streaks of biotite aggregates parallel to it. In some biotite schists the preferred orientation of biotite is exceptionally well developed (see Williams et al., 1977, Fig. 7d). In areas where F_3 is well developed earlier foliations are transposed parallel to S_3, and the S_3 differentiated layering cannot be distinguished from the parallel S_2 (Fig. 8).

Northerly domain: F_4 structures

F_4 deformation features are not observed in areas with a dominant S_2 (N70E) orientation (Fig. 2). The F_3 deformation caused reorientation of pre-existing structures and the development of a N35E trending axial plane foliation (Fig. 10). F_4 chevron folds are common in areas where the distinctly layered S_3 transposition foliation is well developed. The chevron folds generally have a sinistral vergence and their enveloping surface trends SSW. S_4 differentiated layering is developed locally (Fig. 9), and well-developed crenulation cleavage is common in micaceous intercalations. F_4 folds are observed to be restricted to areas where pre-F_4 foliations have a roughly NS trend. It is suggested that F_4 kinking and folding could take place only in those domains where the foliation was oriented roughly NS (which is in accordance with the observations in the Taalintehdas schists).

Pegmatites

Both the Taalintehdas schists and Masugnsträsk gneisses contain pegmatites parallel to the S_2 transposition foliation. Boudinage and fold structures with axial plane surface parallel to S_2 are common. Penetrative axial plane foliation is not visible on mesoscopic scale in the pegmatites, although there is

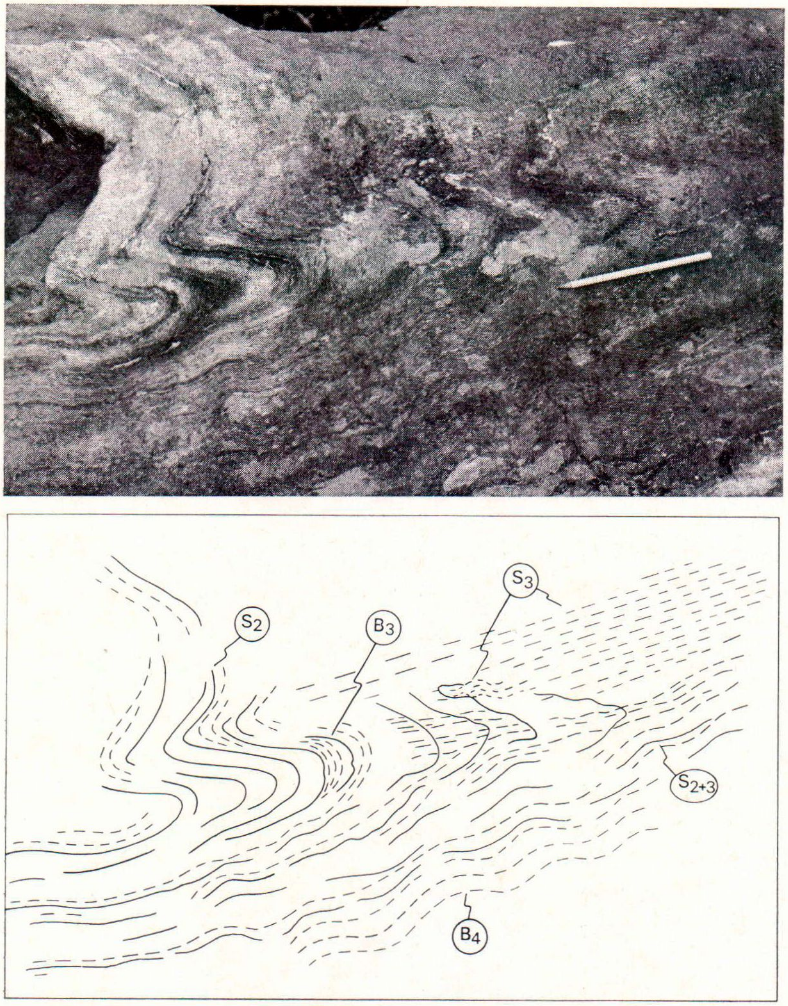


Fig. 8. F_3 fold in quartz-rich gneiss. S_3 differentiated layering has developed in the hinge area (top right). S_2 and S_3 cannot be distinguished in the limbs. Open F_4 crenulations overprint the F_3 fold. Location 66.04—49.00.

some local fracturing parallel to S_2. Other pegmatite bodies cross-cut F_3 fold structures, but no relationship with F_4 structures could be established. Thus a distinction is drawn between pre-F_2 and post-F_3 pegmatites.

