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Weichselian Ice Movements, Sediments and Stratigraphy on Hardangervidda, South Norway

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Based on studies of glacial directional elements and on lithology, palynology and stratigraphy of the Quaternary sediments, the following climatostratigraphy and ice movement patterns are indicated; an early ice-free period, the Hovden thermomer, which is correlated with the Eemian; an early Weichselian glacial event, the Hovden kryomer, characterised by ice movements towards the east from a westerly situated ice divide (Phase I); an Early Weichselian ice-free event, the Førnes thermomer; a continuous glaciation period, the Førnes kryomer, which ended with the Holocene deglaciation. During the Førnes kryomer, the ice divide was situated far to the west in the earliest periods (Phase II). Later, the ice divide was located over eastern Hardangervidda (Phase III) and finally the ice movement diverged from a westerly situated ice divide and a dome located to the north (Phase IV). The environment during the thermomers is discussed as well as the temperature, relief and the reason for the changing ice movement patterns of the ice sheet during the kryomers.

Textural and petrographical studies have been carried out on 485 sediment samples. Emphasis is laid upon ruling out the comminution and dilution of phyllite during glacial and glaciofluvial transport.

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Introduction

The investigated area, shown in Fig. 1, is delimited by the 59° 45′ N and 60° 30′ N parallels and the 6° 35′45″ E and 8° 28′10″ E meridians This 8623 km^2 tract is covered by fifteen 1 : 50.000 topographical map sheets (Fig. 17).

The field work was carried out during the five summers 1970–74. It goes without saying that five seasons field work was not sufficient to satisfactorily cover this extensive area. Within the two southeasternmost maps, Røldal and Haukeliseter, only scattered observations from aerial photographs were made, with no field control. Varying amounts of field work have been carried out within all the remaining maps, though least on the Ullensvang, Eidsfjord, Ringedalsvatn and Skurdalen sheets.

Some results of this work have been published earlier (Vorren 1977 a, b, Vorren & Roaldset 1977).

Bedrock

The most important publications concerned with the regional bedrock geology of Hardangervidda are: Brøgger (1893), Reusch et al. (1902), Rekstad (1903), Kvale (1947), Dons (1960 a, b), Naterstad et al. (1973) and Andresen (1974 a, b, c). Other isolated observations are quoted in these studies.

The greater part of Hardangervidda is Precambrian basement (Fig. 2) with predominantly granitic and granodioritic gneisses.

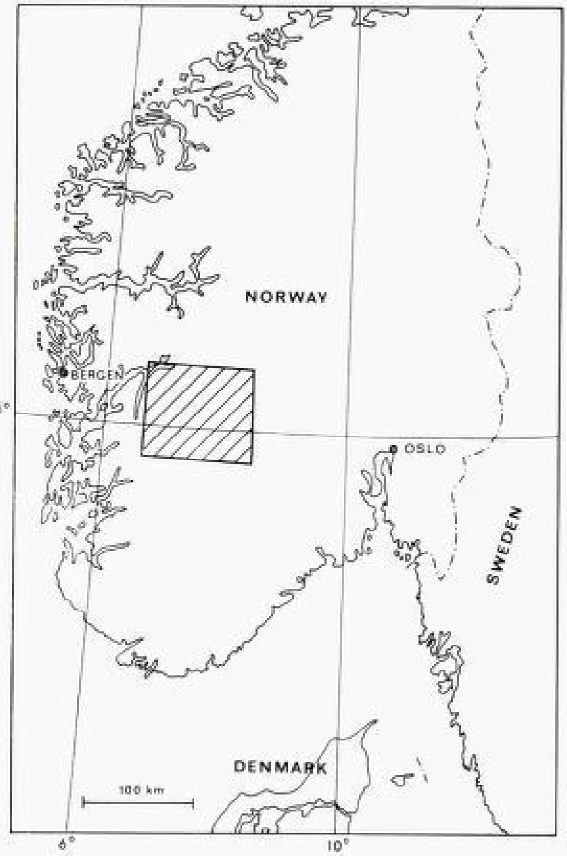

Fig. 1. Location map of the investigated area, hatched.

The Telemark suite is represented by metasediments and metavolcanics. They mainly occur in the west near Sørfjorden (Kvale 1947), and in the easterly parts (Dons 1960 a, b, Holtedahl & Dons 1960).

The Cambro-Ordovician rocks and those of the Holmasjø Formation (here referred to collectively as Cambro–Ordovician) form the main basis for my sediment petrography investigations and will therefore be described in some detail here.

Based upon his own and earlier studies, Andresen (1974a) has proposed the following stratigraphy for the autochthonous Cambro-Ordovician sequence:

grey schist (phyllite),
greenschist with marble zones,
marble,
bluish quartzite,
black shale (Alum shale),
basal conglomerate.

Naterstad et al. (1973, p. 8) consider that the greenschist is the uppermost member. This difference is probably due to Andresen assigning the Holmasjø Formation, consisting chiefly of quartz-rich phyllites (Naterstad et al. 1973,

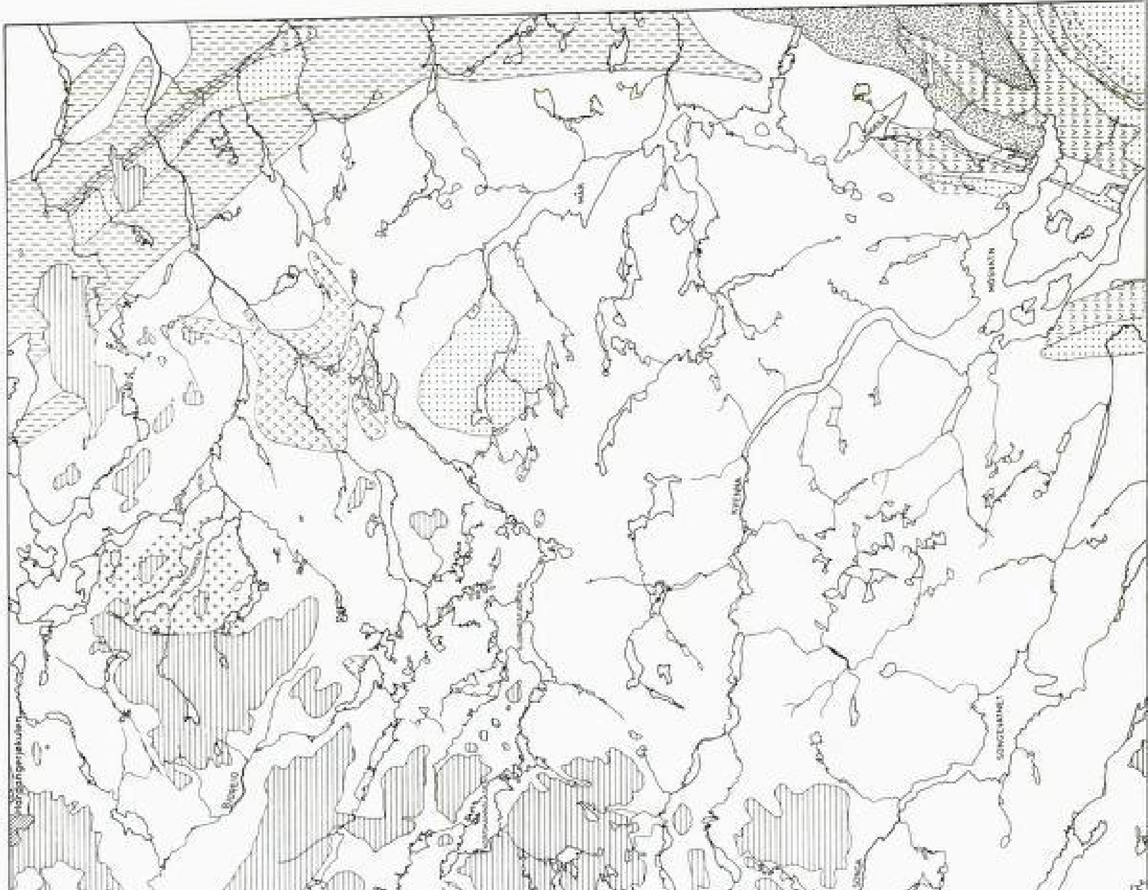

Fig. 2. Bedrock geology map of Hardangervidda based on mapping and publications cited in the text and Løberg (1973).

p. 8), to the autochthonous series as Brøgger (1893) and Reusch et al. (1902) had done.

However, in the present context it is the geographical location and extent of the different outcrops which is significant. Southwest of a line from Sandflott to Kinso river the grey schists predominate and only local remnants of the other sedimentary rocks can be found (Andresen 1974a, p. 43). The bluish quartzite is bounded by this line, to the southwest and by a north-south line along Veigdalen which, south of Grananutane, turns east-wards and continues to Bjornesfjorden before heading south again (Rekstad 1903, p. 16). The marble occurs mainly within the same zone, while grey schists dominate once more in the northern and eastern areas.

The allochthonous nappe rocks are restricted to the southern areas and to Hårteigen and Hardangerjøkulen. They consist mostly of different types of gneiss and meta-supracrustals (Naterstad et al. 1973).

Bedrock Morphology

Most of Hardangervidda lies between 1000 and 1400 m a.s.l. (Fig. 3). Areas exceeding 1400 m are found in the west, often coinciding with nappe rocks, and similarly in the north at Hardangerjøkulen. In addition, there are areas of Precambrian basement between Songevatn and Møsvatn which rise to over 1400 m a.s.l.

The greatest relative relief on Hardangervidda is displayed by the cirques and hanging valleys in the westernmost parts. The cirque topography is especially well developed in the northwest. There are quite considerable altitude variations in the southeast too, but with somewhat gentler forms. Two surfaces are distinguishable here, the Møsvatn basin at about 900–1000 m a.s.l. and the surrounding areas lying at 1200–1500 m a.s.l. (Gjessing 1967, Fig. 16). Generally, though, Hardangervidda is typified by its low relief. The origin of this more or less undulating landscape which is part of Norway's Paleic Surface has been a matter for debate, cf. Gjessing (1967). Reusch (1901) regarded it as a fluvial denudation surface, O. Holtedahl (1960, p. 508) as an exhumed sub-Cambrian peneplain and Gjessing (1967) as the product of subaerial weathering and denudation during an arid or semi-arid pre-Quaternary period.

The drainage routes from Hardangervidda's Paleic surface in the west and to a lesser degree in the southeastern areas, lead into deeply incised younger valleys and fjords. The valleys in the northeast have a more subdued appearance. As far as the excavation of the younger, deeply incised valleys is concerned, their abrupt valley ends have attracted most of the discussion. Reusch (1901, p. 207 ff.) explained them in terms of glaciofluvial erosion by streams descending from the ice surface via crevasses, whereas Ahlmann (1919, p. 77) stressed the role played by subaerial fluvial erosion. Gjessing (1956; 1966, p. 289) considered that glacial erosion had accomplished most while Holtedahl (1967, p. 195) attached more importance to subglacial meltwater erosion.

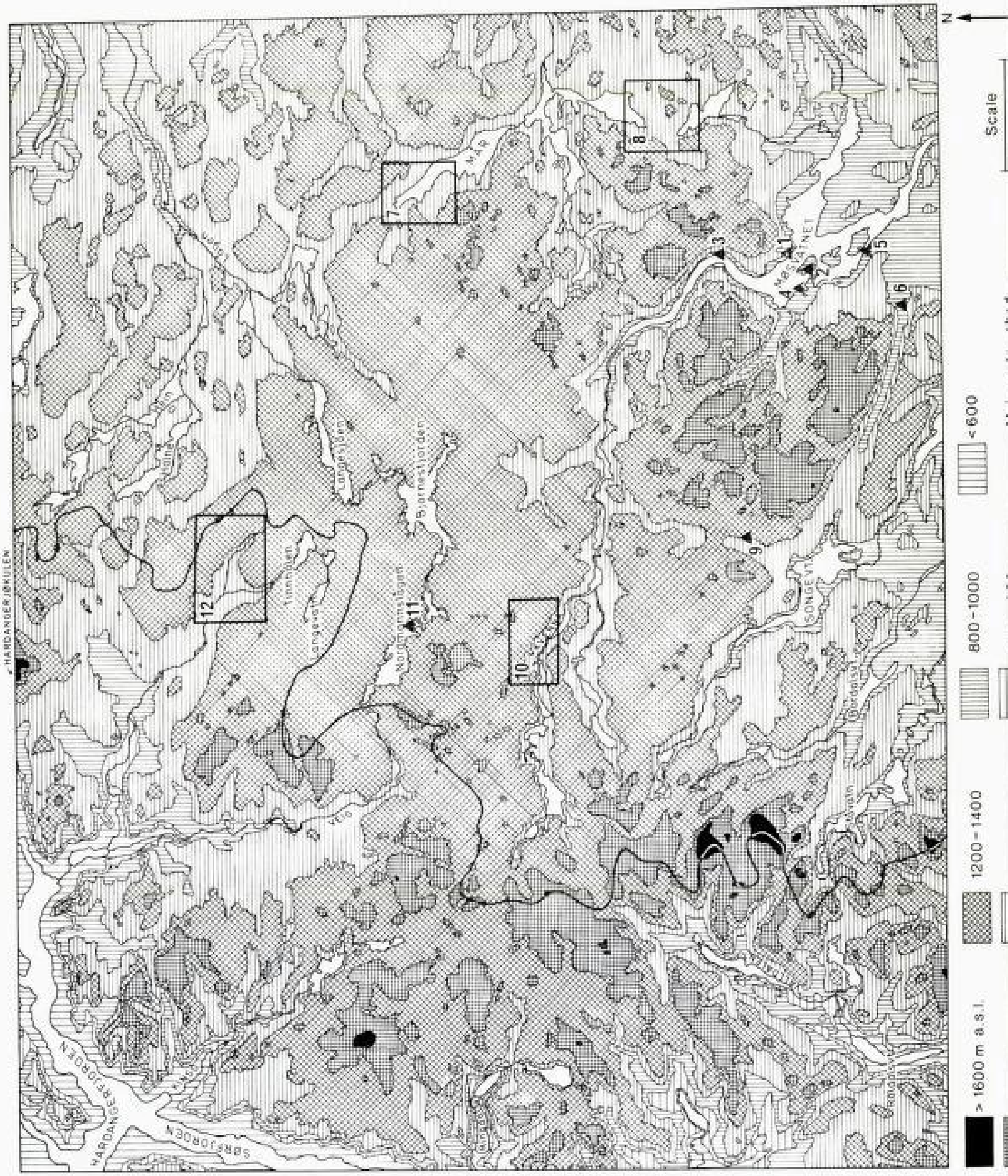

Fig. 3. Relief map of Hardangervidda. Frames and numbers indicate areas of detailed maps and localities where the stratigraphy is discussed. 1: Førnes, 2: Hovden, 3: Trengsle, 4: Laksastøl, 5: Falkeriset, 6: Bitdal, 7: Mår area (Fig. 40), 8: Gøyst area (Fig. 47), 9: Sterra, 10: Hansbu area (Fig. 52), 11: Holsbu, 12: Svinto area (Fig. 53).

Previous Quaternary geology studies

Keilhau did field work on Hardangervidda in 1842 and 1845 (Keilhau 1850,

p. 91). In 1849 Kjerulf visited the area together with Hørbye (Kjerulf 1897, p. 37). The observations made by these three are generally restricted to striae, and occasionally erratics (Kjerulf 1879). Hørbye (1857) plotted the striae data on to a map, and it is interesting to note that both Kjerulf and Hørbye noted crossing striation in the Kvenna valley (Kjerulf 1879, p. 37).

Brøgger's bedrock mapping on Hardangervidda during 1875 and 1877 included scattered observations of erratics (Brøgger 1893).

Dal (1894), who visited Hardangervidda primarily to compile a register of the bogs, also commented on some horizontal lines in Veigdalen which he assumed were comparable to the 'seter' in Østerdalen in central Eastern Norway. However, this is not the case, as the lines are clearly benches following the boundary between the Precambrian and the overlying Cambro-Ordovician sequence.

Hansen (1886), who first described the ice-dammed lakes in central Eastern Norway, applied his hypothesis to the Hardangervidda region too (Hansen 1890, p. 273; 1892, Pl. 2 and 1895, pp. 186–188). He invoked a final ice divide to 'the south of the large lakes here, Møsvand, Totak and Vinjevand'. However, he limited his field work to the southern marginal areas (Hansen 1895, p. 186 ff.).

During 1893 Reusch was mapping the bedrock in the northernmost and central parts of Hardangervidda, where he noted a few striae (Reusch 1894) and recorded scattered observations about the superficial deposits (Reusch 1896).

Occasional pieces of information on the Quaternary geology also appeared as a by-product of the Norwegian Geological Survey's bedrock mapping programme at the turn of the century (Reusch et al. 1902, pp. 25–30 and 49; Rekstad 1903, pp. 38–47). Rekstad (op. cit.) drew attention to the marked difference between the eastern and western vidde with respect to the degree of superficial deposit cover. He had also seen the eskers west of Langevatn and at the western end of Nordmannslågen. Apart from some wrongly interpreted end moraines on the eastern vidde (cf. Holmsen 1955, p. 28) he observed the end moraines round Bjoreidalen (cf. Anundsen & Simonsen 1968). In addition, Rekstad (1903, p. 41) followed Reusch (1894) in assuming that the ice divide had virtually coincided with the watershed and rejected Hansen's ice-dammed lake hypothesis.

Werenskiold (1910a, p. 60 and 1910b, p. 15 ff.) briefly mentions some loose deposits on the extreme eastern margin of Hardangervidda.

Isachsen (1933, p. 438 ff.) reported some observations from the southern end of Mårvatn where, amongst other things, he had found a fluted surface. Furthermore, he maintained that the area's deglaciation had been typified by downwasting of the ice.

The 1 : 250,000 'Hallingdal' Quaternary geology map (Holmsen 1955) covers the extreme eastern parts of Hardangervidda.

During the 1960's several university theses were written on the Quaternary geology of specific areas of Hardangervidda, viz. Anundsen (1964), Kvistad

(1965), Damsgaard (1967) and Myhre (1968 – not seen by present author). Damsgaard (1967) concentrated on the glacial flow direction problem within the southwestern areas. Anundsen (1964), cf. Anundsen & Simonsen (1968), was mainly occupied with ice-marginal deposits around Bjoreidalen–Veig, which were found to be of Preboreal age. The ice advance responsible for these moraines was also reconstructed by Liestøl (1963). The Preboreal age of this advance is further supported by the C^{14}-dating of a juniper branch in the marginal delta at Eidfjord (Rye 1970), since the latter deposit is believed to be contemporaneous with the moraines on the mountain area.

Ice flow directions have been discussed recently by Rye and Follestad (1972) and Sollid (1975).

Popular descriptions of Hardangervidda's Quaternary geology have been published by Sollid (1971) and Vorren (1974).

In the neighbouring Numedalslågen district to the east, a major research project has been under way for some years, dealing chiefly with the mineralogy and geochemistry of the Quaternary deposits (Roaldset 1972, 1973a, 1973b; Rosenqvist 1973, 1975). The glacial geology around Sørfjorden and Hardangerfjorden has been examined in a number of works including Helland (1876), Monckton (1899), Kaldhol (1941), Undås (1947, 1964) and Holtedahl (1965, 1967, 1975).

Pollen diagrams from Holocene deposits on the vidde may be found in Fægri (1945), Hafsten (1965) and Moe (1973, 1977).

The object of the present study is to investigate the ice flow direction, sediments and Quaternary stratigraphy of Hardangervidda. One may summarise the earlier work as being chiefly concerned with questions of ice flow direction and the course of deglaciation. There was thus a certain amount of background data already available on the glacial flow direction, whereas the present writer found that the sedimentary and stratigraphical aspects had scarcely been touched previously.

Ice Movements

DIRECTIONAL ELEMENTS – TYPES AND TERMS

The ice movements studies presented here are mostly based upon *directional elements*. These may be defined collectively as: *Orientated forms resulting from the moulding and sculpturing effects which moving glacier ice and its incorporated debris has on the underlying surface.* This definition therefore excludes internal structures and proglacial phenomena. It also excludes those features where water has been a contributary factor, such as sichelwanne (Ljungner 1930, p. 287) and so-called p-forms (Dahl 1965). The various directional elements are discussed extensively in the literature, e.g. Charlesworth (1957), Flint (1971), Embleton & King (1968).

Since the genesis of many of these features has not been satisfactorily explained, their present classification must of necessity be based on morphological criteria. A primary subdivision into three groups of directional elements will be proposed:

1. Transverse 2. Drumlinoid 3. Linear.

Transverse directional elements. Alternative names for the group of transverse directional elements are crescentic marks (Flint 1971, p. 95), friction cracks (Harris 1943, Embleton & King 1968, p. 142) and chattermarks (Fairbridge 1968, p. 117). An appraisal of the literature and field experience both point to five morphologically discrete kinds of transverse elements, namely crescentic fracture, crescentic gouge, lunate fracture, chattermark and conchoidal fracture. The first four are described in Flint (1971, p. 96), while the conchoidal fracture (muschelbruch) was described by Ljungner (1930).

The commonest transverse elements, which also happen to be most useful for discerning ice movement direction, are the crescentic gouge and crescentic fracture.

Drumlinoid directional elements. Some of the subdivisions previously proposed for large drumlinoid forms are rather detailed, e.g. Chamberlin (1894, p. 522 ff.) and Glückert (1973, 1974). As Flint (1971, p. 104) stressed, there is an apparent continuum of landforms from rock drumlin and roches mountonnée via crag-and-tail to drumlin. A subdivision into discrete types of drumlinoid forms is therefore difficult; perhaps one based on numerically expressed dimensions would be a step in the right direction, such as Chorley's (1959) classification of drumlins according to their width/length ratio (cf. Smalley & Unwin 1968). However, the classical grouping into rock drumlin, roches mountonnée, crag-and-tail and drumlin will be adhered to here. Additionally, there is a minor drumlinoid element called knob and tail, also known as knob and trail (Chamberlin 1888, p. 244), miniature crag and tail, ice shadow (Charlesworth 1957, p. 254) and rat tail (Flint 1971, p. 91).

Linear directional elements have received many names including striae, striations, scratches, scourings, furrows, grooves, fluted surface, fluted rock, fluted moraine, flutings and flutes. The differences between these are not clearly defined, but the following definitions are proposed here:

Scratches, striae, furrows and grooves are linear forms whose width is <1 mm, 1–10 mm, 1–10 cm and >10 cm, respectively. They comprise a group of features collectively called scouring. Innumerable examples are found on bedrock, but they can also occur on sediment surfaces (Westgate 1968).

Fluted surface consist of grooves separated by ridges termed flutes (Boulton et al. 1974, Fig. 1) or fluting (Glückert 1973, p. 16). Such surfaces are called fluted rock (Vorren 1973, p. 3) or fluted sediment, according to their nature.

METHODS

In principle there are three stages in the study of ice flow direction: data collection, analysis of the individual observations, and finally a synthesis of the analysed material. In practice the first two stages are often carried out together.

Collection of data

Directional elements were in part mapped from aerial photographs, in part from field observations. Using the photographs one can identify the larger drumlinoid forms and fluted surfaces. Fluted rock and large drumlins can be seen in the field too, but then they are more difficult to measure accurately. Fluted sediments can only occasionally be identified on the ground.

During field studies scouring and transverse elements are recorded. They were most often examined on horizontal or gently sloping surfaces, to avoid local deviations due to so-called plastic scouring (Gjessing 1965). Many of the observations of scouring are from freshly excavated and washed sites, since weathering has generally destroyed any such marks on exposed rock. When fine scratches were being examined they could be shown up better by blackening the rock surface with a pencil. No attempt was made to collect plastic casts of scouring for laboratory examination (Svensson 1957, Markgren & Frisen 1963). In some instances, however, oriented samples were taken for SEM analysis (Fig. 5).

Analysis of individual observations

Ljungner (1943, 1949, p. 29) divided field data analysis into four stages: directional, where the direction along the axis of flow was determined; successional, where the age relationship between the individual features was established; qualitative, which is related to the plasticity of the ice and therefore, according to Ljungner, its thickness; quantitative, which refers to the duration of the ice movement. Ljungner's methods have later been expanded by Gjessing (1954), Mattson (1954) and Johansson (1956).

In principle the direction and chronology analyses are correct. Doubts are expressed as to whether the quantitative and qualitative analyses really are related to the ice's thickness and duration of the flow.

