NORGES GEOLOGISKE
UNDERSØKELSE

# Studies in the Trondheim Region, Central Norwegian Caledonides II

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## CONTENTS

Page

# I. Geology of the Meråker area

## Introduction

by Fr. Chr. Wolff.

### Aim of study

After several years of geological reconnaissance mapping in the northern part of the Trondheim region, the present writer found the Meråker district to be a rather promising area for a more comprehensive study.

Concequently a very detailed examination has been carried out of a well exposed area wherein the degree of deformation is so low that many of the primary structures have been preserved. The idea has been to apply the results of this work to a more extensive area in an attempt to solve the main stratigraphical and tectonic problems of the entire eastern part of the Trondheim region.

### Location of the area investigated

The area studied in detail is located between lat. 63°20′ and 63°35′ N and long. 1°O′ and 1°30′ east of Oslo. The area is bounded on its east side by the Norwegian — Swedish border and in the west by the valleys of Funn and Lilleåen. The southern and northern boundaries are the southern and northern limits of the map-sheet Meråker (Rektangel 47 D). Fig. 1 shows the location of the area, the centre of which lies about 90 km east of Trondheim.

The southern part of the area is traversed across strike by the new E 75 highway from Stjørdal to Sweden, along which a series of roadcuts displaying most of the rock types and sedimentary and tectonic features of the area may be observed. The western and northern parts are accessible by the minor road to Sulåmo, about 20 km north of Meråker village, whereas the north-eastern part of the area can be reached only by foot.

### Planning and organization of the work

During the field season 1964 the present writer made a detailed study of the roadcuts from Meråker to the Swedish border. Some reconnaissance mapping

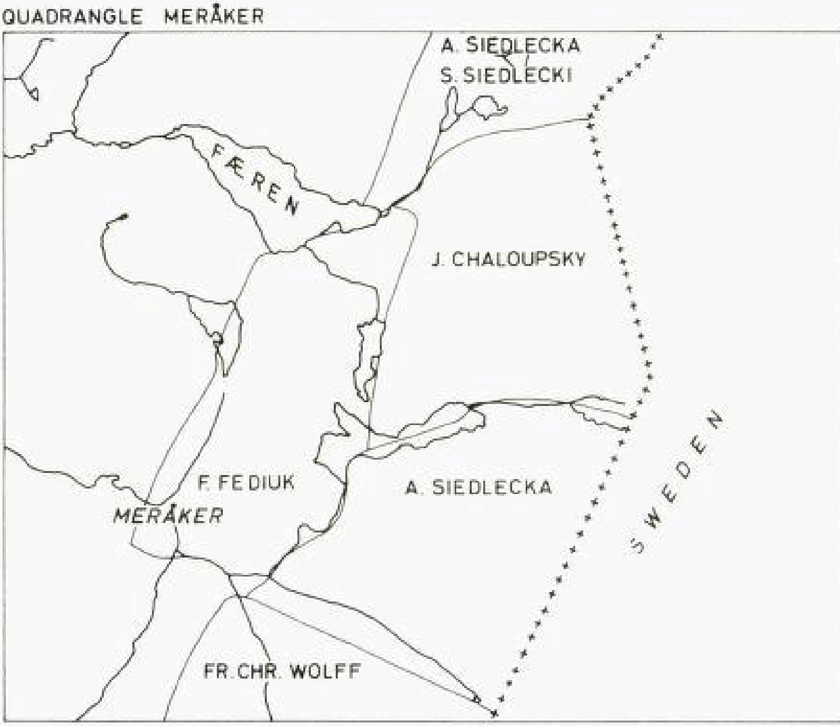


Fig. 1. Map showing the areas mapped by the different members of the team.
*Kart som viser områdene som er kartlagt av de forskjellige medlemmene av gruppen.*

to the north of the highway was also carried out. Along the main profile it was possible to distinguish between series of different sedimentary origin and it was obvious that a more detailed mapping programme would prove fruitful.

During a journey to Czechoslovakia in September—October of the same year the idea arose to invite some of my Czech colleagues to participate in a mapping team in the Meråker area in the summer of 1965. The invitations were given to Dr. Josef Chaloupsky from the Geological Survey of Czechoslovakia and to Dr. Ferry Fediuk from the Charles University of Prague, both of whom kindly accepted, and with the amiable co-operation of Dr. J. Svoboda, director of the Geological Survey of Czechoslovakia, and Dr. ing. Jan Gruntorát, Dean of the Faculty of Science at Charles University, their participation in the work was made possible.

Later, in the winter 1964—65 during the period of more detailed planning of the mapping programme, another two geologists were involved in the team, namely Anna Siedlecka from the Academy of Mining and Metallurgy, University of Cracow, Poland, who happened to stay in Norway for a period, and Dr. David Roberts from Bedford College, University of London, U.K., who is now an associate member of the staff of NGU.

The locations of the areas mapped by the different members of the group are indicated on the map, fig. 1. Dr. Roberts, while mainly occupied with tectonic studies along Stjørdalen between Hegra and the Swedish border, mapped an area to the north of the valley of Stjørdalen. Since the present writer was the responsible leader of the project, a position involving a lot of administrative work during the field season and also, from time to time, collaboration with his co-workers in the field, the area mapped by him was therefore rather small.

### Geological setting

The area investigated is located in the northeastern part of the Trondheim region, Central Norwegian Caledonides. Sedimentary and volcanic rock series found in the area are closely related to rock series in other parts of the Trondheim region. The profile from Stjørdal to the Swedish border near Storlien gives an excellent picture of the general geology across this part og the Caledonian mountain chain, all rock series from the oldest to the youngest being represented. In the east the Cambro-Silurian metasediments overlie a thrust plane east of which rocks of Eocambrian age are found.

### Previous investigations

The first reported geological investigations in the Meråker district are those of K. M. Hauan from the years 1867 and 1870. His beautifully handwritten diaries illustrated with drawings of high standard make for worth-while reading The first printed paper with information on the lithologies and the stratigraphical positions of the different rock series from this area is that of Kjerulf (1883). Kjerulf's observations were supplemented by microscopic descriptions of the rocks by Reusch (1883). Later Reusch (1890) returned to the area and described the profile along the Meråker railway-track. His observations were much more precise than those of Kjerulf (1883) and were accompanied by new petrological studies. Reusch (1890) also commented on the sedimentary environments and emphasized that the facies changes are a great obstacle to the application of the lithostratigraphic method in stratigraphical correlation. An important paper by Getz (1890) shows that the graptolite fauna from Kjølhaugene is of Silurian age. These graptolites are the only fossils of Silurian age in the entire Trondheim region. On the basis of this graptolite fauna, Törnebohm (1896) established the stratigraphical positions of the so-called "Meråker,gruppen", and "Sul skiffer-gruppen" and correlated them with rocks in the western part of the Trondheim region. Carstens (1920) also worked in the vicinity of Meråker, confirming the observations of earlier geologists and stratigraphically correlating the Meråker profile with other areas in the Trondheim region.

In later papers (e.g. Kiær, 1932, Strand 1960 and Wolff 1964) the geology of the Meråker district based on the observations of earlier writers, and in Wolff's case also on his own investigations, is discussed briefly.

**Acknowledgements**

The writer is greatly indebted to the director of the Geological Survey of Norway, Professor Harald Bjørlykke, providing a carte blanche with regard to the planning and carrying out of the present project, and to Professor Chr. Oftedahl for stimulating discussions.

Most of all I would like to express my thanks to my dear colleagues, the participants of the project: Dr. Anna Siedlecka, Dr. Josef Chaloupsky, Dr. Ferry Fediuk and Dr. David Roberts for their inspiring and friendly company in the field and for the splendid manner in which they executed their various tasks.

Special thanks are also due to Meraker Smelteverk A/S, Kopperå, which provided accommodation in a central part of the area.

The responsibility for the chemical analyses occurring in the present paper rests with cand. real. Per-Reidar Graff, leader of NGU's silicate analysis laboratory; one analysis has been done by P. Povonora of the Charles University of Prague. The maps have been drawn by the NGU cartographers Beret Hemming and Astrid Lund. The manuscript has been typed by Mrs. Gunhild Anderssen.

Thin-sections have been prepared partly by the NGU preparants Tom Jacobsen and Egil Iversen, though for practical reasons some have been made in Czechoslovakia.

# Geology of the western and north-eastern part of the Meråker area

J. Chaloupský and F. Fediuk.

**Abstract**

A thick metasedimentary sequence together with a volcanic and hypabyssal intrusive complex is described. While strata dip generally westwards it can be shown that the sucession is, for the most part, inverted; the discovery of a polymict conglomerate stratigraphically above the volcanic series has helped to confirm this view. In the east, the youngest (Silurian) metasediments occupy the core of an asymmetrical syncline, overturned towards the east. Gabbro bodies, mostly sills, occur throughout all except the Silurian rocks: these gabbros are younger than similar basic rocks in the igneous complex. The regional metamorphism of the sequence has been low grade viz. greenschist facies, and is manifested by a well-defined schistosity. Locally, biotite porphyroblastesis post-dates this schistosity.

## Introduction

In the summer of 1965, the authors carried out geological investigations and mapping of an area of approx. 250 km$^2$, lying 90 km E of Trondheim. For mapping purposes aerial photographs on a scale of about 1:35 000 were used. The area is bounded in the S by the Tevla River (between Meråker and Kopperå), the Kopperå—Fjergen road and the Fjergen lake as far as Halsjøen lake, in the E by the state border to the Storsjø-lake NE of Kjølhaugen, from where northern boundary runs to the Færen-lake. In the W the area extends to the tie-line Fundsjø—Meråker. The south-western part was mapped by F. Fediuk and the eastern sector by J. Chaloupský.

The area belongs to the eastern marginal part of the Caledonides of the Trondheim region, and comprises a folded, weakly metomorphosed sedimentary sequence, accompanied by volcanic series and hypabyssal intrusives.

Fig. 2. Alternation of metabasites and quartz-keratophyres in the outcrop along the the road 2 km SSE from the lake Fundsjø.

*Veksling av metabasitter og kvartskeratofyrer i veiskjæring 2 km syd-sydøst for Fundsjøen.*

**Lithological descriptions**

In the several thousand metres thick complex several sedimentary cycles can be differentiated; each of them begins with coarse clastic deposits often with conglomerate at the base and after several repetitions of beds it ends with finer grained pelitic or silty rocks. The earliest deposits occur in the western part of the area, at the contact with the igneous complex.

The igneous complex extends along the western margin of the area, between Meråker and the Færen lake. It consist of alternating basic and acid volcanics. The former show mostly the character of metabasites, viz. greenschist, prasinites and albite-epidote-amphibolites designated by (1) in Plate I. Sporadically, a relict amygdaloidal structure was observed in these rocks. Acid volcanics correspond to quartz-keratophyres (2). The alternation of the two rock types is very rapid, and layering of the order of metres or even decimetres is not exceptional (figure 2). As the layers of both basic and acid volcanics are traceable over a distance of several hundred metres, they can hardly be interpreted

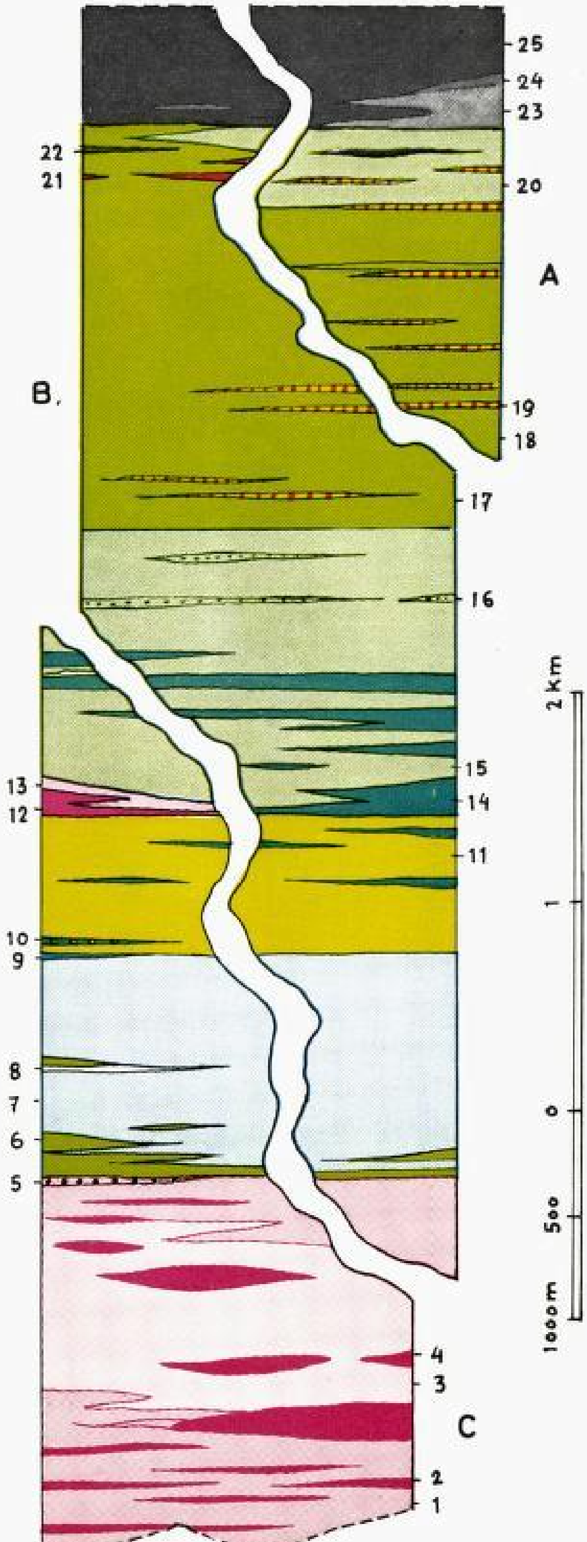


Plate I: Stratigraphical sketch of the mapped area.

*Stratigrafisk skisse av det kartlagte området.*

A. Development of the eastern part of the area.

*Utviklingen av den østlige del av feltet.*

B. Development in the central part.

*Utviklingen i den sentrale del.*

C. Development in the south-western part.

*Utviklingen i den sydvestre del.*

For explanation of numerals at the margin, see the text.

*Forklaring til tallene i margen finns i teksten.*

Fig. 3. Albite granite with graphic texture. From an outcrop 1,5 km N of Meråker railway station. (Photomicrograph by D. Hejdova, magnification X 18, crossed nicols.)

*Albitt-granitt med grafisk tekstur. Fra en blotning 1,5 km nord for Meråker jernbanestasjon. (Mikrofoto ved D. Hejdova, forstørrelse X 18, x-nicoler.)*

as lava sheets, particularly the acid types (due to the high viscosity of acid magmas), and a pyroclastic origin for at least part of the sequence should be postulated. In addition to the above-mentioned types, the igneous complex also comprises substantially coarser grained rocks that can be interpreted as the product of hypabyssal intrusions accompanying the volcanic activity. These rocks have again both acid and basic chemical characteristics. Basic rocks of this group, prevailing greatly over the aid types, are represented in those parts relatively little affected tectonically, by fine- to medium-grained albitized hornblende gabbros ranging up to diorites (3). They frequently pass into more or less schistose varieties. Scarce lenses of ultrabasic rocks (hornblendites) most probably originated by differentiation from gabbros. Acid intrusive rocks (4) are leucocratic and composed of albite and quartz (albite-granite, figure 3), At some places the penetration of acid rocks into basic ones was observed.

Above the igneous complex (structurally below) there is an approximately

Fig. 4. Lille Fundsjø conglomerate — the outcrop on the S. bank of the lake Lille Fundsjø.

*Lille Fundsjø-conglomerater — blotning på sydlig bredd av Lille Fundsjøen.*

1.000 m thick sequence of folded, grey chlorite — sericite up to biotite — sericite phyllites (7), which frequently display platy or sometimes laminar splitting. Some darker coloured varieties have an increased proportion of graphitic matter. In this lower part of the sequence, beds of graywacke-phyllite and laminated phyllitic graywacke are randomly distributed (6). Their boundary with the underlying igneous complex is sharp; S of the Færen lake it is marked by a conspicuous layer of medium- to coarse-grained polymict conglomerate, called the Lille-Fundsjø-Conglomerate (5). This stratigraphically oldest conglomerate band can be followed from the southern bank of the Faeren lake to the northern environs of Meråker. It is best developed near the Lille-Fundsjø lake (figure 4), where it attains a thickness of 20 metres. Towards the N and S of the lake, the thickness decreases to 1 metre. The conglomerate lies immediately or almost directly upon the igneous complex and much of its pebble material has been unquestionably derived from these igneous rocks. An analysis of pebbles provided the following results:

| | |
|---|---|
| Basic intrusives (gabbro) | 49 % |
| Greenschists — greenstones | 23 % |
| Quartz-keratophyre | 13 % |
| Albite-granite | 4 % |
| Graywacke | 4 % |
| Hornblendite | 3 % |
| Limestone | 3 % |
| Erlan | 1 %. |

The mean pebble size is 3—5 cm, but cobbles above 10 cm across can also be found.

At about the middle of the sequence there is another conglomerate band (8); it is not very distinct and has therefore not been well recognized. This fine-grained conglomerate with quartz pebbles is only several decimetres in thickness and tracable for a few hundred metres.

The sericite phyllites are overlain by about 700 m of grey fine-grained, often calcite-rich phyllitic graywacke or subgraywacke (11). In the northern tract of the area investigated, their upper part frequently contains beds of grey sericite phyllite. In the basal part of the sequence thin intercalations of crystalline limestone ("Brenna Limestone") and oligomictic fine- to medium-grained conglomerate, termed the "Brenna Conglomerate", may occasionally be found.

In the limestone (9) a number of quarries are present in the vicinity of Brenna village, where a lenticular, ca. 1 km long and maximum 15 m (on the average 8 m) thick layer is worked. According to DTA results, the admixture of dolomite is negligible so that the rock is almost pure calcitemarble. It is noteworthy that the marble occasionally contains fuchsite; the chromium necessary for its formation was most likely derived from the igneous complex. Near the north-western embayment of the Fjergen lake, another layer of calcitemarble was observed, the maximum thickness of which is however only 3 metres. A third, even thinner carbonate band crops out in the valley of a stream N of the Langen lake. All three carbonate lenses lie nearly at the same, but not quite identical stratigraphical level.

The Brenna Conglomerate (10) can be found in the road cutting S of the limestone quarry near Brenna. It forms several 0,5 m thick bands and contains quartz and subordinate quartzite and grey fine-grained limestone pebbles. The same layer was observed in 1964 by F. Chr. Wolff (personal communication) about 1 km further to the north. The geological position of the Brenna outcrop shows clearly that the conglomerates lie above the Brenna limestone, though not more than 30 m above this horizon.

Fig. 5. Albite granite with graphic texture in a pebble from the Kjølhaugene conglomerate. (Photomicrograph by D. Hejdova, magnification X 18, crossed nicols.)

*Albitt-granitt med grafisk tekstur i en bolle på østbredden av Fundsjøen. (Mikrofoto ved D. Hejdova, forstørrelse X 18, x-nicoler.)*

The next approximately 1.300 m thick sequence of strata is characterized by a monotonous alternation of green-grey phyllitic graywackes to subgraywackes (several metres to a few tens of metres thick) with chlorite-sericite phyllites (15). In the basal part intercalations of sericite phyllite (14) are still abundant. The upper boundary of the sequence is marked by the last thicker bed of feldspathic sandstone (16). To the SSW of Langen lake, a layer of metabasites (12), (13) resembling some varieties of the underlying igneous complex, intervenes between the above-described greenish feldspathic sandstones and the subjacent grey sandstones. This layer attains a thickness of several hundred metres in the S, near Kopperå, but thins out northwards and disappears in the proximity of the Fjergen lake. The rocks of this volcanic horizon show a massive and banded structure. In our opinion, the rocks of the former type (12) originated from diabase effusives and those of the latter type (13) were derived from tuffs.

The sequence of greenish-grey graywackes and subgraywackes is connected

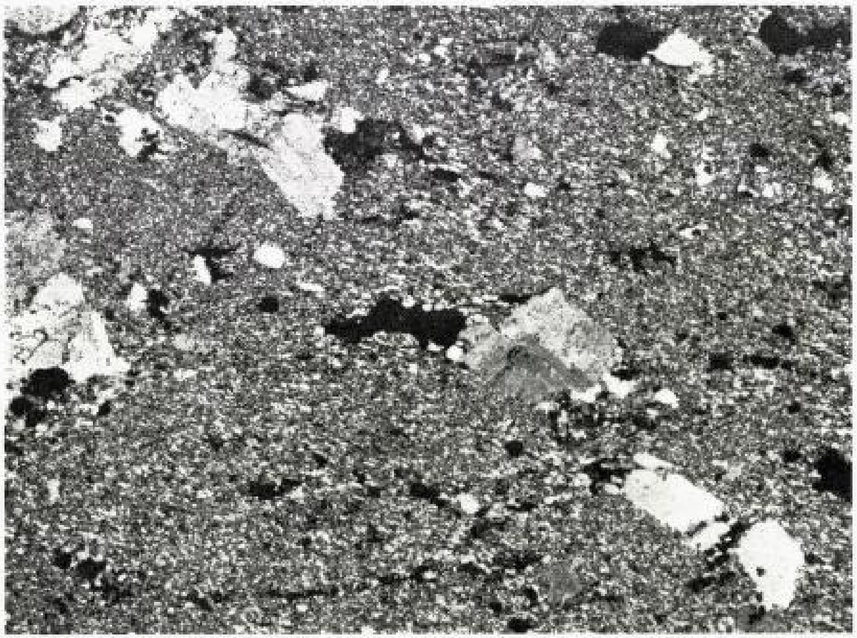

Fig 6a. Quartz-keratophyre from the igneous complex on the E bank of the lake Fundsjø. (Photomicrograph by D. Hejdova, magnification X 18, crossed nicols.)
*Kvartskeratofyr fra eruptivkompleks på østbredden av Fundsjøen. (Mikrofoto ved D. Hejdova, forstørrelse X 18, x-nicoler.)*

by gradual transitions with an almost 2.000 m thick monotonous sequence of greyish-green, fine-grained phyllitic siltstones and chlorite-sericite phyllites (17 and 18). This complex constitutes the main ridges of Kjølhaugene and Blåbergene and in a somewhat altered lithological facies crops out in the eastern part of the studied area in the mountain ranges Liefjeldene and Halsjøfjeld. Typical of the sequence are coarse-grained polymictic conglomerates (19) containing several metres-thick intercalations of graywacke and slate. The pebbles of this conglomerate are well rounded, of elliptical form and, on average, a few cm across. Exceptionally, 40 cm-sized pebbles are found. The pebbles consist prevalently of whitish-grey to grey quartzite and quartz, with lesser quantities of grey and bluish-grey quartz-keratophyre, fine- to medium-grained, more or less schistose leucocratic albite-granite, granite-porphyry and aplite; graywackes, keratophyre greenschists and limestone are subordinate. Most of the pebble material of the Kjølhaugene Conglomerate was undoubtedly derived from the older Ordovician, in particular the igneous complex (e.g. albite-granite

Fig. 6b. Quartz-keratophyre pebble from the Kjølhaugene conglomerate. (Photomicrograph by D. Hejdova, magnification X 18, crossed nicols.)

*Kvartskeratofyrbolle fra Kjølhaugenes konglomerat. (Mikrofoto ved D. Hejdova, forstørrelse X 18, x-nicoler.)*

and quartz-keratophyre, see figures 3, 5, 6a and 6b). North-east of the Fjergen lake, in the uppermost part of the sequence of grey-green chlorite-sericite phyllites to siltstones, the sequence of greenish-grey feldspathic sandstone (20) reappears with gabbro-diabase sills (21) directly above.

Similar beds to those of the Kjølhaugene mountain massif occur in the eastern part of the area, NE of the Skalsvatnet lake. These, however, are the product of a coarser clastic sedimentation and contain numerous intercalations of graywacke to subgraywacke and of conglomerate with nearly the same pebble content as that described above. The younger sequence of green-grey graywacke to subgraywacke (20) is developed in a substantially greater thickness. From this it follows that the source area for the sedimentation was situated to the East.

The youngest member of the sedimentary complex is represented by grey to blackish-grey graywackes in a 1 km to 1,5 km wide belt (in the syncline core) in the depression of the Nordelven River. The basal part of the complex consists of grey fine-gained phyllitic graywackes or subgraywackes, in places calcite-bearing (23, 24), which alternate with dm- to m- intercalations of grey sericite phyllitic slates increasing upwards in number. These rocks occur at the eastern margin of the belt, being strongly reduced or althogether absent from the western border. The upper parts is mostly composed of finely schistose graphitic phyllitic slates, grey-black in colour, which carry thin intercalations of dark siltstones (25).

In many places, sedimentary rocks of nearly all stratigraphical levels, except for the youngest sequence of grey-black phyllitic slates and siltstones, are penetrated by minor bodies, generally sills, of igneous rocks. Their thickness varies from several metres to a few tens of metres, while they may be several hundred metres in length. The rocks correspond in structure to gabbro and gabbro-diabase; mineralogically, they are composed of 50 per cent of a mafic mineral (invariably common green hornblende, never pyroxene) and 50 per cent of fully de-anorthitized plagioclase. During this process a large amount of minerals of the epidote group was generated. The grain-size varies greatly, depending on the distance from the contact and the thickness of the body. With regard to their age, these basic rocks are clearly younger than similar rocks of the igneous complex underlying the Lille-Fundsjø Conglomerate and probably older than the youngest member of the sedimentary complex.

### Notes on stratigraphy, tectonics and metamorphism

The stratigraphical positions of the above-described complexes are very difficult to assess in the absence of fossils. The only recorded ocurrence of fossils (Silurian graptolites) was described from the youngest sequence of greyish-black graphitic phyllitic slates on the eastern slope of Kjølhaugene (A. Getz 1890). All search for other biostratigraphical evidence has unfortunately been fruitless. Thus, the chronological assignment of the complexes can be based only on lithostratigraphical correlation with the analogous, paleontologically proved units of the Trondheim area. The detailed stratigraphical arrangement may be solved from a broad analysis of major areal wholes. Our investigation of a relatively small area permits only to state that the oldest member there — the igneous complex — is most probably comparable with the Lower Ordovician Støren Group. From this presumption an approximate dating of the voluminous sedimentary sequences, extending between the igneous complex and the fossiliferous Silurian phyllitic slates above, can be derived.

At first sight it would appear that proceeding eastwards one is moving down the stratigraphical sequence, but our investigation has corroborated F. Chr. Wolff's conception (1964) that in this area the beds are overturned, so that

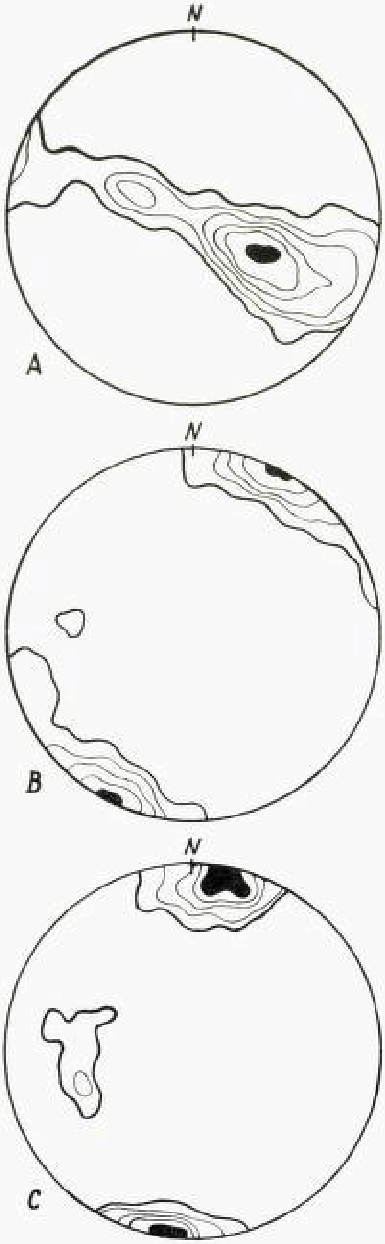


Fig. 7. Structural diagram.
A) — schistosity in the central and eastern parts of the area. (poles of planes). — Isolines: 50—30—20—10—5—2 %, 305 measurements.
B) — lineation in the central and eastern parts of the area. — Isolines: 30—20—10—5—2 %, 135 measurements.
C) — lineation in the south-western part of the area. — Isolines: 20—15—10—5—2 %, 88 measurements.
Equal-area projection,lower hemisphere.

*Strukturdiagram.*
*A) Skifrighet i de sentrale og østlige deler av området, (poler til plan). — Konturer: 50—30—20—10—5—2 %. 305 målinger.*
*B) Lineasjon i de sentrale og østlige deler av området. — Konturer: 30—20—10—5—2 %, 135 målinger.*
*C) Lineasjon i den sydvestlige del av området. Konturer: 20—15—10—5—2 %, 88 målinger. Schmidt-nett, undre halvkule.*

the structurally highest members in the west are stratigraphically the lowest. This opinion has been unmistakably confirmed by the study of conglomerate pebbles, in which rocks typical of the apparently (structurally) overlying sequence have been identified. The observation that, in the west the sedimentary cycles always begin with a coarser clastic sedimentation, is also consistent with the above assumption.

Foliation planes, interpretable as schistosity planes developed from the bedding, strike with a few exceptions at about 010°—040°; in the western and central areas they dip uniformly to the W (strictly to WNW), usually at about 50° (figure 7a). In the eastern part of the area investigated, in the proximity of the Swedish border (SSW of Skalsvatnet lake), the tectonic style is somewhat different. The dip of beds is not only to the west but also to the east at varying angles. From the geological section showing the structure of this sector (Plane II, A-B), it is apparent that the zone of youngest Silurian sediments is followed again by older rock complexes that form the eastern limb of a large assymmetrical syncline.

Linear structures are represented chiefly by the axes of folds of various size, ranging from several hundred metre-folds to phyllitic crumpling. Whereas the axes of large folds trend invariably in one direction, essentially consistent with that of foliation, minor fold axes display two widely different strikes. The predominant lineation system conforms in strike and plunge with large fold axes: in the eastern and central parts of the area it is subhorizontal and strikes at 030° (fig. 7 b); in the SW the strikes change from 030° and the mean plunge varies between 5—10° (fig. 7 c). The maximum of this lineation is nearly identical with the maximum of fold-axes established by graphical construction (fig. 7 a). The apparently monoclinal dip of foliation planes is partly caused by the presence of isoclinal folds (figure 8).

In the area studied, fault tectonics are not very pronounced, although a rather intensive fault activity might be presumed from an examination of the aerial photographs. In addition to a few longitudinal faults, several cross faults of minor importance were ascertained. Of special significance for the geological structure of the area is the reverse fault running along the state border near the Halsjøen lake, along which the Early Palæozoic beds contact the presumed Pre-Cambrian higher-metamorphosed rocks (mainly garnetiferous two-mica gneiss). It is accompanied by a mylonite and phyllonite zone, several tens of metres thick. In truncating a major part of the older complexes, this regional fault produced the assymmetry of the above-mentioned structure.

The joint tectonics are closely connected with the main lineation system.

Fig. 8. The characteristic type of faulting in the graywacke phyllites on the bank of the lake Lille Tjern.

*Den karakteristiske type av forkastning i gråvakkefyllitten på bredden av Lilletjern.*

The greater proportion of joints have the character of ac-planes in relation to the b-axis defined by this lineation.

The regional metamorpism of the Early Palæozoic complexes in the area studied is very weak, not exceeding the grade of greenschist facies viz. the quartz-albite-epidote-biotite subfacies. In metabasites it is distinguished by the assemblage albite-actinolite (common green hornblende or barroisite) epidote-chlorite( or biotite) and in the pelitic-psammitic rocks by that of sericite-chlorite (± biotite) -quartz (± calcite).

The alteration is of a markedly kinetic character, manifested by well-defined schistosity and folding. The intensity of metamorphic crystallization is somewhat higher in the older Early Palæozoic complexes. The youngest (Silurian) sequence occurring in the core of the syncline suffered the lowest-grade metamorphism.

One of the common mineral components is biotite, which in places forms numerous, large porphyroblasts up to several millimetres across. Unlike the remaining minerals, it is almost unaffected by deformation. It is present cheifly in greenish-grey graywackes, subgraywackes, siltstones and chlorite-sericite phyllites, particularly E of the tie-line Kopperå — Langen. Biotite porphyroblasts indicate adequately that their development took place under relatively tranquil kinetic conditions.

**Sammendrag**

En tykk lagpakke av metasedimenter med et kompleks av vulkanske og hypabyssiske intrusiver er beskrevet i denne artikkelen. Til tross for at lagene stort sett faller mot vest, kan det påvises at lagfølgen, de fleste steder, er invertert. Funnet av et polymikt konglomerat, ved Lille Fundsjø, som påviselig ligger stratigrafisk over den vulkanske serien har bestyrket dette syn. I øst opptar de yngste (siluriske) metasedimentene kjernen av en assymmetrisk synklinal som er veltet over mot øst. Gabbrolegemer, for det meste lagerganger, opptrer i alle bergartene unntatt de siluriske. Disse gabbroene er yngre enn lignende basiske bergarter i det eruptive komplekset. Regionalmetamorfosen i lagpakken har vært svak, dvs. grønnskiferfacies, og er uttrykt ved en veldefinert skifrighet. Lokalt fins biotittporfyroblaster yngre enn skifrigheten.

# Geology of the eastern part of the Meråker area

*by Anna Siedlecka*

**Abstract**

Within the mapped area slightly metamorphosed clastic sediments occur, probably of Upper Ordovician and Silurian age. The author describes a metagraywacke-slate association with metaconglomerates and gabbrodiorite sills, and a black-gray metasiltstone-slate association. Observations of preserved sedimentary structures, and microscopic investigations, indicate that these two association are different sedimentary facies. The first consists of beds representing a flysch facies formed by turbidity currents. The second is a black shale facies developed in euxinic environments. The beds are strongly folded and the folds overturned to the east. Silurian deposits form the centre of a syncline which runs from the Kjøllhaugene area south through the mapped area. Regional metamorphism altered the sediments to the greenschist facies, and in part to the epidote-albite-amphibolite facies.

## Introduction

During the summer 1965 I was a member of the geological field party organized by Fr. Chr. Wolff of the Norges Geologiske Undersøkelse, and mapped an area (ca. 150 km$^2$) situated northeast from Meråker. The area is bounded on the south by the main road to Storlien (Sweden), on the west by the Kopperå river, on the north by the Fjergen lake, the Sørelva river and Halsjøen lake, and on the east by the border between Norway and Sweden. I have done least work east of the Kjerringfjellene mountains because of inaccessibility and inclement weather.

Previous investigations in the Meråker area have been discussed by Wolff (pp. 7-8 in this volume). Because of this, the results of these earlier investigations will not be cited here.

## Acknowledgements

I am indebted to N.G.U., and especially to Fr. Chr Wolff for the very good organization of the field party and for the help he has given me.

Further, I wish to express my thanks to K. Birkenmajer for the helpful remarks about sedimentary markings, and to N. P. Lasca and R. P. Nickelsen who kindly corrected the English manuscript. The Institutt for Geologi, Universitetet i Oslo, was most generous in allowing me to use the Institutt's facilities.


## Rock characteristics

Within the mapped area occur monotonous clastic sediments, slightly metamorphosed and folded in tight or isoclinal folds. During the mapping, the following strata have been distinguished:

| | |
|---|---|
| 1. The metagraywacke-slate association, which is subdivided on the geological map into, (a) the predominantly metagraywacke beds with the addition of slates; and (b) the predominantly slaty beds with the addition of metagraywackes.<br>2. The metaconglomerates. | Kjøllhaugene Group |
| 3. The black-grey metasiltstone-slate association which on the geological map includes a separate subdivision for the metasiltstones and metasandstones.<br>4. The gabbro-diorite intrusions. | Slågån Group |

## KJØLLHAUGENE GROUP

*Metagraywacke-slate association*

The rocks were first described (Kjerulf, 1883; Reusch,1883, 1890) as grey and green «lersandstene», «lerstene», and occasionally as «skifre». Carstens (1920) described these rocks as sandstones interbedded with «lerglimmerskifer.» All writers emphasized the very monotonous character of the rocks occurring along the road to Storlien, between Meråker and the border between Norway and Sweden.

Most of the map area consists of the Kjøllhaugene Group. The group forms two zones called here: (1) the western or Bukhammer-Monsklumpene's zone, and (2) the eastern or Kjerringfjellene's zone. Altough the same sediments are found in both the eastern and western zones, there is some difference in sediment character between the zones. The dominantly slaty beds, found in the western part of the Bukhammer — Monsklumpene's zone grades eastward into a dominantly metagraywacke beds. The gradational zone between the slates and metagraywackes is used to mark the contact shown on the map (Pl. II). In the Kjerringfjellene's zone there is an interbedding of slates (which contain some metagraywackes) and metagraywackes (which contain some slates); metaconglomerates occur throughout the zone appearing in both the slate and metagraywacke beds.

Structural and textural features

Characteristic features of the metagraywacke-slate association are as follows:

1) Alternation of metagraywackes, slates, phyllites and metasiltstones. The metagraywacke beds range from 10 cm to 100 cm in thickness; in one case a 5 m thick bed was observed.
2) Rapid lateral variations in thickness and composition of beds are absent.
3) Most frequently the boundary between the metagraywacke and underlying slate is sharply defined. The boundary between the metagraywacke and the overlying bed is usually indistinct; there is often a transition from metagraywacke to slate, or metasiltstone. In such cases, the boundary is usually indicated by rock cleavage, which is distinct in the slates and absent in the metagraywackes.
4) On the surface forming the boundary between the slate and overlying metagraywacke, markings of sedimentary origin occur. These structures were observed only in cross-sections. As outcrops showing the bottom surface of the metagraywacke beds were not found, it was impossible to study the sedimentary markings in three dimensions. The most commonly observed markings are small round- or angular-backed crests, and long mud intrusions in the overlying metagraywacke. Crests and intrusions are generally asymmetric and point in the same direction (fig. 9, fig. 10, fig. 11, fig. 12, fig. 13). The structures seem to be flowage casts (Birkenmajer, 1958), or load casts (also called flow casts by Prentice, 1965; and torose-load casts by Crowell 1955),*) and they show great similarity to "flame-structures" (Walton, 1956; Kelling and Walton, 1957**), or "sawtooth-shaped contortions" (Mellen, 1956).

*) The term "flow casts" for sedimentary structures was used first by Shrock (1948), and the term "load casts" by Kuenen (1953 a, 1953 b), but neither writer had differentiated directional and non-directional markings of such type.

