The Caledonian Mountain Chain of the Southern Troms and Ofoten Areas

Part III. Structures and Structural History

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Gustavson, M. 1972: The Caledonian mountain chain of the Southern Troms and Ofoten areas. Part III. Structures and structural history. *Norges geol. Unders. 283*, 1–56.

The main lines of the structural geology of an area of about 13,000 km^2 in the Caledonian mountain chain of Northern Norway are described. The major structure is characterized by nappes with a thrust front in the east above autochthonous Precambrian basement and Lower Cambrian sediments. The discovery of basal-type psammitic rocks in the westernmost part of the area strengthens the opinion that the basal gneisses to the south and west of the Caledonian nappes constitute Precambrian basement. The basement took part in extensive folding and thrusting during the main Caledonian orogeny, but mobilization features are absent.

The structural history can be summarized as follows:

Early epeirogenic movements are indicated by the Elvenes (–Harstad–Evenskjær) conglomerate zone. This polymict conglomerate is younger than the Evenes marble group.

The main Caledonian deformation commenced with an episode of isoclinal folding, F_1 (in eastern areas an even earlier episode of isoclinal folding has been recognized by other workers), with an associated axial plane schistosity. The main regional metamorphism was coeval with this folding, although some porphyroblastic minerals are younger. F_1 fold trends vary, but are frequently E–W to WNW–ESE or NE–SW. A strong lineation, usually aligned WNW–ESE, is of the same age.

The F_2 deformation produced open to tight folds (e.g. the Ofoten Synform) on all scales, refolding the F_1 schistosity, but without the development of any axial plane foliation. F_2 folds trend both NW–SE and NE–SW. In eastern areas, N–S trends have also been recognized. Evidence of a later episode of flexural folding, F_3, is found in the western and central parts of the map area, with axes trending NNE–SSW.

Thrusting seems to have occurred partly during the F_1 episode, and partly between the F_1 and F_2 fold episodes. In eastern areas some thrust movements also postdate the F_2 folding. Regional comparisons with areas to the north and south of the map area demonstrate great similarities in the sequence of structural units and in the structural history within a considerable part of the Caledonian mountain chain.

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CONTENTS

Previous structural investigations

Tectonic structures have previously been studied to only a limited extent within the map area. Considerable information, however, is available in descriptions of the general geology from parts of the ground, and from adjoining areas.

Foslie's (1949) petrological investigations of the iron ores of the Håfjell area are based on the important structural statement that the metasedimentary sequence in the central Ofoten area forms a fairly regular synclinorium, trending NE–SW. This same author's description of the map sheet Tysfjord (Fig. 1) (1941) also contains a good deal of structural information; the presence of two mutually perpendicular fold systems, trending NNE–SSW and WNW–ESE, is noted, and there are several illustrations of the larger fold structures. The existence of two perpendicular fold trends has been supported by recent detailed investigations by Juve (1967) in the Håfjell area.

Vogt (1941), in a short paper, has described the central and eastern parts of a NW–SE profile north of Ofotfjorden (Fig. 1). The overall structure was interpreted as a series of folds with axial planes dipping about 50° NW and resulting in numerous repetitions of the metasedimentary strata. In the same paper, a brief description of the tectonic influence of Caledonian deformation on the Precambrian Rombak window was presented, as well as a short discussion of a fault along Tjeldsundet (Fig. 1).

Vogt (1918, and posthumous paper, 1967) has also commented briefly on the relations between Precambrian and Caledonian rocks in Eastern Troms. The basement and its cover of autochthonous Hyolithus Zone sediments are described in detail, and it is stated that the upper boundary of these rocks is defined by the thrust plane below the Caledonian allochthonous sequence.

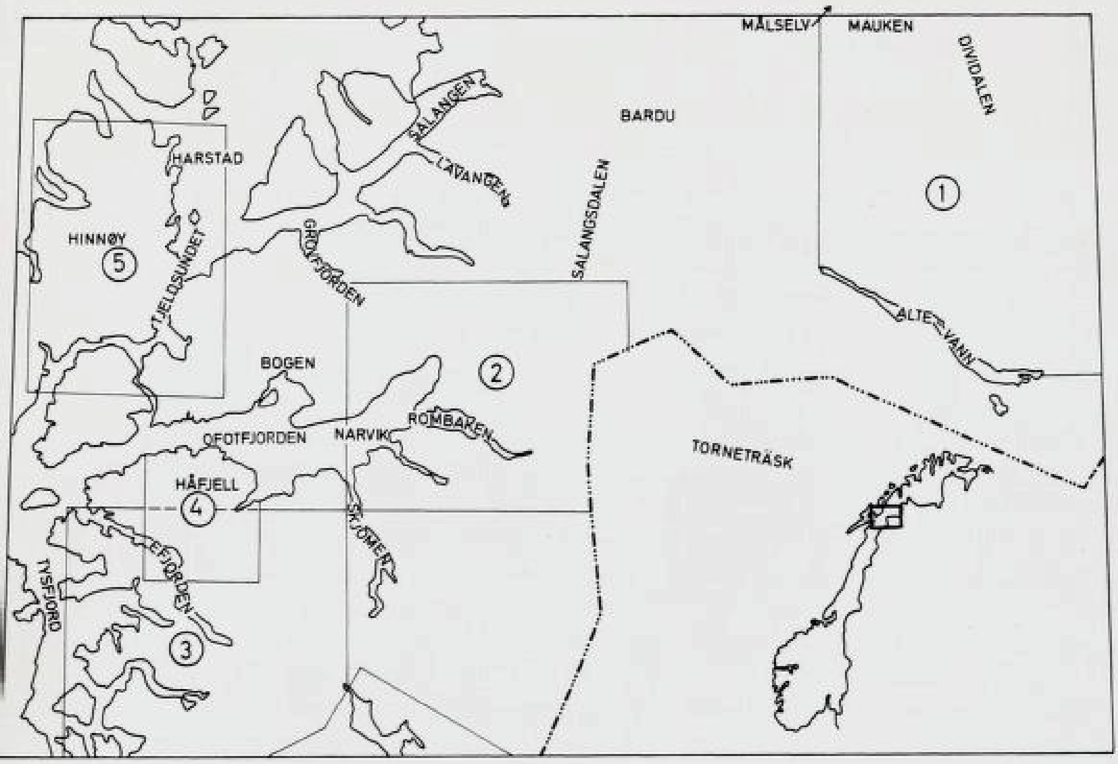


Fig. 1. Localization and sketch of map area with some important geographical names. 1: Part of the area described by Kalsbeek & Olesen (1967). 2: Map sheet Narvik (Vogt 1950). 3: Map sheet Tysfjord (Foslie 1941). 4: Area described by Foslie (1949) and also shown on Fig. 29 of the present paper. 5: The Harstad–Tjeldsund area, shown on Fig. 19 of the present paper.

A survey of the main structural features of Eastern Troms has recently been given by Kalsbeek & Olesen (1967): see Fig. 1. A detailed study of part of this area was carried out by Berthelsen (1967), while Mortensen (1972) has given an account of the main structural features of the area south of Altevann. A special investigation of the hard schists (Mortensen 1970) should also be mentioned.

From the adjoining Torneträsk area of Sweden (Fig. 1), the papers by Kulling (1962, 1964) and Lindström (1955a, b, 1957, 1958a, b) are important. The role of nappe tectonics is stressed in all these papers, and an extrapolation of the nappe boundaries into the Ofoten area was proposed by Kulling (1964, pp. 114–115).

Kautsky (1946, 1953) and Oftedahl (1966) offered structural interpretations based on Foslie's maps from Tysfjord and adjoining areas to the south and on Kautsky's mapping on the Swedish side of the border. Both authors proposed extensive westward extrapolations of the 'Seve' nappe.

No publications dealing with structural geology have been presented from the westernmost parts of the map area.

Present investigations

The main structural problems which have been investigated during the present work and which form the basis of this paper can be summarized as follows:

A. The boundary relations between the basal massifs and the metasedimentary sequence, and the influence of Caledonian deformation on the basal massifs.

B. The major structural features of the metasedimentary sequence; thrust boundaries (nappe tectonics) and major fold structures.

C. The sequence of structural events.

Problems relating to the basal massifs have been discussed in some detail in a previous paper (Gustavson 1966, pp. 27–43); recent mapping has, however, necessitated a revision of some of the earlier statements. For the westernmost areas, on Hinnøy for example, the map presented in the 1966 paper was based on unpublished data from Vogt and his assistants and on my own reconnaissance work. The metasediments were tentatively classified according to the general stratigraphical scheme established for the central Ofoten area. In 1966 and 1967, aerial photographs (scales 1 : 15,000 and 1 : 13,000) became available for most of the area west of Tjeldsundet, and a preliminary study of these revealed structural relationships not previously detected. As the important boundary relations between the basal massifs and the metasedimentary sequence could be studied within these areas, a fairly detailed mapping was carried out there in the years 1968–70.

In most of the earlier literature on the present and adjoining areas the importance of thrust nappe tectonics has been stressed (Vogt 1918, 1967, Lindström 1955a+b, 1957, 1958a+b, Strand 1960, 1961, Oftedahl 1966). Holtedahl (1944), in a discussion of the basal gneiss areas, also mentioned the possibility of alpine-type fold nappe tectonics being present. An exception to these views of large-scale nappe tectonics is that of Foslie (1941, 1942, 1949), whose interpretation contains no regional overthrust in any part of the areas south of Ofotfjorden. Because of this, the present investigations have been directed towards an evaluation of the importance of nappe structures within the area, primarily by regional mapping and comparisons of sequence. Mapping has been carried out on various scales, mostly 1 : 50,000, though special areas, for example the Harstad–Tjeldsund area (see above), have been mapped in greater detail. Minor structures (folds, lineations, etc) have been studied to a certain extent in order to ascertain their relationships to the major features, and also to determine the relative ages of the structural episodes, although the available data are insufficient for any statistical analysis.

The sequence of structural events has been studied chiefly in the westernmost areas (Harstad–Tjeldsundet). Microscopic investigations of metamorphic textures in rocks of these areas are not, however, included in the present paper, which concentrates on a discussion of the main structural features.

Age of the western basal gneiss area and boundary relations to the metasedimentary sequence

A summary of the various opinions concerning the 'basal gneiss' areas and a discussion based on knowledge at that stage of the investigation were given in a previous paper (Gustavson 1966, pp. 27–43). As a conclusion it was pointed out that chemical, mineralogical, and structural relations indicate a close relationship between the basal granite of Tysfjord and the Precambrian Rombak granite. No definite conclusions were drawn as to whether the central parts of the western granite areas were completely recrystallized during the Caledonian orogeny, or if Precambrian structures are still present. The gneiss-granites within the lower part of the metasedimentary sequence were thought to be overthrust Precambrian granite, in accordance with Kautsky's (1946) hypothesis. As far as the Hinnøy area was concerned, the author's own observations showed that thrusting had taken place. It was also suggested that a certain amount of mobilization of the basal granite had occurred in Caledonian time and that this may have caused a porphyroblastesis in the neighbouring mica schist.

Investigations since 1968 have made it possible to decide these questions with greater certainty, not least concerning the age of the basal granite. One of the main points in Foslie's (1941) discussion of the Tysfjord granite was that on a regional scale the granite had a discordant boundary to the overlying schists. It was suggested (Foslie, op. cit.) that northwards along the western side of the Håfjell synclinorium the granite bordered against successively higher stratigraphical units. According to this, the granite should be a Caledonian intrusive massif. Vogt (1941, p. 211) described the granite boundary north of Ofotfjorden as an injection contact ('injeksjonskontakt') and interpreted the granite as a Precambrian granite remobilized during the Caledonian orogeny.

Based on these earlier investigations, and especially on the available maps, the present author then concluded that 'basal type sediments are absent west of the Håfjell synclinorium' (Gustavson 1966, p. 41). Later mapping, in 1968, revealed that these views were not correct:

In the area between Gausvik and Storvann (Fig. 19) *a series of quartzites and meta-arkoses was detected along the granite border, this constituting the basal part of the metasedimentary sequence.* It has been possible to trace this basal series southwards to Sandtorg. Locally, within the series, a quartz pebble conglomerate is developed. The foliation within the margin of the granite-gneiss is parallel to the contact with the metasediments, whereas in the bulk of the massif it shows a variable trend. It seems probable that the foliation *within* the massif could be a Precambrian structural feature, but it is generally too weak and too sporadically developed for any safe conclusion to be drawn at present. The gneiss foliation along the boundary is probably of Caledonian age. There seems little reason to doubt, however, that *in this area we have the Precambrian basement with its cover of basal arkosic and quartzitic sediments.* If it is accepted that the foliation in the granite immediately below the basal sediments is a Caledonian feature, it implies that movements *may* have occurred along the plane of contact. It is likely, however, that such movement took place without causing any great disturbance *within* the lower sequence. The importance of thrusting or sliding at this level is discussed later (pp. 36–37).

Mapping in recent years has not given support to the idea that the basal granite, in these western areas, has intrusive or injection contacts against the metasediments. The supposed mobilization and porphyroblastesis in the Kasfjord area, referred to in the earlier paper (Gustavson 1966, p. 34), has now been reconsidered. What was thought to be a Cambro-Silurian biotite schist with feldspar porphyroblasts is probably a highly deformed and schistose variant of the basal gneiss. The age of the somewhat deformed porphyroblasts cannot be decided, but it might well be Precambrian. At least, it is obvious now that this rock cannot be used as evidence of Caledonian mobilization. On the other hand, within this and other parts of the Harstad–Tjeldsund area, there are numerous examples of Caledonian thrusting of slices of the basement granite (pp. 12–15), this providing confirmatory evidence for previous opinions (Gustavson 1966).

Rb–Sr dating of granites from Ramsund (Fig. 19), carried out by Heier & Compston (1969) lend support to the view that the rocks in question are Precambrian, as the determinations have given an age of 1550 (± 35) mill. years. The occurrence of pebbles and boulders of basal massif granite in the Harstad conglomerate (see later) would also indicate that these massifs are Precambrian basement rocks and not of Caledonian age. Caledonian intrusives emplaced into the metasedimentary sequence before the deposition of the conglomerate (Gustavson 1969, p. 50) are small bodies of trondhjemitic composition, in all respects different from the basal massif granitoids.

As a conclusion, therefore, it can be stated that the Precambrian age of the basal granites and gneisses now seems to be proven beyond doubt and that, locally, psammitic basal metasediments are also preserved in these western areas. Nevertheless, the Precambrian rocks have, to a great extent, taken part in folding and thrusting during the Caledonian orogeny, although there is no definite evidence of remobilization of the granite along this western border of the Caledonian metasedimentary sequence.

Structural geology of the Harstad–Tjeldsund area

LITHOLOGICAL UNITS

A detailed description of all lithological units in the Harstad–Tjeldsund area will not be presented here. Only the main features of the lithostratigraphy are outlined, as a base for the structural description.

Although it is difficult to establish a stratigraphy in such a strongly deformed area as this, it is more than likely that in moving up from the basement, unless one encounters repetitions or signs of major tectonic disturbance, one is passing through a sequence which may reflect the primary stratigraphy. This is even more probable when the same sequence is met with in different parts of the area associated with basement or thrust sheets of basement + metasediments. Stratigraphically important evidence is also provided by the pebble material of the conglomerates at Evenskjær and in the Harstad area. Based on this information and the reasoning outlined above, the following tentative stratigraphy has been established.

Youngest: 6. Mica schist
5. Calcareous mica schist and polymict conglomerate
4. Mica schist and amphibolites, the latter being partly fragmental
3. Marbles, including colour-banded marble, grey calcite marble and white dolomite marble
2. Mica schist, locally graphite-bearing or calcareous
1. Meta-arkose and quartzite

Basement: Precambrian granite and gneiss.

Lithological unit 1, the psammitic rocks (Fig. 2), has its main distribution in the Sandtorg–Storvann area (Fig. 19), as already mentioned (p. 5). Its thickness is some tens of metres, possibly 100 m at the maximum. Similar rocks are found elsewhere within the thrust units in association with Precambrian gneisses (p. 13).

Lithological unit 2 shows facial variations but is usually developed as a normal (garnet)-mica schist.

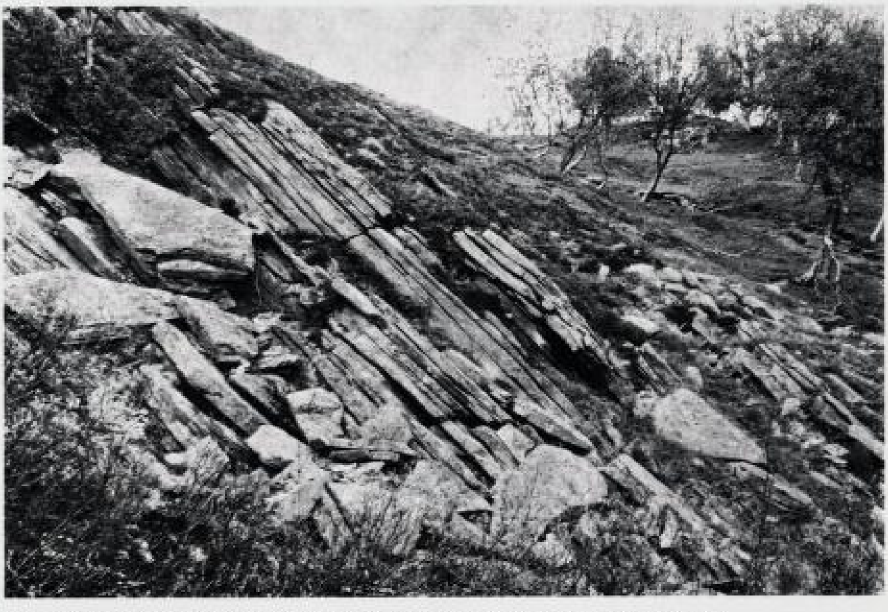


Fig. 2. Quartzite and meta-arkose (lithological unit 1), east of Storvassbotn.