The qualitative part has already been criticised by Johnsson (1956, p. 45) and Gjessing (1965, p. 2), and recent glaciological research seems to confirm their doubts. According to Glen's flow law ($E = AT^n$, where E = strain rate and T = shear stress), ice deformation is dependent on, among other things, temperature (A), and crystal orientation (exponent n). It is also fairly certain that hydrostatic pressure (i.e. ice thickness) does not influence Glen's flow law as long as temperature is measured in relation to the pressure melting point (Rigsby 1958). The ice's thickness will thus only affect the deformation rate to the extent that it lowers the pressure melting point (which is relatively insignificant) and influences basal shear stress. The latter, $T = pgh \sin \propto$ (p = ice density, g = gravitational acceleration, h = ice thickness and $\propto$ = ice surface gradient) usually lies between 0.5 and 1.5 bars (Patterson 1969, p. 91). The ice's plastic deformation is thus dependent upon several factors, of which temperature is important while ice thickness is of much less consequence.

Ljungner's quantitative analysis is also open to objection. The degree of erosion will in addition to the nature of the bedrock and the duration of ice movement, be determined by such factors as ice velocity, supply of abrasive material and temperature conditions. Taking an extreme example, ice frozen to bedrock has little or no erosion capacity (see p. 29).

In spite of these criticisms of Ljungner's analysis method the fact remains that different ice movements have affected the bedrock differently at the very same locality. It is important therefore to also carry out qualitative analysis of directional elements in the sense that their size, shape and composition are noted. In what way these properties are related to those of the agent which produced the landform is another question altogether.

To summarise, there are three main analysis which should be made of directional elements: 1. Directional analysis, 2. Chronological analysis, 3. Property analysis — size, shape, composition.

Directional analysis. Usually the direction of the flow is clearly displayed by the transverse elements (except chatter marks) and by drumlinoid forms. Previous investigations have shown, however, that caution must be exercised here, since some of these features have been observed in a reversed position. Crescentic gouges have been demonstrated to lie in a reversed position in relation to ice movement, e.g. Andersen & Sollid (1971, p. 17). Dreimanis (1953, p. 776) desribes 'crescentic gouges' where the principal fracture plane dips upglacier, implying a form which corresponds to a reversed lunate fracture.

Apart from some weakly developed forms in phyllite which could be interpreted as reversed crescentic gouges, the writer has not seen any similar features on Hardangervidda. Chamberlin (1888, p. 245) has described instances where knob and tail occur in reversed position, a phenomenon he terms 'advanced cones'. Comparable examples have also been noted in phyllite with quartz pods on Hardangervidda. Gravenor (1974) has reported drumlins in reversed position in Nova Scotia. However, on Hardangervidda most of the major drumlinoid forms are of the crag and tail type and thus provided unequivocal evidence of the direction of ice movement.

The main problems are posed by the linear directional elements. For fluted surfaces one must rely on other correlated directional elements to decide which direction along the flow axis is correct. Where scouring is concerned, there have been claims that the flow direction is deducible from the manner in which lines begin and end, e.g. Chamberlin (1888, p. 247), Flint (1971, p. 91). The writer has found this approach quite unsatisfactory and has therefore always sought supporting evidence from the best available combination of other directional elements and/or large and small-scale stoss and lee topography (Fig. 5) and/or directional forms actually within the scour.

Chronological analysis. Based on assumptions that directional elements formed in, or of, sediments, i.e. crag and tails, drumlins and fluted sediment, will be destroyed by a subsequent change in ice flow direction, it has often been pre-

Fig. 4. Small crescentic fractures in quartzite north of Store Vendevatn. The arrows point in the ice movement direction.

Fig. 5. Fine scratches in granite near Halnefjorden photographed with a scanning electron microscope. The largest scratch (ca 0.03 mm wide) in the centre of the picture contains chattermarks. The width of the smallest scratch is ca 0.002 mm (measured on enlarged photograph). Several directions can be seen; the oldest is towards east, uppermost in the picture (Phase II); of the younger directions in the sector WSW–SSW, that towards WSW is oldest and that towards SSW youngest (Phase III/IV).

sumed that such features must reflect the youngest ice movement (Tollan 1964, Flint 1971, p. 479). This has in fact proved to be true in many cases, e.g. Hoppe (1959, p. 204), Holmsen (1964), Sollid (1964, p. 66, Aario et al. (1974 p. 28). However, on Hardangervidda this relationship has often proved invalid,

Fig 6. Crag and tail north of Skiftesjoen, map sheet Bjoreio. The length is about 200 m and height about 20 m. It was formed by an ice movement towards the WSW (towards the right) during Phase III.

especially where drumlinoid forms are concerned. A similar situation has been reported by Virkkala too (1951, p. 58), where drumlins formed during an earlier phase are preserved in spite of younger movements in other directions.

In the present writer's opinion, the best method of determining the age relationship between ice movements is with the aid of scouring. Even then the only reliable criterion is the position of the scouring with respect to stoss and lee topography (cf. Lundqvist 1969). Gjessing (1954, p. 83) has illustrated and interpreted some examples, but the possible situations are multitude.

Property analysis. In contrast to the essentially direct analyses described above, property analysis aims at deducing glacial characteristics, such as rate of flow, temperature and thickness, by inference. However, at the present stage there is insufficient knowledge about which property characteristics of a particular landform are relevant. For example, is it the type of material or shape, or both or neither, which reveal the characteristics of a glacier which created a crag and tail? If shape, is it the length or width or the ratio of the two which is significant? It is such gaps in our understanding of this subject which have often caused property analysis to be neglected. It must be admitted that the analyses presented here are also rather general, for the same reasons. Generally, the analyses are limited to a rough indication of the magnitude of the various directional elements.

DESCRIPTION OF DIRECTIONAL ELEMENTS ON HARDANGERVIDDA

Transverse directional elements. The commonest transverse elements on Hardangervidda are crescentic gouges and crescentic fractures. The former may be observed on most types of the bedrock whereas crescentic fractures are restricted to crystalline rock types. These features are generally 20 to 30 cm in breadth, but do occur quite often on a much smaller scale (Fig. 4). Crescentic gouges and chatter marks (Fig. 5) are found in scourings.

Drumlinoid directional elements. Knob and tails, crag and tails, rock drumlins and roches moutonnées represent this group on Hardangervidda. Well-developed drumlins (sensu stricto) are rather rare. Crag and tails are usually 300 to

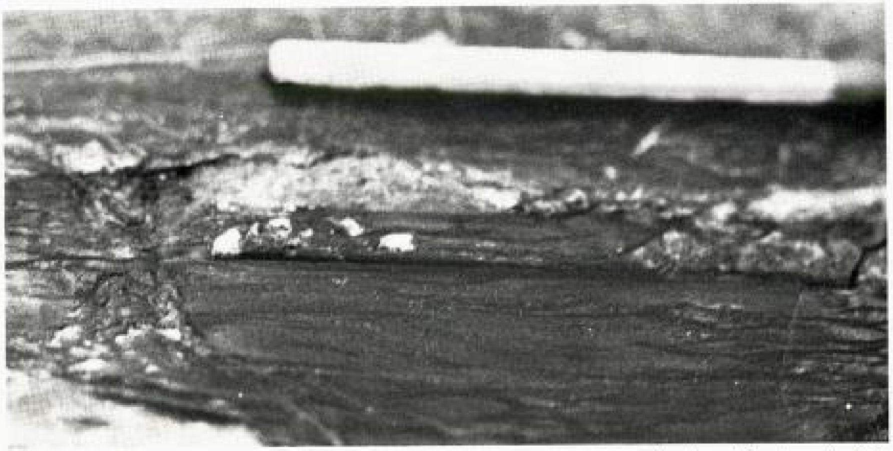

Fig. 7. Knob and tail in phyllite formed by an ice movement towards the right (west). The knob is composed of a small quartz pod. The scale is indicated by the match stick (4.5 cm long).

500 metres long but may vary from 1000 to under 100 metres. Breadth and height are dependent on the size of the crag (Fig. 6); the breadth mainly 50 to 200 metres and the height between 10 and 50 metres at the beginning of the tail. The largest crag and tail areas are found within the Lågaros map sheet (Fig. 11) and at the southern end of Møsvatn.

Rock drumlins have only developed on the softer Cambro–Ordovician bedrock. A gradual transition from rock drumlins to fluted rock is discernible. The best examples of rock drumlins are about 200 metres long and 50 metres wide. Roches moutonnées are found in all types of bedrock, while knob and tails are particularly numerous in phyllitic rocks where quartz pods form the knobs; the size of these quartz pods governs the features' overall size (Fig. 7).

Linear directional elements. Fluted rock, like rock drumlins, is limited to outcrops of the softer Cambro–Ordovician shales (Fig. 8). Roches moutonnées have developed in association with fluted rock (cf. Vorren 1973, p. 3). As already mentioned, fluted sediments are more difficult to identify in the field. Judging from the aerial photographs they are a few metres wide and about half a metre in height. They occur both on even moraine surfaces and on crag and tail formations. Scouring is found throughout Hardangervidda, often as scratches (Fig. 5), especially in the central parts of the area.

Case studies. East of Veigdalen (Fig. 9A), on Fljotdalsfjell, Berakupen, Tverrgavlen and Store Allgarden, there are very distinct fluted rock surfaces and rock drumlins. Bedrock is phyllite. The directional elements are aligned northwestwards (300°–305°), veering towards N (350°) close to Veigdalen. On Fljotdalsfjell the direction changes from 305° to 350° over barely one kilometre. This northward deflection probably reflects powerful ice-streaming down the Veigdal valley.

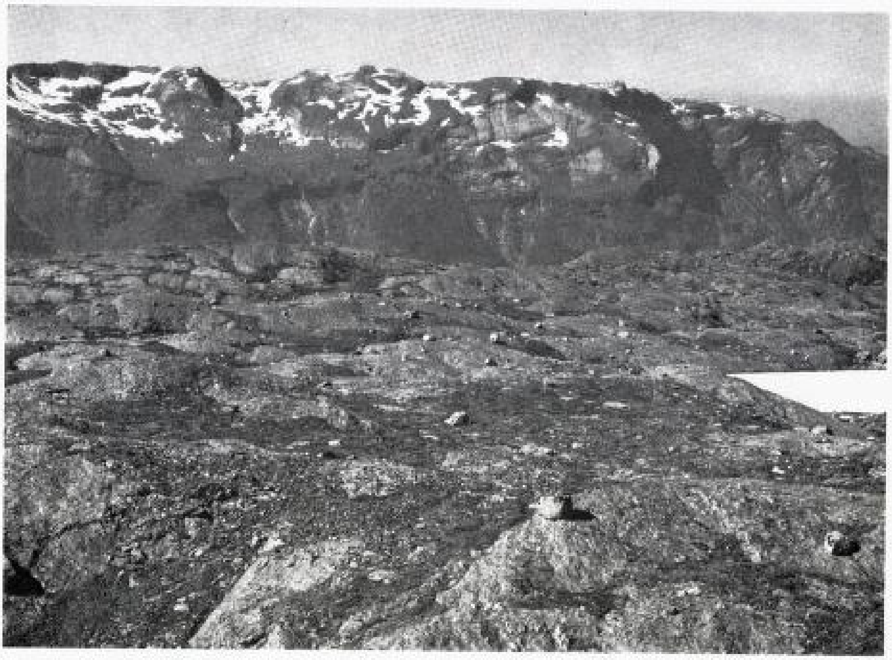

Fig. 8. Photo, looking west, showing part of western Hardangervidda. The depression in the background is Sørfjorden and the mountains behind are on Folgefonn peninsula. Note the sparsely covered bedrock (phyllite), the roches moutonnées and the furrows and grooves eroded into the roches moutonnées. These can be observed on aerial photographs as fluted rock.

On Nasafjell and in the valley to the south there are also fluted rock surfaces and rock drumlins in phyllite (Fig. 9B). On the mountain they are oriented 320°–330° but in the valley follow the latter's alignment of 290°–310°. The relief has clearly influenced the ice's direction of flow. This phenomenon seems to be a counterpart of that where scourings are deflected around roches moutonnées (e.g. Johnsson 1956, p. 47).

At the eastern end of Ringedalsvatn (Fig. 9C) there is a clear convergence of fluted rock towards the west, and once again it is the topography which seems to have determined the alignment of the directional elements.

Around the up-valley end of Mårvatn (Fig. 10) two ice flow directions are visible, an older one towards east and a younger towards south–southeast. The older movement is revealed by crag and tails, on some of which there are similarly oriented fluted sediments. The later flow produced fluted sediments. This locality demonstrates the interesting situation where directional elements in unconsolidated deposits — crag and tails with flutings — have survived a later ice movement diagonally across them (at about 45°) without their alignmen being significantly modified. The youngest movement also demonstrates a rapid directional change from eastward to southeastward.

Viewed together, these case studies show that:

1. The continental ice sheet's movement can change over rather short dis-

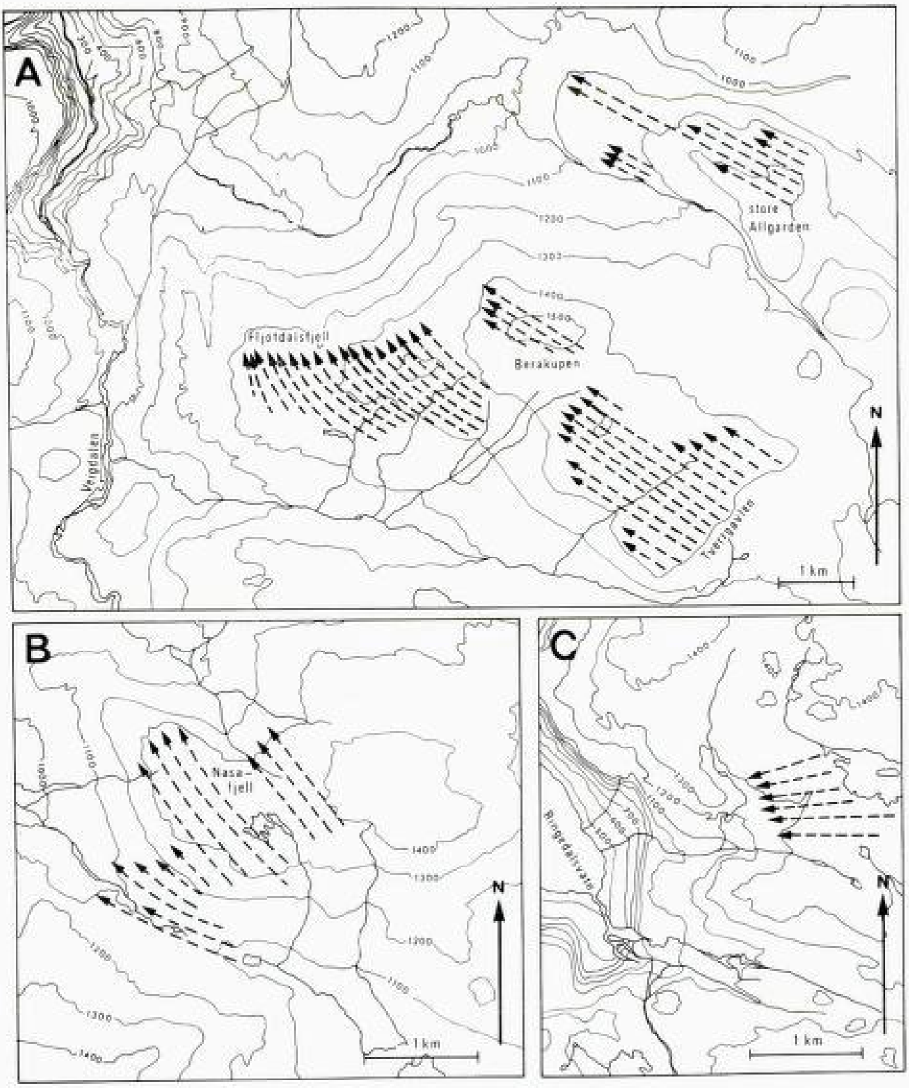

Fig. 9. Directional elements, fluted rock and rock drumlins, observed on aerial photographs from an area east of Veigdalen, map sheet Eidfjord (A); at Nasafjell between Veigdalen and Kinso, map sheet Eidford (B); east of Ringedalsvatn, map sheet Ringedalsvatn (C).

tances. This may be due to competing ice streams as at Veig and Mår, or to plastic divergence or convergence as at Nasafjell and Ringedalsvatn.

2. Directional elements in unconsolidated sediments — here crag and tails with flutings — can be preserved in spite of subsequent ice flow in other directions.

The consequence of these findings is that the utmost care must be taken in attempting the correlation of directional elements in a regional interpretation, especially if the terrain shows a fair range of relief. Furthermore it is unacceptable to assume that oriented features in loose deposits have to represent the

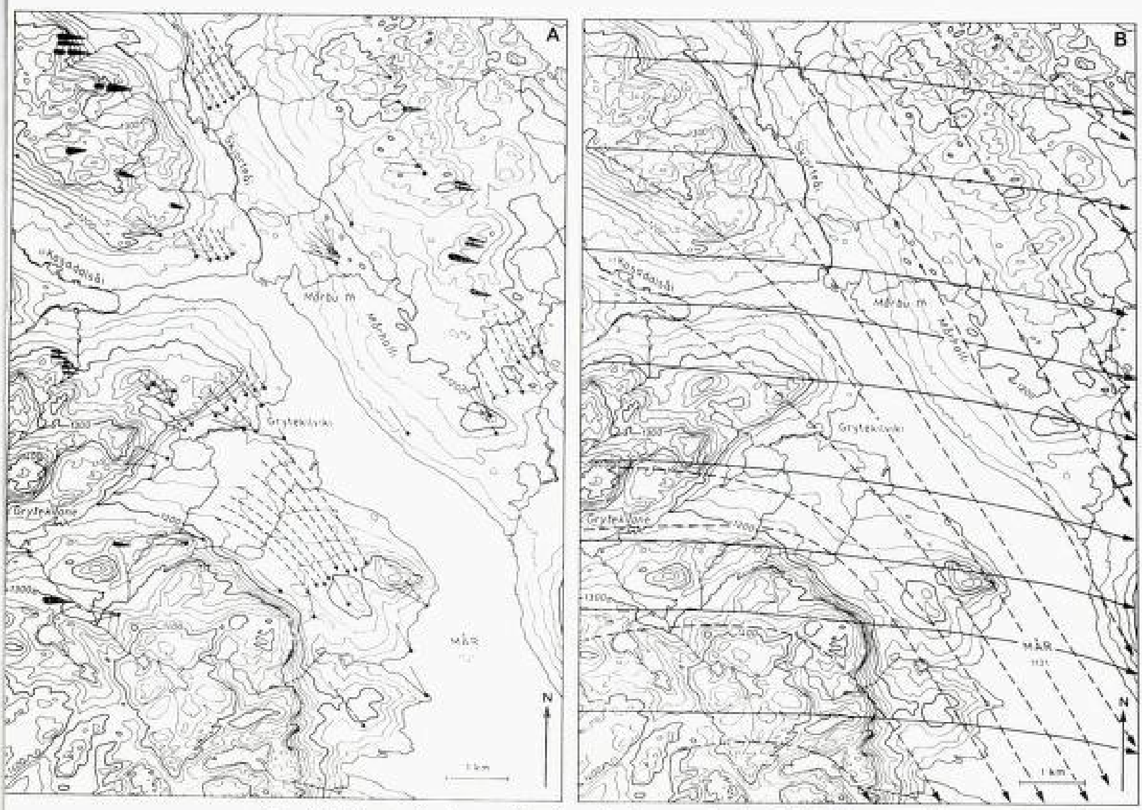

Fig. 10. Figure A shows directional elements in the Mår-lake area. The scouring observations are represented with lines towards the observation points marked with dots; dashed arrows indicate fluted sediments and crag and tails are shown by their own form. Figure B is a reconstruction of the oldest crag and tail forming ice movement which has shaped the flutings, Phase IV (dashed arrows).

youngest ice movement, unless there is confirmation from other evidence within the same area.

RESULTS AND DISCUSSION

Just over a thousand scouring localities have been observed by the writer. The majority of them are plotted on Pl. 1. In areas where few observations were made, the data were supplemented with some observations reported by Kvistad (1965) and Anundsen (1964) (cf. Rye & Follestad 1972) as well as Damsgaard (1967). The latter's observations include all those on the Haukeliseter map sheet which are presented here. Pl. 1 also includes some aerial photograph observations of fluted surfaces and large-scale drumlinoid landforms. Many of these had been noted previously by Rye & Follestad (1972).

A simplified presentation of the directional elements distribution is reproduced in Fig. 11. Each circle represents one, or two or three closely spaced, key localities. Apart from localities e (Rye & Follestad 1972), k, u and v (Damsgaard 1967), this map is based on the writer's own data. In addition, the

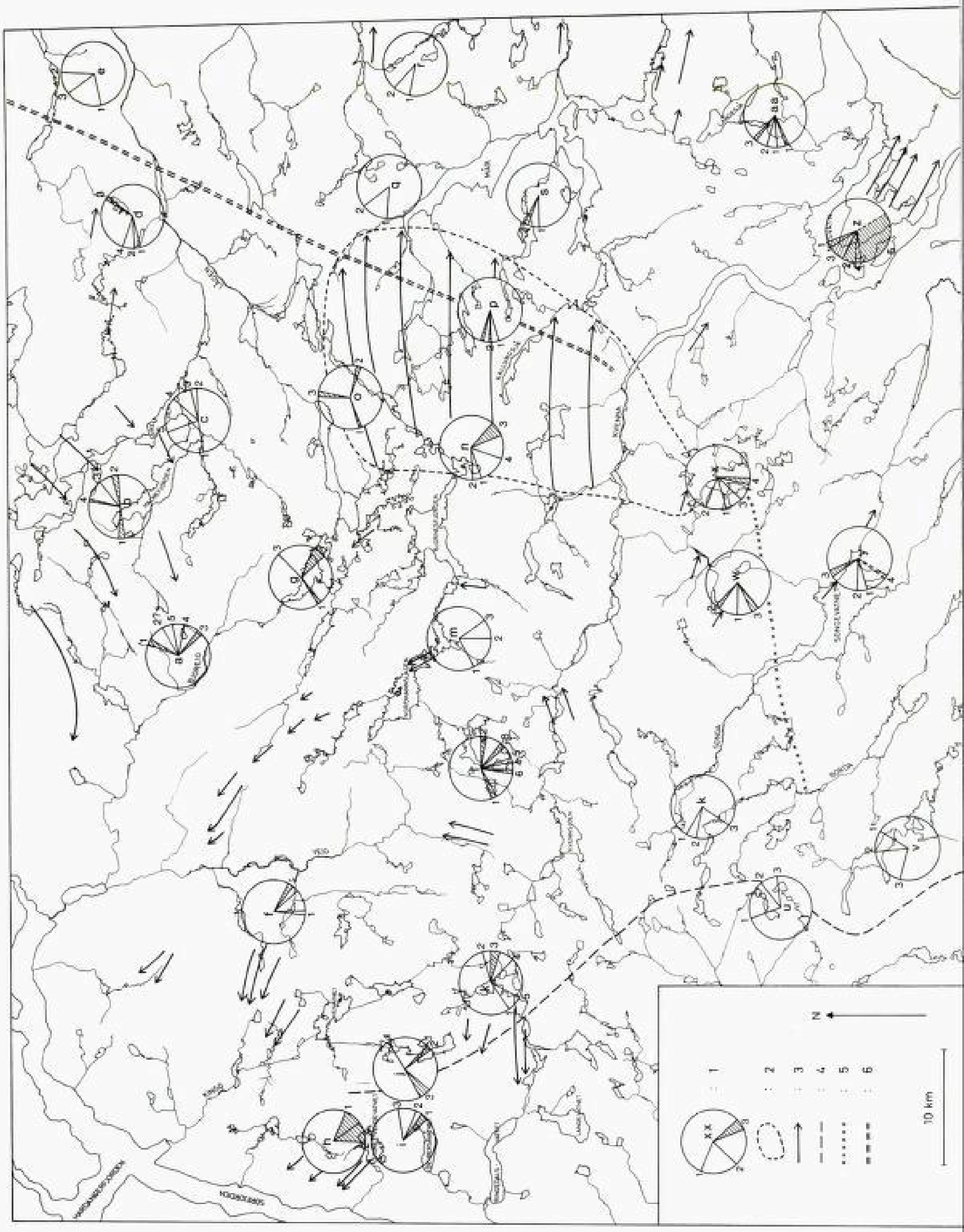

Fig. 11. Simplified map of directional elements. 1: Key localities, a–aa, showing age relationship between the different directions, 1 oldest, etc. The directions should be read from

dominant alignment of erosional oriented landforms, i.e. roches moutonnées, fluted rock and rock drumlins, is indicated. In the crag and tail areas this alignment coincides with that of the rock drumlins. Around the upper Kvenna the roches moutonnée alignments fall into two groups; forms in softer rocks are oriented northwards, those in harder rocks, eastwards. Throughout much of the central and eastern portion of Hardangervidda the superficial deposits effectively rule out such mapping.