**) Interpretated by these writers as "flute-load-casts".

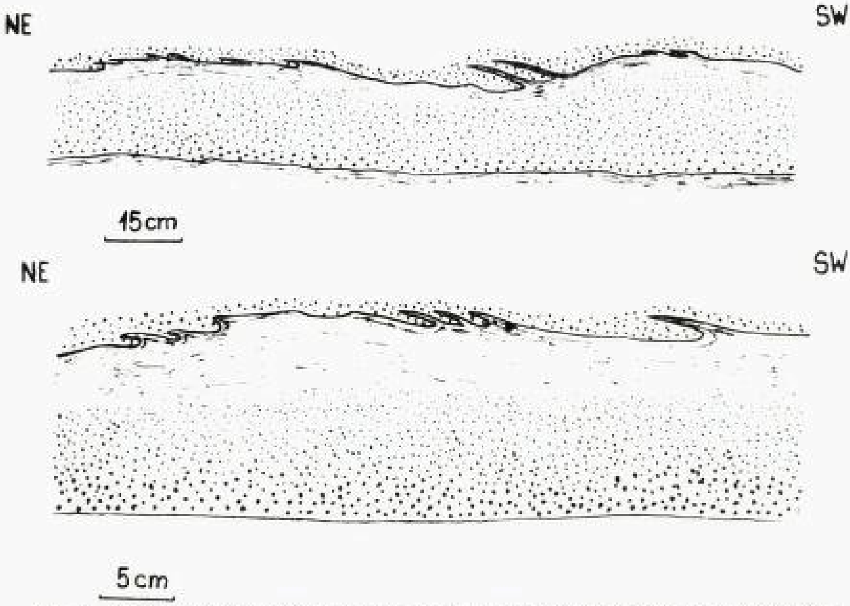


Fig. 9. Flame-structures (flowage casts or load casts). S shoreline of the Western Fjergen Lake.

*Flammestrukturer (flomavstøpninger eller pålastningsavstøpninger), sørbredden av Vest-Fjergen.*


Fig. 10. Flowage casts (load casts). The Grønbekk stream, upper part.

*Flomavstøpninger (pålastningsavstøpninger), øvre del av Grønbekken.*

Similar, non-directional forms have been described by Kuenen (1957) as "load-casted flow marks." Since "flame-structures" in the metagraywacke-slate series show a distinct direction, they are probably formed in one of three ways: (1) by differential loading which accentuates the primary flute casts, or (2) by gravity creep of the overlying soft sandy sediment which incorporates the underyling mud, or (3) by a combination of the two methods. Dzułynski (1963) from work in the Carpathian flysch has described similar directional clay instrusions caused by sand flow.

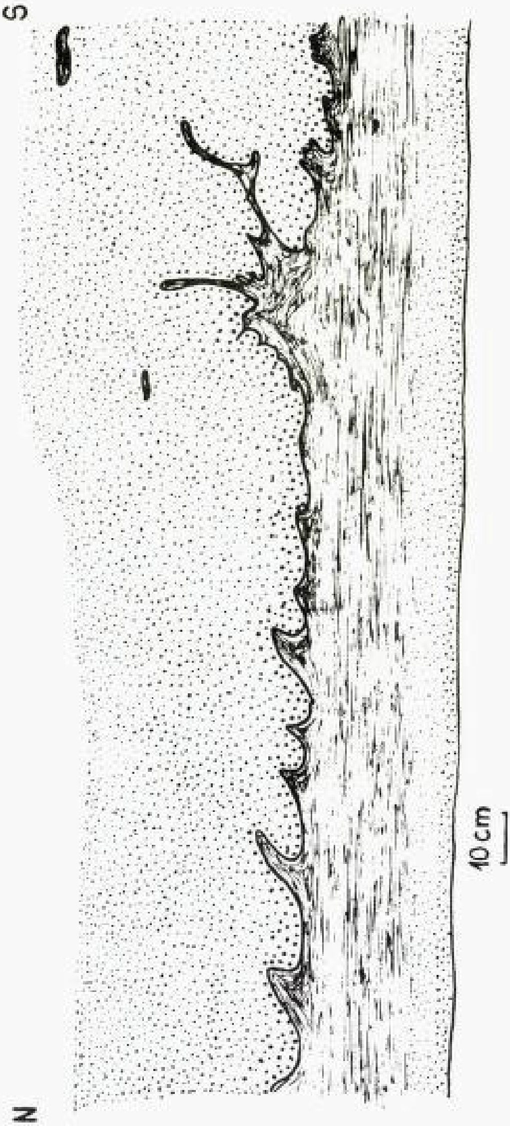


Fig. 11. Flame structures (flowage casts or load casts) and shale clasts floating in a metagraywacke. S shoreline of the Western Fjergen Lake.

*Flammestrukturer (flomavstøpninger eller pålastningsavstøpninger) og skifer filler flytende i en metagråvakke. Sørbredden av Vest-Fjergen.*

Fig. 12. Flame structures (flowage casts or load casts). Main road to Storlien, between Meråker and Grønberg.

*Flammestrukturer (flomavstøpninger eller pålastningsavstøpninger). Mellomriksveien til Storlien mellom Meråker og Grønberg.*

Fig. 13. Flame-structures (flowage casts or load casts). Main road to Storlien, between Meråker and Grønberg.

*Flammestrukturer (flomavstøpninger eller pålastningsavstøpninger). Mellomriksveien til Storlien, mellom Meråker og Grønberg.*

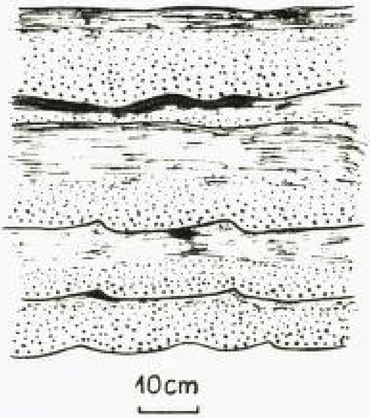


Fig. 14. Ripplemarks (?). The Lillekjerringelva river, upper part.
*Bølgeslagsmerker (?). Øvre del av Lillekjerringelva.*

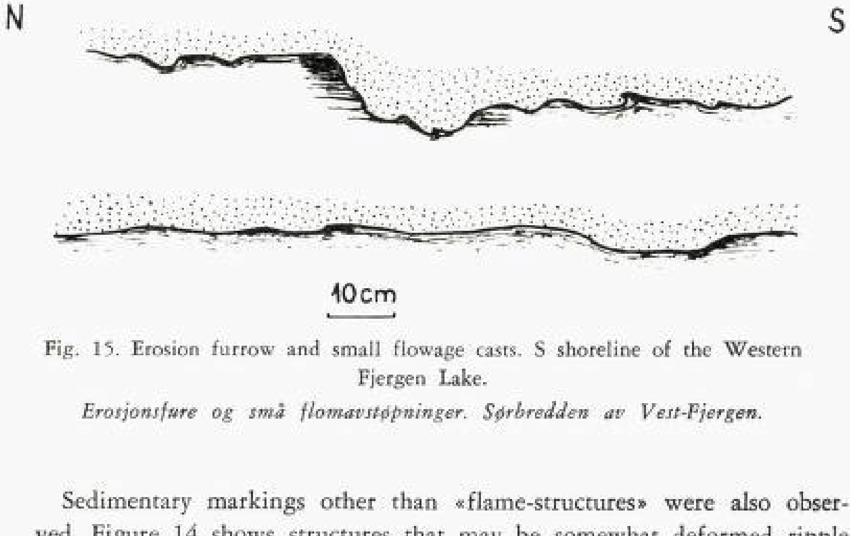


Fig. 15. Erosion furrow and small flowage casts. S shoreline of the Western Fjergen Lake.
*Erosjonsfure og små flomavstøpninger. Sørbredden av Vest-Fjergen.*

Sedimentary markings other than «flame-structures» were also observed. Figure 14 shows structures that may be somewhat deformed ripple marks. In fig. 15, the cross-section of an assymmetrical erosion furrow, and small later developed flowage casts are visible. Shale clasts, floating in the sandy sediment in the lower part of the metagraywacke layers (fig. 11, fig. 16), also indicate that subaqueous erosion of the sea bottom occurred.

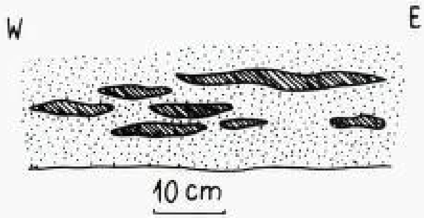


Fig. 16. Shale clasts floating in a metagraywacke. Meråker railway between the Grønbekk stream and Kopperå station.

*Skiferfiller "flytende" i en metagråvakke. Meråkerbanen mellom Grønbekk og Kopperå stasjon.*

5) Graded bedding is common. It is especially well developed in the thick metagraywacke layers. The most common type is asymmetric single normal graded-bedding (Fig. 17) but, more complicated graded bedding types were also observered, e. g., multiple grading in one layer, or asymmetric single inverted graded-bedding (terminology after Birkenmajer, 1959).

6) In the upper part of the fine-grained metagraywacke layers cross-bedding was sometimes noted. From the outcrops observed current direction could not be determined.

7) Metagraywackes are poorly sorted and consist of clayey, silty and sandy material; the clay and silt extend throughout the graded bedded layers. Prior to mechanical analysis the proportion of psammitic grains to the clay-silt matrix was determined. Three granulometric analyses were done following Krumbein's (1935) technique. The boundary between the sand and silt material was established at 0,06 mm (cf. Wentworth, 1922). The following grain size intervals > 0,06 mm were established:

0,06 —0,12 mm
0,12 — 0,18 mm
0,18 — 0,24 mm
0,24 — 0,30 mm
0,30 — 0,36 mm
0,36 — 0,42 mm etc.

Grains of quartz, feldspar, and rock fragments were measured; the biotite porphyroblasts were not. Sericite and chlorite flakes were generally under 0,06 mm in size. The results (see cumulative curves, fig. 18) show that the metagraywackes are poorly sorted and contain much clay and silt. These sediments were silty- and clayey sands prior to diagenesis and metamorphism.

Fig. 17. Graded bedding. Main road to Storlien, between Meråker and Grønberg.
*Gradert lagning. Mellomriksveien til Storlien mellom Meråker og Grønberg.*

The authigenic quartz recrystallization, and tectonic deformation of clastic grains, caused primary grains boundaries to be indistinct. Therefore, grain-size measurements could only be approximated. For the same reason roundness of grains could not be determined. In cases where the boundary between the clastic grain and the secondary quarts rim was distinct, a low roundness class was visible.

Petrology

As a result of microscopic investigations the following rock types has been subdivided:

*A. Feldspathic metagraywackes.* These rocks (Fig. 19—22) are fine grained, most commonly with grain-sizes up to 0,5 mm only. The texture is either massive or parallel. The parallel texture occurs only where the flaky

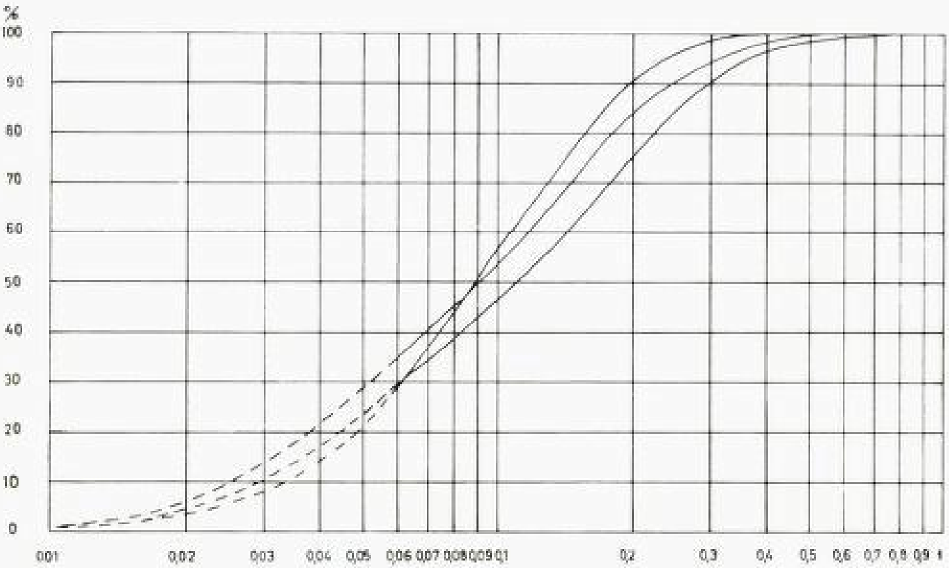


Fig. 18. Cumulative curves showing the granulometric composition of metagraywacke.
*Kumulative kurver som viser den granulometriske sammensetning av metagråvakker.*

minerals have a parallel orientation. Feldspathic metagraywackes consist mainly of quartz and feldspar, with occasional rock fragments. These components are evenly distributed throughout a matrix consisting of microcrystalline quartz, chlorite, sericite and plagioclase. The conventional boundary between coarser grains and matrix (ca. 0,6 mm) is used. Calcareous cement partly replaces the matrix. The secondary metamorphic epidote minerals and biotite occur in varying quantities; opaque minerals such as pyrite and iron oxide also appear.

The volumetric ratios between the constituents of the feldspathic metagraywackes were determined by statistical microscopic analyses*) and are summarized in Table 1 (p. 58).

A description of the constituents of the feldspathic metagraywackes follows:

*Quartz.* Detrital grains of quartz are either isometrical or somewhat elongate. Generally grain boundaries are very irregular due to overgrowths of authigenic quartz. In a few cases the boundary between the surface of

*) The point-count method of Chayes (1949) was used with the linear method as control.

Fig. 19

Fig. 20

a clastic grain and secondary quartz rim was observed. The recrystallization of authigenic silica caused individual grains to join forming larger grains. Some quartz grains show strain shadows when seen under crossed nicols. Most quartz grains contain inclusions, which were not studied in detail.

*Feldspar.* Detrital grains of feldspar are mostly plagioclase. The grains have the subangular shapes which are the result of little mechanical abrasion and cleavage. Plagioclases are relatively fresh and show albite twinning. Grains with both albite and pericline twinnings appear less frequently. Measurements of plagioclases taken on sections normal to 010 show that they contain ca. 10 % An (albite - oligoclase). Potassium feldspars, and perthites have been observed less frequently than plagioclases.

*Rock fragments.* 1) Fine-grained volcanic rocks with intersertal texture. These rocks consist of lath-shaped plagioclase (albite?), or of plagioclase and the allotriomorphic quartz, often with accessory chlorite. Scarcely any fragments of volcanic rocks with porphyric texture, in which the plagioclase phenocrysts are distributed through the fine-grained plagioclase background, were observed. 2) Quartzites. Quartzites are rare; they were visible only in coarser-grained metagraywackes. In fine-grained metagraywackes, quartzite fragments have probably been disintegrated. 3) Fragments of quartz-sericite and quartz-sericite-chlorite schists. These fragments are not common and are distinct only in the coarser-grained metagraywackes.

*Matrix.* The matrix is generally composed of ca. 30 % quartz, ca. 30 % chlorite, ca. 30 % sericite and ca. 10 % plagioclase. Quartz always shows strong regeneration. Chlorite occurs in very small flakes and has optical features similar to penninite.

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Fig. 19. Metagraywacke. Main road to Storlien, ca. 4 km W of the border between Norway and Sweden. (Photomicrograph by O. Brynhildsrud, magnification X 24, crossed nicols.)

*Metagråvakke. Mellomriksveien til Storlien, ca. 4 km vest for riksgrensen. (Mikrofoto ved O. Brynhildsrud, forstørrelse X 24, x-nicoler.)*

Fig. 20. Metagraywacke. Little unnamed lake, ca. 1,5 km NNE of the Bukhammer mountain. (Photomicrograph by O. Brynhildsrud, magnification X 24, crossed nicols.)

*Metagråvakke, lite navnløst tjern ca. 1,5 km nord-nordøst for Bukhammeren. (Mikrofoto ved O. Brynhildsrud, forstørrelse X 24, x-nicoler.)*

Fig. 21

Fig. 22

*Calcareous cement.* Calcareous cement is mostly calcite with subordinate amounts of other carbonates. It is present in distinct anhedral concentrations. Sometimes it also forms thin veins. Calcareous cement occurs in various quantities in the several of the thin sections examined (see Table 1).

*Minerals from the epidote group.* The minerals from the epidote group (epidote, clinozoisite) occur either as small isolated concentrations, or as well developed crystals ca. 0,05 mm in diameter.

*Biotite.* Biotite occurs as porphyroblasts, from 0,05 to 2 mm in size. The smaller porphyroblasts are mostly unoriented. The large ones (often with poikiloblastic texture) are parallel to the slate and phyllite foliation, which is not visible in the metagraywackes.

B. *Slates and phyllites.* These rocks (Fig. 23) are closely related to the metagraywackes. The top part of the graded layer of metagraywacke is often slaty and similar to the matrix in the lower part of the same layer. The mineral composition of slates and phyllites is as follows: quartz (ca. 30—60 %), sericite and muscovite (ca. 10—35 %), chlorite (ca. 5—40%). The minor constituents are plagioclases*), minerals of the epidote group (ca. 1—3 %), porphyroblasts of biotite (1—5 %) and of the opaque minerals (pyrite, magnetite?) (1—2 %). In slates lying near the calcareous metasiltstones, carbonates (ca. 15—20 %) were sometimes observed. These slates therefore form a transition from the slates and phyllites to the calcareous metasiltstones.

Slates and phyllites show parallel or lepidoblastic texture. Chlorite

*) Quantitative determination is difficult because the rock is very fine-grained.

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Fig. 21. Metagraywacke. Ca. 1,5 km NNW of the Skillerfjell mountain. (Photomicrograph by O. Brynhildsrud, magnification X 24, crossed nicols.)

*Metagråvakke. Ca. 1,5 km nord-nordvest for Skillerfjell. (Mikrofoto ved O. Brynhildsrud, forstørrelse X 24, --nicoler.)*

Fig. 22. Metagraywacke. Main road to Storlien, ca. 2 km W of the border between Norway and Sweden. (Photomicrograph by O. Brynhildsrud, magnification X24, crossed nicols.)

*Metagråvakke. Mellomriksveien til Storlien, ca. 2 km vest for riksgrensen (Mikrofoto ved O. Brynhildsrud, forstørrelse X 24, x-nicoler.)*

Fig. 23

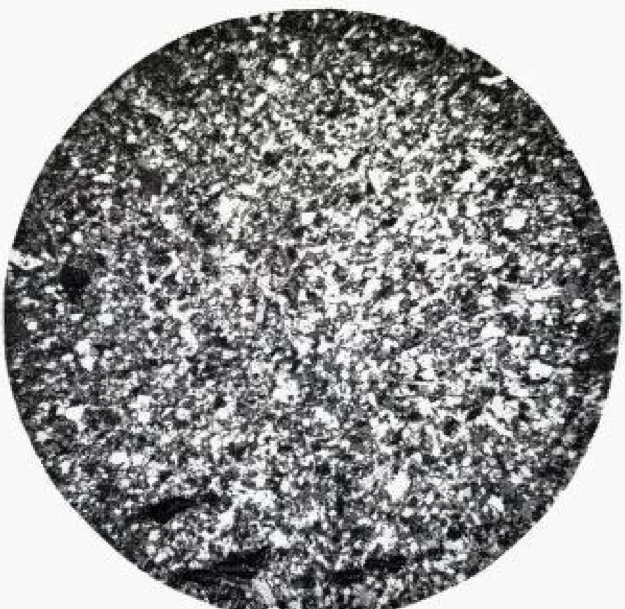

Fig. 24

and sericite flakes are very small but well defined. Quartz is partly regenerated, and some grains are slightly elongated parallel to the flaky minerals. A secondary cleavage crosses the original bedding and causes (1) a displacement and/or contortion of the bedding planes, (2) strain shadow and deformation of muscovite and sericite flakes, and (3) a complete destruction of bedding, if the cleavage planes occur close to one another.

Among the slates and phyllites, especially in the westernmost part of area schists were observed. They have the same mineral composition as the slates and phyllites, but are coarser-grained due to greater metamorphic recrystallization. Muscovite forms relatively large flakes and shows an exellent orientation. Quartz is completely regenerated, and its grains are more distinctly elongated parallely to the muscovite and chlorite flakes.

C. *Calcareous metasiltstones.* The calcareous metasiltstones (Fig 24) occur either as the thin independent layers within the metagraywacke-slate series, or as the upper parts of thick, graded layers of metagraywackes. Texturally they are an intermediate gradation between the metagraywackes, and slates or phyllites.

The calcareous metasiltstones occur more commonly in the Kjerringfjellene's zone. In the Bukhammer-Monsklumpene's zone, it was observed that metagraywackes grade directly into slate or phyllite. Metasiltstones have massive, or indistinct parallel texture caused by orientation of platy minerals. In some cases they show primary lamination, or graded bedding. The main constituents of calcareous metasiltstones are:

*Quartz grains.* Quartz grains (ca. 30—50 %), which reach a maximum of 0.1 mm in diameter, are generally $< 0{,}06$ mm. They are angular and often corroded by carbonates.

---

Fig. 23. Phyllite from the metagraywacke-slate series. The Grønbekk stream, upper part. (Photomicrograph by O. Brynhildsrud, magnification X 24, crossed nicols.)

*Fyllitt fra metagråvakke-skiferserien, øvre del av Grønbekken. (Mikrofoto ved O. Brynhildsrud, forstørrelse X 24, x-nicoler.)*

Fig. 24. Calcareous metasiltstone from the metagraywacke-slate series. Meråker railway, ca. 300 m W of Teveldal station. (Photomicrograph by O. Brynhildsrud, magnification X 24, crossed nicols.)

*Kalkholdig metaleirstein fra gråvakke-skiferserien. Meråkerbanen, ca. 300 m vest for Teveldal stasjon. (Mikrofoto ved O. Brynhildsrud, forstørrelse X 24, x-nicoler.)*

*Plagioclases.* Plagioclases (ca. 2—5 %) occur as small anhedral grains.
*Matrix.* Matrix (ca. 20—35 %) is composed principally of chlorite, sericite and quartz. The platy minerals are more abundant than quartz.
*Carbonates.* Carbonates (calcite and dolomite) occur either as anhedral grains (ca. 0,05—0,1 mm in diameter), which are sometimes twinned, or as simple rhombic crystals ca. 0,03—0,04 mm in size.

Minor constituents of the calcareous metasiltstones are biotite porphyroblasts and opaque minerals up to 1 mm in size. Many of the biotite porphyroblasts show poikiloblastic texture.

*Metaconglomerates*

Many layers of metaconglomerate occur in the Kjøllhaugene Group in the eastern zone. In the western zone they were observed in only one very poor outcrop on the south shore of West Fjergen lake.

Metaconglomerates from the Kjerringfjellene Mts. and similar sediments from the Kjøllhaugene and Halsjøfjell areas (north of the mapped area) were known to previous geologists. The metaconglomerates were reported by Kjerulf (1883), Törnebohm (1896), and Carstens (1920). Törnebohm (1896) mentioned the lateral disappearance to the south and north of the Kjøllhaugene conglomerates. Carstens (1920) first compared these conglomerates with the Lyngesten conglomerate from the Gauldalen valley.

In the mapped area metaconglomerate layers reach thickness of 3 m. Only i one profile, located between mountain tops 1067 m and 1018 m (Pl. II), were two layers greater than 10 m thick observed. Sparsely distributed pebbles were visible in some of the metagraywackes and metasiltstones lying above or below the metaconglomerate layer.

The metaconglomerates are poorly sorted and have white, grey and pink quartz and quartzite pebbles as the principal constituents. Accessory pebbles of limestone and of dark-grey and black volcanic (?) rocks occur. Pebble roundness varies greatly from subangular to well rounded fragments. The fine-grained material is less rounded than the coarse-grained material. Frequently the pebbles are scattered throughout an abundant matrix (conglomeratic mudstone), but in some layers they are closely packed. The coarser pebbles and cobbles, especially in metaconglomerates with a very abundant matrix, are elongated and oriented parallel to the bedding planes. This orientation seems to be in part a secondary, tectonic feature. The matrix of metaconglomerates is sandy and/or silty and is identical to the texture of the surrounding metagraywackes, metasiltstones, and slates.

An important characteristic of metaconglomerates is their lateral change. In the northeast part of the map area (Pl. II) the metaconglomerates are pebbly (material is up to ca. 5 cm in diameter) with a few cobbles as much as 20 cm in diameter. Southward the metaconglomerates gradually become finer-grained; those with closely-packed pebbles disappear, and only the conglomeratic mudstones occur. Further south conglomeratic mudstones grade into gravelites, gravelly metagraywackes or siltstones, and finally as the pebbles disappear, to the metagraywackes, metasiltstones, or slates of the metagraywacke-slate association. Because of the lateral change of the metaconglomerates, five to ten conglomeratic layers, not seen in the profile along the main road to Storlien, are seen in different profiles adjacent to Halsjøen lake. In the profile along the main road to Storlien, in two localities fine-grained conglomeratic mudstones were observed and in one locality a gravelly metagraywacke.

## SLÅGÅN GROUP

*Black-grey metasiltstone-slate association*

The black-grey metasiltstones and slates occupy the central part of the mapped area, between zones of the metagraywacke-slate association (Pl. II), and extend from the northern boundary of the map to the Storlien road. It is a continuation of the black-grey shales of Silurian age occurring in the Kjøllhaugene area. The zone of black-grey shales is visible on the Törnebohm's (1896) map. The rocks of this zone were also described by Reusch (1890) from a profile along the Meråker railway: «*Omtrent 2½ kil. i Ø for vogterhuset Tovmodalen møder man en tyndskifrig, smaarynhet, graa lerglimmerskifer, der holder ved omtrend 800 m. I den derpaa følgende kvidlige sandstenagtige bergart træffes den første dioritiske masse ...*» (Reusch, 1890, p. 14). Reusch did not compare these rocks with the Silurian deposits from the Kjøllhaugene area. Minor occurrences of the black-grey metasiltstones and slates were also observed by me in the eastern zone of the metagraywacke-slate series near the Storkjerring lake.

The boundary between the black-grey metasiltstone-slate association and the metagraywacke-slate association of the western zone is distinct, but not sharp. The boundary between the black-grey metasiltstone-slate association and eastern zone of the metagraywacke-slate association is not distinct. There is a transition between both associations.

Within the mapped area the black-grey metasiltstone-slate association consisst of dark-grey calcareous metasiltstones, fine-grained metasandstones, and dark-grey and black slates and phyllites. Metasiltstones, slates and phyllites when seen in the field usually show a characteristic type of disintegration. They turn rusty-grey in colour, and weather extensively leaving abundant tabular and prolate fragments.

Metasiltstones and fine-grained sandstones commonly occur near the west boundary of the Kjerringfjellene's metagraywacke-slate zone (Pl. II), where they are observed as thin to medium thick, sometimes laminated beds intercalated between phyllites and slates. The sandy or silty laminae were also observed between slaty rocks.

Under the microscope the metasiltstones and fine-grained metasandstones show a massive or indistinct parallel texture, and consist of quartz grains, abundant matrix, many carbonates and accessory feldspars, biotite and opaque minerals. These minerals are described below.

*Quartz.* Quartz grains are commonly equidimensional; grain size reaches a maximum of 0,1 mm, but is most commonly $<$ 0,05 mm. Primary boundaries of clastic grains are usually not recognisable due to overgrowth of secondary quartz and corrosion by carbonates.

*Feldspar.* Clastic grains of feldspar are usually poorly-rounded and in most cases are acid plagioclases.

*Carbonates.* Carbonates form two kinds of concentrations: (1) anhedral grains often twinned, up to 0,25 mm in size (calcite) and (2) rhombic crystals 0,025 — 0,05 mm in size, single or in aggregates, many of them with iron-oxide rims (dolomite).

*Matrix.* The matrix is a massive, very fine-grained mixture of quartz, chlorite and sericite.

*Biotite.* Biotite forms porphyroblasts.

*Opaque minerals.* Opaque minerals occur throughout the matrix in anhedral concentrations which reach a maximum of 0,25 mm in size.

Statistical microscopic analysis of metasiltstone shows the following composition:

| | | | |
|---|---|---|---|
| quartz | 21,3 % | | |
| feldspar | 3,6 % | quartz | ca. 45 % |
| matrix | 33,8 % | sericite | ca. 35 % |
| carbonates | 33,0 % | chlorite | ca. 20 % |
| biotite | 7,5 % | | |
| opaque min. | 0,8 % | | |

Slates and phyllites usually have lepidoblastic texture parallel to the primary bedding. Relict structures, such as fine-graded bedding and lamination (interbedding of flaky minerals and quartz laminae) have been observed. The lepidoblastic texture is crossed by cleavage-planes. Because of the cleavage, bedding planes are displaced and contorted (Fig. 25, Fig 26). Cleavage is not visible in metasiltstones (Fig. 26) and metasandstones.

The major constituents of the slates and phyllites are either quartz, sericite and chlorite, or just quartz and chlorite. A few plagioclase fragments have also been observed. Many porphyroblasts are also present; most of them are biotite porphyroblasts, or poikiloblasts, up to 1mm in size, and porphyroblasts of opaque minerals (mostly pyrite) 0,5 mm in size.

Pyrite crystals lying parallel to bedding and deformed by cleavage were observed. Therefore, we know that the pyrite is older than the cleavage. In only one case have carbonate porphyroblasts and none of biotite been observed. Minor amounts of epidote minerals, zircon and tournalines are also present. The slates and phyllites described probably contain graphite (?) which accounts for the dark-gray and black colours.

*Gabbro-diorite intrusions*

Within the western zone of the Kjøllhaugene Group there occur many gabbro-diorite intrusions. These were known to previous geologists. Kjerulf (1883) was the first to mention numerous masses of diorite and saussurite gabbro from this area. He called the rocks on Midsundstøtten syenite-like rocks following O. Schiötz's usage. Reusch (1890) described the intrusions as diorite masses. Later, Carstens (1929) described these rocks as gabbro intrusions and gabbro-like pegmatite-veins.

The intrusions form sills which are injected into the metagraywackes and slates. The sills range in thickness from a few metres to ca. 100 m. Gabbro-diorite sills on Midtsundstøtten, and in the western part of Grønbæklien (Pl. II), are the largest which I have observed in the area. The gabbrodiorite sills show variations in colour and texture. They are (1) finegrained grey-green (e.g. near Kopperå station) and dark-grey (e.g. at the stream west of the Lillekjerringelva river), or (2) coarse-grained with feldspars and amphiboles up to ca. 2 cm in size (e.g. at Midtsundstøtten, and Grønbæklien, a part of the intrusions visible along the main road to Storlien). The largest coarse-grained sills show gradually finer grained texture towards the contact with the country rock. Contact-metamorphic zones are usually narrow, and represented by hornfelses with carbonate porphyroblasts.

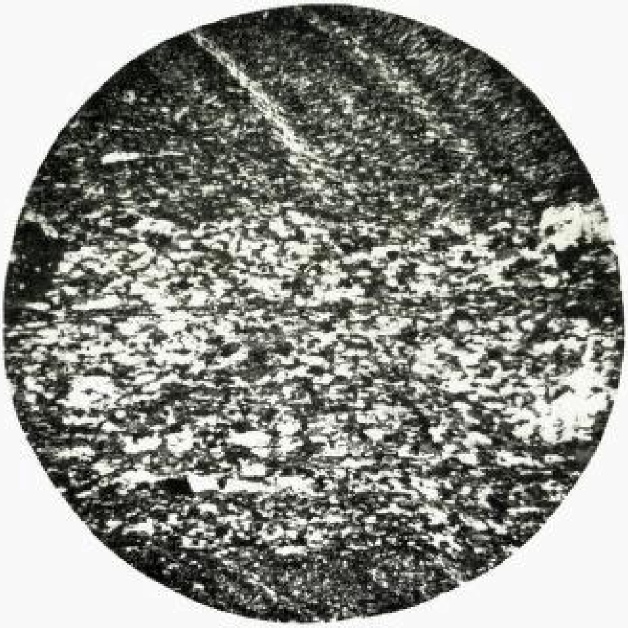

Fig. 25

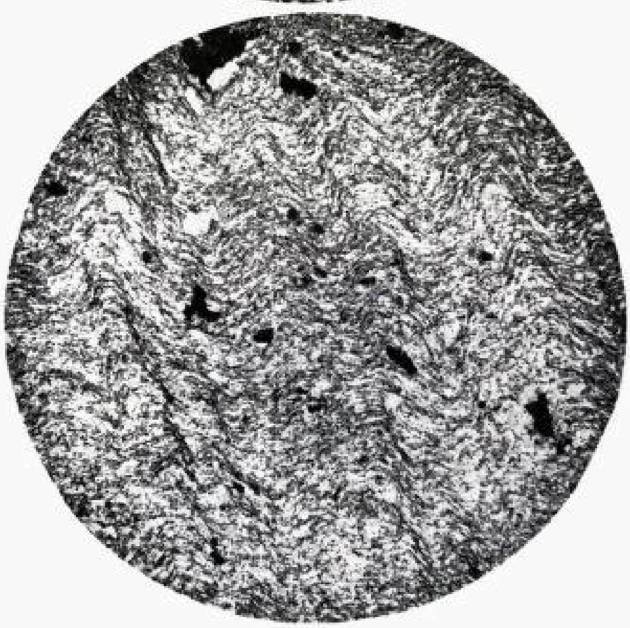

Fig. 26

The principal components of the gabbro-diorites are amphiboles (ca. 30—45 %) and plagioclases (ca. 30—60 %). Minor constituents are: biotite, chlorite, sericite, epidote minerals, quartz, carbonates, titanite and opaque minerals.

The plagioclases (ca. 35 % An) are either allotriomorphic or hypidiomorphic. Many are partly altered to saussurite. Amphiboles are idiomorphic or hypidiomorphic, and many are converted to biotite, or altered to aggregates of chlorite, carbonates, epidote and quartz.

Petrographic descriptions of gabbro-diorite instrusions will be made by F. Fediuk, who has my field samples.

### Development of sediments

As has been mentioned, the metagraywacke-slate association oocupies most of the map area. Characteristics of the rocks have been described in detail (see p. 23 f.). Many preserved primary features are important indicators for the determination of (1) the sedimentary environment, (2) the mechanism of deposition, and (3) the geology of the source area.

The metagraywacke-slate association consists mostly of interbedded metagraywackes and slates. On the bedding surfaces between the metagraywackes and slates, sedimentary markings occur. Graded bedding is common, but sorting is so poor that in the lowermost parts of graded layers fine and coarse material occur together. Matrix is usually very abundant and has the same composition as slate. The sand-size material consists of quartz, feldspar and occasionally rock fragments. The term graywacke facies is applied to the sediments (whether weakly metamorphosed or not) which contain

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Fig. 25. Slate with a silty lamina from the black-grey metasiltstone-slate association. Main road to Storlien, ca. 0,5 km E of the last gabbro-diorite sill. (Photomicrograph by O. Brynhildsrud, magnification X 24, crossed nicols.)

*Skifer med et leirlag fra den gråsvarte metaleirstein-skiferserien. Mellomriksveien til Storlien. Ca. 0,5 km øst for den siste gabbro-diorittgangen. (Mikrofoto ved O. Brynhildsrud, forstørrelse X 24, x-nicoler.)*

Fig. 26. Slate from the black-grey metasiltstone-slate association. Ca. 1,2 km NNE of Monsklumpen. (Photomicrograph by O. Brynhildsrud, magnification X 24, plane polarised light.)

*Skifer fra den gråsvarte metaleirstein-skiferserien, ca. 1,2 km nord-nordøst for Monsklumpen. (Mikrofoto ved O. Brynhildsrud, forstørrelse X 24, planpolarisert lys.)*

the characteristic features mentioned above. The dominance of feldspar over rock fragments in the metagraywackes, seems to indicate a plutonic provenance for the clastic material. It is possible that feldspar was present in greather amounts prior to induration, and that its very fine grains have quickly disintegrated so contributing to the quartz-sericite matrix. The composition of metagraywackes also indicates the predominance of mechanical weathering in the source area. The low sorting index seems to be caused by quick, not selective transportation.

The next important problem to be considered is the environment of sedimentation of the graywacke facies, and its relation to flysch. The term flysch is used by some geologists exclusively as a facies term (e.g., Vassoevič, 1948, 1951: Sujkowski, 1957) by others as a facies and genetic term. In the second case, the term indicates both facies and a definite stage in the development of a geosyncline (e.g. Vassoevič, 1958; Bouma, 1962; Contescu, 1963.) In recent years flysch facies has been defined as a sequense of marine clastic sediments, characterized by an assemblage of positive and negative diagnostic features (e.g. Dzułynski, 1963; Dzułynski and Smith, 1964; Dzułynski and Walton, 1965). A comparison of these features with the characteristics of the metagraywacke-slate association shows that it conforms to the facies definition of flysch.

In many recent papers flysch deposits are considered as deep water sediments, deposited in a geosynclinal trough flanket by tectonically active source lands. The most important agents contributing to flysch sedimentation are thought to be gravity mass movements such as submarine slumps and slides, and turbidity currents (see Kuenen, 1958; Ksiazkiewicz, 1958). Many flysch formations can be called turbidite formations (Kuenen, 1964) if their characteristic features show that they are formed by turbidity currents. Most positive and negative features of turbidite formation (see Kuenen, 1964, p. 16) have been found in the metagraywacke-slate association in the Meråker area, and it is therefore concluded that this association is a graywacke and flysch facies (partly shaly flysch, partly sandy flysch), and a turbidite formation. An abundance of sedimentary markings, which is usually present in flysch and turbidite formations, has not been observed, but future detailed sedimentological studies may provide additional data.

It was not possible to study the directions of sediment transportation in the metagraywackes and slates in detail. The alinement of mud intrusions («flame-structures») indicates transport from the north or northeast, sometimes from the south, along the longitudinal axis of the sedimentary basin.

Metaconglomerates which occur within the Kjerringfjellene's zone of metagraywacke-slate association consists mostly of quartz and quartzite pebbles, and often have a similar character to conglomeratic mudstones. It does not seem possible that the quartz and quartzite pebbles, and the sandy or silty graywacke material, have the same provenance. If the source area was the same the conglomerates should have a polymictic character. A more probable hypothesis is that pebbly material was transported from another direction (e. g. from the E side of the trough), across and subordinate to the principal longitudinal direction. It is possible that pebbles and cobbles were transported from the shallow-water zone along shelf channels, and then along a submarine canyon crossing a continental slope, either by currents and/or by slump movements, to the deep-water zone of the basin. The occurrence of the metaconglomerates within the deep-water flysch sediments, the quick lateral disappearence of the metaconglomerates, and the commonly observed dominance of graywacke matrix over pebbles leads one to conclude that the conglomerates were deltaic sediments deposited at the mouth of a submarine canyon. Although there are several explanations for conglomeratic mudstones (called also tilloid conglomerates), in the Meråker area the most important fact, both as to genesis and stratigraphic interpretation, is that conglomerates occur within geosynclinal flysch sediments. The discussed tilloid conglomerates in the Meråker area are therefore thought to be deep water sediments.

In the eastern zone of the metagraywacke-slate association, the calcareous character of sediments is more common than in the Bukhammer-Monsklumpene's zone. This characteristic of the eastern zone seems to be related to a somewhat shallower water environment lying nearer the continental slope. Most pebbles of the metaconglomerates were transported and deposited in the shallower zone, but some were transported to the deeper western zone.

The similarity between the Meråker metagraywacke-slate association, and both the Silurian deposits of Wales and the Upper Ordovician deposits of Scotland, should be emphasized. The similarities occur in (1) the structural characteristic and mineral composition (see Wood and Smith, 1958), (2) the presence of «flame-structures» (see Walton, 1956; Kelling and Walton, 1957), (3) the common occurrence of graded bedding, and (4) the predominance of longitudinal directions of transportation (see Kuenen, 1957; Knill, 1954, 1960; Kelling, 1964).