Macroscopic interpretation

Figure 2 is a structural compilation of the Taalintehdas area. The macroscopic structure cen be interpreted in terms of the fold generations distinguished on mesoscopic scale. A dominant transposition foliation, S_2, is refolded by a macroscopic fold with a N35E trending axial surface. The large scale folding is matched by the vergence of the F_3 mesoscopic folds. An S_3 transposition foliation is locally developed in the hinge area of the macroscopic fold as outlined by the large quartzite body. F_4 chevron folds are locally developed in restricted areas and domains of various sizes, characterized by a roughly NS orientation of pre-F_4 foliations.

Deformation and metamorphism

The present microstructure of a rock is the end product of many processes operating

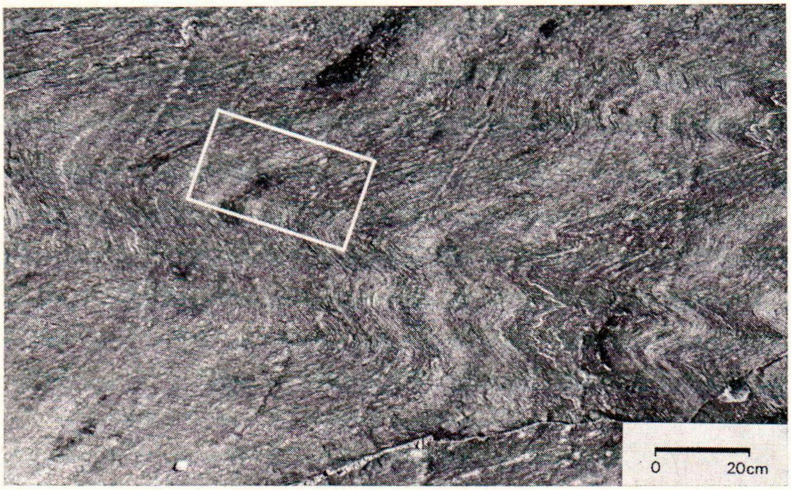


Fig. 9 a. F_4 chevron folds with incipient development of S_4 differentiated layering in quartz-rich gneisses. Outlined area may be compared with Figure 9 b. SW running lineations are glacial striations. Location 65.95—49.09.

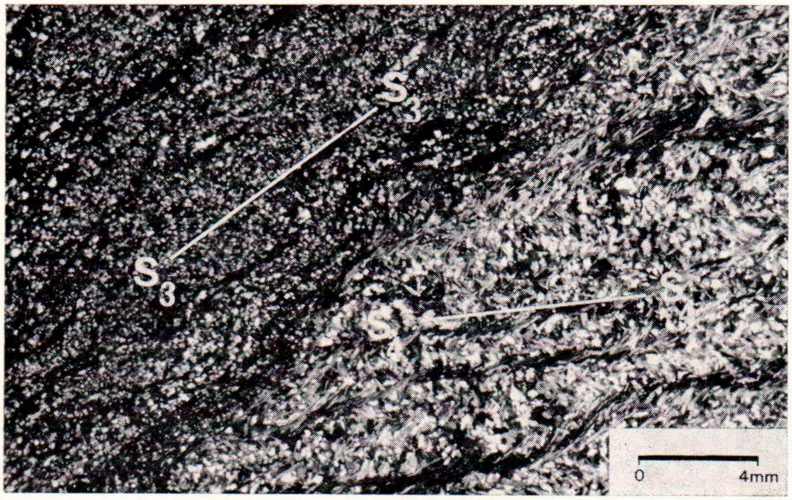


Fig. 9 b. S_4 differentiated crenulation cleavage developed in coarse grained quartz-rich layer, whereas only a slight crenulation of S_3 is visible in the finer grained layer. Crossed nicols. (F517; location 65.80—49.35)

during geological history. To assess the imprint of any of these processes on microstructure development, a careful chronological analysis of the relationship between deformation and metamorphism is necessary (Vernon, 1977 b). An attempt to identify realistic metamorphic reactions is desirable (Vernon, 1976).