The following may be concluded from Pl. 1 and Fig. 11:

1. No directional elements with a westward flow direction have been found east of a line from the northern end of Møsvatn to Skurdalsvatn.
2. No elements directed eastwards have been observed west of a line from Sandfloeggi to Tresfonn.
3. No northward flow directions have been recorded south of a line joining Nupsfonn with the northern end of Møsvatn, except for some elements of probably local character near the southern end of Songevatn and in parts of the Møsvatn district.

Within the central area enclosed by these limits there are clear signs of three major ice movements: the oldest towards the NE–E, an intermediate one towards the SW–W in the north and towards the W–NW–NE in the south, and lastly the youngest movement — towards the S in the north and towards the N in the south. As will be demonstrated later, there have actually been four main ice movements phases. From oldest to youngest, these will be called Phases I to IV. At a number of localities transition phases are also represented which cannot strictly be allocated to the above phases.

Phases I and II

The basis for subdividing the oldest directional elements into two main phases is partly stratigraphic. Phase II, as the younger of these two, has the best preserved directional elements and will therefore be dealt with first.

Throughout the central and eastern parts of Hardangervidda, Phase II is characterised by an eastward ice flow. It was during this phase that the extensive crag and tail area around Kallungsjå (Fig. 11) was formed, while farther east the older scouring is also directed to the east and a clear correlation is apparent. Similarly to the north of the crag and tail area there has been an older movement towards the east, i.e. localities b, c, d and e on Fig. 11.

On the western Hardangervidda the older scouring has a more northwards orientation north of Kvenna. Thus far, the correlation between the directional

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the periphery towards the centre. Shaded sectors indicate that the respective direction is found within the whole of this sector. 2: Large areas with crag and tails. 3: Predominant direction of directional elements in bedrock. 4: Western boundary of directional elements orientated towards the east. 5: Southern boundary of directional elements orientated towards the north, except directional elements at Møsvatn and Songa of supposed local origin. 6: Eastern limit of directional elements orientated towards the west.

elements seem to be clear. The observations gathered from the northwestern and southern parts of Hardangervidda are open to a more subjective evaluation.

In the northwest, locality f (Fig. 11) reveals that the oldest movement was towards the north. This direction may possibly correspond to Phase II, in which case one may envisage an ice divide from the southwestern part of Hardangervidda proceeding northeastwards towards Hardangerjøkulen.

North of Bjoreio, at locality a (Fig. 11), there is scouring directed SSW–SW which is older than Phase III. These scourings could belong to Phase II (sensu stricto) or alternatively represent the stage leading up to that Phase. The latter seems most likely.

Around the middle and southern parts of Møsvatn the scouring pattern is rather complicated. The age relationships between the different directions presented in Fig. 11, locality z, is a synthesis of several observations and thus incorporates a degree of subjectivity. Nonetheless, the connection between the various directions and the ice flow phases seems to be that directions 1 and 2 belong to Phase I. Direction 1 then probably represents the initial stage of Phase I. Phase II is probably represented by some of the southeastwards directions indicated by sector 3. The oldest of these directions is the most southward oriented one and presumedly marks the beginning of Phase II. Some of the directions in sector 3 also represent the succeeding Phase III. The crag and tails within this area must therefore have been formed during one or both of these phases. Direction 4 is then correlated with Phase IV while the directions in sector 5 reflect younger local scouring associated with the deglaciation of the area.

In the southwestern part of Hardangervidda, Damsgaard (1967) found that the oldest movement was towards the SSE, cf. locality u (Fig. 11). South of the western parts of Hardangervidda he has again found the same direction for the oldest movement. This may presumedly be correlated with Phase II.

Based upon the proposed correlations supplemented with the pattern of the other directional elements observed (Pl. 2), an overall picture of the Phase II ice movements is presented in Fig. 12. The most doubtful part of this reconstruction concerns the northwestern area, where the ice divide may have lain farther west than indicated.

Phase I's ice flow seems in general to have followed fairly similar directions to Phase II's. Scouring older than Phase II seems to have a somewhat more northward alignment on the whole though. Thus, in the Kallungsjå crag and tail area there is scouring whose orientation is more northwards than the crag and tails themselves. (Fig. 11, localities n and p). The same applies to localities d and aa, though local flow deviation may have been responsible at aa. At localities w and x there is again an older scouring (direction 1) which predates direction 2 which is correlated with Phase II (sensu stricto). Just north of Kvenna where the Phase II scouring is oriented ENE, there is an older scouring almost due east.

These observations of directional elements alone hardly justify attributing the very oldest scouring to a separate glaciation phase, except possibly near

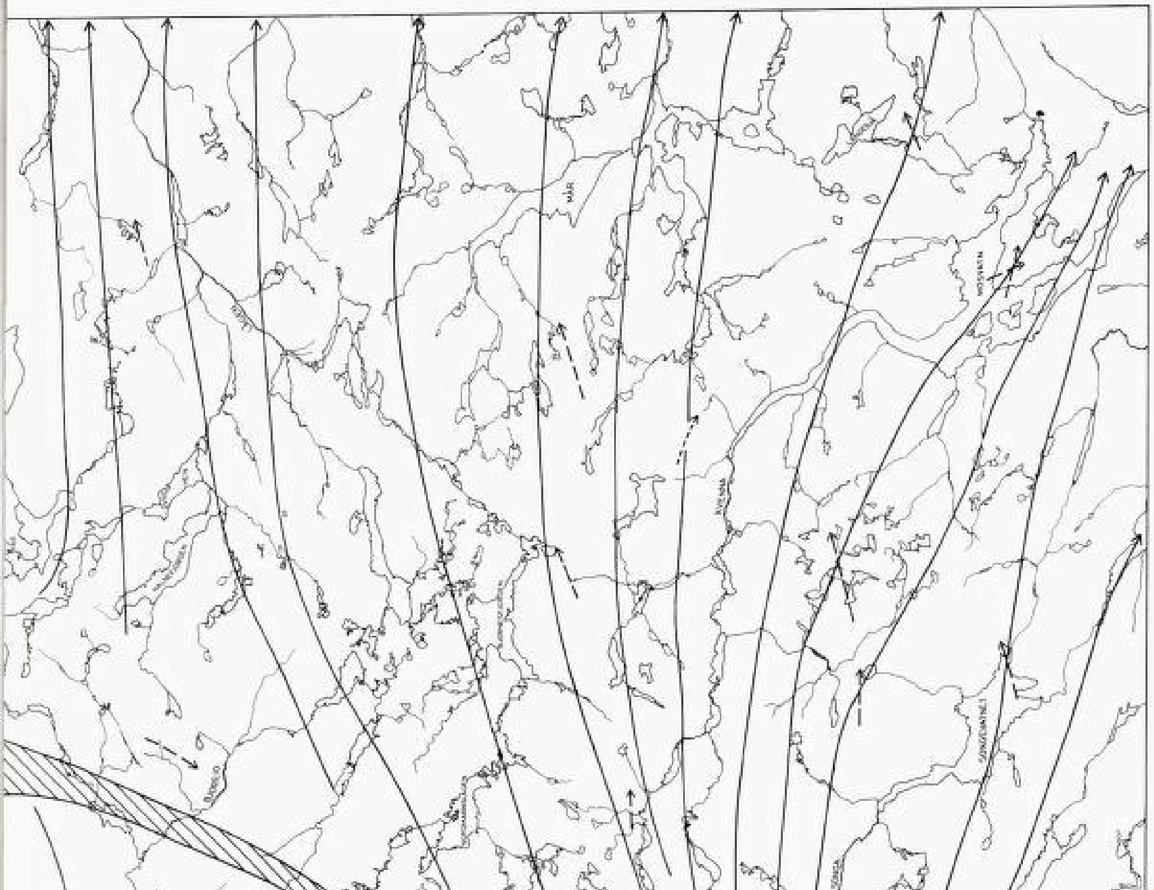

Fig. 12. Reconstructed ice movements during Phase II. 1: Ice divide. 2: Direction of ice movements during Phase II (sensu stricto). 3: Older ice movements. 4: Ice movements intermediate in age between Phase II and III.

Møsvatn where two initial phases are indicated. However, stratigraphical investigations suggest that prior to Phase II there was an ice-free period which itself was preceded by an eastward ice movement. The scouring just mentioned was most likely produced by this early ice movement, Phase I. Unfortunately observations of directional elements which can be attributed with reasonable confidence to this phase are too scarce to allow an overall reconstruction. As pointed out, the evidence suggests a similar flow pattern to that of Phase II, though perhaps with a domed ice surface located farther south than during Phase II (cf. Fig. 16), judging from the more northward component of the Phase I scouring.

Directional elements have also been recorded that are younger than Phase II but difficult to fit into the Phase III flow pattern. Within the crag and tail area around Kallungsjå a couple of localities have revealed scouring with a rather more southward direction. These doubtless reflect younger subphases of Phase II. The SW-oriented elements west of Ringedalsvatn are less easily accounted for, e.g. localities j and k, Fig. 11. Perhaps they belong to a short-lived transition to Phase III.

Phase III

This phase is characterised by ice streams converging towards WNW. Directions ascribed to Phase II include (Fig. 11):

directions 3, 2 and 3 at localities n, o and d,
directions 2 and 2 at localities m and g,
directions 2, 3 and 4 at locality c,
directions 2 and 3 at locality b,
directions 3 and 3 at localities w and x
and directions 2 and 2 at localities j and f.

Where several directions recorded at the same locality are grouped under Phase III, one must distinguish between the Phase III (sensu stricto) and its related subphases.

During the Phase III (s.s.), the ice divide lay relatively far to the east, presumedly near the eastern limit of westward-directed scouring (Fig. 11). In the northern zone of this limit the Phase III scourings are aligned almost parallel with the boundary, e.g. direction 3 at locality d, and direction 2 at locality e. The ice divide clearly lay between these two points. Farther south its position must have been close to the limit for N-oriented scouring (Fig. 11). In the west this ice divide meets a NW-SE aligned divide originally located by Damsgaard (1967). To the west and north of this ice divide, glacial flow has converged on Sørfjorden and Hardangerfjorden, as witnessed by abundant observations of directional elements (Pl. 1). East and south of this ice divide the directional elements associated with Phase II and III are virtually indistinguishable except in the northeast where, as already mentioned there was a southward flow during Phase II. Farther south the flow directions of the two phases must have coincided, i.e. have been towards the east or a little south of east. This should give the reconstruction of Phase III (s.s.) shown in Fig. 13.

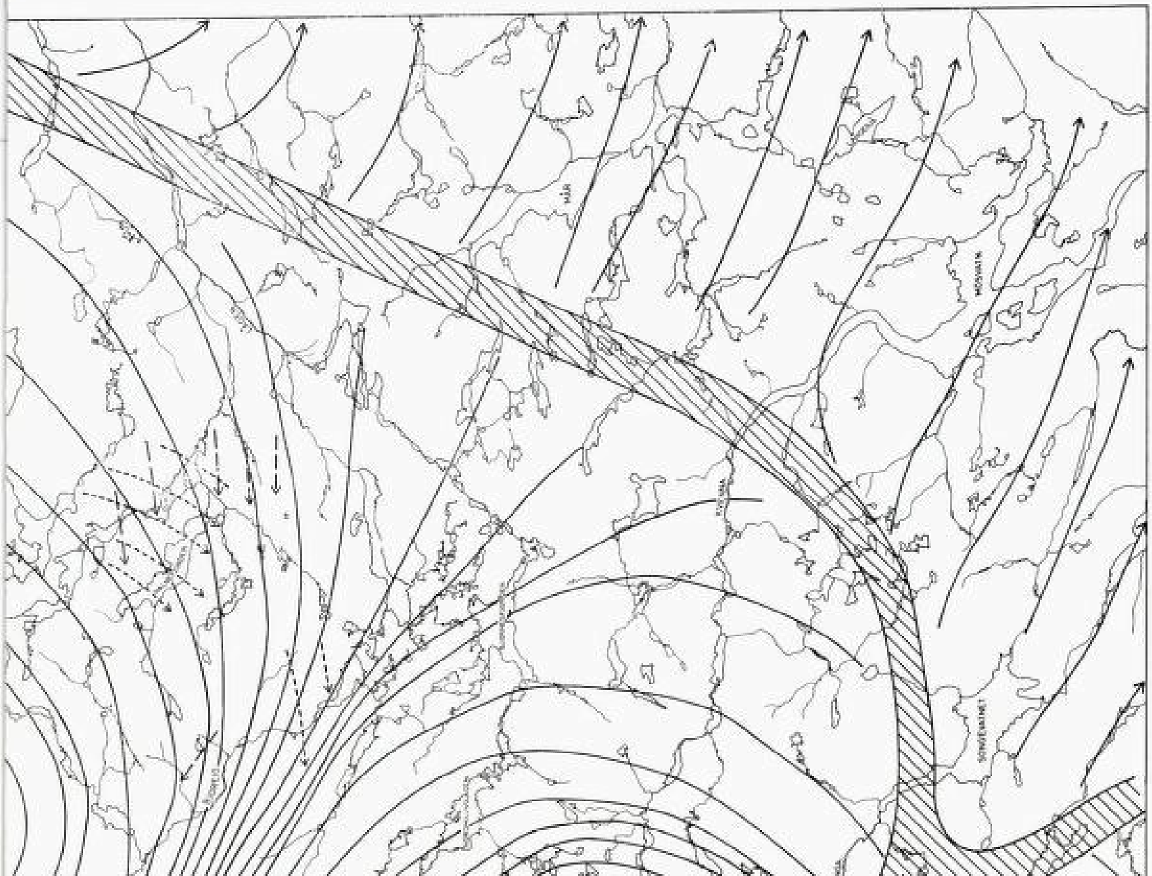

Fig. 13. Reconstructed ice movements during Phase III. 1: Ice divide. 2: Direction of ice movements during Phase III (sensu stricto). 3: Older ice movements which are younger than Phase II (sensu stricto). 4: Ice movements intermediate in age between Phase III (s.s.) and Phase IV.

Suphases of Phase III (s.s.) are particularly distinct in the northern areas. Around Halnefjorden there were older ice movements with a more westward direction than Phase III (s.s.). Furthermore, this same area had later movements more towards the south, marking the transition to Phase IV (cf. Fig. 5).

The southwestward flow in the Ringedalsvatn area, represented by directions 1 and 2 at localities j and k (Fig. 11) was mentioned earlier. Direction 2 at locality i may possibly be related to them. Damsgaard's (1967) material seems to indicate southward flow within the southwestern areas during both the transition period between Phases II and III and that between Phases III and IV.

Phase IV

Phase IV is believed to be the youngest ice movement phase on Hardangervidda. It was characterised by an extended ice divide in the west, located across the highland area (Fig. 3). In the northeast the ice flowed northeastwards from this ice divide, e.g. directions 3 and 5 at localities j and k (Fig. 11). West of this ice divide flow was westwards, and detailed mapping revealed a marked topographical control.

From an area a little east and north of Hardangerjøkulen, the ice radiated southwards with a fan pattern, e.g. directions 4, 5, 3 and 4 at localities b, c, g, and d, respectively (Fig. 11). This movement met opposing ice streams emanating from the southern parts of Hardangervidda. Its directions closely followed that of Phase III at a number of places.

A special problem is presented by the area lying between Mår and Bjornesfjorden. Despite very careful searching, virtually no trace was found of any directional elements younger than Phase III. Either this area was deglaciated after Phase III or the basal ice masses were dynamically inactive during Phase IV here. Since the circumstances at Mår (cf. Fig. 10) show ice moving out of this critical area during Phase IV, the second alternative is presumedly correct. This explanation would also account for the preservation of the large Phase II crag and tail area centred on Kallungsjå, as during Phase III the ice divide lay there so that ice movement was minimal at that time too (see p. 28). The fact that Phase III is represented there only by fine scratches supports this. That some zones of the ice sheet were not reactivated by the changing of ice flow patterns from Phase III to Phase IV, is taken to indicate a fairly short duration for Phase IV.

Within the eastern area, Phase IV movements can in part be traced in detail from the fluted surfaces (Pl. 1), which correspond to the youngest scouring direction (cf. Fig. 10). Ice flow has greatly been influenced by topography – differences in relative relief as small as 200 metres have locally affected the direction of movement.

When the proposed correlations are supplemented with the remaining directional elements, a composite picture of the Phase IV ice flow pattern appears, as presented in Fig. 14. It will be seen that a correlation has also been made with the Preboreal Eidfjord–Osa glacier advance (Anundsen & Simonsen

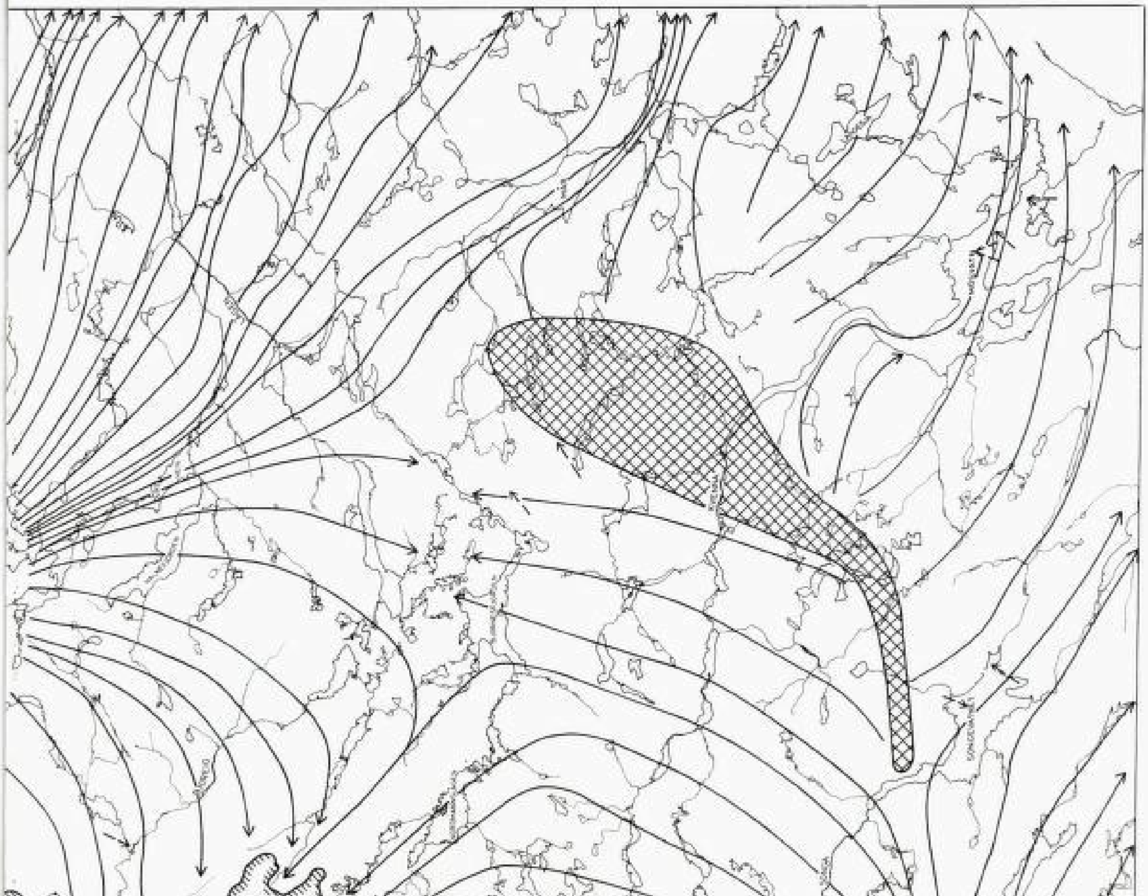

Fig. 14. Reconstructed ice movements during Phase IV. 1: Ice divide. 2: Area with dynamically inactive basal ice. 3: Direction of ice movements during Phase IV (sensu stricto). 4: Younger local directions. 5: Ice margin during the Preboreal Eidfjord–Osa event.

1968) in the areas around Bjoreio and Veig. Two circumstances support this correlation. Firstly both the Eidfjord–Osa advance and Phase IV represent the last period of ice sheet activity; secondly there is an obvious relationship between the ice movement directions and the end moraines.

From Veig to Sørfjorden, though, it was impossible to trace Anundsen & Simonsen's (1968, Pl. 3) reconstructed ice margin which they plotted straight across western Hardangervidda. Directional element data strongly indicates that an ice divide covered this area at the time. My investigations of the deposits also cast some doubt upon their reconstruction. Their marginal deposits near Øvre Bersåvatn could not be found, while the one they plotted at the southern end of Omkjelsvatnene, south of Veivatn, must either represent kame deposits by Øvre Omkjelsvatn, cf. Reusch et al. (1902), or a glaciofluvial/fluvial fan at the inlet of Nedre Omkjelsvatn. None of these features can be related with any degree of certainty to an active ice front. I believe that the well-known marginal deposit at Kinsarvik (Undås 1947), at the mouth of the Kinso, is more likely to mark the ice front position at that time.

Directional elements younger than Phase IV seem to be of local character. Near Songevatnet, younger ice movements led into the lake basin. The scouring towards northerly directions in the Møsvatn district is probably quite local, caused by the flow of minor glacier remnants during deglaciation. Anundsen & Simonsen (1968, p. 26) found apparently younger directional elements north of Bjoreio, while Rye & Follestad (1972, p. 28) describe younger scouring in the direction of the Kvenna valley, i.e. towards the east.

Comparisons with earlier investigations

Until the 1960's there were two views on the ice movements across Hardangervidda. Kjerulf (1879, p. 34), Hansen (1888, Map 2), Reusch (1894, p. 57) and Rekstad (1903, p. 41) claimed that all ice flow had been governed by an ice divide which essentially coincided with the main watershed (cf. Fig. 3). Werenskiold (1910b, p. 16), Isachsen (1933, p. 429) and Holmsen (1955) seem to have supported this concept. Hansen, however, changed his mind later and proposed that the final ice divide lay 'to the south of the large lakes here, Møsvand, Totak and Vinjevand' (1890, p. 273; 1892, Pl. C and 1895, p. 186ff.).

A more refined model was presented by Rye & Follestad (1972). In the writer's opinion their oldest phase is a combination of movements of varying age, namely from Phase II and III. The reason for this, in the writer's opinion mistaken, reconstruction is probably that they have not observed the oldest scouring directed towards E in the northern part of Hardangervidda. Their succeeding phase also has components of different ages – according to the present interpretation, a combination of movements from Phases III and IV and their subphases. Part of the explanation for their interpreting it in this manner may be that they have accredited the scouring within the Lågaros map sheet to an ice flow towards the SE instead of the NW. Their youngest phase

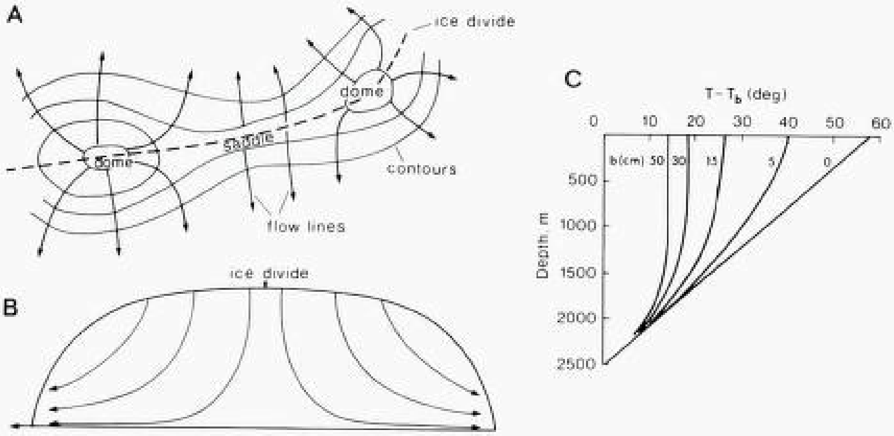

Fig. 15. A: Sketch map indicating trend of surface flow lines, names of relief forms and position of main ice-divide.
B: Cross-section of an idealized ice sheet overlying a horizontal subsurface. Ice particle paths are indicated by long arrows (modified from Dansgaard et al. 1971).
C: Steady state temperature distribution in an ice sheet 2500 m thick, for various values of net mass balance (b) in cm of ice. T = average surface temperature in °C, Tb = basal temperature in °C (from Patterson 1969).

which they have only reconstructed for the northern portion of Hardangervidda, is virtually coincident with the writer's version.