The metagraywacke-slate association grades vertically into dark-grey and black slates and phyllites interbedded with meatsiltstones and fine-grained sandstones. The boundary between both associations is relatively distinct, but not sharp. The slaty-silty dark-grey and black rocks show the following features, important to determination of sedimentary environments: (1) a dominance of shales, (2) a dark-grey and black colour, (3) the presence of pyrite*), (4) usually an abundance of carbonates, and (5) the presence of planktonic and/or epiplanktonic fauna (*Monograptidae* and *Rastrites* from Kjøllhaugene). Sediments with such features are developed in euxinic environments at varying depths and are called black shale facies or euxinic facies.

**Stratigraphy**

Establishment of the stratigraphic position of the described beds is difficult because fossils are lacking, and lithology is very monotonous. The first stratigraphy was based on lithologic similarities between the Meråker area sediments and those of the western part of the Trondheim region. Later the stratigraphy was revised to include the Silurian graptolite fauna found by Getz (1890) in the dark shales of Kjøllhaugene .The graptolites from Kjøllhaugene were verified later by Elles (Kiær, 1932), who established that Getz's descriptions were correct, and that this fauna indicates sediments corresponding in age to the upper part of Middle, and Upper Birkhill time in England and Scotland.

Törnebohm (1896) described the rocks of the eastern part of the Meråker profile as the Meraker Group *(Meraker-gruppen)*, and on the basis of the lithologic similarities compared it with the Hovin Group from the western part of the Trondheim region. The Meraker Group on Törnebohm's (1896) map is bounded on the west by the Sul Schists Group *(Sul skiffres gruppen)*, which Törnebohm compared with the Høiland Group**). Törnebohm has included in the Sul Schists Group the dark shales in which the Silurian graptolites were found. Therefore, both the Sul Schists Group and the Høiland Group were considered Silurian in age by Törnebohm.

Carstens (1920) suggested that the so-called Lyngesten conglomerate from the Hovin district was similar to conglomerates from Kjøllhaugene; further that conglomerates from both areas, and shales with Silurian graptolites from Kjøllhaugene, together with the subjacent sandstones are younger than the Høilanda division (equal to the lower part of the Lower Hovin Series of

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*) Pyrite can be partly secondary and not indicative of conditions of deposition.

**) This view, accepted by Carstens (1920), was later discarded because it was based on an incorrect interpretation of the tectonics in the western part of the Trondheim region.

Kiær, 1932; and Vogt, 1945). Later writers (Kiær, 1932; Vogt, 1945; Strand, 1960: Wolff, 1964) accepted Carstens' correlation and assign the conglomerates from Kjøllhaugene and Lyngesten to the Silurian. The series of subjacent sandstones and shales has been included in the Upper Hovin Group. The age of this group is thought to be Ashgillian because, in the western part of the Trondheim region, shales with graptolite fauna of Caradocian age occur below a similar sandy series and below the Volla conglomerate.

From field observations of the author it has been established that:

1. The black-grey metasiltstone-slate association is a continuation of the Silurian dark shales from the Kjøllhaugene area. No fossils were found in the map area. To the south, the rocks show higher grade metamorphism which may have caused complete destruction of the fauna.
2. The sedimentary structures occurring in the metagraywacke-slate association show that both the eastern and western zones of this series are older than the black shale facies containing the graptolite fauna.
3. There is a continuity of sedimentation between the graywacke facies and black shale facies.
4. Layers of metaconglomerates occurring within the metagraywacke-slate association do not delineate any stratigraphic boundaries.

Since other areas in the Trondheim region were not visited by the author, it is difficult to compare rocks from other parts of the Trondheim region with those found in the mapped area (Pl. II). Some similarities can be drawn between the metaconglomerates of the mapped area and the Lyngesten conglomerate, which has been described in detail by Vogt (1945). The principal similarity between the conglomerates is the dominance of the quartz and quartzite pebbles. However, differences exist indicating different sedimentary environments. The Lyngesten conglomerate (Vogt, 1945) is a basal conglomerate and is an index layer for the stratigraphy. Moreover, « . . *Details previously mentioned by Brøgger display unconformable relations to the substratum, and the conglomerate apparently also overlaps older beds*» (Vogt, 1945, 523). Vogt (1945) regarded the Lyngesten conglomerate as an effect of the Horg disturbance which occurred between Ordovician and Silurian. The Horg disturbance caused the elevation above sea level and the denudation of land masses in the Horg area. Vogt suggested that, in his area, it was a regresion and denudiation that caused a stratigraphic hiatus.

A comparison of features characteristic of both the Lyngesten conglomerate and the metaconglomerates from Kjerringfjellene (see p. 38; 45) show that both have (1) similar composition and (2) different genetic features. These two conglomerates were deposited under different conditions. While the Lyngesten conglomerate is probably a littoral sediment, the metaconglomerates from Kjerringfjellene are deep marine conglomeratic mudstones interbedded with flysch and cannot be used as direct evidence for orogenic disturbance.

The differences between the conglomerates do not negate possibility of contemporaneous deposition, but indicate that stratigraphic correlation cannot be based here solely on the lithologic character of pebbles.

As mentioned earlier, the boundary between the metagraywacke-slate association with metaconglomerates, and the black-gray metasiltstone-slate association of Silurian age is not sharp. Therefore, the stratigraphic boundary between the Ordovician and Silurian may only be shown by a facies change in the Meråker area. The difference in depositional environment, between the northern and western part of the Trondheim region, is probably caused by the greather distance of the Meråker area from the former sea-shore. There was continuous deposition in the Meråker area, first, the rapid, deep-marine sedimentation of flysch, and later the slower deposition of the black shale facies.

From the present work, it is concluded that the stratigraphy of the Meråker area is as follows:

| | |
|---|---|
| Black-grey slates, phyllites, metasiltstones and finegrained metasandstones | Slågån Group = Horg Group = Lower Silurian |
| Metagraywacke-slate association with metaconglomerates | Kjøllhaugene Group = Upper (?) Hovin Group = Upper (?) Ordovician |

The gabbro-diorite sills occurring in the western zone of the metagraywacke-slate association are younger than the surrounding sediments and older than the main phase of Caledonian orogeny. A more exact dating of the sills is very difficult.

Rocks showing similarities to the metagraywacke-slate association have recently been described from the Verdalen valley north of Kjøllhaugene, and from the Blåsjö lake area in Jämtland (Sweden). These rocks may be an extension of the metagraywacke-slate series from the Meråker area. From the eastern part of the Verdalen valley area, Wolff (1960) described the Vera schists as consisting mainly of chlorite schists with biotite porphyroblasts and including horizons of quartzite conglomerates. He correlates the conglomerates with those from Kjøllhaugene and concludes that the Vera schists are Silurian in age. The black-grey Silurian shales known from Kjøllhaugene are not found in the Verdalen valley area; they disappear probably in the northern part of the Kjøllhaugene area (see Törnebohm's map, 1896).

Nilsson (1964) has described the Blåsjö phyllite, which is a calc-phyllite with gabbro intrusions, from the Blåsjö area. "*This calc-phyllite has been deposited fairly rapidly in shallow well-areated sea-water as marls- calciferous sandstones and muds*" (Nilsson, 1964, p. 66). Nilsson regards the Blåsjö phyllite as the lowermost unit in the succession (older than Lower Ordovician). Descriptions of the sedimentary structures are lacking, however, petrographic characteristics seem to indicate that the Blåsjö phyllite could be an equivalent of the metagraywacke-slate association. If future comparative studies indicate that the same rocks occur in the Meråker, Kjøllhaugene, Verdal and Blåsjö areas, as seen by gradually shallower facies northward, the stratigraphy in the Blåsjö area should be revised.

## Remarks concerning the structural geology

Recognition of structures is very difficult because: (1) only two lithostratigraphical units occur in the area; (2) the metagraywacke-slate association probably changes laterally and vertically from shaly flysch to sandy flysch; (3) guide horizons are lacking; and (4) the beds are very strongly contorted and folded. The multiplication of the small folds does not permit measurements of thickness of particular series.

Generally, the beds strike NNE with some local variations to N or NE. In the western part of the area the beds dip 30—100°*) W (27—90°). In the eastern part of the area they dip to the east, the angles of dip are gradually smaller to the east, and some of the beds lie horizontally (E part of Kjerringfjellene Mts.). The relations are expressed in cross-sections (Pl. II).

Two fold structures within the western zone of the metagraywacke-slate association have been observed: the Midtsundstøtten syncline and Grønbæklien syncline (Pl. II). The Midtsundstøtten syncline is very well exposed (Fig. 27); it is asymmetrical with a ca. 90° (81°) steep western limb and gentle, ca. 25° (22,5°), eastern limb. The axial plane of the syncline dips to the west. Moreover, a gabbro-diorite sill which can be traced in both limbs clearly marks the syncline's form. The Grønbæklien syncline is poorly

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*) A Swedish compass with 400° was used. The measurements in parenthesis are recounted to 360°. All measurements of strike and dip shown on the geological map are those of the 400° compass.

Fig. 27. Western limb of the Midtsundstøtten syncline, looking south.
*Vestsiden av Midtsundstøttens synklinal sett mot syd.*

exposed. However, it was possible to trace this syncline because a gabbro-diorite sill, folded conformably to the sedimentary layers, is very distinctly reflected in the surface morphology. A second, nearly completely eroded, gabbro-diorite sill gives a characteristic hill in the central part of the syncline. Many small folds were observed in addition to those of the Midtsundstøtten and Grønbæklien synclines, especially in the slaty incompetent complexes. Such folds (fig. 28, 29, 30, and 31) are mostly tight or isoclinal cleavage folds, disharmonic folds, and tectonic contortions.

Metagraywackes interbedded with some slates occur to the east of the Midtsundstøtten and Grønbæklien synclines (Pl. II). These beds are inverted (as indicated by sedimentary structures) and lie in contact with the metasiltstones and slates of Silurian age to the east. It seems probable that the metagraywacke complex forms an anticline with a tectonically reduced western limb. A second possibility is that the anticline is a folded group of sediments in which a facies change occurred; i.e., the western limb (eastern limb of the Midtsundstøtten syncline) consists of the slates and metagraywackes, while the eastern limb consists of only metagraywackes (see Pl. II). In this case, the anticline

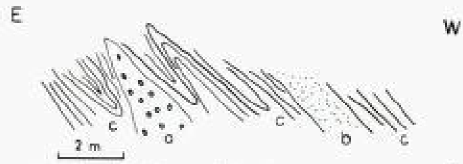


Fig. 28. Tight or isoclinal folds observed in the eastern zone of the metagraywacke-slate association. a — conglomeratic mudstone, b — metagraywacke, c — slate.
*Bratte, tette folder observert i den østre sonen av metagråvakke-skiferserien. a. Konglomeratisk slamstein, b. metagråvakke, c. skifer.*

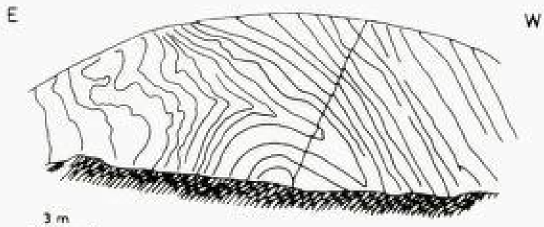


Fig. 29. Folded and contorted beds in the western zone of the metagraywacke-slate association (above S shoreline of the Fjergen lake).
*Foldede og vridde lag i den vestre sonen av metagråvakke-skiferserien (over sørbredden av Fjergen).*

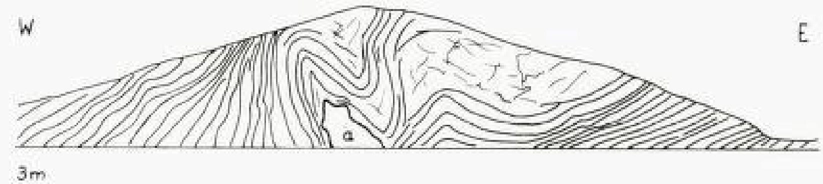


Fig. 30. Folds and joints in the western zone of the metagraywacke-slate association. a - quartz. Meråker railway.
*Folder og sprekker i den vestre sonen av metagråvakke-skiferserien. a - kvarts. Meråkerbanen.*


Fig. 31. Folds in the eastern zone of the metagraywacke-slate association. Cross-section along a fissure cutting the Skillerfjell mountain.
*Folder i den østre sonen av metagråvakke-skiferserien. Snitt langs en sprekk som skjærer Skillerfjell.*

need not have been tectonically reduced. To the east is a relatively large syncline having the younger, Silurian rocks in the trough.

Further eastward tectonic interpretation is based on stratigraphy and on small folds. The Silurian rocks also crop out in the southeastern part of the area and have been interpretated as synclinal folds. In the northeastern part of the area only rocks of the Kjøllhaugene Group crops out, dipping first west and then east. The style of folding seen in the cross-sections (Pl. II, C-D; E-F) is diagrammatic and shows only the general type of folds in the area. Although faults were not observed, many vertical fissures occur and trend east or eastsoutheast. They can be traced for several kilometers. Cleavage is very common, but no measurements were taken as a detailed study of the tectonics in the south of the area has been done by D. Roberts.

Generally, the structure of the mapped area can be interpreted as a fold-compressive type, with the direction of the deformative stress from west to east.

Traces of tectonic phases earlier than the main Caledonian orogeny have not been found.

## Metamorphism

The textures and mineral composition of the rocks indicate that the deposits have undergone a low grade of metamorphism. Cleavage crossing the bedding is well-developed only in the pelitic rocks. The elongation of quartz grains is distinct, especially in the slates and phyllites. In the coarser-grained sediments quartz grains are relatively undeformed; in some parts of the metaconglomerates however, and in a few coarse-grained metagraywackes, the pebbles and quartz grains are elongated. Chlorite and sericite usually show an excellent orientation parallel either to the bedding, or to the cleavage. Chlorite and sericite are probably altered clay minerals, which primarily formed the shale and the graywacke matrix. The new metamorphic constituents of the rocks are biotite and the epidote minerals. Biotite forms the relatively large porphyroblasts and poikiloblastic metacrysts; both are generally parallel to the cleavage.

Carbonates and authigenic quartz are also secondary constituents of the rocks, but their presence could have been caused by diagenesis, not necessarily by metamorphic phenomena. A replacement of silicates by the carbonates could take place during both diagenesis and metamorphism. It is not possible to distinguish between the generations of carbonates. Some of the carbonates have probably originated during the formation of epidote.

The quantities of biotite and epidote seem to increase towards the west, but this increase is not very distinct. This phenomenon could indicate the increasing intensity of metamorphism to the west. The minerals and textures indicate that the rocks belong to the greenschists facies, and perhaps in part to the albite-epidote-amphibolite facies. It is not possible to drawn a boundary-line between the two facies.

The gradual increase of metamorphism toward the west is in agreement with the metamorphic zones described by Goldschmidt (1915) in the area south of Stjørdalen (southern part of the Trondheim region).

## Conclusions

The most important conclusions are as follows:

1) The Upper Ordovician deposits in the map area are developed as the meta-graywacke-slate association which represents a flysch facies and turbidite formation;
2) The metaconglomerates occurring within the flysch facies are also deep marine deposits and show no stratigraphic boundary;
3) The Silurian deposits from the Kjøllhaugene area continue southward through the map area and are a black shale facies.

In the paper sedimentological problems are emphasized. Because the sedimentological observations in the mapped area are of a preliminary character, and because the area is relatively small, no regional conclusions are possible. The continuation and development of sedimentological investigations in the Trondheim region would help to explain in more detail the history of this part of the Caledonian geosyncline, and could also be an useful method in stratigraphic correlation.

Oslo, February 1966.

## Sammendrag

Sommeren 1965 kartla jeg et ca. 150 km$^2$ stort område ved Meråker (se Fig. 1). I området forekommer følgende grupper og ledd:

*Kjøllhaugene Gruppen.*

1) Metagråvake-skifer-serien, bestående av vekslende metagråvaker, skifre og i mindre antall også karbonatholdige metaleirstener (metasiltstones). Metagråvakene består hovedsakelig av kvarts, feldspat, noen få bergartfragmenter, matrix og karbonater. Matrix er en kvarts-kloritt-serisitt-plagioklas blanding. Metagråvakenes sammensetning fremgår av Tab. I. Skifrenes sammensetning

er svært lik metagråvakenes matrix. De karbonatholdige metaleirstener består av kvarts (30—50 %), plagioklaser (2—5 %) og relativt mye karbonater (opp til 30 %). Matrix er som i metagråvakene. Meget karakteristisk for alle disse bergartene er ca. 1—2 mm store biotitt-porfyroblaster. I matrix finnes også epidot og opake mineraler.

Serien viser en del primære strukturer som karakteriserer dannelsesforholdene. Metagråvakene er dårlig sorterte (fig. 7), og graded bedding forekommer ofte. På grensen mellom de graderte lagene finnes sedimentære strukturer (fig. 9, 10, 11, 12, 13) som sannsynligvis er *flowage casts* eller *load casts* og ligner meget på *flame structures* beskrevet fra Skottland (Walton 1956, Kelling og Walton 1957). Strukturer som ligner på deformerte ripple marks og spor av undervannserosjon ble også observert (fig. 11, 14, 15, 16).

2) Metakonglomerater, som forekommer i flere, noen få meter tykke lag i den østlige del av metagråvake-skifer-serien (Pl. II). I konglomeratene dominerer kvarts og kvartsittboller som oftest er opptil 5 cm i diameter, sjelden så store som 20 cm. Bollene blir mindre mot syd, dessuten blir det færre av dem, slik at matrix (som er identisk med metagråvakenes) dominerer mer og mer, og metakonglomeratene avløses ofte gradvis av metagråvaker.

*Slågån Gruppen.*

3) Svartgrå metaleirsten-skifer-serien. Metaleirstenene består hovedsakelig av kvarts, feltspat, matrix og karbonater, blant hvilke er mye dolomitt (statistisk analyse — se s. 40). Skifrenes hovedkomponenter er kvarts, serisitt og kloritt. De gir sort strek og er trolig grafittholdige. I serien forekommer også finkornete, mørk-grå sandstener (se Pl. II).

Serien er en fortsettelse av skifrene ved Kjøllhaugene hvor de siluriske graptoliter er funnet av Getz (1890). Man kan følge denne serien fra Fjergens sørstrand sørover til riksveien til Storlien (se Pl. II).

*Intrusive bergarter.*

4) Konkordante ganger (sills) av gabbro-dioritt. Gangene ble observert bare i den vestlige metagråvake-skifer-serien. De er fra noen få meter til omtrent 100 m tykke. Hovedkomponenter i gabbro-diorittene er amfiboler og plagioklaser. I mindre antall forekommer biotitt, kloritt, serisitt, epidot-mineraler, kvarts, karbonater, titanitt og opake mineraler.

I den undersøkte lagserie ble det ikke observert noen hiatus eller diskordans. Strøket er overveiende mot NNE med fall mot W, i den østlige delen av området delvis mot E.

Følgende hovedkonklusjoner kan trekkes:

A) Metagråvake-skifer-serien representerer dypmarin *flysch facies* og såkalt *turbidite formation*. Konglomerater som forekommer i serien er sannsynligvis avsetninger av deltaer av submarine kanaler. Metagråvake-skifer-serien ligner svært meget på de siluriske avsetninger i Wales og overordoviciske avsetninger i Skottland. Svartgrå metaleirsten-skifer-serien representerer såkalt svart skifer-facies (*euxinic facies, black shale facies*).

B) Metagråvake-skifer-serien med konglomerater svarer sannsynligvis til Øvre Hovin-Gruppe (Øvre Ordovicium). Svartgrå metaleirsten-skifer-serien svarer til Horg-Gruppen (Under Silur). Det er overgang mellom de to seriene. Konglomeratene viser ikke noen stratigrafisk grense.

C) Alle bergarter er sterk overfoldet og danner assymetriske synklinaler og antiklinaler skjøvet østover. Den største synklinalen har siluriske avleiringer i kjernen og den overordoviciske metagråvake-skifer-serien på begge sider (se Pl. II).

D) Sekundære forandringer av struktur og forekomster av epidot og biotitt porfyroblaster tyder på at metamorfosen i området svarer til grønnskifer-facies. Metamorfosen forsterkes vestover og svarer i den vestlige del av området muligens til albitt-epidot-amfibolit-facies.

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*Table 1.* **Mineral composition of metagraywackes**

| | quartz | feldspar | fragments of volcanic rocks | fragments of schists and quartzites | Matrix (< 0,06 mm) | | | | | carbonates | porphyroblasts of biotite | epidote | opaque minerals |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | | quartz | chlorite | sericite | feldspar | | | | |
| Main road to Storlien, S of Teveldal station | 31,4 | 11,2 | 2,0 | 2,8 | 36,3 | 30 | 30 | 30 | 10 | 11,0 | 5,2 | — | 0,1 |
| Little unnamed lake, ca. 1,5 km NNE of Bokhammer ...................... | 36,5 | 9,6 | — | — | 42,3 | 25 | 30 | 35 | ? | — | 8,4 | 1,7 | 1,5 |
| Lillekjerringelva river, upper part ...... | 13,8 | 4,2 | 0,6 | — | 59,5 | 30 | 30 | 30 | 10 | 10,3 | 10,7 | — | 0,9 |
| Main road to Storlien, ca. 700 m from the border between Norway and Sweden .. | 34,7 | 6,5 | 5,0 | 8,7 | 32,5 | 35 | 30 | 30 | 5 | 9,2 | 2,2 | — | 0,2 |
| Unnamed stream W of the Lillekjerring-elva river ...................... | 40,6 | 8,4 | 0,7 | — | 40,2 | 25 | 30 | 40 | 5 | 8,7 | 0,6 | — | 0,8 |
| S shoreline of East Fjergen Lake, near Storfloa ......................... | 37,1 | 3,4 | + | + | 50,0 | 35 | 25 | 30 | 5 | 0,8 | 7,9 | 3,0 | 0.2 |
| Main road to Storlien near first (from E side) gabbro-diorite sill ............ | 35,4 | 8,4 | — | 0,3 | 37,3 | | | | | 3,6 | 15,0 | — | — |
| Valley W of Skillerfjell ................ | 34,6 | 6,6 | 0,9 | 9,0 | 29,5 | 50 | 40 | 5 | 5 | 12,3 | 6,0 | 0,4 | 0,8 |
| Hill ca. 1,5 km SE of Midtsundstøtten .... | 38,0 | 10,2 | — | — | 38,6 | 35 | 20 | 30 | 15 | 9,4 | 3,2 | — | 0,6 |

# Geology of the northernmost part of the Meråker area

By
*Anna Siedlecka & Stanislaw Siedlecki.*

**Abstract**

Four lithostratigraphical groups comprising sedimentary and igneous rocks have been distinguished in the mapped area. There are conformities and gradual transitions between all these groups.

The rocks have been subjected to a low grade of metamorphism, being changed to metabasites and metasediments. They are folded, with fold axes trending approximately NNE. The lithological units form narrow elongated zones dipping to the west. In the eastern part of the area an asymmetrical syncline has been observed.



## Introduction

During two weeks in July 1966 we mapped an area (ca. 75 km$^2$) situated north from the Kjølhaugene mountains, bounded in the west by the Kråkfjell mountain, to the north by the geological map of the Verdalen valley (Wolff, 1960*)) and in the east by the border between Norway and Sweden. As the lakes Sulsjøerne occur within the mapped area, the area will here be referred to by this name.

The stratigraphical and lithological units occurring in the Sulsjøerne area correspond to the units distinguished in 1965 in the other parts of Meråker area. Some lithological differences have been observed and these will be emphasized in the descriptions which follow. The structural geology of the Sulsjøerne area is also in general agreement with that demonstrable in the area investigated in 1965. In view of all these similarities, the description of the geology of the Sulsjøerne area will here be very short.

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*) Wolff, Fr. Chr. (1960): Foreløpige meddelelser fra kartbladet Verdal. N.G.U. Årbok, No. 211, Oslo.

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**Stratigraphy and lithology**

Within the Sulsjøerne area the following strata occur:

*Fundsjø Group*

The rocks of the Fundsjø Group occur in the western part of the area where they form the Kråkfjell mountain. They have not been investigated in detail as only a part of the eastern slope of the Kråkfjell mountain has been visited. This slope is underlain by metabasites and by gabbro-dioritic and granitic rocks.

The metabasites are grey-green in colour and they may, or may not, be foliated. Grain-size is usually ca. 1 mm, sometimes up to 3 mm. The metabasites consist of: amphibole (hornblende), relics of feldspar (ca. $An_{10}$(?)), epidote, chlorite and quartz. In addition titanite and iron-oxides are usually present. These rocks are altered lavas and, in part, probably hypabyssal rocks also.

The gabbro-diorite rocks are dark grey in colour; black and pinkish-white minerals, up to 3 mm in size, may be distinguished macroscopically. Thin sections show that the rock consists of amphiboles and saussuritized feldspar, and is essentially similar to the gabbro-diorite observed as sills within the Kjølhaugene Group in the north-eastern part of the Meråker area.

The granitic rock displays granitic or gneissic textures and is light-grey and pinkish in colour. The main constituents of this rock are mosaic quartz and feldspars, the latter comprising orthoclase, albite ($An_7$), microperthite and microline. Biotite, chlorite, epidote and iron-oxides occur as accessory minerals.

*Sulåmo Group* (ca. 3200 m).

The Sulåmo Group in the Sulsjøerne area is represented by two formations.

a) Dark grey and black slates and phyllites with some metasiltstone intercalations, — ca. 1200 m.

In the slates and phyllites, apart from quartz and phyllosilicates, pyrite, some iron oxides and carbonaceous matter usually occur. In black slates carbonaceous matter is dominant. Under the microscope a secondary cleavage oblique to the primary bedding may be observed in some specimens of phyllite.

Layers of laminated, calcareous metasiltstones occur within the slates and phyllites. The dark laminae in metasiltstones consist of quartz and phyllosilicates, the light bands of quartz and carbonates. Usually both pyrite and iron-oxides are present in these metasiltstones.

Features of metamorphism are seen in the textures of the rocks; porphyroblastic minerals have not been observed.

Within this formation, two narrow (max. thickness ca. 400 m) metabasite horizons also occur: 1) a lower horizon, in the middle of the sequence and 2) an upper horizon at the top. Unfortunately, the part of the area where the metabasites occur is covered by vegetation and consequently they are very difficult to trace. Both horizons disappear towards the south (see geological map, pl. II).

The metabasites of the lower horizon are fine-grained, grey-green in colour and usually foliated. They consist of amphibole (hornblende), epidot, quartz, chlorite and plagioclase, and are interpreted as altered lavas.

In the upper horizon the metabasite is in part analogous to that from the lower horizon and in part coarser grained (up to ca. 3 mm in diameter). The coarser grained type consists of amphibole (hornblende), epidote, chlorite, plagioclase, quartz, carbonates and, rarely, biotite. There is presumably a very gradual transition between the fine-grained and the coarser grained metabasites; they seem to be associated with the same volcanic activity, being partly lavas and partly hypabyssal rocks.

Schistose rocks, adjacent to the metabasite horizons are generally greenish in colour and more compact than other rocks in the Sulåmo Group. They consist either of quartz, chlorite and sericite or of quartz, chlorite, muscovite, epidote and carbonates. Although these rocks are shown on the geological map, too little work has been done for a determination of their precise origin; they are probably partly of sedimentary and partly of volcanic origin.

b) Grey slates interbedded with metasiltstones and metasandstones (ca. 2000 m). On the geological map (pl. II) these are depicted as, 1) slates with intercalated bands of metasandstone and 2) metasandstones with intercalations of slate.

The uppermost slaty complex could, perhaps, be assigned to the next lithostratigraphical unit, the Kjølhaugene Group; a distinct boundary between the Sulåmo Group and the Kjølhaugene Group is lacking.

Grey slates, consisting of chlorite, biotite, sericite and quartz, usually contain small amounts of carbonates. This lithology grades into calcareous metasiltstones and fine-grained calcareous metasandstones. The latter sometimes show traces of primary bedding. A cleavage is not developed in either the metasandstones or metasiltstones.

Coarser grained metasandstones occur rather infrequently in this formation. These metasandstones are very poorly sorted, consisting of abundant fine-grained matrix and scattered larger grains (ca. 1—2 mm in size). The matrix is composed of quartz, feldspar and flaky minerals. Small amounts of carbonates are also present; in some layers larger muscovite flakes occur. Quartz and feldspar form the larger grains and, in addition, rounded, isolated chlorite concentrations are present, the size of which is also ca. 1—2 mm. These concentrations are possibly relics after rock fragments. From these various features it would appear that the described sandstones closely resemble graywackes or sub-graywackes.

*Kjøllhaugene Group* (ca. 2900 m).

The Kjølhaugene Group is represented by metagraywacke-slate association which can be divided into: a) a lower, ca. 800 m, predominantly metagraywacke formation and, b) an upper, ca. 2100 m, predominantly slaty formation. In the latter, two small occurrences of conglomerate have been found.

Metagraywackes and slates are grey-green in colour and show the same characteristic features as in the north-eastern part of the Meråker area (see p. 22 f.). However, the metagraywackes of the Sulsjøerne area are generally finer grained, and because of this, the identification of rock fragments is difficult to determine. Secondary biotite and epidote are common. Sedimentary markings have not been observed.

Conglomerates have been observed north-west of Mærraskarfjell mountain (see pl. II). These conglomerates form two lenses of maximum thickness ca. 2,2 m and ca. 0,7 m. Pebbles of white, yellowish, grey and greenish quartzite, and of white and grey quartz predominate. Subordinate pebbles of porphyry, and of a fine-crystalline albitic rock are present. The pebbles are rounded and are mainly 2—4 cm, rarely up to 10 cm, in size. Shapes of the pebbles are usually spheroidal and ellipsoidal, the long axes of the ellipsoidal pebbles generally being oriented with the plane of bedding, though not showing any alignment.

The matrix of the conglomerates is abundant and its composition is the same as that of adjacent graywackes and slates.

*Slågån Group* (ca. 350m in this area).

The Slågån Group consists of black and dark-grey slates corresponding to the black-grey metasiltstone-slate association in the north-eastern and northern parts of the Meråker area.

A characteristic feature of these slates is their weathering into abundant tabular and prolate fragments. The hills developed within the zone of slates are usually covered by the products of disintegration and contrast markedly in morphology with all other hills and mountains in the area, formed out of the previously described rocks, which have relatively little weathering material.

The slates and phyllites of the Slågån Group consist mainly of quartz, carbonaceous matter and chlorite. Biotite porphyroblasts are common.

The Slågån Group is the youngest group in the whole Meråker area.

### Remarks concerning the structural geology

All the stratigraphical units form narrow, elongated zones, trending NNE—SSW. The oldest division occurs in the western part of the area and the youngest near the border between Norway and Sweden.

The strike is usually about 025° in the western part, and about 035° in the eastern part of the area. The beds dip 35°—50° W. In the eastern part of the area beds dipping to the east have been observed but this is only of local occurrence. There is a conformity between all stratigraphical units.

Little work has been done on tectonic structures. In the easternmost part of the area it is possible to trace a syncline with the grey-black slates of the Slågån Group in its core and the grey-green metasediments of the Kjølhaugene Group along the limbs. Minor folds and corrugations are frequently present, more especially in the slates and phyllites, and a secondary cleavage related to these folds is often quite prominant. Faulting has not been observed.

Trondheim, February 1967.

### Sammendrag

Sommeren 1966 kartla vi et ca. 75 km$^2$ stort område ved Sulsjøerne (se Fig. 1). I området forekommer vulkanske og sedimentære bergarter av Fundsjø-Gruppen, Sulåmo-Gruppen, Kjølhaugene-Gruppen og Slågån-Gruppen. Bergartene viser samme karakteristiske trekk som i andre deler av hele Meråker-området.

I den undersøkte lagserie ble det ikke observert noen hiatus eller diskordans. Strøket er mot NNE med fall mot N. En svak metamorfose forårsaket sekundære forandringer i bergartenes struktur og dannelse av biotitt porfyroblaster.

# Structural observations from the Kopperå-Riksgrense area and discussion of the tectonics of Stjørdalen and the N.E. Trondheim region

*David Roberts.*

**Abstract**

Minor tectonic structures occurring in weakly metamorphosed Hovin and Horg Group sediments along Teveldalen, east of Meråker, are described in the first part of the paper. Three episodes of folding are recognized the respective folds showing differences in style and axial trend. First phase minor fold axes and lineations reveal a large but systematic variation of trend, generally between S.W. and N.W. The one major structure, the Teveldal syncline, is shown to have been produced during the second movement episode. Third phase structures include a penetrative cleavage axial planar to abundant minor folds whose axes trend consistently between N. and N.N.E. Thrusting of the metasedimentary pile in an E-SE direction is related to the concluding stages of the second generation of folding.

The tectonics of the Stjørdalen valley area west of Meråker are then described and particular reference made to the fundamental structure, a fan-anticline here called the Stjørdalen anticline. This anticline, a first generation structure, is seen to dominate the tectonic picture in the northern Trondheim region; in all probability it can be traced further south within the central part of the Trondheim 'depression'. A comparison of the fold episodes in the Teveldalen and Stjørdalen areas is made and minor structures reported from neighbouring districts are considered in relation to the present sequence of fold movements.

The last section deals with proposed major structural correlations across the northern Trondheim region. The Hegsjøfjell area, mapped by S. Foslie and described briefly by J. S. Peacey (1964), reveals a major example of refolding. With a re-interpretation of the nature of the early isocline and further examination of the changing attitude of the Verdal synform (Peacey 1964) and structures across the Verdal region (Wolff 1960), a correlation of (a) the Stjørdalen anticline with the early Hegsjøfjell fold and (b) the Teveldal syncline with the Verdal synform now appears quite acceptable.

Finally, consideration of the thrusting in the Tømmerås-Hegsjøfjell and central Trøndelag-Jämtland areas would tend to support Strand's (1961) suggestion that the metasediments of the extensive Trondheim region are allochthonous. In this regard it is quite possible that Peacey's (1964) upper nappe may be traced down to the southern part of the Trondheim region (see Wolff, present volume).

## A. THE KOPPERÅ - RIKSGRENSE AREA

### Introdduction

During a study of the structural geology of an area between Meråker and Hegra in the summer of 1965, time was spent on adjacent ground in order to establish a more complete regional structural picture. The present account largely concerns the minor structural observations made in the main-road traverse from near Kopperå up to the Swedish border (riksgrense), a tract of ground which forms the southern boundary of an area mapped by Dr. Anna Siedlecka (1967, see accompanying paper in this NGU volume).

Weakly metamorphosed clastic sediments regarded as being of Upper Ordovician and Lower Silurian age occur within the traversed area. Only two major stratigraphical groups are represented ,the Kjølhaugene Group and the Slågån Group (Siedlecka, 1967), these in all probability being equivalent to the Upper Hovin and Horg Groups respectively. The Slågån Group is the younger of the two and its shales can be followed north-north-eastwards along the strike (Siedlecka 1967, Chaloupsky and Fediuk, 1967) to the eastern part of the Kjølhaugene mountains, where a graptolitic fauna shows it to be of Lower Llandoverian age. The eastern limit of the area is marked by a tectonic break, here called the Grense thrust, subjacent to which are quartzites and gneissic schists of Eocambrian age.

Lithologically the Upper Hovin (Kjølhaugen) Group comprises metagraywackes and graywacke-sandstones alternating with grey-green, chlorite-sericite phyllites or sometimes fine-grained phyllitic siltstones. These rock-types are present on both minor and major mappable scales. Where for instance phyllite predominates, metagraywacke is invariably present as subordinate intercalative bands; the converse is also generally true. In the extreme north-west an horizon of greenstone partly with amphibolite and tuffitic greenschist separates the Upper and Lower Hovin Groups. The Horg (Slågån) Group consists largely of dark shale or slate with some metasiltstone and sandstone bands. Sedimentary structures are fairly common occurring mainly in the Upper Hovin Group and provide good evidence of younging. Such features, here graded bedding, load casts and flame structures (Kuenen 1953, Walton 1956) and clasts of shale incorporated in the basal parts of metagraywacke bands are amply documented by Siedlecka (1967).

Throughout the whole length of this traverse of Upper Hovin and Horg Group rocks, indubitable evidence of at least two generations of minor folds can be demonstrated. In addition to their marked differences of style the two minor fold types exhibit notable disparities of axial trend and axial planar attitude. The relative time sequence of folding is proven by the many examples of structures of the later movement episode deforming those belonging to an earlier episode.

For the purpose of this description the fold episodes will be referred to as 'early' and 'late'. It can be shown, however, that the early period of deformation is capable of being divided into two phases. Minor folds of both phases are recognisable although in this small area minor folds of the older phase are by far the more abundant. In the text which follows, the phrase "early minor folds" will refer to minor folds of this older phase.

In general the strike of the rocks is N.N.E.-S.S.W. with beds dipping towards the west-north-west. The major structure dominating the geology of the area is that of an overturned, tight syncline containing the Horg (Slågån) Group in its core. This fold is shown to be a younger phase structure of the early episode of deformation. While the axes of the late folds plunge fairly consistently at some small angle towards N.-N.N.E., those of the minor early folds show far greater variation, trending systematically between S.W. and W.N.W. or even N.W. Furthermore a relationship is apparent between the amount of plunge and the axial direction in the case of these earlier structures.

It should be mentioned that a compass graduated to 360° was used during this present study: all quoted dip and plunge measurements and compass bearings are therefore based on this scale.

### The early minor folds and related structures

Evidence for the occurrence of structures belonging to this early generation of folding is almost ubiquitous. Where folds are lacking related features such as rodded quartz and associated linear structures preserve the identity of a fold episode clearly older than that responsible for the more open later structures which also deform these metasediments.

Invariably the early folds are tight or isoclinal in style (Figs. 32 and 33), the inter-limb or dihedral angle varying within the range 0—25°. In the

Fig. 32. Early isocline deformed by late folds. Interbanded phyllite/metagraywacke, Upper Hovin Group. Teveldal road 1 km south of Kopperå.

*En tidlig isoklinalfold deformert av senere folder. Vekslende fyllitt/gråvakke, øvre Hovingruppe. Mellomriksveien 1 km syd for Kopperå.*

eastern parts of the area a tendency is manifest for a slightly less acute style and such folds may be described as tight to close after the terminology proposed by Fleuty (1964). Style variations dependent on lithology are also demonstrable.

The early folds are clearly of shear origin as evidenced by their similar nature, and lithologies are often thinned or sometimes completely sheared out along limbs. This applies to both phyllites and graywackes. Conversely, fold closures are usually thickened with phyllite frequently simulating an accommodating medium.