Fig. 3. Boundary between calcite marble (lower left corner) and dolomite marble at Leikvik. Note cobbles of calcite marble in the dolomite marble.

The most common rock-type of *unit 3* is a grey calcite marble. Quite prominent, however, are the colour-banded marbles and a white, sugary-grained, dolomite marble. The mutual age relationship of these marble types in the Tjeldsund area is:

Youngest: White dolomite marble
Grey calcite marble
Colour-banded marbles

Along the road at Leikvik (Fig. 19), a boundary between white dolomite marble and underlying calcite marble has been observed and this is interpreted as a primary erosional surface. A type of channel in the calcite marble appears to be filled by the dolomite marble, and boulders or fragments of the former are found within the latter (Fig. 3).

Lithological unit 4 consists of alternating layers of mica schist and amphibolite. Some amphibolites may be intrusives and related to the larger amphibolite bodies which intrude marbles in Blåfjell (Fig. 19), but amphibolites of unit 4 in the Harstad area are partly fragmental and of a special chemical composition. Some of them are hornblende-zoisite schists rather than amphibolites. It is believed that these amphibolites are supracrustal in origin and that the fragmental rocks are possibly agglomerates. Further details of these rocks will be presented in a later paper.

Unit 5 is developed partly as a calcareous mica schist with scattered pebbles of various rock-types, and partly as a typical polymict conglomerate, especially within the town of Harstad and at Evenskjær (Figs. 4, 5). The pebble material varies considerably from one locality to another, but in the Harstad area the

Fig. 4. Evenskjær conglomerate. The flattened pebbles lie within the F_1 schistosity and are further deformed by F_2 folds.

Fig. 5. Granite boulder in calcareous mica schist, Harstad.

most common pebbles are of quartzite, granodiorite or quartz diorite and red granite (of basement type); quartzite and occasional dolomite boulders are present at Evenskjær. The granite pebbles and boulders show that the basement was exposed to erosion in parts of the area at the time of conglomerate

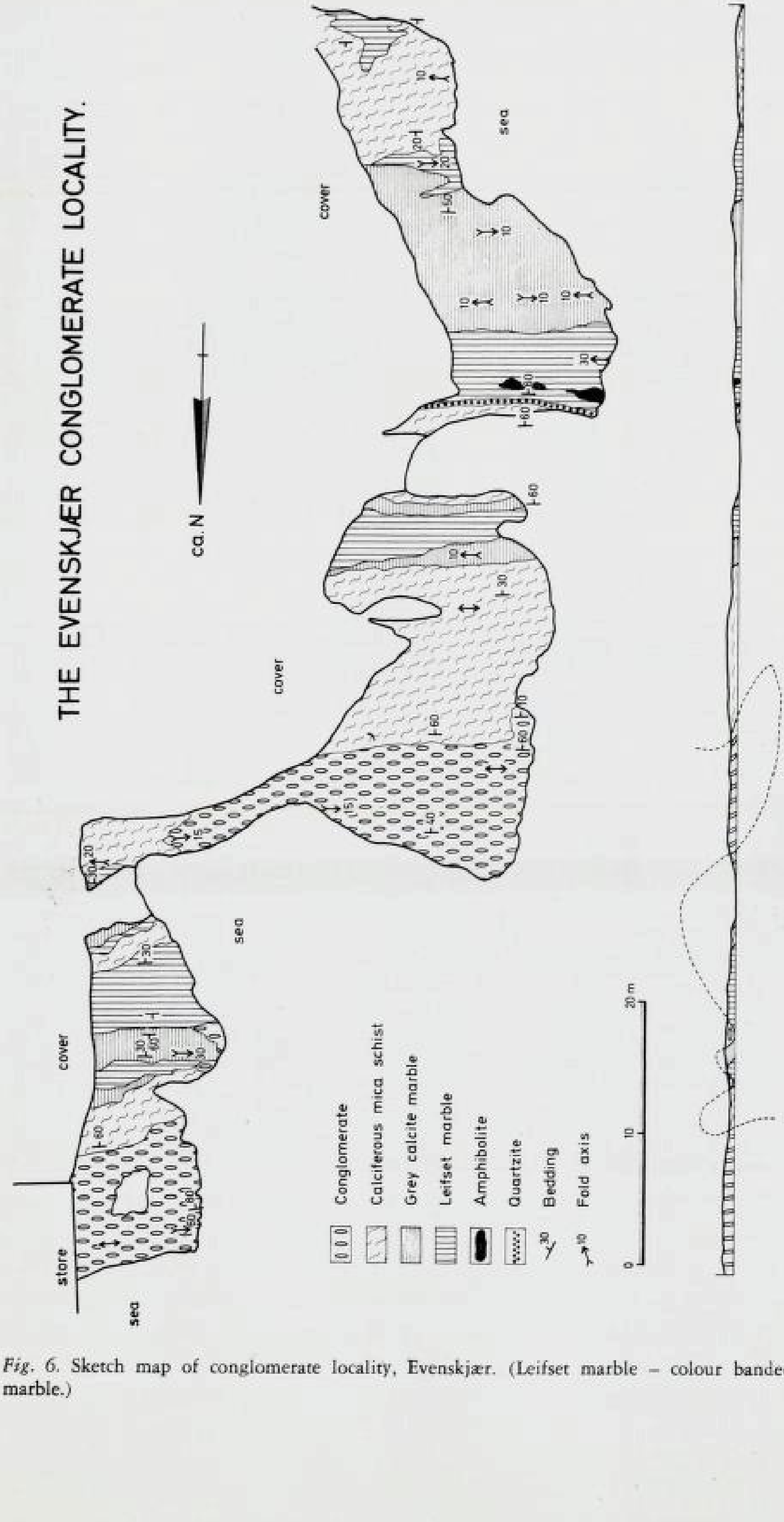


Fig. 6. Sketch map of conglomerate locality, Evenskjær. (Leifset marble – colour banded marble.)

deposition. The relatively strong deformation of the pebbles makes it difficult to ascertain their original shape. The least deformed quartz dioritic boulders, however, show evidence of a fairly high degree of roundness.

Of considerable interest is the presence of colour-banded marble pebbles at a couple of localities. These have been observed in both the Harstad and the Evenskjær areas, and clearly show that *the conglomerate is younger than the (Evenes) marble formation.* This is contrary to Th. Vogt's (1941, pp. 205–208) opinion, which was adopted by the present writer in the description of the regional stratigraphy (Gustavson 1966, p. 44). It should be mentioned, however, that Vogt had earlier (1922, p. 717) interpreted the conglomerate as the younger unit, a view he changed in the 1941 paper on account of his revised interpretation of the structural relations at Evenskjær. Fig. 6 shows the present writer's map and structural interpretation of the locality. The occurrences of colour-banded marble pebbles are not located within the sketch map area but some few hundred metres to the northeast, close to the main road. White dolomite marble, however, is found as pebbles and cobbles within the area covered by Fig. 6, together with the dominant quartzite material.

As no fossils have been found, the absolute age of the rocks in the Harstad–Tjeldsund area is, as yet, unknown. In correlating the sequence with the stratigraphy of the rest of the region (Gustavson 1966, p. 44) the conglomerate zone is of importance. The pebble material and general appearance of this lithology are in favour of a correlation with the Elvenes conglomerate in the central Ofoten area (see Fig. 29). Furthermore, it is possible that the marble unit is equivalent to the Ballangen marbles (p. 31), though it is also likely that this unit is absent east of the Ofoten Synform. With regard to the other units of the Harstad–Tjeldsund sequence no distinctive features have been found which would permit their safe correlation over the rest of the area. The sequence is clearly much thinner than in the central Ofoten area and important metasedimentary formations, for example the thick and characteristic Narvik schists and gneisses, are absent west of Tjeldsundet. This can probably be explained by primary sedimentary causes rather than by tectonic movements.

STRUCTURAL HISTORY

The recent investigations in the Harstad–Tjeldsund area have shown that at least 3 episodes of deformation can be distinguished:

F_1: This is represented by isoclinal folding with the development of an axial plane schistosity. Axes show a variable orientation but would appear to have had a general E–W or WNW–ESE trend before the onset of the F_2 deformation. Important lineations, for example conglomerate pebble elongation, are of F_1 age.

F_2: Most major folds of the area are considered to be F_2 structures. Folding occurred on axes trending NW–SE (or WNW–ESE) and NE–SW. It cannot be directly proved that these two sets of folds are strictly contemporaneous as they occur in different areas, but both sets refold the F_1 schistosity, axial planes, and lineations and the type of minor folds is generally the same for the NE–SW as for the NW–SE folds. A common lineation of F_2 age is a crenulation of the F_1 schistosity (Fig. 18).

F_3: Flexural folds about N–S or NNE–SSW axes were developed during this episode.

In addition to the main Caledonian deformation, an early disturbance is indicated by the Harstad–Evenskjær conglomerates and their supposed equivalent, the Elvenes conglomerate, south of Ofotfjorden. No deformation or fold structures have been observed in the sequence stratigraphically below the conglomerate which are not present in other parts of the sequence. However, the varied pebble material and the presence of basement rocks as pebbles and boulders in the conglomerate, together with its wide areal extent, suggest that the disturbance represented by the conglomerate zone is one of regional significance. The nature of the Evenes marble formation points in the same direction; while it is a relatively thick formation in the westernmost areas, it appears to be completely absent east of the Håfjell synform. As marble pebbles are common in the conglomerate in the latter area too, it seems probable that the lower surface of the conglomerate marks a major discordance in the sequence. Movements thus occurred at an early stage in the Caledonian history of the area, and although the precise nature of these movements cannot yet be established, they were probably of an epeirogenic character.

Thrusting is thought to have occurred between the F_1 and F_2 episodes and possibly during the F_2 deformation. Some evidence also suggests that thrust movements occurred during the F_1 fold episode, as will be discussed below.

Faulting *may* be related to the F_3 episode of folding, as faults west and east of Tjeldsundet are clearly younger than F_2 fold structures. It is quite possible, however, that the faulting post-dates all the fold episodes in this area (see p. 45).

THRUSTS

It has already been noted (p. 5) that the psammites in the Sandtorg–Storvann area represent the basal unit of the metasedimentary succession. As discussed in a later section (p. 37) it seems probable that the lower boundary of these metasediments represents the western continuation of the basal thrust plane of eastern areas. This is supported by the gneissification of the granite along this boundary plane and by the lithology of the lowermost metasediments in the Gausvik area.

The granite or the granite gneiss from Lilleng to Storvann (Fig. 19) lies *above* these metasediments, and is considerably tectonized locally, especially in its southern part. To the north of the granite there is another sequence of metasediments with a thin quartzite at the base. It would therefore seem likely that this granite belt forms the base of a thrust nappe lying above the rocks of the Gausvik area. No simple fold structure, at least, can explain the occurrence of this granite belt and its adjacent metasedimentary sequences.

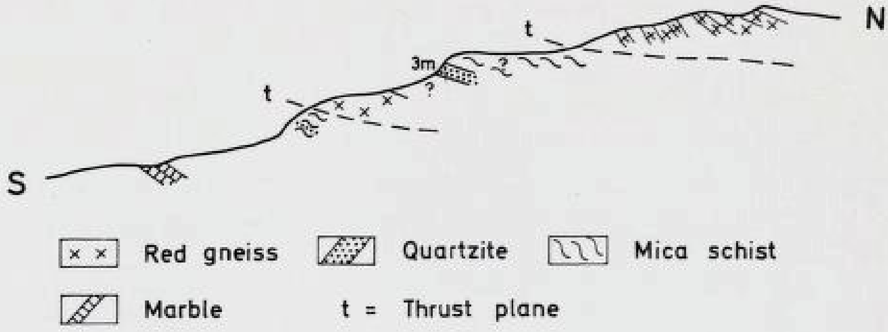


Fig. 7. Profile across the thrust zone north of Storjord.

The thrust plane to this supposed nappe can be traced to the east side of Storvann. Its continuation is then probably beneath the covered area north of Storvann from where it joins up with thrust zones in the area south of Finnslettheia. The lower part of the sequence above the Precambrian occurs in this area around the farm Storjord, although the basal contact is not exposed. Northwards from Storjord, however, the red Precambrian gneiss is again exposed above the metasediments at two levels, with nearly horizontal lower boundaries and quartzitic rocks above each gneiss slice (Fig. 7). These relationships can hardly be interpreted other than the result of overthrusting of the basement granite with its cover of basal quartzite, a similar situation to that described from the Lilleng–Storvann area. Meta-arkoses and mylonitic gneisses cover most of Finnslettheia. The mylonitic rocks, as well as the metasediments north of Finnslettheia, are situated tectonically below the gneisses of Svarthompen–Nattmålsnuten (Figs. 16, 19).

It is probable that the granite of Steinfjell–Kjeipfjell ('S' and 'K' on map, Fig. 19) is equivalent to the lowermost part of the thrust unit described above. Below this basement-type granite, but above the true basement are situated trondhjemite, amphibolite, and metasediments of much the same type as in the Gausvik area. Along the eastern side of the Steinfjell–Kjeipfjell area these metasediments are reduced to a single marble horizon, a feature probably resulting from the thrust movements.

The metasediments and the metamorphosed igneous rocks of this thrust unit occupy the entire area from Lilleng to Harstad. One of the thrust planes again crops out west of Harstad, on the western side of Kjelhusåsen, where a thick tectonic breccia is exposed (Fig. 8). Its continuation to the north and to the south is in covered terrain but can be extrapolated with some degree of certainty. The sequence below and west of this thrust is obviously not authochthonous. Thrust slices of Precambrian granite appear in at least three niveaux; at Gåsvannet, in Sollifjell, and a short distance above the basement at Mølnelva. The thrust in Sollifjell can be followed for some kilometres to the northeast. The sequence below this thrust is much the same as at Gausvik. It is

Fig. 8. Cataclastic granite lenses in the Kjelhusåsen thrust zone.

considered likely that these thrusts were formed contemporaneously with the Kjelhusåsen thrust and that they form a thrust zone, rather than separate thrust planes, of regional importance. The thrust zone then forms the base of a major thrust nappe, which is here termed the Harstad Nappe. In all probability, this can be correlated with the Seve Nappe of eastern areas (p. 37). Its regional extension is shown on the map, Pl. I (at the end of the paper).

The Harstad Nappe dips below the fine-grained gneisses west of Svarthompen (Fig. 19). Although their origin will not be discussed here, these gneisses are interpreted partly as a fine-grained facies of the Precambrian granite, and partly as meta-arkoses derived from the Precambrian rocks. The gneisses alternate with quartz-mica schists and local marble and quartzite, the repetitions being due to folding during the F_2 phase. The rocks of this area, from Nattmålsnuten to Straumen, are interpreted as forming a separate nappe, the Straumsbotn Nappe. This is considered to be of only local importance, because of the direct connection with the basement to the south.

As to the age of the thrusting of the Straumsbotn and Harstad Nappe units it is evident that the latter pre-dates the former. At least this must be the case if it is admitted that there once was a connection between the thrust plane in Finnslettheia and that of Steinfjell (Fig. 19). If this was the younger structure of the two it ought to cut through the Straumsbotn Nappe somewhere, but this is

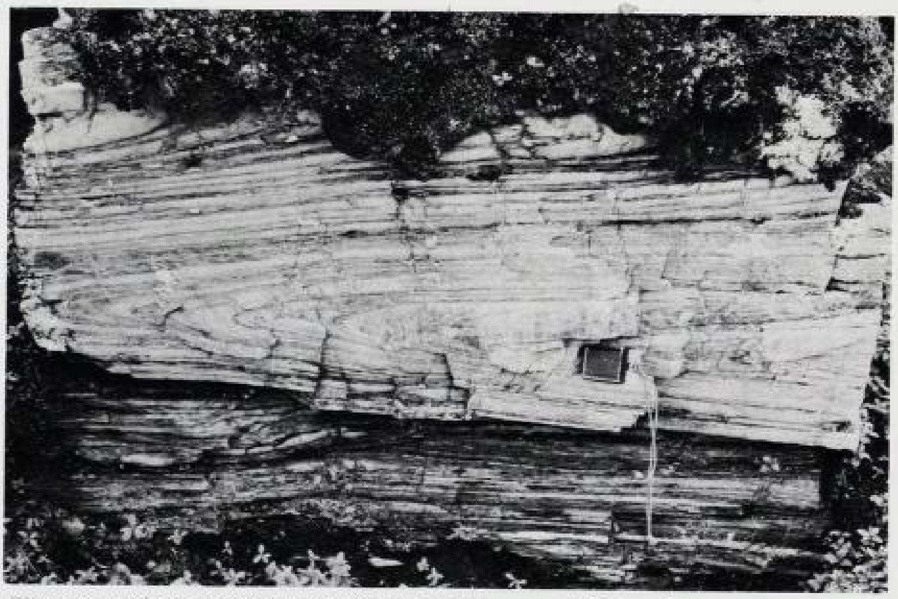


Fig. 9. Recumbent F_1 fold in quartzite, Helleodden, east of Tjeidsundet.

clearly not the case. Both nappes were strongly folded during the F_2 fold episode and cannot therefore be younger than this deformation. Chloritization phenomena at the thrust planes, for instance at Sollifjell and Kjelhusåsen, indicate movements after the main metamorphism, which is F_1. The thrusting here should thus be *post-F_1, pre-F_2*. It cannot be ruled out, however, that movement along the thrusts occurred more than once. In that case the chloritization phenomena may belong to the latest movement phase, and it is then possible that early phases of thrusting occurred *during* the F_1 episode as will be discussed in a later section (p. 40). An argument in favour of F_1 thrusting of the Harstad Nappe is that there is no obvious change in the metamorphic grade at the actual thrust boundary. Relations are thus similar to those found in the Rombak area (p. 39) at the base of the Seve Nappe. As can be seen from the map, Pl. I, and as already noted above, the thrust below the Harstad Nappe is believed to correspond to the Seve thrust of eastern areas.