The predominant microstructures in the schists and gneisses, described above, are differentiated crenulation cleavages or differentiated layerings with quartz-feldspar and mica domains. It is important to know when recrystallization and growth processes operated in the various minerals and when stable metamorphic microstructures could have formed. In the following the microstructures of the individual minerals are described and an attempt is made to correlate the microstructural development of these minerals with the development of the successive foliations.

Quartz occurring with feldspar forms a fine-grained (ca. 0.1 mm), granoblastic matrix. The individual grains have a straight or slightly undulatory extinction. Grain boundaries are slightly curved and generally meet in triple points. A polygonal mosaic ('foam') microstructure is locally present. The grain size varies with the number and size of inclusions, such as mica, dispersed in the rock. Patches and veins of coarse-grained (up

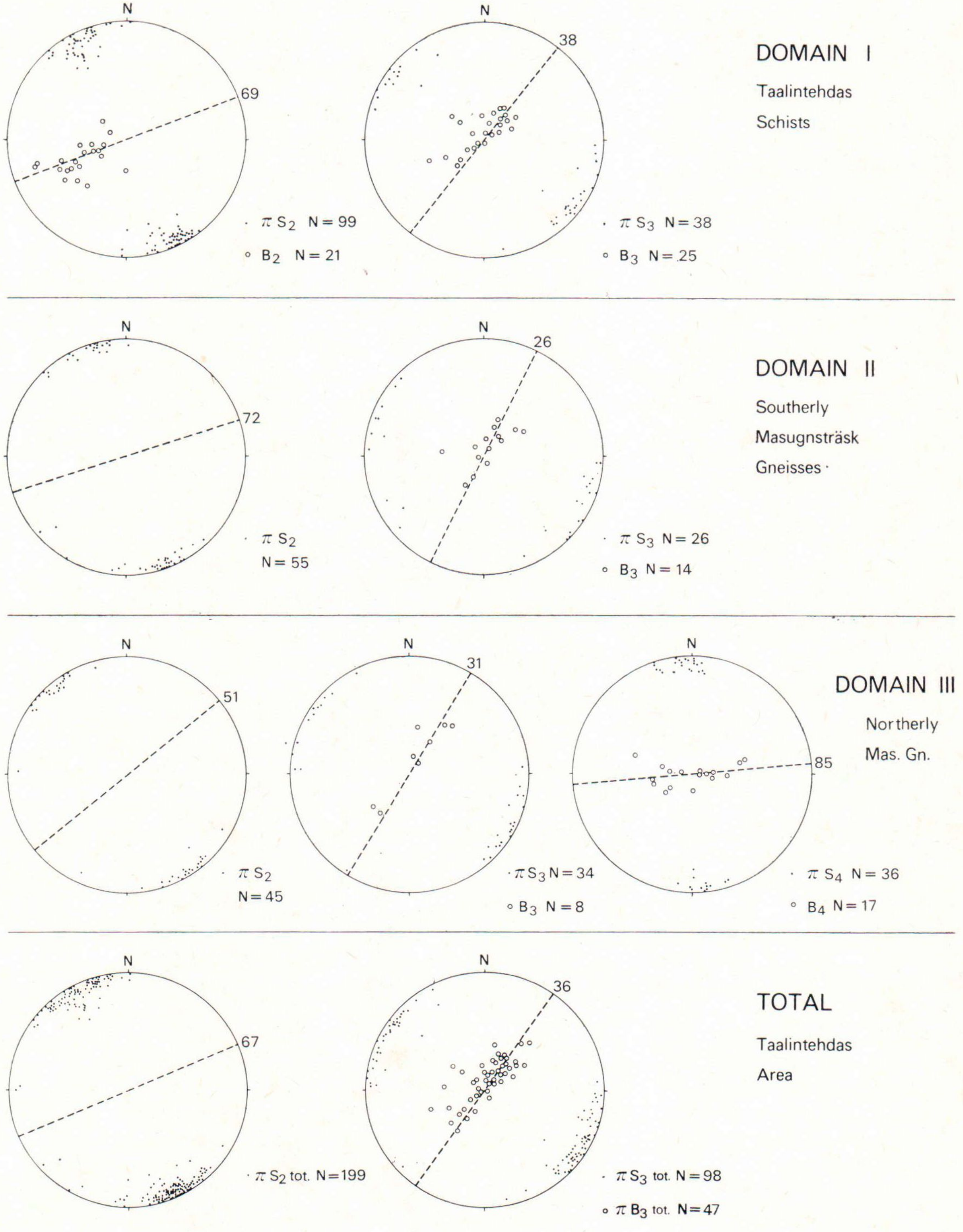


Fig. 10. Stereograms of the Taalintehdas area corresponding to the domains of Figure 2.

