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

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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-

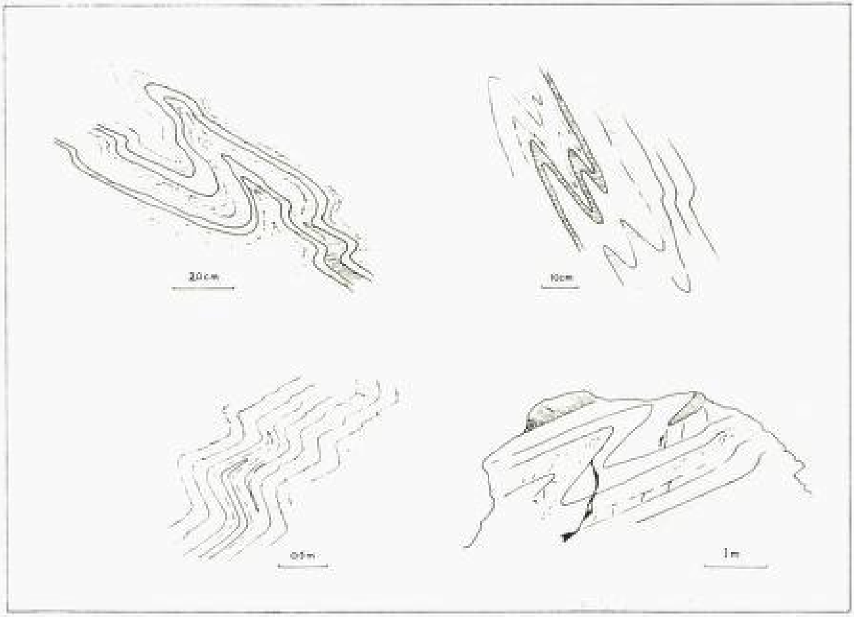


Fig. 33. Early folds deformed by late structures. Interbanded phyllite/metagraywacke, Upper Hovin Group. Teveldalen.

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

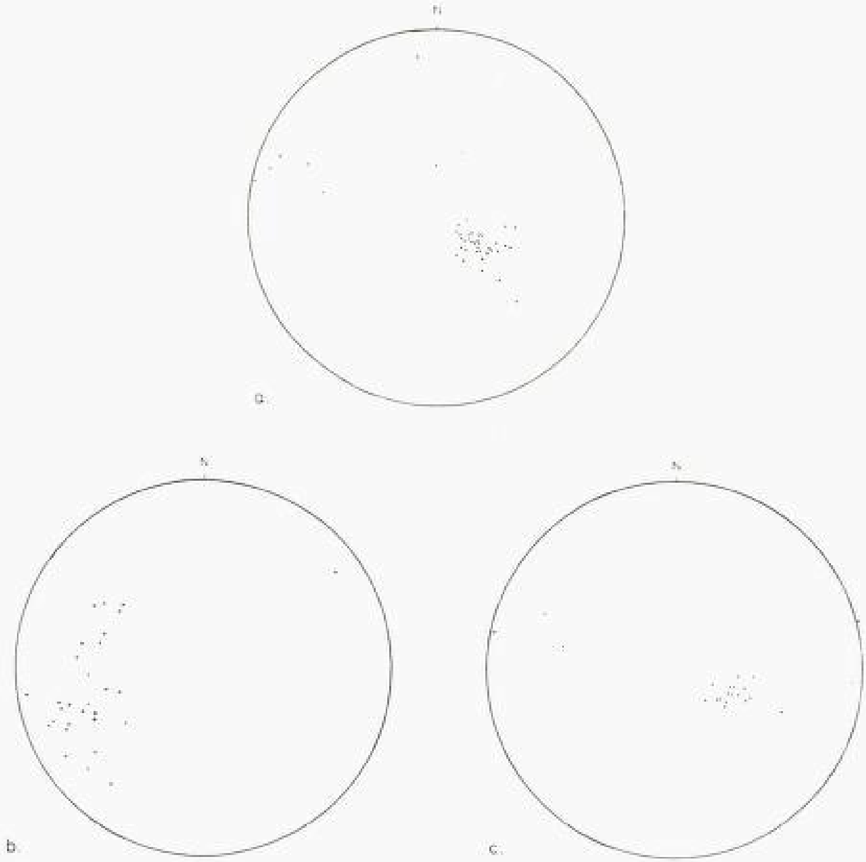


Fig. 35. Stereographic projections of structural data (Wulff net, lower hemisphere) (a) Poles to bedding planes. (b) Early fold axes and related lineations. (c) Poles to axial planes of early folds.

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.

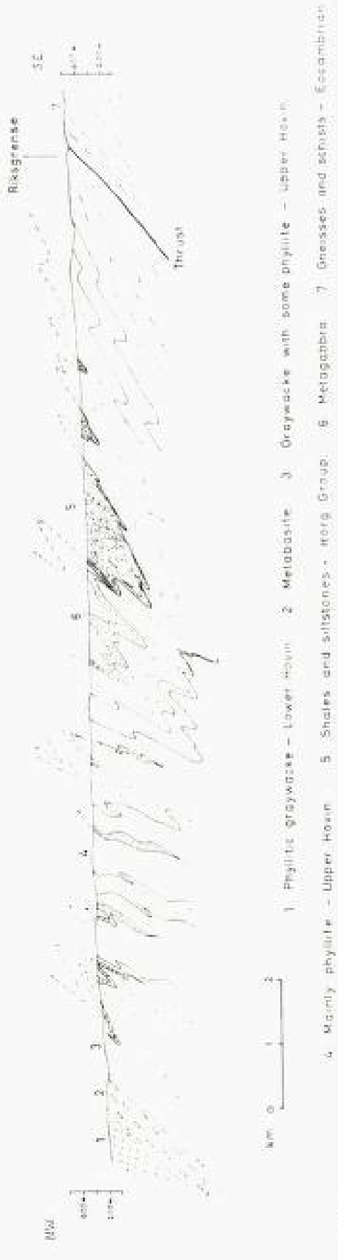
A similar lithological control in the development of quartz-rods was described by Wilson (1953) from the North-West Highlands of Scotland. In this Scottish example the quartz-rods are often abundant in incompetent, semi-pelitic strata and generally lacking in siliceous granulite horizons.