FOLDS

F_1 folds: Although definite major folds of F_1 age cannot be demonstrated in these western areas, minor folds occur frequently. The F_1 minor folds are usually of a similar style, sub-isoclinal and commonly recumbent (Figs. 9, 10, 13, 14). A schistosity axial planar to these folds was developed throughout the area, although some of the porphyroblastic minerals are of post-F_1 age. Because of later refolding, especially during F_2, the orientation of F_1 fold axes is variable (Fig. 15). In many cases, however, the F_1 axes appear to have had an orientation about E–W. In the Harstad area, for instance, it is noted that F_1 fold axes and lineations on the flat limb of NE-trending F_2 folds are oriented approximately E–W or ENE–WSW.

Fig. 10. F_1 folds and lineations, gently refolded by F_2. Calcite marble. Mulen.

Fig. 11. F_2 folding of S_1 schistosity in layers of amphibolite and light granitoid gneiss, east of quarry, Kilbotn.

F_2 folds: The major fold structures in the Harstad–Tjeldsundet area are of F_2 age. In the south their orientation is about WNW–ESE or NW–SE; in the northern area they are oriented NE–SW to NNE–SSW (Fig. 15), that is, they are parallel to the boundary between the basement and the metasedimentary sequence. (See profiles, Figs. 20, 21.)

Fig. 12. F_2 folds with straight limbs and sharp hinges in amphibolite, Harstad.

Fig. 13. F_1 sub-isoclinal folds (lower and right part of picture) refolded by F_2 folds. Meta-arkose, east of Voktor.

F_2 minor folds are open to tight structures (Figs. 11, 12, 13, 14); in some rocks (e. g. amphibolites) they have sharp hinges and more or less straight limbs (Fig. 12), and approach a similar style. They may resemble some F_1 folds but are easily distinguished from these by the fact that the F_1 schistosity is folded around their hinges. Axial plane schistosity or cleavage has not usually

Fig. 14. F_1 fold refolded by F_2 in quartz-rich mica schist. Road-cut north of Lilleng.

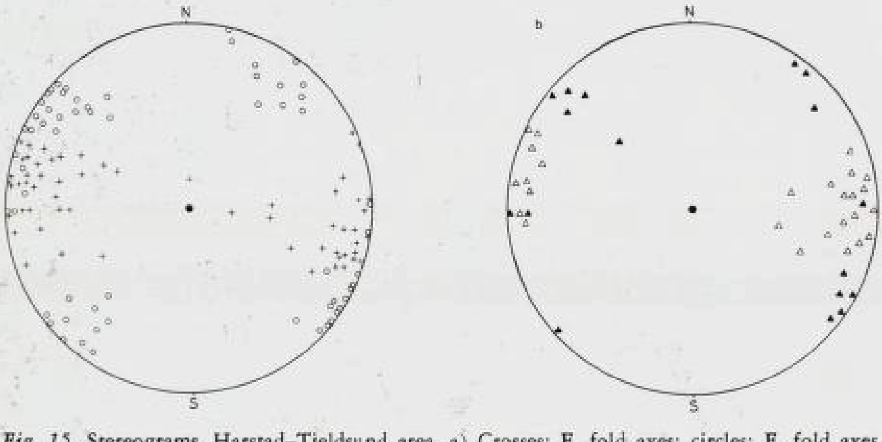


Fig. 15. Stereograms, Harstad–Tjeldsund area. a) Crosses: F_1 fold axes; circles: F_2 fold axes. b) Open triangles: F_1 lineations; filled triangles: F_2 Lineations.

developed during the F_2 episode. The more open F_2 folds are nearer to a concentric style, and not infrequently disharmonic folding can be observed.

The F_2 axes show plunges in the range 10° to 30°. These variations are partly due to the pre-F_2 attitude of the strata before folding, but are probably also the result of younger (F_3) folding.

At Langvann ('L' on Fig. 19), the effects of F_2 deformation are less pronounced in the gneisses of the Straumsbotn Nappe than in the schists below, a phenomenon that may be attributed to the differing competency of the rocks. The folding in the Langvann area (Fig. 16) is revealed by a thin marble, serving as a marker horizon. Erosion along master joints has sometimes produced an

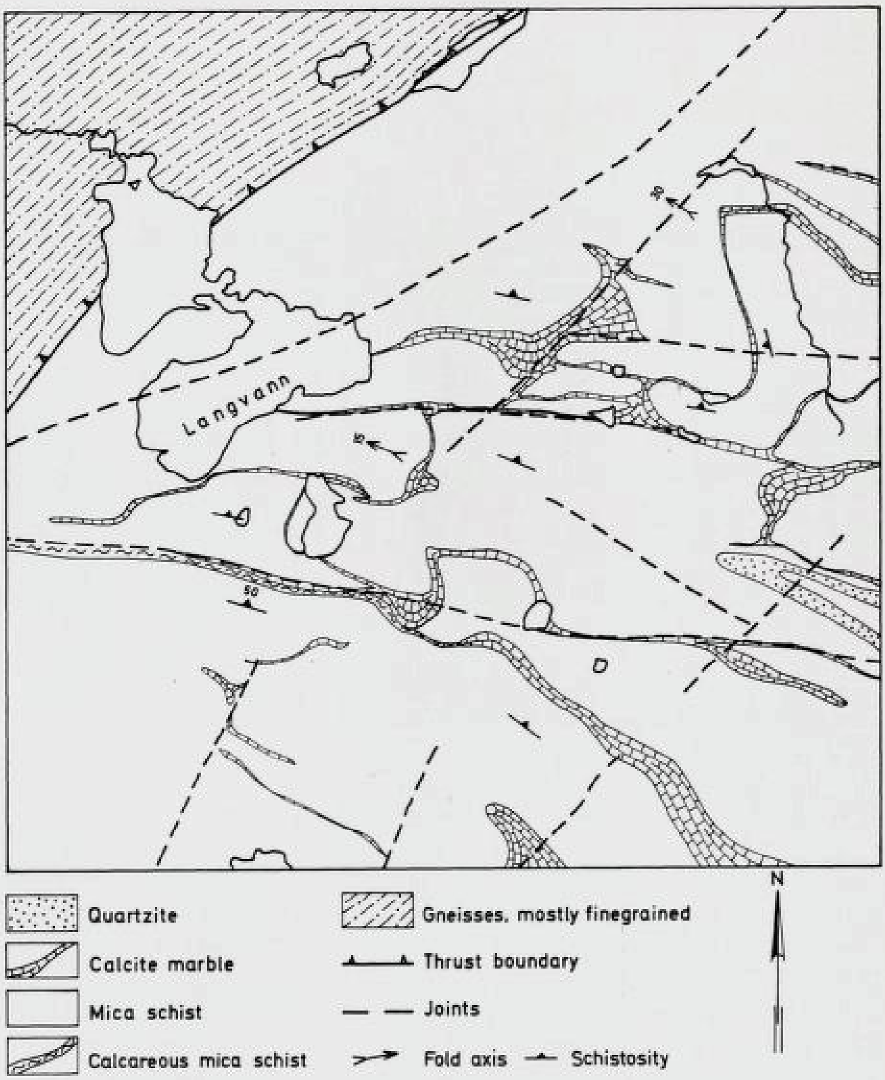


Fig. 16. Geological map of the Langvann area. Scale about 1 : 20,000.

intricate outcrop pattern in this area, but it can be shown that folding was mainly on axes plunging 10–40° to the WNW, with axial planes dipping about 30° to the NNE. Some strike swings are difficult to explain by F_2 folding alone and may be due to effects of the previous F_1 deformation. The greater ductility of the schists in the Langvann area during the F_2 folding as compared with the gneisses above, implies that some movement probably also occurred on the thrust plane below the Straumsbotn Nappe during this episode. Whether the emplacement of this nappe occurred entirely before the F_2 folding cannot be definitely decided. The slight deformation of the thrust plane west of Langvann might be taken as an indication of a relatively late age for the nappe emplacement. However, relations further northwest, in Torskvatsfjell, clearly

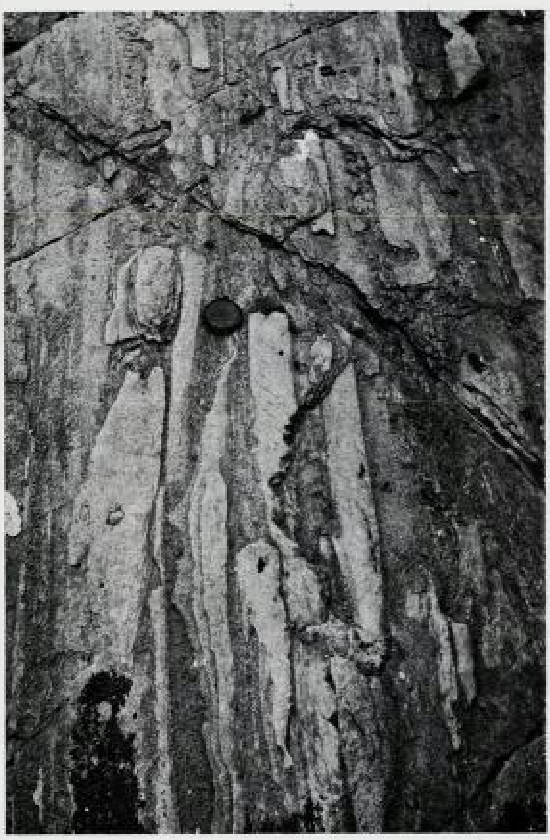


Fig. 17. Pebble elongation lineation in the Harstad conglomerate. The least elongated pebble (beside the lense cap) is granite, the remainder are quartzite pebbles.

show that the thrust plane has taken part in the F_2 folding. Thrusting of the Straumsbotn Nappe must therefore have occurred either during the early stages of the F_2 deformation or between the F_1 and F_2 episodes, possibly as early as at the very end of the F_1 episode.

The sense of overturning of F_2 folds, which is to the south in the Langvann area, is in the same direction also east of this area, towards Tjeldsundet and in the west within the Straumsbotn Nappe. Further south, however, between Lilleng and the bridge over Tjeldsundet, the overturning of F_2 folds is commonly to the north .

The Harstad area is dominated by NE–SW trending folds of F_2 age with an antiform in the central part (Figs. 19, 21). On Gangås, east of the town, the fold trend is more N–S, and although the most prominent structures still appear to be of F_2 age, F_3 folding may play a greater part than realized at present. The older, F_1, deformation in the Harstad area is represented mainly by lineations (Fig. 17) and pre-F_2 schistosity.

F_3 folds: Some open minor folds or flexures have been assigned to a separate and later episode of folding, at which time the rocks involved had attained a rather more rigid state. The F_3 fold axes have a remarkably constant orientation, plunging towards 010°–015°. F_3 flexures deform the F_1 schistosity but apart from this, direct observations of interference relationships between

Fig. 18. F_2 crenulation lineation in mica schist. Road-cut between Lilleng and the bridge across Tjeldsundet.

F_3 and the other fold episodes are scarce. Most observations of F_3 folds are from the areas around Tjeldsundet, but it is not known whether this reflects a special distribution pattern. In general, therefore, this episode of folding is of minor importance in the structural history of the Harstad–Tjeldsund area.

LINEATIONS

Lineations of various kinds are frequent in this area and most appear to be of F_1 age. These are: 1. Elongation of minerals and mineral aggregates. 2. Elongation of conglomerate pebbles. 3. Intersections of S-surfaces. Lineations of F_2 age are usually seen as minor crenulations of the earlier schistosity (Fig. 18). Some less well-defined linear structures may also belong to this episode. The most common F_1 lineation is that produced by intersection between original bedding S_0 and the schistosity, S_1. This always lies parallel to the F_1 fold axes. Mineral lineation (hornblende etc.) is less frequent than the elongation of mineral aggregates. Elongate segregations of quartz, for example, are common in mica schists and quartz schists, these usually occurring together with an S_0/S_1 intersection lineation. Stretched pebbles in the Harstad and Evenskjær conglomerate constitute another type of F_1 lineation. These are partly flattened parallel to the schistosity (Fig. 4) but frequently the extension is in only one direction, producing elongate rods. Locally, this elongation is much more pronounced in quartzitic than in granitic pebbles (Fig. 17). The orientation of the stretched pebbles is parallel to the S_0/S_1 intersection lineation within the conglomerate groundmass, as well as being parallel to local F_1 minor fold axes, and it is clearly a so-called *b*-lineation.

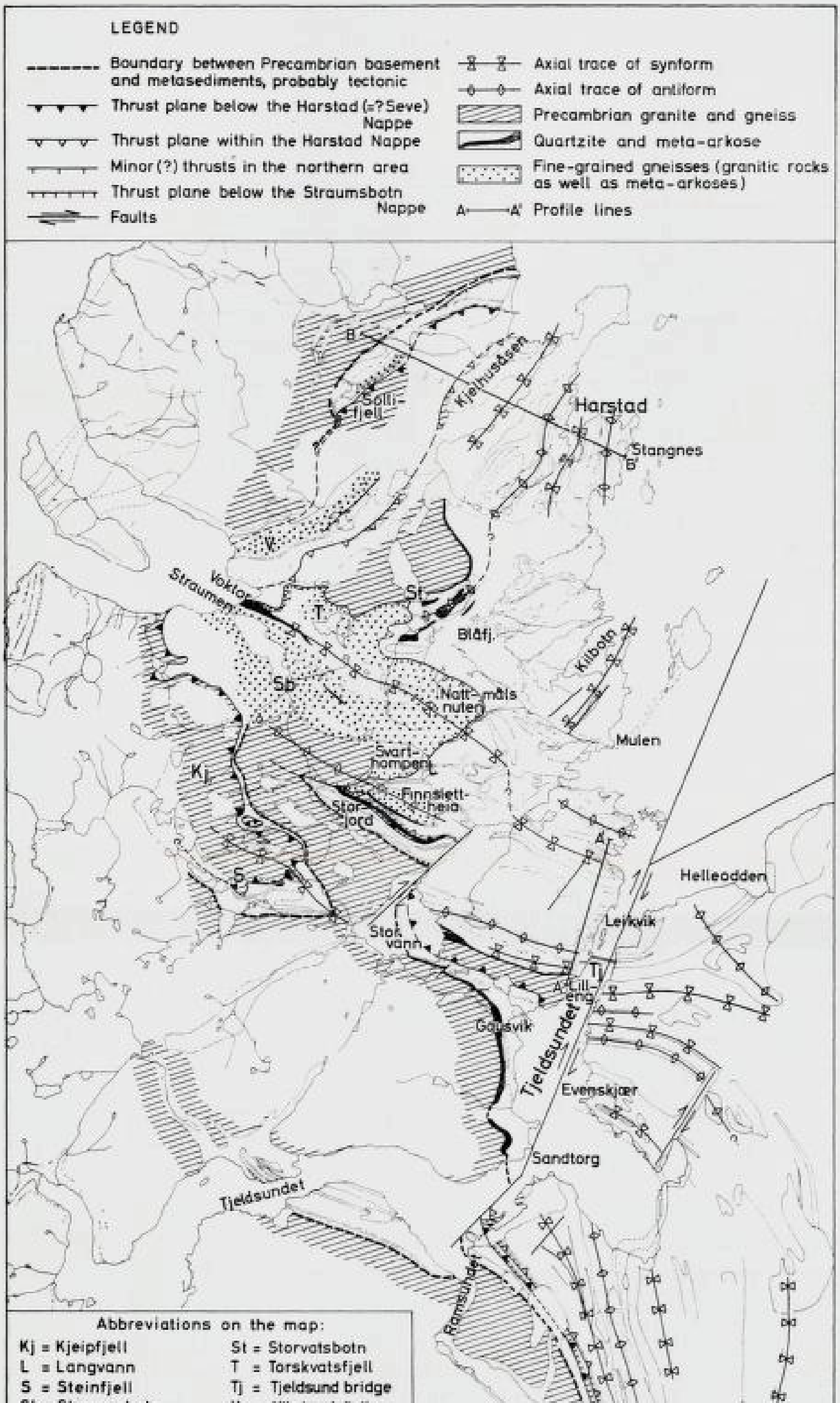


Fig. 19. Structural map of the Harstad–Tjeldsund area. Scale 1 : 250,000.

