6. On the Tectonics of the Caledonides in the South-Western Part of the County of Jämtland, Sweden

By

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ABSTRACT.—Four different nappes have been successively overthrust eastward upon each other. These are: (1) the parautochthonous; (2) the mylonite nappe; (3) the Särv nappe; and (4) the Helags nappe. In the Särv nappe thousands of steep dolerite dikes became inclined and partly deformed during the thrust movements. Only faint traces can be detected of a deformation older than the dolerite dikes. Parts of the primary pre-Cambrian bedrock are included in the Särv nappe. The augengneiss around Tännäs has been formed metamorphically by a late porphyroblastesis in a mylonitized mixture of granites and basic rocks. A wide-spread postdeformative recrystallization has more or less affected the rocks.

Contents

INTRODUCTION

The area surveyed is situated in the south-western part of the county of Jämtland, and extends for about 100 km in a N–S direction along the boundary between Sweden and Norway from latitude 62°18′N to 63°15′N. In an E–W direction the extension of the area is about 80 km.

Within the major part of the area the bedrock consists of Caledonian allochthonous rocks that have undergone metamorphism of varying intensity. These rocks, which have been overthrust as a unit over a long distance, were grouped together by A. E. Törnebohm (1896) as "Stora Seve-täcket", i.e., the Great Seve nappe. It appears practically along the entire length of the Scandinavian mountain chain. In Sweden its southernmost occurrence of appreciable size lies in the south-western part of the county of Jämtland.

The conception of the mountain structure in the southern part of the Caledonian mountain range, to be discussed in the following pages, is based mainly upon my own observations in the field and the mapping results thereby obtained.

The primary observations in the field have been supplemented by petrological and chemical investigations in the laboratory. The elaboration of the material has been limited in so far that, in general, only investigations essential for the tectonic analysis of the mountain structure have been carried out.

One of the aims of the present paper is an attempt towards the distinction and analysis of the different tectonic units that compose the bedrock of the area, and a comparison of the petrographic and tectonic characteristics of these units. Special emphasis has been placed on the study of a great tectonic unit, which I have named the Särv nappe (STRÖMBERG, 1955). The treatment of this nappe and of the zones adjoining it can thus be said to be the main object of the investigation. On account of the great extension of the Särv nappe the study of its occurrence has necessitated the consideration of adjoining tectonic units. This has resulted in the regional character of the work as expressed in the title of the paper. The description of other tectonic units, e.g. the more highly metamorphic rocks in the Helags nappe, has, in accordance with the emphasis placed upon the Särv nappe, been limited to a more general level.

Within extensive regions in western Härjedalen, where no outcrops have been found, tectonic observations have been difficult and the mapping suffers from some degree of uncertainty. Elsewhere, when possible, observations of strike and dip and of linear elements in the rocks have been made. Where larger exposures of schistose rocks were encountered, the greatest possible number of measurements of *s*-planes and lineations have been taken. These measurements have then been compiled in synoptic diagrams, and in a number of cases the type of rock deformation has been determined by an examination of the fabric.

In a previous paper (STRÖMBERG, 1955), which is referred to in the following as (S.55), I have described the parautochthonous rocks in the parish of Hede and the slightly metamorphosed parts of the Seve rocks within the reindeer pastures of Tännäs. In order to facilitate the interpretation of local names, those of lakes are preceded by L., those of rivers and brooks by R., and those of mountains and other heights by Mt. The geological map is provided with a grid reference.

As a topographical basis I have utilized mainly the draft map sheets worked out by the Swedish Geographical Survey in the course of the 1930's. The area in question is covered by the following sheets: 65 Duved, 66 Åre, and 71 Ljusnedal to the scale of 1:100,000 and 72 Åsarne, 76 Tännäs, and 77 Hede to the scale of 1:50,000. The three last-mentioned maps are each divided into six part sheets (NW, NM, NE and SW, SM, SE).

The mapping of Mt. Röstberget at Funäsdalen village has been based upon an enlargement of an aerial photograph.

The results of my mapping in the area surveyed up to the autumn of 1956 have been incorporated in *the pre-Quaternary Rocks of Sweden*, published under the editorship of N. H. MAGNUSSON (1958).

Earlier investigations

The earliest topographical and mineralogical data about western Härjedalen are found in a diary kept by URBAN HJÄRNE during his journey to Härjedalen and Jämtland in 1685 (HJÄRNE, 1763). At this time HJÄRNE was assessor to the Board of Mining, and undertook his journey for the purpose of examining the newly discovered copper ore in Messlingegruvan (the mine of Messlinge) southwest of Mt. Mittåkläppen, and the occurrences of silver and lead ores at Olden.

Of particular interest are the investigations carried out by DANIEL TILAS, a grandson of HJÄRNE, in 1735 and later in the area on either side of the national boundary (Riksgränsen). TILAS (1742) mentions the observation of a black hornblende rock from the easternmost Mt. Svansjökläppen (F1) and (1743) of a polymict conglomerate from some mountains east of L. Fämunden (ZENZÉN, 1930).

Comparatively detailed geological observations were made in western Härjedalen in 1796 by the notary to the Board of Mining, CARL MAGNUS ROBSAHM, who together with ANTON SWAB, supervisor of mining in Falun, travelled to Röros via Härjedalen, and on this occasion made detailed diary notes (H. RICHTER, 1938).

The mountainous regions near Riksgränsen were examined in the middle of the nineteenth century by a.o. J. C. HØRBYE. In 1855 he described the topography and bedrock along the frontier from northern Dalarna to southern Jämtland, and reported observations on schistosity in the porphyry of the Riksgränsen anticline and on the formation of augengneiss at Mt. Skarsfjällen.

Towards the end of the nineteenth century knowledge of the rocks in the southern mountain regions increased rapidly as the result of geological investigations carried out by C. E. SCHIØTZ (1873, 1882, 1883), TH. KJERULF (1873, 1879), A. E. TÖRNEBOHM (1873, 1882), FR. SVENONIUS (1881), and A. G. HÖGBOM (1889). The often very heated controversies in the later decades of the century concerned especially age relations and the tectonic relations between the Cambro-Silurian, the sparagmite rocks, and the Vemdal quartzite.

In 1896 A. E. TÖRNEBOHM divided some of the most important rocks of the mountain range into two large groups, viz. Silurian formations and the Seve group. He split the latter group further into: (1) clastic facies, i.e. sparagmite rocks, and (2) crystalline facies, or Åre schists. At the same time he presented good reasons for this classification by showing that the crystalline facies of the Seve group belongs to a great nappe, which is separated by a long overthrust from the crystalline rocks of the foreland and its sedimentary deposits. These deposits, viz. the clastic facies of the Seve group and autochthonous and para-autochthonous Cambro-Silurian sediments, have retained their primary clastic features.

The combining of both the sparagmite rocks and the more or less crystalline schists of the large nappe into one large Seve group is probably due to the fact

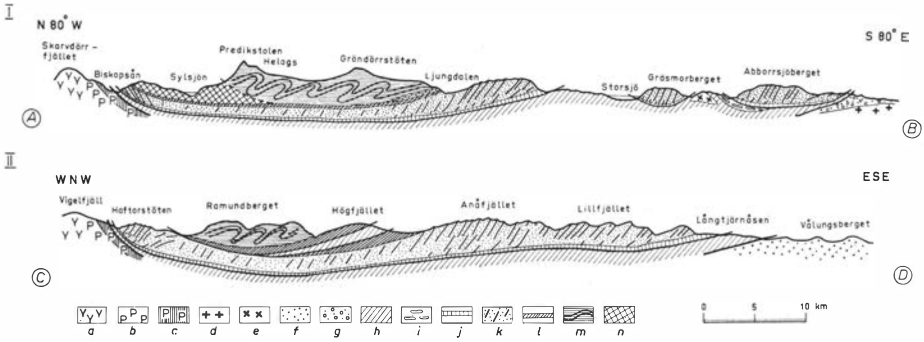

Fig. 1. Schematic, transversal sections through the Caledonides in south-western Jämtland. Section I runs S 80° E (C1–D7), section II ESE (D1–G5), cf. Pl. I. Vertical scale = 4 × horizontal scale.

a = light-coloured microcline-granite, *b* = quartz-porphyry, *c* = schistose quartz-porphyry, *d* = red Rätan granite, *e* = red gneiss granite, *f* = red sparagmite, *g* = blue-black quartzite, *h* = mylonites (mainly granitic), *i* = blastomylonite, *j* = schistose basal zone of the Särv nappe with the Ulvberg complex, *k* = the Särv nappe with dikes of Ottfjäll dolerite (partly sheared), *l* = mylonites and schists in the basal zone of the Helags nappe, *m* = the Helags nappe, *n* = quartzitic schists and amphibolite.

that the mainly psammitic sedimentary rocks, which predominate in these divisions, exhibit several stratigraphic similarities.

At the beginning of the twentieth century the structure of the Caledonian mountain range was the subject of lively discussion among Scandinavian geologists, whose different opinions have been summed up by B. Asklund (1946). During the period mentioned several papers dealing with the geology of the southern part of the county of Jämtland were published. Among these may be mentioned especially the papers by P. J. Holmquist (1919), A. G. Högbom (1920) and G. Frödin (1920, 1921, 1922 a, b).

Among subsequent investigators dealing with adjoining parts of the mountain chain we notice especially C. W. Carstens (1920, 1924, 1928), B. Asklund & P. Thorslund (1935), P. Thorslund (1937), B. Asklund (1938), O. Kulling (1942), P. Holmsen & Chr. Oftedahl (1956), G. Kautsky (1953), G. Stålhös (1956), and K. Ø. Bryn (1959).

The principal tectonic units

The classification of the allochthonous or parautochthonous Caledonian rocks into several different tectonic units, which in most cases have different extent or a different mode of overthrusting to their present locations, can be based upon the presence of one or more of the following criteria concerning the mutual relations between the different tectonic units:

(1) Marked contrast with regard to type and degree of metamorphism of the rocks constituting adjoining tectonic units.

(2) An obvious zone of movement, indicated by the formation of strongly cataclastically deformed rocks with or without crystalloblastesis, separating neighbouring tectonic units.

(3) Repetition due to overthrust, inversion, or other decided disturbances of the stratigraphic sequence at the boundary between the units.

Whenever I have distinguished between different tectonic units in the area surveyed, at least two of the above criteria have been fulfilled.

In this way it has been possible to distinguish four Caledonic tectonic units within the region investigated. In the order of rising tectonic level these are: (1) parautochthonous sedimentary rocks and granitic gneiss; (2) the mylonite nappe; (3) the Särv nappe; and (4) the Helags nappe. Between the Särv nappe and the Helags nappe there is a zone with mylonites and ultramylonites (Fig. 1).

The degree of metamorphism of the rocks increases in general abruptly from unit to unit in the order of their enumeration. The sparagmite south and east of Tännäs, the granite gneiss north of Hede, and the blue-black quartzite between R. Ljungan and Bydalen village (B7) are parautochthonous to allochthonous. The mylonite nappe with augengneiss rests upon the parautochthonous blocks, and projects below the Särv nappe.

This latter nappe rests flatly in the trough-shaped depressions of the mylonite nappe (cf. sections in Fig. 1). In the sedimentary rocks of the Särv nappe the degree of metamorphism is in most cases higher than that in the parautochthonous rock floor, but this nappe has been found to contain also sedimentary rocks with a very slight metamorphism. On Flatruet the southern Särv nappe westwards grades into an area of hard quartz-muscovite-schists.

From the tectonic point of view, granite and porphyry in the anticline at Riksgränsen (B–E1) occupy a special position.

In contrast to the southern part of the Särv nappe the rocks in its northern part are characterized by the formation of biotite coupled with a general recrystallization. A common feature of the entire Särv nappe is the uniform appearance of the Ottfjäll dolerite as intrusions in the form of swarms of dikes usually with a steep westerly dip. At the eastern boundary of the Särv nappe towards the Helags nappe mylonitic rocks form a transition zone several hundred metres wide. This zone is particularly wide at Flatruet (F2–D3) and contains granite mylonite and other mylonites.

The Helags nappe, the highest tectonic unit of the region, contains strongly metamorphic, crystalline calcareous schists and gneissic rocks designated by Törnebohm and others as Åre gneisses. They contain several different intrusives, e.g. several types of amphibolite and isolated dikes of trondhjemitic rocks.

REGIONAL DESCRIPTION AND PETROGRAPHY

The parautochthonous regions

The sparagmite area

The area with sparagmite rocks around Hede has been described and examined to some extent by the present author (A. STRÖMBERG 1952, 1955) and in greater detail by G. STÅLHÖS (1957). More attention will therefore be given here to the less known region with sparagmite-like, light-coloured quartzitic sandstones south of Tännäs.

Immediately north of L. Bolagen, in the valley which descends towards the lake and is bordered by the eastern Mt. Rutenfjällen and Storvigeln (F1), I have found a fine-grained, light-coloured, quartzitic sandstone resting upon quartz-porphyry and apparently not much disturbed tectonically. Often the content of feldspar is low in this sandstone, which is clearly clastic, with scattered grains of white feldspar. Farthest in the north-west the quartzitic sandstone is squeezed in between the underlying quartz-porphyry of Vigeln and the steeply dipping augengneiss which limits the sandstone towards the NE. The proximity of the augengneiss has often strongly influenced the sandstone, producing in it a schistosity and deformation of varying degree. Thus, at the top of Mt. Vättafjället (F1) light-coloured, quartzitic schists with muscovite are exposed. The quartzitic schist on Mt. Tjärnbergsåsen (G3) has been made the object of a special study, and is represented in Pl. III *24*.

At Käringsjövallen and Myskelåsen (H3) I have again found the same fine-grained, light-coloured, quartzitic sandstone. There, immediately below the augengneiss, we find a phyllitic dolomitic limestone at the contacts with the sandstone. Here also, the zone of contact is characterized by a strong schistosity indicating considerable translative movements. Obviously the augengneiss has been thrust over the quartzitic sandstones, although the dips have been rendered rather steep through the action of a folding along axes with a mainly WNW strike (cf. longitudinal section G–H). This folding is to be seen especially distinctly at the boundary between the Särv nappe and the augengneiss west of Tännäs. A closer description will be given in the following.

The granite gneiss

East of Lunnäset village (F6) there is a large area of granite gneiss exhibiting certain characteristics of deformation. Most wide-spread within this area are rocks of granitic composition which are obviously closely related to the Rätan granite, and differ from it only by exhibiting traces of a metamorphism varying in type and intensity from slight crushing to pronounced saussuritization and recrystallization (Fig. 2).

Fig. 2. Cataclasis in granite gneiss from Mt. Fröåsen (F6). The dark mineral in the central part is a broken garnet, the pieces of which have been rotated and separated by translative movements. In contrast to the primary crystal faces the fracture surfaces are free from alteration products, which seems to indicate a comparatively late deformation. The thin-section was orientated in the tectonic *ac*-plane. *Q* = quartz, *Pl* = plagioclase, *Mi* = microcline. × 35.

The Rätan granite which, to judge from its appearance, seems to be entirely unaffected by deformation, is of wide-spread occurrence only a few kilometres north-east of the easternmost outcrops of granite gneiss. The boundary for the participation of the crystalline Archaean rocks in the Caledonian deformations seems to run between the granite gneiss and the Rätan granite east of it. In this case the metamorphism which has impressed itself upon the rocks in the granite gneiss area should be mainly of Caledonian age. This assumption is supported by mineralogical and structural investigations.

In addition to the predominant granite gneiss, the area exhibits basic intrusives, among which some occurrences of ophitic dolerite near Vemvallen seter (F7) have been mentioned in my previous paper (S.55, p. 218). The structure of the dolerite at Vemvallen reminds of the Åsby dolerite, as well as of several occurrences of dolerite in the augengneiss.

As I have earlier emphasized (S.55), the granite gneiss at Mt. Orrhögberget (F7) is thrust upon parautochthonous quartzitic rocks, which were interpreted as Vemdal quartzite.

My observations show that between Mt. Orrhögberget and Mt. Särvfjället the granite gneiss is everywhere involved in and mobilized by the Caledonian translative movements.

Metamorphism seems to be particularly intense at the boundaries of the border region in the south-east and north-west. At the north-western boundary the gneiss constitutes the base of the Särv nappe. In spite of the fact that there the bedrock is hidden by vast boggy regions, experience from the augengneiss zone in the west makes it probable that also here a formation of augengneiss has taken place, and that this metamorphism has at least the same extension as the above-mentioned metamorphism at the south-eastern boundary of the gneiss region towards the Vemdal quartzite.

The boundary for the participation of the Archaean in the Caledonian movements thus seems to run roughly at Mt. Orrhögberget, and then to continue in SSW direction towards Hede and Rånddalen. This agrees with WEGMANN's assumption (1959, Fig. 3), according to which the boundary of the movements of the basement runs from L. Atnesjøen in Norway towards Hede. ASKLUND (1960) considered the thrust at the boundary between the Vemdal quartzite and the granite gneiss to be of late Caledonian age. However, according to my investigations, the imbricate thrusting proves a more intense intermingling and thrusting together of these different tectonic units than ASKLUND has assumed.

PETROLOGY OF THE RED GRANITE GNEISS FROM Mt. FRÖÅSEN (F6).—*Quartz* occurs in three ways: (1) shattered in zones of fracture; (2) as schlieren of strongly undulous "Langquarz" that traverses the rock in undulating ribbons; (3) recrystallized, with weak undulosity, where the rock appears generally recrystallized.

Microcline appears mainly in individual grains 3–5 mm in size. They exhibit a faint microcline lamellation, but often contain inclusions of older plagioclase and microcline, the different fragments of the included mineral exhibiting a mutually homogeneous orientation. The microcline seems to be recrystallized throughout, often in several generations.

Plagioclase is either an oligoclase with the approximate composition $Ab_{85}An_{15}$ or a younger albite with composition $\sim Ab_{95}An_5$. The older plagioclase exhibits varying degrees of progressive saussuritization, starting from an incomplete transformation of the feldspatic mass into clinozoisite, chlorite, etc., with formation of albite in the periphery of the plagioclase and ending with a stage in which the oligoclase has altogether disappeared so that its place is denoted only by well-crystallized accumulations of epidote (clinozoisite) and chlorite. *Microcline pertite* occurs as a newly formed mineral together with albitic plagioclase. *Epidote* occurs in a form that is usually poor in iron, clinozoisite. *Chlorite*, pale, faintly pleochroitic, often appears in aggregates together with epidote. *Ilmenite* usually forms inclusions in aggregates of *titanite*, and in connection with it epidote is of common occurrence. *Rutile* in the shape of isolated, small, short and prismatic crystals is rare.

Garnet. A yellowish brown crystal about 0.1 mm in size has been observed (Fig. 2). It contains inclusions of *apatite*. The garnet shows weak birefringence and is surrounded by epidote, relatively poor in iron, and chlorite. It thus ought to be of the ugrandite type, probably grossularite-andradite (WINCHELL, 1946). The crystal is fractured, parts of it having been carried along by the differential movement. Thust the garnet has been deformed after crystallization, and ought to have been formed already during the primary crystallization of the granite.

TEXTURE.—Three metamorphic stages can be distinguished: (1) cataclastic deformation with preponderant shattering of the granitic minerals; (2) formation of new minerals; (3) postdeformative general recrystallization.

The metamorphic processes which include chemical reactions between the different crystalline phases took place especially during the intermediate stage (2). Among these processes saussuritization of older oligoclase with the formation of albite, epidote, muscovite, and chlorite, seems to have been the most important. The process has frequently proceeded to its final stage. In this connection, the minerals which had been newly formed as the result of the saussuritization have crystallized so as to appear as independent components of the rock, and the common origin of these minerals from older plagioclase is indicated only by the accumulation of epidote, muscovite, and chlorite in elongated aggregates. The formation of titanite and epidote from ilmenite as well as that of epidote and chlorite from garnet seems to have taken place, at least in part, parallel with the saussuritization.

Microcline seems to have been involved in repeated recrystallizations with the result that recrystallized microcline often contains inclusions of older microcline and plagioclase. The quartz has also been subjected to repeated recrystallization, but often appears in long ribbons of tail-over-tail arranged grains of strongly undulous "Langquarz".

The blue-black quartzite and the greywackes

East of L. Storsjö the mylonite nappe rests directly upon blue-black quartzite and greywacke-like schists. The latter are confined to a zone which lies along and to the east of the mylonites from R. Ljungan in the south to L. Buhösten (C6) in the north. In this area the greywacke-schists are in general steeply inclined and often strongly folded with steep axes. This zone continues on the northern side of the valley of R. Arån (C7). It is also strongly folded there (Fig. 4). Up to the present, I have not found any fossils in these greywackes. According to MAGNUSSON *et al.* (1958) the greywackes and slates of Arådalen valley are of Ordovician age.

The basal formations of the blue-black quartzite have been observed in a locality upon Svedjelandet (D6). The granitic bedrock is folded up and dragged along with the blue-black quartzite (Fig. 3). The granite, which on account of cataclastic deformation is in itself strongly crushed, is covered by a residual breccia, about 1 m thick, and which among granite material contains occasional pebbles of quartzite. Above this breccia follows coarse blue-black quartzitic

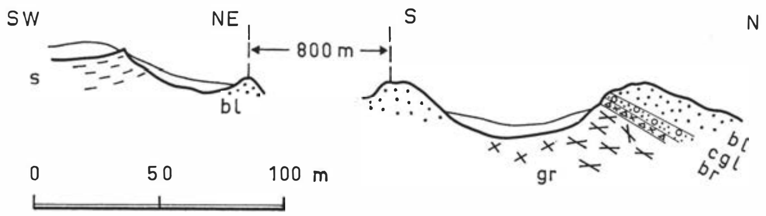

Fig. 3. Base and top of the blue-black quartzite. Sections through Svedjelandet (D6), 2 km SE of L. Nedre Grucken. Vertical scale = 2 × horizontal scale. *s* = quartzitic schist, *bl* = blue-black quartzite, *cgl* = basal conglomerate of the blue quartzite, *br* = residual breccia, *gr* = cataclastic granite.

sandstone, with grains of quartz up to 5 mm in size. This gritty sandstone is about 2 m thick. Higher up, the sandstone passes successively into normal and fine-grained, blue-black quartzite.

The mylonite nappe

General remarks

In most places, where exposures permit observation, the Särv nappe has proved to rest upon mylonitic rocks of varying type. These mylonites are in general several hundred metres thick, and usually lack a direct connection with underlying formations, irrespective of whether these consist of eruptive or sedimentary rocks. Furthermore, as the contacts between the mylonites and their basement always exhibit traces of strong differential movement, there are reasons for distinguishing the mylonitic sole underneath the Särv nappe as a special tectonic unit.

It does not seem possible to decide at present whether or not this mylonite nappe has been transported as a whole over a long distance. The second alternative is supported by observations of local similarities between the material in the mylonite nappe and the immediately underlying rock. Thus the augengneiss west of L. Malmagen (E–F1) rests upon small gneissic nappes of Vigel granite. Similarly, the thin mylonite zone between the Särv nappe and the blue-black quartzite at Svedjelandet (D6) south of R. Ljungan chiefly consists of quartzitic schists.

Furthermore, the schistose and mylonitic basal zone of the Särv nappe often contains a large quantity of basic material. For the origin of the latter we have probably to look not only among the basic rocks in the pretectonic Archaean; to a large extent they ought to be mylonitized dikes of dolerite deriving from the Särv nappe; parts of them could also be intrusive basic rocks injected more or less syntectonically into the thrust zone itself.

The conception of local additions to the mylonite nappe of portions of the basement that have been dragged along in connection with the overthrust,

Fig. 4. Contortion folding in a banded greywacke. *Regel der Stauchfaltengrösse* (thicker layers produce larger folds, SANDER, 1949). In this and the other photos, the head of the hammer is 10 cm and the shaft 40 cm. Glacial boulder at the road near the Lapp village Glän (C7).

however, does not preclude the designation of the mylonites as a tectonic unit, characterized especially by strong translative movements, thrust contacts towards the floor as well as the roof, and a relatively complex structure due to the occurrence of materials with varying texture and chemical composition.

Presumably the nappe is composed in the main of primary, but not isolated, material which has undergone a relatively long transport, and to which have been added secondary elements from the overthrust Särv nappe as well as from the underlying rocks. These secondary elements consist of dragged-along, infolded, brecciated, and mylonitized portions of these different rocks. The numerous observations of quartzitic schists and of dolomite-marble folded into the boundaries of the mylonite nappe point in this direction. The irregular occurrence of the basal formations of the Särv nappe may be partly attributable to such a wear and tear in the lowermost portions of this nappe.

The rocks of the mylonite nappe are influenced by mainly dynamic and regional metamorphism. On this basis most of these rocks can be classified as cataclasites, mylonites, blastomylonites, phyllonites, and blastophyllonites.

Augengneiss along the Riksgränsen anticline

The mylonite nappe, which here consists mainly of augengneiss, protrudes beyond the southern part of the Särve nappe. The augengneiss is several hundred metres thick at Tännäs, but gradually becomes thinner towards the north along the western and eastern boundaries of the nappe.

The northernmost spot at which the augengneiss has been observed lies in the west at R. Biskopsån near L. Sylsjön (C1). At this point the rock has a thickness of about 25 m, and passes in a downward direction into a dark mylonite, this change being due to the gradual decrease of the crystalloblastic "eyes" (Fig. 5). The increasing frequency and growing dimensions of the porphyroblastic feldspars in an upward direction towards the base of the Särv nappe make the connection between the deformation and the porphyroblastic development of the potassium feldspars particularly obvious. A more detailed description of the locality is given on p. 33 in connection with the Särv nappe in the west. North of R. Biskopsån the mylonite nappe is represented by blastomylonites in which quartz, feldspar, and mica are segregated in alternating lamellae.

South of this locality the narrow zone of the augengneiss could be traced in isolated exposures and glacial boulders up to the col between Mt. Bullfjället and, on the Norwegian side, Mt. Skarvdörrfjällen (D1). From this point the augengneiss turns to the Norwegian side of the boundary, where it passes west of Mt. Haftorstöten (E1). At Mt. Gruvsjöhöjden north of L. Stora Glän (E1) the augengneiss reappears on the Swedish side of the boundary. It is several hundred metres thick there and at the same time most of the mylonite, which at R. Biskopsån was found below the augengneiss, has disappeared. As for the rest, the augengneiss is situated relatively close to porphyry and granite of the Riksgränsen anticline for most of the distance from L. Sylsjön (C1) to L. Malmagen (F1).

At Mt. Rutenfjällen (F1), west of L. Malmagen, the granite is gneissic, and in this way forms a transitional zone. This granite gneiss is directly connected with the Vigel granite in Mt. Rutenfjällen. In these mountains transitions between granite and gneiss with porphyroblastic feldspar can be observed. It seems as if these transitions were repeated, most likely due to the capture of several small nappes of granite on the occasion of the overthrust of the mylonite nappe (cf. analyses, Table 2).

About 500 m north of the northern end of L. Bolagen (F1) I encountered quartzitic sandstone of the same type as the light sparagmite. There the

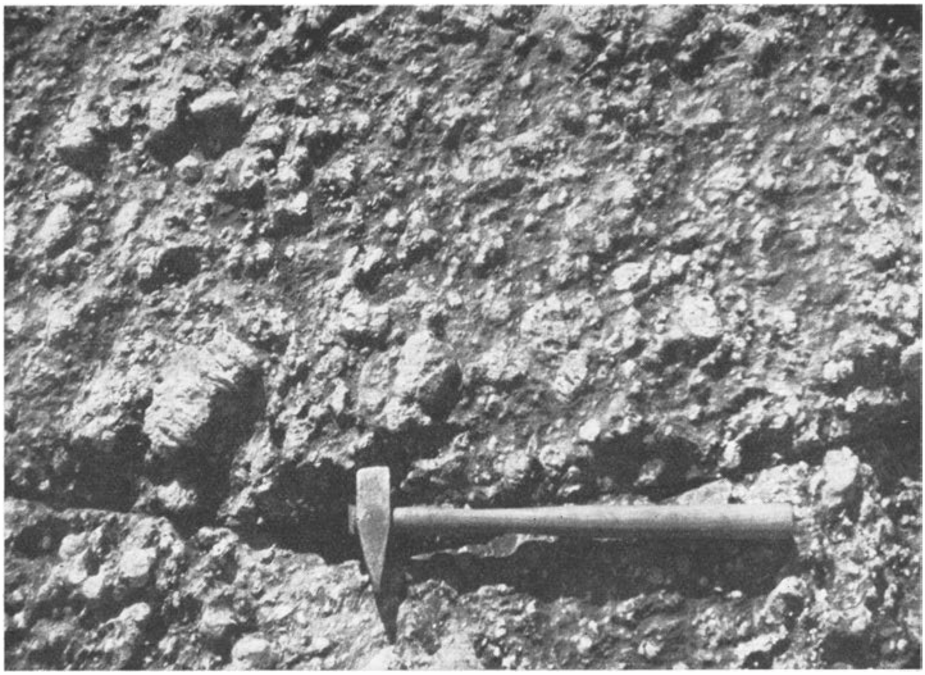

Fig. 5. Augengneiss just below the base of the Särv nappe in the western (Riksgränsen) limb. Water-polished rock surface in R. Biskopsån, cf. Fig. 13 (C1). The microcline porphyroblasts are less eroded and stand out against the surrounding fine-grained crystalline mylonitic rock. The shaft of the hammer points towards N.

porphyry of Vigeln meets the course of the augengneiss at an acute angle, with the light sparagmite squeezed in between its two sides. North of this point and on the Swedish side of the boundary, I have been unable to discover any sedimentary rocks between the granite and the augengneiss. Instead, transitional rocks of a granite gneiss character were found. It thus appears as if in this place all sedimentary rocks have been squeezed out from between the augengneiss and the Archaean rocks of the Riksgränsen anticline. This obviously is in agreement with TÖRNEBOHM's conception as expressed in his general map of 1896. TÖRNEBOHM, on the other hand, has not indicated any occurrence of augengneiss north of Mt. Rutenfjällen between the Archaean of the anticline and the Seve rocks.

G. FRÖDIN (1922, a), who considered the sparagmites to be of Silurian age, has described sparagmitic rocks in the immediate vicinity of the porphyry of the Riksgränsen anticline. He had observed the connection between streaks of granitic composition and the formation of augengneiss, and is of the opinion that occasionally the "eyes" had been formed by compression and disruption of granitic and pegmatitic veins. FRÖDIN admittedly interprets the genuine augengneisses as characterizing zones of pronounced tectonic movements, but supposes at the same time that smaller dikes of augengneiss can be formed by the metamorphism of sparagmite. It seems as if FRÖDIN had not attached any importance to the fact that the streaks and dikes, as well as the formation of augengneiss, which he has described from the regions around the Riksgränsen anticline, are confined to a zone of mylonitic rocks with augengneiss between the schistose marginal zone of the Archaean rocks of the anticline and the overlying sedimentary sequences.

With regard to age and genesis of the eruptive rocks at the Riksgränsen anticline different opinions have been expressed. HØRBYE (1861) and C. W. CARSTENS (1920) have described a felsite-like rock, marginal to the granite of Riksgränsen and connected with the granite by continuous transitions. CARSTENS believed the granite in Mt. Skardörrsfjällen (D1) and at Mt. Sylarna (B1) to be injected into the mica-schist of the Röros group. At the same time a felsitic boundary facies with thickness of several kilometres should have been formed close to the eruptive contact. In this connection CARSTENS refers to TÖRNEBOHM's general map (1896). This very map, however, makes it obvious that the porphyry is not everywhere situated marginally with regard to the schists, but that the latter are occasionally in direct contact with the granite. Among the porphyries TÖRNEBOHM had observed pyroclastic effusive rocks and piercing dikes of the granite of the anticline, and had referred these two rocks to the Archaean. Besides, as observed already by TÖRNEBOHM, the granite and porphyry of the Riksgränsen anticline are traversed by dikes partly of a coarse ophitic dolerite which resembles the Åsby dolerite, and partly of a more fine-grained, uralitized, dark dolerite.

Upon the granite and porphyry of the Riksgränsen anticline G. FRÖDIN (1922 a) observed a marginal weathering zone and, furthermore, powerful tectonic disturbances in this zone, but no marginal transition to a felsitic marginal facies. In accordance with these observations and with the discovery of Cambro-Silurian resting upon the granite west of Mt. Snasahögarna, FRÖDIN interpreted the anticline as a hill of Archaean rocks. CARSTENS (1922) subsequently reports the discovery of a basal breccia on the bedrock of the anticline, and therefore agrees with the interpretation of FRÖDIN and TÖRNEBOHM.

The window of Archaean granite and porphyries at the Riksgränsen anticline is bordered by a zone of cataclastically deformed Archaean rocks. This marginal zone is characterized by a deformation which, together with a strong recrystallization, provides an adequate explanation of the present appearance of the rock. The schistosity in the marginal zone is mainly parallel with the limiting surfaces of the Archaean massif. The nappe of mylonitic schists, which I have found between the schistose marginal zone and the Särv nappe, is additional support for the hypothesis of the pre-Cambrian age of the porphyry and the granite of the anticline.

The Tännäs augengneiss

From the Norwegian border at Mt. Rutenfjällen (F1) and towards the east as far as Lunnäset village (F6) a thick portion of the mylonite nappe is formed by augengneiss and associated mylonitic schists, as well as cataclastic relicts of granite and basic intrusives.

South of L. Malmagen (F1) the upper part of the augengneiss forms a topographically well pronounced ridge which runs in a south-easterly direction, and can be traced for a distance of 4 km. In this augengneiss ridge the *s*-planes dip 45° NE, and it seems as if the overthrust plane also dipped in this direction, though probably at a somewhat smaller angle. It is, however, quite obvious that the thrust-plane has been raised up by deformative forces acting in a NE–SW direction.