Sollid (1975, Fig. 1) has investigated ice movements on the northeastern part of Hardangervidda. His reconstruction of the flow pattern from 'the glacial maximum' shows affinities with the present Phase III, but the ice divide is about 15 kilometres too far west. Sollid's youngest phase, where presented, is virtually similar to the writer's youngest phase, Phase IV.

SOME CHARACTERISTICS OF THE WEICHSELIAN ICE SHEET ON HARDANGERVIDDA

Certain characteristics of the Weichselian ice sheet on Hardangervidda will now be considered, primarily its relief and temperature conditions and the cause of the ice divide displacement.

General discussion

Ice movement and surface relief. According to glaciological theory the surface ice layers will essentially flow normal to the glacier surface contours (Fig. 15A). This has been demonstrated, for example, on Wilke's ice dome in Antarctica (Budd & Morgan 1973, p. 11). Theoretically the flowlines of the basal layers should be vertical projections of those in the surface layers. However, Haefeli (1963) found deviations from this model when he surveyed a minor ice cap, Jungfrauhoch, where the ice summit was offset relative to the basal ice divide. Nevertheless, in the following reconsructions of relative relief the theoretical model will be adhered to. Thus basal flowlines determined from the distribution of directional elements (Figs. 12, 13 and 14) are assumed to reflect the surface flow and relief. The terms concerning relief forms given in Fig. 15A will be used in the text.

At dome summits there are only vertical movements (cf. Fig. 15B) while low rates of horizontal flow will occur from domes towards saddles. If one envisages a continental ice sheet in a steady state, with uniform accumulation, the horizontal flow rate will increase from zero at the dome summits to maximum values at the equilibrium line (Paterson 1969). Since ice sheets usually have little accumulation near the domes and yet are very thick, it follows that the absolute flow velocity in such areas must be slight. Borehole data from both Greenland and Antarctica, where the basal layers proved to be of pre-Weichselian age (Johnsen et al. 1972) support this.

Temperature conditions. Temperature conditions within a glacier are important in several connections, including from a glacial geologist's viewpoint (cf. Schytt 1974). Of prime significance is whether the basal ice layers are below or at pressure melting point, i.e. whether the glacier is cold- or wet-based. Schytt (1974) has considered the temperature conditions in the Weichselian ice sheet in the light of comparisons with Greenland and Antarctica, and says that, ... 'the ice sheet may have been cold and frozen to the bed over at least the major part of its area. It may have been 'warm' at the bottom in the southernmost part, under some central areas with maximum ice thickness, and under the outlet glaciers draning the ice sheet through the Norwegian fjords'.

There are several approaches to the temperature problem based, respectively, on theoretical considerations, surveys of the two recent continental ice sheets, Weichselian glacial landforms and deposits, or a combination of these. Robin (1955) has theoretically derived formulae for the temperature distribution close to the ice divide in steady state ice sheets. The temperature of the basal layers is primarily governed by the thickness of the ice, its surface temperature, net mass balance and geothermal heat supply. Increasing surface temperature and ice thickness cause an increase in the basal temperature. Larger net mass balance lower it (cf. Fig. 15C). Modifications must be made to Robin's formula if frictional heat is being generated at the sole of the ice sheet. Other amendments have to be applied to the formula if the ice sheet in question is not in a steady state, cf. Jenssen & Radok (1961) and Wexler (1959).

Up to now two boreholes, in the Greenland and Antarctic ice sheets respectively, have reached the bottom. At Camp Century in Greenland the ice was 1387 metres thick with a basal temperature of ÷13°C, far below its pressure melting point (Hansen & Langway 1966). The boring at Byrd Station in Antarctica revealed an ice thickness of 2164 metres. Water was registered at the ice/rock interface, which showed that the temperature there was at the pressure melting point, ÷1.6°C (Gow et al. 1968). Both of these temperature observations agree well with theoretical values which could be derived from Robin's (1955) formula (Robin 1972).

In order to determine whether previously existing ice sheets were cold- or wet-based it is necessary to know which glacial geological phenomena are characteristic for these two states. A number of direct observations have been made beneath modern wet-based glaciers, confirming that erosional forms such as scouring and roches moutonnées are produced and that basal till is being deposited, e.g. Carol (1947), Boulton (1970), Vivian & Bocqet (1973) and Boulton et al. (1974).

Observations from cold-based glaciers are much more scarce. Most authors have claimed that such glaciers have a protecting effect, e.g. cf. statements like: 'Alltogether, under cold glaciers frozen to their beds, bedrock erosion probably does not occur', (Embleton & King 1968, p. 12); '. . . and therefore the glacier cannot erode' (Schytt 1974); '. . . bedrock is protected when the glacier is frozen to it' (Patterson 1969, p. 168). These remarks are supported by observations at the base of a cold glacier on Greenland, where Goldthwait (1960, p. 47) found boulders with lichen. Boulton (1972) is less categorical about the erosional capabilities of a cold-based glacier and maintains that differential movements in the lower ice layers can abrade protrusions with englacially transported particles. The existence of englacial material in ice sheets frozen to ground was proved by the Greenland borehole – the lowermost 16.9 metres contained minerogenic material (Hansen & Langway 1966).

Another conceivable erosion mechanism which Boulton (1972) has drawn attention to, is the plucking of large erratics of unconsolidated material where the latter are frozen to a shallow depth, below which the pore water pressure is high. Mercer's (1971) observations on Antarctica tend to support the concept of abrasion and plucking beneath cold-based glaciers under special conditions.

As far as sedimentation below cold-based glaciers is concerned, Boulton (1972) proposed that it might occur against larger obstructions. Nobles & Weertman (1971) believe that one requirement for basal till deposition is that the basal ice layers melt; this would exclude sedimentation from cold-based ice.

Summarising, one may say that if parameters like surface temperature, ice thickness and net mass balance were known for previously existing ice sheets, it would be feasible to derive their temperature conditions by means of glaciological theory. However, such knowledge is as yet quite inadequate and there are insufficient glacial geological observations from modern ice, especially the cold-based variety. Nonetheless, there is apparently agreement about erosional capacity being drastically reduced with a cold-based glacier. Abrasion of protruding rock may occur and plucking of large erratics as suggested by Boulton can probably take place too. With respect to basal till deposition it is doubtful whether it happens on a significant scale beneath cold-based glaciers.

Phase I and II

Phase II will be treated first, as the conditions which prevailed then are better known.

Phase II. The glacial relief reconstruction for this phase (Fig. 16) shows a

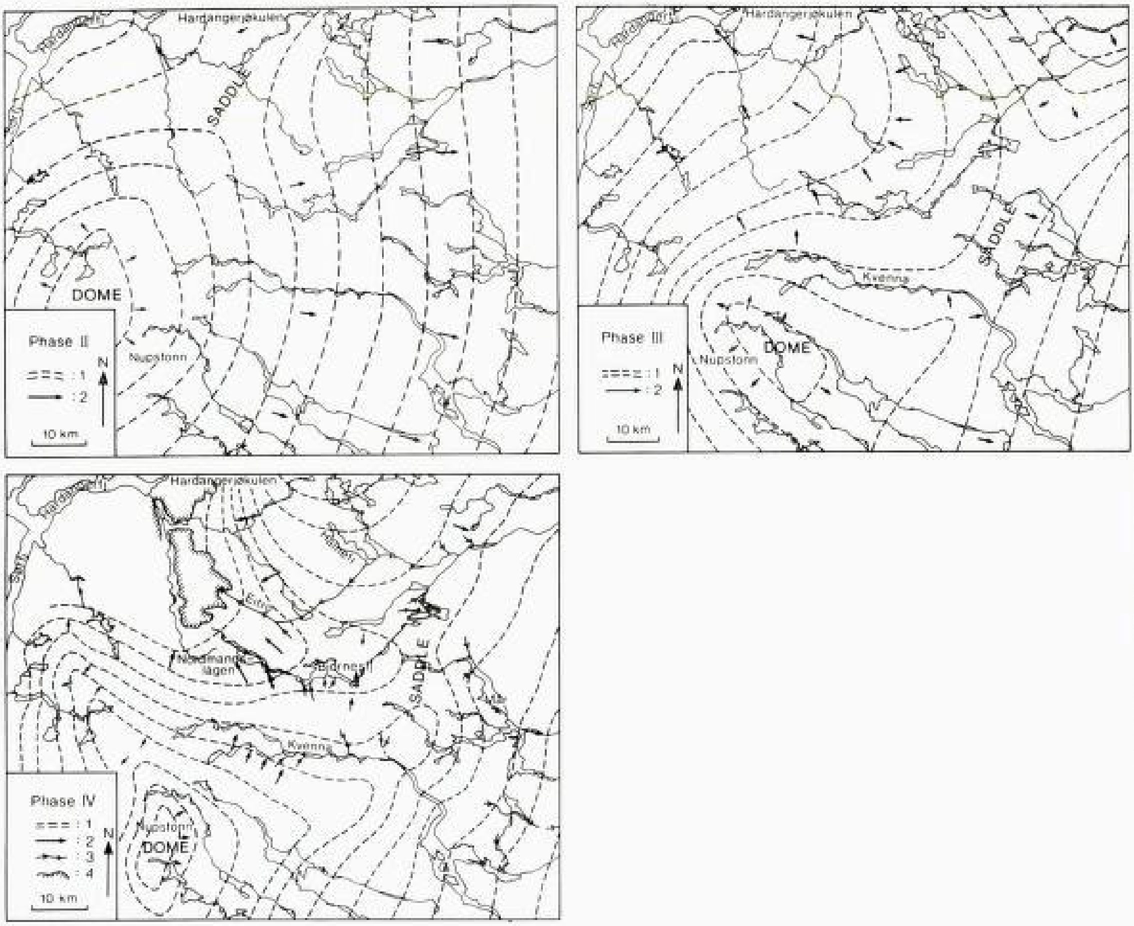

Fig. 16. Form lines (1) indicating relief of the ice sheet surface during Phase II, III and IV. The form lines are not equidistant. The arrows (2) indicate very approximately the relationship between velocity in different part of the ice sheet. Subglacial drainage routes (3) during Phase IV and deglaciation time are indicated (cf. Pl. 2). Ice margin (4) during Phase IV is indicated after Anundsen & Simonsen (1968).

continental ice sheet stage. The ice surface was characterised by a dome over southwestern Hardangervidda. A lower-lying ridge stretched from this dome towards the Hardangerjøkul area, where there may have been another dome. This relief indicates that the continental ice sheet built up in the western areas.

The glaciation limit today lies between 1500 and 1600 m.a.s.l. in the west of the area and at over 1800 m.a.s.l. in the east (Østrem 1964, Fig. 71). The rise eastwards is caused by lower precipitation and higher summer temperatures. Annual precipitation is reported to vary from over 2000 mm in the west to about 600 mm in the east (Østrem & Ziegler 1969, p. 13; Utaaker 1971, p. 27). During winter the precipitation-bearing wind today come chiefly from the southwest (Andersen 1973, p. 7).

The initiation of Phase II was probably a deterioration of the present type of weather pattern, with increased winter precipitation and/or a lowering of summer temperature. This would lead to the western areas being glaciated first. Existing ice caps such as Hardangerjøkulen and Nupsfonn would have expanded and new local glaciers have formed in the west. (The cirque topography there most likely developed during this and similar periods of preliminary glacier build-up.) Eventually the small local glaciers would merge into a single ice cap covering western Hardangervidda. A lowering of the present glaciation limit by only 200 metres would suffice to create such a situation, and with a further lowering the ice caps would expand in all directions, glaciating the entire Hardangervidda.

The location of the southwestern dome (Fig. 16) is rather remarkable in that it lies west of Nupsfonn which in the southwestern area was probably the original centre for glacier formation. This is perhaps due to an increase in the east-west gradient in weather conditions as the glacier grew, so that a greater proportion of the precipitation gradually fell in more westerly areas.

When Phase II was fully developed, landforms such as roches moutonnées, scouring and crag and tails were created. Basal till deposition was sometimes quite considerable, especially in the eastern areas (p. 63). According to the preceding discussion, this is evidence that the ice was wet-based. Thrust slices near Trengsle (p. 72) suggest, though, that the glacier was cold-based in an early phase. This interpretation is dependant upon which mechanism is responsible for thrust slices. Most investigators have assumed that the material was frozen (Moran 1971, p. 134; Boulton 1972, Banham 1975). However, Moran (1971) believed that if the pore water pressure in the underlying layer is sufficiently high then this is no longer a decisive factor. Nevertheless, it does not seem unreasonable that the ground could have been frozen at several places during the onset of Phase II, so that the ice was cold-based in the very early stages. Later on, the basal parts of the glacier could have been warmed up by geothermal and frictional heat, so that the ice became wet-based.

Phase I. Directional elements which can fairly reliably be referred to Phase I are so rare that no detailed reconstruction of flow lines and glacier relief is possible. A general similarity with Phase II may be discerned, though with a more northerly alignment of the directional elements, suggesting an ice dome located farther south than during Phase II. Ice sheet growth in Phase I has probably followed a similar pattern to that described for Phase II above. Indications of thrust slices transport (see chapter on 'Stratigraphy') may relate to at least local cold-based areas of the ice during the Phase I advance.

Phase II/III transition

The transition from Phase II to Phase III was apparently rapid. No gradual change in directional elements can be attributed to this transition period. A replacement of the westerly ice divide by an easterly one has long been recognised for central Scandinavia, and various theories have been proposed to explain this. Högbom (1885, p. 37) invoked an increase of snowfall from westerly winds over the eastern area, on account of the lower temperature there. He also stated that the ice divide's location is always governed by the (mechanical) equilibrium between movement in the two directions — Högbom's law, cf. Ljungner (1949). Enquist (1917, p. 34) explained the shift of the ice divide through precipitation from westerly winds falling on the lee side of the ice sheet axis which was thus displaced eastwards. Ljungner (1945; 1949) explained the ice divide migration as due to cyclones tracking along the southern and eastern sides of the ice sheet as it became larger, which would lead to increased precipitation in the east and consequently a pushing of the ice divide eastwards (cf. Liljeqvist 1957, p. 66).

There probably has been a change in the cyclone tracks. As the ice sheet grew there is reason to presume that the precipitation-bearing cyclones from the west have more frequently taken a more southerly course, along the southern margin of the ice sheet (Liljeqvist 1974). This has led to greater precipitation in those parts but the effect must have rapidly diminished eastwards (Liljeqvist 1974). Is is questionable then whether this was sufficient to cause the ice divide to migrate eastwards.

Weertman (1973, p. 360) has shown theoretically that a very considerable change in annual precipitation on different parts of the ice sheet is required before the ice divide will be displaced, assuming the ice sheet maintains constant area. This, together with reference to present-day conditions in Greenland and Antarctica where the bulk of the precipitation is concentrated in peripheral areas, led Schytt (1974) to point out that the ice divide's location is more a function of the extent of the ice sheet rather than areas of maximum accumulation. He therefore says that,'If . . . the edge is fixed on one side (such as along the Norwegian coast), while moving eastward on the opposite side, the ice divide has to move east as a result of the balance between accumulation and ice movement, which here is the only output factor'.

Summarising, the migration of the ice divide eastwards in central Scandinavia has been explained in terms of changing accumulation conditions and mechanical equilibrium considerations. None of these theories seems to account saticfactorily for the apparently sudden displacement of the ice divide which has occurred on Hardangervidda. A similar event has been reported from Jämtland (Lundqvist 1969). In the writer's opinion there are two possible explanations. The first is that the ice was cold-based during the Phase II/III transition period and did not produce any observable directional elements. During that period the ice divide can have moved eastward for some reason. The other possibility is that the ice divide was displaced eastwards during a large-scale surge out via Hardangerfjord. These two alternatives will be examined more closely.

Since during both Phases II and III the ice sheet was probably wet-based (p. 31, 35), the first alternative implies a wet-based → cold-based → wet-based sequence. The transition from wet- to cold-based can occur with a lowering of surface temperatures, a reduction in ice thickness or increasing net mass balance.

If one presumes that Phase III represents the Weichselian maximum of c.

20.000 BP (p. 104) it is clear that the climate changed between Phases II and III. It is generally agreed that temperaturewise the Weichselian maximum was at the minimum for the whole Weichselian (cf. Fig. 59). Most authorities also claim that this period experienced a relatively continental climate (Nilsson 1972, p. 235). Briggs et al. (1975) have, however, found that the climate of the English Midlands around 27.000 BP was both colder and less continental. Even allowing for the distant locations to which these palaeoclimatic calculations apply, it seems reasonable to assume that there has at least been a fall in temperature between Phase II and Phase III. This still leaves the question of whether this could have influenced the basal ice temperature.

Since ice is a poor conductor of heat, this mechanism alone will not suffice to transfer the reduction in surface temperature through the ice sheet to its base. It would require actual downward movement of ice, i.e. convection. Bearing in mind the low velocity of ice near the ice divide, the efficiency of such a mechanism may be doubted too. The pre-Weichselian ages of the basal layers of the Greenland and Antarctic ice sheets support this. The idea that the ice sheet, having become wet-based, would be able to revert to a cold-based state in the time available, does not seem acceptable.

Turning to the second alternative, it should be noted that surges have been reported from different types of glaciers and climatic regions (Meier & Post 1969, p. 811). One can mention Vatnajøkulen, Iceland (Thorarinsson 1969); Barnes Ice Cap, Baffin Island (Løken 1969); Bråsvellbreen, Sørfonna on Nordaustlandet, Svalbard (Schytt 1969); Otto Glacier, Ellesmere Island (Hattersley-Smith 1969) and several outlet glaciers on Spitsbergen (Liestøl 1969). It is therefore reasonable to expect surging in ice sheets, too. This has also been suggested by T. Hughes (Bull & Webb 1973) who concluded from surface profile measurements from the Ross Ice Shelf to Byrd Station that a surge has occurred during the last 100,000 years. Wilson's (1964, 1969) ice age hypothesis based on gigantic surges of the Antarctic ice sheet is also known.

The phenomenon of glacial surging has yet to be fully explained. For a surge to occur there must clearly be:

a) a damming effect causing the basal stress in the glacier to exceed a critical value,
b) a releasing mechanism,
c) a mechanism which caused the glacier to move at a relatively much higher velocity.

The damming effect may apparently be due to:

1) topography,
2) stagnating ice in the marginal zone (Nielsen 1969),
3) cold ice in the outer zone damming up temperate (wet-based) ice in the inner parts (Schytt 1969),
4) cold ice which itself reduces flow velocity, with the surge occurring at the transition to temperate ice (Robin 1955),
5) glacier ice, e.g. a main valley glacier which dams up tributary glaciers (Nielsen 1969, Stanley 1969).

Robin (1969) has summarised possible releasing mechanisms. He considered the following:

1) temperature change, with transition from cold- to wet-based ice (Robin 1955, 1969),
2) changing stress conditions; in the accumulation zone a change from extending flow to compressive flow can start a surge, while the opposite transition would do so in the ablation zone,
3) increase of subglacial water film to a critical value which, according to Weertman (1969), must equal 'the controlling obstacle size' which varies from 1 to 6 mm,
4) damming up of sufficient water in the release zone due to zero water-pressure gradient (Robin & Weertman 1973),
5) avalanche on the glacier, triggered by earthquake.

The latter theory seems, at the very least, to lack general applicability (Patterson 1969).

The relatively high velocities of surging glaciers seem to be attributable to rapid basal sliding. This is achieved through greater availability of water, either in the form of a continuous film (Weertman 1969) or by the flooding of subglacial cavities as suggested by Lliboutry (Nye, in 'Discussion' of Weertman 1969). Both mechanisms would reduce basal friction and thereby increase ice flow velocity.

Returning to the Phase II/III transition, one can envisage the threshold at the entrance to Hardangerfjord (see Fig. 38 in Holtedahl 1965) having a damming effect during the ice sheet's growth in Phase II. Alternatively, it is conceivable that the peripheral zone was cold-based. Either way, once basal stress passed a critical value a surge was initiated, in turn giving increased basal melting. The water so produced helped reduce basal friction, and thus the surge continued until the thinning of the reservoir ice masses and the lowering of the glacier's surface gradient in the receiving area combined to lower the basal stress once more (cf. Meier & Post 1969).

The consequence of a large-scale surge down Hardangerfjord would be a draining of ice from the western areas of Hardangervidda. The elevation of the ice surface would then be lowered and a new ridge (i.e. ice divide) would be established farther east. This can thus account for the ice divide migration between Phases II and III. The WSW-aligned directional elements to the northwest of Ringedalsvatnet (cf. Figs. 12 and 13) may be explained by this development too. Since the ice surface in the Hardangerfjord area itself was lowered, there may have been a secondary surge down Sørfjorden which would have drained ice from the adjacent parts.

Phase III

The shape of the ice sheet surface during Phase III (sensu stricto), Fig. 16, was characterised by an almost stationary dome between Nupsfonn and Kvenna. From this there was a glacier surface ridge stretching away towards the SSW while another described an arc towards the E and NE. Observations of directional elements (Fig. 13) show that the latter ridge was relatively constant in height in the early stages of Phase III. It was subsequently modified as a saddle area developed between the main dome and a second one which lay to the northeast of the investigated area before gradually migrating westwards during the later stages of Phase III (Phase III/IV transition).

The ice movements from the ice divide show a convergence on Hardangerfjorden, producing relatively high velocities there. Within the divide area the rate of flow was slight, as demonstrated by the scouring, which consists only of fine scratches, and by the survival of Phase II crag and tails.

Occurrences of roches moutonnées, scouring, basal till and subglacially deposited sorted sediments bear witness to a wet-based ice sheet during Phase III.

If certain conditions are fulfilled one may calculate the ice thickness, mean annual temperature and net mass balance at the ice divide during Phase III. These were found (Vorren 1977b) to be 2500–2800 metres, c. –25°C and c. 150 mm (ice), respectively.

Phase IV

In the period between Phases III (sensu stricto) and IV the dome in the north gradually migrated westwards. During Phase IV its centre probably lay north of Hardangerjøkul (cf. Sollid 1975, Fig. 1B). On the western part of Hardangervidda during Phase IV (Fig. 16) there was a ridge (i.e. ice divide) running N–S and generally following the topographical ridge there (cf. Fig. 3). There was another dome above Nupsfonn and a saddle area between Mår and Bjornesfjorden.

The relief and ice movement of Phase IV were partly inherited from Phase III and partly new. In between these two Phases there was probably a period of prolonged ablation with a general lowering of the ice sheet surface accompanied by a retreat of the marginal zones. The ice surface lowering was relatively greater in the eastern area, resulting in the ridge being created in the west. This did not completely erase the eastern ridge, however; a remnant was left between Mår and Bjornesfjorden. Caught between ice masses which were dynamically rejuvenated during the Preboral climatic deterioration, this part remained dynamically inactive in its basal layers. This was obviously an unstable state of affairs and would doubtless have been modified if the ice sheet as a whole had not experienced a renewed and marked downwasting after the short-lived Preboral phase.

Large-scale subglacial drainage during Phase IV is witnessed by eskers and also by other landforms in both bedrock and superficial deposits. The ice sheet was therefore wet-based during this Phase too.