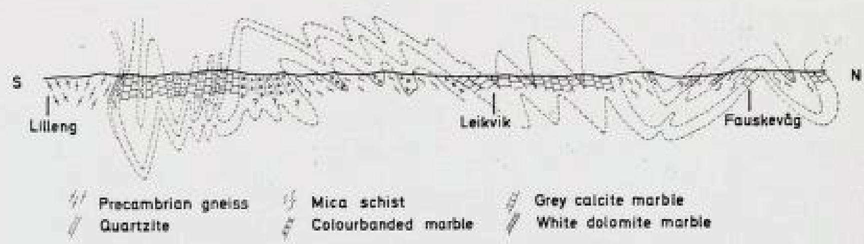


Fig. 20. Profile Lilleng–Fauskevåg (A–A' on map, Fig. 19). Length of profile about 5 kms.

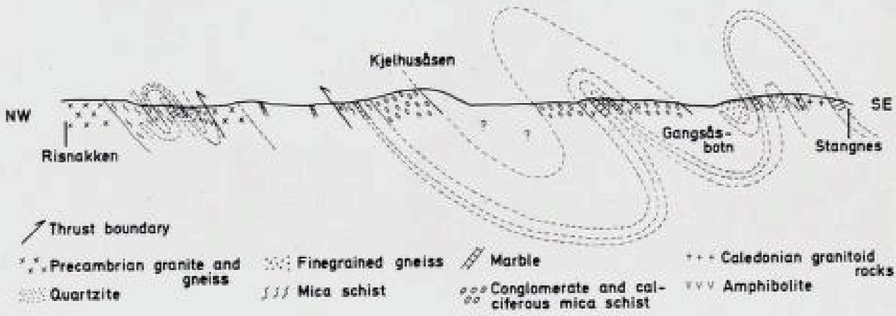


Fig. 21. Profile Risnakken–Stangnes (B–B' on map, Fig. 19). Length of profile about 10 km.

The Ofoten Synform

SHAPE AND EXTENSION

The large synformal fold structure occurring just south of Ofotfjorden was described by Foslie (1949) as the Håfjell syncline ('Håfjellsmulden') (Figs. 1, 29), a term adopted in later publications (Holtedahl 1953, Strand, in Holtedahl et al. 1960). As a result of the present work, this fold structure is now considered to be more complex than previously described. In view of this, and also as the synformal structure can be followed up to Salangen, it is here proposed that it be renamed the Ofoten Synform.

The axial trend of the Ofoten Synform is NE–SW, coinciding with the main trend of the mountain chain in this region. The synformal axis has variable but generally low-angle plunges to the NE or to the SW. Axial culminations are present SE of Grovfjord and in Lavangen, in addition to the major culminations along Efjord and Salangen which mark the southern and northern closures of the structure, respectively (map, Pl. I). The synform comprises the Narvik, Salangen, and Niingen metasedimentary groups. Along its eastern side its boundary approximately follows the lower boundary of the Narvik Group. In the area from Skjomen to Rombaken the Rombak Group also has a westerly sheet dip which is parallel with the general structure of the synform in this area. The western flank of the Ofoten Synform is less well-defined than its eastern part, this being due to the strong influence of EW or NW–SE trending fold structures towards Tjeldsundet and further north. The sheet dip is also flatter and more undulating than on the eastern limb. Only the Niingen Group

Fig. 22. F_1 folding of lime-silicate layer in mica schist, Treldal, Rombaken.

and the upper part of the Salangen Group have fairly regular easterly or south-easterly dips in this area. The northern limit of the synform is defined by a NW–SE trending antiformal structure at Salangen. North of this 'cross fold' the major regional structure generally appears to be dominated by E–W to NW–SE trending folds. Similarly, the southern end of the synform is determined by a culmination, the Tysfjord Culmination,[1] which exposes the granitic basement over a very wide area, and which also trends NW–SE in the areas immediately south of the Ofoten Synform (Efjord area). A simplified cross-section of the Ofoten Synform is shown in Pl. I.

STRUCTURAL HISTORY OF THE OFOTEN SYNFORM

As in the Harstad–Tjeldsund area, 3 episodes of folding can be recognized within the region covered by the Ofoten Synform. An early episode of strong plastic deformation contemporaneous with regional metamorphism was followed by two folding episodes of a less penetrative character. The synformal structure itself appears to have developed during the F_2 folding episode.

F_1 episode: The F_1 folds are frequently recumbent, sub-isoclinal structures and can be seen to affect all Caledonian rock-types except the younger group of igneous rocks. The folds are commonly of similar type, although style differences dependent on lithology can be observed (Figs. 22, 23). While axial planes generally dip at low angles, in certain areas steeper dips are clearly the result of later folding. It does, however, seem probable that most F_1 folds originally had a fairly flat-lying attitude. In the pelitic rocks an axial plane

[1] This term was used by Strand (1961) in his review of the Scandinavian Caledonides. Foslie, in his earlier descriptions, did not use the term explicitly.

Fig. 23. F_1 fold in impure quartzite, Salangen. Axial plane schistosity developed only in the pelitic intercalations.

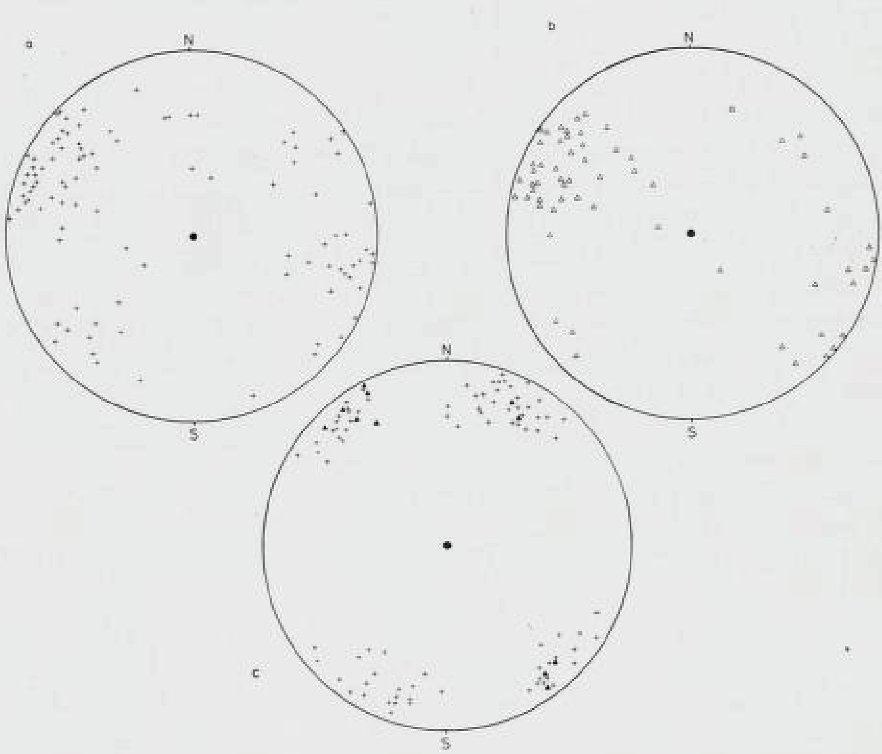


Fig. 24. Stereograms, Ofoten Synform. a) F_1 fold axes; b) F_1 lineations, c) Crosses: F_2 fold axes: filled triangles: F_2 lineations.

Fig. 25. F_2 fold in mica schist, Treldal, Rombaken. The axial planes are dipping steeply to the NW.

schistosity is always present and it is this which represents the regional foliation in the metasedimentary sequence.

All the observed F_1 folds are mesoscopic structures but there are also some indications of large-scale folding during this episode (see below). The orientation of F_1 fold axes is variable, as might be expected in an area of repeated folding, although axial trends about WNW–ESE or NW–SE are more common than other directions. F_1 axes about NE–SW have also been observed (Fig. 24). The observed sense of overturning is mostly to the SE for the NE trending folds, and to the NNE or SSW for the WNW trending folds.

Lineations of F_1 age are common in all lithologies. Several types of lineation are present, including mineral orientation, rodding, and S_0/S_1 intersection lineations. These are always parallel to the axes of associated minor F_1 folds and thus belong to the *b*-lineation category.

The age relationships of F_1 structures to those of other deformation episodes are demonstrated by the folding of the F_1 schistosity and lineations by F_2 and F_3 folds. Refolding of F_1 minor folds by folds of the younger episodes is less commonly observed.

F_2 episode: The F_2 minor folds are close to tight, slightly asymmetric folds, usually not recumbent (Figs. 25, 26). Disharmonic folding from one layer to the next is quite common. An axial planar schistosity is not usually developed and the F_1 schistosity can, as a rule, be followed around the fold hinges of the F_2 folds. Occasionally, small-scale faults occur parallel to the axial planes (Fig. 26).

F_2 fold axial orientation shows two main trends, NE–SW and NW–SE; in the southernmost areas E–W orientations also occur. The same holds for the

Fig. 26. F_2 folds in mica schist with amphibolitic and quartzitic layers, Herjangsholmen. Note small-scale axial plane faults and quartz segregations (partly along axial plane faults, partly along the folded schistosity planes).

F_2 lineations, which are less common than F_1 linear elements and are mostly restricted to small crenulations paralleling the F_2 minor fold axes.

On a larger scale, the dominant F_2 minor fold trends are reflected in the main trend of the Ofoten Synform and that of the open 'cross fold' structures which produced the culminations (map, Pl. I). The F_2 age of the latter structures is indicated by, for instance, the fact that the Salangen Antiform folds the post-F_1 thrust plane below the Narvik Group. It is, perhaps, less obvious that the Ofoten Synform itself is of F_2 age. Folding of post-F_1 thrust planes occur also along the east side of the synform but this deformation is not necessarily F_2 but might be ascribed to the F_3 episode. However, an F_2 age of the synformal structure might be inferred from the analogous orientation with F_2 minor folds. Furthermore, F_3 structures do not seem to have the regional importance and intensity necessary for the development of this large synform. Their orientation (see below) is also deviating considerably from the main trend of the synform.

F_3 episode: Some flexures and gentle folds (Fig. 27), of consistent 010°–015° trend, can be ascribed to a later episode of folding. These occur in the outermost western and eastern parts of the synform. Small-scale faulting has been observed in connection with some F_3 folds. Chlorite has developed along these faults and along tension cracks associated with this folding.

Refolding of F_1 structural elements by F_3 folds is commonly observed, but the relationship between F_2 and F_3 structures is less clear. Because the F_3 folds are gentle to open structures and have a consistent axial trend, they have been grouped into a late deformational episode. It is possible, however, that F_3 could be just a late phase of the F_2 deformation. In an area in the easternmost part

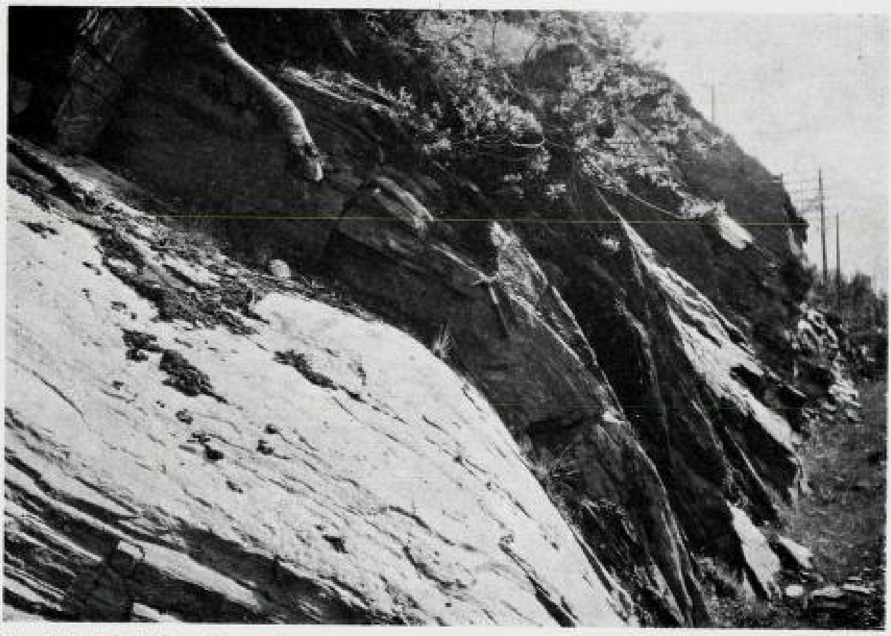


Fig. 27. F_3 fold in mica gneiss, Herjangen.

of the synform, for example, NNE oriented minor folds show similarity of style with the common F_2 folds. These various features would therefore suggest that no great time span separated the F_2 and F_3 deformations, and it would perhaps be equally correct to consider them as two phases of one fold episode as to give them the status of separate deformation episodes.

Major thrusts are situated below and within the Rombak Group, between the Rombak and Narvik Groups and probably between the Salangen and Niingen Groups. For reasons outlined below, a great deal of the thrusting is considered to be of post-F_1, pre-F_2 age:

a) Chloritization and quartz segregation phenomena associated with the thrust movements post-date the main, syn-F_1 metamorphism (Gustavson 1966, pp. 132–133) (Fig. 28).

b) Boundaries between areas of different metamorphic grade in part coincide with the supposed thrust surfaces (Gustavson 1966, metamorphic map, Fig. 2).

c) The thrust planes are deformed by F_2 folds (p. 37).

Some thrust movements, however, may already have taken place during the F_1 episode, and it is quite probable that local thrusting also occurred in later stages of the tectonic deformation, as has been demonstrated in areas further east (Mortensen, 1972). This, and other problems related to thrusting, will be discussed in a later chapter (p. 36).

MAJOR FOLDS

After discussing the main features of the F_1, F_2 and F_3 *minor folds* in this central area, it is important to consider the ways in which they are associated with the

Fig. 28. Tectonic quartz breccia, Neslia, Bardu. Note foliation (centre) and folding (below knife) in rotated gneiss fragments. Large fragments to the right are marble.

major folds related to the different deformation episodes. The Ofoten Synform itself is a complex major fold structure, and it has already been suggested that the bulk of its general development, with a westerly sheet dip on the eastern side and east- or southeasterly sheet dip on the west side, is of F_2 age. As will be shown below, however, there are indications that the development of major folds in this area was initiated during the F_1 deformation. A closer inspection of the local areas where the synform is more complex serves to illustrate this.

The Håfjell area: The geology of this area (Fig. 1) has been described by Foslie (1949, 1 : 50,000 map printed 1930, 1 : 12,500 map of part of the area, 1929) and Juve (1967). At first glance at the printed maps, an impression is gained of a simple and regular synformal structure (Fig. 29). However, as mentioned by Foslie and later confirmed by Juve, the uppermost part of the sequence is folded into a 'double syncline' with a minor anticline in the centre. This double fold is overturned towards the NW with its axial plane dipping southeast at about 40°, and is of a relatively open character (see Juve 1967, p. 12). Juve stated that two distinctive fold episodes could be distinguished, one with fold axes parallel with the main trend and represented by the 'double syncline', the other episode present as a weaker 'cross folding', almost perpendicular to the main axis. No mention was made of the age relationship of these two fold episodes, but in the present author's opinion both may be of F_2 age. The 'cross folds' are gentle NW–SE trending folds paralleling a series of synforms and antiforms which occur at several places along the length of the main synform, and which undoubtedly are F_2 (see discussion below concerning

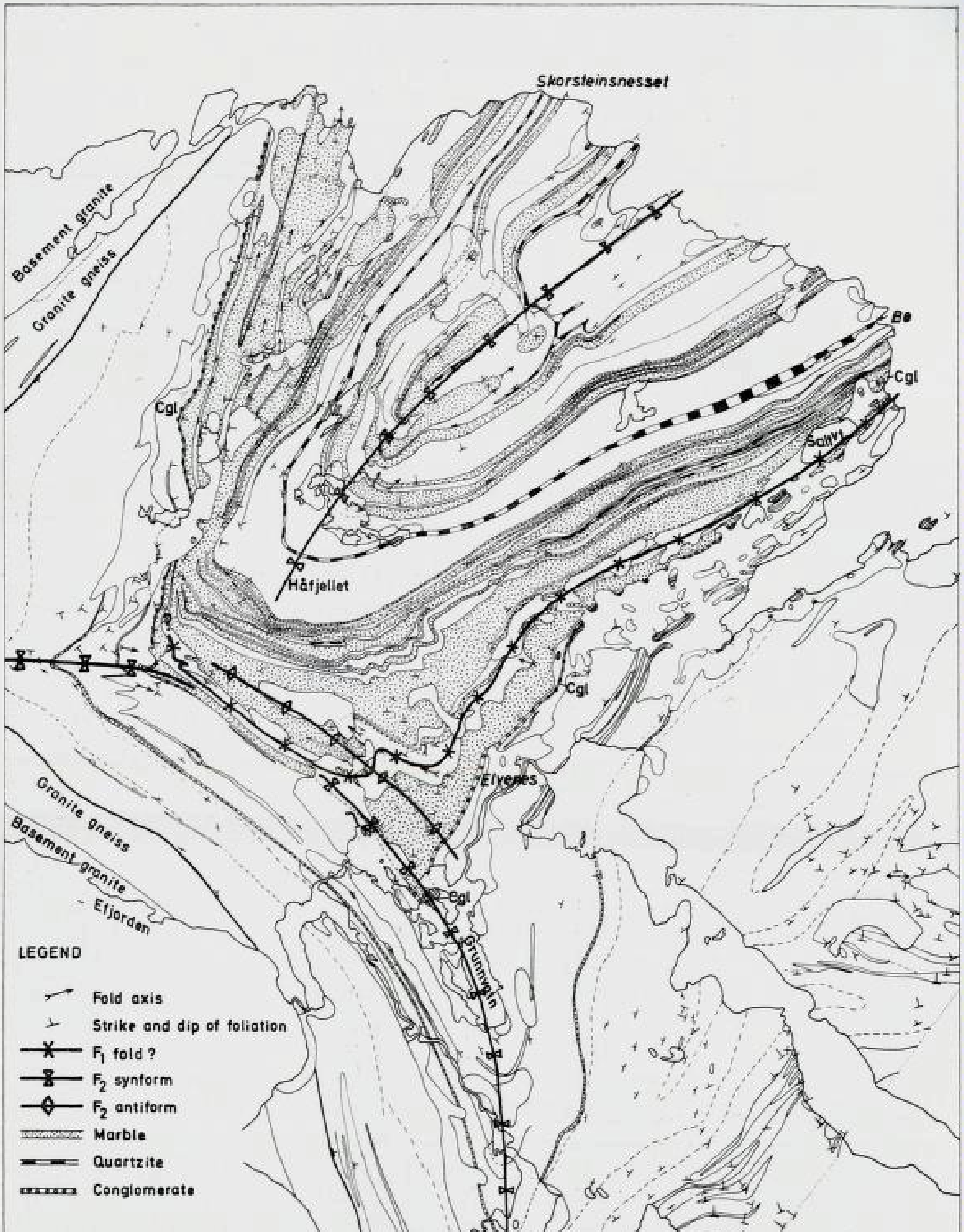


Fig. 29. Structural map of the Håfjell area. Main features of geology from Foslie (1949), structural interpretations partly by the present author. Scale about 1 : 120,000.