Along the valley of the R. Tännån the augengneiss occurs in an anticlinal streak, separating the amphibolitic rocks in Mt. Tännäsbergen (G3) from the corresponding rocks in Mt. Kröksfjällen south of Tännäs village. The deformation which has particularly characterized the augengneiss at Tännäs has a folding axis in a NW–WNW direction, which seems to be of late date. This late folding has made the tectonics west of Tännäs especially complicated. North-west of this place the augengneiss and the Särv nappe have been folded along an axis with a direction about N 55° W. As a result of this folding, both the thrust-plane and s-planes have been raised into a steep position, and most often show dips towards north to north-west. The late refolding of different tectonic units has caused irregularities in the boundaries between the rocks west of Tännäs. This boundary is nevertheless often marked by the occurrence of white quartzite with dolomite-marble (the Ulvberg complex, p. 46).

This folding can be seen not only on the map but also in the section, Fig. 6, which shows the appearance of strongly folded quartzite-schists in the augengneiss of Mt. Långberget west of Tännäs (G3). The quartzitic schists exhibit only a slight tendency towards the formation of porphyroblasts of potassium feldspar. This merits mention, since the surrounding augengneiss is characterized by the presence of well developed porphyroblastic "eyes" of microcline.

From the Opdal region O. Holtedahl (1938) has reported similar observations on sparagmitic flagstone, folded together with augengneiss. In spite of the fact that the *s*-planes of secondary crystallization cross the fold unconformably, no porphyroblasts can be observed in the flagstone. According to Holtedahl this proves the porphyroblastic crystallization to be older than the folding. I. Rosenquist (1944), however, has interpreted the difference in the porphyroblastic structure between the quartzitic schist and the surrounding augengneiss as the result of a primary difference in the chemical composition of these two rocks. He is of the opinion that the dissimilar beds can have become folded together, and afterwards been saturated with solutions containing alkalis. This

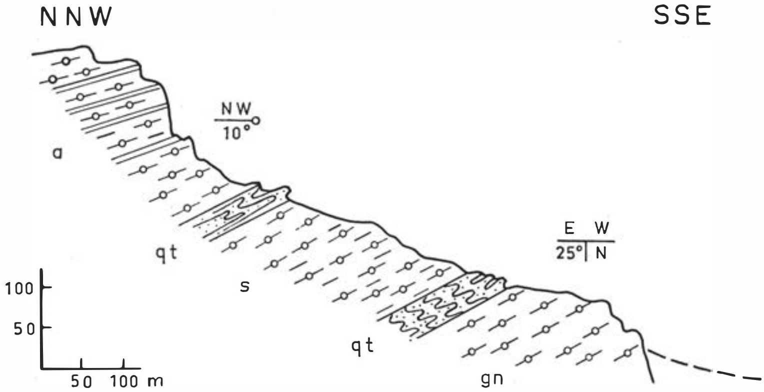

Fig. 6. Quartzitic schists isoclinally folded and imbricated with augengneiss of various type. Section through the southern slope of Mt. Långberget (G3), about 5 km WNW of Tännäs Village. *gn* = granitic augengneiss, *s* = schistose augengneiss, *a* = augengneiss with "anorthositic" bands, *qt* = folded quartzitic schists.

should have led to the formation of porphyroblasts in the bed susceptible to the reaction.

According to my observations, such a susceptibility of a rock to porphyroblastic crystallization of potassium feldspar depends in part upon the primary chemical composition of the rock, but in part also upon the previous existence of an intense deformation.

In the lower part of the southern slope of Mt. Långberget (Fig. 6), the augengneiss is developed with porphyroblasts of microcline, the size of a walnut. These are surrounded by a dark green mass, rich in epidote and consisting of quartz, biotite, and saussuritized plagioclase. In more intensely disturbed zones the porphyroblasts have been flattened into lenses or bands of potassium feldspar, the entire rock assuming a gneissic appearance.

In the uppermost part of the southern slope of Mt. Långberget white bands, an inch thick, appear in the augengneiss (Fig. 7). They are conformable with the dominant *s*-planes of the rock, and consist of a crystalline mass of euhedral zoisite and albite with minor quantities of quartz, potassium feldspar, muscovite, apatite, and fluorite (S.55).

From the Opdal region light layers in the augengneiss have been described by CARSTENS (1922) and ROSENQUIST (1944). CARSTENS, who found them in the augengneiss west of L. Skarsjøen, interpreted these layers as saussuritized dikes of anorthosite. This interpretation is adopted also by ROSENQUIST who is, however, not quite prepared to consider them as being of pre-Cambrian age.

The whole mass of the augengneiss in which these light layers occur on Mt. Långberget is so intensely deformed that it is hardly possible to decide where the material in the white bands might have come from. The designation "anorthositic" layers chosen by CARSTENS assumes, however, a deeper meaning, if it is kept in mind that according to ROSENQUIST the augengneiss in the Opdal

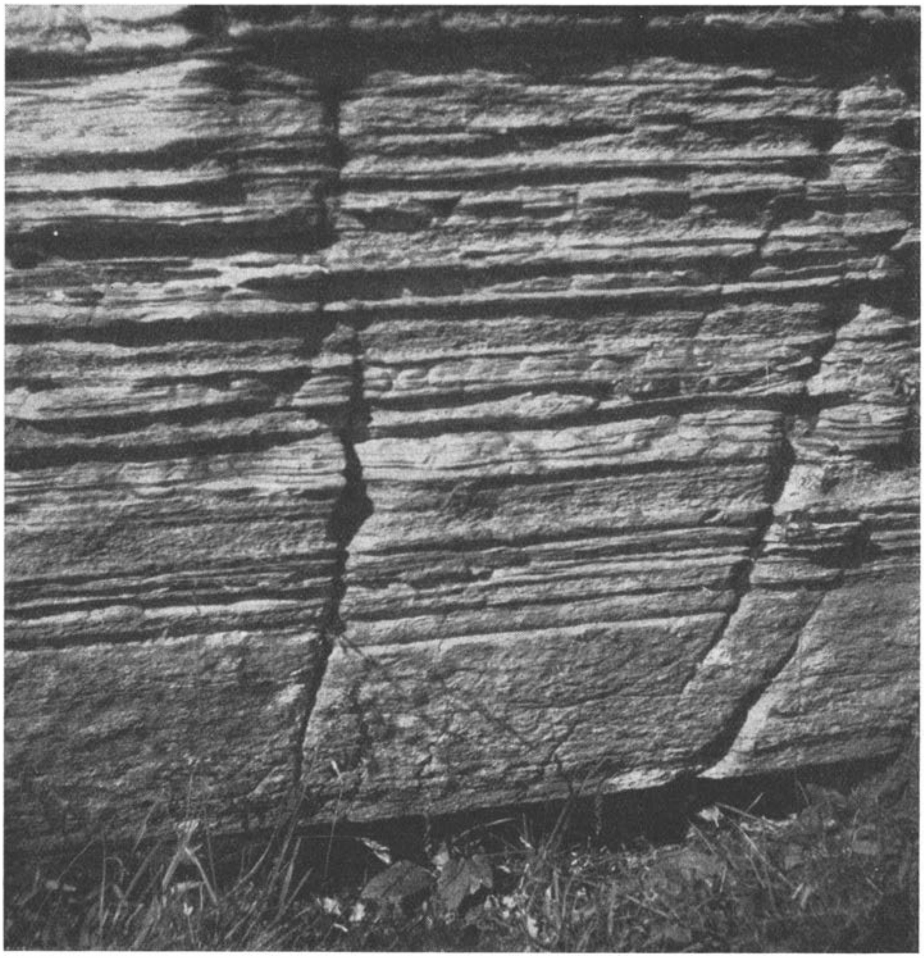

Fig. 7. Augengneiss with white, "anorthositic" bands. Southern slope of Mt. Långberget, cf. Fig. 6 (G3).

region is largely amphibolitic in origin. The amphibolites of the Tännäs complex with feldspar-rich fragments occur as a pre-Cambrian bedrock in the overlying Särv nappe. This Tännäs complex may have been a characteristic component in the bedrock, from which the augengneiss was subsequently formed (cf. p. 83).

Augengneiss east of Mt. Särvfjällen

At Mt. Särvfjällen the easternmost parts of the Särv nappe rest, with only a thin intervening mylonite nappe, upon parautochthonous red granite gneiss similar to the Rätan type of granite.

Large bogs at the foot of Mt. Särvfjället hide, with their glacial and postglacial sediments, the base of the nappe from R. Lunån (F6) as far as Källberget village (E7). A search for exposed rock along this distance gave no result. The region is even devoid of glacial boulders.

The mylonite nappe is, however, clearly developed as augengneiss, both where it disappears south of L. Ned. Särvsjön (F5), and where it reappears at Gavelåsen village (E7). It is therefore reasonable to suppose that augengneiss underlies the Särv nappe in its entire extension east and north-east of Särvsjö village.

North of Mt. Särvfjällen (D6–7) the Särv nappe is thrust almost directly over the Cambrian blue-black quartzite. Schistose dikes of Ottfjäll dolerite in R. Ringbrunnsbäcken show the deformation in the basal zone of the nappe to have been strong. Immediately above the blue-black quartzite upon Svedjelandet (D6), about 1 km east of the brook, a quartz-sericite-schist with a slight westerly dip is exposed (Fig. 2). This schist seems to represent the mylonites in this region.

Similar quartz-sericite-schists occur where R. Ringbrunnsbäcken flows into L. Ned. Grucken (D6), yet also here the thickness of this part of the mylonite nappe seems to be rather small. In the cliffs of Mt. Storberget, immediately to the west of the brook, occurs a light-coloured, quartzitic sandstone with dikes of Ottfjäll dolerite, which seem to be tectonically disturbed at the bottom of the Särv nappe. Not more than 200 m north-east of the mouth of the brook, rocks of the greywacke zone overlying the blue-black quartzite are found.

The area east and north of Storsjö

Eastwards from Storsjö there extends a region of about 100 sq. km in which erosion has created alternating exposures of the basal zone of the Särv nappe and of the mylonites of mainly eruptive rocks, which here constitute the foundation of the Särv nappe.The mylonitic rocks seem to form a continuous nappe extending from this region up to L. Glän (B7). North of this lake it appears in a window in the Särv nappe near the lake system east of Mt. Anarisfjällen (B–C6), and forms likewise an outcrop at Visjövalen (B–C7) between the Särv nappe in the west and the quartzite of Mt. Oviksfjällen in the east (Asklund, 1957).

Within this region the appearance of the schists of the mylonite nappe varies considerably from place to place. Many observations have shown that the rocks which have been deformed into mylonites have been mainly granites of varying composition and with amphibolitic rocks appearing in them.

North-east of Storsjö greenish grey mylonitic rocks appear in an area extending from the highway at R. Bottenbäcken up to Mt. Storåsen (D4–5). These greenish grey, often porphyroblastic mylonites, which include oligoclase-albite in porphyroblasts attaining sizes of up to 1 cm, occur in several outcrops upon Mt. Storåsen, where they alternate with granite gneiss and granite mylonite.

Upon one of the easternmost peaks of Mt. Storåsen these porphyroblastic albite-schists occur in the immediate neigbourhood of a great massif of amphibolite. The amphibolite is of a porphyritic type and only slightly influenced by tectonic deformation in the inner parts of the massif.

The connection between the amphibolite massif and these porphyroblastic schists is so obvious that we are compelled to assume that the porphyroblastic albite-schists have been formed of mylonitized material from the amphibolite.

Near the highway at R. Bottenbäcken there is a small outcrop of a greyish-green greywacke, which seems to belong to the Grönstack complex.

At Mt. Björnhåberget and in its vicinity about 5 km east of L. Storsjö (D5) a pale-weathering kakerite granite is found. As the result of cataclastic deformation of the granite only isolated grains, measuring some millimetres, of potassium feldspar are left as porphyroclasts in a mylonitic groundmass. In the north-west the granite abuts fairly directly against light-coloured quartzite with only moderate disturbances between these rocks. In the east the granite passes into a belt of blastomylonite. This rock is a grey muscovite-schist, which contains laterally orientated, coarsely crystalline secretions. These pegmatitic streaks and secretions contain the minerals quartz, potassium feldspar, siderite (or ankerite), and scaly hematite (cf. Pl. VII a), According to investigations by I. Rosenquist (1951) on similar Norwegian pegmatites, the hematite which occurs as thin plates between the feldspar crystals may contain that iron, which earlier has been bound in the feldspar lattices.

The blastomylonite with the pegmatite-like secretions differs in composition and metamorphism from most other rocks in the mylonite nappe. In my opinion, this may be due to the fact that it has been formed for the most part from material of the relatively salic granite in Mt. Björnhåberget, perhaps with additions of quartzitic material.

In the region north of Hede (F5–6), I found similar rocks as characteristic elements of the uppermost part of the augengneiss and especially as forming the base of the lowermost white quartzite that occurs next to the augengneiss in the basal Ulvberg complex (cf. p. 46) of the Särv nappe. The grey blastomylonite east of Mt. Björnhåberget occupies a tectonically analogous position, since at Falvik, east of R. Ljungan, it is directly overlain by the Särv nappe.

At Norrsätern (D5) strong recrystallization has transformed the granite mylonite immediately below the basal zone of the northern Särv nappe into an almost gneiss-like rock. On the whole, both the mylonites and the entire Särv nappe have, throughout, undergone recrystallization in the northern part of the area surveyed. As a result of this circumstance the rocks of the mylonite nappe are represented at L. Glän and east of it by blastophyllonites, which contain nodules of quartz and are laminated by foliation (Fig. 9).

Everywhere east of Storsjö (D5) the base of the Särv nappe is strongly schistose, this schistosity extending several tens of metres into the overlying nappe. The mylonitization of dolerite dikes in the basal zone and the abundance of

Fig. 8. Dark mylonite with boudinaged amphibolitic layer. 1/10 natural size. Outcrop by the forest road to Tossåsen village, near the southern end of L. Yttre Röversjön (D5).

chlorite and other iron minerals resulting from it, often give the basal schists a dark colour. On account of this, it is sometimes difficult to distinguish the schists of the basal zone, which in the region of Storsjö are wide-spread, and the granite mylonitic schists, which can be seen here and there below the basal zone.

In places where the thrust surface is steeply upfolded, e.g. at the Tossåsen road, the thrust zone between these schists is fairly clear. Pl. VI b shows the strong brecciation of the granite mylonite and of the basic rocks appearing in it. In this rock surface cross-cutting fissures filled with different kinds of mylonitic material are found. Among them we distinguish especially narrow violet-coloured veins of ultra-mylonitic material containing iron oxide ore minerals. As I have mentioned on a previous occasion (S.55), they seem to be especially characteristic of zones of movement. It appears, moreover, as if the formation of these violet veins was conditioned by the presence of granitic material.

The brecciation of the granite mylonite seems to be connected with differential movements that have taken place after the first folding of the entire rock complex. On Mt. Röverhöjden, to the west of the locality just mentioned, the dips are exclusively easterly, both in the granite mylonite and in the basal zone of the Särv nappe. The overthrust zone between these two units is, however, extremely diffuse. It almost seems as if several minor upthrusts in the boundary zone of the nappe had contributed to the blurring of this contrast. After compilation of these observations, the geological development might be visualized in the following way: (1) Formation of mylonite in the mylonite nappe in connection with strong, differential movements; (2) local refolding of the two nappes; and (3) renewed differential movements, causing brecciation in the mylonites and disturbances in the level orientation of the thrust zones.

Granite mylonite proves to be less competent than amphibolite during deformation. Fig. 8 shows a boudinage of amphibolite in granite mylonite, where the granite mylonite has flowed in plastically between the fragments of the torn-off body of amphibolite. This difference in competence might account for the fact that according to numerous observations only amphibolite has been brecciated, while granite has become mylonitized. Thus it is not necessary to adduce the occurrence of relatively slightly deformed amphibolite in granite mylonite as proving the intrusion of the basic rock into already mylonitized granite, although the possibility of such a process cannot be altogether excluded.

A granite mylonite without excessive recrystallization often exhibits a characteristic very intense, crinkly folding along steep axes (Pl. VI a). In granite mylonite that has undergone a more intense deformation and recrystallization the metamorphism has, on the other hand, resulted in a crystallization banding with a separation of lighter and darker minerals in different bands. This makes the above-mentioned small folding less distinct.

The study of the mylonites in the region east of Storsjö has been of great value, as these rocks show relatively moderate recrystallization. A strong crystallization encroaches so much upon the microstructure of the rock that in some cases only a comparison of macro- and microstructures with those of a non-recrystallized mylonitic rock can reveal the recrystallized rock as a primary mylonite. In this way it has been possible to distinguish strongly recrystallized mylonites in the Anaris region and in the Helags nappe.

Mylonites and phyllonites east of Mt. Anarisfjällen

At the eastern foot of Mt. Anarisfjällen the bedrock consists of soft schists, rich in mica and with a varying content of feldspar. These rocks constitute a part of the mylonite nappe which can be traced in the region east of Mt. Anarisfjällen (A–B6) in the shape of an anticlinal ridge with a N–S strike.

The area between Mt. Anarisfjällen and Mt. Oviksfjällen (B6–7) is comparatively poorly exposed, the landscape being flat and to a large extent covered with small tarns and bogs separated by drumlin-like moraines with large boulders. Across this area passes the watershed between Härjedalen and Jämtland. The scarcity of exposures is in a certain measure well compensated for by the local occurrence of often huge moraine boulders.

Most common among the rocks in the mylonite zone extending from Rundvalen (A5) farthest in the north to Mt. Kojåsen at Oldbergsvallen (C6) is a dark blastophyllonite with quartz nodules (Fig. 9). This rock may have been

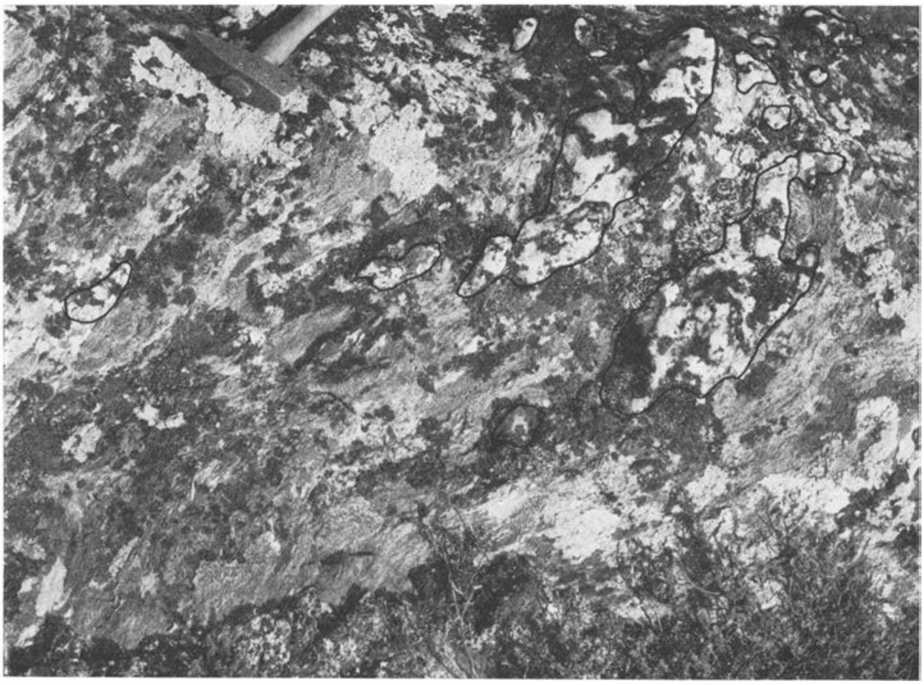

Fig. 9. Blastophyllonite with quartz nodules. Since the surface is partly covered with lichens, a black line has been drawn to outline the quartz nodules. The shaft of the hammer points ESE. Isthmus between L. Södra Långtjärn and L. Anasjön (B6).

a phyllonite (*finely foliaceous tectonite with "skins" of mica*, B. SANDER 1948 I, p. 152), which was completely recrystallized after deformation, and consists of biotite, muscovite, quartz, plagioclase and epidote, all arranged in layers (Pl. X b).

West of the South and North Långtjärn lakes (B6) the nappe contains banded blastomylonitic schists with bands of varying colour. Red bands contain flattened and recrystallized porphyroblasts of potassium feldspar. In addition to them greyish bands with muscovite, oligoclase and quartz are observed.

South from this place, between R. Hällån and L. Anasjön (B6) glacial boulders occur of a moderately coarse red biotite granite, irregularly traversed by a dark eruptive rock. These two eruptive rocks are brecciated together and partly mylonitized. As boulders of amphibolite and boulders exhibiting the contacts are very common, it must be assumed that the amphibolite must be well represented in the granite. Thus granite with amphibolite must occur either in the basal portions of the Särv nappe or in the form of relict core portions in the mylonite nappe.

Near the former Lapp encampment Stora Lövdalen (B6) there occur, in R. Lövån, boulders of a yellow dolomite-marble with long needle-shaped crystals of tremolite. This dolomite-marble seems to be the carbonatic rock that belongs to the basal Ulvberg complex of the Särv nappe. Its occurrence indicates that the thrust zone of the Särv nappe lies between Stora Lövdalen and R. Hällån, where a dark phyllonite with nodules of quartz has been encountered.

Between the northernmost part of Mt. Anarisfjällen and Mt. Falkvalen a narrow anticlinal elevation of mylonite is exposed. This shows that the mylonite nappe probably appears below the Särv nappe also in Rulldalen valley (A5).

At Mt. Falkvalen, Mt. Bredsjöhagen and farther to the south-south-east (B–C6) there appear pale-weathering, dark, macroscopically dense rocks, which exhibit a steep jointing, but no distinct schistosity. Among these rocks a coarse mylonite or microbreccia, with fissures filled with chlorite, has been found. This rock seems to have had primarily a greywacke-like composition. In addition, I found in the same zone a plano-parallelly crystallized blastomylonite, rich in titanomagnetite and titanite, and with porphyroblasts of albite 1 mm in size. It is remarkable that, in spite of a distinct, early cataclastic structure, these rocks are strongly recrystallized, so strongly indeed that, apart from the vertical fissuring, they lack a macroscopically observable schistosity. The rocks themselves give the impression of being massive and only slightly disturbed tectonically. The recrystallization seems to be altogether postdeformative. At Mt. Falkvalen an almost massive, fine-grained, black greywacke rock occurs.

Below the dark crystalline rocks light-coloured quartz-muscovite-schists are found in the bottom of the valleys. The dips in these schists and in the mylonite zone west of it suggest that the dark crystalline rocks in the zone from Falkvalen in the north to Höstanstöten in the south (B–C6) overlie the mylonite nappe, and represent an eastern marginal, tectonically affected part of the Särv nappe. The absence of dikes of dolerite in this part of the nappe need not by itself disprove this interpretation since, according to my observations along the eastern border of the Särv nappe towards the south at Mt. Gavelåsen, Mt. Källberget, and Mt. Håsjöruet (E7), the dikes of the Ottfjäll dolerite appear only seldom so far eastwards.

Still farther eastwards the mylonite nappe reappears, and is thrust steeply over the blue-black quartzite in Mt. Hosjövalen (B7) in the north and Mt. Kokdalsfjället (C7) in the south. At Mt. Sandbodåsen (C7) and in the northern part of Mt. Busjöfjället (C7) the rocks of the mylonite nappe are developed as augengneiss with porphyroblasts of pink microcline perthite. The perthite seems to have been formed by replacement of porphyroblastic albite with microcline (Pl. IX a).

At Tossåsen village (C5) there seems to be a small window of the Särv nappe through which the basal zone and mylonites are exposed.

The Lake Ottsjön area

The highway to Vålådalen passes R. Vålån (A5) near a waterfall, where after piercing the mylonite nappe, erosion has exposed an underlying hump of coarsely crystalline biotite granite.

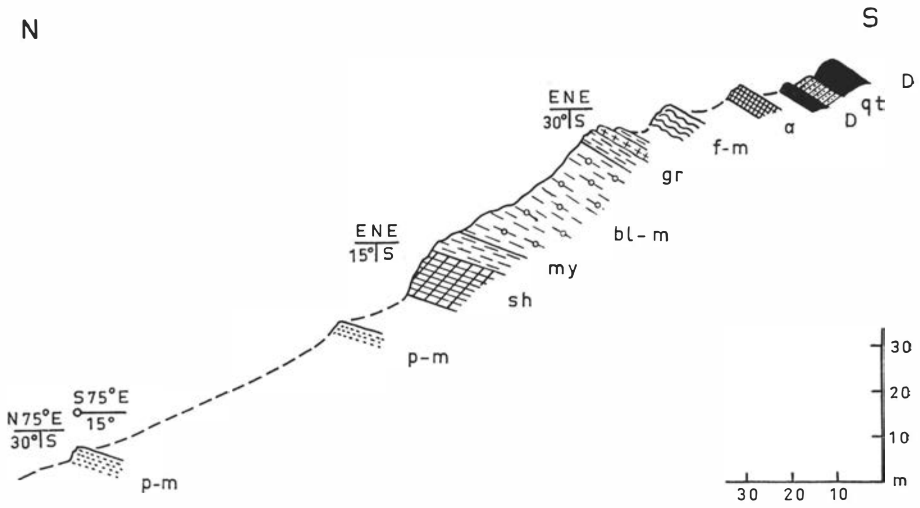

Fig. 10. The mylonite nappe in Mt. Ottfjället. *p–m* = schistose and lineated porphyry mylonite, *sh* = black slate, *my* = light-coloured mylonite, *bl–my* = blastomylonite, *gr* = granite, *f–m* = folded mylonite, *a* = amphibolite, *D* = Ottfjäll dolerite, *qt* = steeply dipping quartzitic sandstone. Section through the northern slope toward the saddle between Mt. Ottfjället and Mt. Rekhuvudet (A4).

Near the bridge downstream, mylonitic schists are found which show exclusively northerly dips. Their degree of mylonitization increases towards the north until the schists gradually become a hard dark mylonite with nodules of quartz. Isolated nodules or bands of carbonatic minerals also appear. The mylonite occurs rather uniformly from a point 100 m below the bridge to the termination of the exposures about 400 m farther down.

Above the bridge the granite can be observed to become increasingly gneissic though the exposures extend only 200 m south of the bridge. Glacial boulders of mylonitic schists suggest the occurrence of these rocks farther towards the south. An outcrop of cataclastic Ottfjäll dolerite occurs in the middle of the river at a point about 600 m south of the bridge. Another 200 m upstream a rocky patch with a bedrock of greywackes is found near the river. The latter contain southward dipping dikes of Ottfjäll dolerite.

It is thus obvious that the granite ridge at R. Vålån is overlain by a nappe of mylonitic schists followed by the typical rocks of the Särv nappe.

At Mt. Rekhuvudet, south of the western end of L. Ottsjön (A4), there is an exposure of quartz-porphyry that has been hardly deformed at all. This has already been reported by HOLMQUIST (1894). On Mt. Ottfjället this quartz-porphyry is overlain by the Särv nappe, being separated from it by thrust planes. The boundary between the porphyry and the Särv nappe passes the mountain ridge in the col between Mt. Rekhuvudet and Mt. Ottfjället. In the eastern part of Mt. Rekhuvudet the porphyry exhibits a mylonitization which increases towards the south, its *s*-planes dipping towards SSE under Mt. Ottfjället.

In the southern part of the col a hard, schistose porphyry mylonite is exposed about 80 m from the almost undisturbed rocks of the Särv nappe. The mylonite nappe is evidently not very thick, and almost disappears against the high porphyry ridge of the Mt. Mullfjället anticline. The quarzite and greywacke of the Särv nappe contain dolerite dikes with a gentle southerly dip.

Farthest down in the northern part of the col lies a hard mylonite schist, formed probably from the porphyry of the basement. On top of it follow about 10 m of a black slate (see section, Fig. 10). This in turn is overlain by 5 m of a light-coloured hard mylonite upon which rest 40 m of mylonite with porphyroblasts of pink potassium feldspar, about 1 cm in size. The uppermost portion of the slope contains a 2 m thick mass of coarsely crystalline granite with hornblende, containing a fissure about one inch thick filled with microcline and quartz. These rocks are overlain by banded and folded mylonites composed of alternating dark and light layers, the dark layers resembling most closely amphibolitic greenschist.

The Särv nappe rests on top of the banded mylonitic schists. In its basal zone the dikes of Ottfjäll dolerite are flat-lying and metamorphosed into an amphibolitic rock in which no traces of the porphyritic structure that is characteristic of the undisturbed dolerite dikes can be observed.

G. Frödin (1922 a) considers a similar type of amphibolite characteristic of the more strongly metamorphic basal and marginal zones in the Särv nappe (by Frödin called the sparagmite-schist field). He reproduces a photomicrograph of the schistose rock, and designates it Seve amphibolite that had been formed by metamorphism of the Ottfjäll dolerite.

I observed moreover amphibolite of this type in several places near the basal zone of the Särv nappe. Both east of Tossåsen village and in the area east of Storsjö, similar amphibolites appear as gently dipping massifs in the basal part of the nappe. In most of the amphibolites in this position there can be distinguished a schistosity and sometimes a lineation. Near the border of the massifs, *s*-planes and lineations are conformable with the deformation structure of the surrounding bedrock. Thus it seems as if the amphibolites of the type mentioned are cataclastically deformed and recrystallized bodies of an eruptive kind of rock which originally may have had some similarity to the Ottfjäll dolerite.

The Särv nappe

The southern part

In the southern part of the Särv nappe the topography is especially characterized by sharply serrated mountain ridges, every top in the ridge being marked by a broad dike of Ottfjäll dolerite. The bedrock is dissected by a system of valleys which have cut through the nappe almost to its base. The valleys dissect the Särv nappe into a number of isolated blocks which are connected only in the lower part of the nappe. It seems as if the valleys mark zones of disturbance in which the more numerous fissures have promoted the erosion. Each of the remaining blocks forms by itself a small mountain massif, examples being Mt. Funäsdalsberget, Mt. Ormberget, Mt. Anåfjället, Mt. Lillfjället, and Mt. Särvfjället.

Zones of tectonic disturbances occur, however, not only along the valleys, but also higher up in the blocks of less disturbed rocks of the nappe. In the places where such disturbed zones could be observed, they are indicated both by a strong schistosity of the sedimentary rocks and by a shearing of the dolerite dikes.

In the southern part of the Särv nappe the bedrock is built up mainly of quartzitic sandstones which are traversed by preponderantly subparallel dikes of dolerite. These can often be followed from one mountain ridge to another, and also across the intervening valleys, where, however, they often change direction. Among the sedimentary rocks there are, beside fine-grained quartzitic sandstones, a coarse sandstone rich in microcline and with conglomeratic layers, and a zone with dark banded slate near the bottom of the series. As basal sediments of the Särv nappe sequence proper, some peculiar fragmental rocks have been observed.

The sedimentary rocks are partly metamorphosed into quartzitic schists, the mineralogical composition of which depends upon the intensity of the deformation and the type of the primary rock. In addition to a general recrystallization of quartz, muscovite has formed out of microcline, together with a finely dispersed potassium feldspar. Especially in the dark arkosic members of the sedimentary sequence a recrystallization has taken place which seems to be younger than the deformation of the complex. In this connection, a formation of chlorite, epidote, and muscovite has taken place in the southern part of the Särv nappe.

In some places, e.g. at Mt. Röstberget near Ljusnedal village (F3), granitic rocks occur in the overthrust bedrock of the nappe. In the southernmost part of the nappe we find, near Tännäs, a number of massifs of amphibolitic rocks, which show a certain analogy with the Bergen-Jotun group. All these eruptive rocks seem to form parts of the primary pre-Cambrian bedrock below the sediments of the nappe.

A ZONE WITH MYLONITIC SCHISTS.—West of the southern part of the Särv nappe extends a region, several kilometres wide, of hard schists, frequently with a gentle westerly dip. In this region the topography is determined by the nature of the bedrock, the landscape being dominated by flat low mountains and bogs. Thus the region of the schists includes areas known by their low relief like Tjölen (E3) with a drumlin landscape and small tarns between Bruksvallarna and Mittådalen, and the low mountain Flatruet (E4) north of Mittådalen.

In the eastern part of the region, from Bruksvallarna (E–F2) across Tjölen and Flatruet, only hard quartzitic schists containing muscovite and chlorite are found. In spite of the metamorphism a blastopsephitic relict fabric with pebbles of quartzite and pink or grey feldspar, 1–5 cm in diameter, can occasionally be observed in these schists. This indicates that prior to the development of schistosity the rocks have been very similar to the light-coloured quartzites common in the southern part of the Särv nappe.

In this region no dikes of Ottfjäll dolerite have been observed. The dolerite dikes which are closest to the region of the schists are strongly schistose. An occurrence of dikes of schistose dolerite and quartzite is found in R. Stadsån at the eastern end of L. Messlingen (E4). Also in the mountains south of L. Messlingen all dikes of dolerite are strongly deformed. This indicates that deformation and metamorphism of the schists of Flatruet have taken place largely after the intrusion of the Ottfjäll dolerite.

Since the quartzitic schists are only slightly folded they exhibit no pronounced β-direction, but a lineation of the type *Striemung* in about N 55° W (Pl. III). In schists upon Tjölen appears also a lineation in a northern direction.

In the west the quartzitic schists merge into granite mylonite which lacks any trace of blastopsephitic relicts, but sometimes contains porphyroblastic potassium feldspar. At the highway in the village of Bruksvallarna this can be observed especially distinctly. In the western part of the village also a granitic breccia has been noticed.

Between the Helags nappe and the Särv nappe the granite-mylonite forms a special, local nappe. It could not be decided, whether the granite mylonite is part of the original basement of the Helags nappe which has been dragged along during its overthrust or if it might perhaps mark a window in the Särv nappe through which an anticlinally folded part of the mylonite nappe has projected, and has in part been thrust secondarily over the eastern portions of the Särv nappe.