Although its axial planar nature is not always perceptible the schistosity displayed by the phyllites can undoubtedly be attributed to this early deformation and folding. A closely interbanded phyllite-graywacke sequence is the most favourable lithology for observing the axial planar schistosity and then, quite often, only when fold closures are present thus displaying the relationship of the S-planes to maximum effect. Where fold closures are absent the schistosity frequently appears to parallel the bedding; alternatively a difference of two or three degrees between bedding and schistosity may be observed on favour-

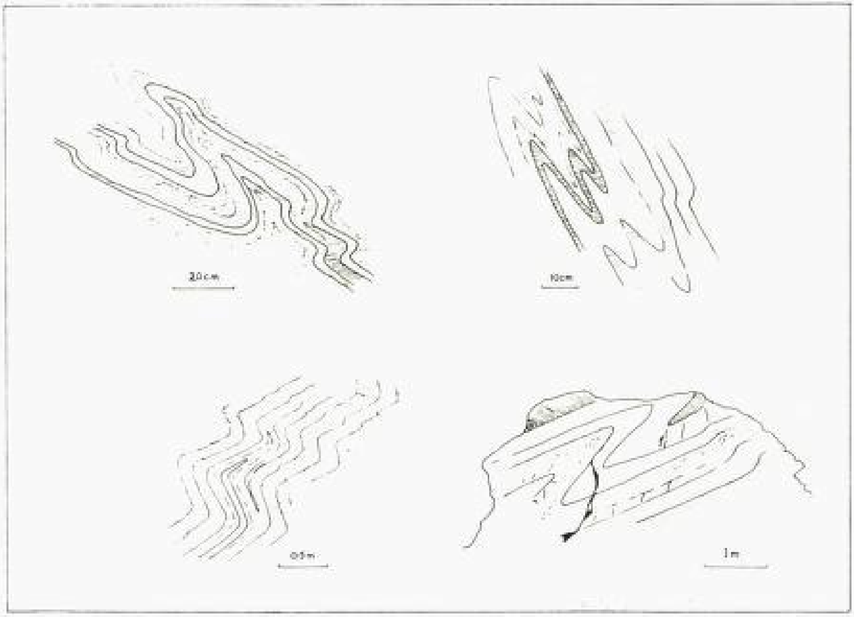



Fig. 33. Early folds deformed by late structures. Interbanded phyllite/metagraywacke, Upper Hovin Group. Teveldalen.

*Tidlige folder deformert av senere bevgelser. Vekslende fyllitt/gråvakke, øvre Hovin-gruppe. Teveldalen.*

able erosion and joint surfaces. With an increasing fold dihedral angle this difference of attitude between bedding and schistosity generally becomes more apparent and a slight fanning of the latter may be evident, whilst in the case of the uncommon, more open, early folds in massive graywacke the cleavage or schistosity may fan quite noticeably around the fold closures. In this massive graywacke early folds sometimes show a tendency more towards a concentric than similar style, in apparent disagreement with the generally accepted shear origin for this generation of structures.

Linear structures, including fold axes, referable to the early deformation episode are here divided into two groups; an uncommon abnormally trending group is discussed later whilst those which are common and pervasive are dealt with immediately below.

Early fold axes and parallel linear structures display an appreciable variation of trend (Fig. 35); the variation is shown to be systematic and from the regional geological point of view is of considerable interest. The lineations generally

Fig. 34. Late cleavage cutting across early folds. Mixed phyllite/greenschists, basal Upper Hovin Group. Turifoss bridge, Teveldal road.

*Sen kløv som skjærer igjennom tidligere folder. Blandet fyllitt/grønskifer ved basis av øvre Hovingruppe. Turifoss bru. Mellomriksveien.*

lie in the trend range 220°—305°. Plunges of these early lineations also show a systematic variation which is dependent on the linear trend. In the north-west of the traversed section for example, the early fold axes and lineations plunge W.N.W. at 25°—36°. Since the axial direction is here trending almost normal to the general strike of the metasedimentary banding and schistosity the angle of plunge approximates to that of the dip.

Moving south-eastwards along the main road the trend of the first generation folds gradually swings from west-north-west through west to west-south-west. Accordingly, as the strike and dip remain fairly constant, the axial plunge now shows a smaller angle. Further south-east approaching the Grense thrust, early fold axes swing into sub-parallelism with the thrust which hereabouts strikes at 054°—058° and dips at some 45° to the north-west. In this area with the early lineations now diverging at only a small angle from the strike, plunges show values in the range 5°—18°.

Although a few irregularities do occur largely due to the local effect of late folds, it must be stressed that this change of linear trend is a noticably gradual

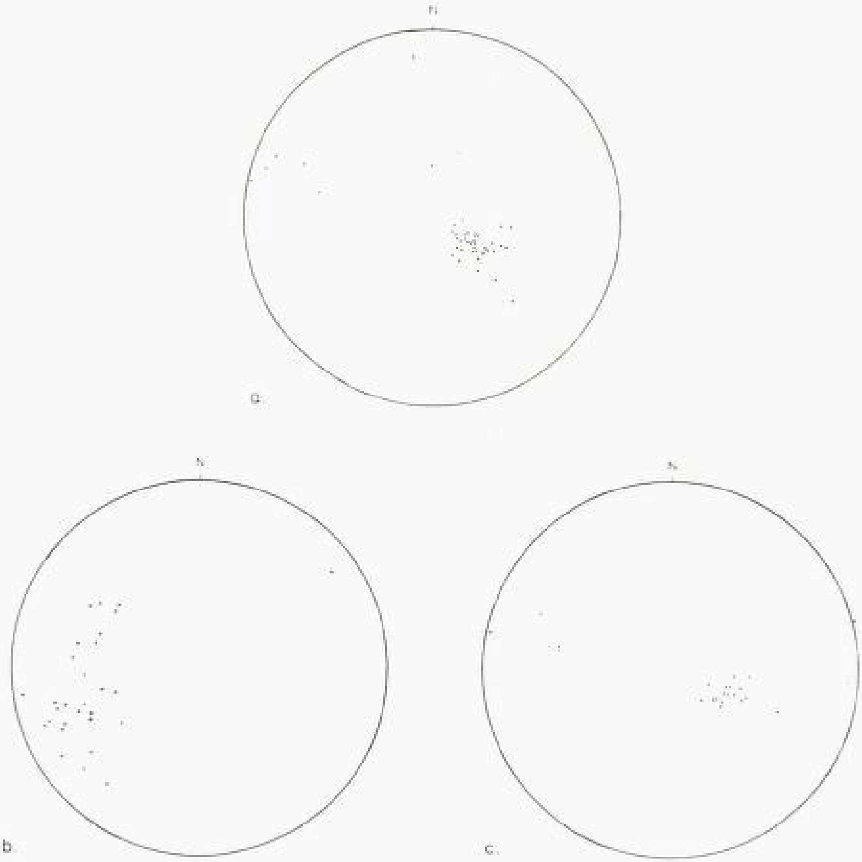



Fig. 35. Stereographic projections of structural data (Wulff net, lower hemisphere) (a) Poles to bedding planes. (b) Early fold axes and related lineations. (c) Poles to axial planes of early folds.

*Stereografisk projeksjon av strukturer eller data (Wulff's nett, undre halvkule). a) Poler til lagflater, b) tidlige foldeakser og tilhørende lineasjoner, c) poler til akse plan for tidlige folder.*

one. Within a relatively short distance along a line almost normal to the Grense thrust early fold axes and lineations swing through ca. 60°.

A co-existence of early linear structures markedly oblique to one another is uncommon. Locally however, such a variation in the trend of linear structures which pre-date the late fold episode may be seen within the limits of one road-cutting in strata with constant dip and strike, although the precise mutual relationship of the two linear elements has nowhere been observed.

in such rare occurrences the 'abnormally' trending lineation (here N.E.-N.N.E.) is represented by the axis of a tight or close fold, while axes of isoclinal or near-isoclinal folds and quartz rods constitute the normal lineation. Axial planes of the N.N.E. trending folds appear to parallel those of the more common isoclinal folds, but a difference is seen in that a minute puckering of phyllite laminae may be discernible at the hinges of the folds with 'abnormal' trend. It would seem therefore that the folds of N.N.E. trend have been generated at a slightly later stage in the deformation sequence than the more pervasive early structures.

Immediately south-east of the Slågån Group the folds which there pre-date the late structures trend between 020°—048° and plunge towards the south-west or south-south-west at 0°—17°. These folds are generally less acute than early folds encountered on the upper limb of the major syncline with fold dihedral angles locally up to 60°.

The normal approximately E.N.E.-W.S.W.-oriented early lineation is here indiscernible, yet it is weakly developed in the shales of the Slågån Group and becomes prominent again some few hundred metres south-east of this group. Further east, isolated examples of more open, less asymmetrical folds may be found: these also pre-date the later movements while at the same time deforming the earlier developed schistosity. Towards the north-east along the strike, such relatively open folds become more prominent (see Chaloupsky and Fediuk's suggested profile, Pl. II). From this various evidence it would seem permissible to divide the early folding into two phases.

Other significant structures belonging to the earliest deformation phase include quartz-rods, diffuse striations and minute plications or crinkles. Rodded quartz is quite common tending sometimes to be profuse in the more phyllitic lithologies; it is much less frequently developed in massive metagraywacke. Where a mixed phyllite-metagraywacke sequence prevails, such rods are almost entirely restricted to the phyllite bands. They form prominent features on weathered surfaces owing to the relative resistance of the quartz to erosion.

A similar lithological control in the development of quartz-rods was described by Wilson (1953) from the North-West Highlands of Scotland. In this Scottish example the quartz-rods are often abundant in incompetent, semi-pelitic strata and generally lacking in siliceous granulite horizons.

Rods are essentially monomineralic consisting of quartz in large or irregularly sized grains, although they occasionally contain a little calcite. In transverse profile they may be oval, near-circular or irregularly lenticular depending in part on whether the quartz is of vein or segregatory origin but also on the extent to which shearing has affected the particular host lithology. At times it is impossible to ascertain the initial form — vein or segregation — of the quartz rods moreso when they occur in relative isolation. Near-planar or flatly lenticular quartz-veinlets whose occurrence and formation was controlled by the stratification would appear to lend support to a vein origin. These may or may not be highly deformed at early fold closures. On the other hand quartz-pods and lenses which are drawn out parallel to the early fold axial planar schistosity are thought to be quartz segregations developed coevally with deformation and metamorphism. Some of these may however, be fragments of deformed veinlets: the field evidence here is frequently ambiguous. The Ben Hutig phenomenon (Wilson 1953) of some segregatory quartz tending to concentrate in the reduced-pressure zones of fold-apices was observed on only a restricted scale in the Kopperå—Riksgrense area although further west, beyond Gudå, such hinge-zone quartz segregation is more pronounced in higher grade schists.

The significant tectonic feature of these quartz rods is that they are demonstrably parallel to the axes of the earliest folds. Occasionally where folds are locally absent, quartz rods or an associated streaky lineation aid in the recognition of the early linear element. On some bedding planes in massive metagraywacke diffuse clots of quartz and calcite parallel this early linear direction.

A close connection between fold axes and quartz-rodding was first recognized by Peach and Horne (1907), again in the Ben Hutig area of North-West Scotland. This occurrence formed the basis of Wilson's (1953) paper in which he emphasized this relationship of the plunge of quartz-rods to that of fold axes and other linear structures and stated that such rods lie at right-angles to the tectonic movement direction. From the structural symmetry point of view rods are monoclinic linear structures with their elongation normal to the plane of symmetry. The rods thus constitute a *b*-lineation: the tectonic coordinates *a*, *b* and *c* as used here are those defined by Jannetaz (1884) and Sander (1930).

In the present area an examination of any isolated exposure would show the rodding as a local *b*-lineation, paralleling as it does the earliest fold axes, but when this is considered in the overall regional setting both within and beyond the traversed area, the tectonic pattern is decidedly more complicated. While the main direction of tectonic movement in the region is known to be towards an east-south-easterly or south-easterly point, this is seemingly incompatible with the evidence presented here of gradually swinging minor fold axes and lineations if one adheres to the 'lineation perpendicular to movement' tenet of the Sander school. Clearly, alternative hypotheses must be considered and are discussed later.

Boudined greenstone and greenschist bands and tectonic inclusions of greenstone occur within the predominantly metagraywacke sequence at the base of the Upper Hovin Group in the north-west of the traversed area. Boudin elongation where recognisable and measureable is again parallel or sub-parallel to the early fold axes. Metagabbroic sheets and lenses occur quite commonly in the Upper-Hovin sequence moreso within the western limb of the major syncline. Although only cursory observations were made it appears evident that these gabbros were deformed by the earliest generations of folds. Details of petrography and field relationships of the metagabbros appear in the accompanying papers of Siedlecka and Chaloupsky/Fediuk.

### The Teveldal syncline

Only one early fold of major proportions can be demonsctrated in this area. This fold, which dominates the structural picture (Fig. 36), accounts for the present disposition of dips and bedding. Horg (Slågån) shales and metasiltstones occupy the core of the fold and are flanked by the metagraywacke-phyllite sequence which, from sedimentary structural evidence, youngs towards the Horg Group. The fold would therefore appear to be synclinal — it is also overturned, asymmetrical and tight or near-isoclinal — and is here called the Tevel-

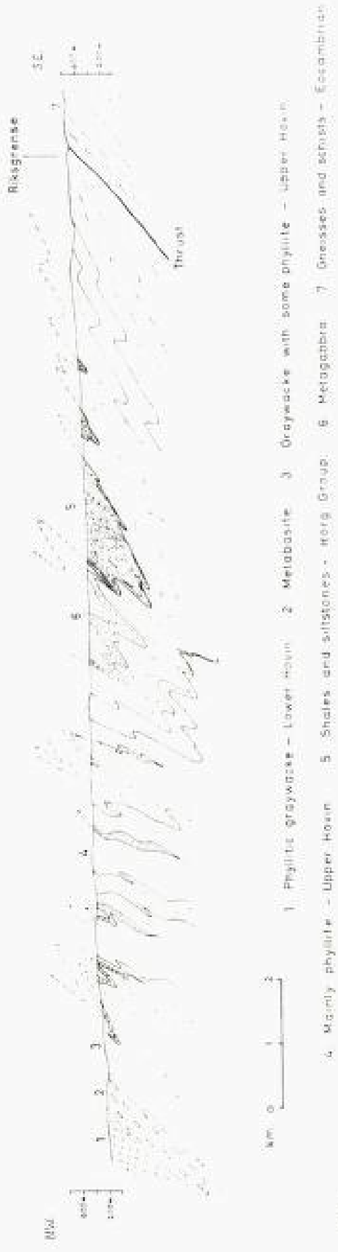



Fig. 36. Simplified geological profile along Teveldalen.
*Forenklet geologisk profil langs Teveldalen.*

dal syncline. South-east from the Horg Group the stratigraphy is for the most part the correct way-up although two narrow zones of shale of identical lithology to that of the main Horg rocks occur on this normal limb and very probably represent the cores of smaller folds congruous to the main structure. Sedimentary structures show that this interpretation is correct for one of these shale bands: poorly exposed ground precludes a more accurate assessment of the second strip of shale.

The width of outcrop of Horg rocks is remarkably constant north-north-eastwards for some 25 km or more to Kjølhaugene. The same rocks are found further north in the Mærraskarsfjell area (Professor S. Siedlecki and Dr. A. Siedlecka, personal communication), but do not occur in the Insvatn-Vera-vatn region of the Verdal map sheet, mapped by Wolff (1960), which is characterized by Upper Hovin metagraywackes and phyllites. It is more than probable therefore that, discounting a possible influence of faulting, a fold closure of Horg rocks exists in the intervening unsurveyed area.

Direct information on the plunge of the main closure of the Teveldal syncline is wanting. Lack of exposures in the tract of ground underlain by the Slågån Group in the Teveldal valley is probably a consequence of the poorly resistant nature of this particular pelitic lithology. Areas in this extensive Meråker region have been mapped largely from a general geological angle so that insufficient attention has been paid to tectonic structures. On the various maps it is therefore usually impossible, from an examination of fold symbols, to distingush one fold phase from another, moreso where folds of sub-parallel trend but different age are thought to exist.

An indirect determination of the attitude of the main fold axis is quite possible however. Since the width of outcrop of the Slågån Group remains reasonably constant when traced north-north-eastwards and dip values show no great variations, a major fold axis disposed near to horizontal can be conjectured. The absence of these Silurian rocks on the Verdal map sheet implies their probable discontinuation at a fold hinge: a gentle axial plunge towards a south-westerly point is then likely. South of the Teveldal valley the mapping is incomplete but the Slågån Group again shows a fairly constant width of outcrop (see Plate III, this volume) though with a possible widening in the Storhusmannsberget area (Z. Pelc, unpublished map 1966). In the Teveldal valley the folds described earlier occurring to the south-east of the Slågån Group and which, though pre-dating the late deformation, are less acute in style than the earliest minor folds, are congruous to the main syncline. Their small or negligible plunge towards a south-westerly of south-south-westerly point is in all probability a fair reflection of the attitude of the major fold axis.

The first reference to a major fold in this Kopperå-Riksgrense area was made by Kjerulf (1875) who, on the basis of local easterly dips on Kjerringfjellene (some 9—12 km north of the Teveldal highway), regarded the structure as anticlinal. He stated that, "en antiklinal linie kan følges fra Store Kjærringå (nær jernbanelinien) i Retningen n.n.o., vest under Kjærringfjeldene vest for Halsjø gjennem de stærke foldinger i Kolkjøndalen og videre øst under Kjølihaugene". His 'anticlinal line' thus follows the outcrop of the Horg Group shales. It is interesting to note here that Kjerulf quotes O. Schiötz as having observed two different schistosity or cleavage directions in this general area.

Törnebohm (1896 — Fig. 56 and Profile 1, Pl. 4) interpreted this same fold as being synclinal although his reasons for doing so appear to be based on what is now known to be an incorrect stratigraphy. From the map and lithological descriptions, it is clear that his Sul Schist Group in the eastern Trondheim region involves a mistaken correlation of the present-day Lower Hovin and Horg Groups.

Högbom (1909) followed Törnebohm in advocating a synclinal fold for the Kjølhaugene area. His section (Pl. 7, fig. 11) is very similar to that published in Törnebohm's important memoir.

Prior to these early interpretations of the main Teveldal fold, Hørbye (1861) noted that to the north of Teveldalen in the general zone now largely referable to the Horg Group, small folds were particularly abundant and dips highly irregular but he made no reference to any large scale structure.

In his paper of 1919, C. W. Carstens indicated the presence of a synclinal fold in the Meråker-Storlien area. On his fig. 1, plate 18, this syncline appears as the complementary fold to a larger anticlinal structure further to the west.

No further mention of a major fold can be traced in the literature until Holtedahl (in Bailey and Holtedahl, 1938) published a comprehensive map of the "Scandinavian Caledonian Zone" on which a syncline is drawn between Meråker and the Swedish border. A similar opinion was held by Bugge (1954).

In a schematic profile drawn from Trondheimsfjord to Storlien, Wolff (1964) follows Carstens in postulating a major syncline east of a larger anticlinal structure, the syncline containing Horg rocks in its core. His argument for this interpretation was based largely on lithostratigraphical correlations. Recent mapping, including the important discovery of a conglomerate (the Lille Fundsjø Conglomerate — see the paper by Chaloupsky and Fediuk) structurally below but stratigraphically above a volcanic series (= Støren Group), has served to strengthen these views.

Fig. 37. Late folds with associated axial planar cleavage. Early isocline present just above hammer-head. Phyllite, Upper Hovin Group. Teveldal road, south of Kopperå.

*Sene folder med tilhørende akseplankløv. En tidlig isoklinalfold sees like over hammerhodet. Fyllitt, øvre Hovingruppe. Mellomriksveien, syd for Kopperå.*

**The later folds**

Although large-scale folds have not been encountered, minor structures belonging to this later tectonic episode can be observed in virtually every roadside outcrop. While minor folds are fairly common the most conspicuous structure is a cleavage which is generally axial planar to the folds and often penetrative.

Many examples of earlier folds deformed by later minor structures are demonstrable (Figs. 32, 33 and 34) the axial-plane cleavage cutting incongruously across the limbs and schistosity of pre-existing folds. Frequently this later cleavage becomes the dominant plane of fissility in the more pelitic rock-types, particularly in those of the Slågån Group. In pelite and psammite alike it may sometimes simulate a major joint.

Concentric folding is characteristic of this episode of deformation. In the more psammitic lithology, minor folds are generally open and asymmetrical with small amplitude relative to wavelength. The perceptible sense of overturning is down-dip, axial-plane cleavage being inclined towards an easterly

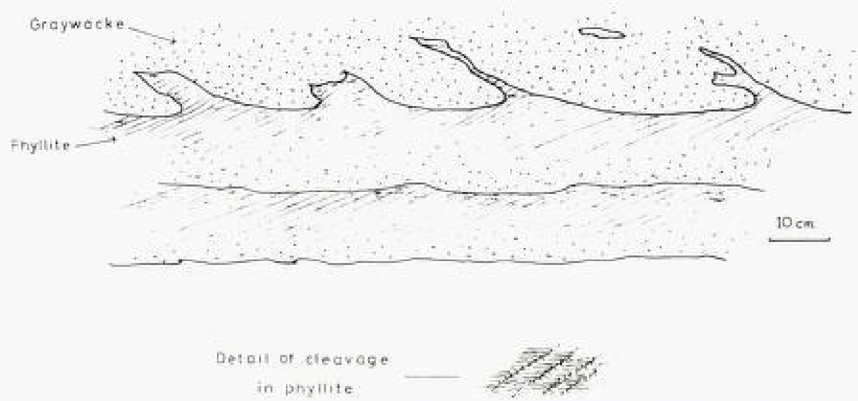



Fig. 38. Load casts, flame structures and late cleavage. Upper Hovin Group. Teveldal road.
*Pålastningsavstøpninger, flammestrukturer og sen kløv. Øvre Hovingruppe. Mellomriksveien.*



point. In a few exposures late minor folds approach a chevron style wherein fold wavelength approximates to amplitude. Such chevron or zig-zag folds, with their straight limbs and fairly sharp hinges, have geometric properties of both concentric and similar folding. The late cleavage, though axial planar to its associated folds, is principally a fracture cleavage and may exhibit a slight fanning around minor fold closures. There is no recrystallization of minerals parallel to this cleavage. Folds are invariably more abundantly developed in the pelitic rock-types, often appearing as a rucking or crumpling of the earlier schistosity with an associated cleavage axial planar to the microfolds.

At one locality where load-casts in metagraywacke protrude into underlying phyllite, the phyllite exhibits a penetrative cleavage which also affects the flame structures between the load-casts (Fig. 38). On close scrutiny the cleavage is seen to be axial planar to late microfolds of some 1—2 mm wavelength, the shorter limbs having been converted into cleavage planes.

An examination of thin-sections of this phyllite verifies the field evidence: the late cleavage is entirely mechanical, deforming the early metamorphic fabric without any concomitant recrystallization or new growth of minerals. No diaphthoretic phenomena were observed, though several of the planes characterized by more intense movement are stained with a brownish-red oxidation product. A progressive development of the cleavage can be traced within any one thin-section, the ultimate stage testifying to a shearing-out of the short limbs of microfolds: in this, the lepidoblastic sericite of the pre-existing fabric is perfectly parallel to the new cleavage plane but is devoid of any alteration or recrystallization.

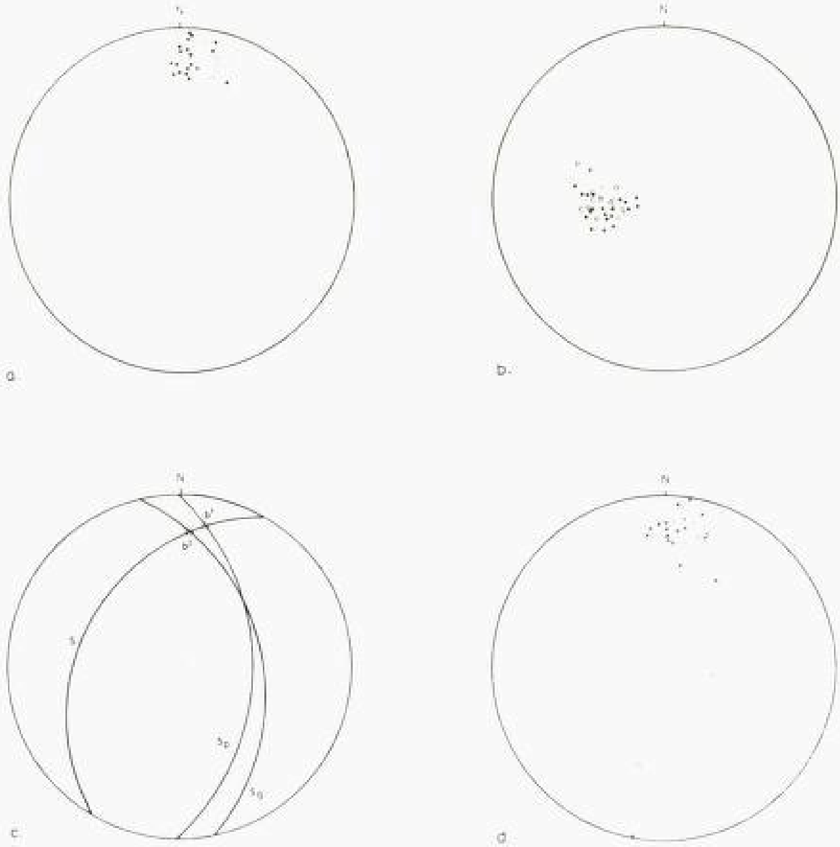



Fig. 39. Stereographic projections of structural data (Wulff net, lower hemisphere). (a) Late fold axes. Circles are phyllite, dots metagraywacke. (b) Poles to axial planes of late folds. Circles phyllite, dots metagraywacke. (c) S-planes in phyllite/metagraywacke. S — bedding plane, Sp — axial plane to late folds in phyllite, Sg — axial plane to late folds in metagraywacke, $b^1$ — intersection of S and Sp, $b^2$ — intersection of S and Sg. (d) diagram of S/Sp and S/Sg intersections. Circles phyllite, crosses metagraywacke. *Stereografisk projeksjon av strukturelle data (Wulff's nett, undre halvkule). a) Sene foldeakser, sirkler er fyllitt, prikker er metagråvakker, b) poler til akseplan for sene folder. Sirkler er fyllitter, prikker er metagråvakker, c) S-plan i fyllitt/metagråvakke. S — lagflater, Sp — akseplan for sene folder i fyllitt, Sg — akseplan for sene folder i metagråvakke, $b^1$ — skjæringslinje mellom S og Sp, $b^2$ — skjæringslinje mellom S og Sg. d) diagram av S/Sp og S/Sg skjæringslinjer. Sirkler fyllitt, kryss metagråvakke.*

Unlike the fold axes of the earliest generation of structures, late fold axes are relatively constant in their orientation plunging regularly at 5°—18° towards 0°—18° with occasional departures to 355°. This N.-N.N.E. trend is a consistent feature of the late fold generation in areas west of Kopperå and Meråker also. Clearly the orientation of the later linear element is markedly different from that of the earliest deformation. While refolding, with its attendant change of orientation of early fold axes and lineations, is demonstrable on a minor scale, it is not present on a large or regional scale in this area.

Cleavage refraction is quite common in the interbanded graywacke and phyllite. It is most prominent on a minor and outcrop scale but can also be demonstrated to exist between mappable lithological units. The effects of this lithological control of late structures are seen not only in variations in the inclination and strike of the later cleavage but also in a small but noticeable divergence of fold axes. This is readily apparent from fig. 39; in general, fold axes in metagraywacke plunge more or less due north whereas the majority of axes measured in phyllites and shales plunge more towards a 006°—018° direction.

Another slight variation in the orientation of late fold axes is a direct result of the refolding of a previously deformed succession. The geometrical complexity of linear patterns arising from polyphase movements is now well known largely due to the work of Ramsay (1958a, 1958b, 1960 and 1962) and Weiss (1959). An appreciation that the orientation of later fold axes will be expected to vary according to the dip of the banding which has undergone refolding is particularly important in this respect. In the present area the late folds affect both limbs of the pre-existing tight syncline. Variations in late fold axial trend would therefore appear likely but account also has to be taken of other factors such as the shape of the early fold and the angle between the surface being folded and the axial plane of the new folds.

From a consideration of these variables in the present area, only very small variations in the orientation of the later folds are likely. Taking the observations of late fold axes on the inverted western limb of the major fold, the orientation averages out at 008° whereas the average for identical axes measured on the shallow eastern limb is 003½°. Plunge values are 11° and 12° respectively. Thus, the theoretical slight variation is confirmed but it must be pointed out that a greater number of measurements are required before a final assessment of the linear divergence can be made.

When $\beta$-points (intersections of s-planes) are constructed from bedding and late cleavage field data, they correspond closely to the observed late lineation trend. This construction is helped in its precision by the fact that throughout this traversed area the late cleavage is nearly orthogonal to the bedding; thus, errors of construction are minimal and spurious points absent or negligible (see Ramsay 1964). Using this $\beta$-diagram technique it was found that $\beta$-points constructed from field measurements in psammitic and pelitic lithologies (Fig. 39) were in close agreement with actual measurements of the respective fold axes and were most useful where folds are locally absent. The divergence of trend related to lithology is thus corroborated.

A few hundred metres west of the traversed section between Kopperå and Meråker, conjugate late folds are fairly common in WNW-dipping Lower Hovin phyllites. These are minor structures, occasionally small enough to be categorized as kink folds and kink bands. Locally the folds overturned towards the south-east (up-dip) are predominant. Important differences of axial trend are inherent in these conjugate folds, axes of the two sets diverging by up to 27° in any one exposure. While the axes of minor folds with axial planes dipping to the north-east plunge towards 0—005° at some 20°—23°, those with axial planes dipping steeply to the west-north-west plunge towards 022°—27° at 10°—13°. This disparity is substantiated in diagrams of the intersections of bedding and the respective cleavages: the constructed kinematic b-axis approximately bisects the conjugate axial angle.

From these observations of conjugate structures it would seem that considering the late fold movement picture, orthorhombic symmetry locally obtains in a region characterized by monoclinic symmetry. However, the intersection of the complementary axial planes of the conjugate folds — the kinematic b-axis — does not lie in the lithological layering and there is thus a non-coincidence of the kinematic and symmetrological co-ordinates. From the point of view of geometry these folds would therefore appear to have a lower, triclinic, order of symmetry (see Ramsay and Sturt, 1963).

## Joints

Joints were largely excluded from the present study since time did not permit the systematic measurement necessary for their inclusion in any comprehensive structural synthesis.

While joints clearly 'ac' to the later folds were prominent locally as well as in parts of the larger area mapped by Siedlecka, these do not form any pronounced maximum on a stereogram. In some exposures joints are present which look to be transverse to the axes of folds belonging to the younger phase of the first deformation episode. Conjugate joint pairs were

Fig. 40. Tension gashes in massive psammite, Upper Hovin Group. 600 m west of Flaten settlement. Teveldal road. Hammer shaft parallel to the bedding.

*Tensjonssprekker i massiv sandstein (psammite), øvre Hovingruppe, 600 m vest for Flaten gård. Mellomriksveien. Hammerskaftet parallelt med lagningen.*

frequently observed; these were quite often noted to strike obliquely to the late folding tectonic co-ordinates but would nevertheless appear to be related to this late folding. Certain of the conjugate joints, one set of which may be better developed than the other, may be infilled with quartz and are discussed more fully below. A jointing direction is also sometimes represented by the penetrative late cleavage.

Towards the east, moderate-to low-angle joints dipping north-westwards become fairly conspicuous. Many of these are parallel or sub-parallel to the Grense thrust in this area, but other oblique low-angle joints are also present. Minor displacements, including both normal and reverse relationships, can sometimes be observed and one clear example of near-horizontal movement along a joint plane is demonstrable. This particular joint, or minor fault, dipping at 54° to the north-east contains a 1—2 cm vein of quartz which shows a pronounced linear grooving akin to slickensides. This linear element makes an angle of only 4° with the horizontal, and steps in the grooving indicate the north-eastern block to have moved S.E. relative to its south-western counterpart. Observations of late folds, cleavage and conjugate joints at this locality strongly suggest that this minor fault is related to the late fold episode.

In a fairly massive metasandstone lithology just west of Flaten (Teveldal) settlement, tension gashes are particularly prominent (Fig. 40) and many are infilled with quartz or quartz-calcite, thus constituting gash veins. The lithological sequence is here the correct way-up, each psammite unit grading perceptibly upwards into a finer grained metasiltstone. It is interesting to note that this lithological change is reflected in the development of these tension fractures since many such features, conspicuous in the coarser sandstone, do not penetrate the finer siltstone or other shaly bands. On the other hand the tension gashes terminate abruptly at the base of any one graded unit.

It would appear that, considering the probable stress distribution responsible for their development, the tension gashes are related to the later generation of structures. While the bedding dips W.N.W. the late cleavage is inclined constantly towards an easterly point and the related folds, not visible in this actual psammite, are everywhere overturned down-dip, towards the west or west-north-west. The tension gashes can therefore be regarded as resulting from the imposition of a shearing couple (Fig. 41a) acting along the top and bottom of each individual lithological unit. This itself would appear to be associated with bedding plane slip, a point emphasized by other workers (Shainin 1950, Wilson 1961).

A modification of these tension fractures is seen where they assume a sigmoidal form (Fig. 41b) due to the rotation of their central portions relative to their extremeties. Such features have been described from both experimental studies (Riedal 1929) and natural occurrences (Shainin 1950, Wilson 1960, 1961) and can be observed to a less prominent extent in the present area.

A further variation of this structure has been noted at the Teveldal locality wherein a major tension gash, sigmoidal in form, is split into 4 smaller gashes (Fig. 41c) three of which are linked by minor fractures, again tensional, developed in response to the localized distribution of stresses. Such a subdivision into minor gashes was nowhere observed to affect contiguous major tensional fractures.

**Thrusting**

With the exception of the minor examples mentioned above, faults were not recognised although the Grense thrust by virtue of its ca. 45° dip can perhaps be classified as a major reverse fault. This forms an outstanding topo-

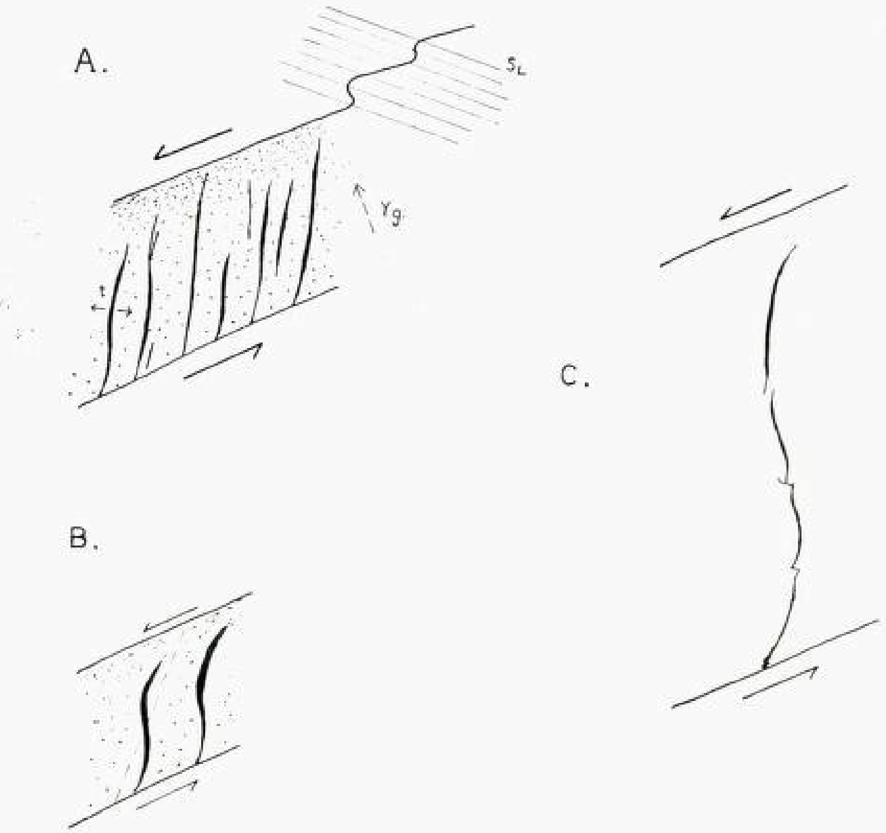



Fig. 41. Tension gashes in psammite. (A) Diagrammatic representation showing probable stress distribution. Yg — younging direction. $S_L$ — late cleavage. (B) Sigmoidal fractures. (C) Division of curved tension fractures into smaller sigmoidal gashes.

*Tensjonssprekker i sandstein (psammite). a) Diagramatisk fremstilling som viser mulig stressfordeling. Yg — oppover i lagserien. $S_L$ — sen oppsprekking. B) Sigmoidale sprekker. C) Oppdeling av kurvede tensjonssprekker i mindre sigmoidale åpninger.*

graphical feature since the mylonite material constituting the thrust zone has proved an easy target for agents of weathering and erosion (Fig. 42). The thrust zone is here at least 50 m thick but poor exposure renders an accurate measurement difficult. A road-cut through the central part of the zone provides excellent exposures sub-parallel to the strike which is here about 054°.

The rock is a dark grey mylonite, in part a phyllonite, with abundant secondary quartz much of which occurs as pods and lenticular segregations.


Fig. 42. View looking S.W. from Storlien road, riksgrense, showing Grense thrust plane. Upper Hovin Group on Stenfjeldet (S): Eocambrian gneisses etc. below mylonite of thrust plane.

*Utsikt mot sydvest fra mellomriksveien ved grensen mot sydvest. En ser skyveplaner. Øvre Hovingruppe (S): Eokambriske gneiser osv. under skyveplanets mylonitter.*

A lineation plunging 7°—10° towards 248° is present, while a weak strain-slip cleavage of indeterminate direction is seen to deform the mylonite locally. The highly sheared metasediments immediately above the thrust zone display a pronounced cleavage trending at 079° and dipping at 42° towards the N.N.W., as well as a lineation near-parallel to that noted in the thrust zone. The late cleavage is here poorly developed and strikes at 166° (dip 25° to E.N.E.) while its associated linear element, mostly small folds and micro-crumples, plunges at 4°—11° towards 355°. An interpretation of these observations in relation to structural considerations on a regional scale follows later.

It is of interest to note that Törnebohm (1896) considered this thrust as a 'minor overthrust-plane' in comparison with his 'great overthrust-plane' which on his profile (Tafl. 4) is drawn below the former. The 'great overthrust-plane' reaches the surface east of the Köli Schists in Jämtland (Sweden): this is the thrust plane of the Great Seve nappe (Asklund 1960, fig. 2) the minor thrust plane being regarded as the tectonic boundary of a sub-nappe to the main nappe.

**Mineralogical notes**

While the metamorphism of the sediments is discussed in the accompanying papers (Siedlecka *op cit*, Chaloupsky and Fediuk, *op cit*), it is pertinent here to comment briefly on certain mineralogical features and their relationship to the tectonic episodes. Within the areas mapped by the above authors, the rocks contain a mineral assemblage indicative of the quartz-albite-epidote-biotite sub-facies of the greenschist facies and perhaps, in part, the albite-epidote-amphibolite facies of regional metamorphism. It is important to note that this metamorphism accompanied the earliest phase of deformation; evidence from areas to the west suggests that it continued partly after the movements had ceased. The schistosity and general metamorphic fabric displayed by the metasediments is clearly associated with, and has originated during, this early folding and has subsequently been deformed by two later phases of folding.

A notable mineralogical feature in these Upper Hovin metasediments, particularly in the more pelitic lithologies, is the profusion of biotite porphyroblasts measuring up to 4 mm across. These biotites, poekiloblastic and overgrowing and containing the metamorphic syn-early fold fabric are deformed by the later microfolds and associated cleavage, although this same cleavage may occasionally be deflected around the porphyroblasts. The trend of the inclusion fabric is perfectly parallel to that of the groundmass schistosity.