the Salangen Antiform, and also p. 44). The NE–SW trending double fold appears to refold an older schistosity, and it can also be demonstrated that an older set of minor folds and lineations are refolded about this 'main' fold axis. In an impure quartzite at Skorsteinsnesset, for example, two directions of folding and associated lineations can be observed. A marked S_0/S_1 intersection lineation and minor fold axes are plunging at 10°–15° to the south, while a younger lineation parallels the main axis and plunges at 15° to the northeast. Rotation of the older lineation into a horizontal plane gives an approximately NW–SE trend, possibly the original trend of these structures. If it is assumed that the folds parallel to the synform axis are F_2, then the older set of lineations and minor folds are clearly F_1 structures. The original trend of F_1 folds in this area is thus more or less parallel to that of the 'cross folds' formed during the F_2 episode.

A closer inspection of Foslie's maps, the main results of which are compiled in Fig. 29, shows that complications to the 'synclinal' structure may also occur in lower parts of the sequence. The Ballangen marbles, which consist of calcite marbles, at least 4 layers of dolomite marble, a number of mica schist horizons, and two quartzite layers (Foslie 1949), lie tectonically above the Elvenes conglomerate. In the summer of 1970 the author found a conglomerate layer, quite similar to the Elvenes conglomerate, NE of Saltvann, in a position higher up in the sequence than previously described (Fig. 29). The sequence, therefore, involves repetitions of layers of identical lithology, repetitions that have possibly been caused by folding.

The three most common pebble types in the Elvenes conglomerate are: (1) meta-trondhjemite, (2) quartzite, and (3) dolomite marble. Of these, (1) occurs mainly, though not exclusively, in a tectonic position below the conglomerate horizon, while types (2) and (3) are common in the sequence *above* the conglomerate, and absent below, within this area. Rosy-coloured marble, which occurs as pebbles in the Harstad and Evenskjær conglomerates (see p. 11), has not been reported from the Elvenes conglomerate zone. It is present, however, in the sequence tectonically above the Elvenes conglomerate as layers within the 'Ballangen marbles'. The pebble material thus favours an inversion of the stratigraphy in this part of the sequence. Tectonic repetition is also suggested by another feature visible on the map – the two thicker calcite marbles, separated by a mica schist layer, run together SW of Håfjell. This map pattern looks very much like the closure of a large fold, although the thin apex by Foslie, (op. cit.) is drawn into connection with marble layers on the west side of the synform (Fig. 29). A continuation of this structure on the west side of the synform is possible as a mica schist zone within marble there seems to indicate isoclinal folding. It would therefore seem possible that repetitions are present which are the result of large-scale, isoclinal folding. These folds show no relationship with the F_2 'double syncline' in the uppermost part of the sequence, but since they resemble F_1 minor folds in terms of style, they are considered to be of F_1 age.

The probable F_1 major folds did not, however, affect the whole sequence.

The regular trend and lateral extent of the Balteskar (Bø) quartzite (Gustavson 1966, p. 100), for example, preclude any major regional folding; nor is there any evidence of large-scale folding within the Narvik Group, below the Ballangen marbles. Major isoclinal folding would thus appear to have affected only that part of the sequence containing the thick marble horizons.

Except for the NE-trending Ofoten Synform, the most conspicuous F_2 structures in this area are the Grunnvatn Synform and its complementary antiform south of the Ofoten Synform (Fig. 29). The Grunnvatn Synform has a N–S trend in its southernmost part, this gradually becoming NW–SE south of Håfjell. The antiform north of the Grunnvatn Synform is less well-defined and is absent in the south. It is of interest to note that these F_2 structures follow the general boundary to the basement SW of the area. As mentioned earlier (p. 27) the NW-trending basement culmination is also considered to be largely of F_2 age (see also discussion on p. 44). A similar parallelism of basement boundaries and major F_2 structures has previously been noted (p. 16) in the Harstad–Tjeldsund area.

The Bogen Area: (Fig. 1). Mapping and interpretations by Vogt (1941) have shown that repetitions of metasedimentary layers in this area are probably due to large-scale folding. Vogt's profiles (op. cit. Fig. 5) show folds varying from open to relatively tight. Judging from the occurrence of minor folds in the area, however, similar-type, sub-isoclinal folding appears to be more common, and as an associated axial plane schistosity is present these folds are probably of F_1 age. The axial directions are not mentioned by Vogt and are not obvious from his map. It seems clear, however, that they are not parallel to the general synform direction. The extension of these folds towards the northeast is only a few kilometres (Vogt, op. cit. Fig. 4). Towards the northeast up to Grovfjord, marker horizons remain parallel so that fold closures appear to be absent; this is perhaps because the actual closures observed at Bogen occur below the surface, and have horizontal axes along the east side of the synform. In the north, parallelism of marker horizons continues over to the west side of the synform.

The area from Bogen to Salangen: Two major NW-trending axial culminations are present at Grovfjord and Lavangen. Their general character and their similarity and parallelism with the Salangen Antiform (see below) suggest an F_2 age for these structures.

The Salangen Area: In this area the NW–SE trending Salangen Antiform is transverse to the Ofoten Synform (p. 23, also Pl. I). The F_2 age of this structure is shown by the fact that the late- or post-F_1 thrust plane below the Narvik Group is folded across the antiform. The antiform is fairly symmetrical with a vertical axial plane and a subhorizontal fold axis.

Structures of eastern areas

Parts of the area east of the Ofoten Synform, mainly the Salangsdalen and Bardu districts, have been mapped during the present investigation; in areas further east, only reconnaissance mapping has been carried out by the author. Information is available, however, from older and recent literature, and a brief summary of the geology is presented here as a background for the discussion in the next chapter.

The Precambrian basement below autochthonous Cambrian and overthrust Caledonian rocks is exposed in a number of tectonic windows (see Fig. 1 in Gustavson, 1966). Details of the Mauken window have been published by Berthelsen (1967), Landmark (1967), and Gustavson (1963, 1966) and of the Divielva window by Vogt (1918), Gustavson (1963, 1966), and Berthelsen (1967). The most interesting structural feature of these windows is their updoming in Caledonian times prior to the overthrusting of the metamorphic rocks, as mentioned briefly by Vogt (1918) and later described by Gustavson (1963) and Berthelsen (1967). Autochthonous Cambrian rocks of the Hyolithus Zone, mainly shales and sandstones, occur above the Precambrian basement around most of the windows. Because of the doming up of basement + autochthonous cover and later overthrusting of the Caledonian nappes, only small remnants are left of the Hyolithus Zone sediments on top of the basement domes, while a thicker and better preserved sequence has partly been found on the 'lee' side of the domes.

The rocks above the Hyolithus Zone were all folded, overthrust, and more or less metamorphosed in Caledonian times. This has been agreed by all workers in these areas. Different opinions arise chiefly when structural subdivisions and the regional importance of thrust planes are brought into the discussion. Some of these problems will be discussed in the next chapter. There is general agreement, however, that thrust tectonics play an important part in the structural geology of these areas. Although extensive deformation by superposed folding has affected the metamorphic rocks, they are, *on a regional scale,* flat-lying or rather gently inclined. This is what Vogt (1941) termed 'the Bardu style' of regional tectonics. Locally, however, bedding and schistosity planes show highly variable dips. The sequence of fold episodes has been described by Kalsbeek & Olesen (1967) for the area between Målselv and Altevann and by Mortensen (1972) for the area between Altevann and the Swedish border. (A paper by Olesen (1971), dealing in some detail with part of the area south of Målselv, arrived when the present paper was almost ready for print. It has therefore only to a minor extent been taken into consideration in the following discussion.)

Kalsbeek & Olesen (op. cit.) subdivided the metamorphic rocks into three 'sequences' which have a fairly similar metamorphic history; this, at least, is the case for the lower two, which constitute most of the succession. The most conspicuous lineation and fold axes trend WNW–ESE. These folds are, in part, large-scale isoclinal structures, and they correspond with D_1 of Mortensen (op. cit.). Although a schistosity has been formed parallel to the axial planes of the isoclinal folds, they also appear to fold an older schistosity. According to Mortensen (op. cit.), the axial plane schistosity of D_1 folds is the regional schistosity in the area south of Altevann. Kalsbeek & Olesen (op. cit.) mention

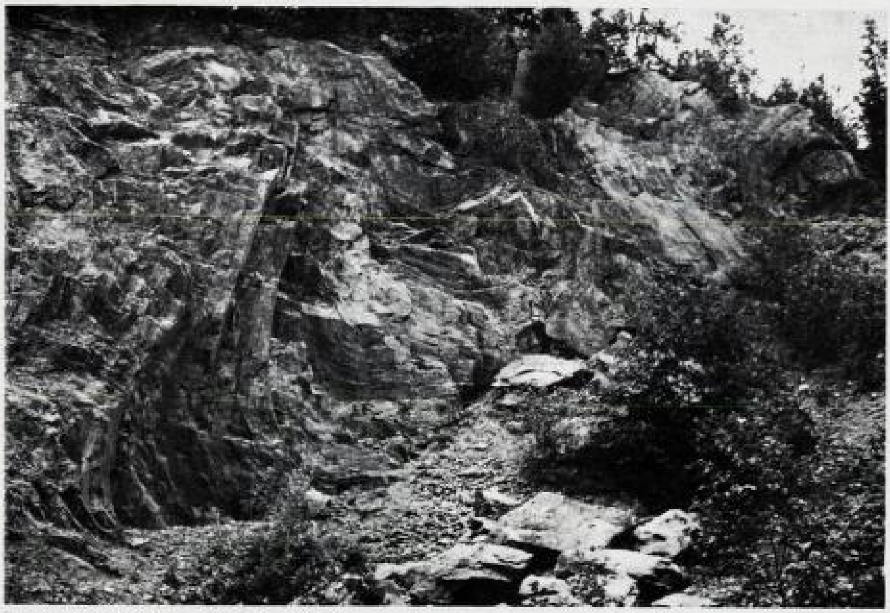


Fig. 30. NW–SE trending F_2 fold in quartzite. Small quarry north of Setermoen.

that folds older than the large-scale recumbent structures occur locally. These folds, too, are isoclinal and have an associated axial plane schistosity.

Folds younger than those with WNW–ESE axes are common in the easternmost areas. They fold the schistosity related to the WNW-trending folds and are usually minor folds, ranging up to some tens of metres in amplitude. Kalsbeek & Olesen (op. cit. p. 260) subdivided these later folds into three groups with axial trends in N–S, NW–SE and NE–SW directions. Characteristically they have subhorizontal and subvertical limbs, and the same style of D_2 folds is described from south of Altevann by Mortensen (op. cit.).

Lineations in these eastern areas are generally of the same age as the important WNW-trending folds. Kalsbeek & Olesen (op. cit.) concluded that lineations (frequently a mineral lineation, occasionally a conglomerate pebble elongation) are *b*-lineations which are older than the thrust movements, a view with which the present author agrees. As to the age of thrusting, it has been stated (Gustavson 1966, p. 16 and pp. 154–155) that thrust movements post-dated the progressive phase of regional metamorphism and the greater part of the fold deformation. This, at least, must be the case with the latest thrust movements, and this is also the view expressed by Kalsbeek & Olesen (op. cit., p. 259): 'It is evident that the metamorphism of the rocks must have taken place before they were thrust on top of the non-metamorphic Hyolithus zone sediments, and a lineation due to the orientation of metamorphic minerals must therefore be older than these thrust movements.' As mentioned above, this lineation is parallel and coeval with an important fold episode with WNW-trending axis, which thus pre-dates the main part of the overthrusting. Mortensen (op. cit.) finds that thrusting probably occurred more than once in the area south of Altevann, for instance in association with his D_1 and D_2 episodes, while the last movements along the main thrust plane post-dated the D_2 folding.

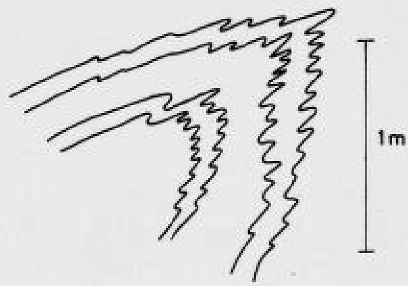


Fig. 31. Sketch of minor F_2 fold in quartz schist, east of Setermoen.

Structures in the *Bardu and Salangsdalen areas* (Fig. 1), mapped during the recent investigation, do not differ much either from those within the Ofoten Synform or from those described above from more easterly areas. It has already been mentioned that generally flat-lying sequences are common in areas east of the Ofoten Synform, and this can largely be ascribed to the presence of *recumbent folds* of smaller or greater size. Such folds are exposed in the steep hillsides east of Sørdalen and Østerdalen in the Bardu area (Pl. 1). As these sections are practically inaccessible, details and precise orientations of the folds cannot be given, but it does seem likely that these structures are parallel and also contemporaneous with minor recumbent early folds in this same area. The orientations of these minor folds fall into two groups, one NE-SW and the other NW-SE. No age difference has been detected between the two groups of early folds, and as they are both of the same style and carry the same schistosity, it seems quite probable that they are contemporaneous. They clearly correspond to the F_1 episode of the central and western areas and the main fold episode (D_1) of eastern areas.

Lineations are conspicuous in several rock-types and frequently oriented about NW–SE or WNW–ESE, parallel to one group of early folds. The lineations are therefore probably *b*-lineations, similar to those occurring in more westerly areas (p. 26).

The subhorizontal F_1 structures are in several places refolded by the younger F_2 episode of folds. The F_2 folds are usually minor folds, only a few being of mappable size, and axial planes, in contrast to those of the F_1 folds, are relatively steep (Fig. 30). Quite commonly the F_2 fold limbs are subhorizontal and subvertical (Fig. 31), much the same as in areas further east described by Mortensen (1972). An axial plane schistosity is generally absent, while the earlier F_1 schistosity is deformed around the F_2 fold closures. The orientation of F_2 folds varies considerably but a detailed examination has not been carried out.

Folds younger than those described above have not been detected. The age and influence of thrusting in these areas are referred to the general discussion in the pages which follow.

The regional extension and importance of thrust planes

THE THRUST PLANE ABOVE THE HYOLITHUS ZONE

There seems to be general agreement between students of Eastern Troms geology that a thrsut plane or thrust zone is located above the autochthonous Hyolithus Zone (Vogt 1918 and 1967, Gustavson 1963, 1966, Kalsbeek & Olesen 1967, Berthelsen 1967, Mortensen 1972). The rocks immediately above this thrust plane are, in part, mylonites and granite kakirites, and in part low-grade quartz schists (frequently of the 'hard schist' type) and dolomites. It has been shown (Gustavson 1966, Mortensen 1970) that a great part of the schists are metasediments. Similar rock sequences have been described from the Torneträsk area by Kulling (1964) as forming the main part of his 'lower thrust rocks', the Rautas thrust complex, and also the 'middle thrust rocks', the Abisko complex. Recent work by Mortensen (1972) south of Altevann has shown that the lowermost part of the allochthonous metasediments in that area ('Girunas schists') most probably belongs to the Abisko complex. The 'Girunas schists' were correlated with similar formations north of Altevann, mapped by Kalsbeek & Olesen, and which roughly correspond to the author's 'Storfjell Group' (Gustavson 1966) of feldspathic quartz schists and dolomite. It is therefore probable that the Rautas thrust complex is absent in Eastern Troms, while the lowermost allochthonous rocks correspond to the Abisko complex. Kulling (1964) placed his 'main overthrust' below this unit, and the thrust below the corresponding unit and directly above the Hyolithus Zone in Troms should accordingly represent a major thrust boundary. This is in accordance with the opinions of Vogt (1918), Berthelsen (1967), and Kalsbeek & Olesen (1967). In the author's first contribution to the geology of these allochthonous rocks (Gustavson 1966), some reservation was voiced concerning the structural relations of the lowermost rocks (the Storfjell Group) in remarking (op. cit., p. 16) that they 'may be truly allochthonous but it is also possible to interpret the Group as parautochthonous'. Work published subsequently (Vogt 1967, Kalsbeek & Olesen 1967) seems to have settled this matter as it has demonstrated the presence in the most easterly areas of masses of granite kakirites, syenite mylonites, and similar rocks. These are obviously comparable to Precambrian mylonitic rocks at the base of the Abisko complex (Kulling, 1964). The boundary above the Hyolithus zone therefore has the characteristics of a real *overthrust* – strong tectonic disturbance of the rocks close to the thrust plane, metamorphic rocks overlying non-metamorphic sediments, and partly with older rocks lying on top of stratigraphically younger ones.