to 1.5 mm) quartz occur where inclusions are rare. These observations may be interpreted as growth (grain boundary migration) and grain adjustment features where the migration of grain boundaries was controlled by the presence of inclusions. Exaggerated grain growth (Vernon, 1976) may be involved in the origin of the coarser grained patches. Growth adjustment fabrics are present in all F_1 to F_4 microstructures of both schists and gneisses. Thus, in the Taalintehdas schists the polygonal quartz fabrics have developed at least until after F_3. Intragranular deformation (undulatory extinction, deformation bands) has been been observed in the vicinity of some F_4 kinked micas. This suggests that recrystallization and grain growth of quartz could not operate there effectively during or after F_4. In the Masugnsträsk gneisses, however, where strain-free and grain boundary adjusted polygonal quartz fabrics are present in the folded rocks (Fig. 9 b), it is clear that growth and grain boundary adjustment of quartz took place at least until during or after F_4.

Feldspars, if present, are part of the quartz-rich granoblastic matrix. Microcline in the Masugnsträsk gneisses has a polygonal fabric in the matrix as well as in microcline aggregates. In it suggested that grain boundary adjustment took place until during or after F_4.

Biotite and *muscovite* show similar structures, but with some morphological differences. Muscovites commonly have larger length to width ratios, and the biotites in a muscovite-rich environment are truncated by (001) boundaries of the surrounding muscovites (Fig. 11). Biotite tends to grow in certain areas, such as the strain shadows of large porphyroblasts. In some F_3- or F_4-kinked muscovite aggregates the muscovite has an undulatory character due to slight bending. If the bending is sharper, serrated kink band boundaries are present (Figs. 7 and 11). In general both muscovite and biotite are more or less strain free and have developed rational impingement boundaries, with (001) of muscovite dominating over rational biotite boundaries (Fig. 11). This means that the present microstructure is a result of growth and grain boundary energy reducing processes (Etheridge and Hobbs, 1974; Vernon, 1977a; Williams et al., 1977) which appear to have operated late in history (until during or after F_4).

The microstructures of quartz, feldspar and mica as described above, strongly suggest that a late modification of the microstructure of the quartz-mica matrix is possible. Hence microstructural observations on these minerals cannot be associated unambiguously with the deformation history established (see discussion). It is possible, however, to study the relationships of critical metamorphic minerals to F_1 to F_4 microstructures (foliations, folds).

Cordierite forms elongated porphyroblasts which are generally parallel to S_2, or parallel to S_1 in the more open S_2 crenulation cleavages. Relations between internal (S_i) and external (S_e) foliations are difficult to interpret since many of the blasts have been pseudomorphed by alteration minerals that obliterate earlier intragranular structures and may even have influenced inclusion fabrics. However, the presence of straight inclusion patterns in some blasts suggests post-F_1 (pre-F_2?) growth of cordierite. Some grains have been folded during F_3 deformation (Fig. 12), and many others show evidence of a rather rigid behaviour such as intragranular cracks parallel to S_3, rotation of elongated blasts parallel to S_3, and the growth of biotite in their strain shadows. The porphyroblasts are therefore pre-F_3.