The biotites are frequently oriented parallel or sub-parallel to the schistosity, though they do occur at any angle to this plane. Pleochroism is generally only moderate: Z and Y pale orange-brown, X straw yellow. In some of the carbonate-rich phyllites and metasiltstones, the porphyroblastic biotites would appear to be phlogopitic.

This porphyroblastesis, in contrast to the earlier main regional metamorphism, is clearly of a non-kinematic origin and occurred either in the static phase separating the second and last deformation phases or immediately preceding the thrusting at the termination of the second phase of the first protracted episode of fold movements. Evidence favouring the latter alternative is that of partial chloritization of biotites in the most easterly specimen of porphyroblastic phyllite yet examined by the writer. This locality is situated some 100—150 m perpendicularly above the postulated downward extension of the Grense thrust plane.

In all other thin-sections examined (further to the west) no trace of chloritization has been observed, not even along late cleavage planes. This would suggest therefore that the diaphthoretic features are intimately associated with the thrusting. In an investigation of the Sylene—Skardørsfjell region 50 km south of this Teveldal area, Schaar (1962) also found chloritization of porphyroblastic biotites in similar metasediments to be restricted to the vicinity of a thrust plane (almost certainly equivalent to the Grense thrust) separating those metasediments from Eocambrian sparagmitic gneisses.

Small pyrite metacrysts up to 1 mm across scattered throughout the more pelitic rock-types would also appear to post-date the early metamorphic fabric. They pre-date the biotite porphyroblastesis however, as several examples of biotite containing this pyrite as inclusions have been observed. While the inclusion fabric within such pyrites is usually parallel to the groundmass schistosity a few examples show this to be oblique to the latter and in one case displays a slight sigmoidal curvature. Moreover, 'pressure-shadows' of quartz are not infrequently present adjacent to opposing sides of the metacrysts. Such a feature is never encountered around even the most idioblastic of biotite porphyroblasts. The inference to be drawn from these observations is that the growth of pyrite began shortly before the first episode movements had actually ceased and continued into the interval of relative quiescence prior to the thrusting.

In the same thin-section in which chloritization of biotites is seen, pyrite displays varying stages of alteration to limonite and hæmatite together with replacement by scapolite to such an extent that scapolite pseudomorphs after pyrite may be well-developed. The periphery of these pseudomorphs is generally outlined by limonite.

Quartz veinlets, with or without calcite and $\leqslant$ 1 cm thick, which post-date the late fold phase are sometimes present in the more phyllitic rock-types. These are remarkably persistent features and generally sub-parallel, or rarely parallel, to the late cleavage. It is of interest to note that identical thin quartz veinlets occur sub-parallel to the late cleavage in the western part of Stjørdalen in similar Lower and Upper Hovin Group shales and phyllites.

An uncommon feature in pelites is that of a fine ramifying network of quartz and quartz-calcite veinlets some of wafer-thin proportions. These thin veinlets may follow the later cleavage planes or joint planes over distances of several centimetres though on the whole they anastomose discordantly and are independent of s-planes. No preferential occurrence in the axial planes of late folds could be detected. Again, similar veining can be observed in the western part of Trøndelag. It would appear evident that the injection of such veinlets occurred towards the close of, or immediately following the cessation of the latest deformation.

**Summary and structural relationships**

The structures in the Kopperå-Riksgrense area indicate that the rocks have been subjected to two distinct episodes of folding, the earlier deformation itself being divided into two phases. The major mappable fold, the Teveldal syncline, can be dated to the younger phase of the early deformation episode. Although minor structures related to this fold are present, the generally pervasive early minor structures belong to the older fold phase. The late episode of folding is characterized by minor folds and an associated penetrative cleavage, these structures clearly deforming the early folds.

The oldest recognisable folds are generally minor structures of a tight or isoclinal nature and of similar style. No general sense of overturning could be determined. Rodded quartz is commonly associated with these folds, the rods paralleling the fold axial direction. A prominant feature, taking the area as a whole, is the gradual swing of these early lineations through more than 60°. In the Kopperå area in the west the linear element plunges towards a west-north-westerly point at a moderate angle. Tracing this lineation towards the Swedish border it swings into an east-west position in the centre of the traversed area, then gradually towards a trend slightly north of east-north-east nearer the Grense thrust.

An explanation for the present disposition of this early lineation can be sought either in one of two mechanisms or by recourse to a combination of mechanisms. Many examples of early lineation trend variations resulting from deformation by newer folds are to be found in the literature. Ramsay (1960) has demonstrated that where an early lineation is deformed by similar folding, the angle between the disoriented early lineation and the axial direction of the similar folds varies systematically across the later fold. The same writer has also shown that during similar folding, fold axes and related linear structures are not necessarily developed at right-angles to the principal direction of movement and can indeed be formed at any angle to the tectonic *a*-direction. This is the first mechanism which could possibly account for the features seen over the present area, and which naturally cannot be rejected as a working hypothesis in any region that has suffered polyphase deformation.

The second possibility explains the curving linear element as a contemporaneous product of the initial phase of folding. This necessarily lends support to the assumption that linear structures under certain conditions may be formed at any angle to the accepted principal direction of movement. In this respect the gradual linear swing is in many ways comparable with that described by Kvale (1948) from the Bergsdalen area of Western Norway:

that area, however, is one characterized essentially by thrust tectonics and so a strict comparison may not be entirely valid. Even so, similarities of linear variation in both the Bergsdalen and Trøndelag areas are of appreciable interest and will be discussed in the final chapter of this paper. Although the conclusions with regard to the relationship of lineation and movement direction are virtually identical in Kvale's and Ramsay's papers, the means of achieving this end are quite dissimilar and genetically unrelated.

The features of the structure of the present area would tend to suggest that not one of these mechanisms alone can fully explain the linear swing. While some degree of combination of mechanisms is therefore postulated, it is necessary to consider structural elements further to the west as well in the general Northern Trondheim region in order to fully confirm this view.

The major Teveldal syncline is demonstrably tight or near-isoclinal and from an examination of congruous minor folds has a broadly similar style. Its main axial plunge is towards a S.S.W. - S.W. point. The later folds, in contrast to the early structures, have fairly constantly oriented axes plunging at a low angle towards north to north-north-east. Small variations do occur and these have been shown to be related to the controlling influence of firstly, lithology and secondly, the disposition of the limbs of the pre-existing Teveldal syncline. Refolding of earlier folds and lineations is present on a small scale but no regional late folds have as yet been found. These folds are parallel folds in contrast to the similar folding characteristic of the earlier deformation phases.

From an investigation of the various structural criteria, the main thrusting appears to have occurred towards the end of the second fold phase. It is thus later than the main regional metamorphism but older than the last recognisable fold movements in this area. Before considering the lines of evidence pointing to this conclusion it is necessary to premise the fundamental observation that, in this general region of the Scandinavian Caledonides, thrust movements have been directed from N.W. to S.E. Local variations occur but this south-eastward direction of transport of thrust sheets and nappes is irrefutable.

Although the thrust zone lithology has not been thoroughly examined petrographically, it is nevertheless clear that the metasediments with their associated schistosity have been severely sheared and mylonitized. Relatively uncommon strain-slip features deforming the mylonite testify to post-thrusting movements. Such strain-slips may or may not be parallel to the cleavage associated with the late folding. This late cleavage is usually penetrative over all the area, but in metasediments immediately above the thrust, it is a notably subordinate structure and less closely spaced than elsewhere. The late fold axes are markedly oblique in trend to the strike of the thrust zone.

These various observations together with those presented earlier provide evidence of movement both prior to and ensuing the thrusting. Reviewing the structures associated with the Teveldal syncline, it is noteworthy that over a region embracing the areas mapped by Siedlecka, Chaloupsky and Fediuk, the fold axial trend and general strike of the bedding is almost perfectly parallel to the strike of the thrust zone. Folds of this age — deforming the schistosity in the phyllites yet older than the late folding and cleavage — are characterized by an increasing fold dihedral angle to the south-east of the axial trace of the Teveldal syncline.

Considering the total evidence, a conclusion that the thrusting was in some way related to the second deformation appears unavoidable. The oldest identifiable folds have all the attributes of plastic deformation whereas the younger of the two early fold phases seems to have been more variable but on the whole less plastic than the former. In the eastern part of the region the plasticity of the folding was quite possibly at a minimum.

In summary, the terminal stage of the second fold phase can be regarded as one distinguished by fairly rapidly decreasing plasticity. With the rocks then acquiring an unaccustomed rigidity, thrusting is thought to represent the ultimate product of this deformation. A final point concerns the latest folding. The sense of overturning of these structures is constantly towards the west, here down-dip, suggesting a westward movement of upper sections of the metasedimentary pile relative to lower units. This late movement direction is approximately opposite to that acceptable for the displacement of the nappes and thrust sheets in this part of the Scandinavian Caledonides and will be considered more fully in the concluding regional discussion.

## B. DISCUSSION OF THE TECTONICS AND POSSIBLE STRUCTURAL CORRELATIONS IN THE NORTH-EAST TRONDHEIM REGION

### Introductory comments

In the light of present knowledge it would be premature at this stage to express unequivocal opinions concerning the sequential aspects of the structural evolution of as broad an area as that of the northern Trondheim region. With one exception (Peacey 1964) detailed structural studies over this region are virtually non-existent since previous workers have, through force of necessity for a general understanding of the stratigraphy, been concerned principally with the basic mapping of lithological groups. Structural interpretation has therefore been relegated to profiles based largely on dip observations and lithological correlation and while this has helped to formulate ideas about the generalized structure, information on the episodic nature of the Caledonian deformation is inadequate or lacking.

The present notes concern the findings of the structural observations in the above-described Kopperå-Riksgrense area and their relationship to major and minor structures investigated over the general Trondheimsfjord-Kopperå area. An endeavour will then be made to correlate the advocated pattern of tectonic events with structures occurring further north in the Tømmerås-Hegsjøfjell area. While the writer is fully aware of the difficulties entailed in attempting any long-range tectonic correlations, the distances involved in this case are relatively moderate and would appear to present a minor hindrance in comparison with the problem of lack of structural information.

The accessibility of Stjørdalen, the deep east-west valley linking Trondheimsfjord to the Jämtland area of Central Sweden, made it the subject and starting-point for a large part of the early geological exploration in this segment of the Caledonides. Since for the most part it trends normal to the strike, the valley provided an ideal cross-section through the various lithologies. Although details of the stratigraphy are given in several publications (Kjerulf 1871, 1875, 1883, Reusch 1883, 1890, Törnebohm 1896, C. W. Carstens 1919 and Wolff 1964), it can be mentioned briefly here that the central part of Stjørdalen and the Trondheim region is characterized by mica schists often containing garnet or hornblende and sometimes bearing kyanite, staurolite or sillimanite (Plate III). This is the Gula Schist Group, previously called the Røros Schist. Reasons for the change of name are discussed in Wolff's paper in this same NGU volume. Both to the west and to the east of the Gula Schist Group there occurs a sequence composed principally of greenstones with subordinate amphibolite and quartz-keratophyre. Differences of lithology are apparent between the western and eastern representatives of this series, the Støren Group. Further to the west and east, Lower and Upper Hovin Group rocks are encountered, these being only weakly metamorphosed. The Horg Group is so far known to occur only in the eastern part of Stjørdalen where it is known as the Slågån Group.

Fossil evidence, as mentioned previously, shows the Horg rocks to be of Lower Silurian age (Getz 1890) and thus represent the youngest sediments in this region. Sedimentary structures point to gradual ageing of the stratigraphy below this group and the pebble content of the basal Lower Hovin conglomerate indicates derivation largely from the Støren Group. The describing of *Dicteonema flabelliforme* in carbonaceous shales from the upper Guldal district (Størmer 1940, Vogt 1940) 60 km south of Stjørdalen is of appreciable interest, although the true stratigraphical position of this fossiliferous horizon has not as yet been clarified. The shales are interbanded with schists and amphibolites and it is quite possible that the horizon is situated within a sequence equivalent to the Støren Group or alternatively, on complex structural grounds, at the top of the Gula Schist Group. Speculation on this point must remain until a detailed investigation of the area is carried out. Størmer (1940) considered the graptolite as dating the shales to the very base $2e_{\sim}$ zone of the Oslo region) of the Ordovician.

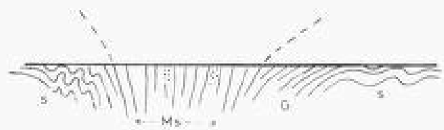
A recent find of a graptolitic shale from the Lower Hovin Group near Løkken, S.W. of Trondheim, has been described by Blake (1962) and was considered to be of Middle Arenigian (extensus zone, pars. of Britain, $3 b\gamma$ of the Oslo region) but Skevington (1963) has queried this and suggests a slightly younger (hirundo zone of Britain, $3b_{\varepsilon}$ of Oslo) age. In his paper Blake suggested the Støren Group to be "at least as old as early Arenigian and quite possibly, Tremadocian".

From this it will be appreciated that a precise dating of the various stratigraphical groups is hindered both by a general lack of fossils and by conjecture about the positioning, in relation to local stratigraphy, of those so far discovered.

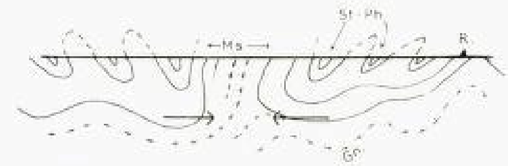
Returning to Stjørdalen, in the west the dip of the bedding of the metasediments is towards in easterly point. Approaching the central Gula Schist zone dips become steeper: a traverse across the Gula Schist Group shows a profusion of minor and larger, sheared, isoclinal folds, the recognisable banding being disposed about the vertical. The Støren Group to the east displays westerly dips becoming less steep as one traverses eastwards. Further east is the Kopperå-Riksgrense area, described above. It will be noted therefore, that over the greater part of the Stjørdalen profile the stratigraphy is inverted.

### Previous interpretations of the Stjørdalen structure

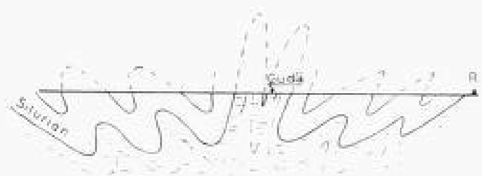
The first descriptions of the Stjørdalen profile were those of Kjerulf (1871 and 1883) who considered the whole as a synclinorium with the schists of the central 'vertical zone' as the youngest rocks (Fig. 43a). Svenonius (1885) went to the other extreme in considering these same schists as the oldest group lying in the core of an anticlinal structure squeezed up as shown in fig. 43b.




A. Kjerulf 1871

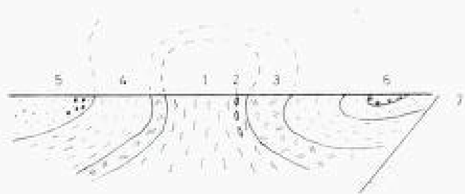



B. Svenonius 1885 (Hypothetical profile)




C. Reusch 1890


D. Carstens 1920




E. Wolff 1964


F. Roberts 1966

Reusch (1890) agreed with this latter interpretation but modified the profile slightly to accommodate the Gudå conglomerate which was then thought to be of Silurian age (Fig. 43c).

Törnebohm's (1896) ideas on this structure are somewhat difficult to grasp as he did not draw a profile directly along Stjørdalen. The impression gained from sections sketched both to the north and south of this area and from his interpretation of the stratigraphical succession is that he was an advocate of the synclinorium hypothesis.

In his lengthy treatise on the geology of the Trondheim region, C. W. Carstens (1919) was firmly in favour of an anticlinorium with the Røros Group (= Gula Schist Group) in its core (Fig. 43d). He applied this interpretation to the whole of the Trondheim region. Holtedahl (in Bailey and Holtedahl, 1938) also indicated this same structure as being anticlinal.

Th Vogt (1940) on the other hand believed the central part of the Trondheim region to be of synclinal character. His profile in the eastern part of Guldalen was more or less a mirror image of Bugge's (1910) conception of the structure around Rennebu: Bugge (1910, 1912) also looked upon the regional structure as a simple syncline but later (1954) reversed his views.

Wolff's (1964) section along Stjørdalen is basically similar to that drawn by Carstens (1919), the oldest rocks occurring in the core of an anticlinal structure (Fig. 43e). The main differences are seen in a revision of the strati-

Fig. 43. Simplified interpretations of the central Stjørdalen structure. (A) *Kjerulf:* Ms — mica schist and gneiss with granite veins. G — greenschist at Meråker. S — siltstone and shale. (B) *Svenonius:* Ms — mica schist. Gn — gneiss, granite-gneiss, quartz schists. St-Ph- — siltstone, shale and phyllite. (C) *Reusch:* R — riksgrense. (D) *Carstens:* 1 — Røros Group. 2 — Bymark Group. 3 — Hovin Group. 4 — Sparagmite formation. 5 — Eruptives. (E) *Wolff:* 1 — Røros Group. 2 — Gudå conglomerate. 3 — Støren Group. 4 — Lower Hovin Group. 5 — Lower Hovin Group. 6 — Horg Group with local basal conglomerate. 7 — Basement. (F) *Present paper:* 1 — Gula Schist Group. 2 — Gudå conglomerate. 3 — Støren Group. 4 — Lower Hovin Group. 5 — Upper Hovin Group, basal conglomerate in west. 6 — Horg (Slågån) Group. 7 — Eocambrian, mostly gneisses and schists. 8 — Possible slide zone.

*Forenklede tolkninger av strukturen i midtre Stjørdalen. a) Kjerulf: Ms — glimmerskifer og gneis med granittårer. G — grønnskifer ved Meråker. S — leirstein og skifer. B) Svenonius: Ms — glimmerskifer. Gn — gneis, granittgneis, kvartsskifer. St-Ph-leirstein, skifer og fyllitt. C) Reusch: R — riksgrense. D) Carstens: 1 — Rørosgruppen, 2 — Bymarkgruppen, 3 — Hovingruppen, 4 — sparagmittformasjonen, 5 — eruptiver. E) Wolff: 1 — Rørosgruppen. 2 — Gudå-konglomeratet. 3 — Størengruppen. 4 — Undre Hovingruppen. 5 — Øvre Hovingruppen. 6 — Hovingruppen med lokalt basalkonglomerat. 7 — underlaget. F) Dette arbeidet: 1 — Gulaskifergruppen. 2 — Gudåkonglomeratet. 3 — Størengruppen. 4 — Undre Hovingruppen. 5 — Øvre Hovingruppen. basalkonglomerat i vest. 6 — Horg (Slågån)-gruppen. 7 — Eokambrium, for det meste gneiser og skifre. 8 — Mulige glidesoner.*

graphy, based largely on lithostratigraphic correlation with the Hølonda-Horg district of the Trondheim region (Vogt 1945), but also taking into account the structural position of the Silurian (Horg) rocks further to the east.

In his important study of the Hølonda-Horg area, Vogt (1945) referred to relatively high-grade schists (his Brek Series) as occurring stratigraphically below the Støren Group. These schists would be equivalent to the Gula Schist Group of the present area. Although no regional structural implications were suggested, Vogt's S.W. Trondheim stratigraphical sequence is not consistent with his earlier opinion (Vogt 1940) on the regional structure.

The above review of the literature shows that there has been considerable disagreement as to the precise nature of the main structure present in the central zone of both Stjørdalen and this part of the Trondheim region. Opinions have oscillated between the syncline and anticline interpretations, mainly it would seem because of confusion over the stratigraphical succession. As knowledge of the region has increased, so the concensus of opinion has appeared to favour an anticline as the dominating structure. Despite this revised view, again based solely on stratigraphical relationships, detailed structural observations either favouring or repudiating it and determining relationships with adjacent structures have been entirely lacking up to the present time.

In the present writer's opinion the major Stjørdalen fold is a fan-shaped anticlinal structure, the core of which is characterized by intense isoclinal folding accompanied by shearing and stretching phenomena (Fig. 43a), such that relics of this early folding are frequently limited to minor fold closures, the limbs having been destroyed. Intrusive bodies of trondhjemite, trondhjemite-pegmatite and associated acidic rock types are locally abundant in these schists. Although the central and western parts of Stjørdalen will form the basis of a subsequent publication, some features of the tectonics and metamorphism are discussed below in relation to the sequence of events advocated for the Kopperå-Riksgrense area.

### Tectonic relationships in the Meråker - Stjørdalen region

In the previous account of the tectonics of the Kopperå-Riksgrense area, it was stressed that the earliest deformation was of a plastic nature and was accompanied by the development of the pervasive schistosity. While the metasediments east of Kopperå were clearly affected by only a low-grade metamorphism, a perceptible increase in grade can be traced westwards moving down in the stratigraphical succession until the more strongly metamorphosed Gula Schist Group is encountered in the core of the major anticline, here called the Stjørdalen anticline. Further west a complementary decrease in grade can be observed on moving out of the zone of schists and up the succession.

Particularly instructive evidence with regard to the sequence of fold phases and the age of the main metamorphism in this central district can be found just to the west of Gudå within the Gula Schist Group. The schists here are garnetiferous and sometimes contain kyanite and fibrolite. Locally the schists contain abundant thin quartzitic ribs thus presenting a somewhat more competent, mixed pelite-psammite lithology in which minor folds and fold relationships are far better preserved than elsewhere within the Group. The Gudå conglomerate, the origin of which has provided a fair amount of controversy (Bäckström 1890, Kautsky 1947), occurs within this generally more psammitic zone.

In these schists isoclinal folds, often of considerable wavelength and to which the schistosity is axial planar, are deformed by folds which, though varying from fairly open to isoclinal style, are generally rather tight structures (Fig. 44). Pods of fibrolite-muscovite lying within the schistosity are stretched in a direction paralleling that of the axes of the earlier isoclines and are bent around the closures of the superimposed tight folds. From these observations it is quite clear that two generations of folds are present, but it is equally important to note that both these fold types pre-date the latest fold episode. Minor structures of this late deformation have not been observed in this small area of Gula Schists west of Gudå but they appear two or three hundred metres further east in greenschists of the Støren Group and become gradually more conspicuous eastwards. Within the variable lithologies constituting the Støren Group both generations of structures pre-dating the late folding are identifiable.

The minor structural sequence in this more central part of Stjørdalen is therefore comparable with that found in the area east of Kopperå, although traceable differences in the style and stages of development of these structures can be observed. More significant are the relations of minor structures of each phase to the major structures and from this it will be shown that the relationship of the Stjørdalen anticline to the Teveldal syncline is not as straightforward as might initially be assumed. Detailed descriptions of the minor structures associated with each of the deformation phases are beyond the scope of this paper, but it is essential to consider certain observations which are of particular significance to the present discussion.

*a) First phase minor structures*

The earliest recognisable minor folds are the isoclines with their associated axial planar schistosity. These structures occur throughout the Gula Schist Group although they are often difficult to observe immediately on account of the intense shearing and monotonous lithology. The most prominent linear element paralleling the fold axes is that of rodded quartz. In the eastern part of this central zone with the banding and schistosity striking constantly N.N.E. and dips generally either steep towards the W.N.W. or vertical, lineations plunge towards S.W.-S.S.W. at a moderate angle. In the vicinity of the Gudå conglomerate zone south of Stjørdalselva (the Stjørdal river) the trend is west of S.W., now markedly oblique to the general strike of schistosity and banding.

The most outstanding feature of the Gudå conglomerate is the extreme deformation of its pebbles, the maximum elongation lying within the plane of the schistosity and ascribed to the earliest deformation phase. Although a detailed examination and quantitative study has not yet been carried out, preliminary observations show the main pebble orientation to trend almost normal to the strike of the schistosity, plunging steeply towards a westerly point. Locally a trend more towards 240° is present, there almost paralleling the axes of the earliest isoclinal folds: however, instances are noted within a metre or two of westerly plunging pebbles, of similar isoclinal folds in a mixed schist-quartzite lithology plunging towards a south-south-westerly point. Thus the earliest lineation in this small area would appear to be quite variable. To some extent this variation can be attributed to the superimposition of second generation folds, but examples of plunge variation associated with a common schistosity plane can be seen in the form of 'eyed' folds in a mixed schist-psammite sequence on the plateau top north of Stjørdalselva. This strongly suggests that some of the variation in the earliest linear trend was synchronous with the development of the earliest folds.

Further to the east in the Støren Group between Gudå and Meråker, the recognisable earliest isoclinal folds and lineations plunge towards a west-south-westerly point. The type-locality of the recently discovered Lille Fundsjø conglomerate (Chaloupsky and Fediuk, 1967) was visited briefly by the writer with a view to obtaining information on possible pebble orientation. It was found that severe stretching akin to that seen in the Gudå conglomerate is absent but a pebble orientation is clearly discernible plunging at 30° towards 270°—278°, almost normal to the strike at this locality.

At another locality 1 km along the strike the pebble lineation plunges towards 282°. A few pebbles were observed fractured normal to their longest

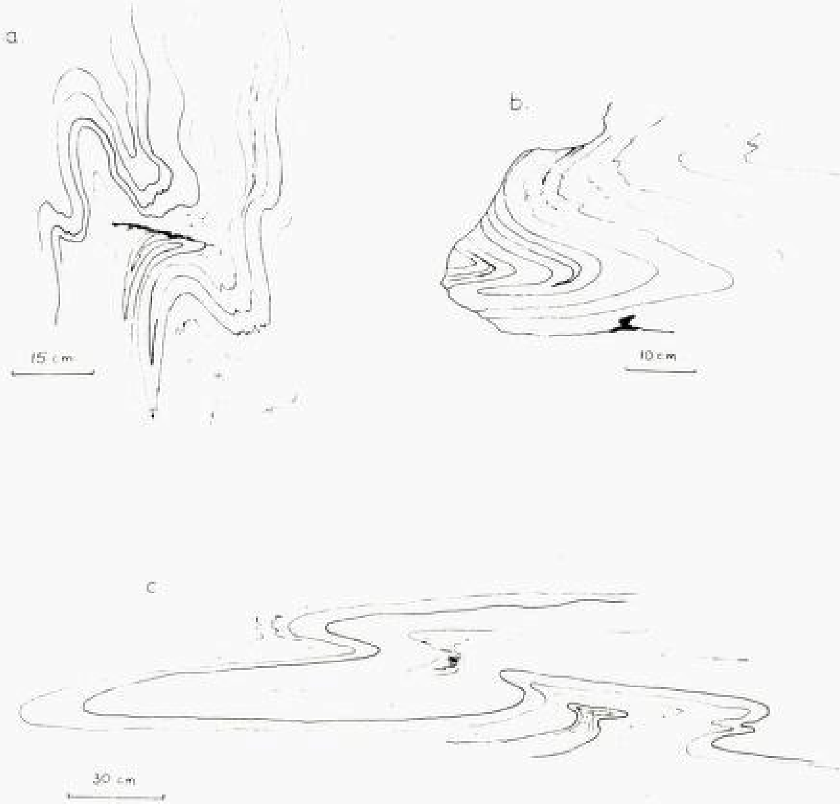



Fig. 44.. Refolding of first-phase isoclines by second generation structures. Interbanded schist-quartzite lithology. Gula Schist Group; just west of Gudå.

*Nyfolding av førstefases isoklinalfolder ved annen generasjons folding. Veksling av skifer og kvartsitt, Gulaskifergruppen like vest for Gudå.*

dimension, the segments sometimes having been pulled apart with the schistosity of the matrix flowing around and partly between them. The fracturing of the pebbles could therefore not have been younger than the development of the schistosity. This various evidence indicates that the pebble orientation is a feature of the earliest phase of deformation.

Eastwards from the central area of the Gula Schist Group a gradual swing of the earliest lineation is therefore manifest, from S.S.W-S.W. through west to a point north of west in the Meråker area; this is achieved without any notable change of strike of banding or schistosity. It will be recalled that in the Kopperå area some 5 km east of Meråker, the early lineations plunge W.N.W. down the dip of the schistosity. Further to the east, a gradual swing back towards a south-westerly plunge was described.

*b) Second phase minor structures*

Folds ascribed to this generation of structures deform the schistosity and related early linear elements. They are generally tight though showing marked variations in style within the Gula Schist Group. Although no cleavage or schistosity attends these folds, a rucking of the pre-existing schistosity at the fold closures is not uncommon.

In this central Stjørdalen district the area of schist with quartzite bands just west of Gudå provides the most prominent examples of second phase minor folds. Further west these structures become inconspicuous. Many such folds in this Gudå area approach isoclinal style though amplitudes are generally small. A stream-section on the south side of the Stjørdal valley some 500 m west of the farm Bitnes displays many examples of refolding, first phase isoclines being deformed by second phase structures (Fig. 44a and b). The host lithology is here an intensely sheared, closely banded alternation of quartzite and schist such that first phase fold hinges are frequently divorced from the sheared and sliced limbs. Indeed in certain places the lithology takes on the guise of a deformed conglomerate. The shearing is a first phase deformation phenomenon. In the present writer's opinion both true conglomerate and pseudo-conglomerate are present in this small area but discussion on this subject must be postponed until a more complete study has been undertaken.

The axes of second phase minor folds plunge fairly regularly at 10°—30° towards S.-S.S.W. In the mixed schist-quartzite lithology, however, although the trend shows little variation some axes are seen to plunge rather more steeply towards SSW whilst one or two plunge to the NNE at moderate angles.

Folds of this generation are irregularly developed in the Støren Group east of Gudå. Where prominent they are of tight to close style with either one or two sets of associated widely spaced shear planes; where two shear directions are present these are parallel to the axial planes and long limbs of the folds. Fold amplitudes may be in the order of several metres.

Between Meråker and Kopperå information on these folds is lacking. They are again noticeable in the Teveldal valley area further east as structures of varying amplitude congruous to the major Teveldal syncline. Plunges there are gentle towards S.W.-S.S.W.

Fig. 45. Late (third-phase) chevron-styled fold showing en échelon arrangement. Phyllite, Lower Hovin Group. Forestry track north of Flornes bridge, Stjørdalen.
*Sen (tredje fase) siksakfolder med trappeform. Fyllitt, undre Hovin-gruppen. Skogsvei nord for Flornes bru, Stjørdalen.*

*c) Third phase minor structures*

Within the Gula Schist Group minor structures belonging to this youngest episode of deformation are almost entirely absent. The only occurrence is to be found in the westernmost part of the schist zone within a locally more phyllitic or slaty horizon. There the deformation is represented principally by kink-bands or kink-folds: the bounding surfaces of such bands, rarely $>$ 15 mm apart, are generally parallel over a distance of a metre or more though they sometimes merge to form a single slide-plane. Such kink-bands only affect the finer-grained phyllitic lithology and are not developed in coarser inter-banded schists.

Westwards traversing greenschists of the Støren Group and Lower Hovin phyllites, a gradual development from the incipient kink-bands to chevron-styled folds of increasing amplitude and wavelength can be demonstrated, until an axial planar cleavage is prominently developed. The sense of over-turning of these folds — frequently arranged en échelon (Fig. 45) — is down-dip towards an east-south-easterly point. Whereas the hinges of kink folds are sharply defined, a gradual rounding of hinges with increased fold development is observed probably due to the coarser more psammitic intercalations in the Lower Hovin rocks behaving in a more isotropic manner. Further to the west beyond the village of Flornes the cleavage becomes penetrative in weakly metamorphosed pelitic Lower and Upper Hovin sediments, a situation analogous to that existing in the Kopperå-Riksgrense area for the late structures.

Fold axes have a relatively consistent orientation, N.N.E.-S.S.W. Plunges are mostly towards N.N.E. at anything up to 20° but in several cases these third phase axes are horizontal or else plunge to the S.S.W. at some small angle. From a detailed study of a well-exposed area in phyllite near Flornes, any one fold axis can be shown to possess a variable plunge. The variation appears to be wave-like and is demonstrable on all scales. Furthermore, the trends and plunges of parasitic minor folds on the limbs of larger chevron folds frequently do not shown a parallelism with the main fold axis.

Axial planes, and cleavage where present, dip consistently at some small angle towards an E.N.E.-E.S.E. point. Further west cleavage dips up to 30°—35° have been noted With an interbanded phyllite (shale) and graywacke lithology, excellent examples of cleavage refraction are encountered.

In the Flornes area a conjugate set of third phase kink-bands or minor chevron folds is occasionally present. Such simultaneously developed paired folds do not appear further west.

On the eastern limb of the Stjørdalen anticline third phase structures appear in the greenschists of the Støren Group as microfolds, phyllitic lineation and sporadic kink folds. A strain-slip cleavage is developed in the vicinity of Meråker and this dips towards E.N.E.-E. Further to the east these late folds become very abundant with a penetrative eastward-dipping axial planar cleavage, as described in the notes on the Kopperå-Riksgrense area. Fold axes throughout this district between Gudå and the Swedish border plunge consistently at some small angle towards N.-N.N.E. The sense of overturning of these late folds is again down-dip, here to the W.N.W., but it will be seen that the sense of movement is here opposite to that which holds for the western part of Stjørdalen (Fig. 46).

Conjugate folds and kink-bands also occur in one part of this eastern district and have been described earlier. It is interesting to note firstly that the conjugate structures tend to be restricted to the thicker horizons of the more homogeneous dark phyllites, though they do occur sometimes in fine-grained greenschist. A second point is that in this general region these paired folds and kink-bands appear to be developed preferentially in the steeper dipping or near-vertical zones, lithology permitting. Paterson and Weiss (1966), reporting on experimental deformation in phyllite, have shown that under compression parallel or sub-parallel to the foliation, conjugate kink folds are readily developed whereas in cases where the phyllite is compressed at moderate to large angles to the foliation, only a single set of kink folds is formed. These conclusions invite speculation with regard to the field occurrences described above, but without a more thorough knowledge of the third phase structures in this Trondheim region it would be dangerous to attempt any correlations. Moreover, experimental stress-strain relationship probably differ appreciably from those under natural conditions. It is nevertheless of interest to note the similar geometrical properties of these experimentally and geologically induced structures.

*d) Relationship of minor and major structures*

The features of extreme deformation displayed by the first phase minor structures are particularly ubiquitous across the central part of Stjørdalen within the Gula Schist Group. Away from this anticlinal core the evidence suggests a gradually diminishing intensity of deformation, this at the same time allied to a decreasing grade of metamorphism.

The pervasive schistosity normally parallels whatever banding may be present, except of course at minor fold closures. In the lower part of the Støren Group at Gudå keratophyric and greenschist horizons have been folded, squeezed and sheared to an extent that it is impossible to determine the original stratigraphy; schistosity is parallel to this banding.

The Stjørdalen anticlinal is clearly a product of the first phase of deformation. The vertical or near-vertical disposition of banding and schistosity across the core of the Gula Schists, associated with innumerable sheared isoclines would appear to testify to this conclusion. Second phase minor folds deform the earlier developed schistosity but are numerically insignificant in the central zone. It is therefore virtually impossible to reconcile these structures with the development of the major fan-anticline.

Second generation minor folds on the other hand increase in magnitude towards the east so much so that, as previously considered, the major Teveldal syncline is ascribed to this phase of deformation. The situation, therefore, is that in this eastern Trondheim region two apparently complementary major structures are regarded as having developed at different times in the protracted deformation history. The time interval separating the evolution of the two structures, though impossible to evaluate, is considered to be relatively short and indeed, as will be discussed, it may be preferable to consider the development of the structures as the product of a broadly continuous deformation.

While the minor folds associated with the second phase of folding are found to be congruous in relation to the major structure in Teveldalen, it is difficult to come to any indubitable conclusion with regard to the relationship of minor first generation folds to their parent structure. This is primarily a consequence of the very nature of the folding and excessive shearing in the central part of Stjørdalen which has tended to mask the actual relationship of minor and major structures. In many exposures it is impossible to differentiate between the long and short limbs of isoclines. Where this is visible the sense of overturning may change quite frequently on traversing across the strike and wide zones are encountered wherein the minor folds appear incongruous towards the major anticlinal structure. It has been noted however, that shearing is most extensively developed along the long limbs of folds of various dimensions so that considering any one fold, minor folds parasitic to this structure tend to be better preserved or indeed restricted to the shorter limb. On this basis the main sense of overturning of recognisable folds in such a highly sheared domain would appear to be incongruous to the accepted major structure. Where the effect of shearing is diminished a more congruous relationship obtains.

The possibility of the occurrence of slides in this region awaits more extensive investigations. To the west of the anticlinal core the stratigraphical sequence is much attentuated, and near Flornes an 80 m thick band of greenschist with some coarser greenstone occurs within a phyllitic or partly phyllonitic lithology riddled with secondary quartz. Minor structural evidence indicates that the greenschist band does not occur as the result of repetition by folding but it is quite conceivable that it has been derived tectonically. A few kilometres further north-north-east in Forradal, the interrelationship of phyllite, Støren greenschists and greenstones and rocks of the Gula Schist Group is exceedingly complex. Although only a reconnaissance survey has yet been made, it is certainly quite possible that a syn-metamorphic slide or slide-zone exists at about this horizon.

Considerable but systematic variations in the attitude of the earliest minor lineations present an important structural feature over this region. In the central area the trend of first generation linear elements broadly complies with the strike whilst further to the east a gradual swing into an orientation normal to the strike of the metasediments is demonstrable. Moving still further east this lineation gradually swings back across the Teveldal syncline into sub-parallelism with the strike close to the Grense thrust.

Alternative explanations for this gradual swing, through some 80°—90°, were discussed earlier, but in the writer's opinion only a combination of mechanisms would appear to satisfy the field observations. The 'variable linear trend due to superimposed similar folding' hypothesis advocated principally by Ramsay can most certainly be invoked in the eastern area since there the major syncline post-dates the early linear element. Further west where the influence of large-scale second phase folding diminishes rapidly, this explanation cannot hold and consequently the curving lineation must there be looked upon as a contemporaneous product of the earliest deformation. To what extent the swing of lineation across the Teveldal syncline can be attributed to this syn-tectonic curvature is impossible to ascertain, but it is reasonable to suggest that it may be a product of combined mechanisms.

A curving lineation produced during one deformation phase naturally implies a variable angular relationship with the accepted principal direction of tectonic transport. This means that the linear structure would be regarded as a 'b' lineation at one locality, an 'a' lineation at another and an oblique lineation in all the intermediate positions. Kvale (1948 and 1953) expressed the view that, "linear structures may, under certain conditions, be formed at any angle to the principal direction of movement", a conclusion arrived at after detailed work in a region dominated by thrust tectonics. The observations in the Stjørdalen region would appear to support Kvale's suggestion, though it can be noted in addition that the curving linear element is here occurring in a region characterized principally by fold tectonics.

Possible causes of syn-tectonic linear curvature and indeed varying strain patterns and fold styles may be found in a number of controlling factors. These include the duration and magnitude of deformation, varying rates of movement and resistance, and the variable anisotropy of the rocks themselves. That the style of the first phase folding changes gradually across this area is easily demonstrated and a not unexpected phenomenon considering the probable operation of more than one of the above variables. A changing fold style and linear trend can also be related to both tectonic level and position with regard to marginal and central areas of the tectonic belt.

The occurrence of isoclinal minor folds with axes paralleling the accepted direction of tectonic transport — i.e., in the main 'a' direction — is however difficult to resolve with the hypothesis of one dominant direction of movement, so much so that the assumption of a uniform transport may be unjustifiable. Some movement along the trend of an orogen would appear likely, perhaps a local plastic flow controlled by pressure gradients. At the present time information is unfortunately lacking on the general sense of overturning of these minor folds in the assumed 'a' direction; this will require a more detailed study over a much wider area.