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-Veravatn 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

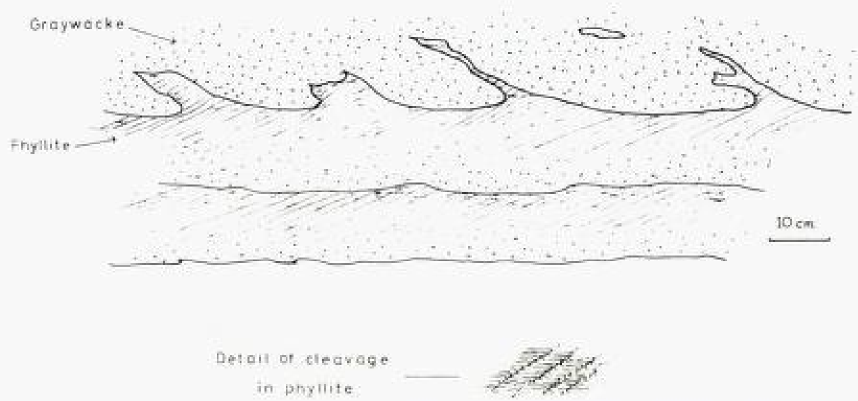


Fig. 38. Load casts, flame structures and late cleavage. Upper Hovin Group. Teveldal road.
Pålastningsavstøpninger, flammestrukturer og sen kløv. Øvre Hovingruppe. Mellomriksveien.

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.

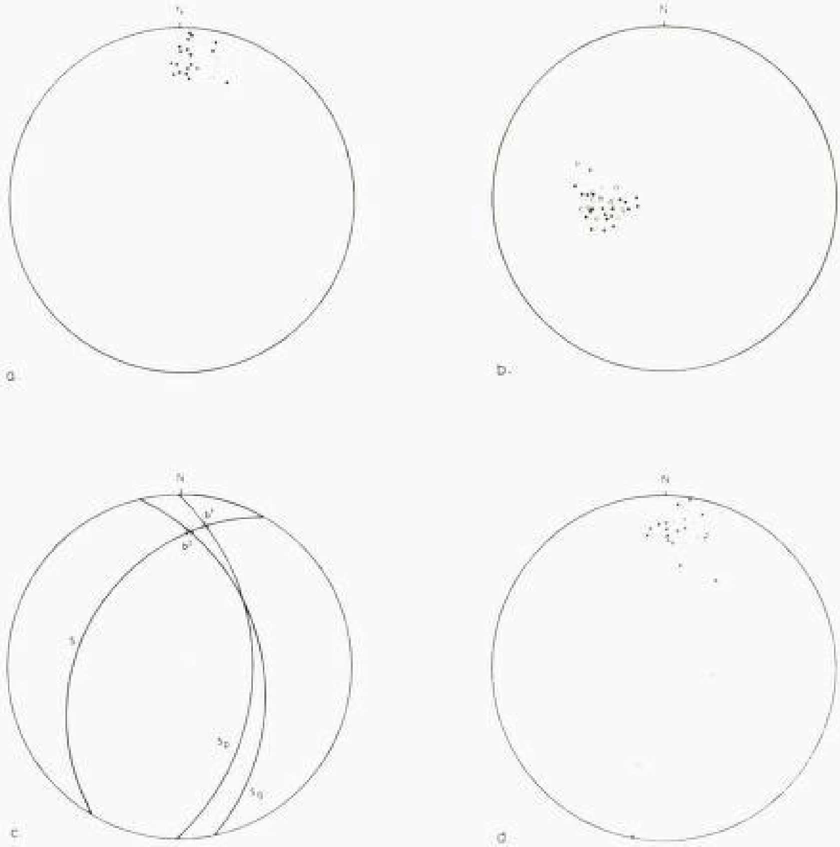


Fig. 39. Stereographic projections of structural data (Wulff net, lower hemisphere). (a) Late fold axes. Circles are phyllite, dots metagraywacke. (b) Poles to axial planes of late folds. Circles phyllite, dots metagraywacke. (c) S-planes in phyllite/metagraywacke. S — bedding plane, Sp — axial plane 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 β-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 β-diagram technique it was found that β-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-