This *basal thrust* along the eastern nappe front can also be recognized around a number of tectonic windows occurring several tens of kilometres to the north-west. In some of these cases, thin Hyolithus Zone sediments are preserved below the thrust plane (Gustavson 1966, pp. 46–48), while in other instances the thrust nappe rests directly on the Precambrian basement.

Turning to the area west of the Rombak window, the basal thrust is also there in direct contact with the basement, and some deformation of the

Precambrian granite is evident. According to Vogt (1941, 1950) Cambrian basal conglomerate and sandstone occur below the thrust north of the window. *Within* the Rombak window, basal sandstone and conglomerate occur together with mica schist of the so-called Rombak Group in small, downfolded areas trending N–S or NE–SW. The thrust plane between the mica schist and the mechanically metamorphosed sediments below was clearly folded after the emplacement of the nappe, and the intensity and style of this fold deformation are suggestive of it being of F_2 age. This therefore means that the basal thrust can be dated to the period between the F_1 and F_2 fold episodes, as it is clearly post-dating the regional metamorphism and F_1 regional schistosity.

Nothing definite can be said about the continuation of this thrust plane from the Rombak area to the south and west. In these areas, non-metamorphic autochthonous rocks are not preserved and it is therefore difficult to evaluate the importance of thrusting. In the Tjeldsund area (p. 5) the basal sediments are clearly metamorphosed. Some movement almost certainly occurred at the boundary between the Precambrian and these overlying metasediments. It is quite possible that this level represents the continuation of the basal thrust plane, but that thrusting in this area occurred without disturbing the original relation between basement and cover.

THE 'SEVE' THRUST

As noted above, it seems possible to correlate the lowermost thrust boundary in Troms with the 'main overthrust' of the Torneträsk area, below the Abisko nappe. In the latter area (Kulling 1964) the top of the Abisko Nappe is defined by the tectonic lower boundary of the so-called 'Seve-Køli Complex'. According to Kulling (op. cit. p. 140) the main features of this complex are: 'Their crystalline schists (sediments) and basic eruptives are characterized by the presence of the metamorphic mineral garnet and the rocks are mainly in epidote-amphibolite facies.' The main rock-types are marbles, mica schists and gneisses of various compositions and appearances, amphibolites, and greenstones. As the Abisko nappe is also characterized by 'schists' there is no great difference in metamorphic grade between the Abisko and the Seve-Køli complexes (Kulling op. cit., p. 23). In some areas, however, the tectonic boundary is marked by angular discordances between the two complexes, for example in areas north of Torneträsk, and there can be no doubt that the thrust at the base of the Seve-Køli complex is one of regional importance.

The area immediately north of the border has been investigated by Mortensen (1972) who stated that he had found no evidence of a thrust at this level. The boundary is, however, relatively sharp as according to Mortensen (in translation), 'The rocks change character within a zone of about 2 metres thickness.' It is noted that the mica schists above this level are coarser than the quartz-mica schists below and, moreover, contain garnet megacrysts. To the present author it seems evident that the rock chemistry within the lowermost schists (Girunas schists), at least in some layers, must have been suitable for the formation of garnet and that the occurrence of this mineral exclusively *above* this actual level is indicative of a real increase in the metamorphic grade. The increased grain-size points in the same direction, as does the common occurrence of chlorite in the hard schists of the lowermost unit since chlorite within the Seve-Køli rocks occurs mainly retrogressively as pseudomorphs after garnet. From a more northerly area, SE of Rostadalen, Mortensen (1970) describes large tectonic lenses or plates of garnet-mica schist within the hard schist formation. To the present author, the above-mentioned mineralogical and tectonic relations together with the important regional analogy with the Torneträsk area point to the probability of a thrust plane at this level. This was indicated in the author's 1966 paper (Gustavson 1966, Table III) where a proposed thrust plane was drawn between the Storfjell Group and the assumed equivalents of parts of the Rombak Group above.

From the areas between Altevann and Målselv (Fig. 1), Kalsbeek & Olesen (1967, p. 258) have reported that, 'There are several indications that intensive thrusting has taken place at many places in the lower sequence, such as the common occurrence of tectonic discordances and mylonite zones, and the local occurrence of granite kakirites (thrust) on top of meta-sedimentary rocks (marbles and normal mica schists). Furthermore, if normal low-grade sediments do occur in the formation 2[1] of the Jerta stratigraphical section, this might indicate the overthrusted position of the more high-grade rocks which overlie them . . .'. Although these same authors state that a thrust plane 'is only rarely evident' in the field, their investigations clearly support the general importance of thrust tectonics in this area. There is also no reason to doubt that low-grade sediments occur in 'formation 2' (broadly the author's Storfjell Group); dolomite and arkosic rocks are clearly present (Gustavson 1966, pp. 51–61) and studies by Mortensen (1970) strongly suggest that the typical 'hard schists' are also metasedimentary rocks.

Evidence from these eastern areas thus shows that the Seve thrust also extends into this part of the mountain chain. The problems to be discussed are, firstly, whether thrusting occurred on one single thrust plane or within a thrust zone, and secondly, the distance of the nappe movements.

Concerning the more westerly districts, Kulling (1964) discussed the possible connections between the Torneträsk area and the areas west of the Rombak window. He compared the lowermost parts of the Rombak schists with the 'hard schists' of the Abisko nappe. Although the present author considers that typical 'hard schists' are not present within the Rombak Group, the correlation certainly finds some support in the general lithology of the sequence. Within the Rombak profile the lowermost part of the allochthonous sequence contains quartz-rich mica schists and, in part, feldspar-bearing layers (see, for example, analysis 1 of Table VI in Gustavson, 1966), which have their counterparts within the Abisko nappe.

A Precambrian granite slice (there are in fact more than one) within the Rombak Group was correlated by Kulling with the big plate of overthrust

[1] Termed low-grade schists, partly 'Hartschiefer'. (Present author's note.)

Precambrian rocks at the base of his Abisko nappe, a correlation which is somewhat difficult to accept. In the Torneträsk area the Precambrian allochthonous rocks are overlain by the hard schist formation, while in the Rombak area the Precambrian granite is on top of the schist sequence which Kulling correlates with the hard schist formation. Moreover, there is no evidence of inversion, due to folding, after the thrusting. It would appear more probable that the granite slices represent the base of *a thrust nappe overlying the supposed equivalents of the Abisko nappe.*

Based on Foslie's map sheets Tysfjord (1941) Hellemobotn and Linnajavrre (1942) and the papers by Kautsky (1946, 1947, 1953), Oftedahl (1966) considered that the granite gneiss horizons within the lower parts of the metasedimentary sequence in the areas southwest of the Rombak window form the base of the Seve thrust nappe. The present author concurs with this conclusion and also includes the granite plates within the Rombak area in this interpretation. For reasons already outlined (Gustavson 1966, p. 26), a Precambrian age for these tectonized granites must be assumed and a thrust plane, or rather, a thrust zone, is obviously present. Regional analogies make it highly probable that this is the base of the great Seve nappe, but a detailed description of the continuation of the nappe boundary to the south and west cannot be given at present. In general, however, it can be said to occur either directly on top of the massive Tysfjord granite, or with a relatively thin zone of mica schist, quartz-mica schist, and quartzite between the Tysfjord granite and the nappe base. In most parts of this area, granitic gneisses form the base of the nappe, evidently representing allochthonous Precambrian basement rocks. If this interpretation is correct it means that, within the Tysfjord area, the Seve thrust and the basal thrust plane are in part coincident, in part separated by a very thin zone of metamorphosed sediments (map, Pl. I).

West of the Ofoten Synform, structural relationships have been investigated by the present author during the years 1968–71. From Ofotfjorden to Tjeldsundet fine-grained granitic gneisses are found in the thrust zone as several thin horizons paralleling the regional schistosity. A zone, consisting of mica schist and marble layers, separates these gneisses at the base of the Seve Nappe from the basal thrust on top of the autochthonous basement. Relations west of Tjeldsundet have been described earlier (pp. 12–15), where it was argued that the base of the Harstad Nappe in these areas most probably corresponds to the Seve thrust.

Nowhere in the area from the Rombak window to Harstad has there been recognized any important difference in metamorphic grade between the rocks immediately above and those below the Seve thrust. In a greater part of the area the granitic rocks at the base of the Seve nappe are fine-grained foliated rocks (see Fig. 11 in Gustavson 1966). This foliation is clearly not a result of mechanical deformation alone, but has formed during the same phase of regional metamorphism which affected the adjacent schists. This is in accordance with Kautsky's (1946, 1947) view, suggesting a phase of metamorphism and deformation post-dating the Seve thrust. As noted earlier, however, in parts of Eastern Troms there is a minor difference in metamorphic grade between rocks of the Abisko and Seve nappe complexes. These somewhat conflicting features may be explained in one of two ways:

1. Movements on the Seve thrust plane occurred at different times within different parts of the area; alternatively, repeated movements occurred at this level in Eastern Troms.
2. The thrust may be of F_1 age – more or less coeval with metamorphism in the central and western areas, while the peak of metamorphism was already passed at the time of thrusting in the peripheral eastern areas.

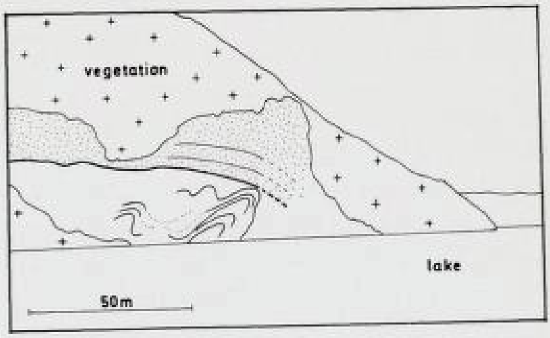
Accepting the latter alternative implies that the Seve thrust is somewhat earlier than the thrust below the Abisko complex, which is clearly post-F_1. In this connection, Mortensen's (1972) investigations in the area south of Altevann indicate that movements *within* the Seve thrust complex occurred in part during the F_1 (D_1) episode, but the same writer also suggested that some thrust movements within the Seve complex are clearly younger, perhaps F_2. The thrusting history in that area might therefore be taken as support for alternative 1 above, as it shows that movements may have occurred more than once.

Even if the thrusting of the Seve nappe was a more or less *en bloc* movement, the relationships to the metamorphic processes may vary somewhat if the metamorphism was not completely synchronous within the region as a whole. Similar diachronous metamorphic features have been reported, for example, from the Trondheim region (Roberts et al. 1970). If the termination of metamorphism was earlier in eastern areas than in the central area and the emplacement of the Seve nappe is assumed to be a syn- or late-F_1 incident, this may explain the relations at the Seve thrust boundary in different parts of the area. In the author's opinion, this explanation is preferable, although absolutely decisive criteria are lacking. What does seem evident at present is that movements on the Seve thrust plane must have occurred at an early stage, syn–or–late–F_1, in the central parts of the area.

THE THRUST PLANE BELOW THE NARVIK GROUP

Above the Seve thrust plane there is no obvious tectonic disturbance before the lower boundary of the Narvik Group is reached. Kulling (1964) proposed a thrust plane, according to him equivalent to the Seve thrust, below the Rombak marble, a suggestion based on the tentative correlation of the Rombak marble with the Nuolja marble of the Torneträsk area. There is, however, no visible sign of tectonic movement at this level (the lower boundary of the Rombak marble is now exposed in new road-cuts east of Treldal), nor is there any reason other than lithological to assume the presence of a major disturbance. Evidence suggests that the lower boundary of the Narvik Group is of a tectonic nature (Gustavson op. cit., pp. 69–70). This is shown by local structural discordances, as for example in the Salangen area (Fig. 32). The boundary is also a metamorphic facies boundary (Gustavson 1966, Fig. 2) along the east side of the Ofoten Synform and as far east as to the Bardu valley,

Fig. 32. Sketch (after photograph) of boundary between strongly folded Rombak Group (below) and Narvik Group at Øvervann, Salangen.



with the higher grade rocks above the boundary. This indicates that the thrust movement is later than the regional metamorphism, i. e., it is post-F_1, although some retrogressive mineral reactions, mainly chloritization of ferromagnesium minerals, also occur along the thrust plane. The regional importance of this thrust is difficult to evaluate. West of the Ofoten Synform, typical Narvik Group rocks are absent, except for a smaller area along the east side of Tjeldsundet, and a thrust plane equivalent to that on the eastern limb of the synform cannot be detected. The relations on the eastern limb, however, appear to indicate that considerable movement occurred at this level.

THE THRUST PLANE (?) BELOW THE NIINGEN GROUP

A thrust plane below the Niingen Group (Pl. I) is indicated by the occurrence within the group of the same type of metasediments and metamorphosed igneous rocks (serpentinites, amphibolites, granitoid rocks) as are present in the Narvik Group, suggesting a large-scale repetition of the sequence. The metamorphic grade within the Niingen Group also appears to be higher than in the metasediments below the boundary; this conclusion, however, is based on comparatively few field observations, and critical structural features have not been observed. Local tectonization phenomena below the boundary in the Gratangen area are no more extensive than those seen locally in other parts of the sequence.

Possible extensions of the Niingen Group to the east and west are not known. The Straumsbotn Nappe (p. 14) in the Harstad area is thought to be only a local nappe, and other allochthonous units above the Seve nappe have not been recognized in that area. In eastern Troms Kalsbeek & Olesen (1967) mapped major thrust planes below their 'middle sequence' and 'upper sequence', respectively. Olesen (1971) however, groups the middle and upper sequence together as one tectonic unit, the 'Upper Nappe', although he states that its lowermost part may be a separate tectonic unit. The rocks of the central and western areas are in direct field connection with the sequence below the Upper Nappe of Olesen (op. cit.). Most probably this Upper Nappe has no direct structural counterpart in the central area. Alternatively the Niingen Group might correspond to it but the rocks are not exactly similar as marbles are quite significant in this eastern nappe unit and absent in the Niingen Group.

DISTANCE OF NAPPE TRANSPORT

An estimation of the distance of nappe movement can be made only for the basal thrust, and in this case only a minimum figure can be given. As unmetamorphosed Hyolithus Zone sediments occur around the Precambrian windows of Mauken (Målselv) and Straumsli (Bardu) (Pl. I), the distance from the northwest side of these windows to the thrust front represents a minimum distance of thrusting. A consideration based on the Straumsli window gives a figure of ca. 50 km, while Berthelsen (1967, p. 63) has estimated the distance of travel from the Mauken window to be more than 70 km. This would therefore seem to be a minimum figure for movement along the basal thrust plane.

A more exact measurement of nappe transport on the basal thrust plane or for higher thrust units is not possible at present. Some estimates have been made for movement on the Seve thrust; Kautsky (1946, p. 594), for example, considered that the roots of this nappe must have been west of Lofoten and the thrust distance accordingly 250 to 300 km, perhaps more. This view was supported by Oftedahl (1966, p. 238). Strand (1961, p. 172) has suggested movement distances of 200 km or more for the major Caledonian nappes. Nicholson & Rutland (1969, p. 82), however, are of another opinion – 'It therefore seems possible that the original site of deposition of the Seve-Køli nappes was in Nordland and not further west.'

In the Southern Troms and Ofoten areas the sequence is metamorphosed to its base south and west of the Ofoten Synform. This would appear to support the view that the nappes could have originated within more or less the same areas in which they now occur. The extension and continuity of allochthonous basement gneiss at the base of the Seve nappe, however, are difficult to explain in this way. Generally speaking, these slices of basement indicate horizontal displacements of a magnitude which would imply that the nappe roots were outside the Ofoten and Southern Troms area. The various relationships are thus best explained by accepting Kautsky's theory of large-scale thrusting from areas northwest of the Precambrian of Lofoten–Vesterålen.

Behaviour of the Precambrian basement during the Caledonian orogeny

Movements in the Precambrian basement during the Caledonian orogeny and the relations of these movements to other tectonic events have been discussed briefly by several authors. These concern the Precambrian tectonic windows in Eastern Troms (Vogt 1918, Gustavson 1963, Berthelsen 1967, Landmark 1967) and the Rombak window (Vogt 1941, Lindström 1958a). Discussion of the Tysfjord granite and the basement of western areas has been concerned mainly with the question of Caledonian mobilization or, alternatively, the presence of a largely unaltered Precambrian basement in these areas.