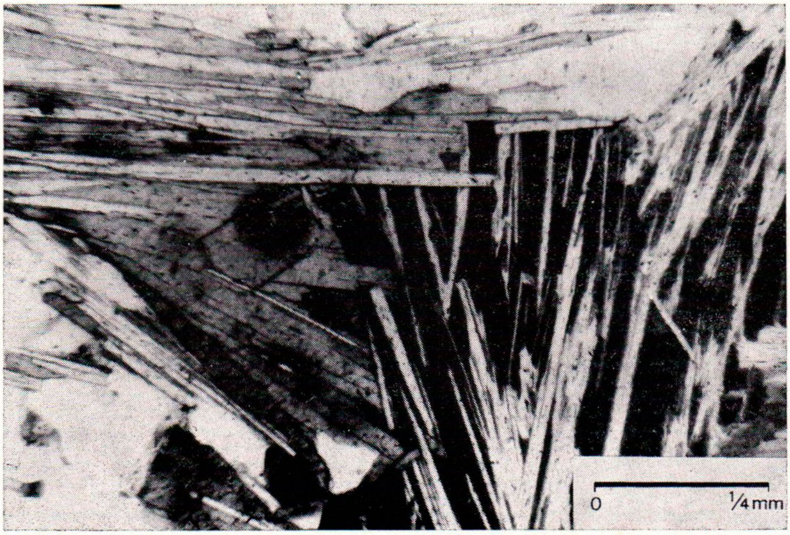


Fig. 11. F_3 kink in mica aggregate from Taalintehdas schist. Individual micas are strain free and have grown out to impingement. Note the dominance of muscovite (001) over biotite (001) boundaries.

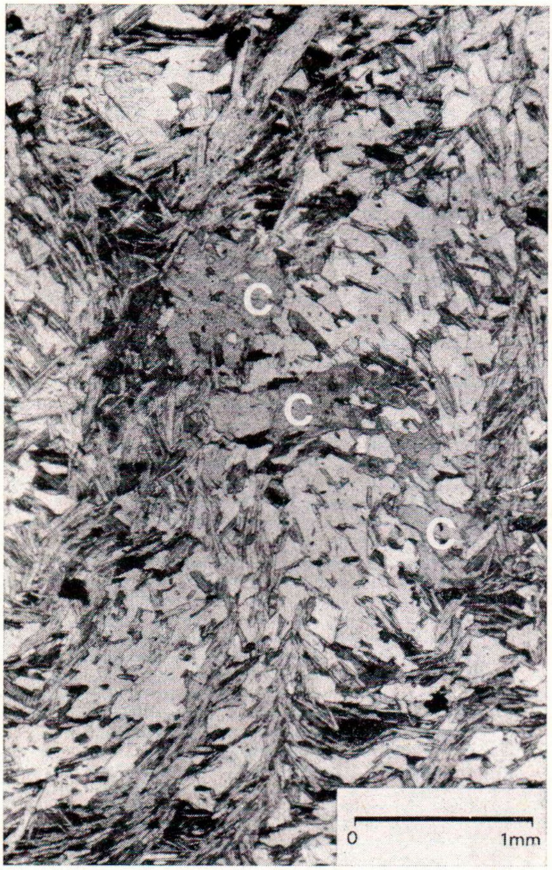


Fig. 12. F_3 crenulated cordierites in quartz-mica schist (Sample F501b; location 65.08—49.79)

A few porphyroblasts show a similar relation with respect to F_2 structures. Hence the orientation of many elongated blasts parallel to S_2 may be the result of transposition during F_2. Nucleation and growth of cordierite can thus be interpreted as pre- or early syn-F_2 (Fig. 14).

Fibrolitic *sillimanite* is present as intergrowths with biotite or quartz, or both, and lack any optical evidence of deformation. Biotite and quartz have a reduced grain size and ragged grain boundaries where sillimanite needles are enclosed. Fibrolite needles pass without deflection of trails from one quartz or biotite grain into another of different orientation. This indicates that fibrolite was present before growth and/or grain boundary adjustment of quartz and biotite took place (Vernon and Flood, 1977). The fibrolite was folded during F_2 deformation (Fig. 13) and is thus earlier than part of the F_2 deformation at least.

Andalusite porphyroblasts are prekinematic to at least part of the F_3 deformation as a rigid behaviour similar to that of cordierite is observed. The curved inclusion fabric, continuous with the external S_2 foliation, that is occasionally present suggests synkinematic growth with respect to F_2. Comparison of fine-grained mica inclusion patterns with coarser-grained external mica fabrics is ambiguous, it being possible to explain the growth as both synkinematic and postkinematic with respect to F_2.