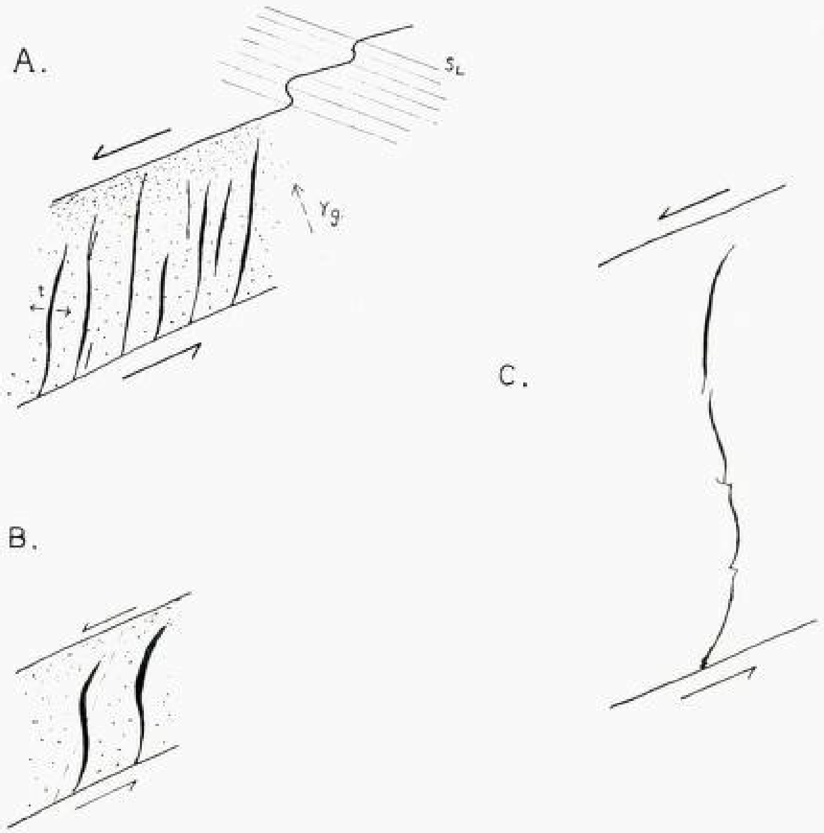


Fig. 41. Tension gashes in psammite. (A) Diagrammatic representation showing probable stress distribution. Yg — younging direction. S_L — late cleavage. (B) Sigmoidal fractures. (C) Division of curved tension fractures into smaller sigmoidal gashes.

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 $2e_{\alpha}$ zone of the Oslo region) of the Ordovician.

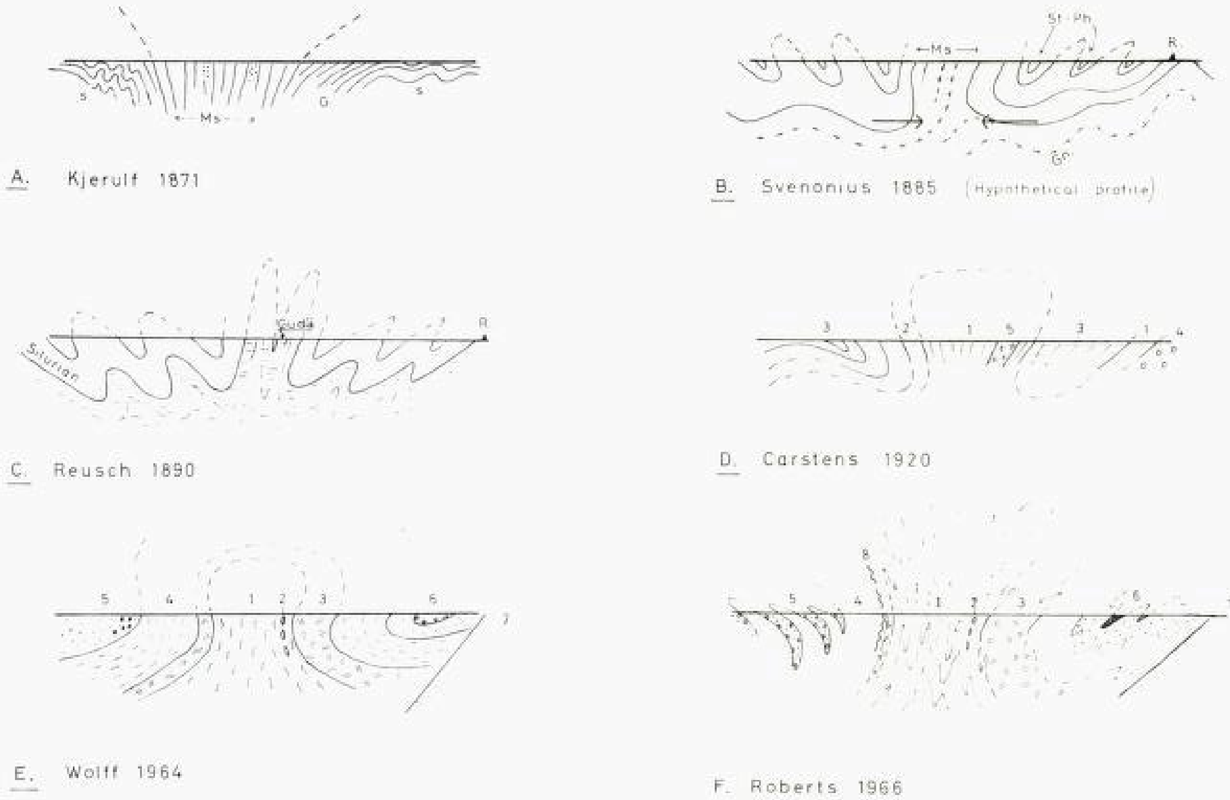
A recent find of a graptolitic shale from the Lower Hovin Group near Løkken, S.W. of Trondheim, has been described by Blake (1962) and was considered to be of Middle Arenigian (extensus zone, pars. of Britain, $3\,b\gamma$ of the Oslo region) but Skevington (1963) has queried this and suggests a slightly 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-

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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