An important feature in any discussion of the basement is the actual direction and nature of the movements. From the previous description (p. 6) it is clear that extensive mobilization in the upper parts of the basement (parts which are now exposed) did not occur. This does not, of course, exclude the possibility that mobilization took place at deeper levels or in more westerly areas prior to nappe emplacement (see the discussion, in Gustavson 1969, of the origin of some granitoid rocks within the Caledonian metasedimentary sequence). In general, the Precambrian basement within the map area behaved as a relatively solid mass during the Caledonian orogeny.

From previous descriptions (literature cited above) it is clear that movements within the basement were mainly concentrated in the vertical plane. Horizontal displacements are restricted to cases where thrust masses of Precambrian rocks were torn loose by the great nappe movements. According to Vogt (1918) the Precambrian peneplain dips about 2½° to the west in the eastern area, and the tectonic windows are more or less elongate, NW–SE trending bulges in this peneplain. The larger Rombak window has a N–S trend while the Tysfjord Culmination and the Precambrian on Hinnøy south of the Gausvik area trend WNW–ESE. West of the Harstad area the Precambrian follows the general NE–SW regional strike in this area. Within the basement bulges, with the possible exception of the Tjeldsund area, faults of any significance are unknown. In conclusion, therefore, the vertical movements must be ascribed to more *flexures* in the basement. This is the opinion favoured by Vogt (1918) for the Dividalen window and later supported by Gustavson (1963) and Landmark (1967). The last-mentioned author also mentions some some minor faults and steep shear planes in the Mauken amphibolite which may have contributed to the movements in that local area.

Although the *upward* movement of some basement areas is the most conspicuous feature in this case, it is also obvious that movements occurred in the opposite direction. It is, for example, highly probable that the basement below the Ofoten Synform moved downward relative to the adjoining areas during the orogeny. Deformation thus probably had some effect on the basement below although there is no direct evidence to show that the Precambrian rocks were involved in really intensive folding in this area. Relations within the Rombak window (see later) are somewhat different and folding may in fact be present.

The timing of movements in the basement is another important point. In the Dividalen area the updoming of Precambrian rocks and its autochthonous Hyolithus Zone cover pre-dates movements along the basal thrust plane (Vogt 1918, Gustavson 1963). Berthelsen (1967) came to the same conclusion and, by means of gravimetric studies, demonstrated a connection between the Dividalen window and the larger Mauken window further northwest. In his own words: 'No doubt, the gravity field betrays a major culmination within the Caledonides caused by a NW–SE directed uplift of Precambrian basement rocks.' The pre-thrusting age of the uplift is thus also valid for the Mauken window. As the thrusting along the basal thrust plane is clearly post-F_1 but probably pre-F_2 (p. 37), the uplift must also, at least in part, pre-date the F_2 fold episode in this area. Similar relationships are probably valid for tectonic windows in the Bardu area, where minor vertical movements in the basement have occurred.

In contrast to these easternmost 'bulges' with their NW–SE trends, the Rombak window is oriented N–S and small tectonic windows in Salangsdalen show that the Precambrian ridge continues northwards for at least 35 km. Vogt (1941) prepared a map of the observed or inferred present-day surface of the sub-Cambrian peneplain in this area. An interesting feature of the Rombak window is the occurrence of some small areas with 'downfolded' autochthonous sediments and allochthonous Caledonian rocks (map, Pl. I). In two respects these small areas are important: 1. It can be seen that strong folding deformed the basal thrust plane below the Rombak Group. 2. The description and map by Vogt (1941, p. 204) show that the fold deformation strongly influenced the immediately underlying Precambrian basement, but did not change the attitude of the regional Precambrian structures. On the contrary, the Caledonian fold structures are seen in part to follow, and (?) to be determined by, the Precambrian structure.

The above-mentioned folding is on N–S and NE–SW axes, and as noted earlier (p. 37) it probably belongs to the F_2 phase. Vogt (op. cit.) considered that uplift and folding of the basement occurred in one phase and that this phase post-dated the thrust movements. This may possibly be true, although the two movements are of principally different character; the uplift appears to be a simple vertical movement while the observed folding points to considerable compression in the horizontal plane. Furthermore, detailed structural studies by Lindström (1958a) east of the Rombak window demonstrate that the basement was, at least to some extent, in its present position already at the time of thrusting. His observations are thus in agreement with the relations described above from the Eastern Troms windows. To the present author, however, it seems unlikely that the thrust nappe should be able to move up on top of and across this basement culmination, leaving it with relatively few signs of tectonical influence, if the basement had already risen to its present height relative to the adjoining areas. The following sequence of events within the Rombak window is therefore proposed:

(F_1 folding, not recognized within the window itself)
Initial uplift
Thrusting
Continued uplift of basement and F_2 folding of the basal thrust plane

Relationships in southern and western areas are comparable to those described above. The F_2 fold structures, especially, follow the pattern of basement culminations in the Hårstad–Tjeldsund area (Fig. 19). As noted previously (p. 24) the Tysfjord Culmination is parallel with a number of NW–SE trending 'cross' folds, one of which is the Salangen Antiform of supposed F_2 age. This suggests that F_2 folding and uplift are contemporaneous or that the F_2 folding was to some extent determined by an older pattern in the basement topography. Thrusting in this western area is clearly pre-F_2 and if the age relationship between thrusting and basement uplift is assumed to be the same as in the eastern parts of the map area, this means that pre-F_2 uplift also occurred. Another point mentioned (p. 12) in connection with the conglomerate description, is that it is probable that movements, involving the basement, occurred at an early stage in the history of the area, before the F_1 episode. As concluded for the Rombak window, however, it is probable that uplift also continued during the F_2 fold episode. The implication, therefore, is that basement uplift occurred over a considerable span of time in this region.

Faults

The most important faults in this region are those in the coastal districts. The *Tjeldsundet fault* (Fig. 19 and Pl. I) was described by Vogt (1941) as a major fault with mainly horizontal movement along a curved fault plane. The main reason for believing the fault to be curved seems to have been the change from a NNE trend along Tjeldsundet to a NE trend north of Tjeldsundet. This can be better explained by assuming that the fault movements occurred along two different planes, one along Tjeldsundet–Vågsfjorden and another along Astafjorden. The former fault appears to have been largely a dextral strike-slip fracture, the horizontal movement being several kilometres. Vogt (op. cit.) suggested about 3.5 km, but it is probably more, perhaps as much as 10 km if the marble sequence south of Fjelldal was originally connected with corresponding marbles north of Lilleng. The Tjeldsundet fault may not be a simple wrench fault, however, but may also have a component of dip-slip. The *Astafjorden fault* (Pl. I) is probably a vertical fracture with maximum vertical displacement in the north, and perhaps with a component of sinistral strike-slip in the south.

Minor faults related to these major faults include a dextral strike-slip fracture in Skånland, east of Tjeldsundet, and a similar fault south of Salangen, displacements being in the region of some tens of metres. Faults appear to be far less common in inland areas.

The age of the fault movements is unknown. Clearly they post-date F_2 folds in the Tjeldsundet area as these structures are displaced by the faults. It is possible that some relationship may exist between the F_3 flexures and faulting, but proof is lacking. The coastal location of most of the faults may, however, indicate that they are younger than the Caledonian orogeny and related to the supposed fault zones along the entire Norwegian coast, for which Holtedahl (1953, p. 375) suggested a Tertiary age. Strong movements of post-Eocene age are known from Spitsbergen, and faulting on Andøya, about 60 km NNW of Tjeldsundet, is known to be younger than the early Cretaceous (Ørvig 1960). Fault movements may indeed have occurred at several times in this area, and Grønlie (1922, p. 11), in a study of postglacial shorelines, has shown that faulting along Tjeldsundet has also occurred in postglacial times.

Structural summary

The main features of the structural history in the different parts of the area show a basic similarity, a summary of which is given in Table 1. The first fold episode (F_1) is associated with a strong regional metamorphism in all parts of the area, a schistosity paralleling the fold axial surfaces. Axial directions vary considerably but are mostly about WNW–ESE, or perpendicular to this direction. A strong *b*-lineation is connected with this folding.

A number of relatively large fold structures belong to the second (F_2) fold episode. Three sets of axial trends belonging to this episode have previously been recognized in the eastern areas; two of these, with NW–SE and NE–SW trends, are also present in the western and central areas.

The F_3 deformation produced weak flexures in the central and western areas. The near N–S axial trend is more or less coincident with one set of Eastern Troms D_2 folds. As these are the oldest group of D_2 folds (Kalsbeek & Olesen 1967, Mortensen 1972) it is unlikely that they can be correlated with the F_3 folds of western areas, even if the latter are merely a late phase of the F_2 episode. Folds older than F_1 have been detected only in the eastern area, but they may perhaps a wider distribution than known at present.

Earlier epeirogenic movements are demonstrated by the occurrence of the Elvenes conglomerate zone (p. 12). The presence of Precambrian granite pebbles in the conglomerate in the Harstad district and the regional extent of the zone suggest that the disturbance may be one of considerable importance.

Except for the Seve thrust, which may be syn-or-late-F_1, the thrusting in the central and western areas seems to be mainly post-F_1 pre-F_2. In eastern areas too, the thrusting is certainly post-F_1 (Kalsbeek & Olesen 1967). It is probable, therefore, that post-F_1 thrusting occurred more or less contemporaneously over the whole area. This does not exclude the possibility that thrust movements occurred more than once and perhaps in close association with fold episodes, as suggested by Mortensen (1972) for the area south of Altevann.

Regional comparisons

TORNETRÄSK AREA

Some structural similarities between the present area and the Torneträsk area in Sweden have been discussed earlier (p. 36). The main difference between the two areas is seen in the absence of the Torneträsk lower thrust rocks, the Rautas complex (Kulling 1964), within the present map area.

The structural history of several areas east of the Rombak (Sjangeli) window has been discussed by Lindström (1955a, b, 1957, 1958a, b). Three main deformation episodes were described, and in certain parts of the region an even older phase could be recognized. Of the three main deformations the oldest, the Palnoluokta phase, has B=b about 030°[1] and tectonic transport towards

[1] Lindström's earliest results, as given in the 1955 papers, were later revised in the 1957 paper.

Table 1 Summary of tectonic events and structural features in various parts of the area

Harstad area	Ofoten Synform	Eastern Troms	
		Målselv – Altevann (Kalsbeek & Olesen 1967)[1]	S. of Altevann (Mortensen 1972)
		I Early isoclinal folds Variable axial directions Early schistosity	
F_1. Isoclinal folds. Trends variable, though frequently E–W. Axial plane schistosity.	F_1. Isoclinal folds, minor and (?) large-scale. Axes trending WNW—ESE to NW–SE, and NE–SW. Axial plane schistosity. Pronounced lineation, mainly WNW–ESE	II Large-scale isoclinal folds. Axes trending mainly WNW. Axial plane schistosity developed. Strong WNW–ESE lineation	D_1. Small-scale isoclinal folds. Axial plane schistosity developed
Basement doming (starting very early, lasting into the F_2 episode).	Basement doming	Basement doming	
Thrusting. (Late ? and post F_1)	Thrusting. (Post-F_1, Seve nappe late-F_1?)	Thrusting post-dates episode II; also younger than III(?)	Thrusting?
F_2. Minor and large-scale folds. Open to tight folds. Axial plane schistosity not common. Final basement doming	F_2. Minor assymmetric folds. Trends NW-SE and NE-SW. Large-scale synforms and antiforms. Axial plane schistosity usually not developed. Final basement doming	III Minor folds with axes NW–SE, NE–SW and N–S. Folds with one subhorizontal and one subvertical limb	D_2. Minor folds of flexural slip type, with variable axial trends Thrusting
F_3. Flexures. Axes about NNE–SSW.	F_3. Flexures. Axes about NNE–SSW		
Faulting			

[1] Data on basement doming from other sources, see text.

ESE. Following this, the Trappsteget phase has B=b about 060° and tectonic transport towards c. SSE. The last (Rajkijokka) phase is a weak, mainly ruptural deformation with B=b about N–S. If Lindström's scheme of tectonic events is accepted it would seem natural to correlate the Rajkijokka phase with the F_3 episode of central and western Troms. Tentatively, the F_1 and F_2 fold episodes could then be correlated with the Palnoluokta and Trappsteget phases. Typical of these phases, according to Lindström (op. cit.), are lineations in the tectonic *a* direction, that is, trending WNW and NNW. In Troms the strong lineations of the F_1 phase, trending WNW, have been interpreted as a *b*-lineation (see Kalsbeek & Olesen, 1967 and this paper, p. 21). It is not the intention here to discuss the genesis of these lineations, but the similarity in direction, appearance, and mutual age relationships would appear to indicate that the F_1 lineations in Troms can be correlated with the lineations of the Palnoluokta phase.

NORTHERN TROMS

Structural descriptions are not available from the ground immediately north of the map area, except for the information on the Precambrian Mauken window given by Berthelsen (1967). From areas further to the northeast, structural data have been presented by Padget (1955), Skjerlie & Tan (1961), and Binns (1967).

According to Padget (1955) the structural geology of the Birtavarre and Skibotndalen areas is determined by major folds trending NNW and ENE, and by thrust tectonics. One major thrust boundary is recognized in Skibotndalen, located on top of the autochthonous Hyolithus Zone. The thrust nappe, with Sparagmitic Schists at the base and Birtavarre Series above, was tentatively correlated with the Seve nappe. An upper thrust boundary, the Cappis thrust, was thought to be of minor importance only.

East of this area, Skjerlie & Tan (1961) described the structural geology of the Reisadalen area. A lower nappe, the Jerta Nappe, is thrust on top of the Hyolithus Zone. This nappe is composed of quartzite, shale, dolomite, and a tillite conglomerate, the occurrence of the tillite conglomerate suggesting an Eocambrian age for these rocks. Above the Jerta Nappe is the Reisa Valley Nappe with mylonitized Sparagmitic Schists at the base and the Birtavarre Series above. Although some differences are apparent between the Birtavarre Series lithologies in the two areas, it is evident that the Reisa Valley Nappe corresponds to the allochthonous sequence in Skibotndalen–Birtavarre. The Jerta Nappe, or a corresponding unit, is thus absent in Skibotndalen, and clearly wedges out between the two areas.

According to Skjerlie & Tan (op. cit.) there is also a thrust plane *within* the Reisa Valley Nappe between the Sparagmitic Schists and the Birtavarre Series. A thrust at this niveau is possibly present further south in the area between Skibotndalen and Signaldalen, as Binns (1967) has divided the allochthonous sequence into a lower and an upper unit, separated by a thrust plane. Of these units the lower one is composed of phyllonitic rocks, which Binns correlates with the Sparagmitic Schists in Skibotndalen. Although the boundary between the Sparagmitic Schists and the overlying sequence was not drawn as tectonic by Padget (op. cit.) in Skibotndalen, it thus seems possible that thrust movements of regional importance occurred at this level.

Correlation between these northern areas of Reisadalen, Skibotndalen and Signaldalen, and Southern Troms, cannot be more than suggestions at the present state of knowledge, but there are similarities in sequence and general structure which may prove to be more than coincidence. A correlation between the Sparagmitic Schists in Northern Troms and the lower allochthonous unit (Storfjell Group) in Southern Troms is supported by similarities in the most important rock-types (see Table V in Gustavson 1966) and correlation of this unit with the Abisko Nappe has already been proposed (p. 36). Its upper boundary is the Seve thrust in the Torneträsk area as well as in the Southern Troms; likewise, a thrust plane occurs at the top of this unit in Northern Troms. If this correlation is correct, the implication is that the Jerta Nappe

Table 2 Correlations of main tectonic units in the areas from Torneträsk to Reisadalen

Reisadalen		Birtavarre – Skibotndalen		Signaldalen – Skibotndalen	Southern Troms	Torneträsk	
Skjerlie & Tan (1961)		Padget (1955)		Binns (1967)	Gustavson (1966, and this paper)	Kulling (1964)	
Birtavarre Series	Reisa Valley Nappe	Birtavarre Series	Allochthonous Unit	Upper Allochthonous Unit	(Overlain by Middle and Upper sequences of Kalsbeek & Olesen 1967) Main unit of Allochthonous rocks (Seve nappe)	Seve-Køli Nappe	Allochthonous structural units
Sparagmitic Schists		Sparagmitic Schists		Lower Allochthonous Unit	'Storfjell Group' (Girunas Schists of Mortensen 1972)	Abisko Nappe	
Jerta Nappe (with tillite)		Lacking		Lacking	Lacking	Rautas Nappe Complex	
Hyolithus Zone		Hyolithus Zone		Hyolithus Zone	Hyolithus Zone	Hyolithus Zone	Autochthonous
Precambrian basement		Precambrian basement		Precambrian basement	Precambrian basement	Precambrian basement	

of Reisadalen (which is absent in Skibotndalen) has no counterpart in Southern Troms. Quite possibly it may be correlated with the Rautas complex (Kulling 1964) of the Torneträsk area which contains similar rock-types, except for the tillite, and is in a corresponding structural position. As noted earlier, the Rautas complex seems to wedge out at the national border and is lacking in Southern Troms. A summary of the proposed correlations is given in Table 2.