The northern part

The northern part of the Särv nappe extends from Mt. Flatruet (E4) in the south to L. Ottsjön (A4) in the north, and in an E–W direction from the lake region east of Mt. Anarisfjällen (B6) to the upper reaches of R. Ljungan and R. Vålån (B–C3) in the west. Dikes of the Ottfjäll dolerite, that are typical of the nappe, are found nearly everywhere within this area.

It must be admitted that the Ottfjäll dolerite seems to occur everywhere rather uniformly in the shape of subparallel swarms of dikes with a western dip, but the dikes seem nevertheless to be thinner in the northern part of the nappe than in the southern. Purely topographically, a distinct difference can be noticed between the southern and the northern parts of the nappe. In the former, the outcropping of the thick dikes often produce sharp serration in the mountain ridges, while in the latter these features are more or less exceptions (Mt. Dunsjöfjället G3), and the mountain peaks exhibit, as a rule, normal rounded shapes.

The southern and the northern parts of the Särv nappe also differ in the fact that the rocks of the latter are in general more highly metamorphosed. This takes the form of recrystallization and partly the formation of biotite. The sedimentary rocks, of which this part of the nappe is composed, seem likewise to exhibit a stratigraphical development that differs from that of the corresponding rocks in the southern part. Thus siltstones or greywacke-like sedimentary rocks are found to a greater extent in the northern part of the nappe. The greywackes often contain magnetite, which as a rule marks the bedding of the rock, but which may occasionally be distributed in spots throughout it. Metamorphism has led to the formation of chlorite, biotite and a partial recrystallization of quartz. These metamorphic sandstones and greywackes sometimes seem to be rather similar to the sequence of blue-black quartzite and greywackes in the eastern parautochthonous region.

In an area between Mt. Tandåsen to Mt. Högrensvålen (C–D4) about 2 km west of Storsjö, crystalline, biotite-bearing, almost gneissic rocks are found. Primarily, these rocks seem to have been sandstones and greywacke-like sediments the present metamorphic state of which is a result of preponderantly postdeformative recrystallization.

In connection with certain inner disturbances, local deviations from the normal metamorphism appear in the rocks of the nappe. A tectonic disturbance of this kind is found in Vålådalen valley (A4), where a relatively strong schistosity occurs. The deformation there has obviously been stronger than elsewhere. This might depend upon movements of blocks which have caused the strike of the dolerite dikes on Ottfjället to run E–W, indicating a possible postintrusive rotational movement counterclockwise through almost 70° with regard to the NNW strike, which is common in most other parts of the nappe (Fig. 24).

It is, however, also possible that the Särv nappe is especially thin just in Vålådalen, where the erosion could be supposed to have reached as far down as the schistose sole of the nappe. The presence also in the northern Särv nappe of such a schistose sole is demonstrated by the conditions at the southern limit of this part of the nappe in R. Storån (C5) where tectonic disturbance, schistosity and shearing of the dolerite dikes increase gradually towards the granite gneiss and hard mylonite schists, which there underlie the nappe.

In the easternmost part of the northern Särv nappe some other deviations from the otherwise rather uniform structure of the nappe also occur. The top of Mt. Tossåsfjället is well exposed, and there we can observe an apparently irregular network of narrow (0.1–1 m) dikes of dolerite traversing the quartzitic sedimentary rock which by recrystallization has acquired a granular appearance. In the valley east of Mt. Tossåsfjället, on the other hand, a large massif of amphibolite is exposed. This rock resembles the rocks of the Tännäs complex which occur also east of L. Röversjön.

Immediately east of this massif of amphibolite, from Mt. Finnsjöfjället to L. Röversjön (B6–D5), there is a large area (C6) with clastic, light-coloured quartzite which markedly resembles the quartzite in Mt. Anåfjället (E–F3). Still farther towards the east, at Mt. Höstanstöten (B–C6), recrystallized and biotite-bearing quartz-muscovite-schists and grey pelitic schists occur.

The tectonic relation between the light-coloured quartzite and the surrounding, more crystalline rocks is still not known with certainty. Rocks, which on the whole are of the same types as those at Mt. Höstanstöten, continue towards the north in the zone mentioned between L. Anarissjön and L. Håsjön (B–C6). In spite of the absence of dolerite dikes, this zone of dark, finely crystalline rocks has been interpreted as an easternmost part of the Särv nappe (cf. p. 24). There may be a thrust plane between the light quartzitic sandstone and the crystalline, dark, biotite-rich sandstones and siltstones with greywackes, which all appear in the northern part of the Särv nappe. Movement of one of these sequences in relation to the other should have occurred at an early orogenetic stage, when the dikes of dolerite had not yet come into existence, as they rather uniformly cut through nearly all the sedimentary rocks of the nappe. Besides the mentioned irregularities in the appearance of the dolerite dikes on Mt. Tossåsfjället no clear evidence of such a thrusting has been observed.

The western limb

The Särv nappe reappears on the western side of the synclinorium in which the Helags nappe rests. This western limb of the nappe is compressed between the pronounced hump of crystalline Archaean rocks belonging to the Riksgränsen anticline and the Helags nappe.

It cannot be proved, anywhere along the 60 km from Fjällnäs village (F1) to Mt. Blåhammaren (A1), that the Särv nappe was completely squeezed out. Wherever sufficient exposures have permitted nearly complete observations, it has, on the contrary, shown itself to be relatively intact, although the deformation of the rocks of the nappe can be very intense locally.

The thickness of the Särv nappe, however, often seems to be considerably smaller in the west than in its occurrence east of the Helags syncline. In the west, metamorphism and recrystallization of the rocks of the nappe are often considerably more intense than in the east (Pl. XII b). This circumstance will receive special attention in the section on metamorphism.

The contrast in the degree of metamorphism between the rocks of the Särv nappe and the Helags nappe has, in the west, decreased to such an extent

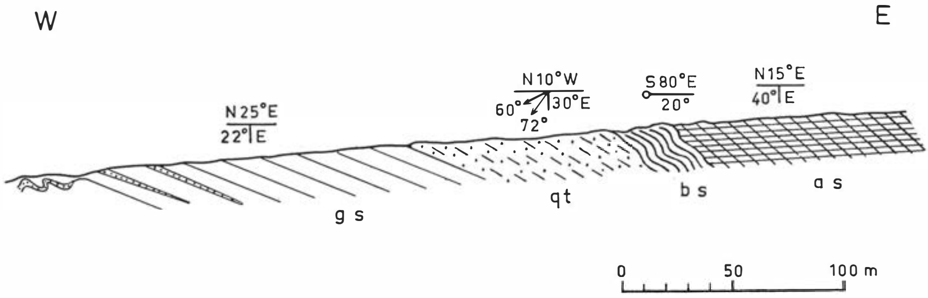

Fig. 11. The Särv nappe underneath the Sylarna amphibolite. *gs* = green amphibolitic schists with bands of quartzite, *qt* = quartzitic schists, in part intensely folded (complex folding), *bs* = intensely folded, banded amphibolitic schists, *as* = dark, crystalline amphibolitic schists. Section along the upper reaches of R. Fiskån tributary on the western slope of Mt. Sylarna between points 1385 and 1379 (B1, Norway). Vertical scale = horizontal scale.

that the determination of the exact position of the thrust zone between the nappes has proved difficult. In this connection the presence of comparatively undisturbed dikes of Ottfjäll dolerite in the Särv nappe has been used as a safe criterion. On the general map (Pl. I) the metamorphic psammitic rocks of the Särv nappe in the western limb have been designated with the colour of biotite-bearing greywackes and quartzites.

Regional description.—Starting from the bend in the complex of nappes at L. Malmagen and for a distance of about 5 km northwards no dolerite dikes could be found in the Särv nappe. There the boundary against the Helags nappe can be considered to be marked by a "dark hornblende-schist", some tens of metres thick and containing nodules of quartz, which is overlain by a "gneiss schist resembling a mica-schist" (A. E. Törnebohm, 1896). These rocks seem to be recrystallized mylonites that belong to the overthrust zone between the Särv and Helags nappes and might for this reason represent the western limb of the granite mylonites which appear in a tectonically analogous position at Bruksvallarna (F2) in the east.

The dolerite dikes of the Särv nappe make their first appearance as far north as near Mt. Glänvålen (E1). The dikes are 10–20 m wide, and strike NNW. The contacts with the surrounding crystallized quartzite dip at about 45° towards the west, but are often greatly disturbed by deformations. Törnebohm (1896) refers to observations of primary sedimentary structural characters in the quartzitic rocks of this region, and also marks a dolerite dike on his map.

At Mt. Grönvålen (E1), north of Gröndalen, alternating dolerite and quartzitic rocks reappear. Here the dikes of the dolerite have a rather flat position which might be the result, at least in part, of an internal rotation towards the southeast. Contemporaneously, the rocks were relatively strongly deformed with the result that the dolerite has been metamorphosed into schistose amphibolite, and the sedimentary rocks into quartz-muscovite-schist.

Near Mt. Haftorstöten (D1), in the neighbourhood of the lake from which R. Ljusnan takes its origin, there is another, fairly undisturbed block of the Särv nappe. The Ottfjäll dolerite occurs in a series of dikes, 10–50 m wide, with a steep western dip. The rocks are only slightly deformed. Still farther to the north, near L. Sylsjön, the bottom of the Särv nappe is exposed in two small cascades in R. Biskopsån (C1). The locality, which can be considered representative of the development of the Särv nappe in this western area, has been mapped in detail (Fig. 13).

At L. Sylsjön (C1) the Särv nappe turns westwards into Norway. The region west of Mt. Sylarna has been studied by the Norwegian geologist K. Ø. BRYN (1959), who describes an outlier of the Helags nappe in Øifjeld, west of the Riksgränsen anticline. He records, however, no observations of rocks which could correspond to the complex of the Särv nappe. It thus seems that there this nappe has been subjected to such a strong metamorphism as to render its identification difficult. A note by A. E. TÖRNEBOHM (1896) on dolerite dikes in sparagmitic rocks in Mt. Gråsiden south of L. Stuesjø shows nevertheless that the rock complex of the Särv nappe occurs immediately west of the Riksgränsen anticline. As has been reported (TÖRNEBOHM, 1896; P. HOLMSEN, 1937) dikes of dolerite, probably of the Ottfjäll type, occur also in Mt. Hummelfjeld.

A series of rocks, which seems to represent the upper portion of the Särv nappe, appears along a ravine in the upper course of a tributary of R. Fiskån between the peaks 1385 and 1379 (B1) (Sheet 43 D, Stuesjø, *Topographical Map of Norway*, 1:100,000). According to the section Fig. 11, on the top occur dark amphibolitic schists, underlain by light-coloured, quartz-rich schists, alternating with amphibolitic schists. Especially near the top, the light-coloured schists contain large crystalloblasts of muscovite. In the lower, western part of the section the folding is extremely intense. Fig. 12 shows a folded contact between quartzitic rock and amphibolite. As a result of its lower competence the amphibolite shows flexural slip folding in a bundle of small-scale drag folds.

Unfortunately this locality offered no possibility to study the lower and most characteristic part of the Särv nappe. On the other hand, it was possible in this region to follow a zone of light-coloured mylonite along the boundary of the porphyry of the Riksgränsen anticline. This mylonite is at least 100 m thick.

THE BASAL FORMATIONS AT R. BISKOPSÅN.—About 500 m SSW from the spot where R. Biskopsån discharges into L. Sylsjön (calculated at high-water level in the dammed-up lake), this brook forms two small cascades where the water has worn a passage through the steeply inclined schists. These rocks consist partly of the bottom formations of the Särv nappe with white quartzite and limestone in which the Ottfjäll dolerite appears in a series of sills and partly of mylonitic schists with amphibolite. The uppermost part of the mylonite nappe is developed as augengneiss.

The entire complex is folded with the axis of folding in the direction

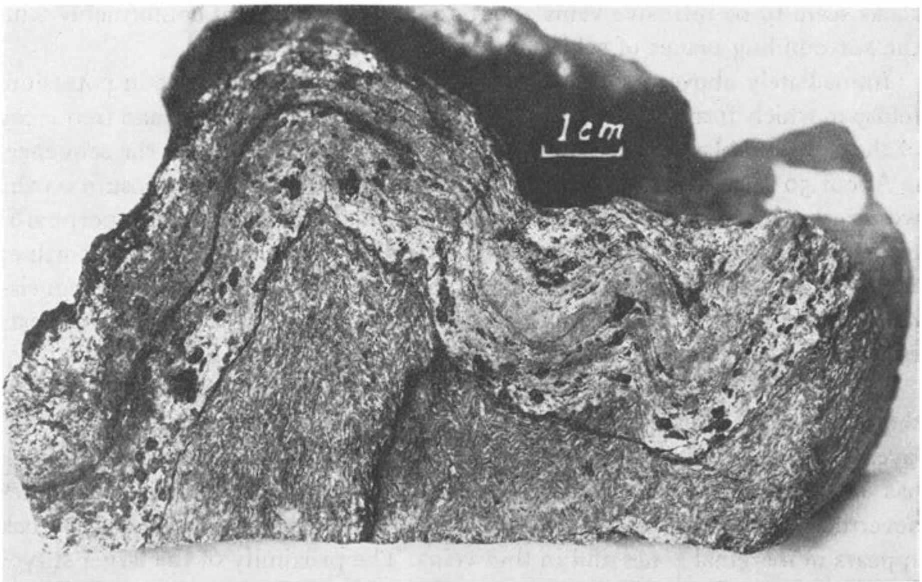

Fig. 12. Folded contact between quartzite and amphibolite. The cleavage in the incompetent amphibolitic rock has developed by shearing of closely spaced drag folds. The axis of folding dips N 40° E/10°. The westernmost part of the section in Fig. 11 (B1). Photo G. Andersson.

N 70° W. The amplitude of the folding amounts to 1–10 m. On account of this folding the contact between the augengneiss and the rocks of the Särv nappe is steeply raised and, like the sedimentary rocks, somewhat tilted towards the south. The distribution of the rocks can be seen in the map Fig. 13.

From the exposures at this locality it cannot be seen how far the mylonitic schists extend westwards before meeting the porphyry of the anticline. The nearest exposure of the porphyry lies at a distance of about 800 m towards the west. There the porphyry has, however, become schistose in a belt several hundred metres wide and is almost mylonitic. This indicates that the mylonite nappe proper can approach these porphyry rocks rather closely. It thus appears possible that the thickness of the mylonite nappe has to be estimated at several hundred metres.

The dark mylonite-like schists appearing farthest downstream along R. Biskopsån all seem to be postdeformatively recrystallized, and contain quartz in the form of bands or elongated nodules. In the lowermost part, the rocks are strongly schistose with only slight and gentle folds. The folding increases towards the south and higher up in the sequence, while at the same time the composition of the schists becomes more variable.

In the outcrops somewhat north of the centre of the map the brook forms a small waterfall. Next to it, the bedrock is strongly folded. The mylonite near the waterfall contains two types of amphibolites, viz. a black amphibolite, poor in plagioclase, and a greenish rock resembling greenschist. These two rocks seem to be intrusive veins about 10 cm thick, injected conformably with the surrounding planes of schistosity.

Immediately above and south of the waterfall the schists contain potassium feldspar which forms porphyroblasts 2–5 mm in size. Both size and frequency of the porphyroblasts increase towards the south, i.e. upwards in the sequence.

About 50 m north of a waterfall higher up in the brook, an exposure on the western bank shows the rock there to be a normal augengneiss with porphyroblasts of potassium feldspar, about the size of a walnut. Some metres north of the boundary towards the Särv nappe the porphyroblasts of the augengneiss reach their greatest size. There the longitudinal axes of the porphyryblasts can attain a length of 10 cm (Fig. 5).

Immediately south of the augengneiss follows the lowermost quartzite of the Särv nappe. This quartzite contains, nearest to its base, strongly folded thin layers or veins of a green amphibolitic rock. The primary structure of the rock has been destroyed by the strong deformation. The amphibolite resembles nevertheless a deformed, fine-grained, and ophitic Ottfjäll dolerite as this rock appears in marginal zones and in thin veins. The proximity of the larger sills of Ottfjäll dolerite favours the assumption that these folded bands of amphibolite are thin veins of Ottfjäll dolerite which, on account of their size and position, have undergone a stronger deformation than the rock in the steep standing thicker sills.

Dolomite-marble appears alternating with the quartzitic sandstone in repeated banks of the thickness of $\frac{1}{2}$–1 m. The dolomite-marbles exhibit thin phyllitic bands. Close to one of the banks of dolomite-marble a 3 cm thick claystone, traversed by fissures filled with calcite, has been observed.

The sills of Ottfjäll dolerite which appear in the sediments are 5–10 m wide and are orientated in close agreement with the *s*-planes of the quartzite which dip N70°W/80–60°N. One of the sills is situated closely below a bank of dolomite-marble. At the contact the dolomite-marble is gently folded, and the dolerite seems to have intruded the folded beds conformably. In one place the sill of dolerite is interrupted near a marked fold in the dolomite-marble (Fig. 26), and continues beyond this fold on the other side of the marble bank. This relation shows that a fold, or at all events a flexure, must already have existed in the dolomite-marble prior to the intrusion of the dolerite. At present the dolomite-marble is here strongly Z-folded according to $B = \text{N}/20°$. At the contact between marble and dolerite a skarn zone of some talc-like material, some centimetres thick, has been formed.

In the sills the Ottfjäll dolerite shows slight traces of deformation, and exhibits the porphyritic texture characteristic of this rock, with phenocrysts of plagioclase in an ophitic fine-grained groundmass.

In the Särv nappe the earlier axes of folding seem to have been dislocated by overturning towards the south during the last deformation which has a *B*-axis in about N80°W (cf. diagrams of megascopic fabric in Fig. 13).

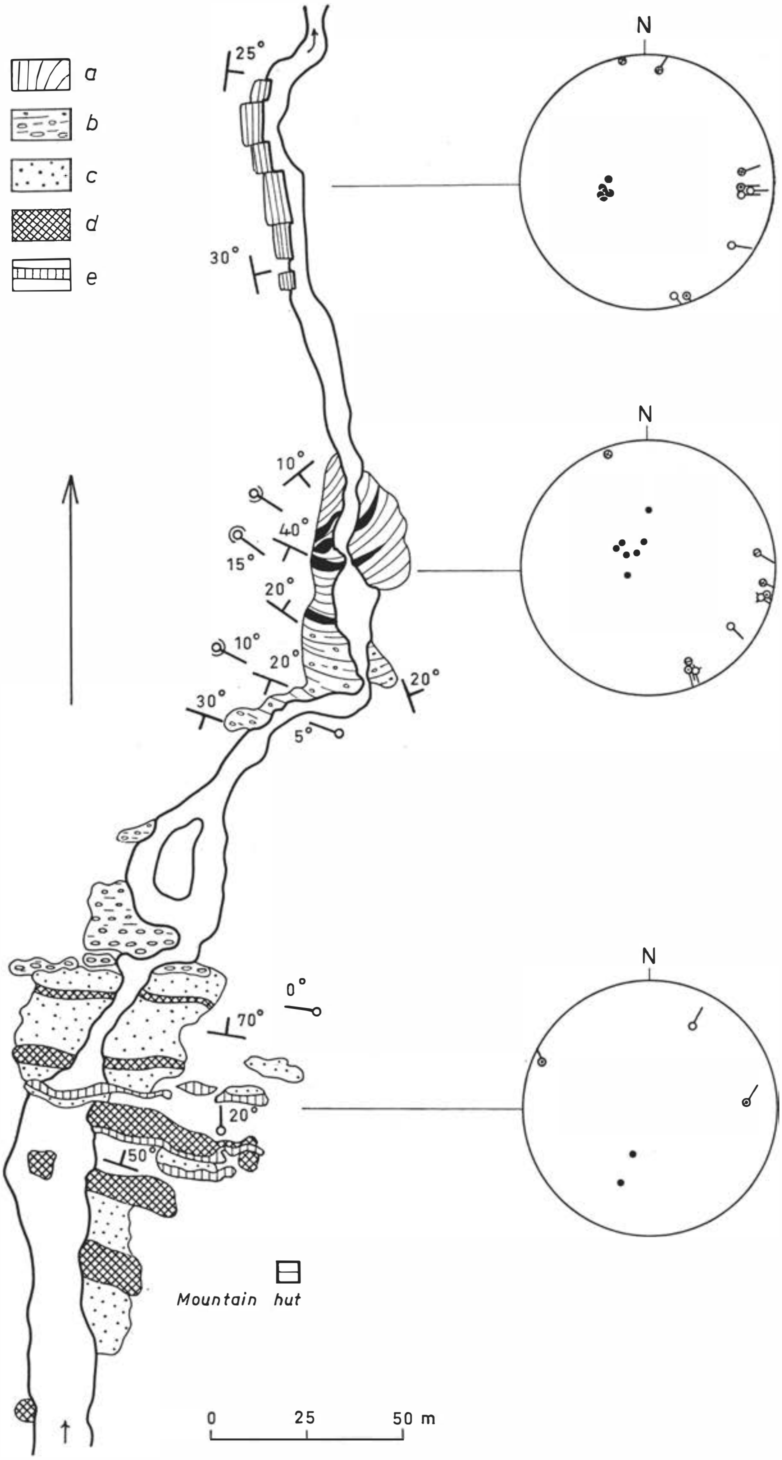

Fig. 13. The mylonite nappe and basal parts of the Särv nappe at the Riksgränsen anticline Bedrock and tectonics along R. Biskopsån, about 500 m SSW from the mouth of the river. *a* = mylonitic schist, *b* = augengneiss (the change in size and frequency of microcline porphyroblasts is indicated by differently sized and spaced markings), *c* = crystalline white quartzite, *d* = Ottfjäll dolerite, *e* = dolomite-marble, black areas = amphibolite. In the diagrams: circles = *B*, black dots = *s*-normals.

SEDIMENTARY ROCKS

The sedimentary rocks in the Särv nappe can be arranged in three groups, viz. (1) the basal complex (S.55), which in the following will be called the Ulvberg complex after its occurrence at Mt. Ulvberget, Midskogen (G4); (2) basal fragmental rocks, which, from their occurrence at Mt. Grönstacksberget (G3), are designated the Grönstack complex; and (3) the series of schists, greywackes, quartzitic sandstones, and quartzites of the Särv nappe proper.

The Ulvberg complex is the lowest of these groups from the tectonical point of view. It is often separated from the Särv nappe proper by a disturbance due to thrust, and therefore in the southern part of the nappe contains no dikes of dolerite. The Ulvberg complex, which consists mainly of white quartzite and dolomite, will be described later in more detail (p. 46).

The Grönstack complex

The fragmental rocks of the Grönstack complex occupy a special position among the sedimentary rocks of the Särv nappe proper.

At the top of Mt. Grönstacksberget and in the northern part of Mt. Tännäsberget (G3) there occurs a peculiar rock which has been described in the field as a chloritic greenschist with an abundance of quartz nodules. The rock occurs in intimate contact with the amphibolitic rocks of the Tännäs complex of which fragments, measuring about 1 m, appear apparently isolated in the rock. The outcrop rock gives the impression of being crinkled and of irregular structure. The abundance of quartz in concretion-like agglomerations is remarkable. The quartz is mainly white vein quartz that occasionally appears in the shape of deformed veins a few centimetres wide. Fragments of a dark quartzitic rock have also been observed.

Microscopical examinations have shown the rock at the top of Mt. Grönstacksberget to contain strongly deformed fragments of small, rounded nodules consisting of epidote and albite, greywacke-like sedimentary rocks and different eruptive rocks. It has thus to be interpreted as a fragmental rock of almost conglomeratic type.

A comparison with the conditions in Heidal, Norway, indicates that the Grönstack complex has been deposited upon a bedrock of the Jotun kindred in the same way as the basal conglomerate of the Heidal Series (STRAND, 1951, p. 56). GJELSVIK (1947, p. 7) describes "metamorphic anorthositic rocks" in an analogous position. The zone with "småöyegneiss" (porphyroblast schist) reported by GJELSVIK in the Heidal Series can indicate a zone of tectonic disturbance. On this account the overlying sediments can nevertheless not be compared with the sediments overlying the Grönstack complex in the Särv nappe. Probably the zone with "småöyegneiss" may indicate the overthrust plane of a higher tectonic unit, possibly corresponding the Helags nappe.

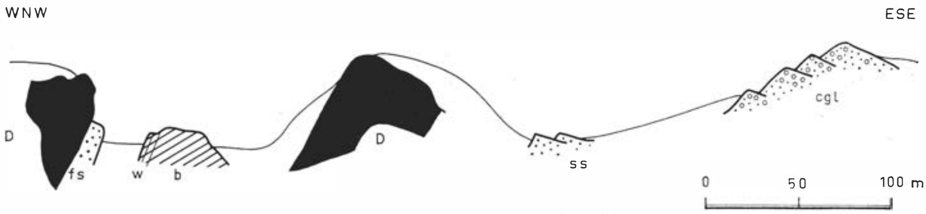

Fig. 14. The red breccia in the Särv nappe, L. Ränntjärnarna (G4), 7 km NW Midskogen Village. Vertical scale = 2 × horizontal scale. *b* = red breccia, *w* = reddish brown, apparently weathered surface of the breccia, *fs* = greyish fine-grained quartzitic feldspar sandstone, *ss* = schistose quartzitic sandstone, *cgl* = coarse quartzitic arkose with conglomeratic beds, *D* = Ottfjäll dolerite.

Immediately west of L. Ränntjärnarna (G4), about 6 km NNW of Midskogen village there is an exposure of a reddish breccia. The outcrop covers only a few hundred square metres, the rock strikes roughly WSW, and shows indications of a folding axis which dips 30° in the same direction. The breccia chiefly consists of angular fragments of a red felsitic rock measuring 0.1–20 cm and containing almost exclusively subhedral quartz and microcline, which are arranged in alternating layers. By this laminated distribution of quartz and microcline the rock thus exhibits S_i in the coarser fragments. The fragments are externally rotated (Pl. V b).

A petrographical investigation has shown that the coarse material in the breccia is strictly uniform, and consists almost exclusively of the reddish felsitic rock, while the more fine-grained material in the matrix is polymict and shows a high percentage of fragments of basic eruptives. Especially the groundmass with the small fragments has undergone a postdeformative recrystallization, and exhibits an exceedingly great abundance of epidote.

The interpretation of the whole sediment as a fragmental rock lies most closely at hand, but its mode of formation can nevertheless not be determined with any certainty. Several possible interpretations can be considered, viz. volcanic breccia, tectonic breccia, or residual sedimentary breccia. The following is a brief discussion of these possibilities.

Volcanic agglomerate. The fine-grained material, rich in iron and lime, seems to consist mainly of fragments of eruptive rocks. This observation supports the above interpretation.

Tectonic breccia. This interpretation is not directly inconsistent with the tectonic conditions. However, a tectonic brecciation must in this case have taken place before the deposition of the overlying sedimentary rocks, since the predominant material in the breccia consists of types of rocks entirely foreign to the sedimentary rocks of the Särv nappe. Moreover, the breccia seems to have suffered a stronger regional metamorphism than the rocks on top of it.

Sedimentary breccia. The coarser fragments are all of the same type and must be derived from a homogeneous rock mass. In this case, this material could be locally detached weathering material, between which detritus rich in lime and iron, and probably derived from basic eruptives, has been deposited by sedimentation. During the sedimentary redeposition of a primary residual breccia, an external rotation of the fragments can have taken place.

The floor of the breccia is not exposed, and it seems to be situated in the core of an anticline with a NNE strike (Fig. 14). In the western part of the cliff consisting of the breccia the rock assumes a red colour. This effect increases towards a sharp limiting surface, which seems to be a contact of weathering against the more fine-grained feldspar sandstone in the immediate neighbourhood. Finds of glacial boulders on Mt. Tramsberget (F4), about 1600 m north of the locality just described, suggest that the breccia might have a certain extension in a northerly direction. This is in agreement with the NNE strike of the anticline in which the observed exposure at L. Ränntjärnarna is situated.

The interpretation of the red breccia as a residual breccia is supported by a comparison with a similar fragmental rock described by G. HOLMSEN (1937), who accounts for an obviously residual breccia consisting of Archaean eruptives observed at Mt. Lillesjöberget, west of Elgådalen in Norway. According to HOLMSEN this breccia passes upwards into a coarse and more polymict conglomerate. This has been examined by N. ZENZÉN (1930) who found that it contained pebbles of fine-grained granitic rocks. As pointed out by ZENZÉN (p. 576), a sharp boundary seems to exist between the conglomerate and the overlying sparagmite, although these two rock members jointly have been subjected to folding and deformation.

At all the Norwegian occurrences of breccia-like or conglomeratic basal formations, these obviously occupy an overthrust position. For this reason their point of origin must be assumed to have been located considerably farther to the west.

Both the red breccia from L. Ränntjärnarna and the conglomeratic chlorite-schist from Mt. Grönstacksberget exhibit a strong recrystallization of most rock fragments, and especially an exceptional abundance of epidote. The conglomeratic chlorite-schist seems to have been deposited directly upon the amphibolitic rocks of the Tännäs complex, while the red breccia ought to be a residual breccia upon a bedrock consisting for the greatest part of a felsitic rock, rich in potassium. Both rocks obviously are basal fragmental rocks containing preponderantly local material derived from the pre-Cambrian rocks of the bedrock.

Among the rocks of the Tännäs complex a streaky felsite, rich in potassium, has been found which resembles the coarse fragments in the red breccia (Pl. XI a).

The overlying sedimentary rocks are unlike the Grönstack rocks—in which I include also the red breccia—both with regard to metamorphism and the sedimentary material. It appears possible that a considerable stratigraphic break exists between the rocks of Mt. Grönstacksberget and the younger sedimentary series of the Särv nappe.

THE RED BRECCIA.—More than 80 % of the rock consist of angular fragments, up to 20 cm long, of a fine-laminated, reddish potassium-rich felsite (Figs. 14, 15). Occasionally these large fragments are so closely packed that adjacent fragments are separated from each other only by a millimetre-thick layer of a material as fine as clay. As a rule, however, the coarser fragments are separated by a layer more than 10 mm thick of a greyish green matrix containing angular fragments of different rocks measuring 1–20 mm. Except in the coarse fragments this matrix is completely recrystallized so that the original margins of the fragments have become diffuse. The crystalline matrix contains epidote in the form of usually elongated prismatic crystals about 0.01 mm long, and quartz grains about 0.03 mm in size. Microcline, chlorite, limonite, and ore minerals occur in subordinate quantities.

The breccia thus contains components of three orders of size:

(1) 20–200 mm. Monomict. Coarse angular fragments of a reddish fine-laminated potassium-rich felsite.

(2) 0.1–20 mm. Polymict. Medium-sized fragments of different rocks embedded in (3) (Pl. V b).

(3) < 0.1 mm. A greyish green recrystallized matrix with a primary grain-size of probably 0.01 mm.

The fragments in the polymict intermediate fraction seem to exhibit an effect of secondary crushing by which numerous small fragments of a certain type of rock have been produced within limited portions of the breccia. Also in the intermediate fraction the most common elements are fragments of the red felsite, which are completely predominant in the coarsest fraction. The way in which the material of the finest fraction has penetrated into cracks in the larger fragments suggests that this material might have been a clay-like powdery substance.

The degree of the polymict composition of the material obviously increases with decreasing size of the fragments. While the coarse fragments are exclusively pieces of red felsite, some other rocks begin to appear among the fragments smaller than 20 mm. The coarsest fragment of a foreign rock that has been observed is a 15 mm long piece of a red felsite rock, in which the potassium feldspar contains inclusions of hematite. In some fragments of the red felsite chalcedone has been observed. Some fragments of greenschist-like rocks are 10 mm long, but most of the fragments of foreign rocks measure only 1–5 mm in diameter. This size group accounts for about 50 % of reddish felsite, while the remaining 50 % are made up of other rocks (cf. Pl. V b).

The mineral composition of some fragments from the intermediate fraction has been determined by planimetrical analysis. The figures within brackets represent the values obtained by determinations on different fragments of the same rock type in one thin section Pl. V b.

1. *Fine-laminated felsite rich in potassium.* The lamination of the rock which determines S_i in the externally rotated fragments is expressed by the bandlike

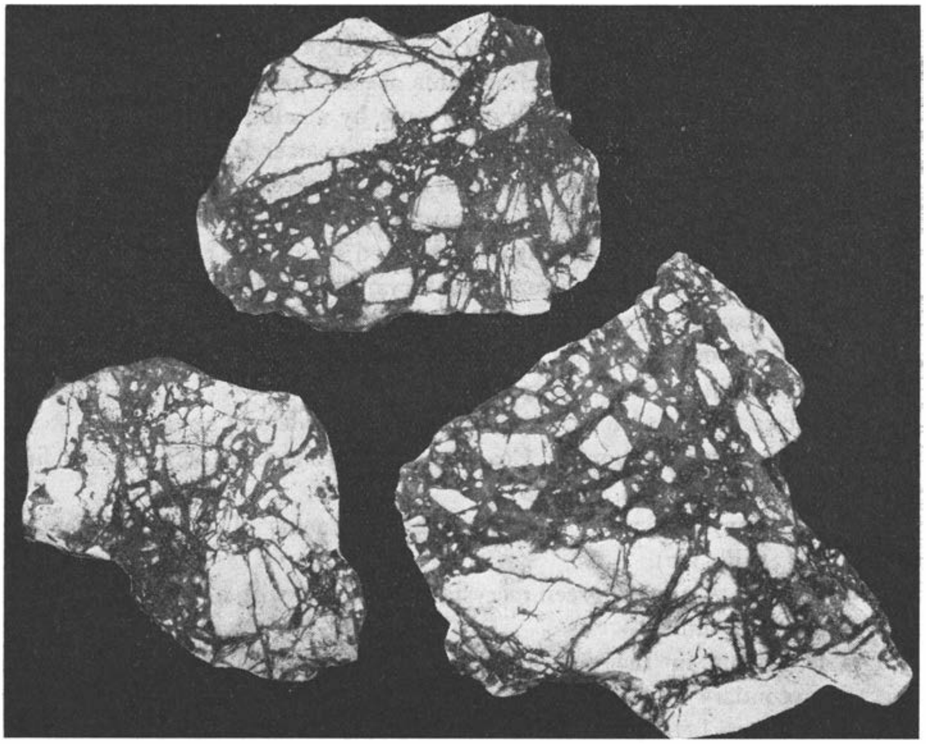

Fig. 15. The breccia from L. Ränntjärnarna, cf. Fig. 14. The white areas are fragments of streaky reddish felsite embedded in a matrix of greywacke-like composition. Polished surfaces of the rock were etched with conc. hydrofluoric acid; after washing treatment with 10 % sodium hydroxide, when residual iron became precipitated as brown hydroxide in the areas of iron-rich matrix. Half natural size. Photo: G. Andersson.

arrangement of quartz and feldspar. The distance between two bands of this kind is about 0.05 mm. This is the same rock which occurs in the coarse fragments of the breccia.