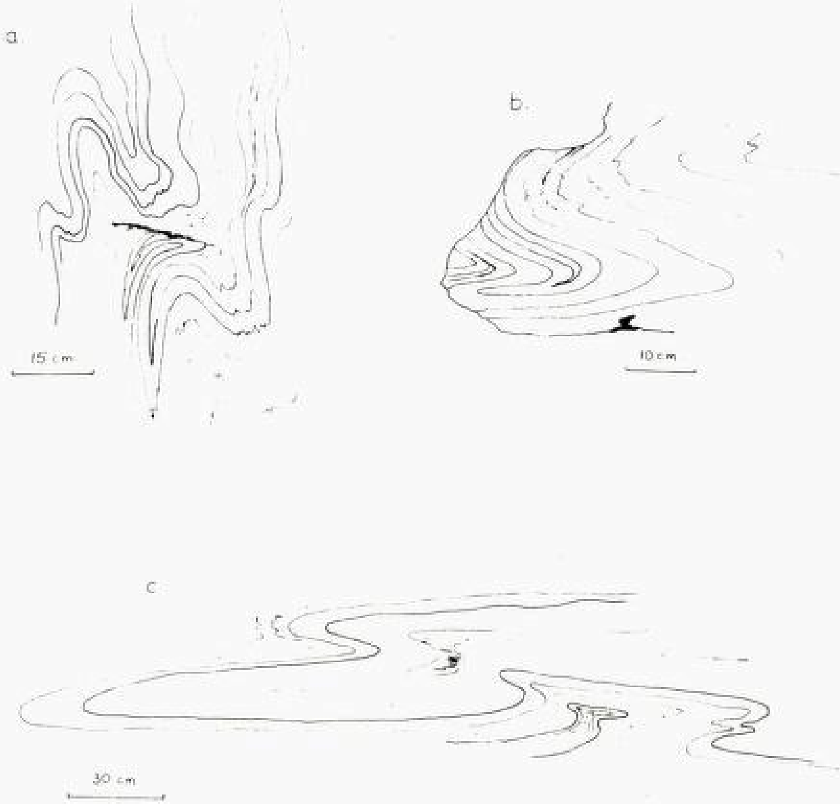

Fig. 44.. Refolding of first-phase isoclines by second generation structures. Interbanded schist-quartzite lithology. Gula Schist Group; just west of Gudå.

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 subparallelism 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-

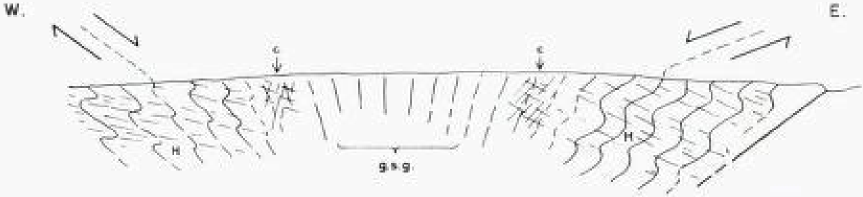

Fig. 46. Schematic profile along Stjørdalen to show sense of overturning of third-phase folds with respect to the attitude of banding. H — Hovin Group: g.s.g. — Gula Schist Group, c — zones with conjugate folds.

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

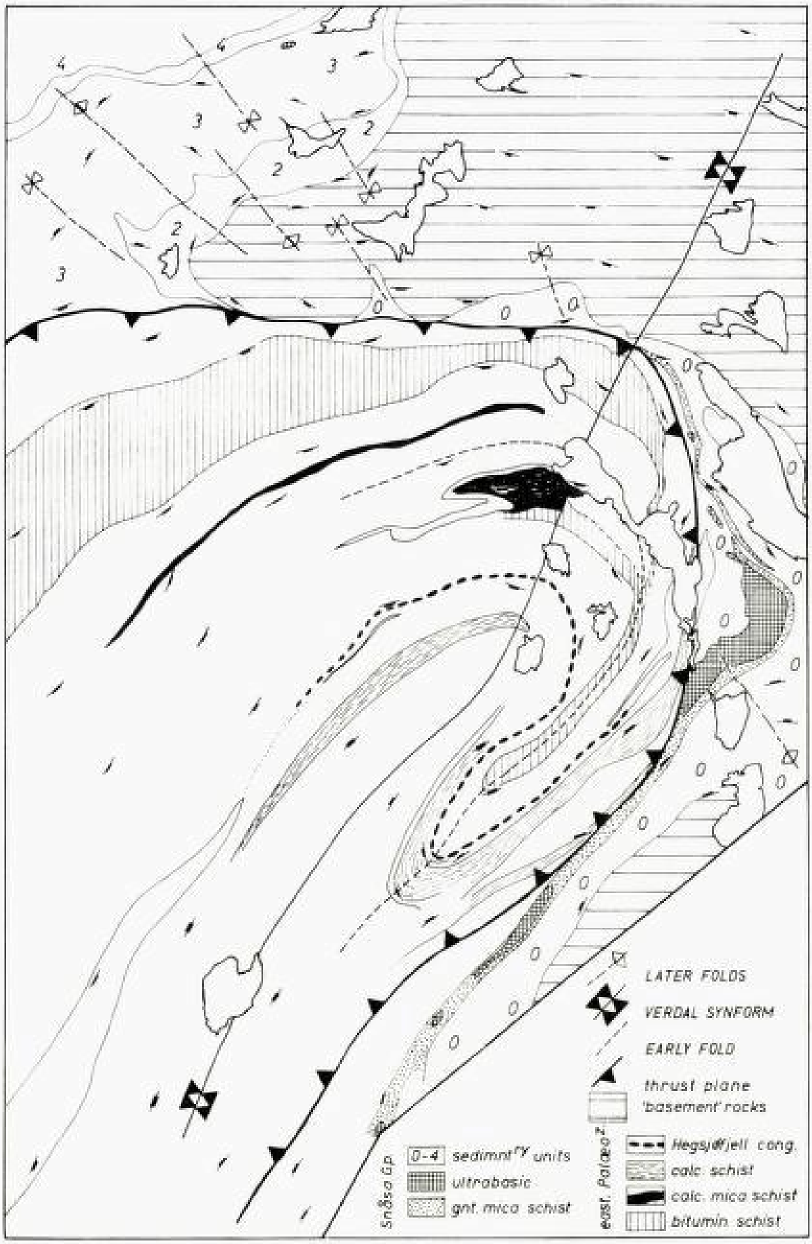

Fig. 47. Simplified stratigraphy of the Verdal synform to show the early fold. Figure from Peacey (1964).