The largest eskers, the Halne-Eitro esker and the Nordmannslåg esker (Figs. 16 and Pl. 2), tend to follow the direction of ice movement but also seem to have been influenced by topography. Both these drainage systems were clearly fed by meltwater from the central Hardangervidda. Perhaps much of

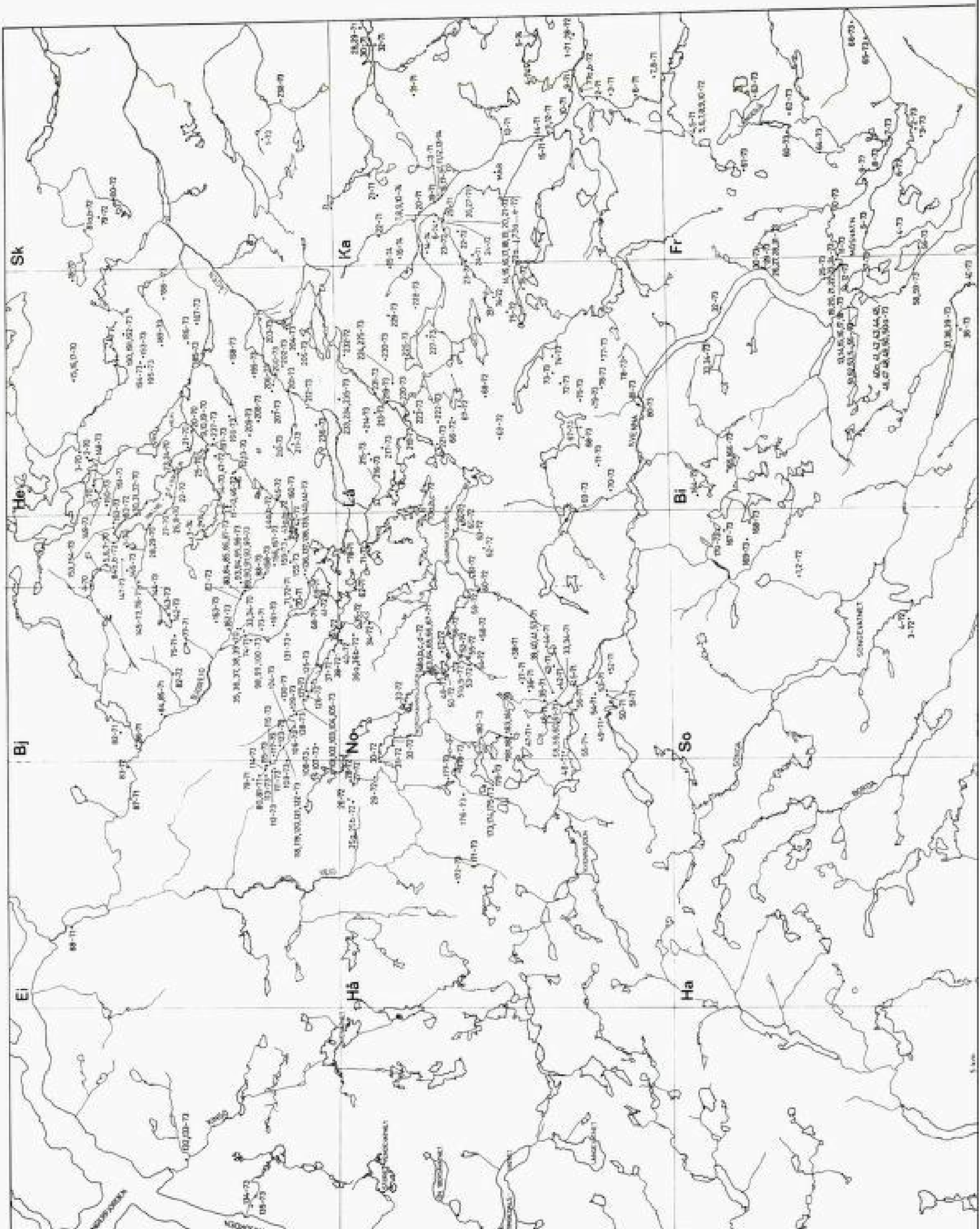

Fig. 17. Map of Hardangervidda showing location of map sheets. The map names are abbreviated to their first two letters: Ul = Ullensvang, Ei = Eidfjord, Bj = Bjoreio,

the ice carried to the central part of Hardangervidda by the two 'colliding' ice streams escaped as subglacial/englacial meltwater?

The drainage systems north of Møsvatn and Kvenna were in part directed against the Phase IV ice movement. This may have been due to topographical factors or to their development later on during the deglaciation.

The final deglaciation of Hardangervidda seems, essentially, to have occurred by downwasting. Traces of lateral drainage at a number of localities indicates that the lower-lying areas became ice-free last. The deglaciation must have been completed by 8460±190 BP which is the age obtained for organic deposits on the eastern vidde (Moe 1977).

Sediments

METHODS

Mapping

The mapping (Pl. 2) was done by means of both field observations and aerial photograph study. Sediment types were determined partly from sections and partly from the geomorphology. The eastern two thirds of the Skurdalen and Lågaros map sheets (Fig. 17) are, with slight modifications, based on Holmsen's map (1955).

The object of this mapping was primarily to provide a regional picture of the clastic sediment distribution. Organic deposits are thus omitted, as are landforms such as crag and tails and fluted sediments. By combining Pl. 1 and 2 one may nevertheless see the distribution of these landforms.

Sampling

Location and type of samples are shown in Fig. 17 and Table 1. Samples were taken from exposures and, more commonly, from specially excavated sections and pits. Slumped material was avoided.

With the exception of five samples of weathered material, sampling was mostly restricted to the C-horizon; the few exceptions were at localities where an excessively thick B-horizon prevented this. In practice this applied to some schist-rich deposits where there had been extensive precipitation of iron.

Samples taken from stratified sediments were, as far as possible, taken from individual units. Samples from unsorted deposits, diamictons (Flint et al. 1960, p. 1809), were taken as representatively as possible for the deposit. Sample size was 1 to 2 kg from diamictons and usually 0.3 to 1 kg from sorted sediments. During sampling the colour, structure, approximate textural composition, depth, degree of consolidation and any important miscellaneous parameters were recorded.

Laboratory processing

Grain-size analyses. The samples were put through a Soil-test CL 280A splitter with 13 mm aperture. Larger grains were apportioned visually Diamictons were normally split down to 0.3 to 0.5 kg. For sorted samples a sample of 0.1 to 0.2 kg was used. Prior to sieving it was necessary to treat 7 samples with H_2O_2 (Ingram 1971, p. 58) to remove humus.

The samples were then wet-sieved through a 1.0 mm or 0.5 mm (0 or 1 phi) screen and a 0.063 mm (4 phi) screen. After drying, material coarser than 4 phi was dry-sieved through sieves with whole phi intervals. The upper grain-size limit used in subsequent calculations was 16 mm (–4 phi). The residue finer than 4 phi, collected from both the wet- and dry-

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He = Hein, Sk = Skurdalen, Ri = Ringedalsvatn, Hå = Hårteigen, No = Nordmannslågen, Lå = Lågaros, Ka = Kalhovd, Rø = Røldal, Ha = Haukeli, So = Songevatn, Bi = Bitdal, Fr = Frøystul. Dots and numbers indicate site and numbers of sediment samples.

Table 1. Sediment samples listed according to different map sheets covering the area from which they are sampled (cf. Fig. 17), and to type: A = ablation till, B = basal till, D = weathering material, or other undetermined diamictons, G = glaciofluvial/glaciolacustrine sediments, S = sorted subtill or intertill sediments

Map \ Type	A	B	D	G	S	Σ
Ul		1	3			4
Ei		1	1	12		14
Bj	12	44	2	36	11	105
He	9	39	1	12	2	53
Sk		6				6
Hå		12		6	2	20
No	9	21	13	20	7	70
Lå	6	33	1	5	1	46
Ka	7	32		7	22	68
So		7			1	8
Bi		26	1		26	53
Fr	5	15		2	5	27
Rj		3				3
Σ	48	240	22	100	77	487

sieving, was homogenised and c. 20 g taken for pipette analysis (Krumbein & Pettijohn 1938, Galehouse 1971a). Prior to analysis these samples were dispersed in Calgon for 7 minutes.

The weights of every fraction analysed were then run on a computer programme prepared by Myhre (1974).

There is a large selection of grain-size scales (cf. summaries by Truesdell & Varnes 1950, Pettijohn 1957) which is still being augmented (Shea 1973). The phi scale (Krumbein 1934) is employed in the present study. Page (1955) has published tables for converting phi values to metric equivalents.

The prime advantage with the phi scale is its extensive use internationally; furthermore, statistical parameters are increasingly based upon it (McBride 1971, p. 110). Other advantages have been discussed by Tanner (1969, p. 809). The chief weakness is that phi values cannot be directly compared with results obtained when using the Atterberg scale which is still very popular amongst Nordic Quaternary geologists. Doeglas' system (1968, Fig. 1) has been followed for the grain-size classes. The grain-size distribution parameters recommended by Folk & Ward (1957) have been employed. Folk (1966) himself gives the best defence for such a choice.

Pertographical investigations. Heavy mineral separation of the 3–2 phi fraction (0.125–0.250 mm) was a standard procedure together with microscope counts of phyllite/black shales and coloured phyllosilisates.

Prior to separation the grains were cleaned by Leith's (1950) method, to remove superficial iron precipitates. Contrary to Leith (1950), this treatment was found to affect the phyllosilicates. To start with, almost all the phyllosilicates are lost during the decanting, while there is probably a chemical reaction within the phyllosilicates as well. Brown mica was white after treatment – this was confirmed by a control test, and may be due to an exchange of Fe and Al ions.

The heavy mineral separation was a gravity separation in bromoform (s.g. = 2.88) (Carver 1971, p. 439). The weight per cent of heavy minerals used excludes phyllosilicates on account of the cleaning treatment, but since the original phyllosilicate content was generally very low (p. 59) the values quoted are virtually equivalent to the true weight per cent values of the total fraction.

The amounts of coloured phyllosilicates and phyllite/black shales (hereafter called 'phyllite') were determined optically. The grains were utreated except in a few cases where cleaning was imperative to enable the phyllite to be identified. The phyllosilicates were then not counted.

Samples were prepared by mounting about 500 to 1000 grains in Clearax (n = 1.666).

A Leitz Ortholux II Pol-BK-microscope was used, with weak transmitted illumination and powerful reflected illumination, and magnification X 160. Between 300 and 500 grains were counted from each sample, using the line method (Galehouse 1971b). Minerals were grouped as phyllite, coloured phyllosilicates and 'other minerals', the latter being mostly quartz and feldspars.

Phyllosilicate identification presented few problems. These minerals are nearly all lying on the [001] plane, giving total extinction in plane polarised light. The majority were brown biotites, many of which displayed a typical hexagonal pattern of inclusions. (White micas were only found in small quantities and sporadically, and were not counted systematically.)

The phyllite was fairly easy to identify too. It is opaque with rounded edges, and under reflecting illumiation reveals a finely granular surface, often with lineations. It can be confused both with ore minerals and with aggregates of clay-silt particles if present. However, ore minerals have a more shiny surface, more angular shape and often prossess conchoidal fractures which are not observed on the phyllite. Fine-sediment aggregates usually differ by having a greyer colour and lack of lineation. In doubtful cases, treated samples were examined, as the cleaning process eliminates such aggregates. One problem posed by the phyllite was that in this fraction (0.125–0.250 mm) it is in part reduced to individual mineral grains, mostly quartz (see below). Grains consisting chiefly of quartz but retaining traces of phyllite were classified as phyllite.

The phyllite content of the other sand fractions and the two finest gravel fractions was determined for 51 of the samples. The 4 to 3 phi and 2 to 1 phi fractions were analysed in the same way as the 3 to 2 phi fraction. The 1 to 0 phi (0.5 to 1.0 mm) fraction was analysed by mounting the grains on a slide without a cover glass, and then using a binocular. The coarsest material (1 to 2 mm, 2 to 4 mm and 4 to 8 mm) was identified under a binocular by Lundqvist's (1952) method, which was also used to identify blue quartz in a few samples.

In the following, corrected phyllite values will also be used. This is because phyllite material is crushed and in part occurs as free mineral grains – mainly quartz – in the finer fractions. The phyllite contents of the 4 to 3 phi, 3 to 2 phi, 2 to 1 phi and 1 to 0 phi fractions have been multiplied by 2.0, 1.5, 1.2 and 1.1, respectively.

The correction factors were calculated by analysing crushed phyllitic rocks in the respective fractions. The quantity of identifiable phyllite in the different samples showed wide variations, viz:

between	46%	and	93%	for	the	4 to 4 phi	frations
»	55%	»	99%	»	»	3 to 2 phi	»
»	82%	»	100%	»	»	2 to 1 phi	»
»	87%	»	100%	»	»	1 to 0 phi	»

The highest values of identifiable phyllite were obtained from alum shales, but it is the relatively quartz-rich phyllite which is the areally predominant type of schist (p. 5). The correction values were therefore weighted with respect to this.

In order to compare this data with the grain-size distribution by weight, it is interesting to recalculate the number per cent of phyllite as weight per cent. It is assumed that the number per cent in the respective fraction equals the volume per cent. Four determinations of the specific weight of phyllitic material gave values close to 2,7, which is fairly near to the specific weight of the granitic material which dominates the sediments on Hardangervidda. The relationship between the phyllite's specific weight and that of the remaining material is therefore set at unity. The weight per cent phyllite in the respective fraction of material finer than 16 mm is thus arrived at by multiplying the number per cent by the weight per cent of the respective fraction and dividing by 100.

TILL

Nomenclature and classification

The classification of till is often based on texture and/or genesis, though stratigraphy and petrographical parameters have been used too.

For a long time a genetic division of till into the two groups, basal till/lodgement till and ablation till, has been recognised (Flint 1971, p. 171).

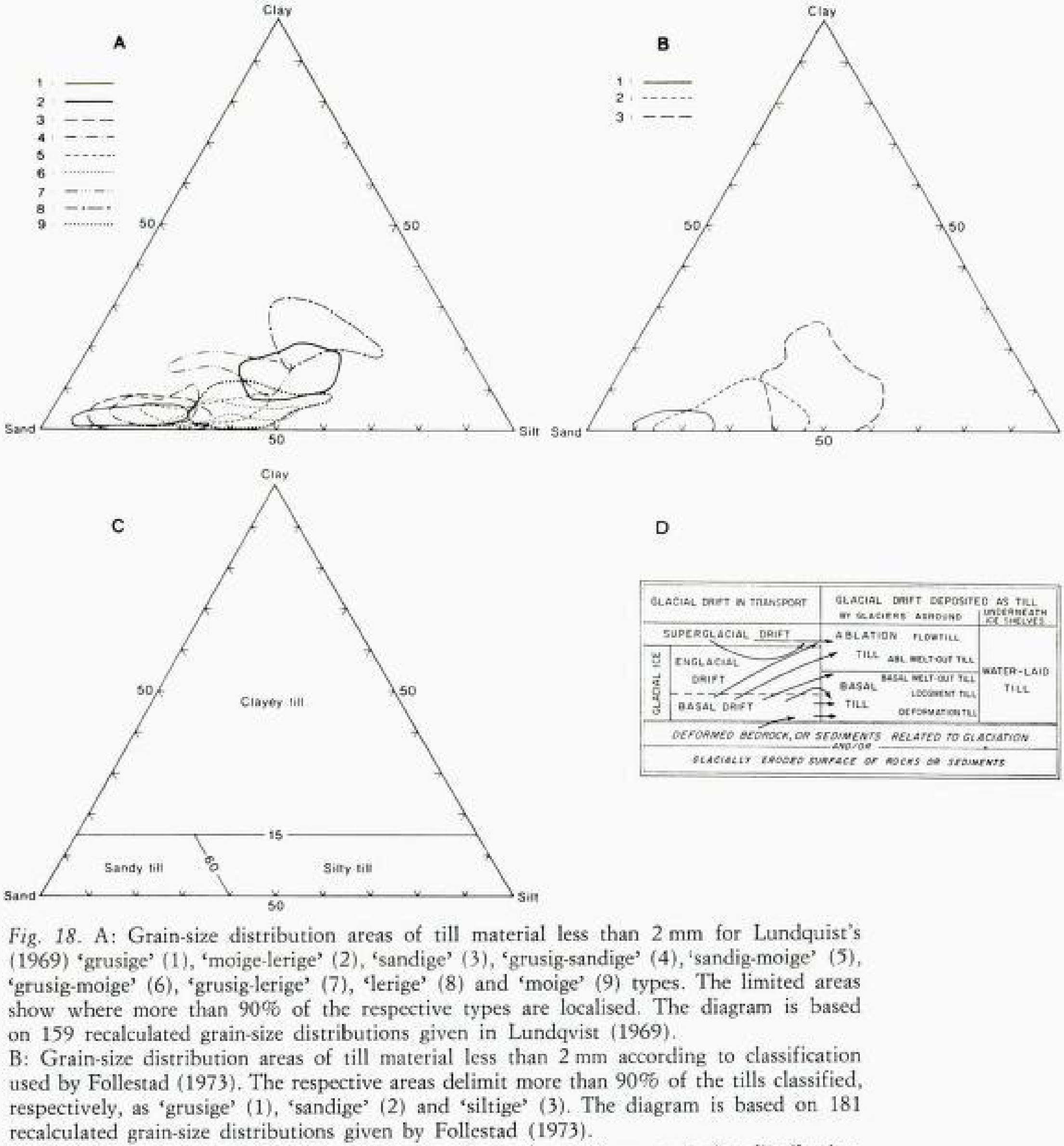


Fig. 18. A: Grain-size distribution areas of till material less than 2 mm for Lundquist's (1969) 'grusige' (1), 'moige-lerige' (2), 'sandige' (3), 'grusig-sandige' (4), 'sandig-moige' (5), 'grusig-moige' (6), 'grusig-lerige' (7), 'lerige' (8) and 'moige' (9) types. The limited areas show where more than 90% of the respective types are localised. The diagram is based on 159 recalculated grain-size distributions given in Lundqvist (1969).
B: Grain-size distribution areas of till material less than 2 mm according to classification used by Follestad (1973). The respective areas delimit more than 90% of the tills classified, respectively, as 'grusige' (1), 'sandige' (2) and 'siltige' (3). The diagram is based on 181 recalculated grain-size distributions given by Follestad (1973).
C: Proposed textural classification of tills based on the <2 mm grain-size distribution.
D: Genetical classification of tills and their relationship to glacial drift in transport (after Dreimanis 1974).

A more sophisticated subdivision was made by Dreimanis & Vagners (1971, p. 239) and later somewhat modified by Dreimanis (1974), Fig. 18D. Francis (1975, p. 50) has also proposed a modification of Dreimanis & Vagners' classification. Since this does not incorporate any improvements, in the writer's opinion, it will not be discussed further.

Dreimanis (1974) has divided ablation till and basal till into two and three subgroups, respectively, and proposed a separate main group for waterlaid till.

In order to distinguish between the various subgroups it is necessary to carry out in part detailed fabric analyses (Boulton 1970, 1971). In the present regional geological investigation the writer had to be content, therefore, with grouping the till into the two main groups, ablation and basal till. Even this can be difficult enough (cf. Drake 1971). The most important criteria used were the morphology and degree of consolidation. In the c. 10 cases where field identification was ambiguous, the grain-size distribution was the decisive criterion. Of the 307 diamicton samples, 240 were classified as basal till and 48 as ablation till.

Examples of textural classifications of till may be found in G. Lundqvist (1940), Elson (1961), Virkkala (1969a) and Follestad (1973). Ideally, all the grain-size fractions should be taken into account in a textural classification. Lundqvist's (1940) classification comes nearest to this, but is weakened, in the writer's view, by requiring so many words to describe certain till types. An extreme examples is: 'normalblockig grusig sandig moig lerig morän' (cf. Lundqvist 1969, p. 412) which in English becomes something like: 'gravelly sandy silty clayey till with normal boulder content' Another objection, which also applies to most of the other classifications, is that 20 mm is an unfortunate choice for a grain-size limit for the phi scale. Many till researchers, particularly Danish and North American, set the upper limit of their grain-size analyses at 2 mm.

One conceivable approach is a textural classification based on:

1) The ratio between coarse (>2 mm) and fine (<2 mm) material.
2) The ratios between gravel, cobbles and boulders within the coarse material.
3) The ratio between clay, silt and sand within the fine material.

(2) and (3) must be described by one word each but the ratio in (1) may be expressed with the noun of the dominant fraction and the adjective of the subordinate fraction. However, the paucity of available data renders such a classification impractical at the present time. One must meanwhile be content to classify according to the proportions of clay, silt and sand.

In Fig. 18 the defined limits for textural types of till are plotted, based on Lundqvist (1969) and Follestad (1973). It is clear that a relatively low clay content is necessary in order to call a till clayey. Furthermore, sand must dominate the material if it should be termed sandy. These features seem to agree with the informal terminology for till texture in Fennoscandia. A simple textural terminology for till will therefore be proposed here, based on the clay, silt and sand content as shown in Fig. 18C. According to this classification 0.4% of the basal tills sampled from Hardangervidda are clayey, 33.8% are silty and 65.8% are sandy. All the ablation tills are sandy except one that is silty.

Another, purely terminological problem which applies especially to till is the use of the terms matrix and clast. Based on his demonstration of the bi- or poly-modality of the individual petrographical components in till, Dreimanis (1969, p. 19) has suggested the boundary should lie between 0.1 and 2.0 mm, where there is often a minimum. Lindén (1975, p. 3) uses 2 mm as the limit.

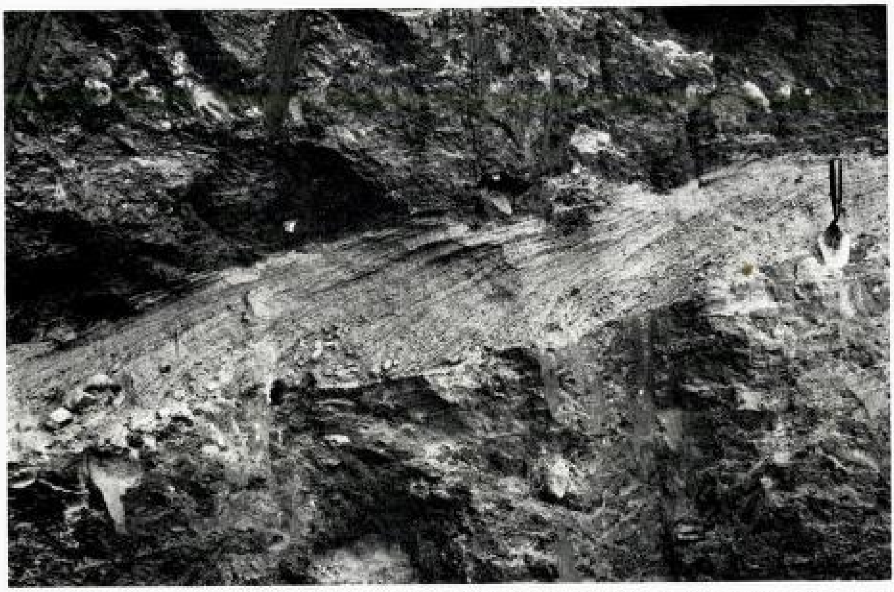


Fig. 19. Sand lens (sample 77b–72) in basal till (sample 77a–72) at Kalhovdfjorden, map sheet Kalhovd. Primary foreset laminae are clearly visible in the sand inclusion. Faint fissility can be seen in the basal till.

To fix a boundary between matrix and clast is contrary to the established meaning of these terms. They refer to the relative sizes of particles in a sediment where no special particle size boundary is defined (Gary et al. 1974, p. 434). Furthermore, bi- or poly-modality in till may have other origins than the comminutation characteristics of the individual petrographical components, e.g. the incorporation of previously sorted sediments. In the writer's opinion the boundary between matrix and clast should be defined specifically for the material being described.

Structures

The basal tills on Hardangervidda can be subdivided according to structure into massive tills, tills with inclusions, fissile tills and layered tills. The massive tills show no sign of structure and are apparently an homogeneous mixture of the different grain-sizes in the till.

Tills with inclusions are virtually as common as massive tills. The inclusions are of sorted material which may be dominated by clay, silt, sand or gravel. Sandy inclusions seem to be commonest. The inclusions are mostly of very local extent and irregular outline. The inclusions themselves may display internal primary structures such as cross-bedding and stratification (Figs. 19 and 20) which reveal original deposition in an aquatic environment. Tills with inclusions are often found near areas with subtill sorted sediments. These were probably laid down subaerially and later incorporated in the till. There are some localities, however, with subtill sediments which are presumed to be of entirely subglacial origin.

Fig. 20. Inclusion of bedded sand and gravel in basal till east of Mårvatn, map sheet Kalhovd. Originally the sediments were probably deposited subaerially during the Førnes thermomer.