Garnet porphyroblasts have developed in biotite-rich parts of the Masungsträsk gneisses. Inclusion fabrics defined by rutile, ilmenite, and sphene show a well-developed S_2 crenulation cleavage in the core of some blasts. Opaque minerals in the rim of these garnets are aligned parallel to and continuous with the external S_3 schistosity. Garnet porphyroblasts in a biotite-rich layer in the Taalintehdas schists have a straight inclusion (quartz, ilmenite, rutile) fabric in the core, which becomes curved towards the edge. It is continuous with the external S_2 schistosity, which was crenulated during F_3 deformation. Many garnets with an elongated habit have been rotated parallel to S_3 and thus their straight inclusion fabric is discontinuous with the external foliation. These observations suggest that garnet growth started during or after the F_2 deformation and was contemporaneous with at least part of the F_3 deformation.

Clino-amphibole porphyroblasts with straight quartz inclusion patterns are generally aligned with their c-axes parallel to S_2 surfaces. They are deformed by F_3 and show undulatory extinction; subgrains and recrystallization have developed locally. Coarse porphyroblasts are surrounded by small recrystallized amphiboles (cf. Bierman, 1977). It is difficult to set time limits to the amphibole growth; nevertheless, it appears to have started before the second deformation phase, whereas recrystallization occurred during the third phase.

Some events younger than the fold generations described above have been recorded in both the Taalintehdas schists and the Masungsträsk gneisses. Their relative age is based cross-cutting relationships observed in thin section. First, small-scale fracturing and mylonitization have created thin zones (ca. 0.2 mm) with an aphanitic matrix, surrounded by strongly kinked, fish-shaped micas and deformed quartz grains that show recrystallization along deformation bands. Quartz grains without any apparent deformation feature, however, occur adjacent to the mylonite suggesting either strongly heterogeneous deformation or recrystallization during or after mylonitization. Strongly kinked mica without any evidence of recovery occurs throughout the area and is considered to be related to this late fracturing.

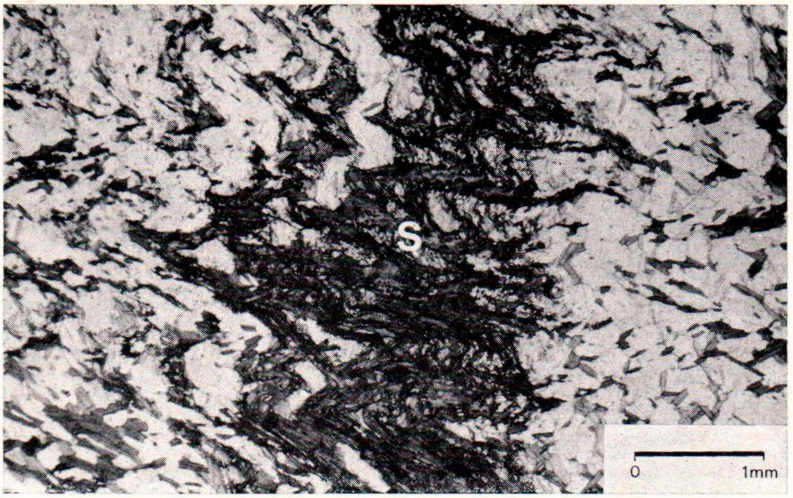


Fig. 13. F_2 folded fibrolitic sillimanite intergrown with biotite. (Sample F551; location 65.00—49.78)

Thin microcline veinlets (0.1—1 mm wide) have formed along irregular cracks crosscutting all the previously described structures. They are generally normal to S_2 surfaces.

Finally, some pyrite veinlets (0.1—0.3 mm wide), partially or completely altered into iron-hydroxides, have formed along matured polygonal matrix grain boundaries and foliations.

Discussion

In interpreting the microstructures of the schists and gneisses, it is important to note the following: (1) Differentiated layerings are dominant fabric elements in both schists and gneisses. Transition from crenulations via crenulation cleavages to differentiated layerings can generally be traced. (2) The matrix minerals (quartz, feldspar, mica) are optically clear and bear hardly any evidence of intracrystalline deformation. Local kinking of micas, undulatory extinction, deformation bands and subgrain development in quartz can be related to late stage (post folding) shearing and fracturing of the rocks. Grain boundary migration and grain boundary adjustment of quartz and micas has been demonstrated to continue until after the fourth fold generation.