Composition: quartz 47 ($\pm$ 4) %, orthoclase 40 ($\pm$ 4) %, opal 5 %, epidote 6 ($\pm$ 3) %, sericite 2 ($\pm$ 1) %, and others. (Mean values obtained from determinations on 9 differently orientated fragments.)

2. *Felsitic metamorphite.* Composition: quartz 25 % (25.6 and 25.15), potassium feldspar 25 % (27.3 and 24.5), epidote 50 % (47.1 and 50.3). (Mean values obtained from two fragments.)

3. *Metadolerite.* Composition: quartz 5 % (5.91 and 1.81), plagioclase 50 % (51.8 and 50.1), diopside 30 % (27.1 and 32.17), epidote 13 % (12.6 and 13.7), ore minerals 2 % (2.6 and 1.6), and others. (Mean value obtained from two fragments.)

4. *Epidosite.* Composition: epidote 88 % (89.0, 85.0 and 91.7), quartz 6 % (7.0, 6.2, and 6.1), serpentine 5 % (3.22, 8.4 and 2.3), limonite (with some other

ore minerals) 1 % (0.8, 0.5, and 0.4). (Mean value obtained from three fragments.)

5. *Schistose fragments* apparently consisting for the greatest part of epidote.

6. Isolated fragments of different types of *metadolerite*, consisting sometimes preponderantly of feldspar and epidote, sometimes like (3), but with different percentages (cf. similar fragments in the conglomerate of Mt. Grönstacksberget, p. 36).

7. Isolated fragments of other types have been observed, e.g. of a relatively coarsely crystalline metamorphite, the composition of which ought closely to approach that of (2), and a fragment of pure quartzite.

Three types of fissures appear in the red breccia. The oldest group are *brecciating fissures*. These appear independently in every fragment, and seem to be connected with pressure and movement, which have influenced the sediment at an early stage. Occasionally it can be observed that after such a partial brecciation the parts of a broken fragment have been dragged apart in a certain direction, and that simultaneously material of the matrix has plastically entered the widened fissures.

The *fissures filled with quartz* have been formed after the consolidation of the rock. They often traverse portions of the rock filled with matrix.

Open joints have finally been formed at a late date. They traverse the rock irregularly with or without relation to older inhomogeneities.

Elongated or flat fragments are in general arranged with their greatest dimension in a certain plane. There exist, however, exceptions to this rule, e.g. when small pieces of schist lie conformably to the surfaces of adjacent larger fragments. In this case the orientation can differ considerably from the dominant *s*-plane.

Quartzite and greywacke sequences

The sedimentary rocks in the Särv nappe proper form two great series. One of them, which for this reason ought to be called the quartzite series, is dominated by quartzites and arkose-like rocks. The other series contains a great quantity of dark, fine-grained siltstones which according to current Swedish usage are called greywackes. On this account this latter series is called the greywacke or siltstone series. In the southern part of the nappe the quartzite series is almost the only one present.

In the northern part of the nappe both series occur side by side, even if the greatest part of the surface is occupied there by the greywacke series. The stratigraphic age relation of the series has not been worked out in a unambiguous way so far. In the east the greywacke series is underlain by the quartzite series, while farther to the west, at Ljungdalen (D4), the quartzite series appears west of and on top of the greywacke series. The quartzite series is probably older than the greywacke series. Weak and apparently very early tectonic

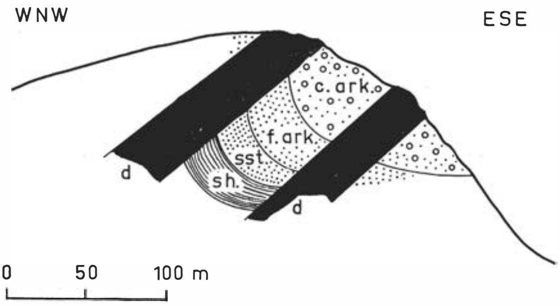

Fig. 16. The rocks in the southern face of Mt. Orrstädjan (F6). Vertical scale = 2 × horizontal scale. *sh* = banded shale, *sst* = dark arkosic sandstone with intercalations of clayey material, *f.ark* = fine-grained arkose, *c.ark* = coarse arkose with conglomeratic beds, *d* = Ottfjäll dolerite.

disturbances occasionally occur at the contacts between the series as is shown by observations from Mt. Tossåsenfjället (C5), where irregular narrow dikes of dolerite occur in folded quartzitic schists.

The *greywacke series* contains fine-grained impure sandstones, greywackes, or siltstones. These rocks are, however, often more or less recrystallized.

As an example of the rocks in this series a partly recrystallized dark-grey rock is described. Its composition, determined by planimetric analysis of a thin section perpendicular to the plane of sedimentation is: quartz 27 %, feldspar 22 %, epidote, ore minerals, etc. 12 %, intergranular matter recrystallized to biotite, muscovite, and chlorite 39 %. The average grain size is 0.07 mm. Owing to the fact that coarser rock fragments are missing the rock ought to be labelled siltstone, in spite of the circumstance that the high content of detritus produces a resemblance with a greywacke. Locally agglomerations of magnetite occur in portions of rock measuring about 1 cm. The composition of such a portion rich in magnetite is as follows: quartz 21 %, magnetite 35 %, feldspar 25 %, epidote 4 %, and intergranular matter 15 %.

In the most south-easterly part of the Särv nappe the *quartzite series* contains the following rocks:

Light-coloured, feldspar-bearing quartzites	ca. 150 m
Light-coloured, coarse, feldspathic quartzite with gravel conglomeratic beds, containing pebbles of quartz, quartzite, and microcline	70 m
Greyish arkose-sandstone	50 m
Dark-coloured arkose with clay laminae	20 m
Grey, banded shale with beds of limestone and bluish grey quartzite	10 m
Total thickness	ca. 300 m

The given thicknesses are estimated mean values. There occur great local variations.

The series is well exposed in Orrstädjan, one of the southernmost peaks in Mt. Särvfjället. In the southern slope of the mountain occur grey banded shales and sandstones and arkoses (Fig. 16). In this locality the quartzitic member of the series is richer in feldspar than is usual. The light arkose-like quartzite contains coarse grains of white or pink feldspar and light-blue quartz. It is these mineral grains which are especially characteristic for the light-coloured quartzites. Also the light, arkose-like quartzite contains pebbles of white or grey quartzite. The dark sandstone on top of the shale contains interspersed grains of dark quartz, and exhibits laminae and galls of a greyish black, clayey material. The banded shale contains light layers of silty material with cross-bedding and graded bedding. According to these sedimentary structures, the sediments ought to rest in normal position. All sedimentary rocks are folded with great amplitudes, and exhibit an average dip towards ESE. This packet of sedimentary rocks is traversed by two dikes of Ottfjäll dolerite with a westerly dip.

The same rocks or isolated members of the series are represented to somewhat different extent in different parts of the southern Särv nappe. The lowermost members of the series occur mostly at the southern and eastern boundary of the nappe, as at Mt. Rönningsåsen (F5), Mt. Svartuggen and Mt. Bankhammaren (F6), and at Gravarvallen seter (G4). At Rönningsåsen I have observed centimetre-thick bands of magnetite in light-coloured feldspathic quartzite. These bands of ore appear to be concordant with the bedding plane of the quartzite. Higher up in the nappe light, medium-grained (0.5–2 mm) feldspathic quartzites are widespread, and seem to be the only member occurring in the western part of the southern Särv nappe. In these quartzites the amount of feldspar and the degree of metamorphism are variable. On the top of Mt. Anåfjället, for instance, the sedimentary rock is a pale-blue quartzite with an inconsiderable amount of feldspar (cf. S.55, Fig. 18). These rocks always contain interspersed grains of pale-blue quartz, even if the content of feldspar is somewhat higher than in the pure quartzites. In addition, isolated pebbles of quartzite and of coarsely crystalline pink feldspar are occasionally found. Thus the similarities with the light-coloured quartzites of the series are obvious. Owing to the fact that the light-coloured conglomeratic or arkosic quartzite seems to be especially characteristic of the quartzite series, this rock will be dealt with in greater detail.

About 300 m east of the locality of the red breccia that has been described in connection with the Grönstack complex appears a coarse, light feldspathic quartzite containing gravel-conglomeratic layers with pebbles of quartzite (Fig. 14). The area between the breccia and the conglomerate contains at least one dike of Ottfjäll dolerite. The conglomeratic layers of the arkose indicate a plane of sedimentation that dips about 35° east. With regard to this dip and the space occupied by the intervening dolerite the conglomerate can be estimated to lie at a stratigraphic level about 50 m higher than that of the red residual

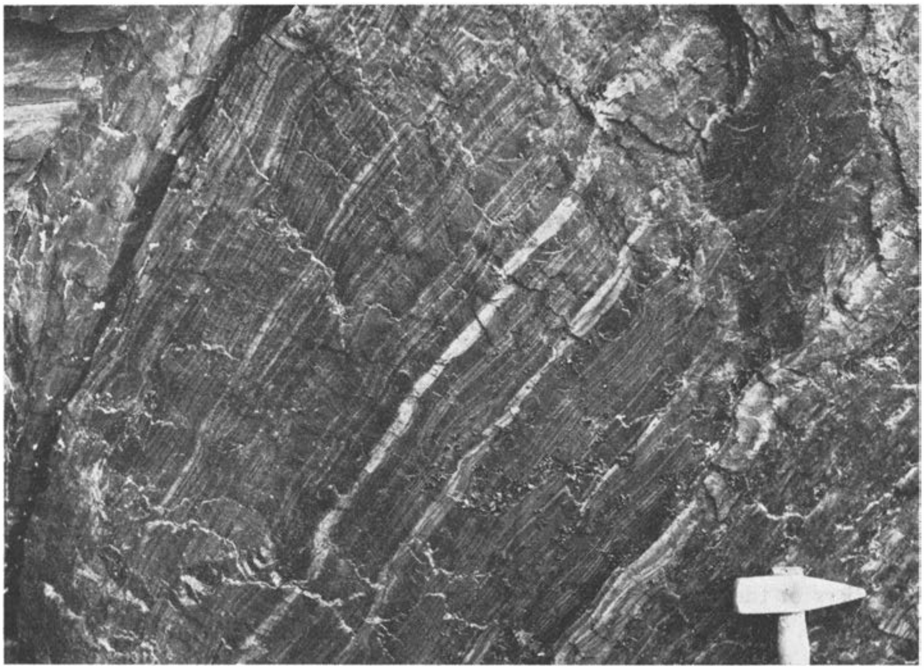

Fig. 17. Banded shale from Mt. Röstberget (E3) cf. Fig. 19. Flow casts and irregular lamination indicate variable conditions during the sedimentation. Some layers exhibit micro-lamination where the graded bedding proves that the sequence is inverted, the base of the layers being on their left-hand side (= NE).

breccia, provided that no unknown disturbances exist. Between these two rock members occurs a fine-grained, somewhat schistose quartzitic sandstone.

The coarse quartzite exhibits the characteristic feature, with interspersed grains of pink alkali feldspar and bluish quartz, and occasionally isolated larger grains of greyish blue feldspar. The poor sorting of the rock might indicate that its clastic material has undergone only slight transport, and perhaps is derived from a single rock.

In the region of Vigelen and the Archaean windows south-west of it as far down as the window of Brydalen occurs a granite of a very characteristic type with bluish quartz, red alkali feldspar, and often greenish (saussuritized) plagioclase. This granite was called "granit trikolor" by Kjerulf (Holtedahl, 1953). Possibly it was just a granite of this type, which seems to be widespread in a westerly direction, that has supplied the detritus material for the formation of the light-coloured quartzite.

The coarse feldspathic quartzite.—The rock consists of well-rounded grains, 1–5 mm in size, mostly of quartz and feldspar. The grain-size varies in different beds of the sequence. Also dispersed large pebbles occur, occasionally giving to the rock the appearance of a gravel conglomerate (Fig. 14).

According to a planimetric analysis of one thin section of the coarse arkose-sandstone from Gråvålen (G4) the rock contains: Quartz ca. 70 %, microcline and microcline-perthite ca. 18 %, oligoclase ca. 7 %, ore minerals ca. 1 %, and sericitic matrix ca. 4 %. In the thin section examined, the matrix appears especially in angular spaces between the rounded grains. The quantity and perhaps also the composition of the matrix are variable. A macroscopical examination sometimes creates the impression that the matrix of the rock contains saussuritized plagioclase.

The content of quartz is so high that the rock consequently should be called sandstone. As will be seen below there is sometimes less quartz.

A planimetric analysis of a similar arkose from Rönningsåsen (F5) in one section gave the following result: Quartz ca. 57 %, microcline and microcline-perthite ca. 17 %, oligoclase ca. 11 %, saussuritized plagioclase ca. 11 %, and sericitic matrix ca. 4 %. The content of plagioclase is thus higher in this sandstone than in that from Gråvålen. Isolated grains of oligoclase contain an abundance of Fe_2O_3 which gives the mineral a pink colour. Certain scattered grains of quartz contain fine needles of sillimanite. This sillimanite may be responsible for the light blue colour that is often observed in the clastic quartz of this sandstone.

Especially at certain levels the arkose contains scattered pebbles of quartzite. An examination in the field of the material in these pebbles showed that the majority of them consists of light-coloured quartzites, while isolated smaller monomineral pebbles consist of grey or white feldspar. This interpretation of the material in the scattered conglomeratic pebbles applies to all the localities in the eastern part of the Särv nappe, in which I have observed the coarse arkose.

Of 44 pebbles collected from Mt. Gråvålen three proved to consist of feldspar, while 41 were of quartzite or vein quartz. Calculated in per cent this material of pebbles with the size 1–5 cm consists of:

Feldspar (microcline or perthite) ca. 7 %;
Quartz from veins or nodules ca. 30 %;
Clastic quartzites (white or light grey) ca. 54 %;
Metamorphic quartzites (with recrystallization or crystalloblastese) ca. 9 %.

In a thin section a rounded fragment of feldspar porphyry measuring 2 mm has also been observed.

The pebbles in the conglomerate are rarely well-rounded, but often flat and elongated, and sometimes exhibit obliquely orientated flat surfaces together with well rounded edges. Since certain effects of recrystallization, especially in the finer clastic material, distinctly indicate a deformation of the rock, it is possible that the pebbles of the conglomerate have also been subject to some tectonic influence. The shape of the pebbles seems nevertheless largely conditioned by relict clastic conditions (cf. TURNER, 1950 p. 151). A rather small number of observations at Gråvålen seems to indicate that the longitudinal axis of the pebbles is roughly orientated in NE/10°.

The Ulvberg complex

In the thrust plane between the augengneiss and the Särv nappe appears a series of metamorphic sedimentary rocks. It is characteristic of the position of these rocks that they have never been observed in the Särv nappe proper. If this series had any direct connection with the other sedimentary rocks of the nappe, it could be expected that it should have been folded up among them, since the rocks of the Särv nappe often exhibit steep dips, and consequently must be either steeply folded-up or overthrust in an imbricate manner. In the majority of cases, however, the steep dips in the nappe depend upon overturns *en bloc* towards the east, with translations in local zones of disturbance.

In the basal series, called the Ulvberg complex from one of the most typical occurrences, no dikes of dolerite have been observed in the southern part of the Särv nappe. The rocks in the southern face of Mt. Ulvberget at Midskogen village (G4) have been described on a previous occasion (S. 55).

In addition to the localities described in my earlier paper (S.55, pp. 230–231) rocks belonging to the Ulvberg complex occur in the region around Funäsdalen (F2–3) and in the area east of L. Storsjö. Within the last-named area the Ulvberg complex has been encountered at Aborrtjärnvallen seter (D5), at R. Ekorrbäcken (C–D6), at Mt. Stensjöberget (D6), and at Stora Lövdalen (B6). In the western limb of the Särv nappe rocks belonging to the Ulvberg complex have been observed at R. Biskopsån, where they contain sills of Ottfjäll dolerite (cf. p. 34). The series seems to be most completely developed in the region Funäsdalen–Midskogen village (F2–G4).

The Ulvberg complex contains the following rocks:

Greyish white feldspathic quartzite	ca. 20 m
White dolomitic quartzite	10 m
Yellowish white dolomite-marble	5 m
White feldspathic quartzite	50 m

The thickness of the complex decreases towards the west and the east, the dolomite-marble and the white quartzite being the most persistent members. The decreasing thickness of the series probably depends upon the circumstance that it has for the greatest part been deposited in a confined basin. Locally the wedging-out of the series might also find an explanation by the assumption that the sedimentary rocks have either been squeezed out on account of their position in the thrust plane or have become excessively schistose. East of Lunnäset the Ulvberg complex is altogether missing. Thus at the eastern border of the Särv nappe the sparagmite schists are in direct contact with the augengneiss.

In many localities the mode of occurrence of the complex bears witness to very intense tectonic disturbances. Small occurrences of dolomite-marble and white quartzite are thus found west of Tännäs (G3). These are either wedged

Fig. 18. Dolomitic sandstone of the Ulvberg complex cut by and over-ridden by a body of Ottfjäll dolerite. Outcrop near Åbergstjärnvallen Seter on the old Tossåsen Road (D5). The lower right corner of the photo shows a probably postintrusive lineation in the dolomitic sandstone. This lineation dips NW/45° and roughly conforms with the surface of the dolerite body.

into or close to the augengneiss, e.g. in the southern slope of Mt. Furuberget and in a region about 2.5 km north-west of Tännäs church. In the steeply raised blastomylonite of the mylonite nappe occur about 1 dm thick banks of white dolomite-marble about 1 km south of the western end of L. Lossnen.

No fossils have been encountered in the rocks of the Ulvberg complex which are always strongly deformed and crystalline.

The Ulvberg complex can be related to the surrounding rocks in either of two ways:

1. The sedimentary rocks of the Ulvberg complex are linked stratigraphically with the other sedimentary rocks of the Särv nappe. In this case the mutual position of the rocks would suggest that the rocks of the Ulvberg complex are older than the rest.

2. The Ulvberg complex and the other sedimentary rocks of the Särv nappe do not belong to the same stratigraphical series, and are separated by an overthrust. Apart from this the mutual relation in time is altogether undefined.

It might be counted in favour of the first alternative that the second alternative postulates a long overthrust of the Särv nappe proper in relation to the Ulvberg complex. The phenomena of schistosity, folding, and metamorphism which have been observed in the zone of contact between the Ulvberg complex and the Särv nappe proper at Mt. Ulvberget, Mt. Röstberget, Mt. Tännäsberget, and in other localities are nevertheless indicative of considerable translative movements in this zone.

The second alternative is supported by several observations. The rocks of the Ulvberg complex have never been observed in the nappe, as already pointed out. In places where the Särv nappe contains pre-Cambrian rocks or basal formations, the slates and quartzites of the nappe are deposited directly upon this basement of Archaean rocks or upon basal formations of the Grönstack type. At Mt. Kröksfjällen the rocks of the Tännäs complex, which there form the bottom of the Särv nappe proper, are thrust on top of the Ulvberg complex. Also the following observations at Mt. Röstberget bear witness to an overthrust above the Ulvberg complex.

On account of the conditions mentioned, the Ulvberg complex has obviously to be interpreted as a distinct stratigraphic series, separated from the Särv nappe proper by an overthrust of considerable order of magnitude.

THE BASAL ZONE AT MT. RÖSTBERGET.—The bedrock of Mt. Röstberget (F3) has been examined earlier by A. E. TÖRNEBOHM (1873), who has given a section both of this mountain and of Mt. Funäsdalsberget. According to it, a steeply raised dike of dolerite in Mt. Röstberget traverses a sedimentary series consisting of clayey shales, limestone, and coarse clayey shales. G. FRÖDIN (1922 a) objects to this by stating that he had observed in Mt. Röstberget a thick sill which, on account of the covering masses of soil, could have been interpreted as a steeply dipping dike by TÖRNEBOHM.

In 1956 I observed granite upon the top of Mt. Röstberget close to the contact of massive dolerite, and came to the conclusion that a detailed mapping of the region was desirable. In the course of this mapping, which was carried out on the basis of an enlarged aerial photograph to the scale 1:3000 (Fig. 19), broad dikes of Ottfjäll dolerite were found to strike with a slight western dip in a northerly direction over the top of the mountain. The westernmost of them, which ought to be the dike of dolerite mentioned already by TÖRNEBOHM and FRÖDIN, appears to be bent in the southern slope of the mountain, the lower part of the dike having lagged behind in the thrust movement. This dike rests with a partly undisturbed traversing contact against a dark coarse-grained banded shale, which exhibits only slight tectonic deformation. The banding is produced by an alternation of pale coarse, and dark, fine-grained layers 2–10 mm thick. The soft sediment seems to have suffered subaquatic sliding movements and deformations (Fig. 17), which made it difficult macroscopically to determine the up and down in the sedimentary bedding of the shale. Microscopic examination of three sections from orientated hand specimens with micro-lamination unmistakably showed that the base of the now steeply raised packet of shale faces north, since the microlayers exhibit a rather distinct

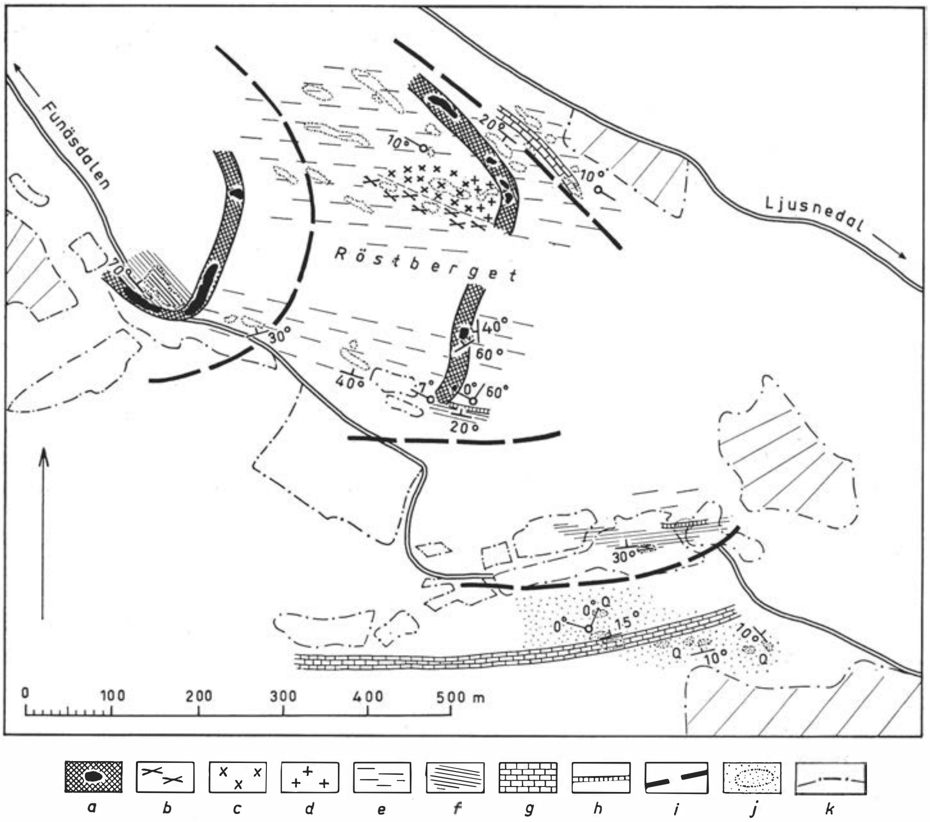

Fig. 19. Geological map of Mt. Röstberget (F2–3). *a* = Ottfjäll dolerite (outcrops black), *b* = granite mylonite, *c* = light-coloured microcline-granite, *d* = intermediary, red granite, *e* = greenish mylonite, *f* = banded shale, *g* = white dolomite-marble, *h* = grey dolomite-marble as lenses in the banded shale, *i* = thrust, *j* = outcrop (Q = white quartzite), *k* = field or clearing.

graded bedding. To the north-east, thus below the shale, occurs granite that is described below.

The banded shale and the dike of dolerite on Mt. Röstberget appear to be orientated so as to form an almost right angle with each other, and to be tilted together towards southeast. On this account I have attempted to rotate the plane of sedimentation of the schist into the horizontal plane.

The existing average dip of the raised shale is NW/70° E, and the present dip of the dolerite dike at the contact with the shale can be estimated to be N 10° E/ 20° W. Reconstructional rotation of these two planes around the axes N 60° E(14°) and N 30° W(110°) gives a theoretical primary dip of 0° for the shale and N 30° W/88° E for the dolerite dike. This result might indicate that, prior to the tilting, the rocks can roughly have had this mutual orientation. It is, however, possible that an external horizontal rotation of a large block in the nappe might also have contributed to the movement that has changed the strike of the dolerite dike.

It can be added that the rocks of the neighbouring Mt. Funäsdalsberget exhibit analogous changes of orientation. At the top, the *s*-planes dip about 50° towards the south, while their dips are almost vertical in the lower parts of the southern slope of the mountain. The dolerite dikes at the base of the nappe are at the same time dragged along, with the result that they exhibit a rather flat western dip.

In addition to an intermediary to mafic red granite the outcrops of granitic rocks in the north-eastern part of Mt. Röstberget contain a light-coloured microcline granite showing an obvious similarity to the granite from Nedalen valley in the Riksgränsen anticline (C1). The red granite which appears in the easternmost of the granite outcrops contains inclusions of fragments of finely crystalline amphibolite, and seems to be of rather varying structure and chemical composition. Occasionally this granite exhibits more felsic and, at the same time, coarser crystalline portions.

The nearest exposed dolerite is a relatively coarse uralite-dolerite, without porphyritic grains of feldspar. In this dolerite fissures 1 cm wide occur. They are filled with a substance rich in quartz which is deformed conformably to the boundaries of the fissure. This may be a filling of fissures with quartz and feldspar which is comparatively common in the disturbed basal zone of the Särv nappe.

In a downward direction the banded shale passes into a slate with grey dolomite-marble. Folded dolomite-marble and grey slate of this type occur in several outcrops in the south-eastern part of the mountain.

On the southern side of Mt. Röstberget many exposures of white quartzite are found. The white dolomitic marble, which usually occurs in it, discloses its existence by a row of glacial boulders of dolomite-marble along the southern side of the mountain, and occurs also in a 3 m high cliff about 100 m south of the largest potato-field. In this locality the dolomite-marble is yellowish white with relatively slight schistosity. The white quartzite, on the other hand, is rather markedly schistose and folded.

No dolerite dikes have been observed in the region of quartzite and dolomite-marble. This fact is of the greatest interest, since it provides an indication of the relation between the basal series of dolomite-marble and white quartzite on the one hand, and the sequence of the Särv nappe proper with grey shale containing marble, banded shale, arkoses and quartzites on the other. The basal complex at Mt. Ulvberget (G4), which I have described previously (S.55), is obviously not only an intrinsic element in the real sedimentary series of the Särv nappe. The nappe is separated from the basal complex by a thrust which is specially marked, since most often dolerite dikes are missing in the basal complex.

At Mt. Ulvberget (S.55, p. 227) this overthrust is located at the phyllonites and blastomylonites which overlie the white quartzite. Schistose and truncated, occasionally fragmentary, dikes of dolerite appear *above* this level, while below it dolerite is altogether absent.

Törnebohm (1873, p. 36), who described a conglomeratic schist below Mt. Funäsdalsberget (F2), has already pointed to the existence of a stratigraphic break between the basal zone and the sequence of the Särv nappe proper. The conglomeratic schist consists of a clayish matrix with isolated pebbles of quartzite and granite. The conglomeratic schist would thus be situated exactly in the thrust zone between the Särv nappe and its basal zone. In Funäsdalen (F2) I observed a schistose rock with fragments, including some of dolomite-marble, measuring about 1 m. In spite of the fact that in the field I interpreted this rock as a tectonic breccia it is nevertheless possible that it might be the counterpart to the conglomeratic schist mentioned by Törnebohm.

Holmsen & Oftedahl (1956, p. 96) also record an analogous observation from the Koppang field, viz. the probable existence of two different sedimentary facies. Of these one contains quartzitic sparagmite, dolomite-marble, and conglomeratic schist, while the other consists of dark schists and sparagmite.

In the general map (Pl. I) the big overthrust, for economy of space has been concentrated to the boundary between the basal zone and the augengneiss. This overthrust is, however, a complex feature being composed of several minor overthrusts, distributed not only between the Ulvberg complex and the mylonite nappe but also between the Särv nappe and the Ulvberg complex.

Carbonatic rocks

Carbonatic rocks occur both in the Ulvberg complex and in the sedimentary series of the Särv nappe proper. Owing to the fact that these rocks are situated in close proximity to the overthrust zone the dynamic metamorphism is strong, and makes a stratigraphic distinction between the different carbonatic rocks difficult. At the boundary between the Särv nappe and the Helags nappe I have also found several marbles south of Ljungdalen (D3). To compare the content of dolomite, these different carbonatic rocks have been chemically analysed for calcium and magnesium.

Metasomatism and deformation, e.g. flattening, folding, and brecciation often have produced an intense intermingling of these rocks with material from the surrounding silicate rocks. In this connection the carbonatic rocks have obtained very varying contents of quartz and silicate minerals. It must therefore be taken into account that the local content in the carbonatic rocks of elements other than calcium and magnesium has come to depend to a large extent upon a number of external factors, such as the degree of the mechanical and chemical addition of silica, etc., and for this reason is variable according to circumstances. This remark can be said to apply in the same way to the total content of calcium and magnesium. For this reason the chemical analyses of the carbonate rocks have been limited to the ratio between calcium and magnesium in the respective

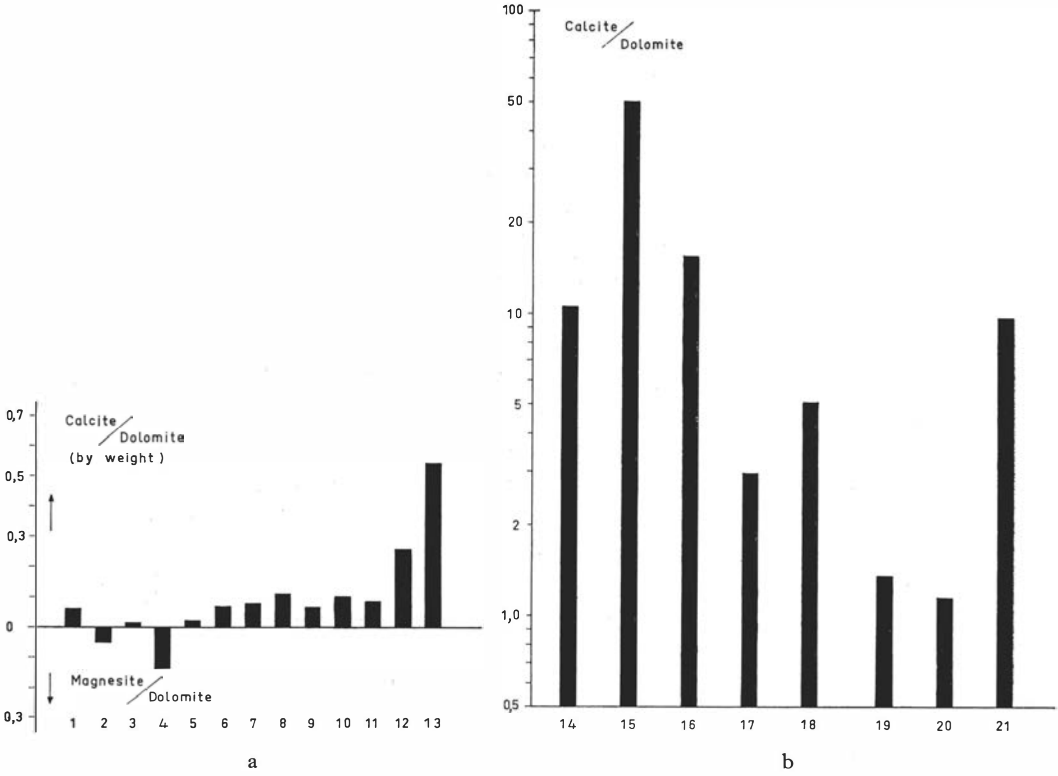

Fig. 20. Composition of the carbonates in some carbonatic rocks from south-western Jämtland county. Calculation made after chemical analyses on Ca and Mg. Analyst: B. Almquist, Min.-Geol. Inst., Uppsala.

(*a*) *Linear scale.* Calcite/dolomite or magnesite/dolomite ratios in dolomite-marbles. The localities are: ULVBERG COMPLEX: *1*, Vegalsberget I (G3), *2*, Vegalsberget II (G3), *3*, Valmåsen (G3), *4*, Åbergstjärnvallen (D5), *5*, Röstberget I (F3), *6*, Röstberget II (F3), *7*, Matsäckstjärn (D6), *8*, Biskopsån (C1), *9*, Funäsdalen (F2), *10*, Ulvberget (G4), *11*, Grundsjön (F5), *12*, NW Unsjön (F5), *13*, St. Lövdalen (B6).