Fig. 21. Fissile basal till (sample 17–71) at Mårvatn, map sheet Kalhovd. The fissility here may be due to desiccation.

The structures termed fissility (Figs. 19 and 21) probably correspond to Virkkala's (1952) 'laminated structure' and Boulton et al.'s (1974) 'platy structure'. This is a group of structures probably due in part to consolidation effects (Boulton et al. 1974), but also to ' . . . accretion of successive layers of drift

Fig. 22. Layered till exposed at Gøystdalsvassåi (cf. Fig. 48). The scale is 0.5 m long.

from the base of the glacier as the ice moved forward' (Flint 1971, Virkkala 1952) and to post-sedimentary soil structures (Flint 1971).

The expression layered till as used here excludes till beds of various ages derived from different areas, and tills of different genetic origin. Only distinct layering within one and the same till unit are considered. Such structures were observed only at one locality on Hardangervidda (Fig. 22). Here there are c. 5 cm-thick layers lying more or less horizontally within the till. These layers are better consolidated than the rest of the till and possess a distinctly higher clay content (Fig. 48). The till concerned here overlies sorted sediments, from which a large part is derived, judging by the grain-size distributions (Fig. 48). The relatively large clay content in the thin layers can perhaps be explained in terms of the basal ice having shear zones where debris suffered more crushing than in the adjacent zones. Later the basal ice with these zones has become separated. Virkkala (1952) has described similar structures and his interpretation was summarised by Flint (1971, p. 159) as '. . . these layers were made by successive stagnation of thin overloaded basal layers of ice in the terminal zone of the glacier. The base of the moving glacier shifted upwards in jumps, leaving beneath drifts with interstitial ice.'

Phyllite content

Phyllite content in different grain-size fractions. In Fig. 23 39 till samples have been grouped according to their location on, or transport distance from, phyllitic bedrock. With several different ice movements the transport distance cannot always be estimated accurately. The samples selected are located where the transport distance will be within the given interval for all the ice movement phases. For example, samples 30–, 31– and 32–70 from Halnefjorden would have been transported 1 to 3 km whether the ice flowed eastwards (Fig. 12), westwards (Fig. 13) or southwards (Fig. 14), but in each case a different source area of phyllite would have been involved. One ambiguous factor which cannot be avoided is that the till material may have been transported by several different ice movements prior to deposition. The cited transport distances are therefore really minimum values. The following can be seen from the corrected phyllite value curves in Fig. 23:

— There is a tendency towards bimodality in samples with a high phyllite content. One mode, the matrix mode, lies close to 3 to 2 phi, while the other, the clast-size mode, lies over –1 phi (except for sample 144–73 which has a slight maximum between 0 and 1 phi).
— The pronounced clast-size mode is lacking in samples with lower phyllite contents (sample transport over 1 km from phyllite bedrock).

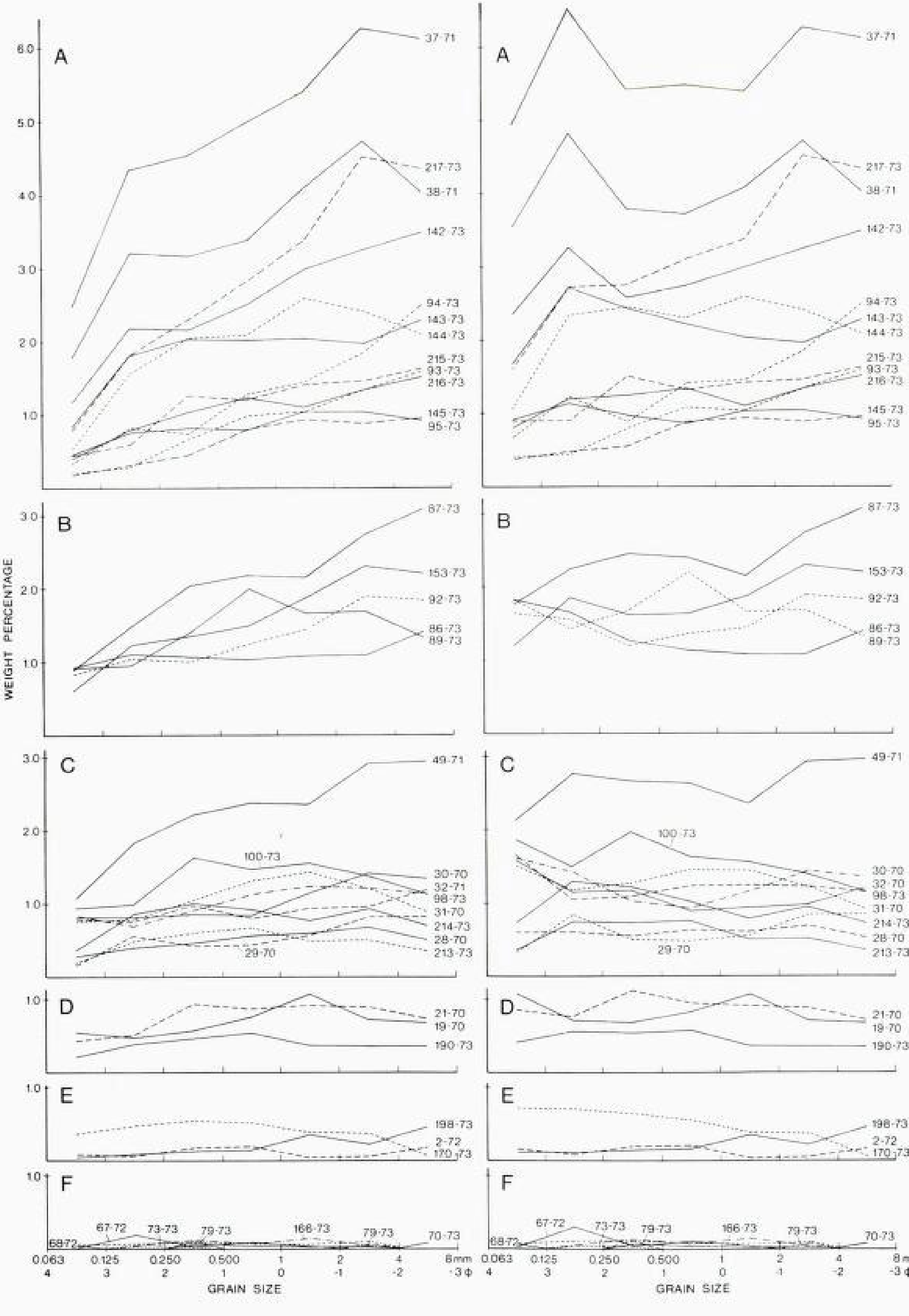
Furthermore one can see that:

— The variation in phyllite content between different samples is greatest in samples collected above or near to phyllitic bedrock.
— There is a clear tendency towards decreasing phyllite content with increasing distance from the phyllitic outcrop. With more than 15 km transport the phyllite content in till is sporadic, normally less than 0.1 weight per cent which corresponds to c. 1 number per cent.

The recorded phyllite bimodality in the phyllite-rich till is interesting. Where the clast-size mode is concerned, it lies in the –1 to –2 phi fraction in samples 37– and 38–71 and 92–, 158– and 217–73. Four other samples which indicate a different position for this mode were analysed with respect to material up to –6 phi (64 mm) (Fig. 24). Sample size was about 100 kg. In addition to a mode in the –2 and –3 phi fractions, sample 142–73 has a secondary mode at –5 to –6 phi. Sample 143–73 has a weak mode at –2 to –3 phi, whereas sample 144–73 has a mode at 0 to –1 phi. Sample 143–73 with a generally low phyllite content has a single diffuse mode centred around –1 phi. It seems therefore that phyllite-rich tills have a phyllite mode in the clast fraction between 0 and –3 phi with the bulk between –1 and –3 phi (2 to 8 mm).

With respect to the matrix mode, this lies between 3 and 2 phi (0.125 and 0.250 mm) for phyllite-rich tills, as already mentioned. Several samples with lower phyllite content show a tendency to a phyllite mode below 3 phi, but this is not always the case.

Dreimanis (1969) and Dreimanis & Vagners (1969, 1971 and 1972) have derived a rule for bimodal and polymodal distributions of petrographical components of till, with one mode in the coarse fractions and one or more modes in


WEIGHT PERCENTAGE
A
B
C
D
E
F
37-71
217-73
38-71
142-73
94-73
143-73
144-73
215-73
93-73
216-73
145-73
95-73
87-73
153-73
92-73
86-73
89-73
49-71
100-73
30-70
32-71
98-73
31-70
214-73
28-70
29-70
213-73
21-70
19-70
190-73
198-73
2-72
170-73
68-72
67-72
73-73
79-73
166-73
70-73
0.063 0.125 0.250 0.500 1 2 4 8mm
4 3 2 1 0 -1 -2 -3 ϕ
GRAIN SIZE

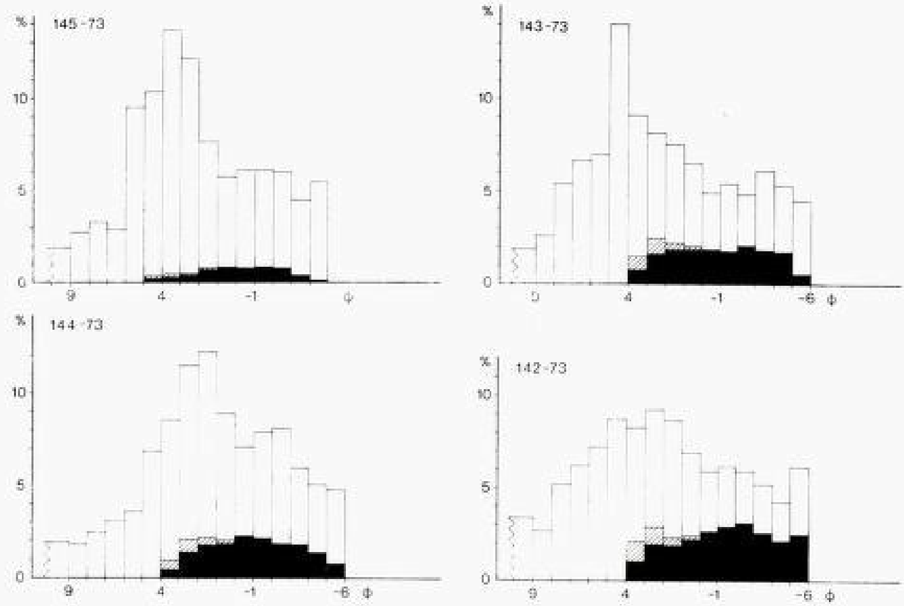


Fig. 24. Histogram showing grain-size distribution by weight of grain less than 64 mm (–6 phi) for four basal tills. The sample sites are on phyllite bedrock, map sheet Bjoreio. Black areas indicate weight percentages of phyllite in the fractions 4 to –6 phi. Shaded areas together with black areas indicate corrected phyllite values for the 4 to 0 phi fractions.

the fine fractions, according to the bedrock's mineralogical composition. The mode or modes in the fine fractions occur in certain grain-sizes typical for the respective minerals. These grain-sizes have been named terminal grades, because according to Dreimanis & Vagners these represent the final product of glacial comminution.

The clast-size mode for phyllite is clearly very short lived. It is only observed in the extremely phyllite-rich till on or close to the phyllitic bedrock and evidently disappears during the first kilometre of transport. By comparison, Dreimanis & Vagners (1972, p. 73) found that with dolomite the matrix concentrations appear as soon as the ice has incorporated the rock, and the clast-size mode is rapidly reduced. With metamorphic and igneous rocks the clast-size mode is still identifiable after several hundred kilometres transport (Dreimanis & Vagners 1971, Fig. 3). The reason for the relatively rapid disappearance of the phyllite clast-size mode registered on Hardangervidda must be that this rock can be easily comminuted because of its schistosity and softness.

With reference to Dreimanis and Vagners' investigations it could be tempting to interpret the mode in the fine fraction between 3 and 2 phi as a terminal grade. However, this mode also disappears after a transport of less than 1 km. Artificially crushed phyllites show a marked increase in single mineral grains from 2 phi downwards (p. 39).

Fig. 23. Uncorrected (left) and corrected (right) weight percentages of phyllite/black shale in different phi-fractions given on the abscissa. The samples in diagram A represent basal tills on phyllitic bedrock. The samples in diagrams B to F represent basal tills transported away from phyllitic bedrock, minimum 0–1, 1–3, 3–9, 9–15 and 15–30 km, respectively.

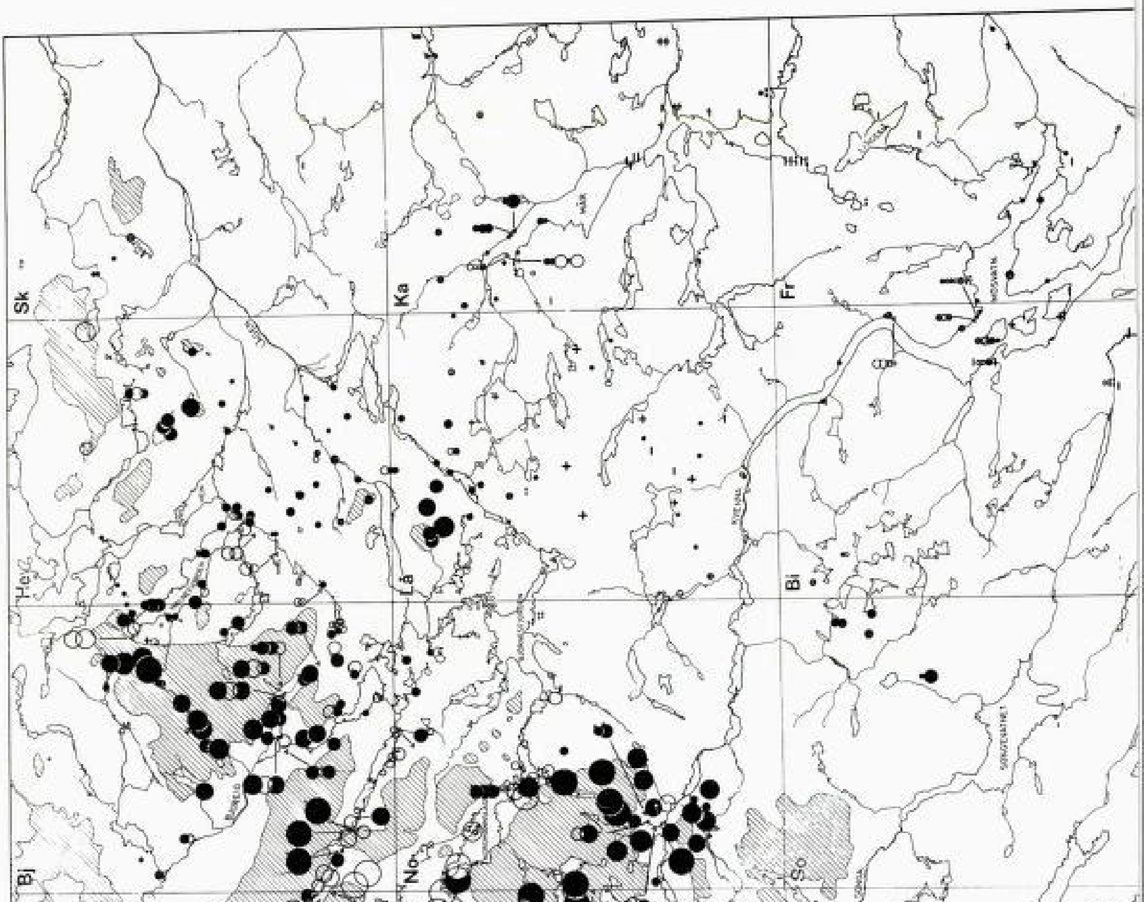


Fig. 25. Number percentages of phyllite/black shale in the 3–2 phi fraction in the different sediment samples. Values placed vertically above each other represent samples from the same sections; uppermost value shows uppermost sample, etc. Some values of the sorted

Table 2. Maximum, minimum, mean (X) and standard deviation (S) of number percent phyllite in 3–2 phi fraction in basal tills from, a: isolated Cambro–Silurian bedrock localities on the northeastern part of Hardangervidda, b; Cambro–Silurian bedrock area between Halnefjorden and Bjoreio, c; Cambro–Silurian bedrock area between Nordmannslågen and Kvenna. N is number of samples

	max.	min.	X	S	N
a	23.5	3.3	9.8	5.9	9
b	46.9	1.6	15.7	12.5	14
c	62.7	22.3	43.3	11.9	13

The observed mode is probably due to the initial increase in resistance to communition: but most of the single mineral grains from phyllite are smaller so the terminal grades most likely lie in the coarse silt/very fine sand fraction (for the quartz component) and in the fine silt/clay fraction (for the mica component). Dreimanis & Vagners (1972, p. 75) found that quartz derived from fine-grained clastic rocks has terminal grades in the 4 to 5 phi and 7 to 8 phi fractions. The increase in phyllite content below 3 phi which was noted in half of the samples from beyond phyllite outcrops, may represent a rise towards a terminal grade around 4 phi.

Summarising; the comminution of phyllite seems to occur in two steps. In the first step there is comminution and enrichment, i.e. mode, in the 3 to 2 phi and (0) –1 to –3 phi fractions. Second stage comminution then leads to the disappearance of these modes, the 3 to 2 phi mode being displaced into even finer fractions. The transition from the first to the second step is rapid and usually requires less than 1 km transport. With further comminution there is a fairly even reduction in all the fractions examined. Nevertheless the data suggest that the coarser fractions disappear first. Of the six analysed samples of material transported 15 to 30 km there was only one which retained any phyllite in the coarsest fraction (–2 to –3 phi).

Regional variations in the phyllite content. No systematic variation in phyllite content between ablation and basal till within the respective areas is indicated (Fig. 25). This is in agreement with Drake's (1971) results from New Hampshire, where no petrographical differences were found between ablation and basal till. To ensure a valid comparison, however, only basal tills will be considered here. First, those overlying phyllitic bedrock will be dealt with, then the ones above Precambrian bedrock.

If the basal till samples overlying phyllitic bedrock are divided into one group for the isolated eastern phyllitic areas, another for the phyllite tract between Bjoreio and Halnefjorden and a third for that between Nordmannslågen and Kvenna (Fig. 25), one discovers they have quite different phyllite content characteristics (Table 2). One of the reasons is probably the erosion mechanism. If one compares the area between Bjoreio and Halnefjorden with

sediments are average values of two or more samples. Ablation till is indicated with a line in the upper right quadrant, while diamictons, excluding basal and ablation till, are indicated with two lines in the same quadrant. Basal till has no such special markings.

the similarly sized Nordmannslågen–Kvenna area, the latter possesses far greater relief differences. Erosion by plucking may be assumed to have been much more active here. Since this process provides more material than abrasion would (Flint 1971, p. 112), it may partly account for the higher phyllite content in the till in the Nordmannslågen–Kvenna area. This is hardly the whole explanation though; the supply of foreign material must be taken into account. The ice has flowed from the west or south across this area. Much of these neighbouring areas also have phyllitic bedrock so that a good deal of the foreign material was phyllitic too.

The dominant ice movements, insofar as they can be deduced from erosion landforms (Fig. 11), have crossed the area between Bjoreio and Halnefjorden from chiefly Precambrian zones to the east and north. Material derived from these rocks has thus mostly been transported into the phyllite area and has suppressed the phyllitic component in the till. Some of the samples from here are completely dominated by Precambrian material. The samples 76–71 and 145–73 (Fig. 17) were taken from a deformed drumlinoid feature deposited by ice flowing towards the WSW. Precambrian rocks lie 3 km away. The phyllite content is scarcely more than the background level, 1.6% and 2.1% respectively, in the 3 to 2 phi fraction. There has therefore been almost no incorporation of material for a distance of 3 km.

The till sample 93–73 was found overlying phyllitic bedrock which is scoured towards the west — Phase III. It must therefore have been deposited by Phase III or Phase IV ice movements, which implies transport across 4 or 14 km of phyllitic bedrock, respectively, The phyllite values fail to exceed the background values in this case as well.

The significance of the volume of foreign material is shown most markedly by the phyllite content of tills on the isolated phyllite outcrops in the east (Table 2).

One may therefore conclude that the phyllite content in till overlying phyllitic rock is partly dependent upon the relative relief which determines the erosion mechanism and thus the amount of local material supplied, and partly on the amount of foreign material already being carried by the ice and related to the extent of the source rock outcrops.

For till samples from Precambrian bedrock areas the question of the significance of minor relief variations for the phyllite content will be discussed first. Afterwards the influence of transport distance and other factors on the dispersal of phyllite will be discussed.

Near Svinto there are two localities (loc. a and b, Fig. 53) which both lie about 1 km from phyllitic bedrock. At both localities there is a lower till deposited during Phase III and an upper till deposited during Phase IV. The prime difference between these two tills (Fig. 27) is the higher content of phyllite and other material in the coarser fractions in the upper till. The writer suspects that this is due to the minor landforms, i.e. roches moutonnées, offering little resistance to the similarly aligned ice flow of Phase III, so that abrasion, rather than plucking, was dominant. The youngest, Phase IV, movement

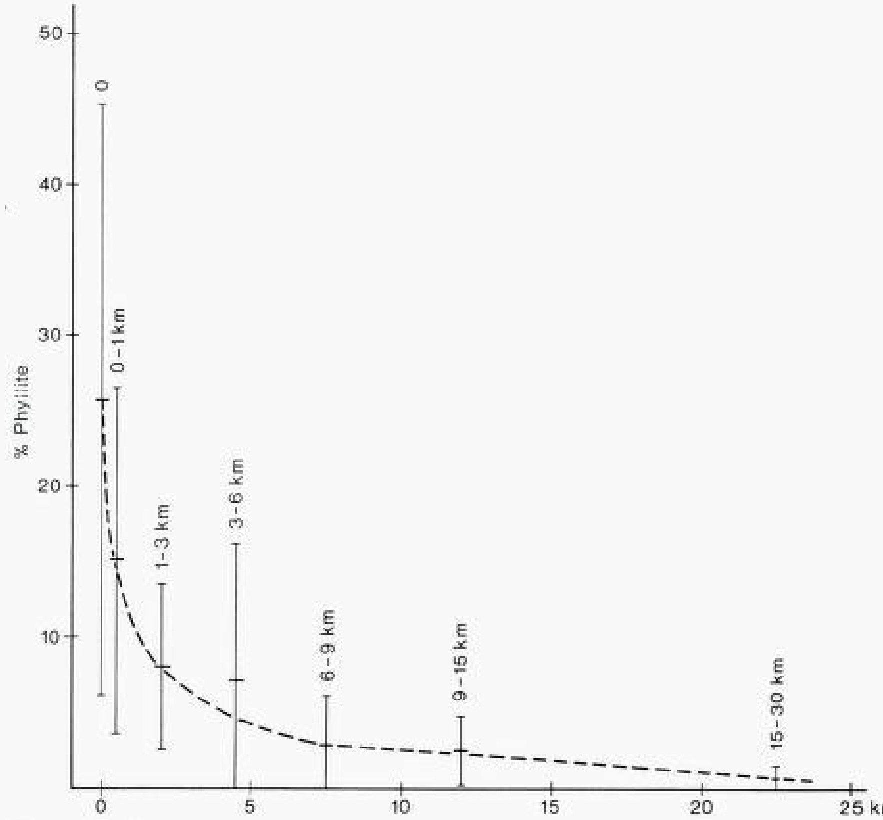


Fig. 26. Phyllite content in the 3 to 2 phi fraction (uncorrected) versus transport distance for basal tills. Each column shows mean and standard deviation of samples according to different transport distances away from phyllitic bedrock (indicated by figures above columns, 0 = samples on phyllite).

was relatively short-lived and in this area was roughly at right-angles across the existing landforms. It is reasonable to expect that the resultant lack of adjustment between glacier sole and ground surface would lead to more active plucking processes.

Dilution and comminution during transport is considered one of the most important factors for the petrographical composition. On Fig. 26, 223 of the basal till samples are grouped according to their assumed transport distance, taking into account all the known factors such as stratigraphy and ice movement direction. Even with respect to these there are many uncertain classifications and for 17 of the samples it was impossible to define the transport distance.