Concerning the *structural history,* data from the above-mentioned areas of Northern Troms are relatively scanty. The structural picture appearing from the maps of Skjerlie & Tan (1961) and Padget (1955) is one of relatively open major folds trending NW to NNW and with the main lineations in the same direction. Transverse folds are also present. All folds seem to be ascribed to one major episode of deformation, while a variable style and orientation of folds and lineations are believed to be due to the varying competency of the rocks (Padget op. cit., p. 98) or to local deviations in the movement pattern. Tectonic transport is considered to have been towards SE or SSE and the most conspicuous folds and lineations are those in the tectonic *a* direction. From this description the structural history would seem to be relatively simple with folding, more or less contemporaneously with regional metamorphism (Padget op. cit., p. 101), followed by thrusting on the major thrust plane with movements continuing in about the same direction.

Binns (1967) presents a more complex history for the Signaldalen–Upper Skibotndalen area; 'At least one early period of isoclinal folding is recognized, and the main thrusting is associated with this. These structures were refolded by a system of open NW–SE folds and a possibly later system of gentle NE–SE flexures.' At least two fold episodes are therefore present. The first of these can probably be correlated with the F_1 episode of Southern Troms. The open folds of the second generation, trending NW–SE, and possibly also those trending NE–SW, would appear to correspond to the F_2 fold episode of Southern Troms. They also show similarities in style and orientation to the predominant fold systems in the Birtavarre and Reisadalen areas, which may therefore be F_2 folds. It is possible that F_1 and F_2 are broadly coaxial fold systems in these areas, as has been shown to be the case in the Ofoten Synform (p. 26) as well as in the coastal district of North Troms. Hooper (1968), when dealing with the lineation in the Reisadalen and Birtavarre areas, states that 'reconnaissance work by the writer and his colleagues leaves no doubt that it is parallel to the axes of early tight to isoclinal folds (Fleuty 1964) and is a continuation of the F_1 lineation mapped in detail on Skjervøy (Ash, in press) and the Kvænangen area.' It may therefore well be that the predominant lineation in the Birtavarre and Reisadalen areas is F_1, while the major folds, trending in the same direction as the lineations, are F_2.

A summary of the structural history of the Lyngen area has recently been given by Randall (1971). A threefold subdivision of the structural history is proposed, with an early episode of recumbent folding, a second folding on NNE axes with major folds responsible for the main distribution of the rock groups, and a third deformation which produced a 'cleavage of similar strike to that of the second episode and some minor upright folds'. Description is too brief for any safe conclusions to be drawn but it seems possible that these episodes can be correlated with the three deformation episodes described in this paper from Southern Troms.

The various suggestions put forward in the above discussion thus appear to indicate that there is a good deal of correspondence between the structural history of Northern Troms and that of the Southern Troms and Ofoten areas.

AREAS TO THE SOUTH

Based on work in the Bodø–Sulitjelma region, Nicholson & Rutland (1969) suggested correlations of major structural units as far north as to the present map area. Although several aspects of these proposals are of considerable interest, the present author regards our present knowledge of structures of the intervening areas to be too sparse for any safe correlations to be drawn. However, a few brief comments will be given on some regional structural features which constitute important points in Nicholson & Rutland's paper:

a) The thickness relations of the different structural units
b) Differences in the boundary relations of structural units between eastern and western areas

According to Nicholson & Rutland (op. cit., see e.g. Fig. 15), the most conspicuous structural feature on a regional scale is the westward thinning of all major structural units. It is shown that these thickness differences are not related to the major synforms and antiforms. It is also stated that 'structural isopachs trend NW–SE across the N–S trending fold axial traces. The thinning evidently antedates these folds'. However, as the thinning is usually in the western limbs of the late major synforms and the marble groups are thickest in the synformal areas, there is, according to the authors, evidently 'some correlation in space between the siting of the late structures and the earlier thinning and it may be that the late structures were initiated at a very early stage'.

A similar westward thinning of structural units is present in the Køli of Västerbotten, as emphasized by Zachrisson (1969, p. 30 and Fig. 6).

In the Southern Troms and Ofoten Areas somewhat similar relations are valid for part of the sequence. The lowermost structural unit, comparable with the Abisko Nappe, shows a thick sequence of psammites, phyllonites, and, in part, cataclastic Precambrian rocks in the easternmost area. Westwards it is composed of a much thinner sequence, chiefly of metamorphosed psammitic rocks, and from the west side of the Rombak window it becomes extremely thin and is partly absent below the Seve Nappe. West of the Ofoten Synform this basal structural unit reappears in the Gausvik Area with increasing thickness. This increase seems to take place mainly in the basal psammitic part of the unit.

Within the Seve Nappe the relationships are different in different parts of the sequence (see profile on Pl. I). The Rombak schists and marble are thickest north of the Rombak window, but thin markedly westwards below the Ofoten Synform. The same is true of the Narvik Group, which has its thickest development of schists and gneisses in Herjangen, north of Narvik. For this unit a thinning northwards to Salangen is also present (Gustavson 1966, p. 70). The Evenes Marble Group, which is on top of the Narvik Group and below the conglomerate zone at the base of the Salangen Group, has a thick development in the Harstad–Tjeldsund area and in Skånland west of the Ofoten Synform. It is absent east of the Synform although it may be represented in the Håfjell area by the Ballangen marbles (p. 31), and is therefore perhaps thinning in the opposite direction to that of the units below (see profile, Pl. I). For the Salangen Group no great thickness change can be proved and the same is true of the Niingen Group, but as these rocks are found only in the central area, with a relatively small east-west extension, no conclusions can be drawn about their thickness variations on a regional scale.

The cause of this general thinning of the main units is thought, by Nicholson & Rutland, to be tectonic. In their own words, 'The writers have inferred that this characteristic thinning of the western limbs of major late synforms is of early tectonic rather than original depositional origin . . .'. For the Ofoten and Southern Troms areas the reason may be partly tectonic, partly depositional. A tectonic thinning seems probable for the greater thickness variations in the psammitic unit, especially for the piling up, to considerable thicknesses, in the eastern areas. With regard to the thickness variations in the Rombak and Narvik Groups, these may, at least in part, be due to original variations in sedimentation. There is, for instance, an increase in the number and thickness of the marble layers northwards from the Rombak window which probably cannot be ascribed to tectonic causes. Similarly, a thickness decrease northwards in the Narvik Group must be ascribed to sedimentation changes (Gustavson 1966, p. 70). The thickness variation in the Evenes Marble Group is, at least partly, a primary feature, as marble pebbles in the Elvenes–Harstad conglomerates indicate that the Marble Group was, in places, removed by erosion rather than by tectonic movements. In general, therefore, the thickness changes in the present map area are less uniform than seems to be the case in the Bodø–Sulitjelma region, and in several cases they seem to be of primary depositional origin.

Another important aspect of Nicholson & Rutland's (op. cit.) discussion is the boundary relations between the different structural units. While the structural units are disjunctive and clearly allochthonous in the easternmost areas, they are to a large extent conjunctive in the west. The sparagmitic cover of the western basement areas is described as pseudo-autochthonous. About this cover it is said that it could be 'essentially autochthonous though there may have been major relative translations of stratigraphic units without disturbing the overall succession'. In this respect the similarities to the Ofoten area are obvious; the major thrust planes are clearly disjunctive in the eastern areas, especially the major thrust plane below the lower allochthonous unit. These same basal psammitic (sparagmitic) rocks west of Tjeldsundet are, however, conjunctive on top of the basement. As there are no Hyolithus Zone sediments beneath, the psammites may well represent the original sedimentary cover. The Seve Nappe boundary above seems to be conjunctive in a greater part of the area, and as described by Kulling (1964) in Sweden discordances are only locally present. However, because of the presence of Precambrian granite at its base in some areas, for example west of Tjeldsundet, there can be no doubting its importance as a regional nappe boundary.

The conjunctive boundary relations in the west, and in part in the eastern area, can be explained in at least two ways. One possibility is that the thrust movements are synmetamorphic. It is reasonable to assume that a certain structural parallelism would have been maintained during movement under such 'hot' conditions. This appears to be a possibility for the Seve Nappe, which in the present area is characterized by either a relatively small or no difference in metamorphic grade across its basal thrust plane. The second alternative is that thrust movements followed planar structural weaknesses which already existed before thrusting. In the case of post-F_1 thrusting, such structures could have been the foliation surfaces. As F_1 is characterized by recumbent folds, the schistosity planes would be generally flat-lying and parallel to primary lithological boundaries. Thrusting would result in conjunctive boundaries within the metamorphosed areas while disjunctive relationships would develop against unmetamorphosed or weakly metamorphosed sediments in the eastern areas. For the 'basal thrust' this would appear to be the most probable explanation.

In addition to the cited paper by Nicholson & Rutland (1969), the structural history of central parts of Nordland has been described by Rutland & Nicholson (1965), Bennett (1970), and Wells & Bradshaw (1970). The paper by Rutland & Nicholson gives a survey of the Bodø–Svartisen area, while the two other papers give detailed descriptions of smaller parts of the same area. An early episode of isoclinal folding, F_1, with the development of axial plane schistosity, is recognized by these workers. The orientation of F_1 fold axes and lineations is usually E–W to NE–SW. Some structures denoted as F_1 by Rutland & Nicholson (op. cit.), e.g. the large Krokvatn Fold, are termed early F_2 by Wells & Bradshaw (1970). F_2 folds are generally coaxial with F_1 folds, but while Rutland & Nicholson (1965, p. 76) state that F_2 are 'much more open asymmetrical structures' Wells & Bradshaw (1970, p. 53) found that 'there may be considerable uncertainty whether the earliest lineations and minor fold axes in any area are early-F_2 or relics of F_1'. Bennett (1970, p. 13) states that the F_2 folds 'are less obviously intense than the F_1 folds, and the schistosity can be traced around their hinges . . . In some lithologies, however, differentiation between F_1 and F_2 minor structures has not proved possible, and they have been mapped as undifferentiated early phase structures.' Both episodes seem to have a complex deformation history, and the two may grade into each other. The basement granite gneisses were involved in the deformation during the F_2 episode. In the words of Nicholson & Rutland: 'There thus seems to be a prolonged early phase of deformation producing the large-scale nappe structures and the regional lithological sequences which were then disturbed by the later phases, whose effects tend to dominate local outcrop patterns, . . . These later phases include the granite gneiss tectonics that may be examples of density-controlled gravity structures.' An F_3 phase of folding deforms F_1 and F_2 structures, resulting in large antiforms and synforms with a NE–SW or N–S trend. The present structural grain of the areas in question was determined by these late F_3 folds (Bennett, op. cit.).

A direct correlation of the 3 deformation episodes with those of Southern Troms is not obvious, especially as some differences clearly exist in the distinction between F_1 and F_2 folds by different authors. Furthermore, Wells & Bradshaw (op. cit., p. 51) believe the F_1 deformation to be 'essentially premetamorphic', which would be in contradiction to the present author's opinion of F_1 in the Southern Troms area. But, as noted (p. 34), folds older than my F_1 have been observed in Eastern Troms by Kalsbeek & Olesen (1967). The relations between F_1 and F_2 seem to be much the same in the two areas; F_1 axial plane schistosity was folded around the hinges of F_2 folds, which did not themselves usually develop axial plane schistosity. Development of porphyroblastic minerals occurred in the interval between the F_1 and F_2 episodes in both areas. Lastly, some similarities in axial direction of F_1 and F_2 are present: in both areas the two episodes are partly co-axial, but less pronounced in Southern Troms and Ofoten where one set (NE–SW) of the F_2 trends coincides with the F_3 directions of central Nordland. A similarity between fold style of the NE–SW trending F_2 folds of Ofoten and F_3 of more southerly areas seems to be present (compare Fig. 24 of Wells & Bradshaw with Fig. 25 of this paper). F_3 folds shown by Bennett (his Fig. 14) however, are also like F_3 folds of the present area, and a similar trend (about N–S) may occur. Wells & Bradshaw (op. cit. p. 14) state that F_2 and F_3 'almost certainly overlapped in time.' The same authors describe a case where two sets of F_3 trends (N–S and WNW–ESE) occur more or less in the same manner as the two sets of F_2 in Southern Troms.

The authors conclude: 'It is reasonably certain in this case that the two trends developed by contemporaneous cross-folding.' It can be stated, therefore, that *great similarities exist between the early and late structures in Southern Troms and Ofoten and those of central Nordland.* The similarities comprise axial directions as well as fold style.

An interesting question is whether the proposal by Nicholson & Rutland (op. cit) of density-controlled gravity structures can be applied to the Southern Troms and Ofoten areas. It has been noted already (p. 32) that the trends of F_2 fold structures are parallel to the strike of the basement/meta-sediment boundary in the Harstad–Tjeldsund area and south of the Håfjell area. Some genetic relation, therefore, may exist, and it is tempting to explain basement doming and fold structures (at least the F_2 folds) as the result of gravity forces, as Ramberg (1967) proposed in his model of the Scandinavian Caledonides. The F_2 minor folds west of Tjeldsundet (north of Lilleng, Fig. 19) face down-dip, away from the basement culmination, in accordance with such an explanation. However, a further evaluation of the significance of this theory for all folds and for the whole map area is not possible at the present state of knowledge.

Acknowledgements. – I thank Professor Chr. Oftedahl for introduction to the area, Dr. David Roberts for constructive criticism of the first draft of the manuscript and for improving the English, and various members of the staff at NGU for valuable help in the preparation of the manuscript.

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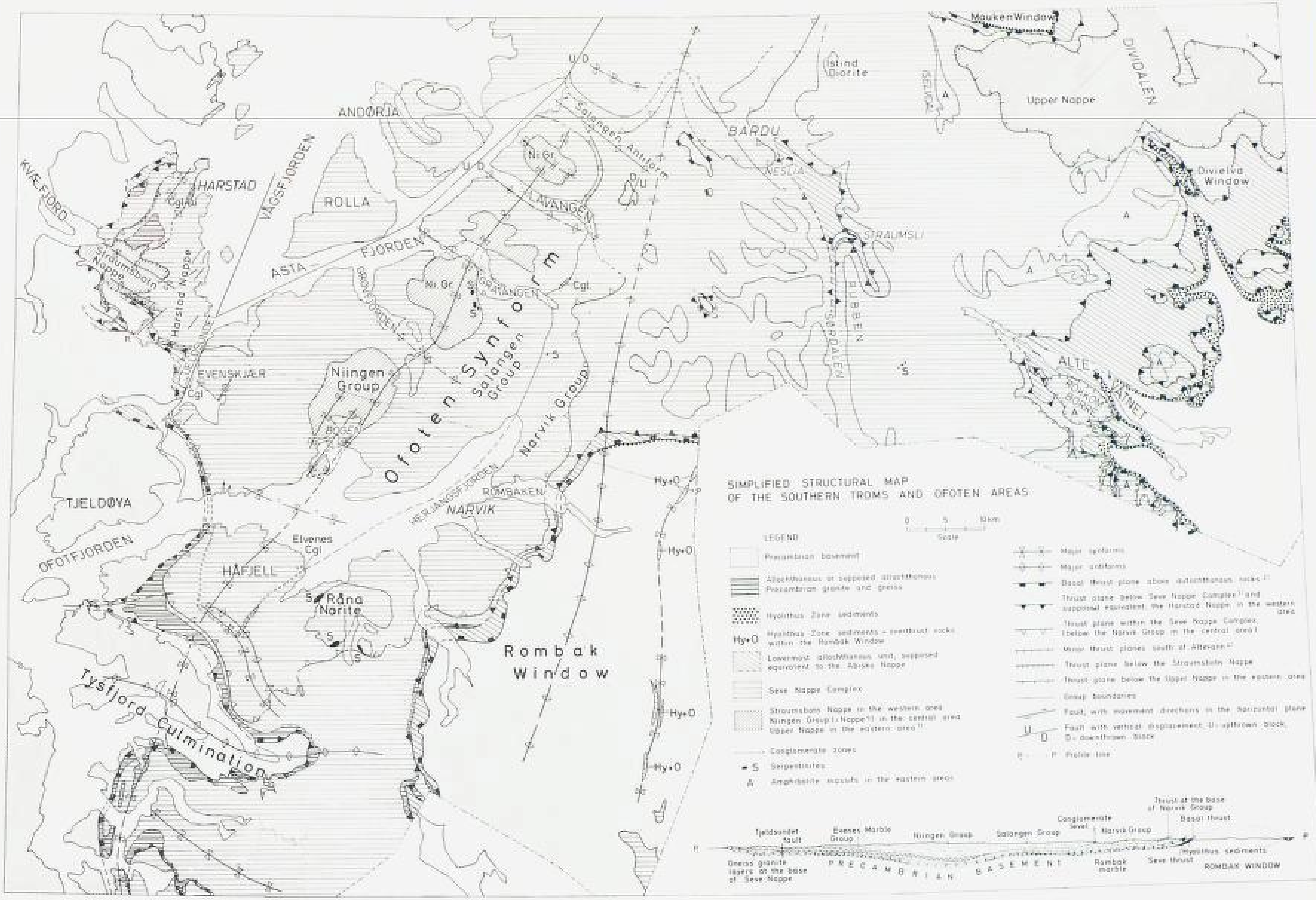

Pl. 1. Map of major structural elements in the Southern Troms and Ofoten areas.
Notes in the legend:
1) Data from Kalsbeek & Olesen (1967) and Olesen (1971).
2) Location partly from Kalsbeek & Olesen (1967) in eastern area.
3) Lithological boundary in the eastern area from Kalsbeek & Olesen (1967), tectonic interpretation according to the present author.
4) Data from Mortensen (1972).