(*b*) *Logarithmic scale.* Calcite/dolomite ratio in some marbles and meta-limestones. The localities are: EASTERN SÄRV NAPPE: *14*, Tottoåsen (F5), *15*, Fröedeberget (F5), *16*, Svartåsvallen (D7), *17*, Lunndörren (C5), *18*, Stensjöberget (D6); HELAGS NAPPE: *19*, S. Ljungdalsberget (D3), *20*, Uggvallen (D3), *21*, S. Ljungdalen (D3).

rocks. This ratio has been recalculated as the ratio calcite: dolomite or magnesite: dolomite, when there was a deficiency of calcium.

The determinations were restricted to carbonates which are soluble in 4 *n* HCl, and calculated for samples dried under ordinary room conditions.

Owing to the fact that in the rocks examined the ratio calcite: dolomite varies from 1 to 50, a logarithmic scale has been used in Fig. 20b to represent this ratio. In Fig. 20a, where the ratios calcite: dolomite and magnesite: dolomite are 0–0,5 and 0–0,2, respectively, these ratios have been represented on a linear scale.

From the illustrations it becomes evident that the carbonatic rocks can be classified according to the composition of the carbonates.

Thus, the white marbles of the Ulvberg complex are always dolomitic (Fig. 20 a). The phyllitic meta-limestones, which were found in connection with the banded shale near the bottom of the quartzite sequence in the eastern part of the Särv nappe, generally have a much higher calcite content (Fig. 20 b, 14–18). In this manner these meta-limestones show a certain conformity to the Hede limestone, in which the calcite: dolomite ratio was earlier shown to be about 3.0 (Ca: Mg = 5.77, STRÖMBERG, 1952).

Some calcite-marbles from the bottom of the Helags nappe are included in Fig. 20 b, cf. p. 61.

IGNEOUS ROCKS

The Ottfjäll dolerite

As far as the primary magmatic crystallinity of the rocks can be distinguished the Ottfjäll dolerite always exhibits a porphyritic texture. If in the observed dark varieties of the dolerite the usual xenoblasts of plagioclase cannot be distinguished either in the hand-specimen or in the outcrop, the explanation of this fact is to be found mainly in the occasionally very dark colour of the plagioclase. In spite of the occurrence of minor local varieties the rock is thus remarkably uniform with regard to content of feldspar and texture. Occasionally a minor part of a larger dike of dolerite is very rich in feldspar, but in this case mainly the size and the frequency of the porphyritic interspersed grains of basic plagioclase increase with the consequent transformation of the dolerite into a porphyritic anorthositic rock.

In the augite HOLMQUIST (1894, p. 188) had observed deformation and undulose extinction. This he attributes to inner tensions in the crystal structure independently of any external pressure. In the dolerite, however, undulose extinction of the plagioclase and other signs of deformation are frequently observed. With respect to the thorough tectonic engagement of the entire block of the nappe subsequent to the solidification of the dolerite, these effects of pressure must naturally be considered as of secondary nature and of being directly connected with the deformative movements.

In some localities, where several generations of dolerite have been intruded successively, the mechanism of the intrusion could be studied. At the highest peak of Mt. Håckelberget (F2–3) an almost vertical dike of dolerite, about 20 m wide, occurs. Here, a younger dike, some cm thick, of fine-grained dolerite has been intruded conformably along the eastern border of the former, which is more coarse-grained. The contacts between the rocks are blocky, a very common phenomenon in the Ottfjäll dolerite. Both dikes of dolerite contain interspersed porphyritic grains of basic plagioclase, and resemble each other very much, with the exception only of the coarseness of the texture.

Fig. 21. Gently dipping (NNW/20° SW) dikes of Ottfjäll dolerite intersecting a light and dark-banded quartzite (W/30° N). (Horizontal lineations on the overlying dolerite dike are glacial striae.) Outcrop near the base of the Särv nappe by the forest road to Falvik (D5).

On the southern slope at the southern end of Mt. Blåfjället (C4), I have observed in the Ottfjäll dolerite a less common solidification structure. A 2 cm thick slab of uralitized dolerite borders with intrusive contact on fine-grained sandstone. The recessed fissure structures which are represented on Pl. VIII a ought to have been formed during the solidification of this thin vein of dolerite that had been intruded primarily into the sandstone. Subsequently, an intruding larger mass of dolerite has filled the contraction cracks in the primarily intruded slice of dolerite, and formed a dike of dolerite with a width of about 10 m. The figured depressions have no counterparts in the sandstone next to the dolerite. The sandstone contact is formed by a surface of joint, being part of a system of pre-intrusive joints with the orientation N 30° W/80° W. The orientation of the plane of sedimentation is N 30° E/40° S.

As mentioned before, the boundaries of the Ottfjäll dolerite dikes always exhibit chilled margins. The contacts are blocky and matched. The inclusions are, like the wall rock, often altogether clastic without signs of tectonic influence (cf. S. 55, Fig. 18). At the very contact between the wall rock and the intrusive, apophysis-like fillings of microscopical cracks can occasionally be distinguished. Also microscopical zones of shattering are parallel with, and often coincide with, the surface of contact. Evidences of a distinct contact metamorphism have nowhere been observed in the wall rock.

Table 1. The chemical composition of some amphibolites and dolerites in the Trondheim area, Norway and its surroundings.

	I	II	III	IV	V	VI	VII
SiO_2	46.62	47.11	46.71	45.96	47.15	47.97	48.48
TiO_2	—	0.67	—	0.15	1.56	1,68	1,38
Al_2O_3	15.15	19.75	18.15	15.42	15.11	11.26	18.07
Fe_2O_3	—	2.30	1.23	3.75	2.84	9.09	3.08
FeO	12.85	4.59	8.39	6.81	9.31	5.46	7.51
MnO	0.59	0.08	—	—	—	0.87	0.19
MgO	9.84	7.73	10.43	10.98	7.73	3.95	3.8
CaO	11.81	11.67	11.24	11.72	11.53	11.76	11.91
Na_2O	2.60	2.80	2.35	2.55	2.51		3.64
						5.14	
K_2O	0.50	0.80	0.35	0.42	0.51		0.84
P_2O_5	—	0.10	0.02	0.02	0.19	—	0.15
S	—	0.02	—	—	—	—	—
H_2O-	—	0.07	0.08		0.07		0.80
				1.75		2.54	
H_2O+	—	1.72	1.55		1.46		—
Total %	99.96	99.41	100.50	99.57	99.97	99.72	99.89

I: Saussuritized gabbro, Hestekletten, Storvarts, Röros (according to TH. HIORTDAHL–FR. LINDEMANN [V. M. GOLDSCHMIDT, 1916, p. 16]).
II: Amphibolitic schist, Naversnes, Finnö, Stavanger region (according to V. M. GOLDSCHMIDT, 1916, p. 15).
III: Epidotic amphibolite from Reinskloster, Rissa (according to H. RAMBERG, 1943, p. 19).
IV: Uralitic gabbro from Selnes, Lensvik (H. RAMBERG, 1943, p. 19).
V: Amphibolite from Rødsjødalen (H. RAMBERG, 1943, p. 24).
VI: Ottfjäll dolerite from Mt. Ottfjället, Jämtland (P. J. HOLMQUIST, 1894).
VII: Ottfjäll dolerite from Mt. Ulvberget, Härjedalen (A. STRÖMBERG, 1955).

Analyses I and II give the composition of the Caledonian lavas, III–V are amphibolites from the Fossen area at Vestranden, while VI and VII are material from the dolerite dikes in the Särv nappe.

There is a distinct difference in the emplacement of the dikes of dolerite in the higher and the lower parts of the Särv nappe. While in its upper parts the series of dikes are steeply inclined and fairly regular, the dikes at the base of the nappe are thinner, have a flatter dip, and often a more irregular distribution (cf. Figs. 21 and 23). Nevertheless, also these dikes in general intersect the wall rock. Small veins, presumably of Ottfjäll dolerite, conformable with the *s*-surfaces, have been observed exclusively at R. Biskopsån (C1). In spite of the probability that the flat position of the dikes of dolerite near the base of the nappe is the result of postintrusive deformation in connection with the overthrust, as has been described at Mt. Röstberget, we have to reckon with a certain difference in deformation anisotropy between the higher and the lower parts of the nappe at the time of the intrusion.

Deformations of the Ottfjäll dolerite are especially common at the base of the nappe. In this connection the gradually more intense deformation has produced in the dolerite the following metamorphic series:

Ottfjäll dolerite—amphibolite with blasto-porphyritic texture—amphibolite with uniformly fine-grained texture, often with pencil structure—chloritic phyllonite.

Certain similarities exist between the green Ordovician effusives in the Trondheim area (GOLDSCHMIDT, 1916) and some of the amphibolitic rocks from Vestranden. In Table 1 I widen this comparison by adding two analyses of the Ottfjäll dolerite to the data given by H. RAMBERG (1943). The table gives a certain background to the discussion on the origin and age of the Ottfjäll dolerite even if the deviations of the various analyses are too considerable to allow the attribution of this simple comparison a decisive significance as regards the relations between the rocks.

The Sylarna amphibolite

In the western slopes of the Norwegian Mt. Sylarna (B1) we find at a height of some hundreds of metres above the floor of the valley of R. Fiskån an alternation of mica-schists and amphibolite, which is illustrated in the section Fig. 11. Higher up, the mica-schists give way to amphibolite which otherwise occupies the upper and eastern parts of the entire Mt. Sylarna massif.

From the southern side, the boundary between a brown coloured finely folded mica-schist and the overlying amphibolite can be observed in the south-easterly precipices of Mt. Storsola and Mt. Ekorrhammaren. Thus the amphibolite of Mt. Sylarna lies in the form of a cover more than 500 m thick on top of the mica-schists. The sheet of amphibolite is about 10 km wide, and exhibits a slight dip towards the east. Towards the south-east the amphibolitic cover can be traced as far south as Mt. Njåmeltjarve (C1). On the eastern side of this mountain the amphibolite abuts directly against a steeply raised brown mica-schist with a steep ESE dip and an equally steep easterly axis of folding in the contact surface. The dip of the *s*-planes is different for the two rocks, those of the schistose amphibolite dipping NW/30° SE, while the metamorphic schist shows *s*: N 30° W/20° NE and *B*: N 65° E/17°.

East of Njåmeletjarve amphibolite appears in the shape of 1–10 m broad dikes with a dip NW/40° NE. The wall rock of the dikes is a grey crystalline quartzite. In his paper of 1855 (p. 89) J. C. HØRBYE reports from the foot of Mt. Predikstolen, the most south-westerly peak of Helags, the find of an amphibolite, presumably of the same type as that of Mt. Sylarna. HØRBYE mentions that at this peak, apart from the amphibolite, he found a quartzite with some mica. He furthermore presumes that this rock might constitute the equivalent of the sandstone in Mt. Dörrfjällen (B5) in the east. HØRBYE evidently conjectured the direct tectonic connection, which I now have shown to exist between the rocks in the eastern and western occurrences of the Särv nappe.

In the area between Gröndalen and Mt. Blåhammarfjället the boundary between the Särv nappe and the Helags nappe is still little known. Along this distance I have been able to observe such a direct boundary in a few places only. In Mt. W. Helagsskaftet a row of boulders of a pegmatitic, white, trondhjemitic rock with garnets occurs north-east of the above-mentioned quartzite with amphibolite. Still further towards the north-east Åre gneiss makes its appearance.

The texture of the amphibolite in the massif of Mt. Sylarna often recalls, both in the outcrops and in hand-specimens, the texture of a somewhat schistose porphyritic Ottfjäll dolerite. The microscopical examination shows the pseudo-porphyritic grains to be filled with detrital minerals which often are lost in thin sections, but which sometimes contain chlorite. The rock consists of green hornblende, plagioclase (andesine), this mineral occurring in porphyritic aggregates with fine-grained epidote, chlorite (prochlorite), likewise in large aggregates, and brown titanite (cf. Pl. XI b).

A comparison with the rock in a dike of amphibolite from an environment, which almost certainly can be referred to the Särv nappe (the western slope of Mt. Blåhammaren, at a brook leading into R. Enån, [A1]), shows obvious similarities with the amphibolite from Mt. Sylarna. The dike rock thus contains green chlorite in fairly large aggregates, while the intervening mass consists of green hornblende, epidote, quartz, oligoclase, and some titanite.

Whether or not the amphibolite of Mt. Sylarna is entirely a Caledonian intrusive rock seems to be somewhat uncertain. Its tectonic position is near the overthrust zone between the Helags and Särv nappes, maybe partly replacing the latter.

The Tännäs complex

In his map of 1896, A. E. Törnebohm indicates only dikes of Ottfjäll dolerite in the augengneiss at Mt. Tännäsberget. In doing so, Törnebohm presumably interpreted the rock in Mt. Tännäsberget as a somewhat deformed, light-coloured Ottfjäll dolerite.

In their structure and composition the eruptive rocks in the massifs around Tännäs village (G3) exhibit certain points of agreement with the rocks of the Bergen-Jotun group. Thus an apparent agmatitic structure is very common, together with a great variation in the content of dark minerals. On account of this great variation one might be tempted to describe the rock as some kind of agmatitic breccia (Sederholm, 1934) with deformed fragments of different eruptive rocks which in this case all seem to contain potassium and may be consanguineous (leucocratic and melanocratic variants).

The tectonic influence upon the rocks of the eruptive massif can be divided into two distinctly distinguishable deformations. The younger of these deformations is of an obviously Caledonian type and appears in definite zones which are concentrated to the peripheral parts, while the intervening portions of the massifs have remained relatively undisturbed by this deformation. The Caledonian deformation has thus produced strong schistosity in the margins of the massifs, e.g. towards the augengneiss west of Tännäs (Pl. III *25–28* and Pl. IV 29–32).

The older deformation is of an entirely different and more plastic type. It has resulted in the brecciation and smearing out of already consolidated rocks. This deformation has been followed by recrystallization of the entire rock complex. The old deformation has produced steep lineations of undulating light and dark bands (Pl. V a). In the Tännäs complex the following rocks occur in the form of fragments: black pyroxenite, anorthosite, potassium-rich felsite, and some kinds of rock derivations in which appear different transitions into the brecciated rocks. A gradual change in the content of dark and light minerals makes the boundaries of the fragments in the breccias diffuse.

The eruptive rocks in Mt. Tännäsberget contain plagioclase and microcline perthite in delimited, pale areas. Between these lie dark bands with clinozoisite, chlorite, biotite, and garnet. Accessorily appear apatite, titanite, magnetite, rutile, and zircon. The clinozoisite and chlorite may have been formed from pyroxene by a retrograde metamorphism. The perthite contains spindles of plagioclase. In the western slopes of Mt. Kröksfjällen S. of Tännäs (G3) granitic rocks and felsite have been observed (Pl. XI a).

Between the anorthosite in Heidal and the complex of Tännäs some similarities are obvious. GJELSVIK (1947) has shown that in Heidal all rocks of the Bergen-Jotun group contain garnet which has often been altered into biotite. In the rocks of the Tännäs complex the garnets are frequently distributed in a mass rich in biotite. The fact that the small garnets are remarkably fresh and genetically progressive suggests the contribution of polymetamorphic factors, so that older garnets have been deformatively shattered and perhaps corroded, while the fragments, on a change of the pressure-temperature conditions, have again started to grow.

In Mt. Tännäsberget, as well as in Heidal, the micro-perthite occasionally occurs in aggregates measuring several centimetres. On these I have observed in the southern slope of Mt. Tännäsberget a vortex structure in the shape of spiral arms. Plagioclase and perthite are often assembled into macroscopically visible bands. Between these pale portions irregular areas, bands, or lenses of the dark minerals are situated. Analogous observations in Jotun rocks have been reported by GOLDSCHMIDT (1916), GJELSVIK (1947), and STRAND (1951). These effects have been interpreted as protoclastic streaks in a flowing, partially crystallized magma.

In 1947 GJELSVIK reported that the anorthosites in Heidal are tectonically disturbed at their contacts with the surrounding rock floor. The secondary structures which have been formed during the Caledonian metamorphism traverse, however, the primary structures with agmatitic breccias and streaks. GJELSVIK (*op. cit.*, p. 33) finds great similarities in structure and mineralogical composition between the rocks of the Heidal complex and the rocks in the anorthosite complex of the Adirondacks described by R. BALK (1931). Like this author, GJELSVIK interprets the schlieren structure of the anorthosites as formed protoclastically in a flowing, partly crystallized magma. BALK's interpretation of the structures has been opposed by BUDDINGTON (1933).

The present structure of the rocks of the Tännäs complex may be explained as the result of intense metamorphism of a series of eruptive rocks. This metamorphism can have included deformation in several phases and complete recrystallization, possibly with potassium metasomatism. In this connection, the late retrograde metamorphism is not taken into account.

In the easternmost part of the southern slope of Mt. Tännäsberget I have found a dense, dark dolerite which traverses the Tännäs rocks in the form of a disrupted dike about 1 m wide (Pl. V a). The microstructure of this dolerite, however, does not resemble that of the normal Ottfjäll dolerite, but has the appearance of a somewhat porphyritic amphibolite. Amphibolites of a similar kind have been observed at some other localities in the area investigated as, for instance in connection with the amphibolite of Sylarna as well as in narrow veins and marginal zones on dikes of Ottfjäll dolerite.

The occurrence of Ottfjäll dolerite in close connection with the rocks of the Tännäs complex is proved by the presence of this dolerite in the western portion of the southern slope of Mt. Tännäsberget. Since it is difficult to distinguish in the field between Caledonian deformed Ottfjäll dolerite and primarily deformed Tännäs gabbroid rocks, the contact relations could not be determined.

From the Sel-Vågå area pebbles in a basal conglomerate in the lower, Eocambrian part of the sedimentary Heidal series have been examined by T. STRAND (1951, p. 16). He considers this conglomerate to overlie the Rudihö complex, which belongs to the Bergen-Jotun group, and the Heidal anorthosite forming part of it. STRAND has identified some pebbles in the conglomerate as eruptive rocks from the underlying Rudihö complex. Accordingly, he refers the entire Rudihö complex to the pre-Cambrian.

For the Tännäs complex there are two possible explanations. These rocks can either form early intrusives into the basal zone of the Särv nappe, or they may represent the primary pre-Cambrian base upon which the sediments of this nappe have been deposited.

In Mt. Grönstacksberget (G3), at the northern part of Mt. Tännäsberget and at Fröstsjöberget (H3–4), a chlorite-schist with fragments of eruptive rocks and sedimentary rocks has been observed. This fragmental rock, which is described on p. 36, seems to have been deposited immediately upon rocks of the Tännäs complex.

With the further support of STRAND's observations in the Sel-Vågå area, I arrived at the conclusion that the anorthositic-gabbroidic rocks of the Tännäs complex would thus form the original basement of the Särv nappe which now crops out as a narrow zone at the margin of the nappe from S. Kröket to Mt. Grönstacksberget.

Beside this occurrence, the Tännäs complex has been observed along the boundary of the Särv nappe north of L. Östersjön (F2) and south of L. Malmagen (F1). In the northern part of the nappe potassium-rich rocks, which seem to belong to the Tännäs complex, occur near the bottom of the nappe 2 km E of L. Yttre Röversjön (C–D5) and at L. Brattjärnarna (C5).

In the area between the rivers Glommen and Rena, Norway, there is an easternmost part of the Lower Jotun nappe described by P. HOLMSEN and C. OFTEDAHL (1956). Augengneiss is there thrust over sparagmitic rocks, but is itself overlain by the Jotun rock complex, which contains gabbro, pyroxenite and anorthosite. Upon the Jotun rocks were in some places found green, greywacke-like sedimentary rocks, containing fragments of green lava rocks. Quartzitic rocks, which belong to the Kvitvola sequence were found upon the Jotun rocks, which effect by the above authors was presumed as being the result of an imbricate structure.

The Helags nappe

As pointed out in the introduction, I have placed particular emphasis on the investigation of the rocks of the Särv nappe. In consequence, other tectonic units have received a more summary treatment, and this applies especially to the rocks of the Helags nappe. For this reason a definitive stratigraphic distinction of the rocks of the Helags nappe is not possible at present, the observations being too widely spread. We shall thus content ourselves with a general survey of the rocks entering into the sequence of this nappe.

A great part of the Helags nappe consists of metamorphic, mainly postdeformatively recrystallized gneisses and gneiss-like schists which, in contrast to the crystalline rocks of the Särv nappe, are especially characterized by the appearance of calcite and biotite. In the rocks of the Helags nappe calcite seems to be of such common occurrence in the rocks that it could be assumed to have developed during the regional metamorphism in the Helags nappe. However, a great variation in the lime content, resulting in the occurrence of muscovite schists which are poor in lime, e.g. along R. Mittån in the upper Mittådalen valley, shows that the carbonatic matter mainly originates from calcareous metasandstones and calcite marbles which occur on various levels in the sequence.

Among the sedimentary rocks in the Helags nappe several metamorphic marbles are observed. One of these is found in Ljungdalen (D3) very close to the base of the nappe. Another marble is associated with a black slate south of this locality at the highway near Uggvallen (D3–4). Analyses of these marbles give a weight ratio calcite:dolomite of almost 10 for the marble from Ljungdalen (Fig. 20b, No. 21), and of ca. 1 for two samples of the marble at Uggvallen (Fig. 20 b, Nos. 19–20).

At Uggvallen the marble is traversed by a dolerite dike and ought thus to be most closely connected with the Särv nappe, while the marble from Ljungdalen is covered and underlain by calcareous crystalline schists which, like the marble, doubtless belong to the Helags nappe.

The stratigraphical series near the base of the Helags nappe, in which the marble at Ljungdalen lies, can be studied at the brook from Mt. Torkelsstöten in the southern part of Ljungdalen village (D3). This series contains the following rocks:

100 m crystalline schist with nodules of quartz.
10 m calcareous phyllite with lenses of marble.
50 m phyllonitic green schists rich in muscovite and containing concretions of quartz and calcite.
Overthrust.

Mylonites

On the eastern side the overthrust of the Helags nappe over the Särv nappe is marked by a distinct mylonite zone which I could follow from Tänndalen (F2) in the south to Mt. Kyrkstensfjällen (A3–4) in the north. These mylonites were encountered for the first time in the eastern slope of Mt. Gunnarstöten (E3). At Mt. Högfjället (E2) the violet veins, which are characteristic of the zone, appear. These veins are up to 1 dm wide, and consist of an extremely fine-grained mylonitic material (cf. also Pl. XII a) which contains porphyroclasts of the similarly mylonitic but coarser wall rock. The mylonitic wall rock can be described as a porcellaneous, white, flinty rock (cf. S.55).

In the region west of Mts. Flatruet and Tjölen, where the localities mentioned are situated, the formation of mylonite seems to have been especially intense. There the overthrust runs west of a large area with granite-mylonite, which locally forms a small nappe between the Särv nappe and the Helags nappe. These observations support the hypothesis expressed on p. 21 that the formation of the violet veins is favoured by the presence of granitic material.

North of the valley of R. Mittån similar mylonites have been observed more sporadically only, while at the same time the rocks of the zone of contact were more intensely recrystallized. The boundary between the two great nappes is marked above all by the fact that towards this boundary the complex of dolerite and quartzitic rocks of the Särv nappe exhibits increasing deformation and schistosity. Effects of this kind have been observed at Björnskallen south of Ljungdalen (D3), along the highway leading from the south towards Ljungdalen, and along the course of R. Ljungan in western Ljungdalen (G3) and north-west of this place.

To the west of these schistose dikes of dolerite and mylonites lie metamorphic schists. Among them are also found, as in the other parts of the Helags nappe, amphibolitic schists, but no low-metamorphic dikes of dolerite of the Ottfjäll type. The disappearance of the dikes of dolerite is thus an additional phenomenon characterizing the boundary between the Särv nappe and the Helags nappe. In the spots where the strongest differential movement has taken place in the dolerite bedrock, greenschists have been formed instead of light-coloured quartzitic mylonites. In the mylonite zone from Ljungdalen in the south to Rekådalen (A3) in the north an alternation between greenschists and quartzitic mylonite can thus be observed.

Igneous rocks

The bedrock of the Helags nappe contains several different types of eruptive rocks. The phyllonitic greenschists, which are often encountered near the base of the nappe, and which above have been described from Ljungdalen village, might possibly represent transformed effusive rocks. Metamorphism has, however, destroyed all primary structural features that might reveal such an effusive origin.

Several basic intrusive rocks occur. One of them appears in the form of strongly deformed bodies some metres thick. It exhibits, both in outcrop and in the hand-specimen, a blasto-porphyritic texture, and thus greatly resembles uralitized Ottfjäll dolerite. G. FRÖDIN (1922 a) records observations of metamorphic Ottfjäll dolerite in the crystalline Åre gneisses of the Helags nappe. In view of my own observations, I endorse FRÖDIN's view about that some deformed amphibolite dikes may have been original Ottfjäll dolerites. However, these rocks occur mainly near to the base of the Helags nappe.

Conformably to the schistosity, two other types of basic intrusives occur in the crystalline Åre schists of the upper Mittådalen valley (D–E2) in the shape of banks with a thickness of 1 cm to 1 m. Both types of rock are schistose and completely recrystallized. One of these intrusives is a dark amphibolite which contains limonitized titanomagnetite in the form of grains measuring several millimetres. This rock has been encountered, e.g., at the foot of Mt. Mittåhammaren (D2).

The other amphibolite is of a lighter colour, contains unusually dark biotite, and has been observed in R. Mittån (E2).

Owing to the strong metamorphism of these two intrusive rocks, their primary magmatic mineral composition and structure cannot be determined.

If we take the higher metamorphism of the Helags nappe into consideration, these two kinds of amphibolites are rather similar to those occurring in the blastomylonitic schists of the mylonite nappe at R. Biskopsån (p. 34).

Only one kind of acid intrusive rock has been observed. This was found at Mt. W. Helagsskaftet in the shape of a dike about 2 m wide, with a northern strike. It consists of a coarsely crystalline, white, trondhjemite-like rock containing albite, quartz, muscovite, and some small, scattered, red garnets. On account of the high degree of metamorphism and recrystallization throughout, it was not possible to decide whether or not the other larger occurrences of gneissic rocks, found in several places within the nappe, could be intrusives.

Ore-bearing zone

Widely scattered, most often euhedral crystals of pyrite are found occasionally in the rocks of the Helags nappe. A special zone of the nappe is however particularly rich in sulfide ore minerals. This zone, which has been observed only at the eastern boundary of the nappe, lies at a distance of about 1 km from the boundary, and could be traced from Mt. Ramundberget (E2) in the south to Mt. Smällhögarna (B3) in the north. The ore zone runs rather conformably to the boundary of the nappe. This seems to indicate that it occupies a distinct tectonic or stratigraphic level.

The indications of ore are often associated with a zone rich in quartz in which the latter appears either in the form of nodules or lamellae in calcareous chlorite schists, or the adjoining rock may be altogether a fairly pure quartzitic schist (cf. TEGENGREN, 1924). The ore minerals always occur in "schlieren" or veins conformable to the *s*-planes of the schist. Magnetite often accompanies the sulfide minerals, which as a rule consist of pyrite, pyrrhotite, chalcopyrite, and, less frequently, sphalerite and galena. F. R. TEGENGREN (1924) points out that the schlieren of ore often alternate with schists rich in quartz. J. H. L. VOGT (1887) is of the opinion that these occurrences of sulfide ores in western Härjedalen ought to belong essentially to one single level in the complex of schists. In the Trondheim area C. W. CARSTENS (1920) distinguishes between several topographically and genetically different types of ore. One of them, the Leksdal type, alternates as sedimentary layers with lava rocks and beds of jasper, which generally belong to the Bymark group of the Trondheim area.

In the spring 1961 there were in the ore-bearing zone the following mining claims:

1) Lillsjö copper mines at the southern end of L. Öjön (D3).
2) Vargtjärn copper mines on Mt. Vargtjärnsstöten (C3).
3) Herrångsskal mine in R. Härjångsån valley (B3).

TECTONICS

The tectonic investigations have been carried out with the aim of finding data for the regional tectonic development within the area examined. For this reason, special attention was paid to the Särv nappe with its peculiar tectonic structure. In this case, the particular problem was to determine the time of the intrusion of the Ottfjäll dolerite in relation to the Caledonian phases of deformation, folding and overthrust movements.

Within the area investigated the different tectonic zones exhibit locally peculiar structures. Traces of deformations of different ages are sometimes met with in the same area, but the direction of the axes of these deformations can be strongly rearranged locally, cf. Pl. III (esp. *19*, *30*, *31*) and Pl. IV 29–35. A direct computation of the deformation in the entire region might give misleading results, unless the peculiar features characteristic of the different deformations are taken into account.

I will therefore start with the description of petrofabric and deformation within some selected areas. This will guide us in the study of the regional tectonics, which has been investigated by means of a fairly large number of measurements on *s* and *B* structures, concentrated to limited areas within the area examined (Pl. III).

Fabric and deformation

The granite gneiss

At the watershed between the valleys of the rivers Särvån and Lunån north of Hede an obvious and sudden change in the bedrock makes its appearance in the band of augengneiss which has an E-W strike. While west of the watershed practically only augengneiss or granite mylonite with strong crystalloblastesis is found, mainly granite gneiss with cataclasis of varying intensity and certain zones of movement with more pronounced metamorphism and occasionally a tendency towards the formation of augengneiss have been observed to its east.

In the southern elongation of the line of contrast a N-S fault has been found west of Hede. Its throw amounts to at least some tens of metres, the eastern block being the higher one (S. 55; Stålhös 1957). The fault can be supposed to continue towards Mt. Särvfjället with the result that the eastern block has been raised, and the zone of the augengneiss together with the boundary of the Särv nappe displaced in northerly direction. This interpretation is supported also by the features of the map, since it is just at the mentioned line of contrast that the area of the granite gneiss begins with a decidedly greater N-S extension than the band of the augengneiss.

In my earlier paper I have shown that, at its south-eastern boundary at Mt. Orrnäsberget, the granite gneiss is thrust on top of, and imbricately into, the sedimentary rocks of the Vemdal quartzite (S.55, p. 206, and Figs. 3 and 4). The thrust plane has a fairly steep dip, viz. 20–30° NW–W. The steep thrust planes are explained by the tendency towards imbricate structure, since the overlying granite contains thrust-in flat masses of quartzite, while similar small nappes of granite mylonite appear in the quartzite below. Farther to the west, several zones of movement occur in addition in the granite gneiss. They are indicated by the formation of augengneiss. A particularly distinct zone of augengneiss of this kind is found in R. Södra Vemån at Fallkojorna, 14 km north-west of Vemdalen (F7).

DEFORMATION.—In many places within the granite gneiss north of Hede, two *s*-surfaces are particularly obvious, viz. N–NNE/W 80° and NE/NW 30°. The *s*-surfaces with a steep westerly dip appear especially in those localities where the rock has undergone strong metamorphism with a tendency towards the formation of augengneiss. In a greenschist within the area 2 km south-east of Vemvallen appear axes of folding of the directions N 60° W, N 75° W, and N 10° W. The greenschist is gently folded, and contains inclusions of quartz and calcite. Its main dip is W/N 35°.

At Fröåsen the granite gneiss, the petrology of which is described on p. 9, exhibits *s*-surfaces with mainly two orientations, viz. NW 30° and W 80°. Local differential movements occur // ENE. In the N-S striking *s*-surface, the quartz has been squeezed out into long, strongly undulating bands. The diagram (Pl. IV 1) shows the quartz to have a preferred orientation according to a type of *S*-tectonite with a weak girdle around a *b*-axis with a strike of about ENE.

FABRIC.—In a red granite gneiss from Mt. Fröåsen (F6) I have determined the fabric of the quartz. This determination has been carried out on two thin sections forming between each other an angle of about 40°. Subsequently, the separate diagrams have been orientated into the same plane, and added up. A plane of schistosity (*s*) is marked by squeezed-out *Langquarz* and by a ribbon-like distribution of epidote and other minerals (cf. Fig. 2 and Pl. IV 1).

The tectonic *ac*-plane exhibits four maxima which show the crystallographic *c*-axis of the quartz to be orientated mainly in this plane and preponderately at an angle of about 30° towards the plane of schistosity (cf. SANDER, 1948, p. 151, and D29, p. 360). The rock can be interpreted as an *S*-tectonite with some *B*-tectonite. As a result, the freedom of the axes of the quartz, allowing a certain rotation around the tectonic *c*-axis, which as a rule appears in an *S*-tectonite, has been suppressed. Instead of it a weak zonation around *b* has developed.

Linearities observed at Mt. Fröåsen strike between N and ENE. This results in a certain coincidence between the macroscopic fabric and the fabric of the quartz in the examined tectonite, the *b*-axis of which runs about N80°E.

An analogous fabric of the quartz in a gneissic granite has been described by F. J. TURNER (1948). According to this author, this slightly gneissic granite from Lake Manapouri, New Zealand, has a pronounced lineation parallel to *b*, and in the fabric diagram shows a tendency towards concentration of the axes of the quartz into two sectors forming angles of about 30° with the plane of schistosity and *a*, respectively. TURNER calls this kind of fabric "preferred orientation correlated with flattening".

Caledonian tectonites with the same kind of preferred orientation have been shown by F. C. PHILLIPS (1937) to occur in schists of the Moine Series, Kyle of Tongue, Scotland.

Similar quartz fabric has been found in sedimentary rocks from the Lochcarron area by M. R. W. Johnson (1960). Prominent quartz girdles striking NNE–SSW were also found in strongly deformed mylonites.

The augengneiss

DEFORMATION.—On account of the very coarse texture of the augengneiss, its planes of schistosity and its linear elements appear most often irregularly, and are determined only with difficulty. Steeply inclined joints traverse the rock in several directions. For this reason, structural examinations on *s*-surfaces and *B*-directions have preferably been carried out on schists which occur in the augengneiss. Most of the schists are of basic type. Diagrams I–IV in Fig. 22 represent structural conditions in such bands of schists in the augengneiss north-west of Hede.