Despite these uncertainties one may state categorically that there is a reduction in the amount of phyllite in the till with increasing transport distance. Further, the phyllite appears to be diluted/comminuted relatively quickly. Of the 14 samples in the 6–9 km group, 13 contain less than 5% phyllite and 2 of those have no phyllite. After more than 15 km's transport

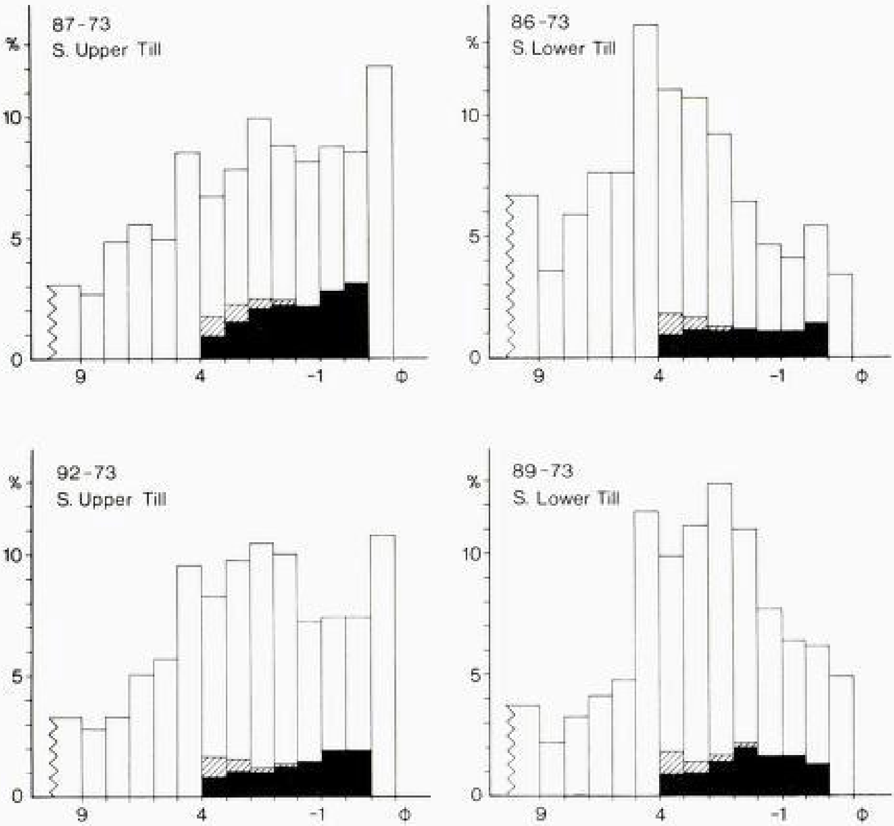


Fig. 27. Histogram showing grain-size distribution by weight of grains less than 16 mm (–4 phi) for four basal till samples from the Svinto area, see Fig. 53. Black areas indicate weight percentages of phyllite in the 4 to –3 phi fractions. Shaded areas together with black areas show corrected phyllite values for the 4 to 0 phi fractions.

there are only traces or no phyllite left in 42.7% of the till samples. 74.8% contain less than 1% phyllite and 85.4% have less than 1.5%.

When making comparisons with other investigations one must take into account that the rock's quality influences its capacity for glacial dispersion, see for example G. Lundqvist (1952) and Gilleberg (1967). Generally shales and limestones disintegrate fastest.

It is well known that indicator boulders have been transported several hundred kilometres (e.g. Flint 1971, p. 111). These often make only a very small contribution to the total till material, as shown by Marcussen (1973), and their mode of transport has not been fully explained. For the Danish indicator boulders Marcussen (1973, p. 181) intimates that they were transported during several glaciations and perhaps not only by glacier ice. Gry (1974) maintains that the boulders must have been carried englacially.

Harrison (1960) who has analysed 16 fractions in till from central Indiana, found that over 90% of the material has been transported more than 160 km. Goldthwait (1971, p. 15) has expressed doubts about the petrographical determinations in this investigation.

Gravenor (1951, p. 69) found that after 47 km's transport the crystalline rock content in Ontario was 18–20%, falling gradually to 4% after 280 km.

J. Lundqvist (1952, Fig. 22) found that the combined Caledonian rock content in the gravel fraction in Kopparbergs Län was 30–50% after less than 15 km's transport and below 5% after 50–150 km, disappearing completely after 90–160 km.

Holmes (1952) found that over 90% of the till material in the New York area had been transported less than 80 km.

Virkkala (1969b) reported that the majority of gabbro fragments larger than 20 cm at a Finnish locality had been transported less than 5 km. In Norway, Låg (1948) found till which had virtually formed in situ.

Lindén (1975, p. 42) found that more than 85% of the material coarser than 20 mm in a Swedish granite area had been carried less than 3.5 km.

Gilleberg (1965, Fig. 22) found that the Cambro-Silurian shales in the 2–20 mm fraction were diluted/comminuted to less than 1% after 70–80 km in Västergötland and Östergötland. Comparable necessary transport distances for the same type of material were found to be 25–30 km in the Närke area (Gilleberg 1967, Fig. 7). His results are interesting seen in relation to the writer's, particularly since he also analysed shales (although in coarser grade). The greater transport distances which he reported may be due to more constant ice movement directions across his study areas.

Relatively few investigations have been made of transport distances for material finer than gravel. Gross & Moran (1971) analysed the 0.125 to 0.172 mm fraction and found that over 50% of the till material on the Allegheny Plateau was derived from within 32 km. Gravenor (1951), Gilleberg (1967), Virkkala (1969), Dreimanis & Vagners (1971) and Lindén (1975) reported that the fine fractions are normally transported farther than the coarse ones.

Summarising, the studies referred to give a varied picture of till material transport distances. There is a tendency, though, to indicate relatively short distances with the bulk of the material being deposited/comminuted within 10 to 50 km and only sporadic occurrences after 100 km or more. The phyllite transport on Hardangervidda seems to be amongst the shortest by comparison with this, especially when one remembers that here a fine fraction (3 to 2 phi) was examined, which normally would be transported farther than the coarser fraction.

It is evident from Fig. 25 that certain till samples near Møsvatn, the northern part of Mårvatn and one locality north of Songevatnet, Sterra, show abnormal values for phyllite content compared with the general pattern.

The Møsvatn localities lie more than 30 km from the nearest phyllitic bedrock and the most likely minimum glacial transport distance was 40–55 km. Some of the samples near Møsvatn with anomalously high phyllite content (5–, 25–, 29– and 54–73) are shown in Fig. 28 along with the envelope curves for the 'normal' phyllite content for till transported 9 to 15 and 15 to 30 km (cf. Fig. 23). In one case the Møsvatn till's phyllite content exceeds that met with in till transported only 9 to 15 km and several other examples show values

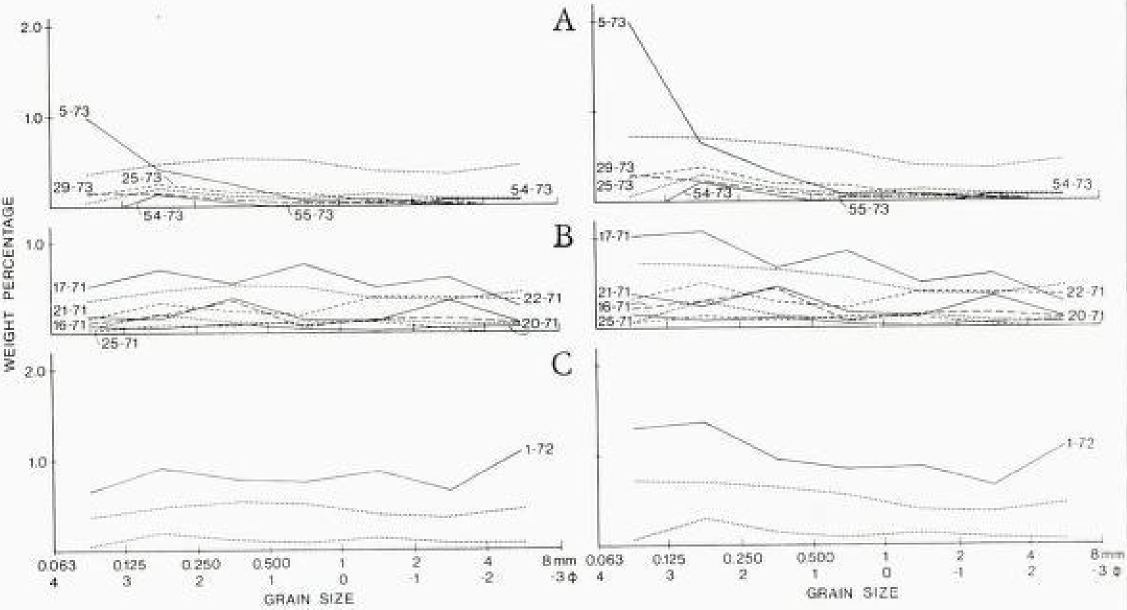
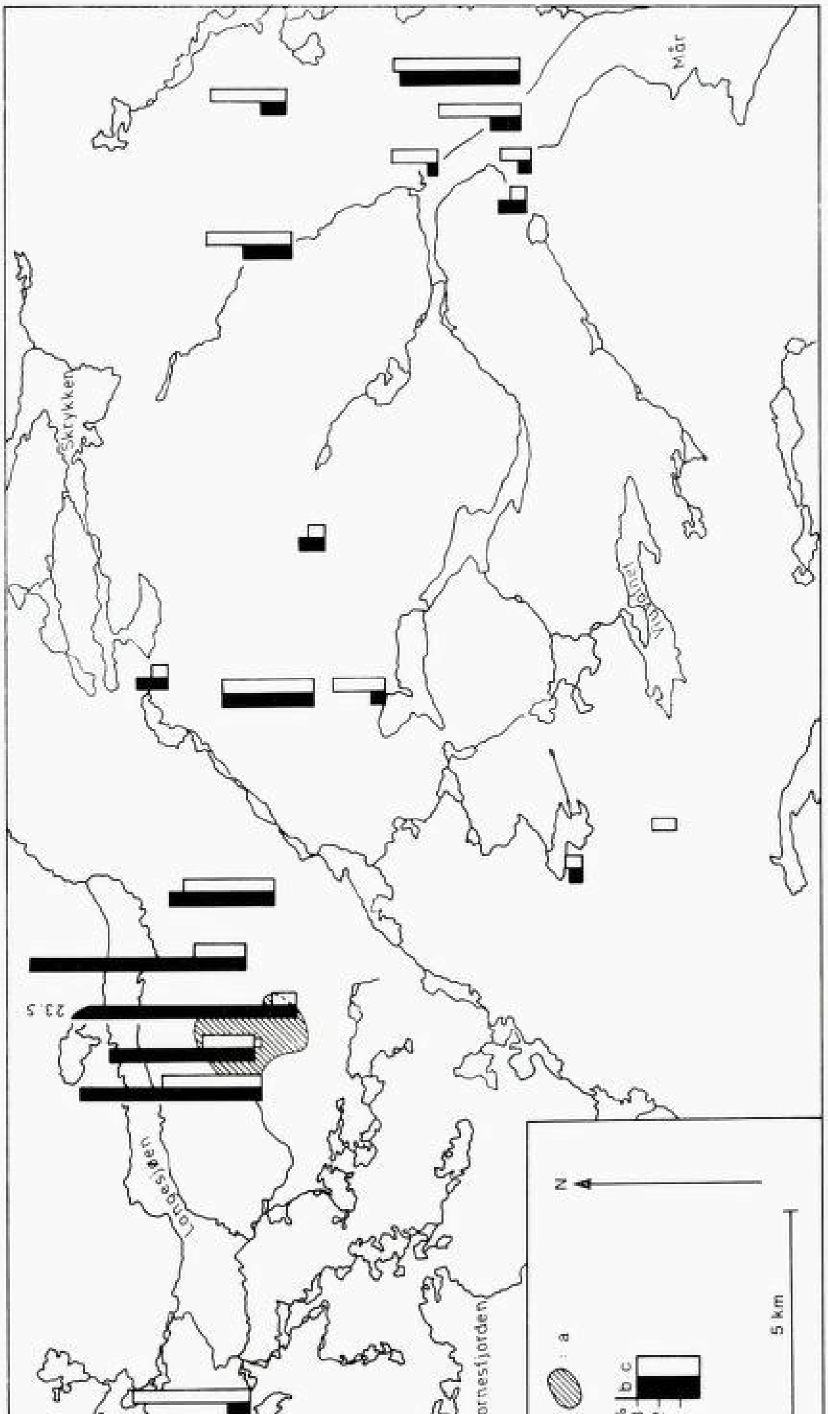


Fig. 28. Uncorrected (left) and corrected (right) weight percentages of phyllite/black shale in different phi-fractions given on the abscissa. The samples represent basal tills from Møsvatn-area (A), Mårvatn area (B) and Sterra (C). Dotted lines give maximum values for basal tills transported 9–15 km (upper line) and 15–30 km (cf. Fig. 23).

higher than those for 15 to 30 km. Most conspicuous though is the phyllite content increase in the finer fractions, particularly in sample 5–73. (Sample 54–73 deviates from this pattern by its lack of observable quantities of phyllite in the 4 to 3 phi fraction.) By comparison with the 'normal' samples there thus seems to have been a relative enrichment in the finer fractions, and it is tempting to attribute this to fluvial transport. The Kvenna river which runs into Møsvatn has its sources on phyllite (see Fig. 25). Presumedly much of the phyllite now found in till near Møsvatn was originally carried into the area fluvially during ice-free periods. The fluvial sediments have subsequently been incorporated in the till. Such an interpretation is supported by grain counts in the subtill sediments, sample 31–73, where the corrected number per cent values in the 4 to −3 phi fractions were 7, 6.6, 4.4, 2.4, 3.1, 1.4 and 1.5, respectively.

Possible sources of the phyllite in the till with abnormally high values near Mår are the phyllite outcrop south of Langesjøen, Fig. 29 (18–22 km transport), or the area of phyllite between Nordmannslågen and Kvenna (30–40 km). The ice movement analyses favour the latter area. To decide the the question, the blue quartz content of some of the till was examined (Fig. 29). Blue quartz is absent from the Cambro–Ordovician rock south of

Fig. 29. Content of phyllite (b) and bluish quartzite (c) in basal tills. The content of phyllite is represented as number percent (uncorrected) of the 3–2 phi fraction while the content of bluish quartzite is given as average number percent of the three phi-fractions 0– −3 phi. «a» denotes area with phyllitic bedrock.



Mår
Skrykken
23.5
Bjornesfjorden
a
b c
% 3 2 1
N
5 km

Langesjøen, but does occur in the Nordmannslågen–Kvenna area. One can see from Fig. 29 that the blue quartz content in the till overlying the phyllite south of Langesjøen never exceeds the 'background value' given by the sample at the western end of Langesjøen. Blue quartz is relatively abundant in the phyllite-rich till around the upper end of Mårvatn. One of the samples even contained more than those from the phyllite area south of Langesjøen. Since the blue quartz must have been derived from the Nordmannslågen–Kvenna area, so must the phyllite too, indicating 30–40 km transport for the Mår samples

Several explanations may be proposed for the anomalously high phyllite of these samples, viz:

1) Interglacial/interstadial fluvial transport.
2) Unmapped phyllite outcrops close to the area.
3) Modes of transport characterised by slight, or absence of, dilution/comminution of material.
4) 'Abnormally' large supply of phyllite.

Re–(1), it is difficult to envisage fluvial transport comparable to the Møsvatn example, since the phyllite has a 'normal' distribution throughout the individual fractions (see Fig. 28). The present drainage system does not support such an explanation either. Re–(2), the extent of phyllite outcrops has been given considerable attention and the writer rules out the possibility of unknown outcrops anywhere near Mår. A now-eroded phyllite source seems highly unlikely too, remembering the blue quartz content. This latter rock is, as mentioned (p. 5), restricted to the western Cambro–Ordovician areas. Re–(3), of much of the material has been transported as larger or smaller thrust slices, this would account for the anomalously high phyllite content. G. Lundqvist (1947, p. 89) and J. Lundqvist (1952 p. 24) propose the same process for certain boulder frequency anomalies in Sweden. Re–(4), an explanation implying an abnormally large supply of phyllite is to some extent supported by the fact that the most phyllite-rich till near Mår presumedly dates from the first Weichselian glaciation (p. 83). One can thus envisage that after a prolonged period of subaerial weathering there would be an abundance of weathered phyllite material. If this its correct one should, perhaps, expect a wider distribution of phyllite-rich till. Explanation (3) is therefore considered more probable, involving the en bloc transport of phyllite-rich till. The mechanism for incorporating and transporting glacial thrust slices has been discussed in the literature, e.g. Moran (1971), Boulton (1972) and Banham (1975).

Sample 1–72 north of Songevatn, near Sterra, also contains very much phyllite. The minimum transport distance for the phyllitic material is 11 km, and similar explanations are valid here as at Mår. Again, fluvial transport is excluded (Fig. 28) along with unmapped outcrops. Like the Mår deposits this till was also deposited during early Weichselian (p. 89). The explanation could be transport as thrust slices or an unusually large supply of phyllite at the end of the Eemian. It may be mentioned that pieces of phyllite-rich till up to 1 m^3 occur in granitic till 10 km south of Songevatn. These observations endorse a thrust slice explanation.

Collectively the analyses of phyllite content of till overlying Precambrian bedrock thus show that:

- minor landforms are significant for the phyllite content since they influence the type and degree of erosion,
- after 15 km transport the majority (74.8%) of the till samples contain <1% phyllite in the 3 to 2 phi fraction,
- positive anomalies of the latter occur and are due to, (a) interglacial/ interstadial fluvial transport with subsequent incorporation in the till (Møsvatn), (b) transport as thrust slices (Mår, Sterra) or (c) less likely, a large supply of weathered phyllite after a long ice-free period.

Heavy minerals

All the till samples were analysed to determine the weight per cent of heavy minerals (s.g. ≥2.88) in the 3 to 2 phi fraction (Fig. 30). In addition the heavy mineral content in the 4 to 3 phi fraction was determined in 23 of the samples. The average content in the 4 to 3 phi and 3 to 2 phi fractions was 5.7% and 4.2%, respectively. Only one of these samples revealed a lower content in the 4 to 3 phi fraction than in the 3 to 2 phi fraction. There is thus a clear tendency for higher concentrations of heavy minerals in the finer fractions (cf. Virkkala 1969b, Rosenqvist 1975, Table 4).

The heavy minerals have not been systematically identified but while counting the phyllite and phyllosilicates it was noticed that amphiboles and magnetite predominated. In general, one can say that there is a paucity of heavy minerals in the Hardangervidda tills. Basal till on average contains 3.84% and ablation till 3.67% (Table 3); in other words they are roughly the same. However, 'phyllitic tills', that is tills with >20% phyllite in the 3 to 2 phi fraction (Vorren 1977a), on average contain less, 2.28%.

Table 3. Maximum, minimum, mean (x), standard deviation (s) of weight percent heavy minerals in the 3–2 phi-fraction for; phyllitic basal tills (F), ablation tills (A), basal tills (B) and basal tills sampled from areas covered by the map sheets Bj . . . Rj (cf. Fig. 17) respectively. N is number of samples

	max.	min.	X	S	N
F	4.1	1.1	2.28	0.68	29
A	8.6	1.1	3.67	1.72	48
B	58.1	1.1	3.84	4.08	240
Bj	5.0	1.5	2.90	0.91	44
He	7.1	2.2	3.67	0.9	39
Sk	5.5	3.9	4.6	0.63	6
Hå	5.0	1.3	2.46	1.06	12
No	11.5	1.1	3.51	2.34	21
Lå	4.9	1.4	2.62	0.80	33
Ka	7.1	1.2	3.53	1.32	32
So	4.8	2.1	3.37	0.76	7
Bi	8.3	2.4	4.55	1.34	26
Fr	7.9	3.0	5.00	1.57	15
Rj	7.9	4.1	6.23	1.59	3

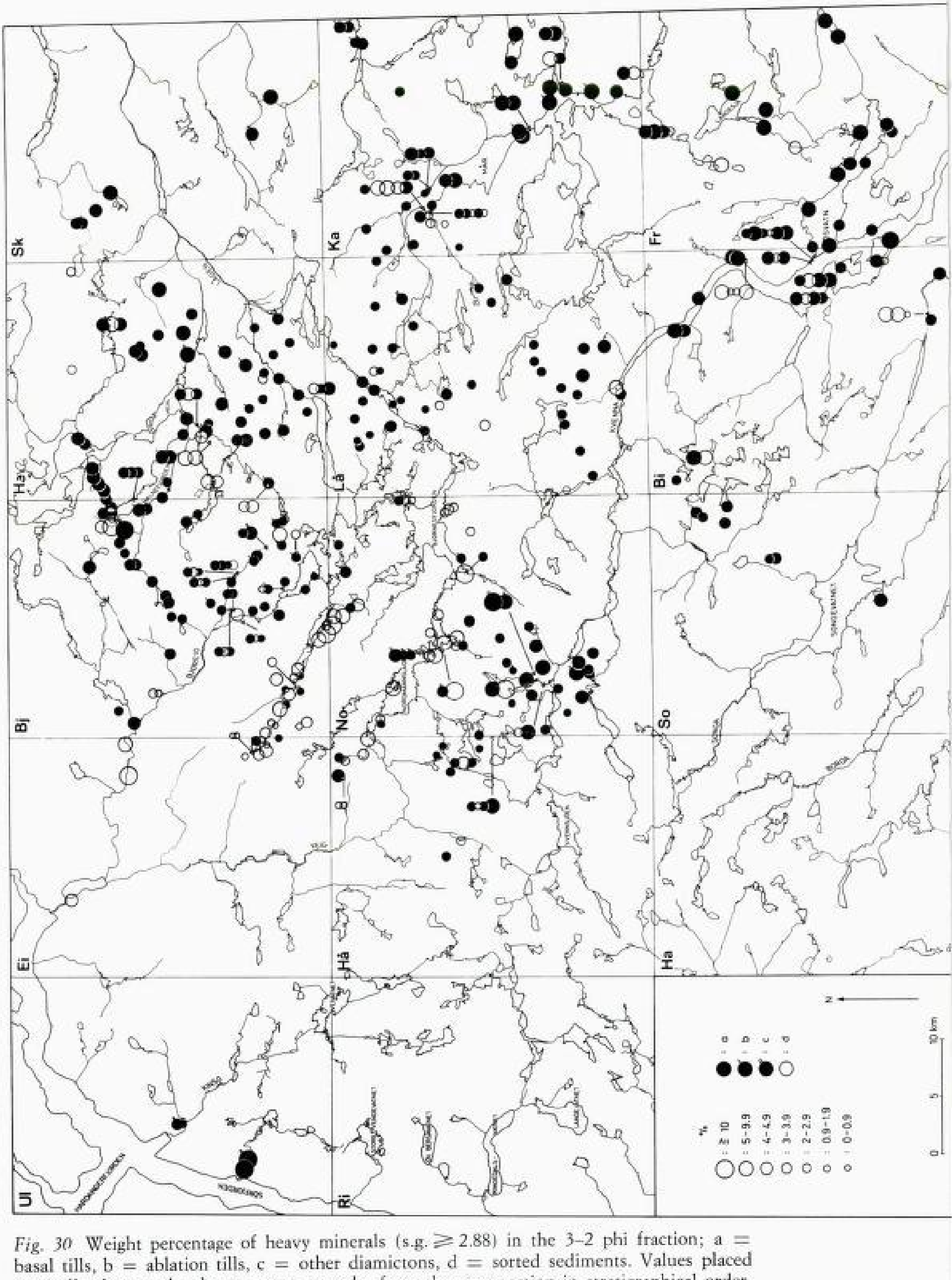


Fig. 30 Weight percentage of heavy minerals (s.g. ≥ 2.88) in the 3–2 phi fraction; a = basal tills, b = ablation tills, c = other diamictons, d = sorted sediments. Values placed vertically above each other represent samples from the same section in stratigraphical order. Some values of the sorted sediments are average values of two or more samples.

Regional variations. Fig. 30 and Table 3 reveal that the tills within the Bjoreio and Hårteigen map sheet areas have relatively low heavy mineral contents. This is probably because much of the material here was derived from phyllite which contains very little of heavy minerals (cf. heavy mineral content of the phyllitic tills.) For the same reason a low value was anticipated in the Nordmannslågen map sheet area, but here the mean value was raised by a few samples with relatively high contents, up to 11.5%. These probably come from an area with unmapped basic rock outcrops, perhaps biotite-hornblende schits (see below). The relatively low values in the Lågaros map sheet area are probably attributable to some of the material being derived from the local quartzite (Fig. 2).