A detailed analysis of the development of a layered crenulation cleavage by Marlow and Etheridge (1977) similar to the analyses described in the present paper showed that chemical, mineralogical, grain size, and preferred orientation changes are involved. These authors suggest that processes of the 'diffusive mass-transfer' type play a major role in the development of differentiated layering. A major argument in favour of this theory is the absence of phenomena indicative of intracrystalline deformation. If intracrystalline dislocation flow mechanisms (cf. Nicholas and Poirier, 1976) should have been dominant, extensive annealing recrystallization would have been required to remove all evidence of intracrystalline deformation.

The development of a polygonal foam microstructure in microcline in the schists and gneisses around Taalintehdas can be explained if the development of differentiated layering involved dominant diffuse mass-transfer processes. In rocks where quartz is deforming by intracrystalline plasticity, microcline is generally known to exhibit brittle behaviour (e.g. Bossière and Vauchez, 1978;

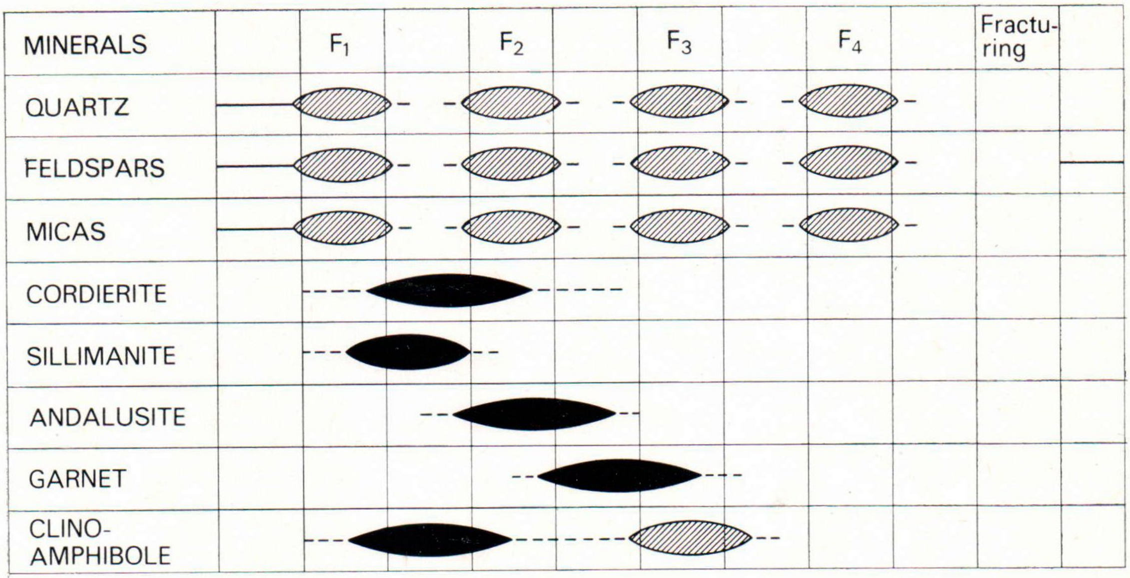


Fig. 14. Schematic diagram illustrating the relationship of deformation and metamorphism as inferred from microstructures. Quartz, feldspar and mica have recrystallized during each deformation phase. Microcline veinlets have developed after the fracturing period. The highest probability of nucleation and growth is estimated at places where fish are thickest. Clinoamphibole recrystallized during F_3.

Lister and Price, 1978). Since both quartz and microcline develop foam microstructures, a deformation regime not dominated by intracrystalline dislocation flow in quartz alone is indicated. Diffusive mass-transfer processes seem to be required for microcline to develop a foam microstructure.

The development of differentiated layering during each of the four folding events places constraints on the interpretation of fabrics of structures older than F_4. Growth and grain boundary adjustment during F_4 deformation may have affected some and perhaps all of the features characteristic of deformation during earlier phases (F_1, F_2, F_3).