Mt. Svartåsen lies about 8.5 km north of Långå. The entire mountain consists of a coarse ophitic dolerite embedded in augengneiss. This dolerite seems to be of pre-Caledonian age, and in the previous paper (S.55) I have called it Åsby dolerite. In the western part of the mountain the dolerite encloses a flat mass of red microcline granite of the Rätan type.

At the south-eastern foot of the mountain the dolerite has become strongly schistose. This had led to the formation of a greenschist with a north-westerly dip. In this schist, which is exposed for an area of 100 × 200 m, I have measured the orientation of the *s*-surfaces, the poles of which have been entered into Diagram I, Fig. 22. Unfortunately, as no tectonically determined direction could be observed, the diagram does not contain any linear elements.

Around the direction N 55° W a tendency towards a π-circle can be observed. The diagram, however, lends itself also to another interpretation. It is possible that the *s*-surfaces have first been arranged as hol-planes around an axis of folding in a direction N 12° W. The π-circle formed might then have been broken up in groups of poles (1, 2, 3), which were rotated in a varying degree around an axis of folding which was almost perpendicular to the first axis. This might point to a folding according to two axes $B \perp B'$ in the directions N 12° W–N 78° E.

In several localities with schistose augengneiss in the region around Vispvallen, 7 km west of Lunnäset village, I have determined *s*-planes and lineations. Thus the material for Diagram II in Fig. 22 has been collected from an area of about 4 sq. km north and south of Vispvallen. The schists exhibiting only rarely minor folding, the determined directions have been observed either in the form of parallelly arranged aggregates of porphyroblastic feldspar and quartz or as a pencil structure caused by intersecting shear joints.

Diagram II from the Vispvallen region indicates tendencies towards orientation of poles of *s*-surfaces around the β-directions N 20° W and N 70° E. These directions obviously do not coincide with the two clusters of linear elements in about N 55° W and N 72° W in the diagram. The directions N 20° W and N 70° E form two β-axes of the $B \perp B'$ kind, which nearly coincide with the two β-axes N 12° W–N 72° E that have been described from Svartåsen in the foregoing. Since the two β-axes indicated in Diagram II by the poles of the *s*-surfaces do

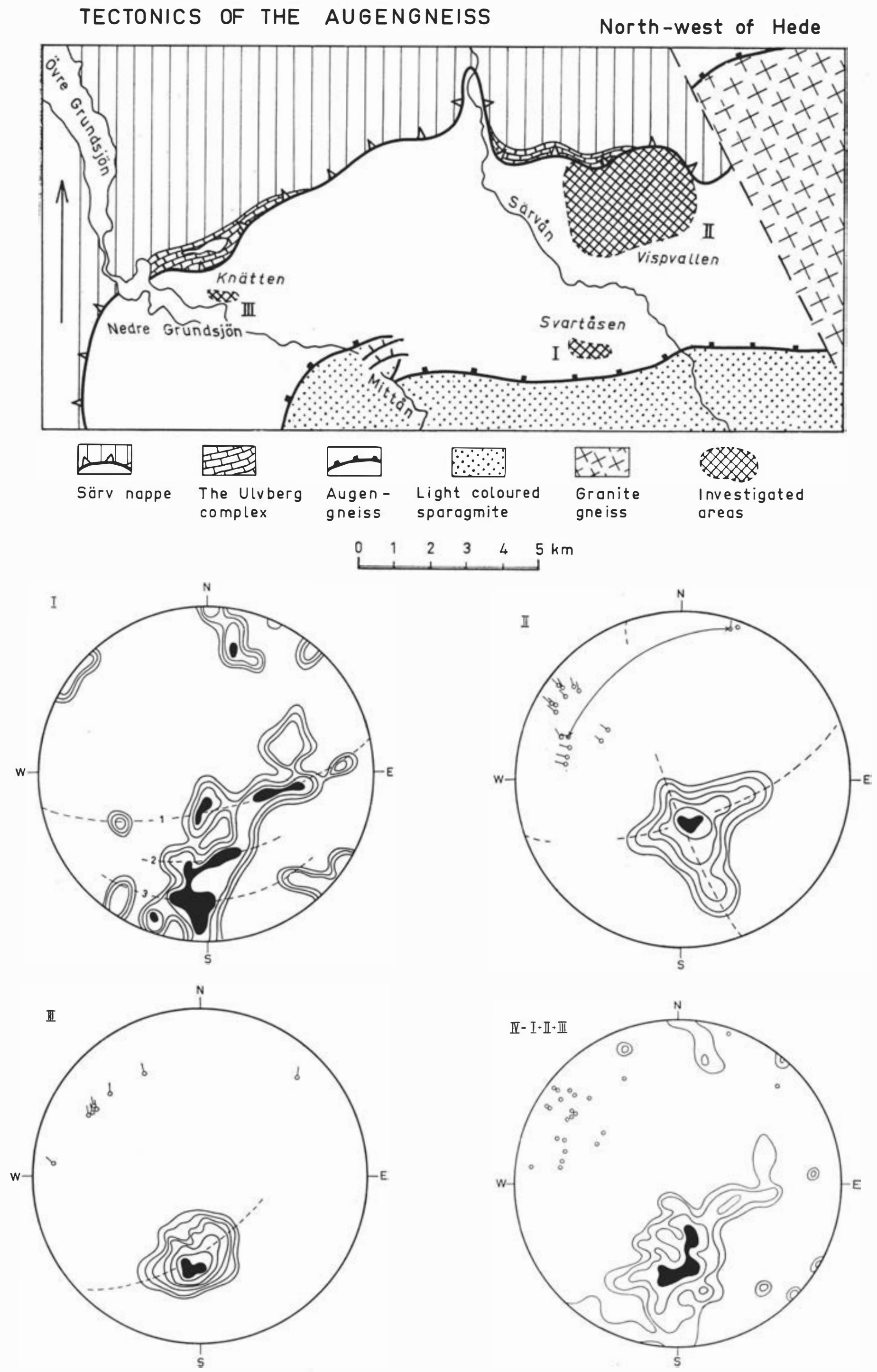

Fig. 22. Structural features of the augengneiss NW of Hede (F5–6). In the diagrams rings indicate *B*-observations. Contours of *s*-normals are: I (55 π_s) 25, 16, 8, 4, 0. II (36 π_s) 30, 20, 12, 8, 4, 2, 0. III (54 π_s) 40, 30, 20, 12, 8, 4, 2, 0. IV (145 π_s) 30, 12, 4, 2, 0.

not seem to be connected with any of the B-maxima, and since a change in the position of B with regard to β need hardly exist, we can venture the assumption that the augengneiss in the area of Vispvallen carries the imprint of several perhaps not contemporaneous deformations.

In Mt. Knätten near L. Nedre Grundsjön the augengneiss contains a comparatively thick bank of greenschist. The observed lineations are expressed in the form of a pencil structure which in this rock ought to have been caused by intersecting shear joints. The measurements have been taken within a comparatively small area (Fig. 22). Diagram III shows a weak indication of a B-maximum in a direction about N 50° W. The determined s-surfaces lie very close together, and therefore give no certain indication of a π-circle (= zone-circle of s-poles). The folding in the planes of schistosity is thus relatively weak.

The three diagrams, I–III, have been collected in Diagram IV. A rather distinct B-maximum appears in about N 55° W with some extension towards N 72° W.

The s-poles are rather spread, but determine no definite π-circle. Upon addition of the diagrams, the π-circles which are distinguishable in Diagrams II and III have practically disappeared. The distribution of the poles of the s-surfaces might most closely hint at a β-maximum that might be connected with the B-maxima in NW–NNW.

THE FABRIC.—For the augengneiss of Midskogsbygget (G4) the fabric of quartz, biotite, and apatite has been determined. Also in this locality the megascopic fabric of the rock exhibits several systems of s-planes producing by their intersection a pencil structure, as is often the case with the augengneiss.

The diagram (Pl. IV 3) which shows the fabric of the *quartz* contains a double ac-girdle with maxima about 50° from the ab-plane. The *mica* diagram (Pl. IV 4) shows a distinct maximum around c. This is subdivided into several sub-maxima that can belong to several sets of s-planes. The patterns of preferred orientations of quartz and mica might in part be homotactic, since the double girdle in the quartz diagram agrees in a certain measure with the maximum of the mica diagram around c.

In the diagram (Pl. IV 2) the *apatite* exhibits maxima close to b and also a girdle in the ac-plane. The individual crystals of apatite measure 0.01–0.05 mm, and occur mostly in biotite. This determination suffers from some uncertainty, since the weak birefringence of the apatite has resulted in diffuse orientations at extinction. The presumption (S. 55, p. 217) that the crystallographic c-axes of apatite show a preferred orientation with a maximum in b has further been fully corroborated by H. J. KOARK (1961).

The examined augengneiss can obviously be interpreted as a B-tectonite in which intersecting sets of s-planes have produced a B'-axis in the tectonic a-direction (SANDER, 1948). This B' points roughly towards WNW, while $b = B$

is directed NNE. Only very rarely is it possible megascopically to observe in the augengneiss any other B-direction than that pointing WNW. This direction is, on the other hand, the much more common in the way described above, while the real $b = B'$ appears only indirectly as $B \perp B'$.

Basal zone of the Särv nappe

FABRIC OF A MYLONITIC SCHIST AT THE TOSSÅSEN ROAD.—East of L. Storsjön a well-defined boundary between the mylonite nappe and the schistose basal zone of the Särv nappe is rarely to be seen. A recrystallized mylonitic schist from this boundary region at the Tossåsen road bridge over R. Västerån (C5) has been examined as regards the fabric of quartz and mica. The megascopic fabric at this locality is illustrated in Pl. IV 25.

In the diagram (Pl. IV 6) the 001 poles of the *muscovite* show a distinct preferred orientation in c, perpendicular to the prominent plane of schistosity, s.

The crystallographic c-axes of *quartz* in Pl. IV 5 are oriented to form a girdle around a b-axis. The preferred orientation of the quartz is heterotactic in relation to the fabric of muscovite with an amount of about 30°.

A comparison with previously determined fabric of schists in about the same tectonic position (S.55, Pl. II, diagrams VII–X) reveals certain analogies between the diagrams as regards the patterns of preferred orientation and the degree of heterotacticity between the fabrics of quartz and mica, respectively. An important difference lies, however, in the fact that the b-direction points about NW at Mt. Husberget and Mt. Ulvberget, but has a north-easterly direction at R. Västerån.

The localities with b in about NW are situated in strongly deformed zones immediately below dolerite dikes, whereas at R. Västerån a more homogeneous deformation is found. Already earlier I observed such a difference in the orientation of the $b = B$ and even β (cf. S.55, Pl. I, diagrams I–III and V in contrast to diagrams IV and VI).

At R. Västerån the fabric is distinctive also of the late age of the observed pair of $B \perp B'$ in N40°E and N50°W. In the megascopic fabric this pair of B-axes is determined as younger than the simultaneously observed pair of B-axes $B \perp B'$ in a direction NNW–ENE (Pl. IV 25). Another order of deformations has been found at Morslättvallen, Pl. IV 24. This may be due to a blockvise rotation in parts of the Särv nappe. Such a rotation is indicated by the orientation of joints in dolerite and quartzite, cf. Pl. IV 26–28.

Top zone of the Särv nappe

FABRIC OF A QUARTZ-MUSCOVITE-SCHIST FROM FLATRUET.—The rock is a fine-grained crystalline schist in which quartz and muscovite are postdeformatively recrystallized. A strong precrystalline deformation seems to have occurred in the form of a mylonitization. The rock belongs to an area with mylonitic schists rich in quartz, which tectonically are situated above the Särv nappe proper and underneath the Helags nappe. The megascopic fabric at Mt. Flatruet (E4) is distinguished by a pronounced foliation. On the *s*-surfaces lineations exist in directions of about N55°W and N5°W, of which the lineation in N55°W is the most prominent. In the surrounding areas of Mt. Särfvålen (D4) and Tjölen (E3) analogous lineations (Pl. III, and Pl. IV 37) appear.

The fabric diagram for crystallographic *c*-axes of *muscovite* (Pl. IV 8) shows a strong maximum which coincides with the pole of the foliation plane *S*. From this maximum around *c* the (001) of the micas are somewhat spread in the *ac*-plane, and thus indicate hol slip surfaces which are oriented tautozonally around a *b*-direction.

The *quartz* of the rock has developed a preferred orientation with the crystal axes in a particular direction in the *s*-plane. At the same time, the quartz axes are distributed in a diffuse girdle which, with the mica girdle, determines the quartz maximum as an *a* maximum (cf. also KVALE, 1945, who in a similar case, however, interpreted the quartz maximum as a *b* maximum). The homotactic girdles of quartz and mica show a tendency towards *B*-tectonite in the fabric of the rock. The lamination of the rock, the mica maximum around *c*, and the quartz maximum in *a* are nevertheless indicative of a pronounced *s*-surface connected with a "flattening".

There is some similarity between the fabric of this recrystallized mylonitic schist and that of a slickenside mylonite of Melibokus, Odenvald, as described by SANDER (1948, pp. 228–231, D22–27). The only difference between the preferred orientation in these rocks is that in the fabric of the quartz-muscovite-schist from Flatruet, a certain tendency towards *B*-tectonite can be seen.

Inner parts of the Särv nappe

TECTONICS OF THE SÄRV NAPPE.—The Särv nappe is built up of gently folded series of rocks in which swarms of dikes of Ottfjäll dolerite appear. The conditions prevailing at the lower boundary of the Särv nappe have earlier been described (S.55). According to these observations the Ottfjäll dolerite was intruded at a relatively early phase of the Caledonian orogenesis, whereupon considerable thrust movements took place.

The Särv nappe receives its particular structure mainly from the thousands of dikes of Ottfjäll dolerite. For this reason, I have chosen to date the deformations in this nappe in relation to the time for the intrusion of the Ottfjäll dolerite, which on the whole seems to have taken place simultaneously in the entire nappe. In this way pre-intrusive, synintrusive, and postintrusive deformations can be distinguished. Thus the following tectonic phenomena are distinguished.

PRE- OR POSTINTRUSIVE DEFORMATIONS.—Folding, thrusting, or regional dynamic metamorphism could conceivably have influenced part of the rocks of the nappe before the time of the Caledonian thrust movements or the Caledonian regional metamorphism.

The metamorphism that characterizes the rocks of the Tännäs complex presents an instance of deformation which, in all probability, is older than all the other deformations in the nappe. This could possibly apply also to some part of the deformation which can be distinguished in the polymict conglomerate with detritus of basic eruptive rocks in Mt. Grönstacksberget (G3).

In the fragments of potassium-rich felsite, which enter into the composition of the red breccia from the Uggvallen region (G4), we can always trace a distinct lamination with a pronounced *s*-plane. This may, like the agmatitic brecciation in the rocks of the Tännäs complex, be attributed to a pre-Caledonian deformation. On account of the great predominance of the angular fragments of felsite in the breccia, we must suppose that these fragments have undergone a slight sedimentary transport, and constitute the remains of an older and partly deformed bedrock.

Both the light-coloured sandstone in the Särv nappe and the polymict conglomerate in the red sparagmite at Hede contain isolated pebbles of rocks with pre-Caledonian deformations, yet these deformations can be of a very early date.

For the relative dating of the Ottfjäll dolerite the pre-intrusive deformations are of special importance. For this reason, I have always looked for distinct traces of such deformations in the course of the field-work. In this connection, however, I have generally found that no deformations occur in the immediate vicinity of the dolerite dikes in those parts of the nappe where most of the surrounding sedimentary rocks have become completely schistose. Thus blastopsammitic relict textures are often retained only in the direct contact between dolerite and the sedimentary rock (cf. Turner, 1948).

It appears as if the stiff, thick plates of the dolerite had reinforced the bedrock with the result that deformation and metamorphism had become less intense locally. In the bedrock between the dolerite dikes, one might expect to find traces of deformations which had taken place prior to the intrusion of the dolerite. In order to decide whether or not some of the traces of deformation in the sedimentary rocks near the dolerite dikes are older than the latter, I have examined the structure in some suitable places.

I. *Mt. Bläckberget.* In my previous paper (S.55, diagrams XI and XII), I accounted for the quartz and mica fabrics in a moderately coarse, somewhat recrystallized arkose-sandstone from Mt. Bläckberget.

The arkose-sandstone, which is of the same type as that described from Mt. Gråvålen, is distinctly schistose with a preponderantly flat northern dip of the *s*-planes, and slightly folded. The amplitude of the folding is about 10 m, the direction of the folding axes roughly north and east ($B \perp B'$). At intervals of about 50 m the sandstone is traversed by three dolerite dikes about 5 m wide, which dip about /NW 50°, and thus cut the *s*-planes at an acute angle.

In spite of the considerable recrystallization of the schistose arkose-sandstone, the preferred orientation of its quartz is only slight. The most pronounced feature in the fabric of the rock is found in the heterotactic orientation of quartz and mica. The *s*-plane, which is expressed by the preferred orientation of the mica, is also observable macroscopically in the rock. Owing to the fact that this *s*-plane, which is intersected by the dolerite dikes, seems to be older than the latter, preferred orientation of the mica can also be assigned a comparable age. The mica shows a tendency towards zonal development about the direction N70°W. This zone of mica thus indicates the direction of a transverse deformation which is possibly older than the dolerite dikes.

The orientation of the quartz is mainly influenced by postintrusive deformations. An analogous pattern of preferred orientation occurs also in other localities in the Särv nappe, and will be discussed in greater detail in the following.

II. *Mt. Gråvålen.* One of the westernmost peaks of Mt. Gråvålen (G4), about 1200 m south-west of Uggvallen seter, contains two parallel dolerite dikes, which are separated by an interval of 5 m, and traverse a banked, quartzitic, coarse feldspar sandstone.

Diagram Pl. IV 22 shows 20 *s*-planes in the intervening sandstone. These planes exhibit a tendency towards tautozonal orientation around two poles in the ENE and the SE, respectively. The 17 lineations seem to be developed mainly by intersection of *s*-surfaces and to a less degree by microfolding on the *s*-surfaces.

From these conditions the conclusion can be drawn that the slight folding of the *s*-planes is directly connected with the impression of the linearities upon the *s*-planes. In spite of the fact that the angular difference between them amounts to only about 80°, these *B*-axes seem to belong together, and most likely are to be interpreted as a pair of axes of folding $B \perp B'$.

At this locality the dip of the dolerite dikes is N30°W/65°W, which is indicative of some connection between the system of dolerite dikes and the pair of deformation $B \perp B'$. The angle between the dolerite dikes and the presumably sedimentary *s*-planes is about 75°.

A connection between the orientation of the dolerite dikes and the deformation in the adjacent rocks can depend upon two factors. Either the dolerite has been intruded into a system of joints formed during the same deformation which caused the folding in the *s*-planes and the impression of the *B*-directions, or the existence of the dolerite dikes has guided a postintrusive deformation by mechanical anisotropy in such a way that its *B*-axes have been twisted into conformity with the dikes.

In the quartzitic sandstone the fabrics of quartz and mica have been determined. The *mica* (diagram Pl. IV 18) partly shows an orientation with the crystallographic *c*-axes perpendicular to the dominant *s*-plane, and is partly collected into a zone with the *b* roughly in the north.

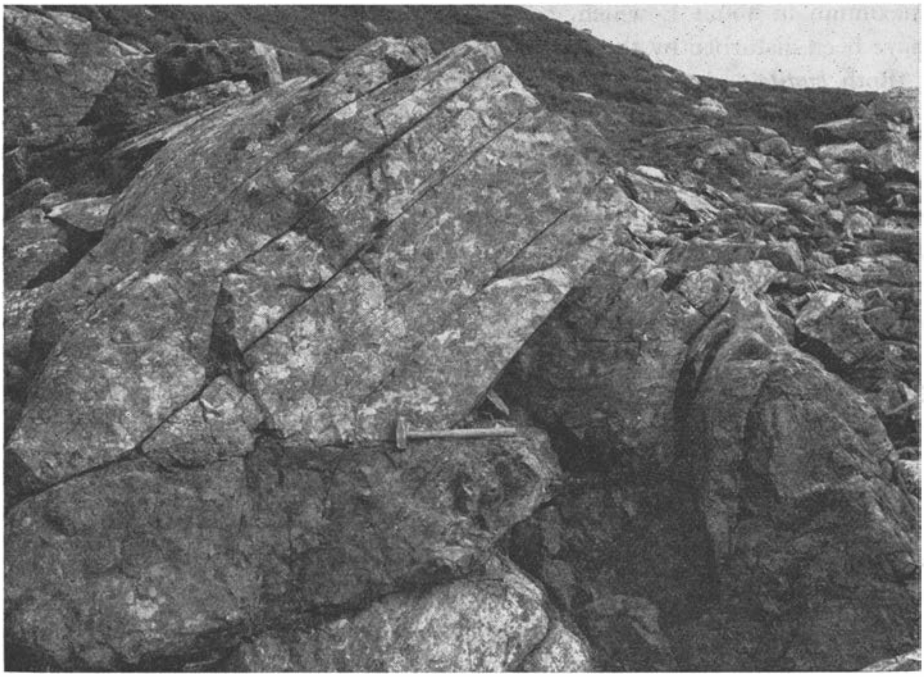

Fig. 23. Dike of Ottfjäll dolerite, partly cutting the *s*-planes in a sericitic quartzite schist, partly occurring as a somewhat deformed sill. The quartzite is less schistose in the contacts to the dolerite. The shaft of the hammer points E. Outcrop near the base of the Särv nappe in the eastern part of Mt. Anarisfjällen (B5).

The fabric of the *quartz* is more complex. Here, too, an analogous double zone can be noticed, which, like the fabric of the mica, indicates a *b*-direction roughly N (Pl. IV 17).

Thus the fabrics of quartz and mica do not show any direct agreement with the megascopic fabric. This bears witness to a possible difference in age between the deformations expressed in the macro-texture and the fabric. The mica fabric seems, at least partly, to have reproduced the pre-existing *s*-plane in the rock, and in this way would be partly pre-intrusive in character.

III. *R. Bruddabäcken.*—A biotite-bearing, recrystallized sandstone from R. Bruddabäcken (B5) was made the object of petrofabric investigation. The somewhat schistose sandstone is intersected by Ottfjäll dolerite dikes, which are about 10 m broad and dip NNW/40°W (Pl. IV 23). The *s*-surfaces of the sandstone show slight dips towards E or ENE. Lineations are gathered in two maxima with the directions ENE and E. The *s*-poles are spread in a zone circle (Z), indicating a π-pole ($=\beta$-maximum), which shows a remarkable coincidence with the strike of the dolerite dikes. These circumstances seem to indicate that the β and the ENE *B* form a pair of axes $B \perp B'$, which is closely connected with the orientation of the dolerite dikes. There is also another *B*-

maximum in about E which, contrary to the B-maximum in ENE, seems to have been disturbed by the above deformation $B \perp B'$.

Both *biotite* and *muscovite* show patterns of preferred orientation connected with "flattening" (Pl. IV 20–21). The muscovite (20) is partly crystallized earlier (bent crystals) than the biotite, which shows postdeformational crystallization. Furthermore, the *b* of the biotite diagram (21) seems to be somewhat rotated in a northerly direction relatively to the muscovite fabric. The crystal axes of *quartz* show a rather diffuse preferred orientation with a faint *b*-girdle and some other girdles of uncertain significance (Pl. IV 19). Consequently, the fabric of quartz is not homotactic with those of the micas.

It is probable that the deformation with $b (= B)$ towards ENE is entirely postintrusive and is due to a "flattening" of the complex of dolerite and sandstone. The disturbed B in about E seems to be older than this "flattening" and may be either postintrusive or possibly pre-intrusive.

POSTINTRUSIVE SCHISTOSITY.—A schistose feldspathic quartzite is found at *L. Ränntjärnarna*, about 2 km west of Uggvallen seter (G4). The outcrops are shown at *ss* in the section (Fig. 18). The megascopic fabric exhibits lineations in about N40°E and N50°W, together with β-maxima in the same directions.

The 001 poles of *muscovite* (Pl. IV 14) are gathered in a maximum around *c*, nearly perpendicular to the main plane of schistosity *ab*.

The crystallographic *c*-axes of *quartz* show a pattern of preferred orientation which includes a double girdle, the *b*-axis of which has an orientation of about N50°E (Pl. IV 13). Thus the lineation in about N40°E seems to be a $b = B$-direction, probably connected with the lineation in about N50°W as a pair of axes $B \perp B'$.

In this rock lengthened coarse quartz grains occur, which seem to be recrystallized only marginally. The preferred orientation of the optic axes in these *Langquarze* is shown in Pl. IV 15. However, the fabric of this quartz is rather similar to the fabric of the more fine-grained quartz (Pl. IV 13).

Some of the quartz grains contain very fine needles of *sillimanite*. These sillimanite needles are preferably orientated in the tectonic *ab*-plane, with a tendency towards a maximum in *b* (Pl. IV 16). However, this maximum may be somewhat uncertain, as the investigation was made on one single *ac* thin section and the frequency of determined sillimanite needles with orientations near the *ac*-plane may have been suppressed.

Pronounced schistosity in the Särv nappe has been observed in connection with sheared dolerite dikes and is then proved to be of postintrusive age. Very weak deformations only could possibly be pre-intrusive. In accordance with these facts the described, rather strong deformation at Ränntjärnarna is supposed to be postintrusive.

At *Mt. Svartuggen*, south of L. Nedre Särvsjön (F5), I have noticed a crystalloblastic feldspathic quartzite including an abrupt transition between a

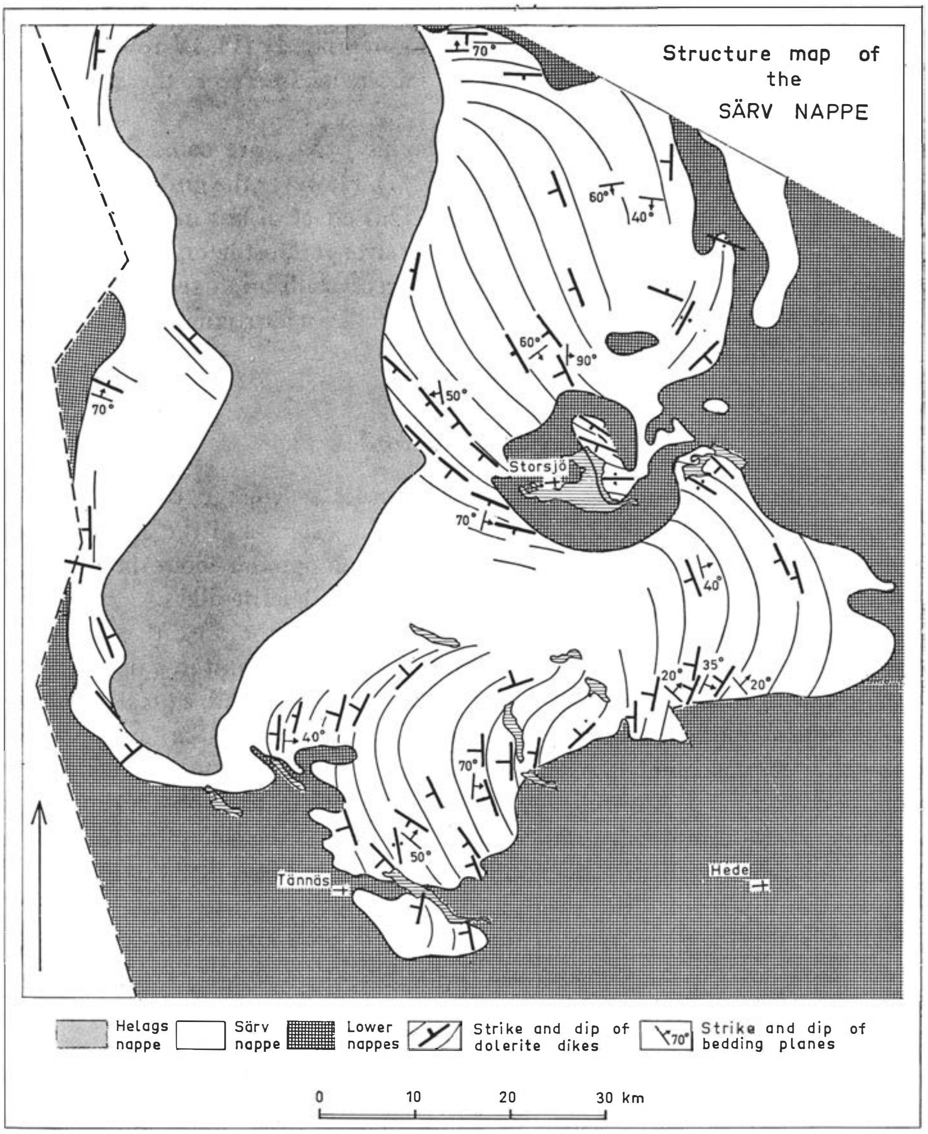

Fig. 24. Dolerite dikes and bedding planes in the Särv nappe. The dip of the dikes varies between 25° (longest dip line) and 90° (a point on either side of the strike line).

coarse crystalline rock and a fine crystalline rock of about the same composition. The sample was collected from a boulder in the talus near the mountain face, and for this reason could not be orientated. By means of a late small thrust movement rocks from two layers with some difference in crystallization have been placed side by side so near together that they could be studied in the same thin section.

In both rocks *muscovite* shows about the same pattern of preferred orientation of the 001 poles, viz. a rather strong maximum around *c* (Pl. IV 10 and IV 12). The *s*-plane of this recrystallization of muscovite intersects the small-scale thrust (Pl. IX b).

Also the fabrics of *quartz* (crystallographic *c*-axes) are somewhat similar, with girdles in an *ac*-plane (Pl. IV 9 and IV 11). However the girdle is weaker in the quartz fabric of the fine-grained rock. This effect of less distinct preferred orientation seems to have been caused by the stronger postdeformative crystallization in the finer crystalline rock. This recrystallization seems to be connected with a preferred orientation of the quartz axes in maxima in the tectonic *a* and *c* directions.

Regional tectonics

Several cases of deformation which in most cases are younger than the Ottfjäll dolerite have been studied. A knowledge of the extent and the orientation of these deformations has made it possible to form an opinion about the degree of the deformation at the time of the formation of the dolerite dikes.

In the tectonic treatment of the different parts of the area, the scarcity of exposures has made it impossible to cover every tectonic unit in a uniform way with points of observation. Where the bedrock is well exposed, a greater number of determinations has been carried out instead. Work in the Särv nappe which from the tectonic point of view is of special interest, has to a certain extent been concentrated on a search after localities in which tectonic observations were possible. Every integral diagram on Pl. III represents the tectonic conditions in a special locality often covering relatively small areas of the order of magnitude of 10^4 sq.m.

There are marked contrasts in the appearance of *B*-axes in the different tectonic units. In the augengneiss and in the basal zone of the Särv nappe a *B*-direction of N57°W predominates (Pl. IV 36). The inner parts of the Särv nappe exhibit a distribution of lineations along the periphery of the diagram, which may be due to block-wise horizontal rotations of parts of this nappe (Pl. IV 38). A certain preponderance of a *B*-maximum in about ENE may be distinguished. In the mylonite zone below the Helags nappe and to the west of the main part of the Särv nappe two *B*-axes appear, viz. in the directions NW and N (Pl. IV 37).

Dolerite dikes and bedding-planes in the Särv nappe

Quite a number of observations prove the existence of a certain connection between the orientation of the dolerite dikes and the dip of the planes of sedimentation in such a way that the latter often form an angle of almost 90° with the former. This circumstance suggests that the system of joints into which

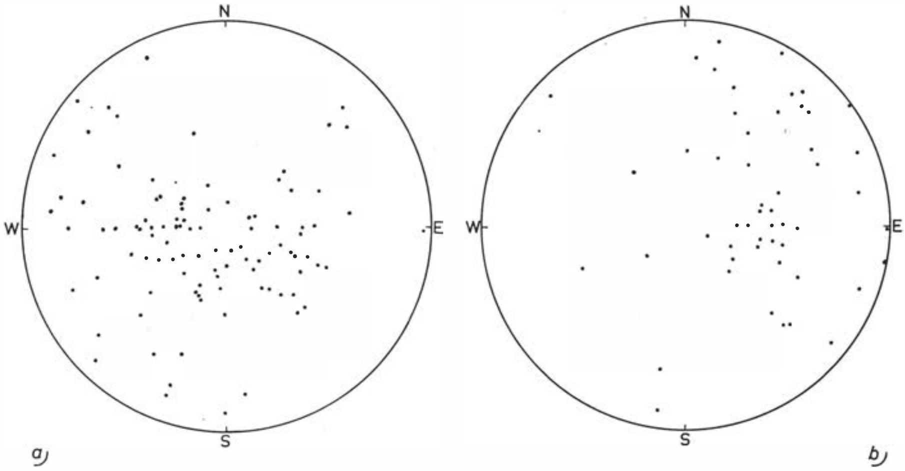

Fig. 25. Bedding planes and dolerite dikes in stereographic projection according to Lambert (lower hemisphere). (*a*) Normals of 102 *s*-surfaces of supposed sedimentary origin in the Särv nappe. (*b*) Dipping planes (normals) of 52 Ottfjäll dolerite dikes or groups of dikes in the Särv nappe.

the dolerite was intruded was at right angles to the sedimentary bedding-planes irrespective of whether the latter were undisturbed or had been folded prior to the intrusion.

Fig. 25 *a*) represents about hundred observations of the orientation of the *s*-planes in the sedimentary rocks of the Särv nappe in a projection according to Lambert-Schmidt. These *s*-surfaces most frequently constitute planes of sedimentation which I was especially keen to determine. On account of the uncertainty implied by such an identification, it is necessary to consider the possibility that a minor part of these *s*-surfaces can have been tectonically influenced, and can have an orientation which deviates from the plane of sedimentation. This material comprises only scattered observations in the field, but not the numerous determinations of the *s*-surfaces that have been carried out upon especially suitable exposures within the nappe. The diagram represents normals to the *s*-surfaces. They exhibit a mainly equatorial accumulation into a great circle (zone of *s*-normals, cf. SANDER, 1948), the normal of which, the π-pole, is orientated in a direction of about N 10° E. This π-pole marks the position of the axis of folding (B) of the large-scale folding that especially characterizes the sedimentary rocks of the Särv nappe. The amplitude of the folding has the order of magnitude 100 m. The strike of its B-axis is almost coincident with the Caledonian main axis.