The direction of extreme stretching in the Gudå conglomerate is largely subnormal to the strike, but it is difficult to say here to what extent this corresponds to the tectonic 'a' direction since other first-phase lineations in this particular small area are irregularly orientated. In view of the nature of the closely compacted isoclinal folds with near-vertical axial planes, it is also plausible to regard movement during the most intensive stage of the deformation as having been directed upwards so that in this case it is not unreasonable to think of the pebble stretching as locally approximating to the 'c' direction. This is a possibility to be considered. However, during the development of the first generation structures, particularly in this zone of extreme deformation, many variations in both movement and stress direction were probably operative so that a complex rather than simple relationship between lineation and transport has to be envisaged. Turning to the Lille Fundsjø conglomerate, the picture there appears somewhat simpler as the deformation was less intense and pebble orientation conforms near perfectly to the overall 'a' direction.

Late or third phase folds are nowhere found to attain major proportions, although recent work has indicated that this generation of structures is better developed in the western districts of Stjørdalen. It was pointed out earlier that, considering the overall picture of easterly dips to the west and westerly dips to the east of the Stjørdalen anticline axial trace, the sense of overturning of the late folds in these two sub-regions is directly opposed (Fig. 46). Discounting the local development of conjugate structures, the overturning is everywhere down-dip or towards the core of the anticline.

In the Jämtland region of Sweden the Köli Schists, equivalent to the metasediments of the Kopperå-Riksgrense area and representing part of a thrust unit, are situated in a shallow depression centred on the Tännfors area. In these weakly metamorphosed sediments the penetrative easterly dipping late cleavage so common in Teveldal is notably absent (Dr. Arne Strømberg, personal communication).

A solution both to the apparent restriction of this generation of structures to the so-called Trondheim-field and to their pattern of overturning, probably lies in a consideration of the operative stresses. Within the Trondheim meta-

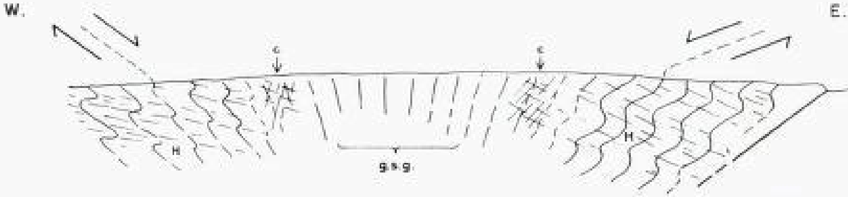



Fig. 46. Schematic profile along Stjørdalen to show sense of overturning of third-phase folds with respect to the attitude of banding. H — Hovin Group: g.s.g. — Gula Schist Group, c — zones with conjugate folds.

*Skjematisk profil langs Stjørdalen for å vise hvordan folder av tredje fase veltes over i forhold til båndingen. H — Hovingruppene. g.s.g. Gula skifergruppen. c — soner med konjugerte folder.*

sedimentary complex, if any reaction were to occur as a result of the nappe development and thrusting, movement would tend to be opposed to that operating previously. Late movement can therefore be envisaged as being towards the deeper rooted anticlinal core and a gravitational effect may conceivably be regarded as an additional factor in achieving this end. On frictional and resistance considerations this movement could probably be resolved as a regional shear couple as depicted in fig. 46. Since the Köli schists lie in a shallow saucer-shaped depression, it would be difficult in this case to imagine any reactionary or elasticoviscous movement or gravity sliding taking place so that structures of this phase would not be developed.

### Consideration of minor structures reported from neighbouring areas

In view of the several factors controlling the development of regional fold styles or strain patterns, a consistency of minor fold types and trends over any sizable area is most unlikely. In addition, the Trondheim region is seriously lacking in detailed structural studies so that any comparison of minor structural observations must lie within the realm of speculation. Nevertheless, certain reported observations are now of renewed interest following the work in Stjørdalen.

Kisch (1962), working in the S.W. Tydal area some 50 km S.S.W. of Meråker, has noted the orientation of strongly deformed quartz pebbles from the Usmadam-Bukhammerfjell metaconglomerate which has been correlated with the Gudå conglomerate by Wolff (1964). Of the stretching direction he remarked that, "the lineation is about E.-W., 45°W., roughly perpendicular to the average strike". Kisch also notes that the few small-scale folds measured in this part of his map area plunge either towards 340°—010° or towards

200°. Mention is made too of a lineation plunging at 25°—30° towards 210°—220° in certain schists of this group, and of isoclinal microfolds in part of his Amphibolite Group (= Støren Group) although no measurements of the latter are available.

The particularly interesting observation here is that of the deformed pebble orientation which is analogous to that in the Gudå conglomerate. Of the other observations it is difficult to come to any conclusions on account of the lack of information on fold styles, amplitudes, refolding, etc., but the 'lineation' in the schists plunging to the S.S.W. is possibly equivalent to the first phase lineation in the Gula Schists of Stjørdalen.

Schaar (1962) working in the Sylene-Skardørsfjell region to the east of Kisch, has described structures from his Stuedal Schist (correlative with the biotite-porphyroblastic Upper Hovin rocks of Teveldalen). He refers to 'small scale folds' with axes plunging fairly gently towards either N.-N.N.E. or S.-S.S.E. which are slightly overturned in an easterly direction. Although no details of style or axial planar altitude are given, these structures are probably equivalent to the late folds of the Kopperå-Riksgrense area since Schaar correlates them with the drag folds described by Bryn (1959) from similar metasediments in the nearby Essandsjø area.

The folds referred to and pictured in Bryn's (1959) article closely resemble those which the present writer ascribes to the third phase of deformation. Their axes plunge gently N.N.W. while the actual folds, with quite sharp hinges and dihedral angle ca. 90°, are overturned towards the W.S.W. Thus the sense of overturning is opposed to that noted by Schaar. Bryn attempted to relate these structures to the main easterly directed movements producing the thrusting. He explains the anomalous sense of fold overturning by assuming a friction-reducing well-lubricated band, perhaps a graphitic schist, to have existed beneath the metasediments such that during the eastwards-directed principal movement, the lowermost layers in the sedimentary pile were transported to the east at a faster rate than those above. Unfortunately such a graphitic layer has not been found. Moreover, structures correlatable with first and second phase folds from the Kopperå area are not described from this Essandsjø district, so that Bryn's explanation and relating of his drag folds to the major eastward movements is clearly based on the observation of folds of only one generation. In this same paper, Bryn mentions the occurrence of small folds with axes trending N.W.-S.E. in the south of his area. No further description is given but he refers to these as cross folds to other uncommon N.N.E.-S.S.W. oriented structures.

From the S.W. Jämtland area of Sweden, Strömberg (1961) has described a final deformation which is clearly subsequent to the thrusting of his Helags nappe (= part of the Seve nappe). He suggests a correlation of structures developed during this deformation with the drag folds reported by Bryn (1959). It is therefore likely that the late folding recognised in S.W. Jämtland is broadly synchronous with the latest deformation observed by the present writer in the Stjørdalen area.

Structures in the Selbu area S.E. of Trondheim have been discussed briefly by Torske (1965). He noted two fold episodes, the earlier of which is represented both by minor isoclinal folds in a phyllitic horizon and by a stretching of pebbles in a polygenous conglomerate (= basal Lower Hovin). The pebble orientation is E.-W., here sub-parallel to the strike of banding. Early lineations in general plunge towards N.N.E.-E. The later structures would appear to be kink-folds or slightly larger folds with an axial planar cleavage, and plunge at small angles to the S-SSE. These structures are most probably equivalent to the third phase folds of Stjørdalen whilst the earlier structures may possibly be equated with those of the first deformation phase.

Sixty km north of Stjørdalen the Skjækerstøtenes conglomerate, correlative with the Gudå conglomerate (Wolff 1964), displays features of extreme deformation. Quartzite pebbles are stretched in a direction almost normal to the local strike of banding and schistosity (Wolff 1960).

Quite the most consistent feature of these various minor structures is the extreme stretching of pebbles in metaconglomerates belonging to the same stratigraphical horizon — Usmadam, Gudå, Skjækerstøtenes — the stretching being normal to the local strike. This elongation, recognisable in areas some 130 km apart, would appear to correspond to the principal direction of tectonic movement.

In the extreme north of the Trondheim region, the basement rocks of the Tømmerås anticline and their adjacent metasedimentary successions have been described by Peacey (1964). Two episodes of folding were recognised, a so-called regional folding and a later folding of a more brittle mechanical nature. Lineations, including pebble orientation in the Steinkjer conglomerate (= basal Lower Hovin), are the dominant structure of the earlier deformation and display a gradual swing of trend from N.W.-S.E. in the south to N.E.-S.W. in the far north of the area. Later fold axes trend mainly E-W, their orientation controlled by the earlier fold limbs. In the north, however, a plunge towards a north-easterly point is evident. Peacey also recognised the existence of a fold generation older than the 'regional folds'; this was apparent from a study of Foslie's maps east of the Tømmerås anticline.

Without detailed information on the tectonics of the broad region between Stjørdalen and this Tømmerås area, it is impossible at this stage to attempt any overall correlation of minor fold types. Indeed, it is highly unlikely that at any one time and in any one lithological horizon, folds of consistent style and trend would have been developed over such a wide area. The possibility that similarly styled folds may have developed at different times must not be overlooked, moreso as this phenomenon is demonstrable in Stjørdalen.

## Proposed major structural correlations across the northern Trondheim region

The factors which govern the variable development and style of minor folds are equally applicable to regional structures so that the tracing of major folds or fold systems across large areas must be effected with caution. With the unravelling of the structural picture in the Stjørdalen-Teveldalen area it is nevertheless now possible to suggest extensions of the major folds northwards into the Tømmerås-Hegsjøfjell region.

In her Tømmerås paper, Peacey (1964) delineated three tectonic units separated by major thrusts. The Seve nappe, comprising Lower Palæozoic Eocambrian and Pre-cambrian rocks, overlies the Olden Nappe (Oftedahl, 1955) but is itself overlain by an upper nappe of allochthonous Palæozoic metasediments. The situation is best seen in Peacey's fig. 34, the thrust plane separating the Seve and upper nappes describing a wide arc concave to the south-west and dipping inwards on all sides.

This same map together with her fig. 31, reproduced here as fig. 47, depicts the Verdal synform which affects the upper nappe and dies out northwards towards and possibly across the basal thrust plane. The Hegsjøfjell area, forming part of S. Foslie's map sheet "Jaevsjø" published by NGU, exhibits a remarkable large-scale example of refolding. With the Hegsjøfjell conglomerate as a marker horizon, it can be seen that a tight or isoclinal structure is deformed by the Verdal synform (Fig. 47, Plate III). Since the latter is one of Peacey's 'regional folds' the refolded fold is therefore representative of an even earlier deformation. The axial trace of the early fold on the western limb of the Verdal synform is drawn within the northward extension of the Gula Schist Group. In this area, Peacey interpreted the early structure as a fold closing eastwards, seemingly on the basis of the Hegsjøfjell closure indicating an antiform plunging to the S.W. (Fig. 48A).

In his section across the Verdal map-sheet Wolff (1960) indicated the presence of an anticlinal fold containing the Skjækerdals schist (= Gula

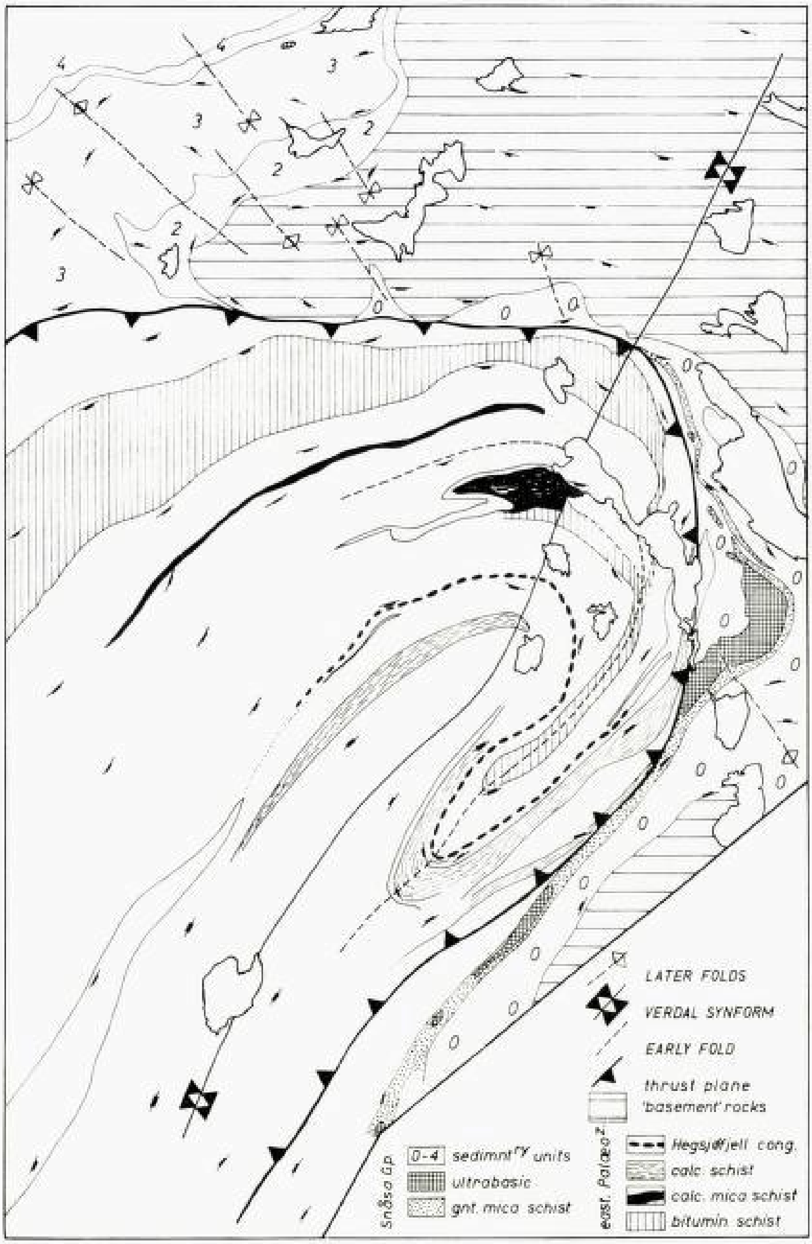



Fig. 47. Simplified stratigraphy of the Verdal synform to show the early fold. Figure from Peacey (1964).

*Forenklet stratigrafi fra Verdalssynklinalen som viser den tidlige fold. Etter Peacey (1964).*

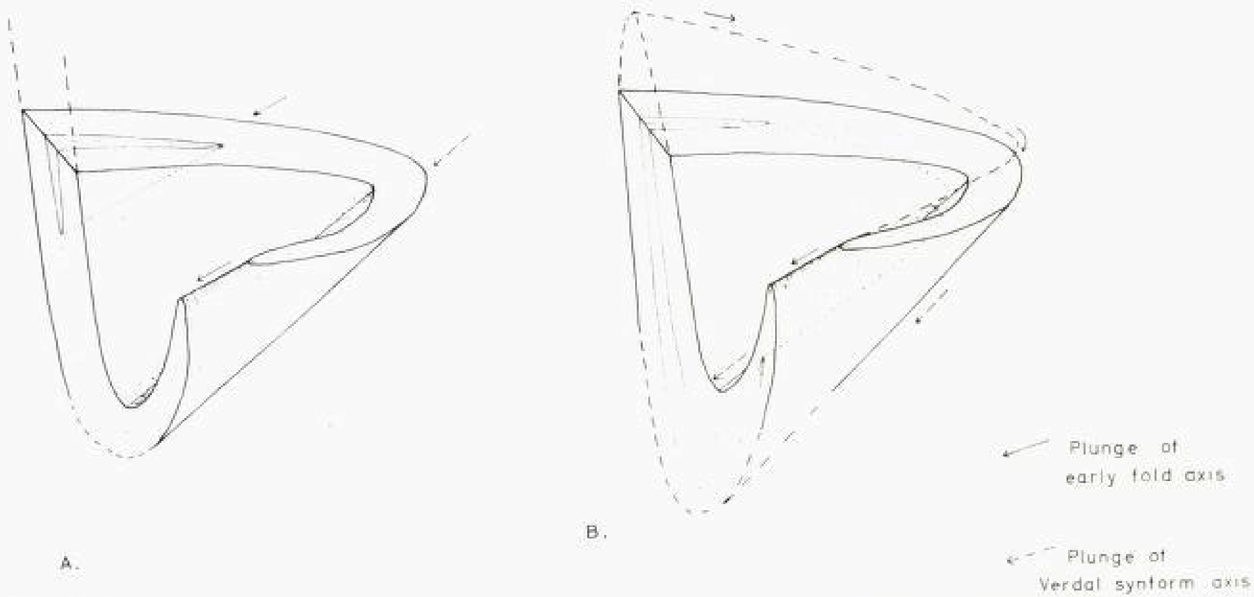



Fig. 48. Diagrammatic representation of the Hegsjøfjell — Imsdalsfjell early fold. (A) Peacey's interpretation drawn from the accompanying sections (1964). (B) The present interpretation. For explanation see text.

*Diagram som viser* ***Hegsjøfjell — Imsdalsfjell's*** *tidlige fold.* ***A)*** *Peacey's tolkning* ***tegnet etter et medfølgende profil*** *(1964).* ***B) Tolkningen i det nærværende*** *arbeid. Forklaring finns i teksten.*

Schist Group) in its core. In this Verdal area dips within these same schists have steepened, and in the south of Wolff's map-area the eastern limb of the anticlinal structure dips W.N.W. having turned over through the vertical. Further to the south along the strike, this links up directly with the major anticlinal structure in Stjørdalen.

With regard to Peacey's representation of the early structure, the Hegsjøfjell closure is most certainly antiformal (or anticlinal since the stratigraphy is now better known), the axis plunging to the south-west. In the present writer's interpretation the whole structure is seen as an anticlinal fold plunging towards a north-easterly point: as a consequence of the subsequent folding of this structure by the Verdal synform, that part of the early fold on the eastern limb of the Verdal synform now plunges south-westwards (Fig. 48B). The Stjørdalen anticline can then be correlated with the early Hegsjøfjell fold despite changes of style and possibly of plunge along the axial direction.

In this northern region the Verdal synform develops as a recognisable fold in the vicinity of the basal thrust of the upper nappe. Followed south-westwards along its axial trace, a remarkable change of style is manifest within a distance of some 20—25 km. The axial plane first dips steeply to the S.E.; it then becomes vertical and, further south, dips towards a north-westerly point. At the same time the fold changes from an open to near-isoclinal style (Fig. 49). Further to the south-west on the Verdal map-sheet Wolff's (1960) initial interpretation of the structure was of a provisional nature and based on tentative stratigraphical correlations. His current view (personal communication) is that a synclinal fold is present in this eastern Verdal region involving Hovin Group rocks.

In the earlier part of this article the Teveldal syncline was described and traced in a north-north-easterly direction to beyond the Kjølhaugene mountains. It is the writer's opinion that this fold is the southward continuation of the Verdal synform. One of the main conclusions from the work in the Stjørdalen-Teveldalen region concerned the fact that the Teveldal syncline is younger than the Stjørdalen anticline. The tracing of these structures into the Hegsjøfjell area thus corroborates this conclusion, since there the Verdal synform actually deforms the major early fold on a grand scale (Plate III).

The thrust plane at the base of the upper nappe is almost certainly equivalent to the Grense thrust. It will be recalled that the Grense thrust was considered by Törnebohm and subsequent workers as a thrust subjacent to a 'sub-nappe' of the underlying major Seve nappe. Thus Peacey's upper nappe is most probably equivalent to this Seve sub-nappe. While the position of the

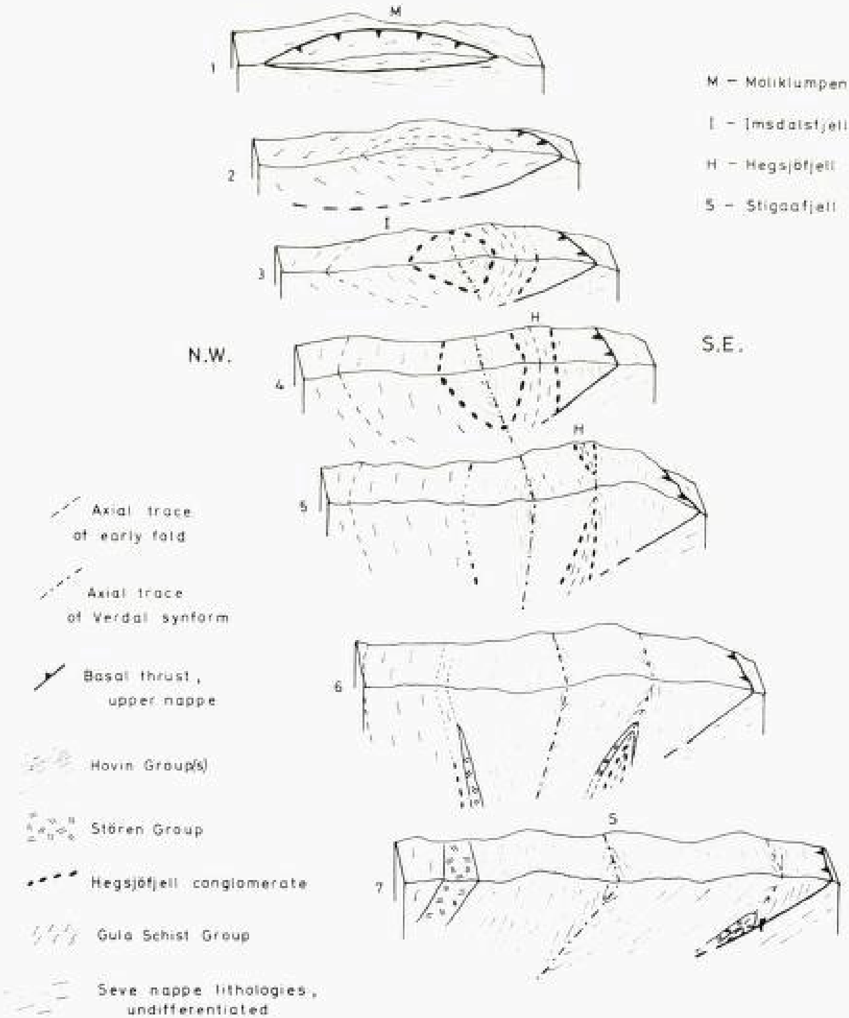



Fig. 49. Sections across the Verdal synform to show (a) its changing development and (b) the early fold. Distance between sections 1 and 7 about 22 km.

*Profil tvers over Verdalens synform som viser a) dens vekslende utvikling, b) den tidlige folden. Avstand mellom profil 1 og 7 omkring 22 km.*

thrust plane is known both from Peacey's investigations in the north and from the present work in the south of the region, its precise positioning in the intervening areas is uncertain. Wolff's (1964) map depicts a definite tectonic break between these two areas whereas the thrust plane should have been drawn as an inferred feature (Wolff, personal communication).

The actual trace of the thrust plane over this intervening area is probably to be found mostly on the Swedish side of the border perhaps approximating to Törnebohm's boundary between his Åre schist and Røros Schist Group, the former Eocambrian and the latter now known to be equivalent to part of the Upper Hovin Group (Upper Ordovician). Moreover, the Köli Schists of the Tännfors area are also floored by this same thrust plane and are part of the upper nappe or Seve sub-nappe.

As to the age of this thrusting, studies in the Kopperå-Riksgrense area have indicated that it can be related to the concluding stages of the second fold phase, that is, the deformation producing the Teveldal-Verdal fold. Peacey (1964) was of the opinion that the thrusting preceded the generation of the Verdal synform, since the thrust plane appears to be folded across the synformal axis. It can at least be stated with certainty that the thrusting post-dated the first folding — part of the major early structure is truncated by the thrust plane.

From an examination of strike symbols and lithological boundaries on Peacey's and Foslie's maps, only a very gentle swing of strike is perceptible across the axial trace (the trace of the Verdal synform axis as drawn by Peacey) at or below the horizon of the thrust plane. North of the thrust the drawing of an axial trace of a definite major structure would indeed seem to be rather conjectural despite the slight bend in the stratification of the leptites and quartz porphyries underlying the upper nappe. South of the thrust plane a pronounced strike swing becomes increasingly apparent such that the axial trace of a major synform can be drawn with little difficulty.

The suggestion is, therefore, that the Verdal synform as an obvious major structure can be largely, though not wholly, restricted to the upper nappe. It is quite feasible that the degree to which the Palæozoic metasediments were folded differed markedly from that of the basement leptites and porphyries, possibly because of differences either of tectonic level or lithology, or both. During deformation such a difference in lithology may conceivably have assisted in promoting the thrusting: the regional arcuate form of the basal thrust can thus be seen partly as mimicking the second phase synformal structure although to some extent being controlled by the pre-tectonic disposition of the basement. Prior to the onset of the Caledonian orogeny the Tømmerås massif, for example, was considered to have been a submarine ridge (Peacey 1964, p. 63) which influenced the siting of subsequently developed structures. On the other hand, accepting Oftedahl's (1955) view of the upheaval of the Grong culmination as being a relatively late phenomenon, this can also be envisaged as influencing the present disposition of the thrust since such a transverse ridge of positive movement would clearly result in a south-westward tilting of structural elements in the Tømmerås-Hegsjøfjell region.

Further to the north-west the Tømmerås anticline and Snåsa syncline are clearly of a later age than the earliest folding and have been regarded as complementary structures to the Verdal synform. The writer's investigations in the Snåsa area confirmed that the Snåsa syncline deforms the earlier developed schistosity and therefore the Seve nappe and basal thrust plane; this is best seen on Peacey's fig. 34. Perhaps the ultimate answer to the question of the age of the thrusting of the upper nappe lies in the recorded fact that nappe emplacement and thrusting has not taken place in one single episode of movement. This is known from various parts of the Caledonian mountain chain. Strömberg (1961), for example, has demonstrated that in Jämtland the Helags nappe, a higher sub-nappe of the Great Seve nappe, has been thrust at a later stage than the underlying Särv nappe, the basal division of the Great Seve nappe.

It is not unreasonable therefore, to suggest a similar sequential thrusting in the northern Trondheim region, the basal thrust of the upper nappe being slightly younger than the main Seve thrust. The upper nappe thrust could then be conceived as developing during the later stages of the second folding as suggested earlier, and locally the folding may indeed have continued for a short time after the thrust plane had been generated, thus helping to explain some of the anomalies. The proposed sequence of events in the general Hegsjøfjell area can now be summarized as follows:

1. First folding, syn-tectonic metamorphism and nappe development. Generation of the early Hegsjøfjell fold. Seve nappe developed during this period.
2. Second episode of folding. Though essentially one episode, this can be sub-divided into: —
   a) Main deformation, producing the Verdal synform: folding less readily impressed on basement lithologies.
   b) Thrusting — basal thrust plane of Peacey's upper nappe.
   c) Continuation of folding (now weakening) subsequent to the actual thrusting.

3. Upheaval of the Grong culmination — possibly initiated late in the second deformation episode. (The late folds recognized in the Tømmerås area (Peacey, 1964) and the third phase folds of the Stjørdalen-Teveldalen region clearly deform structures of the first two episodes but their precise time relationship to the Grong culmination is not known.)

In an instructive article on the nappe tectonics in the Grong region Oftedahl (1955) described two movement episodes, the major thrusting being related to the older episode whilst folding and further thrusting were ascribed to a younger deformation. With reference to the younger episode, Oftedahl described the thrust plane of a nappe (thrust plane D2 on his map, Plate 1) which cuts the Gjersvik nappe. This younger nappe was thought to be part of either the Gjersvik nappe or a more strongly metamorphosed western nappe — the western nappe of Strand (1953). The upper nappe was not then recognized as a separate tectonic unit but it is possible that the thrust plane of this upper nappe is of approximately the same age as Oftedahl's (1955) thrust plane D2 associated with the Gjersvik nappe.

However, such speculative correlations would involve a more comprehensive discussion on much wider topics and problems within the mountain chain as whole, a subject which is beyond the scope of this paper. It would for instance, necessitate discussion of the Rødingfjell nappe of Nordland and its possible extension into the Trondheim region (Oftedahl 1966), as well as a review of the movement and metamorphic episodes over a large tract of the Scandinavian Caledonides. In this regard it is interesting to note that the detailed work in Nordland over the past decade has demonstrated the presence of three episodes of folding (see, e.g., Rutland and Nicholson, 1965). While this invites comparison with the polyphase movements demonstrable in the Trondheim region, major differences particularly with regard to thrusting and the acme of the main metamorphism in relation to movement phases are thought to exist. In order to resolve such differences, future inter-regional tectono-metamorphic correlation may necessarily involve a concept of non-synchronism of particular events. Indeed there is no rule requiring that particular tectonic or metamorphic events occur simultaneously in different parts of an orogen.

Discussion of suggested major structural correlations across the northern Trondheim region has at this stage purposely been restricted to the easternmost part of the region simply because large areas south and south-west of the Tømmerås anticline require either detailed mapping or remapping. It is more than likely that one or two early isoclines are present along the flanks of the Tømmerås massif which can be traced in a south-westerly direction. Sliding or thrusting is also thought to be of no mean significance in explaining the present disposition and rapidly varying thicknesses of certain lithologies.

### Concluding discussion

Structural investigations in the Stjørdalen-Meråker-Riksgrense region of Trøndelag have demonstrated the presence of folds and associated minor structures belonging to three distinct episodes. Two folds of major proportions are present. The Stjørdalen anticline, a fan-shaped structure displaying features of extreme shearing and isoclinal folding, was developed during the oldest phase of deformation. The Teveldal syncline further to the east is shown to have been generated during the second movement phase. A penetrative cleavage axial planar to fairly open minor folds is the most conspicuous feature of the third and latest deformation.

Minor folds axes and lineations of the first phase reveal a wide but systematic variation of trend, generally between S.W. and N.W., the gradual swing being explained by a combination of mechanisms, namely syn-tectonically generated curvature and superimposed similar folding.

The Teveldal synclinal axis plunges gently to the S.W.—S.S.W., while its axial plane dips at 35—40° towards a west-north-westerly point. Fold axes of the latest deformation phase plunge consistently at some small angle towards N.—N.N.E. They are normally overturned down-dip, axial plane cleavage being inclined eastwards, though conjugate structures occur in phyllite where the foliation is disposed steeply or near to vertical.

There is evidence to suggest that, while the main regional metamorphism is largely coeval with the first isoclinal folding as indicated by the common schistosity axial planar to these folds, some mineral growth continued after the movements had ceased. This appears to be true of garnet and muscovite in schists from the Gula Schist Group, but little petrographical work has so far been carried out. Biotite porphyroblastesis in the east of the region would appear to be dated to the latter part of the second fold phase, prior to the actual thrusting.

Although time did not permit their closer investigation, observations of the irregular bodies, dykes, veins and pods of trondhjemite in the Gula Schist Group furnish valuable evidence regarding the relationship of emplacement to movement phases. It would seem that such bodies were emplaced both during and subsequent to the first folding but prior to $F_2$; some may be pre-$F_1$. Both injection and replacement features are present and many veins of trondhjemite are of a non-dilational nature. The relative abundance of trondhjemite in the core of the Stjørdalen anticline strongly suggests a close relationship of emplacement, main metamorphism and first folding. Thus, during the isoclinal folding and intense shearing representative of the oldest deformation phase granitic material can be visualized as moving upwards in the fold core, this being more or less coeval with the acme of metamorphism.

Correlations of the Stjørdalen and Teveldal folds with major structures occurring further north in the Tømmerås-Hegsjøfjell area have been proposed. In this latter area the mutual relationship of these folds is depicted in a large-scale example of refolding. (Plate III).

It is noteworthy that the initial major folding is concentrated in the central zone of the 'depression' containing the Trondheim eugeosynclinal sediments. To the east the Teveldal syncline has been shown to be younger than the Stjørdalen fold. Relatively little has been published on the deformation sequences in the western parts of the Trondheim Cambro-Silurian. It is reasonable to suppose that the Hølonda-Horg syncline (Vogt 1945) S.S.W. of Trondheim may be a second generation struture: this needs to be re-investigated.

Structurally the Trondheim-field has constantly been referred to as a synclinorium despite the variable interpretation of the structure in the central parts of the region. Previous writers have rarely explained their reasons for adhering to the synclinorium concept and it seems that opinions have been influenced firstly by Kjerulf's original interpretation and secondly, and perhaps more significantly, by the fact that one is dealing with a supposed sedimentary trough-shaped depression or segment of an original geosyncline. Thus the 'syncline' part of 'geosyncline' has dominated structural thought to the extent that the synclinorium or syncline idea has rarely been questioned.

Taking into account present-day stratigraphical and tectonic knowledge this old concept would appear to be misleading. It would be hard to disagree with Strand's (1961) statement that there is, "no reason to consider the sediments of the Trondheim region to be in an autochthonous or parautochthonous position". Peacey (1964) has demonstrated the presence of two tectonic units of Palæozoic rocks in the extreme north of this region. The present writer's investigations across Stjørdalen lend support to the idea of a southward continuation of the upper nappe or sub-nappe of the Great Seve nappe. Moreover, from a compilation of maps and field data reported by several geologists, Wolff (see accompanying paper in this NGU volume) has traced the thrust plane of this upper nappe into the extreme southern part of the Trondheim region. It is therefore highly probable that a large part of the Trondheim-field sequence is allochthonous and Wolff has tentatively adopted the term 'Trondheim nappe'. The major structure, despite is complexities, would however, appear to be an anticlinorium with recumbent folds overturned northwestwards in the west (Peacey 1963) and towards the E-SE in the east, and with the oldest and generally more strongly metamorphosed rocks in its core.

Ramberg (1966) has published results of experiments involving the centrifuging of models of unstable geosynclinal belts. Parts of some of his illustrations of structures produced in these experiments show a remarkable resemblance to the general structure advocated by the present writer in Stjørdalen. It would be unsafe to assume that gravity has played a predominant part in the development of the Stjørdalen structure however, although its possible influence in generating the movements cannot be ignored. Aubouin's (1965) subdivision of complex geosynclines into 'divergent' and 'convergent' on a symmetry basis is interesting in this respect as the Caledonian geosyncline in its entirety has all the attributes of his divergent or centrifugal type.

Evidence regarding the age of the deformation in this part of the Caledonides is insufficient for more than general comment, but it is important to remember that Horg Group rocks of Lower Llandoverian age are affected by the first phase of folding. From evidence on the island of Hitra west of Trondheimsfjord, Strand (1961) has noted that "the main Caledonian orogeny in the central parts of the mountain chain was ended before the beginning of Devonian time". In Jämtland, Thorslund (1960) has reported the Ekeberg Graywacke of the Föllinge nappe as being of Wenlock age: therefore the thrusting of that nappe must be post-Wenlock. In Swedish Lappland the youngest rocks in the Seve-Köli nappe are dated by graptolites to upper Middle Llandovery (Kulling, 1960).

It is thus reasonable to regard the main Caledonian deformation and metamorphism in Trøndelag as having occurred sometime between the Upper Llandoverian and Lower Ludlovian of the Silurian. A more precise dating is not possible just now. Ordovician crustal movements in the Trondheim region have been well documented (Holtedahl 1930, Vogt 1945) and are manifested in conglomeratic horizons which tend to be thicker and more extensive in western and north-western districts. Holtedahl has noted the increasing importance of these crustal disturbances towards the more western (and central) parts of the Caledonian zone. A particularly important break occurs above the volcanic sequence of the Støren Group (the "Trondheim disturbance" of Holtedahl) in the west and it is not improbable that folds may have been produced still further west in the central parts of the geosyncline during late Tremadocian or early-Arenigian times.

Tremadocian or early-Arenigian times. Another significant point is that the Lille Fundsjø Conglomerate and other basal Lower Hovin conglomerates contain pebbles of Støren Group (and older) greenschists and intrusives. This would appear to suggest that the pre-Hovin sediments and volcanics were partially metamorphosed prior to the deposition of the conglomerates, as hinted at by Vogt (1945).

**Sammendrag**

I den første delen av artikkelen beskrives tektoniske småstrukturer i svakt metamorfoserte sedimenter tilhørende Hovin- og Horg-gruppene langs Teveldalen øst for Meråker. Det er påvist tre foldeperioder som alle viser forskjell i stil og akseretning. Første fase av småfoldeakser og lineasjoner fremkaller en stor, men systematisk variasjon i retning vanligvis mellom sydvest og nordvest. Den ene hovedstrukturen, Teveldalsynklinalen, viser seg å være dannet under den andre bevegelsesperioden. En viktig struktur som tilhører tredje foldefase er en gjennomtrengende akseplankløv i de hyppige småfoldene med akseretning mellom nord og nordøst. Forskyvning av sedimentpakken i retning øst til sydøst henger sammen med det siste stadium av andre foldingsgenerasjon.

Heretter følger en beskrivelse av tektonikken i Stjørdalen vest for Meråker, og det vises spesielt til hovedstrukturen i området i en vifteformet antiklinal, kalt Stjørdalsantiklinalen. Denne antiklinalen, som er en førstegenerasjonsstruktur, dominerer det tektoniske bildet i den nordlige del av Trondhjemsfeltet. Etter all sannsynlighet kan denne strukturen spores videre sydover i den sentrale del av Trondhjems-'depresjonen'. Det foretas en sammenligning av foldeepisoden i Teveldal- og Stjørdalsområdet, og småstrukturer kjent fra tilstøtende områder blir studert i forhold til de forskjellige foreliggende foldebevegelser.

Den andre delen av artikkelen beskjeftiger seg med sammenligning av foreslåtte hovedstrukturer tvers på den nordlige del av Trondhjemsfeltet. Hegsjøfjell-området, som er kartlagt av S. Foslie og kort beskrevet av J. S. Peacey (1964), fremviser et eksempel på gjentatt folding i stor skala. Ved nyvurdering av karakteren til den tidligere isoklinal og videre undersøkelse av Verdalssynformens skiftende stilling (Peacey 1964) og strukturene tvers over Verdalsområdet (Wolff 1960), synes en sammenligning av (a) Stjørdalsantiklinalen med den tidlige Hegsjøfjell-folden og (b) Teveldalsynklinalen med Verdalssynformen, å være meget akseptabel.

Til slutt synes betraktninger av skyvninger i Tømmerås-Hegsjøfjell og sentrale Trøndelag-Jämtlandområdene å støtte Strand's (1961) antakelse om at metasedimentene i det utstrakte Trondheimsfeltet er alloktone. Etter disse betraktninger er det meget mulig at Peacey's (1964) øvre dekke kan spores ned til den sydlige del av Trondheimsfeltet (se Wolff f.f.).