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

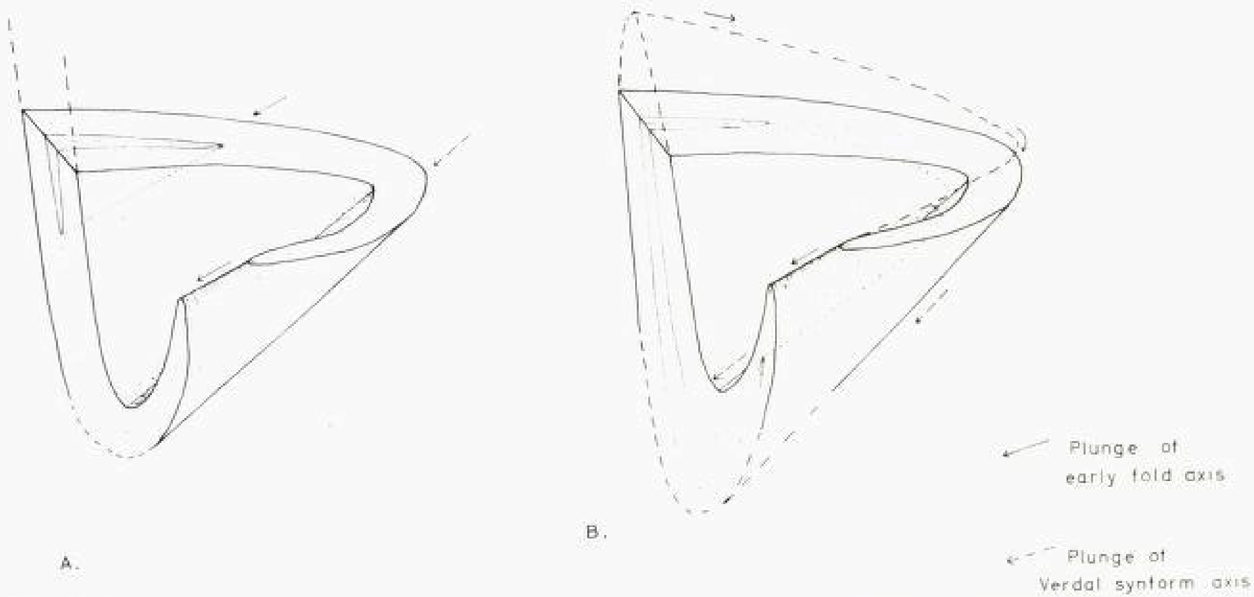

Fig. 48. Diagrammatic representation of the Hegsjøfjell — Imsdalsfjell early fold. (A) Peacey's interpretation drawn from the accompanying sections (1964). (B) The present interpretation. For explanation see text.

*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

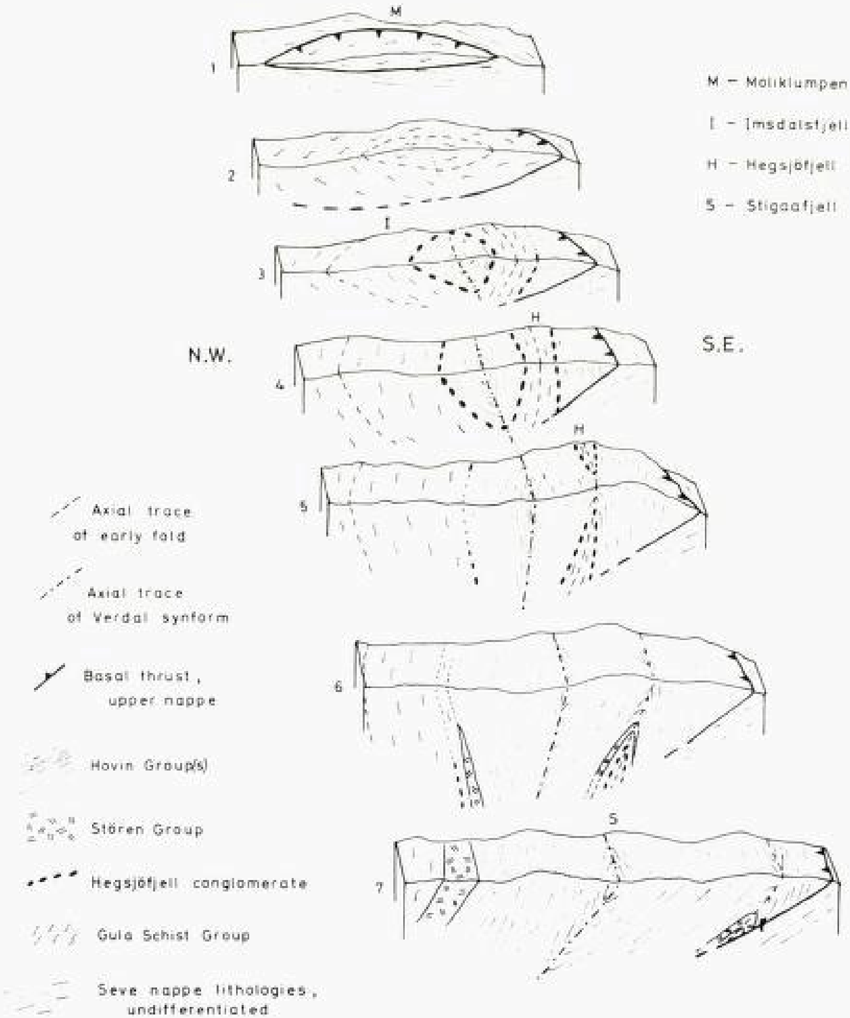

Fig. 49. Sections across the Verdal synform to show (a) its changing development and (b) the early fold. Distance between sections 1 and 7 about 22 km.