In addition to the large-scale folding mentioned, the diagram also shows a spreading of the poles of the steeper *s*-planes towards the south and north. The fact that exactly the steepest *s*-planes have suffered from this effect might be due to their fan-shaped distribution around a sub-vertical axis of folding of unknown age. A partial overturn towards the east may have taken place. Their participation in this disturbance indicates at the same time that the affected parts of the bedrock of the nappe have been very mobile, and thus could be exposed also to a folding according to an east-westerly *B*-axis, e.g. the *B*-direction in about N 75° E indicated in Fig. 25 *b*).

Fig. 25 *b*) shows that the roughly E–W striking dikes of dolerite are more steeply inclined than others. This might be explained by the assumption that the dip of the former could not be essentially changed by an overturn to the west. Fig. 24 shows likewise a geographical connection between the dolerite dikes with an E–W strike and the anticlinally protruding portions of the basement of the nappe. Such an effect might have been produced in connection with the overthrust by the rotation of a whole block of the nappe owing to the resistance of the basement. Simultaneously a folding along sub-vertical axes ought to have taken place, as indicated in diagram 25 *a*). Since the dikes of dolerite often occur in groups with mutually sub-parallel orientation, the material is rather restricted in the field. For this reason the diagram does not provide statistic certainty. If this circumstance is kept in mind, certain conclusions can nevertheless be drawn. The dolerite dikes obviously have a preponderating orientation in N–S/35° W. Starting from this main strike a spreading of the poles of the dikes occurs partly in an arc of a circle about the direction N 75° E and partly towards steeper and sometimes vertical dips. It is not possible to decide what part of these scattered orientations is due to joints existing at the time of intrusion.

In an attempt at summing up the mentioned conclusions based upon the two diagrams one might say that the *s*-planes in the Särv nappe are orientated tautozonally around a *B*-axis about N 10° E, while the dikes of the Ottfjäll dolerite seem to exhibit a feeble tautozonality around a *B*-axis about N 75° E which is possibly younger.

METAMORPHISM

The total metamorphism within the region examined is due to varying contributions of dynamic metamorphism and regional thermal metamorphism. Thus in the large nappes the rocks are characterized by a preponderant thermal metamorphism, while the rocks in the boundary regions between the units, and especially in the thrust zones, have been particularly influenced by dynamic metamorphism. The mylonite nappe with its strong dynamic metamorphism constitutes an exception to this rule.

As a result of the relation existing between the metamorphism and the tectonic position of the rocks, a distinct zoning can likewise be distinguished. It must be pointed out, however, that this division into tectonic zones implies a rather strong generalization of the existing conditions. The forms of metamorphic effects mentioned constitute subjectively chosen examples of the metamorphism which I have found to be the principal character of the zone in question.

	Metamorphism
THE HELAGS NAPPE	Albite-epidote-amphibolite facies.
Bottom zone	Mylonitization, formation of blastomylonite and blastophyllonite.
Zone of movement	Recrystallized mylonites with secretions of quartz and calcite. Veinlets of ultramylonite.
Top zone	Shearing of dolerite dikes and rotation of dolerite fragments together with development of schistosity in adjoining quartzitic rocks.
THE SÄRV NAPPE	Greenschist facies (biotite-chlorite to muscovite-chlorite subfacies).
Bottom zone	Shearing of dolerite dikes, development of strong schistosity in them and in quartzitic rocks. Veinlets of ultramylonite.
Zone of movement	Strong differential movements with crystalloblastesis of quartz and feldspar.
Top zone	Formation of pegmatitic secretions containing quartz, microcline, hematite, and siderite. Strong crystalloblastesis.
THE MYLONITE NAPPE	Mylonitization and recrystallization, strong crystalloblastesis especially of microcline. Greenschist facies.
Bottom zone	Cataclasis.
Zone of movement	Mylonitization. Veinlets of ultramylonite.
Parautochthon	Cataclasis and recrystallization.

Zones of intense movements

In relation to the surrounding bedrock, cataclastic rocks and especially mylonites are probably as a rule formed at a late stage. We have to reckon with the possibility that many of the rocks which have been recrystallized in the course of the processes of regional metamorphism may have suffered at least a partial crushing in an early phase. Such a temporarily crushed rock can either have been altered immediately and continuously through successive or repeated growth of the crystals, provided that conditions favourable to recrystallization existed, such as a high content of volatile components and increased pressure and temperature, or it can have been induced to recrystallization through a renewed change of the environmental conditions, after having existed as a mylonite during an anorogenic intermediate period.

From the thermodynamical point of view, a finely ground mixture of minerals represents an unstable compound which tends to change into a more stable stage at the first opportunity. This process is combined with the growth of the crystals or, under altered external conditions, the formation of new crystals. One of the laws governing these processes is a rule which is of fundamental importance for the growth of crystals, viz. that the solubility of a certain crystal depends on the radius of curvature in the crystal particle. In a simplified form this relation can be expressed by saying that the solubility of the crystallized substance decreases with increasing size of the crystal. In consequence, the bigger crystals in a mixture of crystals or fragments of such will tend to grow at the expense of the smaller ones.

An important proof for the intensity and the long duration of the movements in the thrust zones is supplied by the dynamic metamorphism which generally characterizes these zones.

However, a postdeformative crystallization, which will be described in the following, has within large areas regionally metamorphosed mylonites and phyllonites into the present crystalloblastic schists. Although in such cases the structures of the rocks reveal a strong precrystalline deformation, it is nevertheless difficult to express an opinion on the intensity of a previous mylonitization.

The mylonite zone underneath the Helags nappe contains apparently fresh mylonites as compared with the mylonites in other zones of movement within the investigated area. This is one of the evidences of relatively late thrusting movements of the Helags nappe. These flinty, white mylonites contain veinlets of violet ultramylonite (S. 55), which occasionally can be 10 cm broad.

K. Ø. BRYN (1960) has described similar ultramylonitic veins from a mylonitic zone about 3 km NW L. Essandsjø, Norway, which seems to be situated in an analogical tectonic level.

Origin of the augengneiss

Scandinavian geologists have for a long time held different views on the formation of the augengneiss.

During the later part of the nineteenth century the augengneiss was frequently considered as marking a stratigraphic horizon in the sparagmite Series (HÖRBYE, 1855; TORELL, 1888; TÖRNEBOHM, 1882; HÖGBOM, 1920). In 1896, however, TÖRNEBOHM described the augengneiss as a pre-Cambrian intrusive rock. HÖGBOM, who had noticed the cataclastic structure of the rock endorsed the opinion of TÖRNEBOHM.

O. E. SCHIÖTZ (1891) had, however, arrived at the conclusion that the augengneiss should be an intrusive rock of Caledonian age. This opinion was accepted by G. HOLMSEN and J. SCHETELIG and gradually also by K. O. BJÖRLYKKE (1905). GOLDSCHMIDT (1916, p. 120) stressed the difficulty of reaching a final decision about the age of the augengneiss, but appeared most inclined to consider it, like BJÖRLYKKE, as a Caledonian intrusive rock.

In 1924 C. W. CARSTENS described augengneiss from several localities at boundaries between different complexes of rocks, and on this occasion used the designation "veins" of augengneiss. In his paper of 1928 (p. 236), however, CARSTENS points out that field-work and excursions, especially in Drivdalen and neighbouring regions, have unequivocally shown most augengneisses in the marginal zones of the Trondhjem region to represent transported nappes of the bedrock, probably nappes of the Archaean. The elongated eyes of feldspar may simply be products of the deformation of older augen granites. ASKLUND (1960, p. 19) seems to agree with this opinion as far as the augengneisses of Jämtland are concerned. This opinion, which is due mainly to TÖRNEBOHM, receives its most powerful support from the fact that the augengneisses generally appear as nappes of Archaean material. This has been particularly stressed by WEGMANN (1926, 1959).

ROSENQUIST (1944) has emphasized that the structure of the augengneisses is mainly postdeformative and thus of late Caledonian age. In his endeavour to point out that no reason exists for considering the augengneisses as pre-Cambrian intrusives, ROSENQUIST (1944) turns to supporting the view that the augengneisses are admittedly metamorphic, yet exclusively Caledonian rocks, presumably connected with the series of the green effusives. T. F. W. BARTH (1938) has, however, shown that a strong porphyroblastesis of potassium feldspar can occur also in sparagmite rocks (flagstone).

In the area examined I found the augengneiss to be a metamorphic rock' the formation of which seems to have been influenced by the following three factors: (1) the primary chemical composition; (2) strong deformation; (3) a mainly postdeformative crystalloblastesis with abundant access to aluminium and potassium (cf. ROSENQUIST, 1944).

Between these three conditions there exist mutually compensating relations. Thus the surplus of Al and K mentioned under (3) may be attributed mainly to the primary chemical composition of the rock (1). The deformation according to (2) implies in a similar way a mobilization and transport of the components of the rock which facilitate the final crystalloblastesis according to (3).

Sometimes, albite-porphyroblast-schists seem to have been formed primarily. This is illustrated by the example of an augengneiss from Mt. Busjöfjället (C7). The rock, which is situated close to an exposure of schistose amphibolite, contains reddish perthite porphyroblasts in which microcline seems to have replaced albite (Pl. IX a cf. K. HEIER, 1955).

My observations in the great masses of augengneiss east and west of Tännäs show that in this region the primary material of this rock has consisted to an overwhelming degree of pre-Cambrian (Archaean) rocks with granites and pre-Cambrian basic intrusives. In spite of the fact that we have no positive proof for the assumption that Caledonian intrusive rocks have also contributed to the augengneiss, nevertheless this opinion appears most probable. Thus in the augengneiss no undisturbed dikes of Ottfjäll dolerite but marginally schistose massifs of Åsby dolerite have been observed (S.55).

The augengneiss always contains a comparatively large quantity of mafic material, which at all events may have been derived at least in part from basic rocks intruded into the augengneiss (ROSENQUIST, 1944). It cannot be decided whether the mafic material in the augengneiss has been introduced prior to, or in more direct connection with, the precrystalline deformations in the augengneiss. It

Table 2. Comparison between the chemical composition of augengneisses, granite, and a gabbroic rock.

	Granite Ekornaaen (C 1), C. W. CARSTENS (1920, p. 29)	Augengneiss, Rien (E 1), C. W. CARSTENS (1920, p. 29)	Augengneiss Midskogs-bygget (G4),*	Gabbro, mean composition according to R. A. DALY
SiO_2	76.50	68.95	66.65	48.24
TiO_2	0.14	0.51	0.73	0.97
Al_2O_3	11.22	10.72	14.51	17.88
Fe_2O_3	1.55	6.44	2.04	3.16
FeO	0.57	1.35	2.24	5.95
MnO	n.d.	n.d.	0.08	0.13
MgO	0.74	1.29	1.68	7.51
CaO	1.21	2.29	2.30	10.99
Na_2O	3.03	3.70	2.90	2.55
K_2O	4.98	4.34	4.78	0.89
P_2O_5	n.d.	n.d.	0.21	0.28
H_2O	0.50	0.71	1.02	1.45
Total %	100.44	100.30	99.14	100.00

* Central analytical laboratory, Uppsala 1960.

is possible that intrusions of basic rocks in a granitic environment have created especially favourable conditions for the metamorphism of the complex of rocks into augengneiss.

The suggestion of such an intrusion at the pronounced zone of movement is supported by WEGMANN's (1959) assumption that the supply of basic intrusives has taken place mainly via movements zones below the large nappes. This would explain, according to WEGMANN, why some of the feeding dikes for the great quantities of certain intrusives, e.g. the Ottfjäll dolerite and others in the Trondheim region, have never been observed. My own opinion is that the last-named circumstance would not exclude the possibility, as I have already shown, that the rock floor of the Särv nappe might have been pushed forward over a considerable distance after the intrusion of the dolerite, when existing sup-plying dikes would have been left behind farther in the west.

In this connection, a comparison between some analyses of augengneisses can elucidate these conditions (Table 2).

In Table 2 the values obtained from the analysis of the augengneiss from Midskogsbygget (G4) and another analysis of augengneiss from Rien (E1) have been compared with those of an acid and a basic rock. The comparison has to be taken merely as an attempt to investigate whether or not augengneiss of the chemical composition expressed by the analyses could have been formed by mylo-nitization and porphyroblastesis of granite and a basic rock. From a purely *chemi-cal* point of view, this would obviously be possible, especially if it is taken into consideration that the actual composition of the two original rocks can have differed from those now compared. Thus the granite from Ekorrnaaen (C1) in the Riksgränsen anticline contains rather little Al and much Si.

The augengneiss from Rien (E1) belongs to that part of the mylonite nappe which lies near the granite of the Riksgränsen anticline. C. W. CARSTENS (1920) had there two rocks analyzed in order to show the consanguinity between them. In the augengneiss from Midskogsbygget the admixture of material from basic rocks to the granite mylonite seems to have proceeded one step further.

There may also exist some connection between the occurrence of augengneiss and the rocks of the Tännäs complex. These rocks have a high content of potassium, which beyond doubt would favour the growth of microcline porphyroblasts in the augengneiss. Coarse augengneiss seems to occur everywhere in the neighbourhood of rocks of the Jotun group (cf. TÖRNEBOHM, 1896, Tafl. 1.). A near connection between the coarse augengneiss and the Lower Jotun nappe has been presumed by P. HOLMSEN & CHR. OFTEDAHL (1956).

Post-tectonic recrystallization

In the foregoing I have shown that the Ottfjäll dolerite has been intruded into the Särv nappe at an early date or pretectonically, and that the dolerite dikes exhibit great uniformity in their occurrence and distribution within the nappe. This proves that the processes which have taken place in the nappe in the course of postintrusive overthrusts and deformation must have been very uniform. In spite of the circumstance that a syntectonic deformative activity thus must have been on the whole uniformly distributed within the nappe, different portions of the nappe exhibit nevertheless distinct differences in crystallization and degree of metamorphism. In the western and northern portions of the Särv nappe the bedrock is more crystalline than in the south-eastern part. At the same time traces of local deformations are more distinct in the south-east than in the western and north-western parts.

Also the difference in the degree of deformation between the overthrust bedrock of the nappe and the pronounced zones of movement between the nappes, where the overthrusts have taken effect is most distinctly visible in the east. On the other hand, in the west it is difficult in most cases to distinguish the corresponding zones of intense movements, e.g. between the Särv nappe and the Helags nappe. On this account it might be supposed that in the west the nappes should have undergone a more universal deformation which might there have resulted in a uniform recrystallization. Such an assumption is, however, opposed by the occurrence of slightly deformed dikes of dolerite in the western limb of the Särv nappe. This circumstance, together with the observation of recrystallized mylonite rocks at the Riksgränsen anticline, proves the occurrence of overthrusts *en bloc*, and makes it probable that also in the west pronounced zones of deformation ought to have occurred, e.g. between the Helags nappe and the Särv nappe. In the west the zone of the overthrust has obviously been masked by an intense regional recrystallization (Pl. XII b).

Thus there seems to have occurred a regional post-tectonic recrystallization which generally shows an increasing intensity towards the northern and western parts of the area.

GEOLOGICAL EVOLUTION

The *sedimentary rocks of the Särv nappe* have been deposited upon pre-Cambrian bedrock somewhere to the west of the Riksgränsen anticline. The bedrock of the sediments seems to have consisted of light-coloured granites, together with a pre-Cambrian complex containing dark migmatitic gneisses and agmatitic breccias of various eruptive rocks which have suffered a pre-Caledonian deformation. The rocks of this complex, named the Tännäs complex after its type locality, are remarkably rich in potassium, and may be compared with the rocks of the Bergen-Jotun group. The light-coloured granites, rich in microcline, which have been encountered in the Särv nappe, exhibit similarities with granites of mainly western extension. Thus, the granite which occurs in Mt. Röstberget resembles very much the granite in the Riksgränsen anticline.

Upon this pre-Cambrian bedrock have been deposited some types of fragmental rocks, the Grönstack complex, consisting of local breccias or conglomerates with a greywacke-like matrix. In spite of the fact that no direct contacts have been observed it seems, nevertheless, probable that near above this rock a banded grey shale was deposited, containing quartzitic intercalations and lenses of limestone. Considering the association of these types of rock, viz. breccias and conglomerates with greywacke-like matrix, and a banded shale, lithologically rather similar to certain facies developments of the Biri series, the possibility of eventually glacigenous origin of some of these sediments should not be overlooked.

Anyhow, after the formation of the Grönstack rocks a great change in the condition of transport of the clastic material did occur. The material in these basal fragmental rocks is mainly derived from basic eruptive rocks. After the banded shale only sandstones have been deposited. The abundance of pale-blue quartz and pink or greyish blue feldspar indicates that the clastic material can have been supplied by microcline granites of the type encountered in the southern as well as in the northern part of the Särv nappe.

East of the area where the sequences of the Särv nappe proper were formed, the Ulvberg complex, which contains white quartzites and dolomitic marble, was deposited. The bedrock of this complex seems to have been a reddish granite similar to the Rätan granite.

In the northern part of the Särv nappe, the series of the quartzites is overlain by a series of greywacke-like dark sandstones and siltstones. The absence of

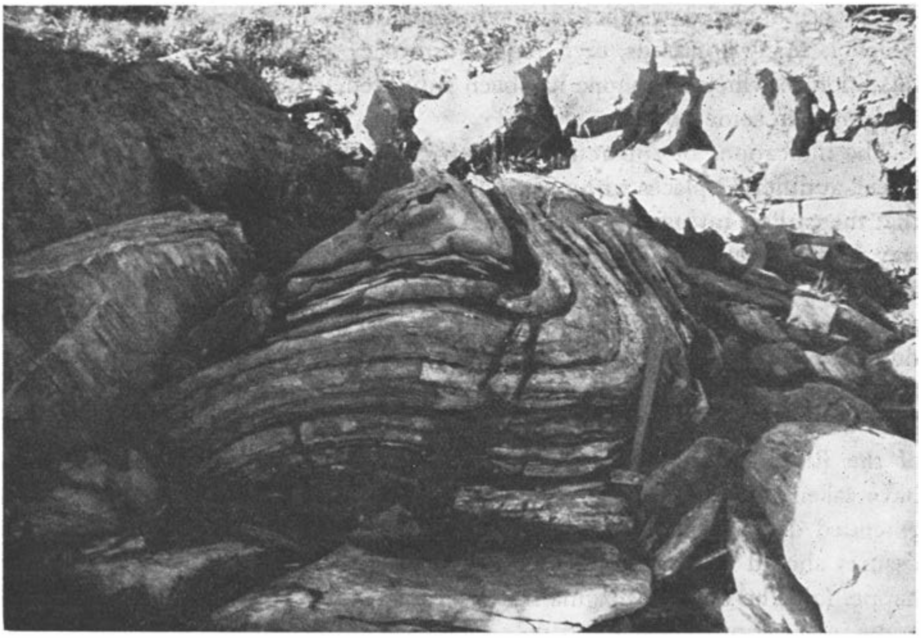

Fig. 26. Probably pre-intrusive folding in dolomite-marble, cut by an intrusion of Ottfjäll dolerite. R. Biskopsån at the Riksgränsen anticline (C1). Axis of folding dips N/20°.

fossils makes the dating of these rocks uncertain. The tectonic disturbances observed at the boundary between the quartzite sequence and the greywacke sequence are too local phenomena to be interpreted as indications of an overthrust of the latter sequence over the former. As long as no more stringent proofs of a real, and in this case pre-intrusive, overthrust are available, the transition between the two series of rocks has to be interpreted as a primary sedimentary deposition of the greywacke series upon the quartzite series.

At present an opinion about the age of the non-fossiliferous sedimentary rocks can be based upon comparison with analogous series of rocks only. From the tectonic point of view A. E. TÖRNEBOHM (1896) connects the southern part of the Särv nappe with the Kvitvola nappe in Ytre Rendal, Norway (overthrusts of the first order). He seems to have based his opinion upon observation of petrographical similarities between rocks in these isolated nappes. The rocks in this region were described by G. HOLMSEN and CHR. OFTEDAHL (1956).

The intrusion of the *Ottfjäll dolerite* has taken place at a time when the greatest part of the Särv nappe was situated west of the Riksgränsen anticline. At this time the sedimentary rocks of the nappe occupied a mainly horizontal position, and were exposed to a tension in an east-westerly direction, which produced a system of sub-vertical tension joints in the tectonic *bc*-plane with a N–S strike. This was the system of joints into which the dolerite magma was subsequently intruded.

The accumulated width of the subparallel dikes of dolerite, the number of which is many hundreds, each 10–100 m broad, gives a rough measure of the dilatation to which the nappe has been subjected, a dilatation which must have amounted to some 20 km.

The indications of apparently pre-intrusive deformations which thus can be traced at different places within the nappe might depend upon the circumstance that the earliest intrusions were located to the inner parts of the nappe. In this way the more peripheral parts have been subjected to syn-intrusive deformation, and these deformed parts have later been traversed in their turn by dikes of dolerite. In the foregoing, observations of several generations of dikes of Ottfjäll dolerite have been described.

In the rock floor beneath the Särv nappe, feeders of Ottfjäll dolerite have been observed neither in the mylonite nappe nor in the granite and porphyry of the Riksgränsen anticline. The emplacement of the dolerite magma can have taken place in either of two ways: (1) through a system of fissures which extended downwards into the primary base of the nappe. Such a system of fissures should thus correspond exactly with the system of dikes in the Särv nappe. (2) The dolerite magma has been distributed to the numerous tension joints in the nappe via sills parallel to the base of the nappe. These sills in turn received the magma from feeders in the bedrock. Structural arguments contradict the former alternative. At the time of the deformation, a great difference in competence between the sedimentary rocks and the crystalline basement must have existed. Even on the improbable assumption that the deforming external forces might have been of comparable intensity in the nappe and in its early basement, the formation of fissures can hardly have been of similar development in both types of bedrock.

The occurrence of sills at the base of the Särv nappe bears direct witness in favour of the second alternative. Thus at Biskopsån (Fig. 13), the Ottfjäll dolerite has been intruded in the shape of a series of sills which, together with the sequence of strata, have been steeply raised and even overturned towards the south by a late folding according to a *B*-axis in about WNW. Similar sills ought to have occurred also in other parts of the bottom zone of the Särv nappe, although they have been deformed into mylonitic schists in connection with the overthrusts.

The uniformity of the primary orientation of the dolerite dikes suggests a corresponding uniformity of the deformation to which the joints are due. This early formation of tension joints seems to be connected with a moderate regional deformation with an axis of symmetry striking from north to south. The Ottfjäll dolerite seems to have been intruded during a primorogenic intrusive phase which has preceded the Caledonian thrust movements.

Overthrusts.—The present position of the Särv nappe is the result of a long-range overthrust in an easterly direction which has displaced the nappe over at least 50 km from its original position. This is the shortest way by which the pre-Cambrian rocks dragged along in the nappe could be found somewhere near their presumed original position. The length of the overthrust is probably greater than this minimum value, and can have reached an order of magnitude of 100 km. An overthrust of this magnitude would imply that a sheet with a thickness equalling roughly 0.5% of its width would have been moved across an uneven base over a distance exceeding the width of the slice. Considering the plasticity of the material, when units of this size are being displaced, it is hardly imaginable that such a thin nappe could be moved so far solely by pressure from behind, without suffering more profound destruction of its structure. Yet, in the Särv nappe, the original structure is on the whole well preserved, existing disturbances being of minor size and of local occurrence. If a force acting from the rear is replaced by forces uniformly distributed all over the nappe, the tendencies towards a deformation of the nappe ought to be considerably lessened. Gravitation is the only force which could act in this way (cf. R. W. v. BEMMELEN, 1950).

In connection with the overthrust of the Särv nappe a strong mylonitization has taken place in its bottom zone and in the underlying crystalline rocks. In this way a whole nappe of mylonitic rocks with changing crystalloblastic structures has developed underneath the Särv nappe.

After the first great overthrust of the Särv nappe, the Helags nappe has become thrust over the former. This chronological order of the thrust movements is proved by the fact that several intrusive rocks, among them a trondhjemitic rock, appear in the Helags nappe, but are missing in the underlying Särv nappe. In connection with this overthrust, a strong deformation with shearing of dolerite dikes, schistosity, and mylonitization has taken place in the uppermost part of the Särv nappe.

In the course of the overthrusts the dolerite dikes in the Särv nappe were overturned in different directions, mainly eastwards, and were deformed locally. In this connection also entire blocks of the nappe were rotated around vertical axes. Such a rotation has taken place at Vålådalen (A4), where the resistance of the Mullfjället anticline caused an anti-clockwise rotation of a northern block of the nappe of about 70° in the horizontal plane. Another rotation in the nappe has occurred between the northern and southern parts of the nappe east of L. Storsjö (D5). Also here the resistance of an anticlinal part of the basement seems to have caused an anti-clockwise rotation with the result that the vertical dikes of dolerite now strike from east to west (Fig. 24).

It is possible that this sudden resistance during the final phase of the overthrusts, which has resulted in the rotation of blocks in the Särv nappe, can have been produced by a late horst-like upheaval of the anticlines.

A final deformation with a *B*-direction in WNW is clearly evident in the bedrock and appears also on the geological map. This deformation implied, *inter alia*, a relative movement between a northern block, including the northern Särv nappe, and a southern block, including the southern Särv nappe. Between these two blocks the boundary passes in a west-north-west direction from L. Storsjön (D4) to L. Sylsjön (Cl); the southern Särv nappe has been pushed towards the east and east-north-east relatively to the northern Särv nappe. In connection with this movement, the Helags nappe and the bedrock between the parts of the Särv nappe have been deformed. At the same time, overturns of the planes of folding and the dikes of dolerite have occurred towards the north-east in the eastern parts of the southern Särv nappe, and towards the south-south-west at L. Sylsjön (cf. Fig. 24 and Pl. III). Along the continuation of this line of deformation towards WNW, overturns of planes of folding in drag folds have been observed by K. Ø. BRYN (1959) near Öifjeld. It may be this late deformation with a *B*-axis in WNW and compression in the direction NE–SW which has been reproduced in the augengneiss and in the basal zone of the Särv nappe, and which has caused the folding together of the tectonic units west of Tännäs.

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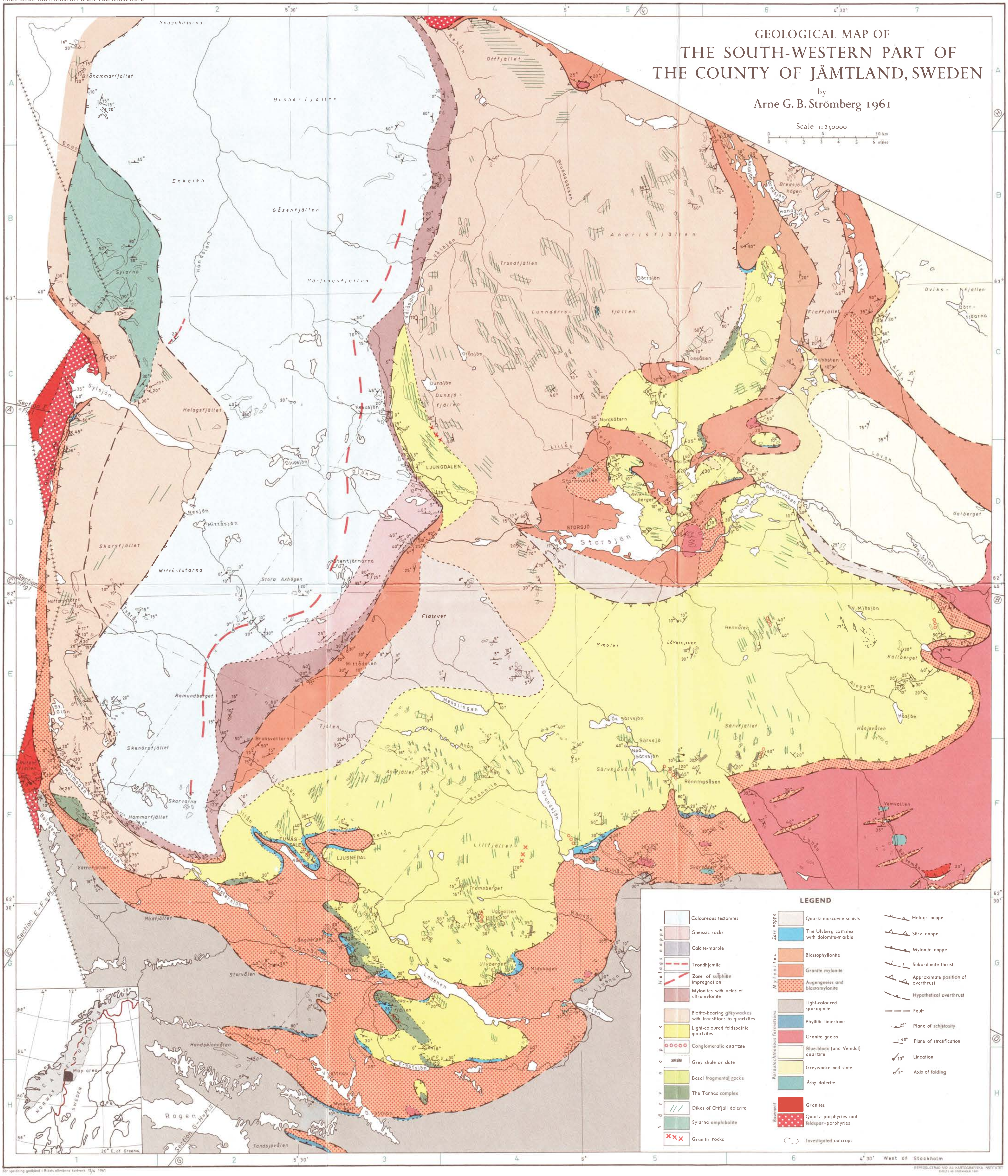
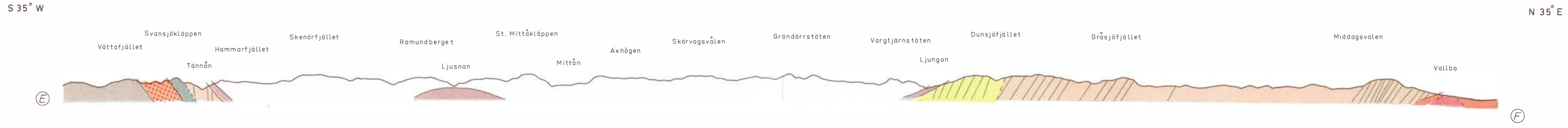
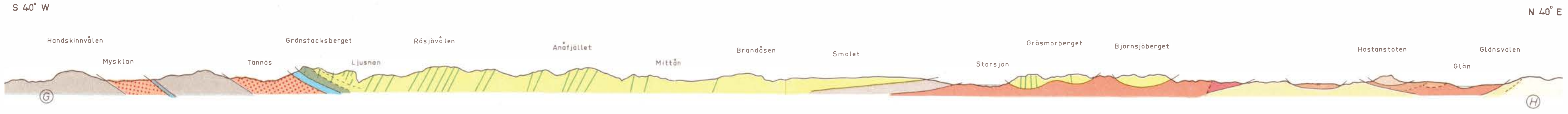


För spridning godkänd i Rikets allmänna kartverk 12/4 1961
REPRODUCERAD VID AB KARTOGRAFISKA INSTITUTET
ESSELTE AB STOCKHOLM 1961

LONGITUDINAL SECTIONS E–F AND G–H ACROSS THE SOUTH-WESTERN PART OF THE COUNTY OF JÄMTLAND, SWEDEN

By Arne G.B. Strömberg 1961





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0 5 10 km
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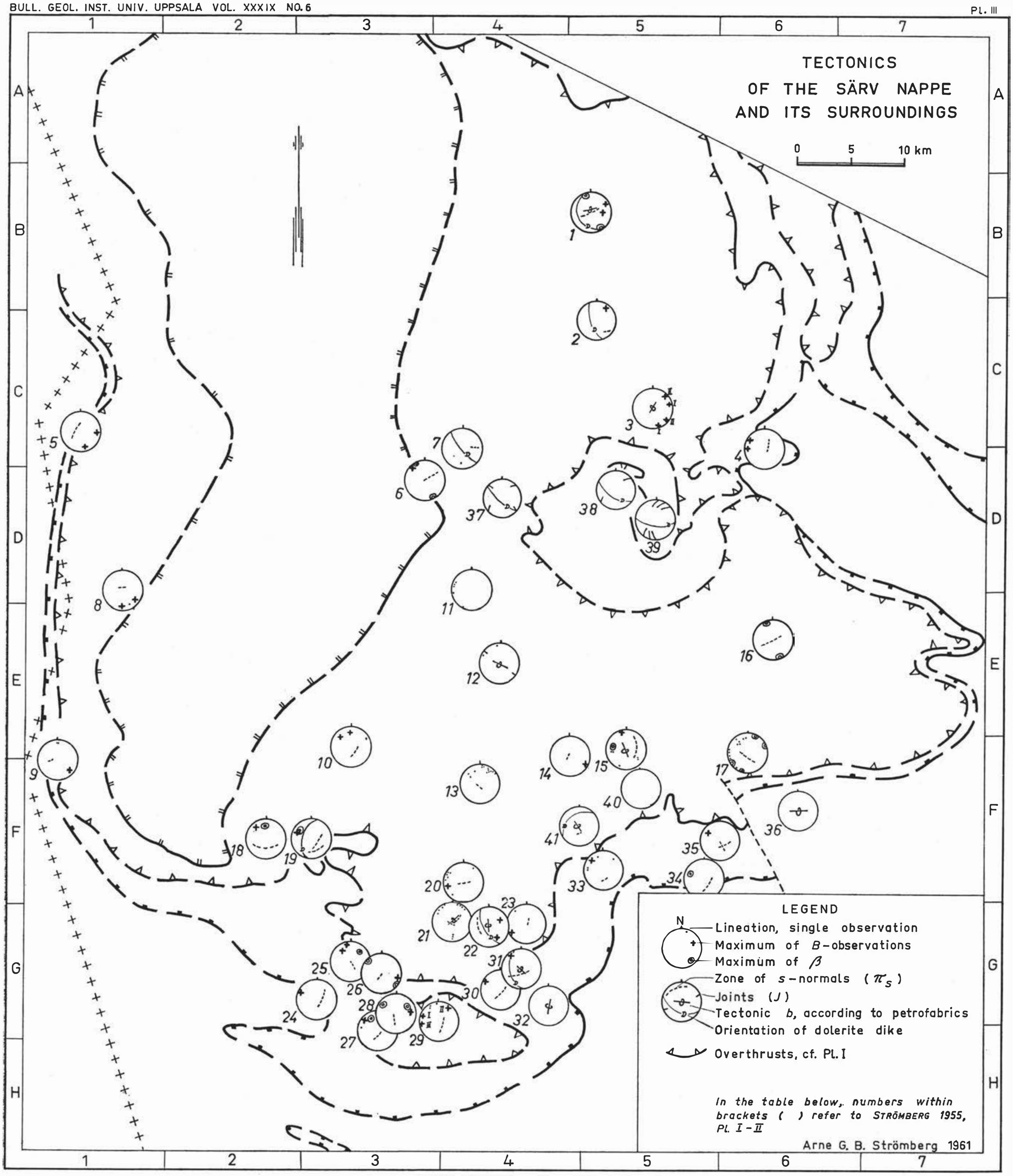


		π_s	B	J
1)	Pl. IV 19-21, 23	45	28	
2)		20	19	
3)	Pl. IV 5-6, 25	8	22	
4)		6	11	
5)	Fig. 13	13	18	
6)		17	15	
7)		9	1	
8)	Pl. IV 37a	6	13	
9)	Pl. IV 37b	15	16	
10)	Pl. IV 37c	17	19	
11)	Pl. IV 37d	8	5	
12)	Pl. IV 37e 7-8	5	2	
13)		25	13	
14)		11	6	
15)	(V, IX-X)	26	20	
16)		7		
17)	(VI)	56	11	
18)		14	6	
19)	Pl. IV 35	20	7	
20)		57	25	
21)	Pl. IV 13-15	21	11	
22)	Pl. IV 17, 18, 22	20	18	
23)		16	7	
24)		17	14	
25)	Pl. IV 29	28	20	
26)	Pl. IV 30	7	10	
27)	Pl. IV 31	17	17	
28)	Pl. IV 32	11	9	
29)	Pl. IV 24	18	49	
30)	Pl. IV 33 (VII-	126	20	
31)	Pl. IV 34 VIII)	150	40	
32)	Pl. IV 2-4			
33)	Fig. 22 III	54	8	
34)	Fig. 22 I	55		
35)	Fig. 22 II	36	16	
36)	Pl. IV 2			
37)	Pl. IV 27			10
38)	Pl. IV 28			8
39)	Pl. IV 26			7
40)	Pl. IV 9-12			
41)	(XI, XII)			

Petrofabrics

(In *quartz* the preferred orientation of optic axes and in *mica* normals to [001]-flakes have been determined. The contours are given in % per 1% area. Schmidt net, lower hemisphere.)