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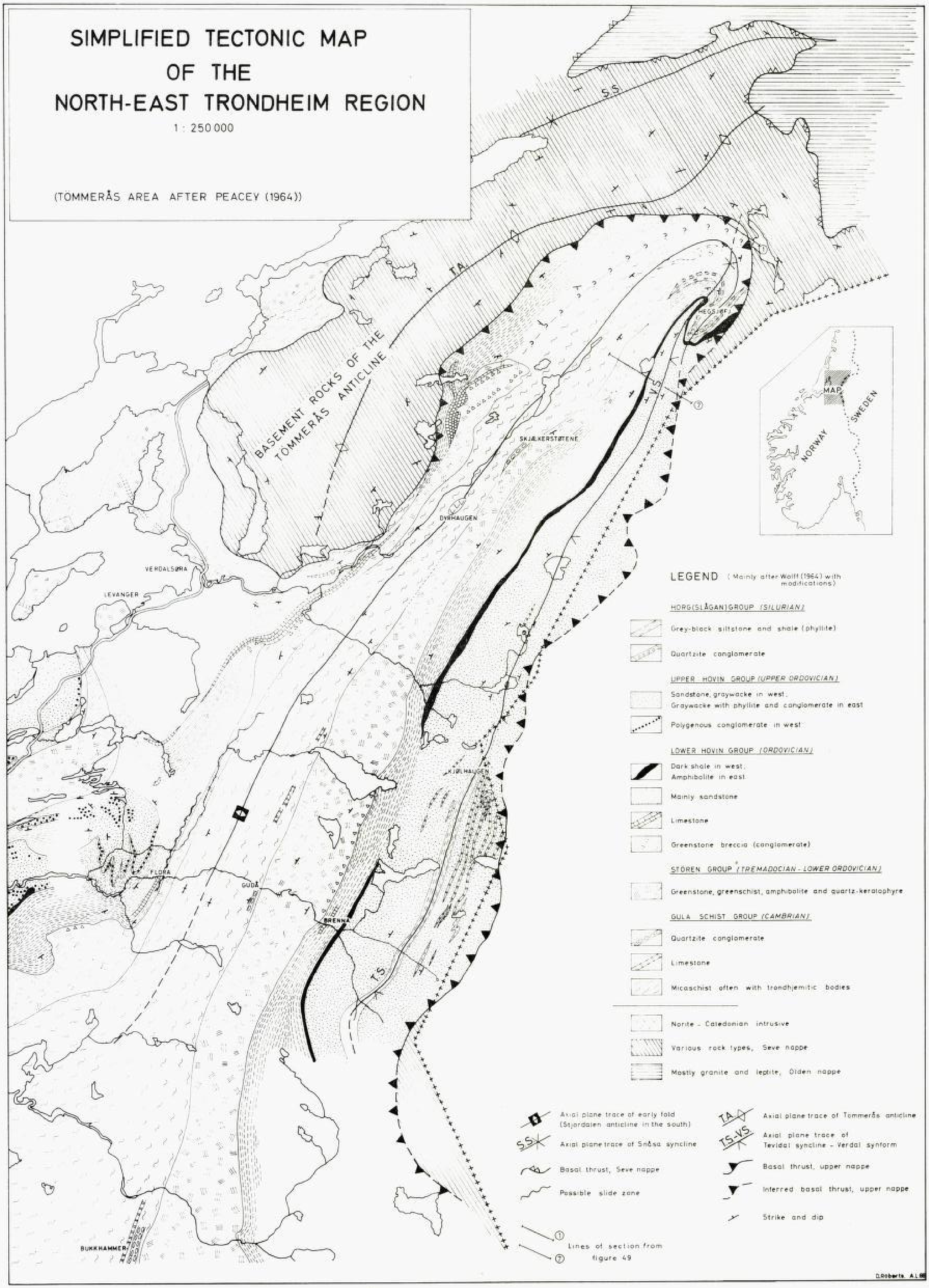
SIMPLIFIED TECTONIC MAP
OF THE
NORTH-EAST TRONDHEIM REGION
1 : 250 000
(TÖMMERÅS AREA AFTER PEACEY (1964))
BASEMENT ROCKS OF THE TÖMMERÅS ANTICLINE
TA
S.S
V.S
T.S
HEGSJØEN
SKJÆKERSTØTENE
DYRHAUGEN
VERDALSØRA
LEVANGER
KJØLHAUGEN
FLORA
GUDÅ
BRENNA
BUKKHAMMER
MAP
NORWAY
SWEDEN
LEGEND (Mainly after Wolff (1964) with modifications)
HORG(SLÅGAN)GROUP (SILURIAN)
Grey-black siltstone and shale (phyllite)
Quartzite conglomerate
UPPER HOVIN GROUP (UPPER ORDOVICIAN)
Sandstone, graywacke in west.
Graywacke with phyllite and conglomerate in east
Polygenous conglomerate in west
LOWER HOVIN GROUP (ORDOVICIAN)
Dark shale in west.
Amphibolite in east
Mainly sandstone
Limestone
Greenstone breccia (conglomerate)
STÖREN GROUP (TREMADOCIAN - LOWER ORDOVICIAN)
Greenstone, greenschist, amphibolite and quartz-keratophyre
GULA SCHIST GROUP (CAMBRIAN)
Quartzite conglomerate
Limestone
Micaschist often with trondhjemitic bodies
Norite - Caledonian intrusive
Various rock types, Seve nappe
Mostly granite and leptite, Olden nappe
Axial plane trace of early fold (Stjørdalen anticline in the south)
S.S Axial plane trace of Snåsa syncline
Basal thrust, Seve nappe
Possible slide zone
Lines of section from figure 49
TA Axial plane trace of Tommerås anticline
TS-VS Axial plane trace of Tevidal syncline - Verdal synform
Basal thrust, upper nappe
Inferred basal thrust, upper nappe
Strike and dip
D.Roberts A.L.68

# II. Geology of the Meråker area as a key to the eastern part of the Trondheim region

by F. Chr. Wolff.

**Abstract**

The present paper deals with a discussion of the main tectono-stratigraphical problems of the eastern Trondheim region. Based on detailed studies made in the Meråker area in the last three summers and available manuscript maps and personal communications from geologists working in the southern parts of the region, it has been possible to establish a general stratigraphy for this eastern region and revise the stratigraphical nomenclature, distinguishing the Gula Schist Group from the Røros Schist. It is also pointed out that on evidence principally from the northern half of the region, the former 'Trondheim synclinorium' has to be regarded as an anticlinorium. It is shown that a major thrust plane occurs along the larger part of the border of the Trondheim region and the term 'Trondheim nappe' is suggested for the allochtonous metasedimentary sequence above this plane. The occurence of serpentinite bodies along the thrust plane is discussed, as is the distribution of the pyritic and chalcopyritic ores over the entire Trondheim region.

**Introduction**

In the last decade a good deal of mapping has been carried out in different parts of the eastern Trondheim region, from the Sel area in the south to the Jævsjø area in the north. No attempts to compile all this data have been made until now.

The present writer is fully aware of the difficulties involved in performing such a compilation and while he accepts the risk of making errors on a small scale, the conviction that such work would be of considerable value to those pursuing studies in this field has tended to outweigh all the risks involved. It may well be that certain of the conclusions which here will be drawn from the different observations made over this broad region will later be proved to be incorrect, but on the basis of present-day knowledge the writer is of the opinion that an attempt to put all this data together will point out the main problems and assist in determining where and how to start the most fruitful investigations in the years to come.

**Previous investigations**

In this chapter mainly the investigations made previous to the last decade will be mentioned.

In 1838, 1844 and 1850 B. M. Keilhau published his revue of Norwegian geology, "Gæa Norvegica", in three volumes. In part four of the last one he gives a description of the area which is today known as the Trondheim region. Keilhau shows a general knowledge of the principal rock types and the broad outlines of the strike and dip of the schists of the area. Although he expresses himself in a rather vague manner with regard to the question of the tectonic development of the area the principle conclusion to be drawn is that he considers the Trondheim area as folded in a large syncline. This can be seen from the following statement, "Zwischen Vaage und Læssøe (Vågå and Lesja) fängt mit nördlichem und nordwestlichem Fallen eine Schichtenzone an, welche mit nordwestlichem Fallen über Foldalen (Folldal) im Tønset (Tynset) hinein fortsetzt, in dem H a n g e n d e n (present writer's italics) derselben folgt eine eben so lange vertikale Zone, und jenseits dieser wieder Schicten, die südlich- und südöstlich einschiessen, dann aber in der Gegend am Stu-See und um Bubakken in Qvikne sich gegen Norden mit östlichem Fallen schwingen, und noch nördlicher sogar umbiegen, um beinahe nord-nordwestlisches Streichen mit Fallen gegen ONO anzunehmen. Eine Art Regelmässigkeit in der Schichtenstellung findet sich also in dieser besonders im Süden und Osten an Dovrefjeld angeschlossenen Gegend, und die Regel führt zu einem Streichen ganz quer über die Kette, die man sich hier zu denken pflegt."

After Keilhau Th. Kjerulf worked in the Trondheim region, from 1866 with K. Hauan as his assistant. The results of this work was published in two papers; "Om Trondhjems stifts geologi", 1871 and 1872, which gave a lot of descriptive information about the Trondheim region. With regard to the stratigraphical and tectonic problems Kjerulf also favoured the idea that the whole sequence was folded in a syncline. He published a stratigraphical table suggesting the "Gulaschists" as the youngest, Støren, claystone and schists, conglomerate and sandstone layers as the middle and Røros and Trondheim schists as the oldest part of the sequence. Later in his paper, "Udsigt over det sydlige Norges geologi" (1878), he still held the same opinions about both stratigraphy and tectonics, though he then discussed (p. 176) the possibilities of an inverted system with the oldest rocks in the middle of an anticlinal structure. However, since he was not able to locate the thrust planes necessary for such an interpretation, he abandoned this alternative. Largely the same view is also found as a basis for Kjerulf's interpretation in his next paper, the "Merakerprofilet" (1882).

In contrast to this opinion are the views of the Swedish geologist A. E. Tørnebohm (1872) and his countryman F. Svenonius (1885). Svenonius discusses the opinions of Th. Kjerulf and Tørnebohm and concluded that the Meråker profile must be interpreted as an anticlinal structure with the youngest rocks in overturned synclines on either side of an older central core zone, an opinion which seemed to be accepted by the head of the Norwegian Geological Survey, Hans Reusch, in his paper of 1890. At this time Reusch was able to involve A. Getz's discovery of graptolites at Kjølhaugene in Meråker in the same discussion.

A. E. Tørnebohm in his memoir on Central Scandinavian geology (1896) discussed his ideas for the general geological structure of the Trondheim region. He pointed out the NNE — SSW trend of the mountain chain and declared that amongst all the synclines and anticlines in the region there are two "nuclear lines", the syncline in the east along the Swedish—Norwegian border and the syncline along the Trondhjemsfjord. But as his picture of the stratigraphy is partly misleading he was not able to clear up the overall tectonic picture of this broad district.

Carl Bugge (1912) rejected the views of the Swedish geologists and supported Kjerulf's old opinion that the "Gula Schists" must be the youngest stratigraphical group. Bugge drew a profile across the mountain chain which indicates a syncline with the "Gula-schists" in the middle (fig. 50). Gunnar Holmsen (1915) held the same view.

C. W. Carstens (1920) revived the opinion of Svenonius, Reusch and Tørnebohm but repeated the interpretation of all previous authors in placing the Røros schists in the lower part of the sequence. C. W. Wegmann (1925) regarded the central Gula schist zone as a "grand lambeau de recouvrement", while Gunnar Aasgaard (1927) again stuck to the old opinion of Kjerulf.

Th. Vogt (1940) emplyed a revised study of an old find (by the late J. H. L. Vogt, his father (1888)) of *Dictyonema flabelliforme* at Nordaunevold to try to strengthen the idea of a synclinal structure for the Trondheim region. He also supported the opinion that the Røros schists were older than the Gula schists by mistaking the schists at Nordaunevold as belonging to the Røros schist division. In this attempt he succeeded so well that no one working

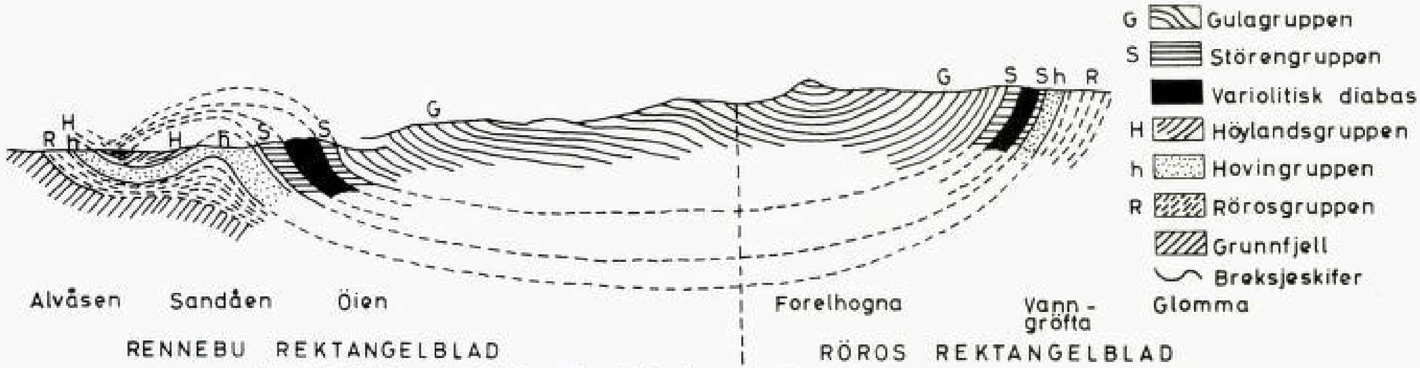


Fig. 50. C. Bugge's (1912) section indicating a syncline with the Gula Schists as the youngest beds.
*C. Bugge's (1912) profil som antyder en synklinal med Gulaskifrene som de yngtse lag.*

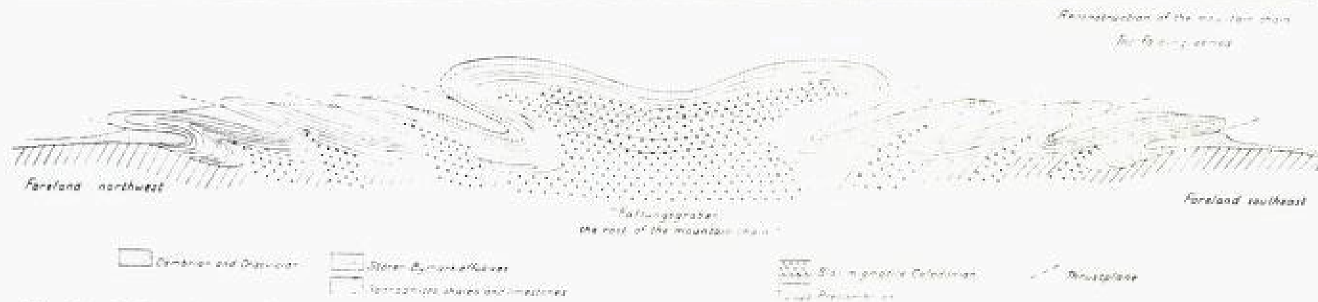


Fig. 51. C. Bugge's (1954) section across the Norwegian Caledonian mountain chain from NW to SE with reconstruction of the folding. The section also tends to show that the "Faltungsgraben" (the root of the mountain chain) to a certain extens is a depressed geanticline.
*C. Bugge's (1954) profil gjennom fjellkjeden fra nordvest til sydøst med rekonstruksjoner av foldingene. Profilet skal også vise hvorledes foldingsgrøften kan tydes som en delvis sammensunket geantiklinal.*

in this area opposed him until Carl Bugge did in 1954. Bugge published a profile across the mountain chain (Fig. 51) and a map demonstrating his idea that this eastern part of the Trondheim region is analogous to the western part. He correlated the greenstone of the Folldal — Haltdal area with the Støren greenstones and concluded that the greenstones are overlain in a syncline by the Røros schists, which are equivalent to the Hovin — Hølonda Group in the western Trondheim region. He also stated that although the greenstones overlie the Røros schists, in his opinion this is only a tectonic feature and in reality the Røros schists are younger than the greenstones. Unluckily Bugge's views were never accepted (this is demonstrated by the fact that his ideas were never lectured nor cited) before the present writer (1964) arrived at a similar, but better established stratigraphical and tectonic picture based on accurate field observations from detailed studies in the northern half of the Trondheim region.

Using the evidence from these northern areas the present writer was able to re-analyse the available data from the southern half of the region, based mainly on manuscript maps from his former colleague state geologist Johs. Færden, who had worked for a period of 10 years in the Røros district (Quadrangles: Stuesjø, Aursunden, Røros, Haltdal, Tynset and Kvikne).

### Acknowledgements

Thanks are first of all due to my former colleague Johs. Færden for permission to use his manuscript maps and report on the Røros area, and to Hans Heim, Folldal, for similar reasons.

A student group under the guidance of Professor J. A. W. Bugge is carrying out a programme of ore research in the Killingdal area. The present writer is indebted for data obtained from members of this group.

### Discussion based on recent investigations

We shall now turn to the discussion of the broad tectono-stratigraphical picture in the eastern Trondheim region, see table I, a picture which can be constructed from the studies made over this region in the last ten years.

*The Stjørdal - Meråker area*

As this area has been so thoroughly described earlier in this NGU volume, it is necessary here only to sum up what the writer considers to be the main results of these studies with regard to the present discussion.

As the present writer points out in an earlier paper (1964) he regards the profile across the Caledonian mountain chain from Stjørdal to the Swedish border as a fan-shaped anticline with the oldest rocks, the Gula Schist Group in the central part and the younger Støren and Hovin rocks in overturned synclines on either side. In view of the resulting controversy surrounding this statement, it was felt that certain more specialised and detailed studies in the area would help to clarify various points and perhaps settle the main arguments. The results of the small scale tectonic studies are given by D. Roberts in his article. He has been able to support the anticlinal interpretation by much stronger arguments based on a better and detailed knowledge of the structure. The finding of the Lille Fundsjø conglomerate by F. Fediuk proved the inversion of the sequence in the eastern area as did the detailed study of the sedimentary structures by A. Siedlecka. The views of the present writer are therefore more firmly established today than in 1964.

*The extension of the thrust-plane*

As pointed out by the present writer (1960 and 1964) and by Peacey (1964) there is a thrust plane separating the basement from the Cambro-Silurian rocks in the northern part of the Trondheim region. Peacey (1964) moreover was able to interpret the rocks resting on this thrust plane as constituting an upper nappe above the Seve nappe. Oftedahl (1966) states that, "As yet it remains to be seen if this upper nappe can be compared with the Rødingfjell nappe farther north and if this new nappe can be traced into the central parts of the Trondheim district".

Our mapping over the last two summers has shown that this thrust plane is traceable southwards as far as half-way down the quadrangle Essandsjø (Pl. IV). On Quadrangle Stuesjø it coincides with the thrust plane marked on Schaar's (1961) map, although this writer interprets the rocks overlying the thrust-plane as belonging to the Seve nappe. Færden (personal communication*) regards the augen gneiss zone extending from the lake Langen on quadrangle Stuesjø to Brekken on quadrangle Aursunden, as being derived from metamorphosed mylonites belonging to a thrustplane. This zone, which links up with the above-mentioned thrust plane of Schaar, extends southwestwards across the quadrangles Aursunden and Røros and joins up with the augen gneiss zone of P. Holmsen (1950) on the quadrangle Tynset. Heim (1966) also marked this augen gneiss zone on his map, quadrangle Folldal, and as the thrustplane is overlain bye the Røros Schists and underlain by the

*) Hereafter personal communication will be cited as p.c.

Eocambrian Sparagmites it is possible to extend the trace of the thrust plane along the border between these two rock-types across the map sheet Dovrefjell (Heim 1966 p.c. on manuscript map). This takes the thrust plane into the Sel and Vågå map area of Strand (1951) where it links with the northern part of his Otta nappe. In this southern area the trace of the thrust plane swings round and proceeds northwards across the western part of the map-sheet Dovrefjell (Heim's manuscript map) and then links up with the augen gneiss zone of P. Holmsen (1960) on the quadrangle Oppdal map-sheet.

At the moment it is not possible to trace it further northwards but it is quite probable that this will be achieved by more detailed mapping in the future.

As can be seen from the map (Pl. IV) the thrust plane surrounds what for a long time has been called the Trondheim region and the present writer therefore suggests the *Trondheim nappe* as a suitable name for the metasedimentary pile occuring above this major thrust plane. This will include the upper nappe of Peacey (1964) in the north and the northern part of the Otta nappe of Strand (1961) in the southwest.

It then remains to be proved whether or not the Trondheim nappe is equivalent to the Rødingfjell nappe in the north (Oftedahl 1966) or the Lower Jotun Nappe in the south (Strand 1951).

*The Gula Schist Group*

The rocks of this group are characterised by different types of mica schists often containing garnet or hornblende. Some horizons also contain staurolite, kyanite or sillimanite (Roberts present volume).

Kjerulf (1871) uses the term Gula Schists for this rock series and Røros Schists for the rocks appearing in this paper as belonging to the Røros Group. As the studies of the Trondheim region proceeded the term Gula Schist lost ground in preference to the latter, and in the recent papers rocks of both groups appear as one, namely the Røros Group. As it has now been possible to distinguish between the two, the olds terms should be re-established. The Gula Schist Group then refers to the mica schists described above and the Røros Group to the metagraywackes of the Røros — Meråker zone in the east which are equivalent to the Upper Hovin Group of the Hølonda — Horg area in the west.

In the northern part of the Trondheim region these rocks contrast markedly with rocks of the Røros Group, the latter containing phyllites and graywackes of the lower part of the greenschist mineral facies. As the degree of metamorphism increases southwards and the rocks of the Røros Group, often contain porphyroblasts of biotite, garnet and hornblende in the Haltdal — Røros area, it becomes increasingly difficult to distinguish between the two groups. For this reason the mapping of these rocks has been problematical in these southern areas. Færden (p.c., manuscript maps 1965) for example, designates the metasediments of both these groups as mica schists.

As it is clearly much easier to distinguish between these two types of schist further north in the vicinity of Meråker, this obvious difference has proved advantageous to the present writer in his examination of the rock-types occurring in the Haltdal — Røros area. Consequently it has been possible to follow the two metasedimentary groups along the strike between the Meråker and Røros areas, despite their changing metamorphic condition.

As already pointed out by Strand (1960), "Objections can be raised to the validity of the term Røros Group..." taken as a group name for the oldest Cambro-Silurian beds of the Trondheim region. As discussed later in the present paper the schist of Røros Group type is almost certainly equivalent in age to the rocks of the Upper Hovin Group in the Hølonda — Horg area (Vogt 1945). The term Røros Group is therefore invalid for the schists below the lower greenstone (Støren Group) in the central zone of the Trondheim Caledonides.

The schists typical of this group outcrop all along the Gauldalen valley from Ålen to Støren and the term *Gula Schist Group* is therefore quite appropriate.

Since the northern distribution of the Gula Schists (then called the Røros Group) has already been outlined in an earlier paper, Wolff (1964), it is only necessary to discuss their distribution south of the Tydalen valley in this chapter.

South of Gauldalen valley Birkeland (p.c. 1966) report rocks of the Gula Schist Group as far southwards as the area east of Dalsjøhøgda, quadrangle Røros. In the western half of the same quadrangle Færden (p.c. manuscript map 1965) has indicated the presence of mica schists. On quadrangle Folldal Heim (manuscript map 1966) indicates that the different mica schists of the "Storhøschieferserien" most likely represent the Gula Schist Group in this area. By comparison with the Sel Micaschist to the west, Strand (1951) Heim interprets this rock series as belonging to the Hovin Group. In the present writers opinion while the comparison with the Sel Micaschist is valid as the Storhø Series continues across the map-sheet Dovrefjell and links up with the Sel Micaschist on quadrangle Sel (Strand 1951), the interpretation of the stratigraphical position is incorrect since the Storhø Series is almost certainly equivalent to the Gula Schist Group.

In the western part of the southern Trondheim region, rocks of this group are also known from quadrangle Oppdal (Holmsen 1960) and Kvikne (manuscript map Færden 1965).

*The extension of the Gudå conglomerate zone*

In an earlier paper, Wolff (1964) the present writer interpreted the Gudå conglomerate zone, represented at different localities in the central part of the northern Trondheim region as lying near the base of the lower greenstone. The distribution of these localities is shown on the map (Pl. IV). These quartzite conglomerates are often accompanied by a band of crystalline limestone. This limestone occurs in the southernmost part of the previous map at Bukkhammeren and is also known south of this locality, on Vollfjell (Volfjell limestone, Vogt 1941). Sørbye (p.c. 1967) reports a deformed, discus-shaped polygenous conglomerate near Vårhus in Hesjedalen. A similar conglomerate is also reported from Harsjøfjell and Øvstubekken on quadrangle Røros (Birkeland p.c. 1966).

A conglomerate of similar type (the Husum conglomerate) appears on Heim's manuscript map Folldal together with a crystalline limestone. As will be seen from the map (Pl. IV) it's stratigraphical situation is indicative of the position of the Gudå conglomerate zone.

A conglomerate located at Buåi in Grimsdalen on map-sheet Dovrefjell has been correlated with the Husum conglomerate (Heim p.c. 1966). The present writer also believes thatt he Skardshø quartzite conglomerate, appearing on quadrangles Sel and Vågå (Strand 1951), belongs to this zone.

No definite reports are known of any conglomerate which could be interpreted as belonging to this zone in the southwestern part of the Trondheim region. A conglomerate near Hjerkinn, quadrangle Dovrefjell (Heim p.c. on manuscript map 1966) is somewhat dubious since it contains boulders of both quartzites and greenstones.

*The lower greenstones — the Fundsjø Group*

The rocks oft this group have been correlated with the Støren Group (Wolff 1964) of the Hølonda — Horg area (Vogt 1945). This zone consisting of basic and acid volcanics as well as gabbroic intrusive bodies, is very easy to trace north and south of the Meråker area. It's extension to the north has already been pointed out in an earlier paper (Wolff 1964). The map accompanying this paper depicts its extension towards Tydal south of the Meråker area, where it links up with the amphibolite group of Kisch (1962). This zone proceeds southwards on the manuscript maps Haltdal, Røros and Tynset of Færden (p.c. 1965). Færden reports pillow structure at Lille Øyebekken near Nygjeltvold on the quadrangle Røros.

The rocks of this group are also met with on Heim's manuscript maps Folldal and Dovrefjell (p.c. 1966). Heim also regards these rocks as belonging to the Støren Group, an intepretation favoured by Strand (1963) in the adjacent Sel area, where a similar zone of greenstone links up with the greenstones of the southern part of the map sheet Dovrefjell.

The greenstones appears repeatedly possibly due to the folding in this area.

The western limb of this greenstone fan disappears in the northwestern part of the quadrangle. Vågå, but reappears on the map-sheet Dovrefjell (Heim manuscript map 1966) and then continues northwards on to quadrangle Oppdal (Holmsen 1960). The exact borders of the lower greenstone are not properly given on Holmsen's map, but the existence of such rocks is undubitable since rock-types such as pillow lavas, chert, greenstones and greenstones conglomerate appear on the map.

*The Lille Fundsjø conglomerate zone*

The best demonstration of the inversion of the beds at Meråker is provided by the Lille Fundsjø conglomerate discovered by Fediuk (Fediuk and Chaloupsky this volume). Grammeltvedt (p.c. 1966) reports a similar conglomerate from the same stratigraphical position on map-sheet Essandsjø. A polygenous conglomerate is also reported from quadrangle Haltdal by Rui (p.c. 1966) but it's stratigraphical position is not quite clear. It might be equivalent to the Lille Fundsjø conglomerate but it is more likely to represent the position of the Brenna conglomerate stratigraphically above the first one.

Birkeland (p.c. 1966) reports polygenous conglomerates containing boulders of trondhjemite and greenstone at Rensjøen and at Grøtåa near Hesjedalen quadrangle Haltdal. These conglomerates are bordered by the lower greenstone Fundsjø (Støren Group) to the west, and are thus in a comparable position to the Lille Fundsjø conglomerate.

The same holds for the Grimsa conglomerate (Heim, p.c. 1966) on quadrangle Folldal although Heim interprets this conglomerate as underlying the lower greenstone thus representing the Gudå conglomerate zone (Wolff 1964). Heim's interpretation is due to his belief in the assumption of the Trondheim *synclinorium*, (Strand 1951). As to the age of the greenstone conglomerate on quadrangle Sel (Strand 1951), there is no doubt that this conglomerate is overlying the greenstone of this area interpreted as being equivalent to the

Støren Group by Strand (1964). Moreover this conglomerate is overlain by the serpentine conglomerate dated by fossils to the lower part of the Lower Hovin Group, (Yochelson 1963).

From the west side of the anticlinorium a conglomerate of a similar type and position is known from the Grønbakken area, Holmsen (1960). Some of the "various polygenous conglomerates" depicted on Holmsen's map (1960) might also occupy this position.

A serpentine conglomerate is also found near Brekkebekk, quadrangle Folldal (Marlow 1936). The most obvious assumption is to correlate this conglomerate with the serpentine conglomerate near Sel, but as it occupies a position which seems to be the same as the upper greenstone this parallelization is still somewhat doubtful. The extension of the Brekkebekk serpentine conglomerate was reported by Törnebohm (1896) who claimed that it continues northwards for 80 km to Sætersjøen, on quadrangle Røros, as a quartzitic conglomerate. Near Sætersjøen the conglomerate is overlain to the east by the zone of an upper greenstone. The present writer is therefore inclined to correlate this conglomerate with the Brenna conglomerate at Meråker.

*The extension of the Sulåmo-Group*

The rocks of the Sulåmo-Group have been correlated with the rocks of the Lower Hovin Group (Fediuk and Chaloupsky, present volume). Their extension to the north is clear (Wolff 1964), but to the south their presence is still somewhat dubious. Grammeltvedt (manuscript map 1966) has been able to trace the beds of this group to the lakes Lødølja on the quadrangle Essandsjø. South of this lake a limestone is reported by K. M. Hauan (diary 1870) together with a grey to black phyllite, rocks typical of the sequence in question.

On quadrangle Haltdal, Rui (p.c. 1966) holds that the sandstone to the west of the upper greenstone belongs to the Sulåmo Group. Because of the thrust plane mentioned in the next paragraph, there is no mappable connection between these two localities (Pl. IV) which in effect represent the same zone. Birkeland (p.c. 1966) reports phyllites, partly graphitic, and graywackes south of the Gauldalen valley on map-sheets Haltdal and Røros, which also may belong to this same horizon.

Færden (manuscript map Røros 1965) describes these rocks as mica schists, but as they occur between the upper greenstone in the east and the lower greenstone (Fundsjø — Støren-Group) in the west it is most probable that these mica schists are equivalents of the Sulåmo-Group.

On P. Holmsen's map Tynset (1950) this zone is depicted as phyllites;

there can therefore be no arguments against including them in the same group.

On an excellent map of the quadrangle Folldal, Heim (stensil, Diplomarbeit University of Mainz, F.G. 1966) notes the occurrence of a "Graue phyllitische Glimmerschiefer" in this zone. Heim's sketch map Dovrefjell (p.c. 1966) shows this same rock continuing across the map sheet towards the west where it links up with the schists indicated as the Heidal Series on Strand's map Sel and Vågå (1951). Strand's interpretation of the Heidal Series, in this part of the map, as being situated beneath the greenstone (Støren Group) is contrary to the fact that according to his map the schists in the area in question are obviously overlying the greenstone while the border between the two is occupied by a double greenstone-serpentine conglomerate which is proved to be younger than the greenstone. The serpentine conglomerate is dated by fossils to 3c$\beta$—3c$\gamma$ of the Oslo region (Yochelson 1963) a position equivalent to the Venna conglomerate of the Hølonda — Horg area (Vogt 1945). This questionable point is due to the fact that Strand (1951) with the information then available had to assume that, "In the northern part of the Vågå map area a synclinorium widening towards the north forms the southern end of the large synclinorium of the Trondheim Region", while today, working with the present knowledge of the northern Trondheim region, one has to accept the possibility that the main structure of this area is similar to that in the northern parts i.e. anticlinal. This leads to the assumption that the schists in the western part of this area, indicated as the Heidal Series with a question-mark on Strand's map, belong to the Sulåmo (Lower Hovin) Group.

*The upper greenstone*

At several localities in the eastern Trondheim region a greenstone of minor thickness is found to the west of the rocks of the Røros Group (Upper Hovin). This greenstone was found by Foslie (Foslie and Oftedahl 1959) during his mapping of quadrangles Jævsjø and Bjørkvassklumpen and followed across the quadrangle Verdal by the present writer (1960). Stratigraphically it has been placed near the horizon of the Volla conglomerate (Wolff 1964). As mentioned in Fediuk's and Chaloupsky's article (this volume) it has also been traced across the quadrangle Meråker. Grammeltvedt (p.c. 1966) has only been able to follow this greenstone for a short distance southwards on the quadrangle Essandsjø where it thins out. U. Bjørlykke's (1963) map indicates that this greenstone reappears along the river Lødølja. Rui (p.c. 1966) reports a similar greenstone horizon further to the south on the quadrangles Stuesjø and Haltdal, the inversion of which is demonstrated by well-preserved flame structures in a tectonically overlying sandstone to the west. To the east of this greenstone Færden (examination thesis 1949) and Rui (p.c. 1966) report the presence of a thrust plane. This thrust plane is probably sub-parallel to the main one but of a minor order. Its extension both northwards and southwards is not properly mapped at the moment; until this mapping has been completed several questions must remain unanswered. Although several manuscript maps by Færden and others indicate the extension of an easterly greenstone, most likely belonging to this zone it is impossible at present to give an exact picture of it on the map. Færden (p.c.) also reports this green stone from Branumshøgda west of Kongens grube quadrangle Røros and from a drill hole at Lergruebakken east of this mine. This observation is particularly noteworthy as it is a strong indication that the upper greenstone underlies the Røros schists in an inverted anticline in this area. (Færden p.c. profile (fig. 52).). It should be pointed out here that an accurate mapping of this greenstone, in the present writer's opinion, would be of great value to the stratigraphical understanding of this region since it is most likely equivalent to the Volla conglomerate in the Hølonda — Horg arc (Vogt 1945), thus marking the border between the Lower and the Upper Hovin Groups. Carstens 1919 reports a similar greenstone niveau in the western Trondheim region, the Jonsvann greenstone. Heim (p.c. 1966) reports an "untere grünschieferzone" on Bukletten, quadrangle Folldal, which may belong to the upper greenstone since he is not aware of the inversion of the sequence. Farther to the west there are no reports of this greenstone horizon.

*The Røros Schists*

Since Kjerulf (1876) introduced the name "Røros-skifer" this rock series has been regarded as the oldest part of the Trondheim suite. As mentioned previously C. Bugge (1954) was alone in opposing this opinion although T. Strand (1960) pointed out that, "The Røros Group consists of mica schists and can be defined as a stratigraphical unit only by its position below the greenstones of the Støren group and above the underlying sparagmitic schists, which may perhaps be an original basement". Strand also states that, "objections can be raised to the validity of the term Røros Group as the mica schists in the surroundings of Røros, which ought to be the type area of the group, are of undetermined stratigraphic position". It must be kept in mind that the Røros Group as mentioned here, is taken as a group name which includes both the Røros Schists and the Gula Schists.

As the studies in the Meråker area proceeded the present writer became

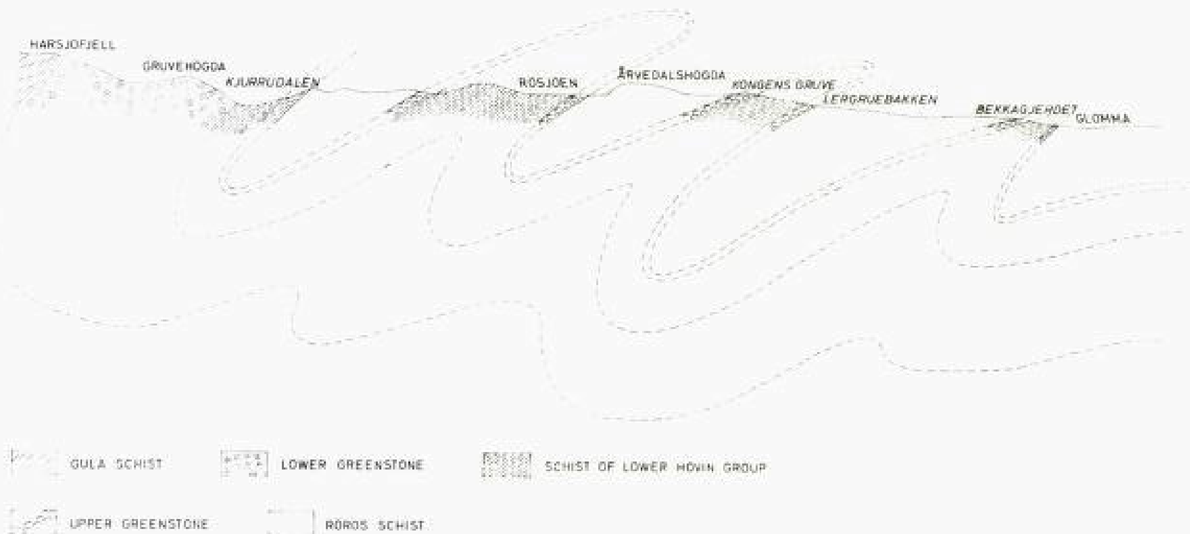


Fig. 52. Section from Harsjøfjell to Glomma (the western part redrawn after Færden (p.c.) showing the present writer's interpretation of the fold structures in this area.

*Profil fra Harsjøfjell til Glomma. (Den vestlige delen er omtegnet etter et manuskriptprofil av Færden) som viser det nærværende arbeids tolkning av foldestrukturen i området.*

more and more suspicious about the character of the Røros Schists as they were described in the literature. It appeared from the more detailed description that the schists mentioned by several authors (Reusch (1890), Carstens (1920), Bryn (1961) and others) as "Stuedalsskifer" and "garbenskifer" were, because of their containing porphyroblasts of biotite and amphibole, being designated as mica schists or garben-schists; this served merely to camouflage their primary sedimentary character. A couple of short visits to Tydal (in 1965) and Røros and Folldal (in 1966) demonstrated that the Røros Schists (Stuedal Schists and garben-schists) are slightly higher metamorphic equivalents of the metagraywackes and slates met with in the Meråker area, as they almost everywhere showed more or less well-preserved sedimentary structures of the Meråker type. The most convincing example was found in a stone quarry be a sideroad to Storvarts east of Røros where gritty graywacke of the Meråker type was observed.

The present writer is therefore convinced that Bugge was correct in suggesting the Røros schists to be equivalent to the Hovin — Hølonda rocks of the western Trondheim region and that a more detailed study of the Røros area in the future will provide additional data in support of this conception.

This view is based on the assumption that the Kjølhaugen Group is equivalent to the Upper Hovis Group, thus placing the Røros Schists also within this latter group.

*Silurian and Devonian sediments*

The Silurian beds encountered in the Meråker area (Siedlecka and Chaloupský presentv olume) proceed southwards and are traced along the strike to midway down the quadrangle Essandsjø. Their extension further south is also probable, but the area is not yet properly mapped.

The present writer is of the opinion that in the easternmost part of the region near the large thrust zone there is a fair chance of finding beds younger than the Røros Schists. R. Falck-Muus (Map 1936) indicates bituminous alum-shale at the two localities Dalvola and Tronsmyren on map-sheet Aursunden. At Dalvola a dark limestone also occurs. These beds might be of Silurian age.