In summary, if we assume that intracrystalline plasticity was important during a large part of the deformation history, annealing recrystallization must have played a key role too, at least during and/or after the last deformation phase, because of the presence of growth and grain boundary adjusted fabrics in F_4 deformed rocks. It seems inevitable that the pre-existing F_1, F_2, and F_3 structures should have been affected. If, on the other hand, we assume that processes of diffusive mass-transfer were dominant, then only growth and grain boundary adjusted fabrics would have developed (cf. Marlow and Etheridge, 1977) and the influence on the fabric of older structures is more difficult to assess from microstructural observation.

The present paper shows that critical metamorphic minerals grew at different stages of the metamorphic and structural history (Fig. 14). Since differentiation processes took place during each deformation phase, quartz, feldspars and micas are inferred to have recrystallized during successive deformation events. In multiple deformed and differentiated metapelites matrix minerals have a good chance of reequilibration during recrystallization. This holds not only for the rocks described above but also for metapelites from other parts of Kemiö Island (and any other comparable area) where only one (transposition) foliation is dominant. Studies on element partitioning between coexisting minerals should provide information on the possible loss of chemical equilibrium in recrystallized matrix minerals, e.g. in biotite-garnet and biotite-cordierite relationships.

Regional interpretation

The structural history established in the Taalintehdas area offers a basis upon which to interpret of the surrounding areas. To test the applicability of the Taalintehdas structure to adjacent areas, a number of localities were selected on Kemiö Island. The location of these areas is marked in Figure 1. The data obtained at these localities are based on field observations only; the analysis of microscopic structures has yet to be done.

In many of the localities where structural analysis has been carried out a similar succession of four fold generations has been established. As S_2 transposition foliation is the dominant foliation in the Taalintehdas area, it is inferred that the dominant foliation, which has a similar orientation in adjacent parts of Kemiö Island, is also related to F_2. This is affirmed at most of the localities investigated.

As described in the section on geological setting, north-south trending foliations prevail west of Taalintehdas. Structural analysis in this area shows that local F_3 folding reoriented older foliations into a north-south direction. Additional F_4 deformation is locally present. The dome structure around Bergö is formed by the overprinting of a fourth generation structure over older ones (P. F. Williams, C. R. van Staal, pers. comm.). A sequence of metamorphic gabbros, pillow basalts and agglomerates near Vestlax shows four generations of folds, which have been correlated with F_1 to F_4 structures at Taalintehdas (J. H. M. Coenen, pers. comm.). Gabbroic amphibolites on Kemiö Island occur in at least two distinctly different geological settings. The majority of the amphibolites are intercalated in the supracrustal gneiss series as massive bodies transected by local shear zones. A metagabbro body and the surrounding supracrustal rocks just NW of the Taalintehdas area were investigated in detail. The supracrustal gneisses show a sequence of F_1, F_2 and F_3 folds which have been correlated to folds defined by the contact of both rock units and by shear zones inside the gabbro body (Verhoef, 1977). The body north of Pederså shows overprinting of large scale F_1 and F_2 folds (Verhoef, op.cit.). These observations suggest that these magmas were emplaced before the first deformation phase.

Microstructures in rocks from the 'synkinematic' rock suite are more difficult to date. The gabbroic rocks show a penetrative amphibole lineation and evidence of local S_2 shear zones parallel to it (Dietvorst, 1977). The tonalitic and granodioritic rocks have a penetrative E-W foliation that continue into the surrounding supracrustal rocks (Simonen, 1960) and into the shear zones in the metagabbros (Dietvorst, op.cit.). $F_1 — F_2 — F_3$ overprinting may be present at a locality in migmatitic tonalite north of the gabbros in the northwest of Kemiö Island. It is generally very difficult to establish the existence of F_1 deformation structures in the synkinematic intrusives.

Our preliminary findings can be summarized as follows. The supracrustal rock series, including the intercalated metagabbros, have undergone at least four folding events, with F_2 structures dominant at most places. The 'synkinematic' tonalite suite is of a pre-F_2 age. The F_2 deformation is characterized by extensive transposition of earlier foliations and rock bodies parallel to the S_2 axial plane schistosity. This explains the simple EW pattern of foliations on a macroscopic scale (Fig. 1). F_3 deformation is only locally developed and in the southern part of Kemiö Island F_4 structures have a dominating influence on the macrostructural pattern visible on the map. If F_3 is responsible for pre-F_4 variation in S_2 (as in the Taalintehdas area), then F_3 by orientating S_2 favourably localizes F_4.

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