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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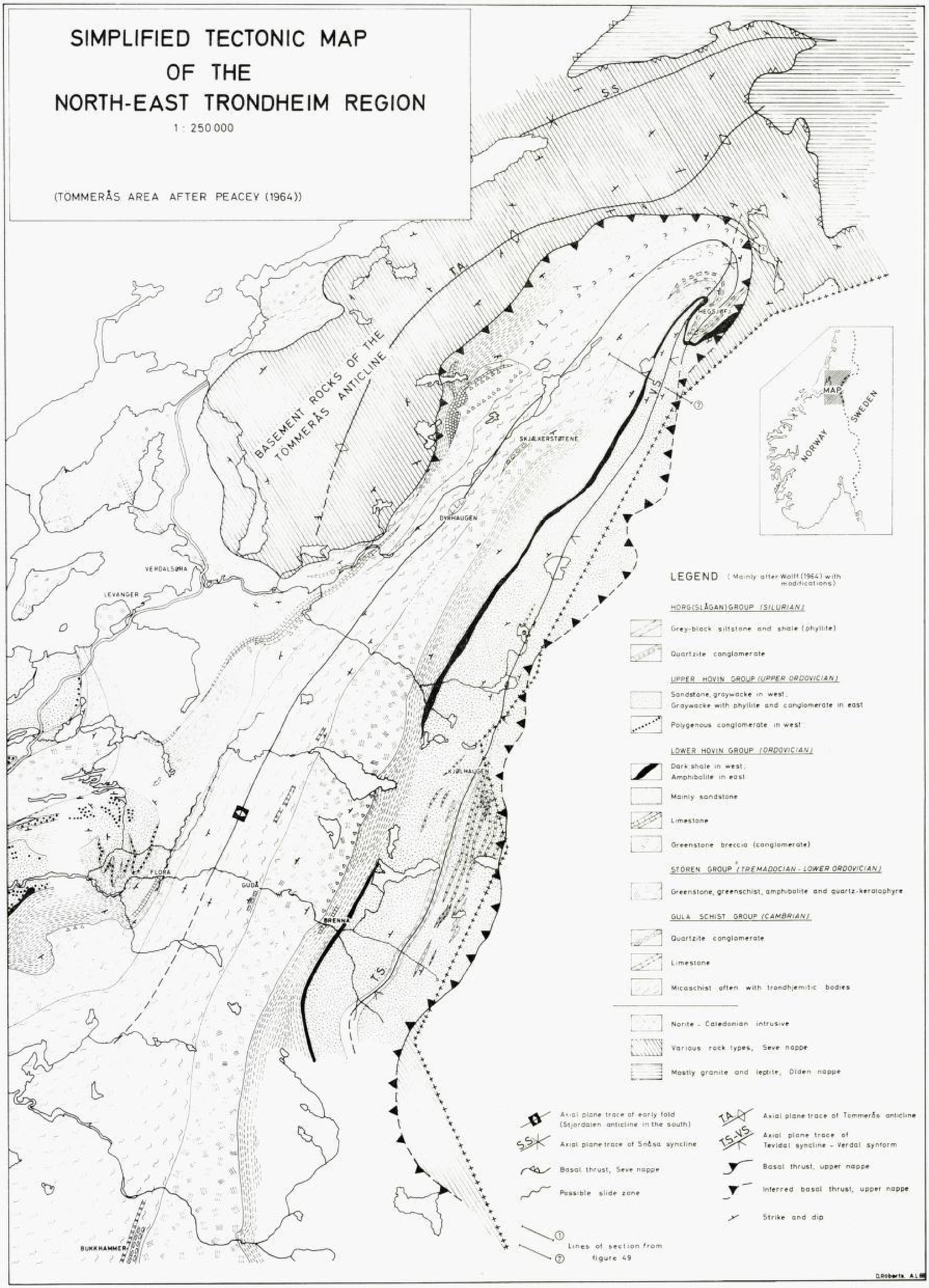