Strata of Devonian age are well known from the locality near Røragen (Goldschmidt 1913). A short visit to this locality convinced the present writer that there is an undoubtable sedimentary succession from the Røros Shists in the west to the Devonian beds in the east, although the contact shows a clear discordance between them. Goldschmidt described the border in this way, "Das Basalkonglomerat wird hier fast ausschliesslich von ausgewitterten

Quarzlinsen des Rørosschiefers zusammengesetzt, es liegt auch direkt auf Rørosschiefer: die unmittelbare Grenze is gut aufgeschlossen. Man erkennt deutlich die Diskordanz, indem der Rørosschiefer nach Nordwesten fällt, das Konglomerat hingegen nach Südosten." This statement is repeated by Holmsen (1962) who found that the Devonian beds were also folded.

*The location of the different serpentinite bodies*

Ultrabasic peridotitic bodies are known from a long series of localities in the Trondheim region. They are usually metamorphosed into serpentinites and occasionally even to soapstone.

Strand (1960) states that, "In all parts of the Scandinavian Caledonides where the stratigraphic relations are known, the peridotites occur in the older part of the stratigraphic sequence only corresponding to the Røros and Støren Groups of the Trondheim region. This seems well enough established to enbale one to take the occurrence of peridotites as a strong indication of an old age of the sediments enclosing them". Interpreting the Røros Schists as equivalent to the Upper Hovin Group, the present writer is forced to oppose this statement. As seen from the map (Pl. IV), the distribution of the serpentinite bodies is closely related to the large thrust plane of the Trondheim nappe or to smaller thrust zones such as the zone of Færden and Rui (p.c.) at Kjøliskarvene, quadrangle Haltdal. As the ultrabasic bodies are intruded into beds of different age, the occurrence of such rocks can hardly be taken as a "strong indication" of an old age for the surrounding beds.

The present writer regards it as more probable that the emplacement of these rocks was associated with the development of the larger trust planes in this region, thus permitting their emplacement into beds of different age. As to the question of the mechanism of the emplacement it might be possible that the fissures of the thrust plane caused pressure to be released at great depts and opened up transport channels up which the relatively viscous ultrabasic magma has been squeezed. This idea for the origin of some of the serpentinite bodies in the Trondheim region is supported by the statement of Turner & Verhoogen (1960) that, "It is not surprising, therefore, to find that major intrusions of peridotite and serpentinite tend to be located along zones of strong dislocation or at least to be bounded by faults of great magnitude."

Another strong support for this idea is the fact that the lower part of the Devonian beds at Røragen contains no boulders of serpentinite while such boulders dominate in the upper part, indicating that the serpentinite masses did not exist in early and middle Devonian time. Consequently the serpentinite bodies were most likely emplaced just before the deposition of the upper Devonian beds.

There is, moreover a distinct difference in the particle size of the sediments in this niveau indicating an abrupt change in sedimentation conditions. This is thought to have been brought about by an uplift of the land block in the east caused by a movement along the thrustplane also at the end of middle Devonian time (Svalbardian fold phase).

*The distribution of the ore deposites*

Mines and smaller occurrences of ore are scattered throughout the Trondheim region and several places have been established as mining districts since the middle of the 17th century. Much data has been gathered from all these deposits, but very little has been done to systematize the available data. This chapter will therefore be devoted to an attempt at correlating the data obtained from the studies of ores with the geology of the regtion.

By plotting the mines on the geological map it is manifest that the various ore occurrences are connected with different rock zones. From Foslie's (1925) list of the South-Norway mines and ore occurrences, it is clear that there is a certain difference in the ore mineral assemblages of the different deposits, a difference which is connected with the surrounding geology. This trend will be seen from the following lists, Tables II, III and IV, compiled both from Foslie's list and NGU archive reports, wherein the prevailing mineral occurs to the left and the secondary mineral to the right. The numbers before the names of the mines refer to the numbers on Foslie's list.

A. Mines situated in the lower greenstone

The trend of this group is very clear as 98 of the 126 occurrences are dominated by pyrite and another 13 by iron-quartzites, hæmatite- and magnetite layers. See Table II. Only 9 are dominated by chalcopyrite and 6 by pyrrhotite. According to Vokes (1962) the two latter can be regarded as one group since there is a marked tendency in the sulphide ore bodies of the Norwegian Caledonides for these to occur together. Thus 111 of the 126 occurrences in this group are dominated by noncuprous sulphides while only 15 are dominated by sulphides containing copper: moreover there are several among the 15 where the genetical connection with the lower greenstone is dubious as bodies of hornblende gabbro may occur nearby. This point will have to be investigated more closely by the mining geologists working in this region.

B. Mines situated in the upper greenstone

The trend in this group is not so marked as that in the lower greenstone as only 12 of the 20 occurrences are pyrite-dominated, 2 being dominated by chalcopyrite and 6 by pyrrhotite. See Table III. This will need a more thorough explanation. On quadrangle Meråker the upper greenstone is often found to be intruded by bodies of hornblende gabbro. This might be the reason why some of the occurrences are deviating from the main trend.

C. Mines situated in the Røros Schists near bodies of the hornblende gabbro

The trend within this group is quite clear as 28 of the 31 occurences are dominated by chalcopyrite, see Table IV.

D. Final remarks

The result of the plotting of the ore occurrences shows a marked tendency for pyrite to be concentrated along zones of greenstone and for chalcopyrite to be concentrated near bodies of hornblende gabbro intruded as sills mainly in the sediments of the Røros Group but also in older sediments and volcanics.

A discussion of the genesis of the ores is beyond the scope of this paper, but as the greenstones are known to contain layers of acid volcanics the theory of exhalative-sedimentary ores of Oftedahl (1958) should be kept in mind when studying the occurrences.

A connection between the hornblende gabbro and chalcopyrite seems to be so close that the present writer is inclined to believe that the genesis is associated with the existence of the gabbro sheets.

**Sammendrag**

I denne artikkelen diskuteres de tektono-stratigrafiske hovedproblemer i det østlige Trondheimsfeltet. Basert på detaljerte studier i Meråker-området de tre siste somrene og på tilgjengelige manuskripskart og personlige meddelelser fra geologer som arbeider i de sydligere deler av feltet, har det vært mulig å bygge opp en generell stratigrafi for det østlige Trondheimsfelt tilsvarende stratigrafien for det vestlige. Det er også foretatt en revisjon av den stratigrafiske nomenklaturen slik at Gulaskifergruppen brukes om de deler av Rørosgruppen som ikke innbefatter Røros-skifrene.

Med utgangspunkt i hva som er påvist i den nordlige del av Trondheimsfeltet pekes det videre på at den eldre oppfatningen av et 'Trondheim synklinorium' må endres til et antiklinorium. Det vises også at det fins et hovedskyveplan langs størsteparten av grensen for Trondheimsfeltet, og betegnelsen

'Trondheimsdekket' foreslåes for den alloktone metasedimentpakke over skyveplanet. Forekomstene av serpentinlegemer langs skyveplanet diskuteres. Til slutt blir fordelingen av svovel- og kopperkisforekomstene i Trondheimsfeltet diskutert med hensyn på geologien.

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Table I. *Correlation table for beds discussed in this paper.*

| | Sworkmo — Lundamo<br>Carstens 1954 · Vogt 1945. | Stjørdal — Meråker<br>(West)<br>Carstens 1960, Wolff 1964 | — <br>(East)<br>— Getz 1890, Wolff 1966 | Ålen<br>I. J. Rui 1966 (p.c.) |
|---|---|---|---|---|
| Horg Group | Shale and sandstone (The Sandå beds)<br>Quartzite conglomerate (Lyngstein) | Absent | Dark shale and sandstone<br>Quartzite conglomerate (Kjølhaugene) | Absent |
| Upper Hovin and Røros Group | Rhyolite (Grimås)<br>Sandstone<br>Polygenous conglomerate (Volla) | Polygenous conglomerate<br>Rhyolite tuff<br>Polygenous conglomerate and sandstone alternating<br>Polygenous conglomerate (Hopla) | Graywackes<br>Absent | Graywackes — often with biotite or amphibole<br>Phyllites<br>- - - - Thrust plane - - - - |
| Lower Hovin Group | Dicranograptus black shale<br>Rhyolite tuff (Esphaug, Hareklett)<br>Sandstone and shale (Krokstad)<br>Limestone (Hølonda)<br>Fossiliferous shale (Langeland)<br>Greenstone conglomerate (Venna) | Dark shale<br>Rhyolite tuff (Muruvik)<br>Sandstone<br>Limestone (Tautra, Forbordfj., Flora)<br>Fossiliferous shale (Leksdal)<br>Greenstone breccia (Stokvola) | Amphibolite (Turifoss)<br>Sandstone and conglomerate<br>Limestone (Brenna)<br>Shale<br>Polygenous conglomerate (Lille Fundsjø) | Dark biotite schist<br>Amphibolite (Kjøliskarvene)<br>Limestone<br>Polygenous conglomerate<br>Sandstone<br>Dark phyllite<br>Conglomerate (Rensjøen) |
| Støren Group | Greenstones | Greenstones | Greenstones | Greenstones, with some sediments |
| Gula Schist Group | Mica schist<br>Quartzite conglomerate (Svorkmo)<br>Crystalline limestone<br>Mica schist | Mica schist<br>Absent<br>Absent<br>Mica schist | Mica schist<br>Quartzite conglomerate (Hegsjøfjell, Skjækerstøtene Gudå and Bukkhammer — Usmadam)<br>Crystalline limestone<br>Mica schist | Black-shale (Dictyonema 2eα)<br>Quartzite conglomerate (Øvstubekken)<br>Crystalline limestone (Vollfjell)<br>Mica schist |

Table II. *Mines situated in the Lower greenstone.*
Py = pyrite. Cpy = chalcopyrite. Po = pyrrhotite. Sph = sphalerite.

*Quadrangle Dovrefjell, western part.*

| | | |
|---|---|---|
| 150. | Elgsjøbækk occurence | Py |
| 152. | Vårstigfjeld (with Skålebekk occurrence) | Py |
| 152a. | Drivdalen occurence | Py |
| 148. | Fundberget occurrence | Py |
| 149. | Elgsjøtangen occurrence | Py |
| | Tverrfjellet | Py |

*Quadrangle Vågå.*

| | | |
|---|---|---|
| 80. | Vasenden mine | Py |

*Quadrangle Dovrefjell, southeastern part.*

| | | |
|---|---|---|
| 180. | Værkensdalen field | Py |
| | Gåshovda occurrence | Py |
| | Tværliseter occurrence | Py |
| | Grimsa occurrence | Py |

*Quadrangle Folldal.*

| | | |
|---|---|---|
| | Einundalens occurrence | Py |
| 175. | Grimsdalen mine | Py Cpy |
| 176. | Grev Moltke mine | Py Cpy |
| 177. | Folldal main mine | Py Cpy |
| 179. | Juliane Marie & Godthåb mine | Py Cpy |

*Quadrangle Tynset.*

| | | |
|---|---|---|
| 201. | Nebyvoll occurrence (St. Olaf's mine) | Py Cpy |
| 202. | Hvaltjernåsen mine | Py |
| 203. | Nonsvola occurrence | Py |

*Quadrangle Røros.*

| | | |
|---|---|---|
| 204. | Kvitstein mine | Py |
| 205. | Vingelen mine | Py Cpy |
| | Vingelsvola occurrence | Py |
| 208. | Åseng mine | Py |
| 209. | Vandgrøften occurrense | Cpy Po Py |
| 210. | Fredrik IV mine | Cpy Py |
| 233. | Harsjø mine | Py Cpy |

*Quadrangle Haltdal.*

| | | |
|---|---|---|
| 249. | Rognså occurrence | Py Cpy |
| 247. | Storvold mine | Cpy Po Py |
| 248. | Hesjedalen mine | Py Cpy |
| 245. | From mine | Py |
| 246. | Rogn mine | Py Cpy |
| 251. | Lillerena occurrence | Po Cpy |
| 269. | Rødhammer mine | Py |
| | Grønfj. occurrence | Py |
| 270. | Hultrå mine | Py |
| 268. | Skjellåfjell mine | Po Cpy |
| 252. | Kårslått mine | Cpy Po |

*Quadrangle Essandsjø.*

| | | |
|---|---|---|
| 349. | Torsbjørk mine | Py |

*Quadrangle Meråker.*

| | | |
|---|---|---|
| 350. | Mandfjell mine | Py Cpy |
| 351. | Gruvbekk and Bakbekk occurrence | Py Cpy |
| 352. | Fondfj. mine and Løvlibekk occurrence | Py Cpy |
| 353. | Finskar occurrence | Py Cpy |
| 355. | Lillesætervold occurrence | Cpy |
| 356. | Krogstad occurrence | Po Cpy Sph |

*Quadrangle Verdal.*

| | | |
|---|---|---|
| 372. | Årstad occurrence | Py Po |
| 373. | Storstad occurrence | Py Po |
| 375. | Åkervold mine | Py Cpy Po |
| 376. | Malså mine | Py Cpy |
| 377. | Vetringshallen mine | Py Cpy |
| 378. | Gulstad & Mok mines | Py Cpy |

*Quadrangle Selbu, south-east.*

| | | |
|---|---|---|
| 266. | Rokne occurrence | Py |
| 265. | Selbu mines | Py Cpy Po |

*Quadrangle Selbu, north-west.*

| | | |
|---|---|---|
| 298. | Renå occurrence | Py |
| 299. | Dragsten mine | Py |
| 301. | (Nonsh. & Venen mines) | Fe |
| 302. | Viken (Løvådal) mine | Py |
| 303. | Sandoret occurrence | Py Cpy |
| 304. | Grøttemsvold field with Kirkelid mine | Py |
| 305. | Engvold occurrence | Py |
| 306. | Fuglemvold field with Langjon mine | Py |
| 307. | Ingridvold occurrence | Py |
| 319. | Vottafjell occurrence | Py |
| 320. | Damtjern occurrence | Py |
| 321. | Røsbæk occurrence | Py Fe |
| 322. | Sesåsvold occurrence | Py |

*Trondheim.*

| | | |
|---|---|---|
| 282. | Leinum (or Mo) occurrence | Py |
| 283. | Leinstrand mine | Fe-quartzite |
| 293. | Bratsberg occurrence | Py |
| 292. | Lien occurrence | Py |
| 290. | Vikåsen occurrence | Py |
| 291. | Stene occurrence | Fe |
| 288. | Kobberdammen occurrence | Py |
| 284. | Klefstadåsen occurrence | Py Cpy |
| 285. | Flåkahaugen occurrence | Py |

| | | |
|---|---|---|
| 286. | Svartdalsbæk (or Klemetsaunet) occurrence | Py |
| 287. | Holstvolden — Bratløfta occurrence | Py |
| 289. | Fagerli (or Ilsviken) occurrence | Py |

*Melhus.*

| | | |
|---|---|---|
| 97. | Åmot mine | Cpy Py |
| 98. | Stor-Næve occurrence | Py Cpy |
| 275. | Leberg occurrence | Py |
| 276. | Kvål og Skjerdingstad mine | Py Po |
| 277. | Flå (or Vasfjeld) mine | Py |
| 278. | Lerli and Løvset occurrence | Py |
| 279. | Havdøl occurrence | Py |
| 294. | Bratstigen occurrence | Fe - quartzite |
| 295. | Viken occurrence | Py |
| 296. | Lervik occurrence | Py |
| 297. | Tangvoldodden occurrence | Py |

*Rennebu.*

| | | |
|---|---|---|
| 105. | Gorset occurrence | Py |
| 106. | Jordfjeldets mine | Py |
| 107. | Lillevandåsen | Po |
| 111. | Mærk occurrence | Py |

*Rindal.*

| | | |
|---|---|---|
| 81. | Solås - Midtgård occurrence | Py |
| 82. | Nergårdsmo occurrence | Py |
| 83. | Trøkna mine | Py |
| 84. | Lommunda mine | Py |
| 95. | Dragset mine | Py Cpy |
| 103. | Holum occurrence | Py |
| 96. | Skjøtskift—Jordhus occurrence | Py |
| 100. | Løkken mine | Py Cpy |
| 99. | Høidal mine | Py Cpy |
| 101. | Grefstad mines | Py Cpy |

*Stjørdal.*

| | | |
|---|---|---|
| 309. | Brandåsen occurrence | Fe & Py |
| 310. | Flensberg mine | Fe |
| 311. | Næverå mine | Fe |
| 312. | Hinberg occurrence | Fe |
| 313. | Kleptjern occurrence | Fe |
| 314. | Bjørn mine | Fe |
| 315. | Næver mine | Py |
| 316. | Vikvold mines | Py |
| 317. | Klep mine | Fe |
| 318. | Grønli mine | Fe |
| 323. | Rangåvold & Vindsmyren occurrence | Py Po |
| 348. | Sonvandets mine | Cpy Po |
| 358. | Renåbolet occurrence | Po Cpy |

*Levanger.*

| | | |
|---|---|---|
| 364. | Tingstad mine | Cpy Py |
| 366. | Rokne occurrence | Py |
| 367. | Kolberg occurrence | Cpy Py |
| 389. | Nordvik occurrence | Py |
| 388. | Jørstad mine | Py |
| 386. | Falstad, Eines, Stangerholt mines | Py |
| 387. | Ytterøens mines | Py Cpy |
| 390. | Sundsetnes occurrence | Po |
| 380. | Hø occurrence | Py Po |

*Trollhetta.*

| | | |
|---|---|---|
| 104. | Reisfjeld occurrense | Py |

## Table III. *Mines situated in the upper greenstone.*

*Quadrangle Tynset.*

| | | |
|---|---|---|
| 200. | Nyberg occurrence | Py |
| | Storbekken occurrence | Py |

*Quadrangle Røros.*

| | | |
|---|---|---|
| 213. | Oscar II mine | Py Cpy |
| 222. | Salå mine | Po Cpy |
| 223. | Lomtjøn mine | Py |
| 230. | Kongens and Årvdals mine | Py Cpy |
| 231. | Sekstus mine (Christian VI) | Po Cpy |
| 232. | Muggruben | Py Cpy |

*Quadrangle Haltdal.*

| | | |
|---|---|---|
| 243. | Skar (Skårdals) mine | Py |
| 242. | Killingdals mine | Py Cpy |
| 255. | Svensk - Menna mine | Py Cpy |
| 256. | Røros - Menna mine | Po Py Cpy |
| 257. | Guldals mine | Po Cpy |
| 254. | Folldals occurrence | Po Cpy |

*Quadrangle Stuesjø.*

| | | |
|---|---|---|
| 258. | Kjøli mine | Py Cpy |
| 260. | Midt (Jens) mine | Py Cpy |

*Quadrangle Meråker.*

| | | |
|---|---|---|
| 340. | Sag and Røsås occurrence | Py Cpy |
| 342. | Dalemo occurrence | Cpy |
| 344. | Stadsås mine | Po Cpy |
| 345. | Vægterhaug mine and Angeli occurrence | Cpy Po |

## Table IV. *Mines situated in the Røros Schists near bodies of hornblende gabbro.*

*Quadrangle Røros.*

| | | |
|---|---|---|
| 214. | Kvernskal mine | Cpy |
| 255. | Isak occurrence | Cpy |
| 226. | Storvarts mine | Cpy Po |
| 227. | Hestkletten, Quintus, Nyberg, Solskinn mine | Cpy Po |
| 228. | Klasberget mine | Cpy Po |
| 229. | Sletmo occurrence | Cpy Po |
| 235. | Skarv (Ole Iversa) mine | Cpy Po |

*Quadrangle Haltdal.*

| | | |
|---|---|---|
| 241. | Sørosen occurrence | Cpy |

*Quadrangle Aursunden.*

| | | |
|---|---|---|
| 218. | Lossius and Sara mines | Cpy Po |
| 238. | Klinkenberg occurrense | Cpy Po? Py |

*Quadrangle Stuesjø.*

| | | |
|---|---|---|
| 239. | Sødals mine | Cpy |
| 237. | Mads (Mathis) mine | Cpy Py |
| 236. | Fjeldgjelt occurrence | Cpy Po |
| 240. | Bønskneppen occurrence | Cpy |
| 259. | Lillegula occurrence | Cpy |

*Quadrangle Essandsjø.*

| | | |
|---|---|---|
| 263. | Esna mine | Cpy |
| 264. | Vorrevik occurrence | Cpy Py |
| 331. | Ramfjell mine | Cpy Po |
| 332. | Gilså mine | Cpy Po |
| 333. | Bjørneggen occurrence | Cpy |
| 334. | Dronningen mine | Cpy Po |
| 336. | Lillefjell mine | Py Cpy |
| 337. | Storhusmannsberget (Dudu) mine | Cpy Py Po |
| 338. | Væråsvold occurrence | Cpy Po |

*Quadrangle Meråker.*

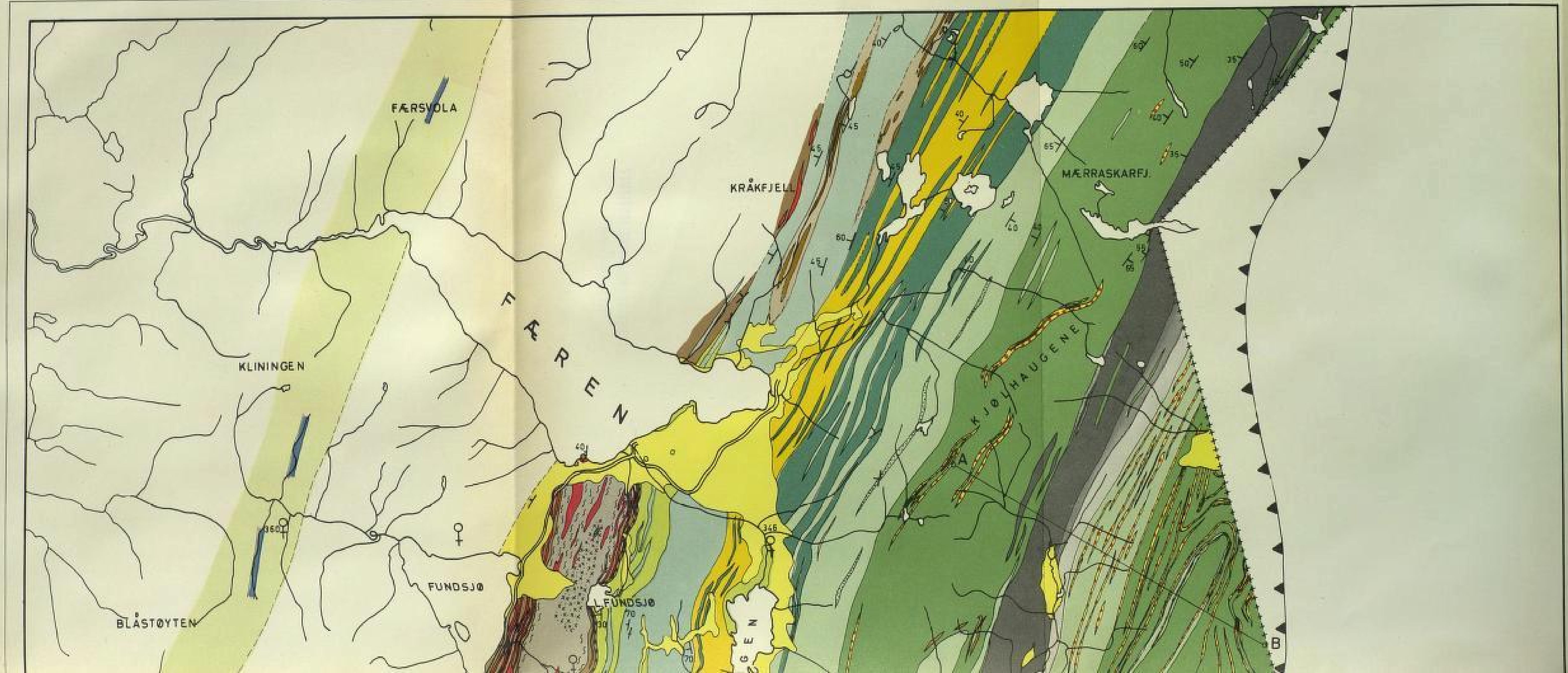
| | | |
|---|---|---|
| 339. | Langsund | Cpy Po |
| | Davola occurrence | Cpy |
| | Dalvolavollen occurence | Cpy |
| | Brenthaug occurrence | Cpy Py |
| | Navelhaug occurrence | Cpy |
| | Langen occurrence | Cpy |
| | Hammerskallen occurrence | Cpy |


KJØLHAUGENE

# GEOLOGICAL MAP OF THE MERÅKER AREA

## GEOLOGISK KART OVER MERÅKER

Scale 1:100 000




**LEGEND**

Quaternary.

**Slågån Group (Silurian).**

Grey to grey-black phyllite, slate and metasiltstone.

Grey slates with intercalations of metasandstone.

Grey metasandstone with intercalations of slate.

**Kjølhaugen Group (Upper Ordovician).**

Grey-green slates and phyllites with intercalations of metagraywacke.

The Kjølhaugene quartzite conglomerate.

Grey-green metagraywackes with intercalations of slate (dotted: thicker beds of subgraywacke).

Grey phyllite.

**Sulåmo Group (Middle Ordovician).**

Metabasite with banded structure.

Metabasite of massive structure.

Grey phyllite.

Grey calcareous metasandstone.

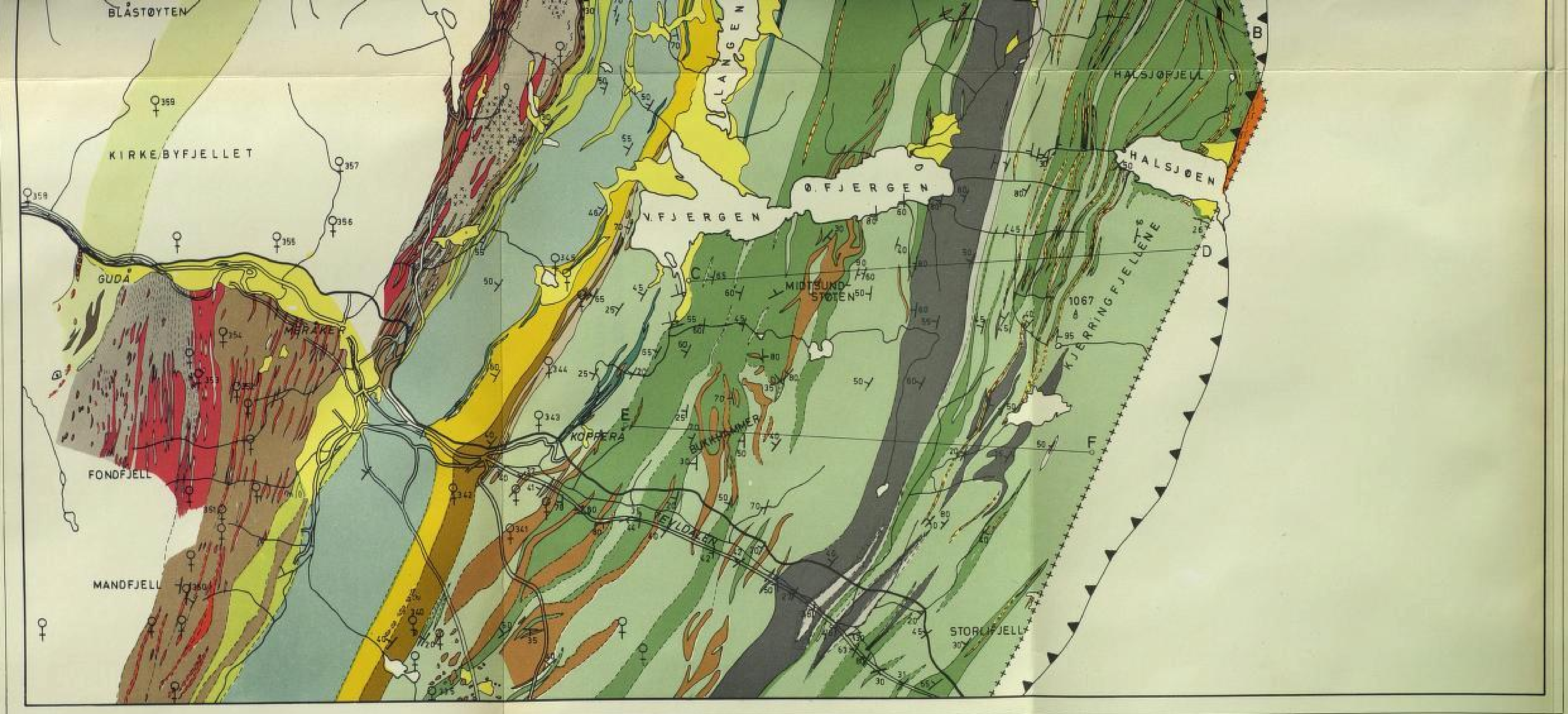
The Brenna conglomerate.

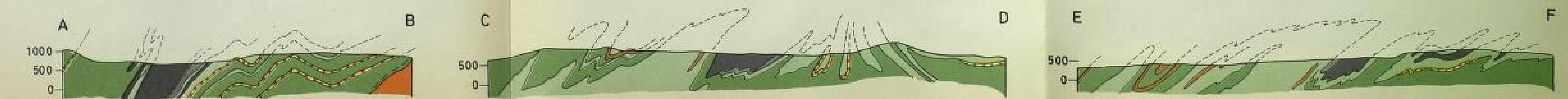
The Brenna limestone.

Grey and black phyllite.

Grey phyllites and graywackes.

The Lille Fundsjø conglomerate.




Grey phyllites and graywackes.

The Lille Fundsjø conglomerate.

**Fundsjø Group (Lower Ordovician).**
Metabasites.

Quartz-keratophyre.

**Sonvatn Group (Cambrian)**
Mica schists, often with garnet.

Alternating amphibolites and schists.

The Gudå quartzite conglomerate.

Limestone.

**?Eocambrian.**
Schists and gneisses.

**Caledonian intrusives.**
Granitic rocks.

Fine- to medium-grained gabbro.

Fine- to medium-grained gabbro, without preferred orientation.

Fine- to medium-grained gabbro, strongly schistose.

Hornblende gabbro.

**Structures.**
Strike and dip.

Lines of section.

Foliation, lineation.

Mylonite zone.

Thrust plane.

Occurrences of pyritic and chalcopyritic ore.

# GEOLOGICAL MAP OF THE TRONDHEIM REGION

GEOLOGISK KART OVER TRONDHEIMSFELTET

1:500000

COMPILED BY FR.CHR.WOLFF AFTER:
SAMMENTEGNET AV FR.CHR.WOLFF ETTER:

T.BIRKELAND, C.W.CARSTENS, H.CARSTENS, J.CHALOUPSKY, G.GRAMMELTVEDT, F.FEDIUK, M.FIŠERA, S.FOSLIE, J.FÆRDEN, A.HAUGEN, H.HEIM, P.HOLMSEN, H.J.KISCH, GHR.OFTEDAHL, J.PEACEY, Z.PELC, D.ROBERTS, I.J.RUI, G.SCHAAR, A.SIEDLECKA, S.SIEDLECKI, T.STRAND, TH.VOGT, FR.CHR.WOLFF.

LEGEND
*TEGNFORKLARING*


RÖRAGEN BEDS (DEVONIAN)
*RÖRAGENFELTET (DEVON)*

CONGLOMERATE AND SHALE
*KONGLOMERAT OG SKIFER*

SLÅGÅN GROUP – HORG GROUP (SILURIAN)
*SLÅGÅNGRUPPEN – HORGGRUPPEN (SILUR)*

DARK SHALE AND SANDSTONE
*MÖRK SKIFER OG SANDSTEIN*

KJÖLHAUGEN GROUP – RÖROS GROUP – UPPER HOVIN GROUP (UPPER ORDOVICIAN)
*KJÖLHAUGGRUPPEN RÖROSGRUPPEN ÖVRE HOVINGRUPPEN (ÖVRE ORDOVICIUM)*

PHYLLITE, METAGRAYWACKES, WITH INCREASING AMOUNTS OF BIOTITE, HORNBLEND AND GARNET TOWARDS THE SOUTHEAST, PARTLY CONGLOMERATIC
*FYLLITT, METAGRÅVAKKER MED ÖKENDE MENGDER AV BIOTITT, HORNBLENDE OG GRANAT MOT SYDÖST, DELVIS KONGLOMERATISK*

POLYGENOUS CONGLOMERATE
*POLYMIKT KONGLOMERAT*

SULÅMO GROUP – LOWER HOVIN GROUP (MIDDLE ORDOVICIAN)
*SULÅMOGRUPPEN UNDRE HOVINGRUPPEN (MIDTRE ORDOVICIUM)*

DARK SHALE AND RHYOLITE TUFF IN WEST, GREENSTONE IN EAST
*MÖRK SKIFER OG RHYOLITT TUFF I VEST, GRÖNNSTEN I ÖST*

GREY CALCAROUS SANDSTONE AND GREY TO DARK PHYLLITE
*GRÅ KALKHOLDIG SANDSTEIN OG GRÅ TIL MÖRK FYLLITT*

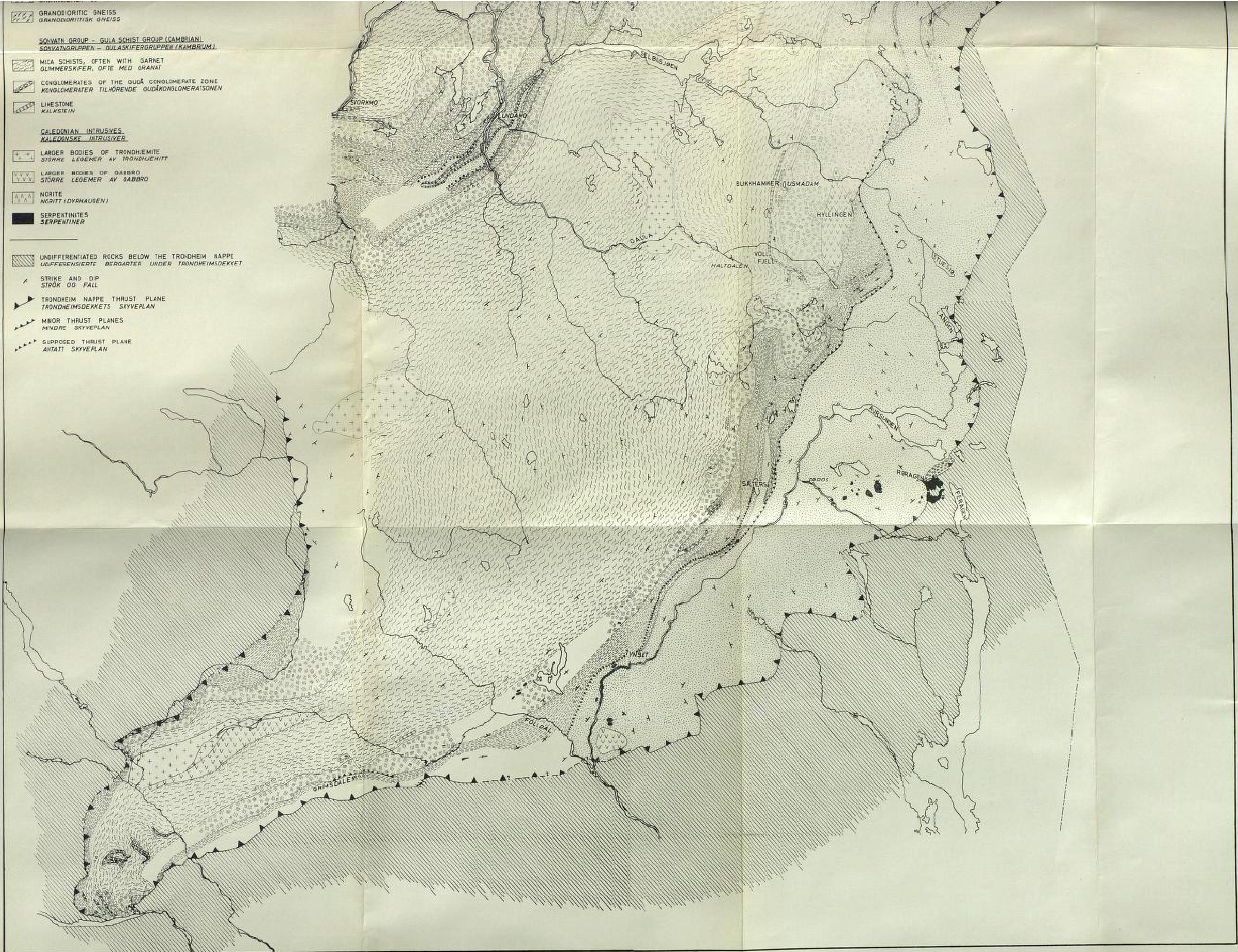
HÖLONDA, TROMSDALEN, BRENNA AND SIMILAR LIMESTONES
*HÖLONDA, TROMSDALEN, BRENNA OG LIGNENDE KALKSTEINER*

VENNA, STOKKVOLA, LILLE FUNDSJÖ AND SIMILAR CONGLOMERATES
*VENNA, STOKKVOLA, LILLE FUNDSJÖ OG LIGNENDE KONGLOMERATER*

FUNDSJÖ GROUP – STÖREN GROUP (LOWER ORDOVICIAN)
*FUNDSJÖGRUPPEN – STÖRENGRUPPEN (UNDRE ORDOVICIUM)*

GREENSTONES AND QUARTZKERATOPHYRES
*GRÖNNSTENER OG KVARTSKERATOFYRER*

GRANODIORITIC GNEISS
*GRANODIORITTISK GNEISS*


GRANODIORITIC GNEISS
GRANODIORITTISK GNEISS
SONVATN GROUP – GULA SCHIST GROUP (CAMBRIAN)
SONVATNGRUPPEN – GULASKIFERGRUPPEN (KAMBRIUM)
MICA SCHISTS, OFTEN WITH GARNET
GLIMMERSKIFER, OFTE MED GRANAT
CONGLOMERATES OF THE GUDÅ CONGLOMERATE ZONE
KONGLOMERATER TILHÖRENDE GUDÅKONGLOMERATSONEN
LIMESTONE
KALKSTEIN
CALEDONIAN INTRUSIVES
KALEDONSKE INTRUSIVER
LARGER BODIES OF TRONDHJEMITE
STÖRRE LEGEMER AV TRONDHJEMITT
LARGER BODIES OF GABBRO
STÖRRE LEGEMER AV GABBRO
NORITE
NORITT (DYRHAUGEN)
SERPENTINITES
SERPENTINER
UNDIFFERENTIATED ROCKS BELOW THE TRONDHEIM NAPPE
UDIFFERENSIERTE BERGARTER UNDER TRONDHEIMSDEKKET
STRIKE AND DIP
STRÖK OG FALL
TRONDHEIM NAPPE THRUST PLANE
TRONDHEIMSDEKKETS SKYVEPLAN
MINOR THRUST PLANES
MINDRE SKYVEPLAN
SUPPOSED THRUST PLANE
ANTATT SKYVEPLAN
SELBUSJØEN
SVORKMO
LUNDAMO
BUKKHAMMER
USMADAM
HYLLINGEN
GAULA
HALTDALEN
VOLL FJELL
STUESJØ
AUGUNDEN
RØRAGEN
RØROS
SÅTERSJ.
FERAGEN
VIGSET
FOLLDAL
GRIMSDALEN