1. Granite gneiss from Mt. Fröåsen (F6), *quartz*, 250 grains 4, 3, 2, 1, ¼, 0.

Augengneiss from Midskogsbygget (G4). **2.** *Apatite*, optic axes of 97 grains: 5, 4, 3, 2, ½, 0. **3.** *Quartz*, 133 grains: 8, 5, 3, 2, ½, 0. **4.** *Biotite*, 112 flakes: 10, 7, 4, 2, ½, 0.

Recrystallized quartzitic schist, R. Västerån at Tossåsen road (C5). **5.** *Quartz*, 155 grains: 5, 4, 3, 2, 1, 0. **6.** *Muscovite*, 61 flakes: 25, 15, 10, 6, 3, 0.

Quartz-muscovite-schist, Flatruet (E4). **7.** *Quarts*, 248 grains: 6, 5, 4, 3, 2, 1, 0. **8.** *Muscovite*, 171 flakes, 10, 8, 4, 2, 1, ½, 0.

Feldspathic quartzite, Mt. Svartuggen (F5). 9–10: fine-grained part, 11–12: coarse-grained part of the same thin sect. **9.** *Quartz*, 300 grains: 4, 3, 2, 1 ⅓, 0. **10.** *Muscovite*, 154 flakes: 20, 10, 5, 1, 0. **11.** Quartz, 280 grains: 4, 3, 2, 1 ⅓, 0. **12.** *Muscovite*, 100 flakes: 20, 15, 10, 5, 1, 0.

Schistose feldspathic quartzite, L. Ränntjärnarna (G4). **13.** *Quartz*, 215 grains: 3, 2, 1, ½, 0. **14.** *Muscovite*, 104 flakes: 10, 6, 3, 2, 1, 0. **15.** *Langquartz*, 88 grains: 3, 2, 1, ½, 0. **16.** *Sillimanite* in quartz, crystallographic *c*-axes of 120 needles: 10, 6, 3, 2, 1, 0.

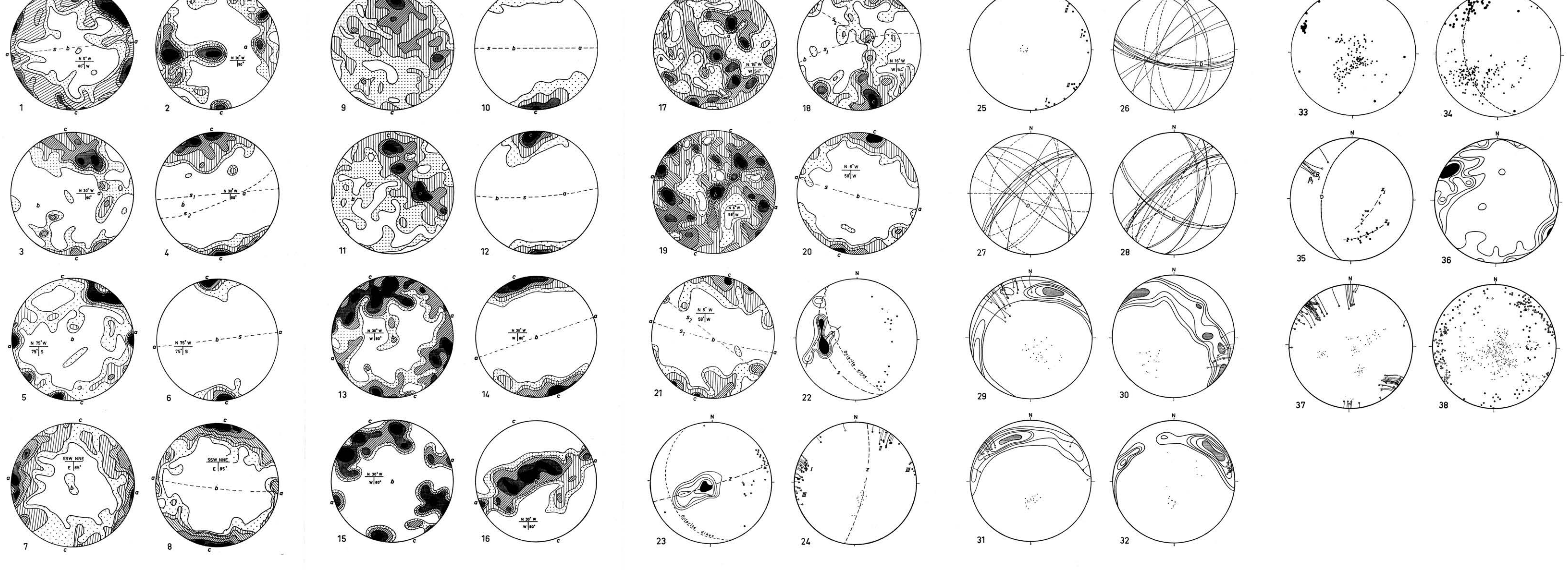
Quartzitic sandstone between dolerite dikes, Mt. Gråvålen (G4). **17.** *Quartz*, 372 grains: 2½, 2, 1½, 1, ¼, 0. **18.** *Muscovite*, 90 flakes: 6, 4, 3, 2, 1, 0.

Biotite-bearing sandstone near a dolerite dike, R. Bruddabäcken (B5). **19.** *Quartz*, 404 grains: 2½, 2, 1½, 1, ½, ¼, 0. **20.** *Muscovite*, 202 flakes: 10, 5, 2, 1, 0. **21.** *Biotite*, 103 flakes: 10, 5, 2, 1, 0.

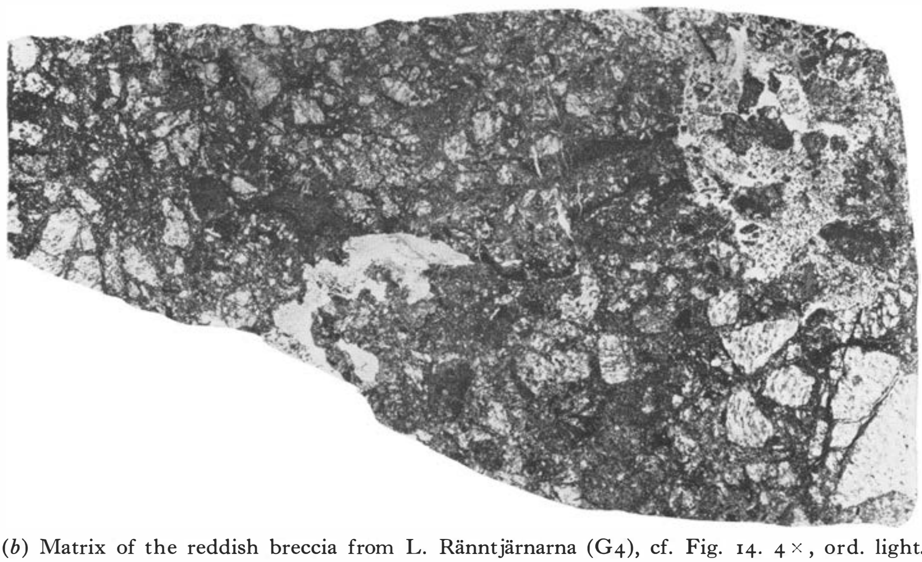
Megascopic fabric

(Schmidt net, lower hemisphere, contours in % per 1% area. Z = zone of *s*-normals.)

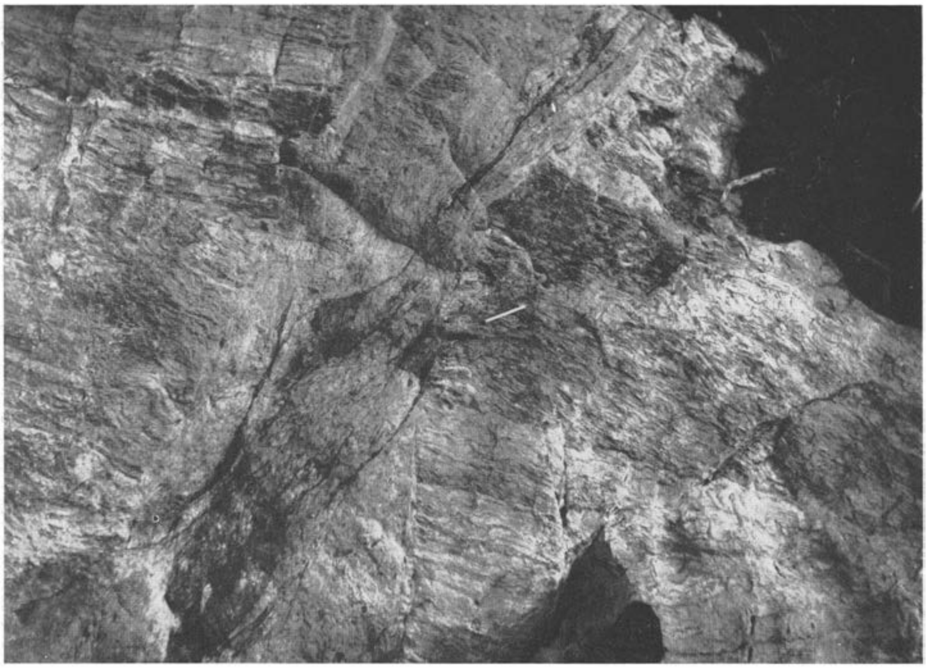
22. Quartzitic feldspar sandstone between two parallel dolerite dikes. Mt. Gråvålen (G4). 17 *B* (circles), 22 π_s: 25, 15, 10, 5, 0. **23.** Biotite-bearing, recrystallized sandstone, R. Bruddabäcken (B5). 28 *B* (circles), 45 π_s: 20, 10, 6, 4, 2, 0. **24.** Quartzitic schists, Morslättvallen (G3–4). 18 π_s (dots), 49 *B* (circles). **25.** Recrystallized, quartzitic schists, R. Västerån, Tossåsen Road (C5), 8 π_s (dots). 22 *B* (circles). **26.** Joints, 14 in dolerite (full lines), 4 in wall-rock quartzite (broken lines). D = dolerite dikes. Mt. Åvikberget, 5 km NE of Storsjö Village (D5). **27.** Joints, 10 in dolerite and 7 in quartzite. Mt. Långmyrberget (D4), 4 km E of Ljungdalen Village. **28.** Joints, 14 in dolerite and 6 in quartzite. At the highway near R. Storån, 4 km NE of Storsjö Village (D5). **29.** Western part of Mt. Tännäsberget (G3). 28 π_s (dots), 20 *B* (circles), 280 β: 10, 8, 5, 3, 1, 0. **30.** Eastern part of Mt. Tännäsberget (G3). 17 π_s (dots), 10 *B* (circles), 137 β: 6, 4, 2, 1, 0. **31.** Quartzitic schists below the slope of Mt. Western Kröksberget (G3). 17 π_s (dots), 17 *B* (circles), 126 β: 15, 10, 7, 4, 1, 0. **32.** Greenschists in the slope of Mt. Western Kröksberget (G3). 11 π_s (dots), 9 *B* (circles), 44 β: 15, 10, 7, 4, 1, 0. **33.** Quartzitic schists lowermost in the slope of Mt. Ulvberget, Midskogen Village (G4). 126 π_s (small dots), 20 *B* (large dots). **34.** Greenschists at the sheared dolerite dikes in the slope of Mt. Ulvberget, Midskogen Village (G4). D = orientation of dolerite dikes. 150 π_s (small dots), 40 *B* (large dots). **35.** R. Ljungan at highway bridge, NE of Funäsdalen. 6 π_s (crosses, Z_I), 7 *B* (circles). D = orientation of dolerite dikes. 14 π_s (circles with dots, Z_{II}) at the sheared dolerites in the river above the bridge. **36.** Synoptic diagram of 119 scattered *B*-observations in the augengneiss and the basal zone of the Särv nappe. Contours: 10, 8, 6, 4, 2, 0. **37.** Synoptic diagram of 70 *B* (circles) and 68 π_s (dots) from the mylonite zone below the Helags nappe and west of the main part of the Särv nappe, including Pl. III, *8–12*. **38.** Synoptic diagram of 292 π_s (small dots) and 197 *B* (circles) in the Särv nappe.



(*a*) Dike of dolerite cutting across rocks of the Tännäs complex, affected by pre-Caledonian deformation. Southern face of the eastern part of Mt. Tännäsberget (G3).

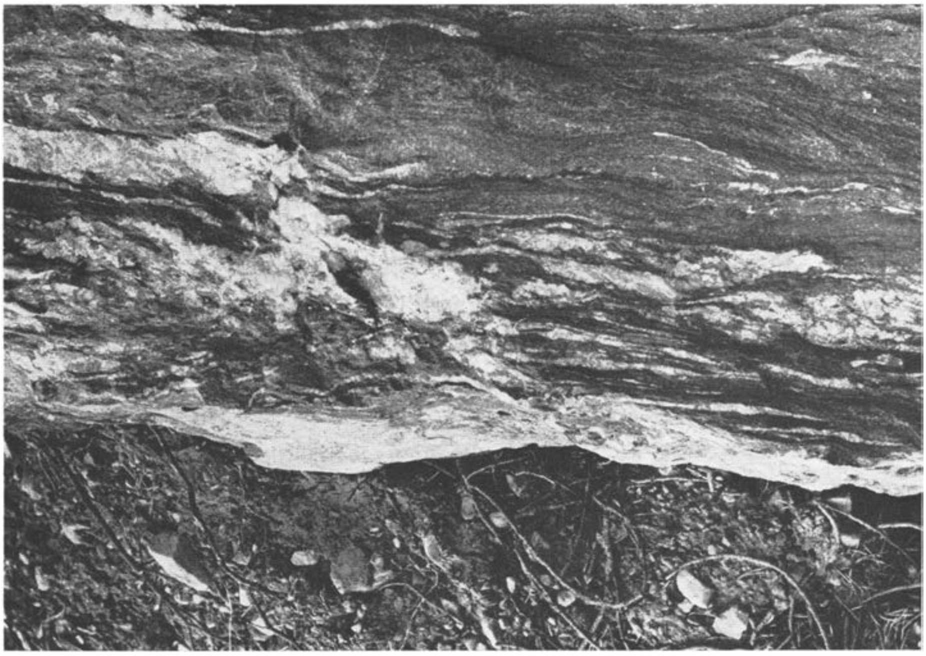


(*b*) Matrix of the reddish breccia from L. Ränntjärnarna (G4), cf. Fig. 14. 4 ×, ord. light.



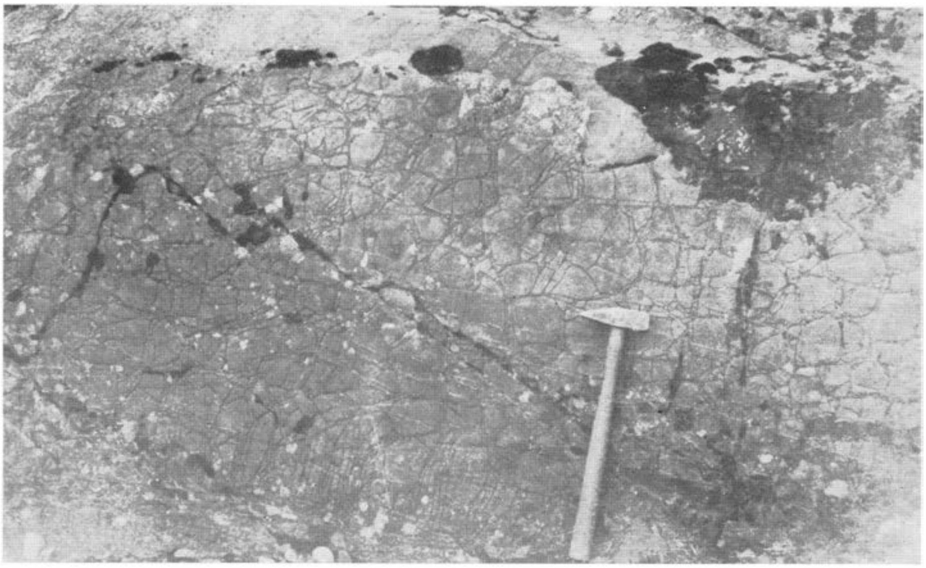
(*a*) Granite mylonite, cut by a vein of subsequently mobilized mylonitic material. Rock surface cleaned by brushing with 10 % sodium hydroxide solution. By the forest road to Utansjön (D5).

(*b*) Granite mylonite with brecciated fragments of amphibolite. Highway near the forest road to Tossåsen Village (D5).

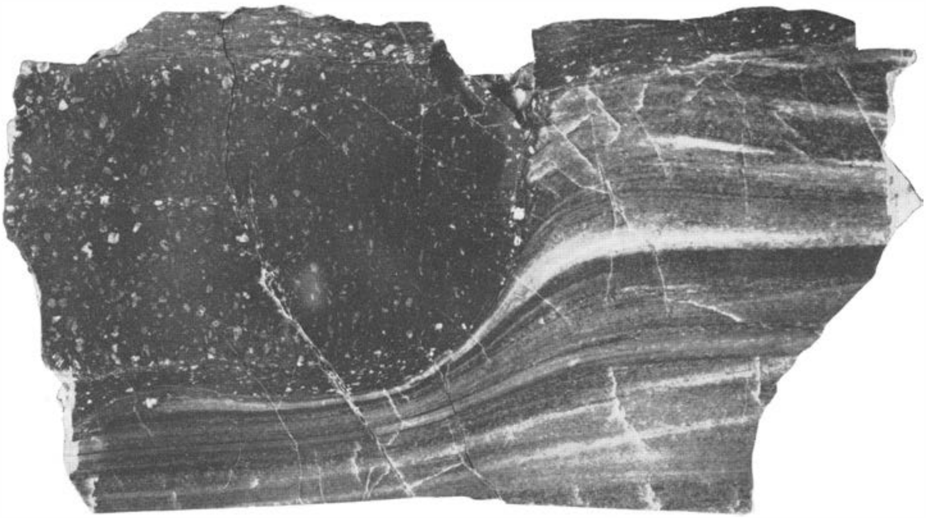


(*a*) Rocks of the mylonite nappe with coarse-grained secretions containing quartz, potassium feldspar, siderite, and scaly hematite. By the forest road to Falvik (D5–6).

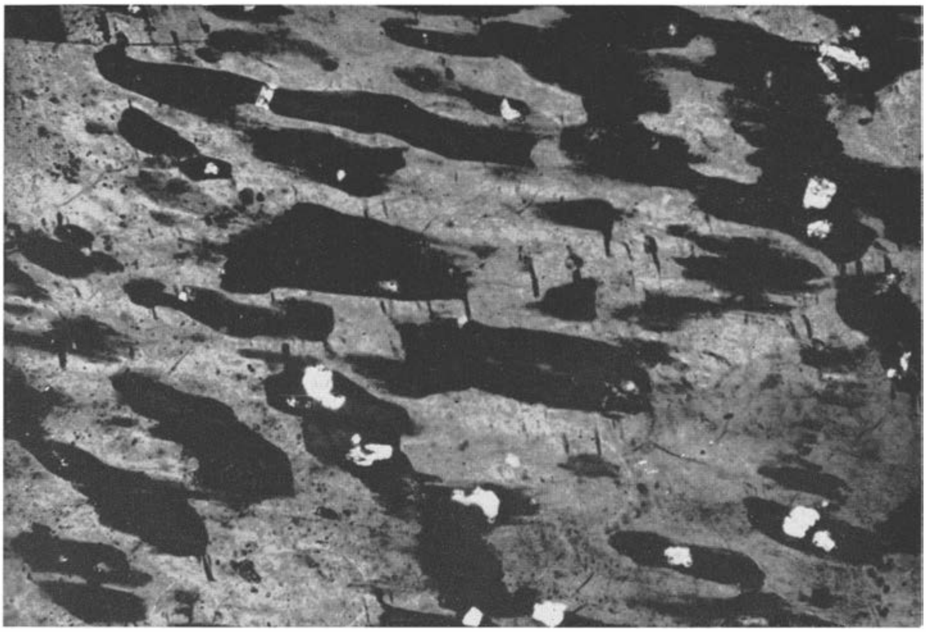
(*b*) Coarse-grained mottles containing quartz and calcite in the mylonites below the Helags nappe at Ljungdalen Village (D3).



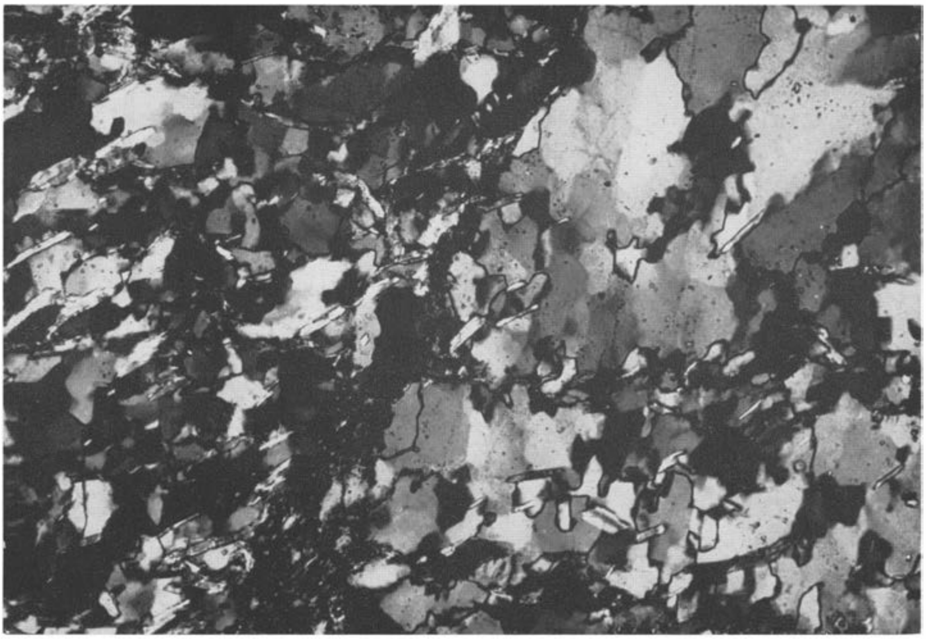
(*a*) Contraction cracks in an early intrusion of Ottfjäll dolerite. This dike forms the margin of a later and broader dolerite dike. Southern slope of Mt. Blåfjället, peak 1358 (C4).



(*b*) Margin of an Ottfjäll dolerite dike, partly intersecting and partly conforming with the banding in the wallrock quartzite. Post-intrusively the dolerite body has been moved to the right, squeezing the quartzite. Highway bridge at R. Storån (D5). About half natural size. Photo: A. Sagerholm.



(*a*) Perthite porphyroblast with microcline and symplectitically enclosed albite (dark). The perthite grain has margins of pure microcline and seems to have formed by replacement. Both feldspars are corroded by veinlets of quartz. Augengneiss, Mt. Busjöfjället (C7). +nic., × 120.

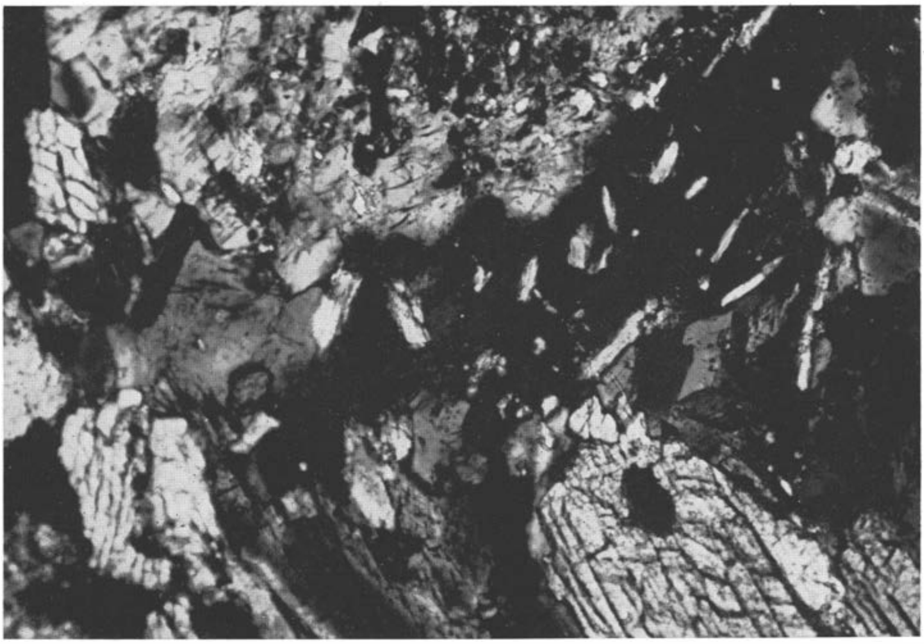


(*b*) Coarse-grained and fine-grained feldspathic quartzites, brought into direct contact by a small-scale thrust, and intersected by an *s*-plane of a later mica crystallization, cf. fabric diagr. 9–12. Microcline either forms porphyroblasts or occurs as grains elongated conformably to the late *s*-plane. Mt. Svartuggen (F5). +nic., × 120.

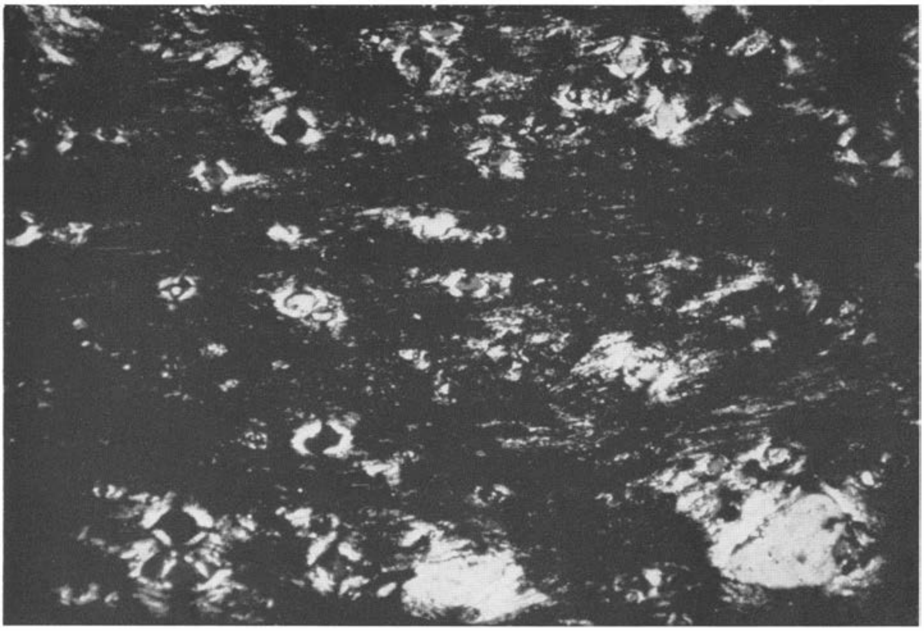
(*a*) Syncrystalline deformation in a muscovite-rich schist. Postdeformatively crystallized quartz and potassium feldspar fill the interstices between muscovite bundles. Mt. Vebuåsen, N of Ljusnedal Village (F3). +nic., × 120.

(*b*) Blastophyllonite, consisting of quartz, albite, muscovite, biotite, and epidote, which are segregated in undulating layers. The crystallization of the minerals seems mainly to be postdeformative. Outcrop as Fig. 9 (B6). 1 nic., × 120.

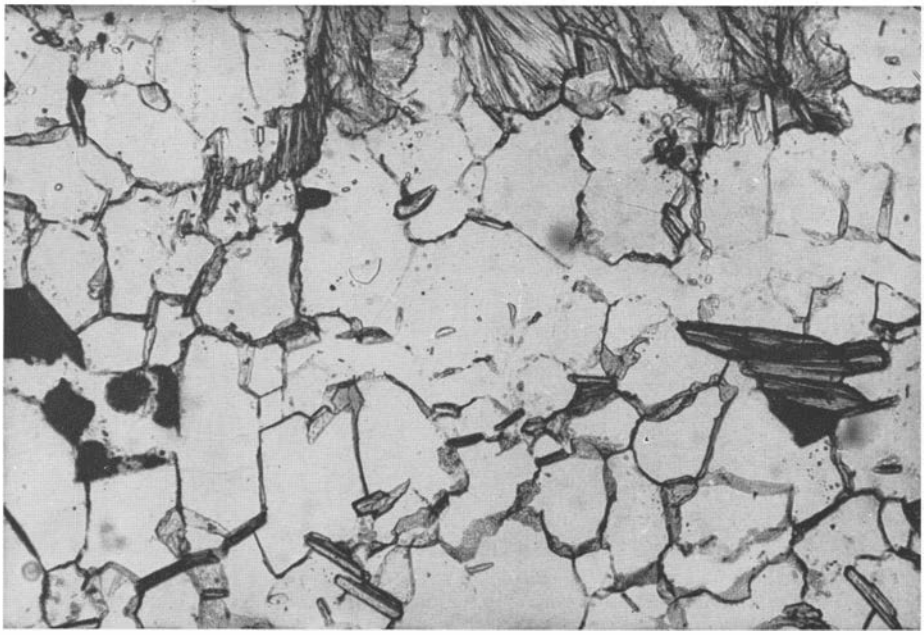
(*a*) Felsite exhibiting a streaky texture due to the parallel arrangement of alternating quartz and feldspar bands. To the left is a fragment of the rock rotated about 70°. The Tännäs Complex, western slope of Mt. Östra Kröksfjället (G3). +nic., × 120.



(*b*) Labradorite, studded with grains of epidote and surrounded by a zone of albite containing broken and dislocated fragments of amphibole below. The Sylarna amphibolite, western slope of Mt. Sylskafthuvudet (B1). +nic., × 120.



(*a*) Chloritic mylonite containing small porphyroblasts of albite-oligoclase. The chlorite shows homogenous extinction and pleochroism, and is preferrably orientated with (001) in the tectonic *ab*-plane. "Interference figures" are produced by the arrangement of chlorite conformably to the outlines of the rounded feldspar porphyroblasts (the surrounding chloritic mass in extinction position). The mylonite, which contains magnetite and limonite, occurs as violet streaks at the sheared surface below the westernmost dolerite dike in Mt. Ulvberget (G4), (cf. S. 55, fig. 17). +nic., ×150.



(*b*) Postdeformative crystallization in muscovite-bearing quartzite of the Särv nappe in the western (Riksgränsen) limb. Western slope of Mt. Blåhammaren (A1). 1 nic., ×120.