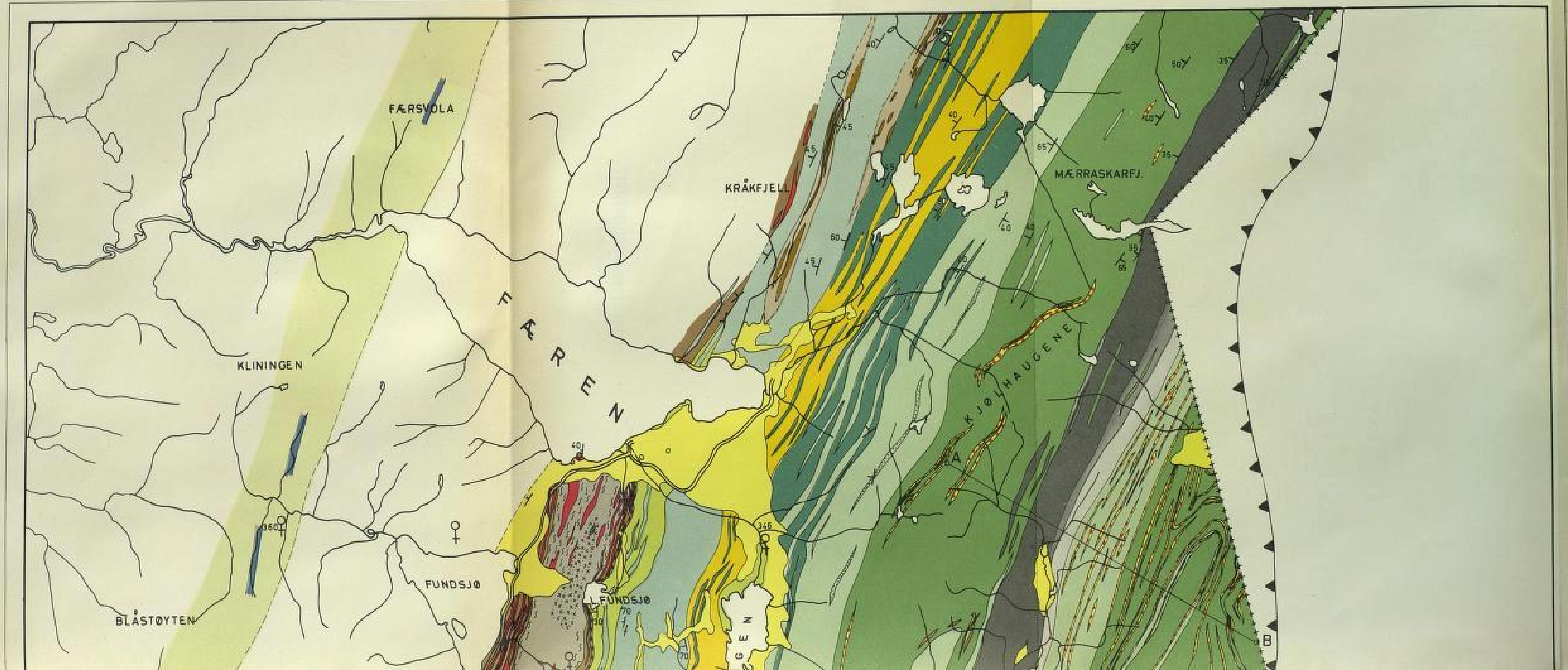
SIMPLIFIED TECTONIC MAP
OF THE
NORTH-EAST TRONDHEIM REGION
1 : 250 000
(TÖMMERÅS AREA AFTER PEACEY (1964))
BASEMENT ROCKS OF THE TÖMMERÅS ANTICLINE
TA
S.S
V.S
T.S
HEGSJØEN
SKJÆKERSTØTENE
DYRHAUGEN
VERDALSØRA
LEVANGER
KJØLHAUGEN
FLORA
GUDÅ
BRENNA
BUKKHAMMER
NORWAY
SWEDEN
MAP
LEGEND (Mainly after Wolff (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 (Stjordalen 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.


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.

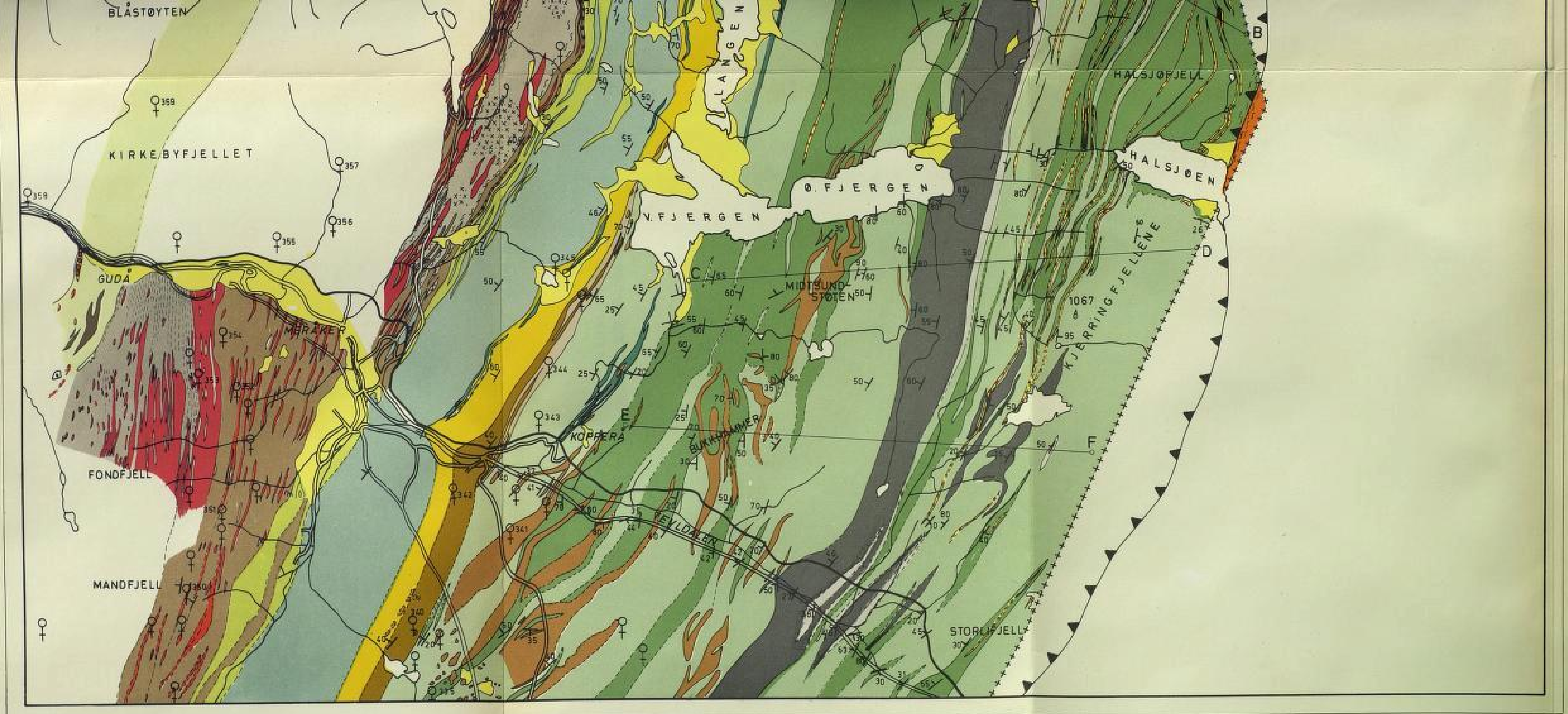
The Brenna conglomerate.

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.

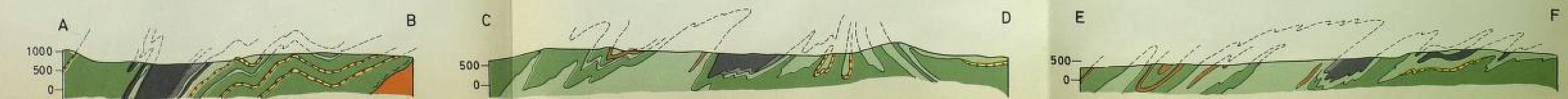
Lines of section.

Foliation, lineation.

Mylonite zone.

Thrust plane.

Occurrences of pyritic and chalcopyritic ore.



GEOLOGICAL MAP OF THE TRONDHEIM REGION

GEOLOGISK KART OVER TRONDHEIMSFELTET

1 : 500 000

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

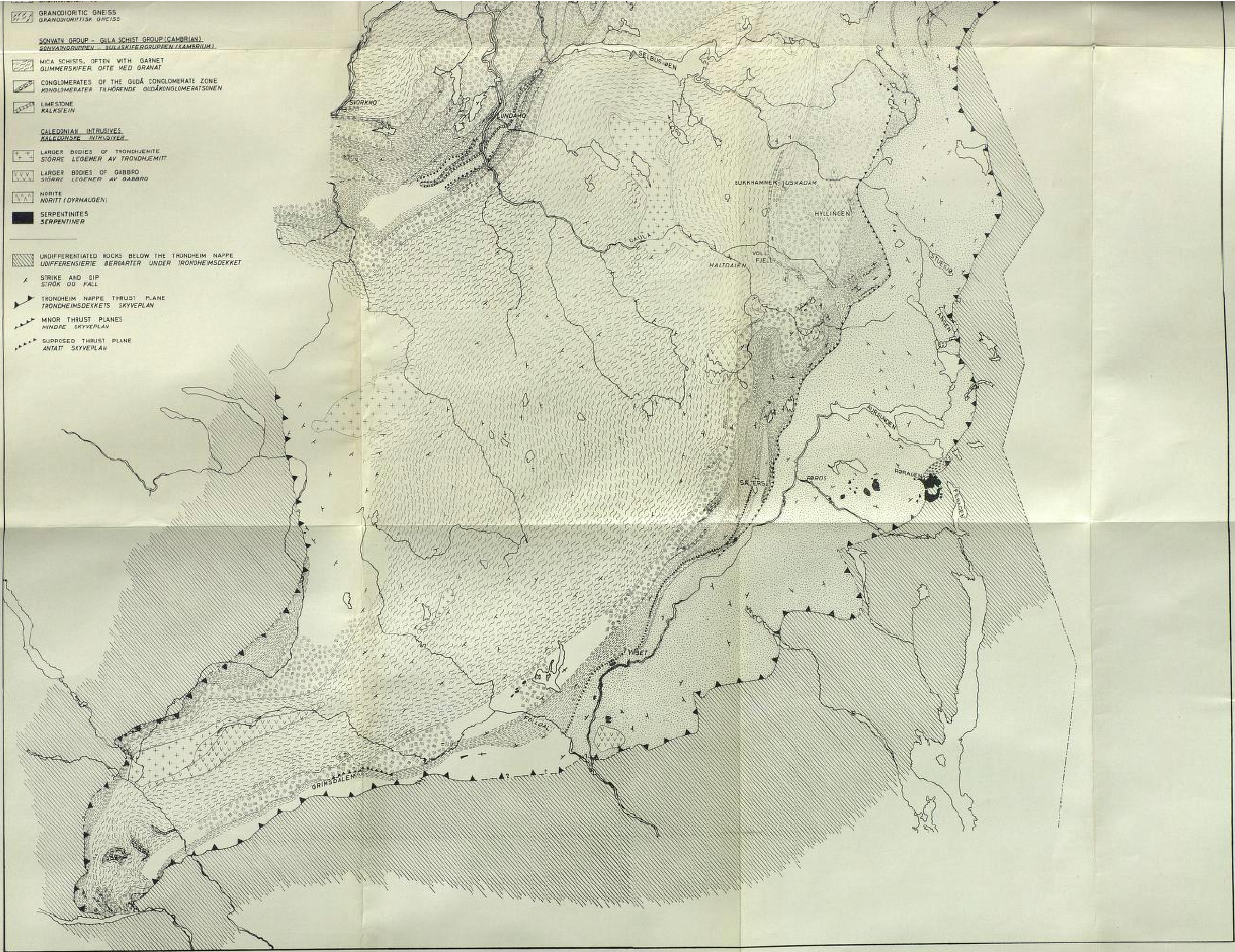
HÖLONDA, TROMSDALEN, BRENNA AND SIMILAR LIMESTONES
HÖLONDA, TROMSDALEN, BRENNA OG LIGNENDE KALKSTEINER

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
ÅRGEN
AURSUNDEN
RØRAGEN
RØROS
SÅTERSJ.
FERAGEN
TYNSET
FOLLDAL
GRIMSDALEN