In the Halnefjorden area, Hein map sheet, the heavy mineral content is above average. The heavy minerals probably come from the granite in this area (Fig. 2) which is relatively hornblende-rich; a couple of crushed rock samples gave an average of 4.8% heavy minerals in the 3 to 2 phi fraction.

The highest observed value, 58.1%, was provided by a till at Sørfjorden, sample 134–73, which is completely dominated by local hornblende schists. With this one exception the highest values are restricted to the eastern areas, cf. the average values for till within the Skurdalen, Bitdal and Frøystul map sheet areas. This was expected since here one is in contact with the basic rocks of the Telemark suite, of which the mapped outcrops are shown in Fig. 2. The surprising aspect though is that the heavy mineral content for certain of the samples is not even higher. If local material had been dominant one would have expected comparable values to that of sample 134–73. If the till samples in the area of basic rocks around Lågen, southeast of Halnefjorden (Fig. 2), are considered together as one group, they have a heavy mineral content approximately the same as the surrounding samples. Clearly the till here is virtually totally dominated by foreign material. The reason may be that the local rocks are relatively resistant and the landforms subdued. Together this leaves few vulnerable points for erosion.

Coloured phyllosilicates

On average the basal till contained 1.25% and the ablation till 1.60% coloured mica minerals (Table 4). The phyllitic till has lower contents – 0.71%. This may seem remarkable at first sight since the phyllite is chiefly composed of mica minerals. These are present, however, as minute sericite flakes and therefore do not contribute free minerals in the 3 to 2 phi fraction, cf. the existence of phyllite rock fragments in this fraction.

The highest phyllosilicate content, 27.1%, was recorded in sample 134–73 in the Ullensvang map sheet area near Sørfjorden. As already mentioned, this sample is dominated by the underlying hornblende schist, and phyllosilicate is represented by chlorite here. On Hardangervidda the brown micas, in particular, biotite, dominate. There is a tendency towards an increase in green/olive-green micas at the expense of brown varities in the eastern areas.

Except for the sample near Sørfjorden and an area within the Nordmanns-

Table 4. Maximum, minimum, mean (x), standard deviation (S) of number percent coloured phyllosilicates in the 3–2 phi-fraction for phyllitic basal tills (F), ablation tills (A), basal tills (B), and basal tills sampled from areas covered by the map sheets Bj,. . . ., Fr (cf. Fig. 17), respectively. N is number of samples

	max.	min.	X	S	N
F	2.1	0.0	0.71	0.62	28
A	7.8	0.0	1.60	1.75	47
B	27.1	0.0	1.25	2.46	233
Bj	6.6	0.0	0.99	1.10	44
He	4.4	0.0	1.22	1.06	37
Sk	2.5	0.4	1.32	0.89	6
Hå	3.8	0.0	0.89	1.11	9
No	17.5	0.0	2.90	4.72	21
Lå	2.1	0.0	0.83	0.54	33
Ka	3.2	0.0	0.70	0.65	32
So	5.0	0.0	2.06	1.59	7
Bi	2.8	0.0	0.76	0.69	24
Fr	4.6	0.0	1.07	1.21	15

lågen map sheet, there are no major regional variations in the total content of coloured phyllosilicates. The Nordmannslågen map sheet area samples with high mica content lie near Kvenna, and also are fairly rich in heavy minerals. The material is thus probably derived from an unmapped biotite-hornblende schist outcrop.

Grain-size distribution – regional variations

The grain-size distribution and its relation to the genesis of the tills and to the petrography is discussed in Vorren (1977a).

Fig. 31 depicts the average basal tills within the respective map sheet areas. Some of the regional variations are explicable in terms of differing petrographical composition; the greatest deviation from the average till is in areas of phyllite-rich tills. The tills within the Hårteigen map sheet area have an average composition approaching that of the phyllitic tills (Vorren 1977a), as do the tills from the Nordmannslågen map sheet area which locally contain a great deal of phyllite. The tills in the Bjoreio map sheet area are also influenced to a noticeable degree by their phyllite content, giving a rather large gravel fraction.

The basal tills within the Songevatn and Lågaros map sheet areas have a relatively high clay and very fine to medium silt content. These samples were mainly collected from Phase II crag and tails. The relatively large content of fine material might be due to glacial incorporation of older sorted sediments, but the writer would prefer to explain it as a result of a long, stable ice movement (p. 104). Thus, the minor relief forms were quite favourable aligned for this movement and abrasion was the main mode of erosion, which produces the finer grades which are enriched here (cf. Fig. 5).

The average tills from the Bitdal, Frøystul and, to some degree, Kalhovd map sheet areas are characterised by a high sand content and comparatively low clay and gravel contents, though the clay content varies considerably. On

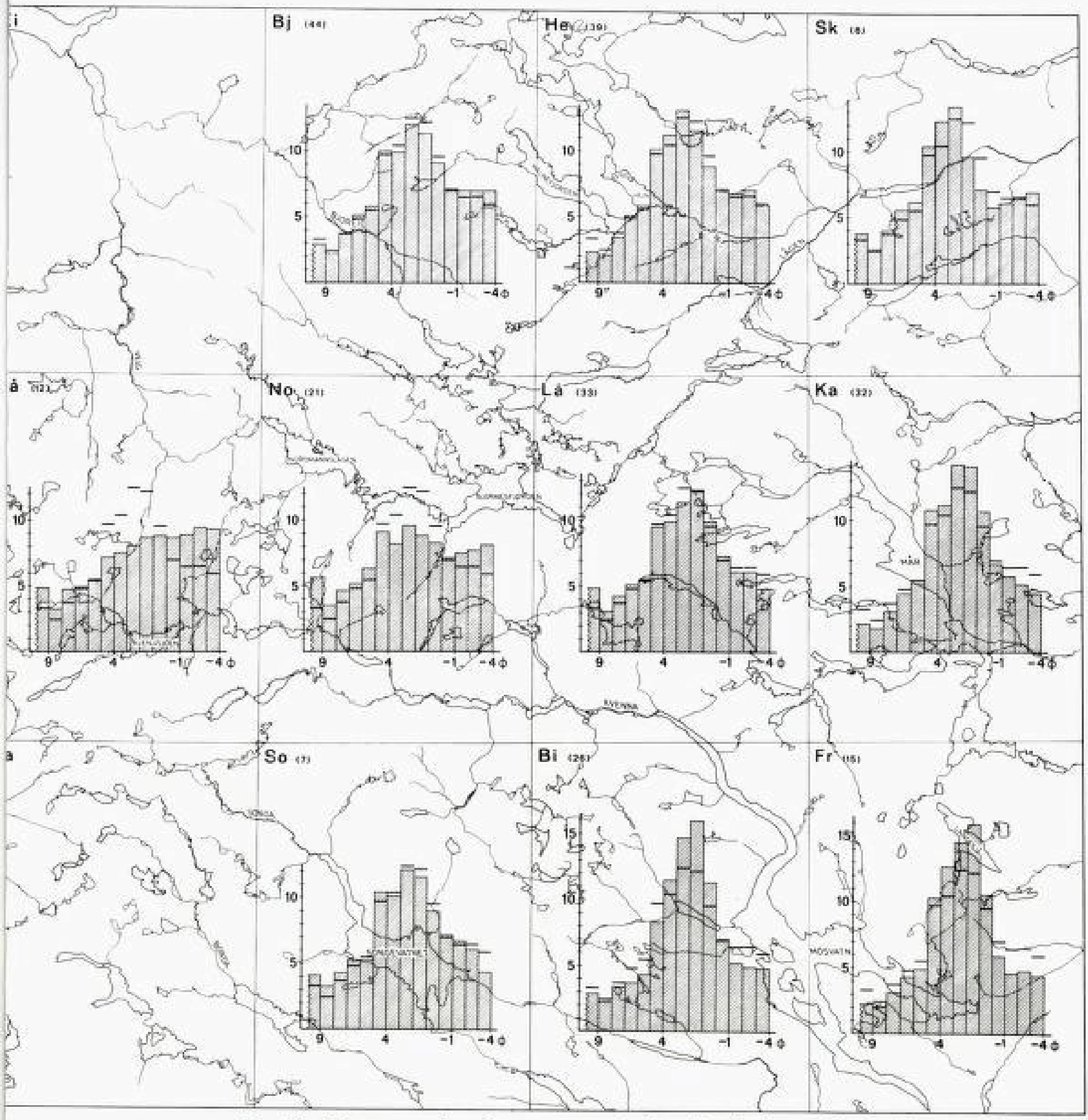


Fig. 31. Histograms showing average grain-size distribution of basal tills sampled on the different map sheets. The number of samples on which the histograms are based is shown in parenthesis. The average grain-size distribution of all basal tills is shown for comparison on each histogram by short horizontal lines.

account of its relatively large phyllite content the tills around Møsvatn have already been interpreted as containing a good deal of resedimented material. This is further indicated by the frequent inclusions of sorted sediments both here and within the Kalhovd map sheet areas (Figs. 19 and 20). There are several localities in these areas where subtill sediments (sand and coarse silt) have been found – see chapter on stratigraphy. The author therefore believes that the large sand contents are the result of earlier deposited sorted sediments being incorporated. Where the subtill sediments are clayey the basal till possesses a high clay content, as at Hovden, Møsvatn (Vorren & Roaldset 1977).

The regional variations in the basal tills' grain-size distribution are therefore believed to be due to several factors, namely their petrographical composition, the mode of erosion and transport distance, and the incorporation of sorted sediments.

Origin of the till

Till and ice movements. The results presented here are partly based on the morphology of the superficial deposits, partly on their lithological composition and partly on stratigraphical investigations (which will be reported in detail later).

During Phase I the phyllite-rich tills near Mår were probably deposited, along with those near Sterra (p. 57). At Møsvatn there are some localities with comparatively clay-rich tills believed to date from the very first Weichselian ice movement phase (Vorren & Roaldset 1977). The tills in the extensive crag and tail field within and around the Lågaros map sheet area (Fig. 11) were deposited during Phase II.

In the area east and south of the Phase III ice divide all the various ice movements have been more or less eastwards. The till deposits here clearly had to have been derived from districts farther west. It is most likely that the bulk was deposited during Phase II since there is little reason to expect very much transport and sedimentation so close to the ice divide during Phase III. As far as Phase IV is concerned it appears to have been of short duration and most likely reworked the upper layers rather than supplied fresh material in this area.

In the northern areas there are signs of Phase II till material too. Samples 27–70, on the west side of Halnefjorden, and 2–74 (Fig. 17) east of the phyllite tract between Halnefjorden and Bjoreio, both have relatively high phyllite contents (Fig. 25). This suggests that they were derived from the nearest phyllite areas to the west. These samples are also distinctive on account of their comparatively high clay contents.

In the area to the east and southeast of Kvennsjøen (Nordmannslågen map sheet) in the Kvenna valley and the neighbouring valley to the south, there are two till types. One has rather a lot of phyllite while the other has only a little (Fig. 25). The low-phyllite samples were probably deposited during a movement towards the east, i.e. Phase II, down the valleys which lie in Precambrian rocks, while the phyllite-rich tills were laid down by a northerly directed ice movement, Phase II and/or IV.

Within central Hardangervidda it is difficult to evaluate which ice movements deposited the till. Frequent occurrences of blue quartz (see for example the sample from the western end of Langesjøen in Fig. 29) do indicate, however, that a not insignificant portion of the till was deposited during Phase II.

Phase III has clearly accounted for quite significant transport and sedimentation of till in the northern areas. Precambrian rocks have been transported into the phyllite area between Bjoreio and Halnefjorden, including quantities of the characteristic granite from around Halnefjorden (Fig. 2). In the areas between Nordmannslågen and Songa the landforms and phyllite distribution show that an appreciable amount of till was deposited during Phase III (and IV?).

Phase IV seems to have been the one during which there was least sedimentation of till. Definite signs of Phase IV till are restricted to the northern areas. For example, at the northern end of Halnefjorden a till section reveals a relatively phyllite-rich till (sample 154–73, Fig. 29B) overlying a till virtually devoid of phyllite (sample 153–73). The upper till must have been transported southwards from the area due north, whereas the lower till was laid down during an ice movement towards the west, i.e. Phase III. Till from Phase IV has also been observed in the Bjoreio area (p. 97). Elsewhere on Hardangervidda some of the recently described transport and sedimentation with northward-flowing ice in the district between Nordmannslågen and Songa may also have occurred during Phase IV.

To summarise, there are scattered localities with till which almost certainly date from Phase I. A significant proportion of the till in the eastern, southern and perhaps central parts of Hardangervidda was deposited during Phase II, and tills from this phase have also been recorded in northern areas. During Phase III, sedimentation occurred in the northern, southwestern and perhaps also the central areas. Till from Phase IV is chiefly restricted to the northern part of Hardangervidda.

Age of the till. There are two schools of thought regarding the age of the Norwegian tills. It is usual to consider the Norwegian tills as erosion and sedimentation products essentially from the last ice age. In contrast to this Rosenqvist (1975) has recently claimed that a 'fair amount' of the till material is derived from Tertiary chemical weathering products. He bases his conclusion to a large extent on the observation that, 'Quartz and phyllosilicates are enriched relative to the other bedrock minerals'. It is clear from his own data, though, that quartz is enriched in relation to the other rock-forming minerals in the sand and coarse silt fractions, while the finer silt and clay fractions are relative quartz-deficient. For example, the 2 to 1 phi fractions have from 45.6% til 60.0% quartz while the clay fractions have only 2–20%. Where phyllosilicates are concerned only the clay fractions are mentioned. The present writer's investigations of coloured phyllosilicates show an average content of 1.25% in the 3 to 2 phi fraction. (When the white mica minerals are included the average will be increased by one tenth or two of a %.) This is far lower than the phyllosilicate content of the clay fraction.

These differences are partly due to different minerals being enriched in different fractions, as shown for example by Dreimanis & Vagners (1971) and Virkkala (1969b). However, some of the clay-sized phyllosilicates probably really are chemical weathering products (Rosenqvist 1975). As demonstrated by Vorren & Roaldset (1977) there has been an advanced degree of soil profile development, with clay mineral formation, during the last interglacial. Furthermore, SEM studies of glacially crushed quartz grains (Fig. 58) reveal that even quartz (probably from a Weichselian interstadial) has been exposed to chemical attack. The chemical weathering process involved does not therefore need to be of Tertiary age.

In the writer's opinion essentially all the till deposits are of younger Pleistocene age, and most of the possible anomalies between the mineralogy of the till and of the bedrock is explicable in terms of Quaternary weathering.

SORTED SEDIMENTS

As Table 1 shows, 177 samples were collected from sorted sediments. Seventy-seven of them are sub- and inter-till sediments, four are glaciolacustrine, seventy-six are from eskers and twenty from glaciofluvial terraces (some are possibly of normal fluvial origin).

Grain-size distribution

The grain-size data are presented in Fig. 32. The overall picture is not truly representative though, as sampling was selective with a bias towards sandy facies of glaciofluvial deposits to facilitate petrographical comparisons. Furthermore, the finer-grained representative of sub- and inter-till sediments were preferred because of their greater potential for microfossil preservation. The frequencies of the different textural types sampled is thus hardly a reflection of their occurrences in the field.

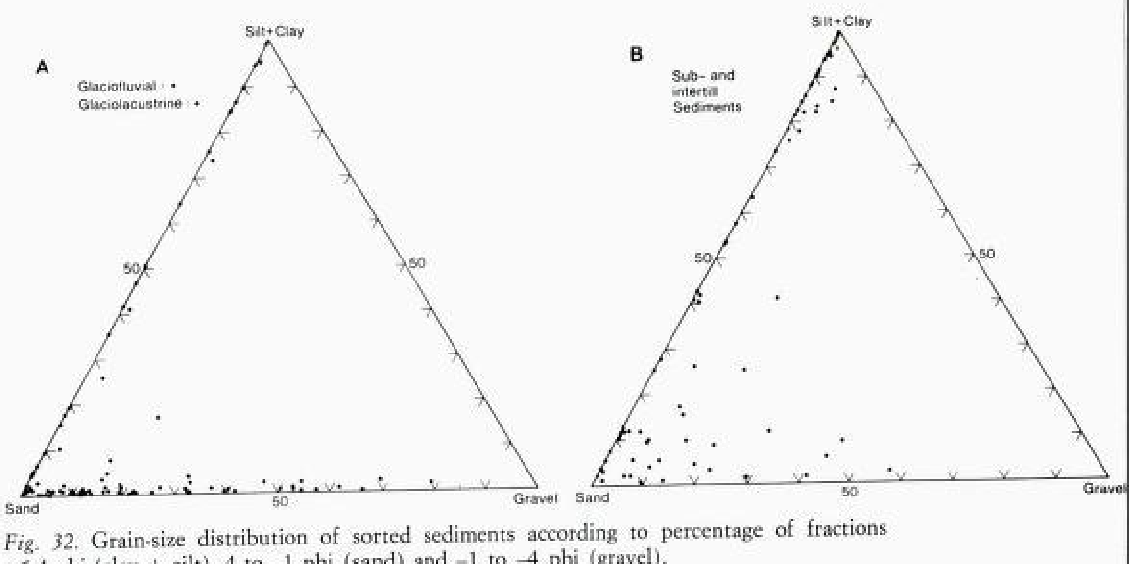


Fig. 32. Grain-size distribution of sorted sediments according to percentage of fractions < 4 phi (clay + silt), 4 to –1 phi (sand) and –1 to –4 phi (gravel).

The diagrams, nevertheless, do indicate the large dispersion of grain-size distributions which exists in the sub- and inter-till sediments and points to a variety of origins for them. Some of the finest-grained examples (chiefly sub-till) are most likely of glaciolacustrine/lacustrine origin. The remainder were probably originally deposited as glaciofluvial/fluvial sediments.

Fig. 32 shows that, naturally enough, glaciolacustrine sediments are rich in silt/clay, but also that glaciofluvial sediments can be rather fine-grained too. The other glaciofluvial deposits are predominantly sandy and gravelly.

Grain-size distribution parameters

In Table 5 the extreme values for the different grain-size distribution parameters are listed. A wide range is revealed between the various parameters for all sediment types, with the exception of the Mz and σ values for glaciolacustrine sediments — which are represented by just 4 samples. The selective sampling invalidates any discussion of average parameter values and possible concentrations, but the relations between the different parameters can be examined independently of this. The four parameters are plotted in pairs. (Samples containing > 15% clay fraction are omitted as extrapolation is too unreliable for the lower percentages used in the formulae.) The Mz – σ and Mz – Sk relations are the only apparently systematic ones and the discussion will be limited to these.

Mz – σ. As Fig. 33 shows there is a clear trend, the best sorted sediments having a mean in the fine to medium sand grades, between 3 and 1 phi. Sub- and inter-till sediments have a σ-minimum between 2 and 3 phi, while for glaciofluvial sediments the corresponding minimum is about 1.5 phi. Sediments with Mz in finer or coarser fractions are generally less well sorted. There are distinct exceptions, however, chiefly from intertill sediments where one suspects that originally sandy deposits have become contaminated with other fractions during glacial transport.

That fine sand sediments are the best sorted ones has been demonstrated earlier for fluvial, eolian, littoral and marine sediments, e.g. Inman (1949), Folk & Ward (1957), Walger (1962), Dyer (1970) and Folk (1974). The same has been shown to be partly true for glaciofluvial outwash sediments (Slatt & Hoskins 1968). This relation seems to be almost ubiquitous for sorted sediments, and requires a universal explanation.

Inman's (1964) explanation is based on hydrodynamical considerations. Folk (1974) discovered that the curve expressing the Mz – σ relationship is sinus-

Table 5. Maximum and minimum values of grain-size parameter from glaciofluvial (A), sub/intertill (B) and glaciolacustrine samples (C)

	Mz min	Mz min	σ min	σ max	Sk min	Sk max	Kg min	Kg max
A	6.86	– 1.86	0.45	2.72	– 0.39	0.52	0.64	1.48
B	> 10	– 2.26	0.69	3.32	– 0.42	0.42	0.80	2.13
C	7.29	5.44	1.36	2.12	– 0.01	0.16	0.78	1.32

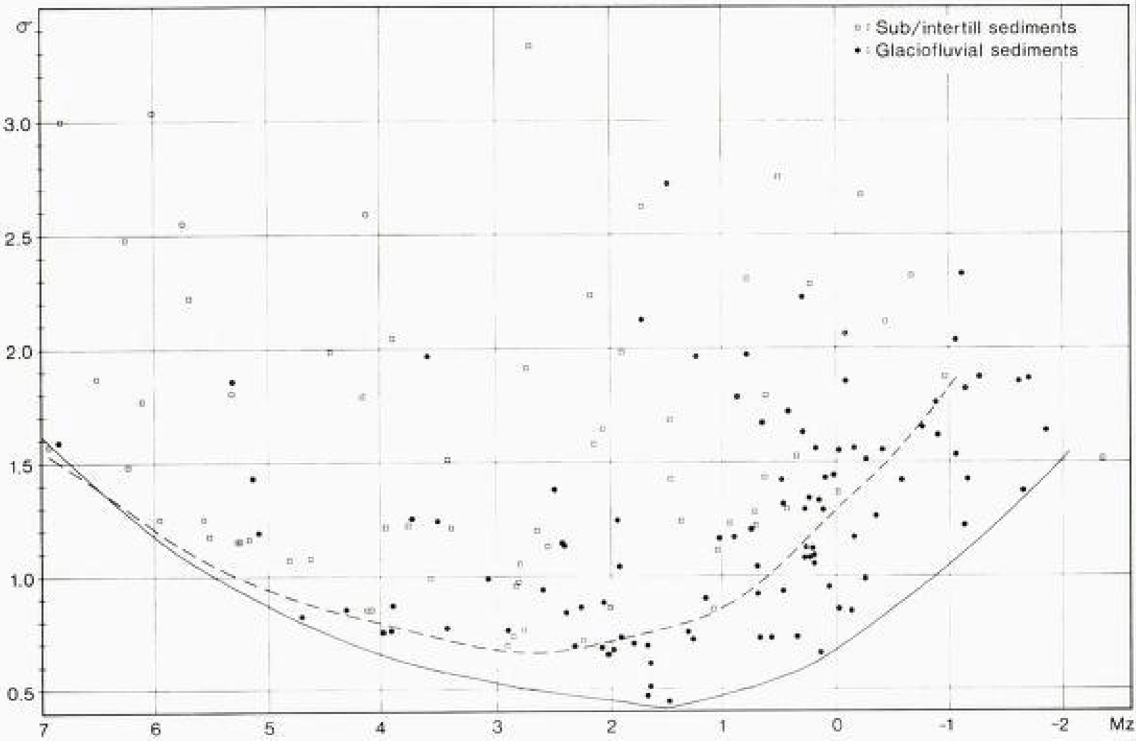


Fig. 33. Scatter plot of mean size (Mz) versus sorting (σ) for 61 sub/intertill samples and 96 glaciofluvial samples. Dashed lines demarks the lower boundary of sub/intertill samples; continuous line the lower boundary of glaciofluvial samples.

oidal, with poor sorting associated with mean values in the fine gravel/coarse sand fractions and in the silt fractions. He pointed out that in nature there is usually a paucity of these latter grain-sizes, so that samples whose mean is located there must be mixtures either of gravel and sand or of sand and clay. This would result in poor sorting. Perhaps both explanations are right. Concerning Folk's proposal, it is interesting to note that the till from which much of the sorted sediment is probably derived often has the –1 phi minima he drew attention to (Vorren 1977a).

Mz – Sk. This relationship also is defined by a sinusoidal type of curve (Fig. 34). A curve resembling that in the right-hand part of the diagram was earlier reported by Folk & Ward (1957) from fluvial sediments whose mean lay between 3 and –3 phi. For Mz values larger than 6 phi and around 1 to 0 phi, most Sk values are negative (Table 6). The first of these minima is attributed to a tail of coarser particles, interpreted as ice-drop material in glaciolacustrine environment. The other minimum is mainly represented by sandy glaciofluvial material containing gravel.

Skewness is usually positive for Mz values between 6 and 2 phi and between –1 and –3 phi. The former group probably comprises sediments whose coarser fractions were transported as bed load while the finer-grained tail was deposited

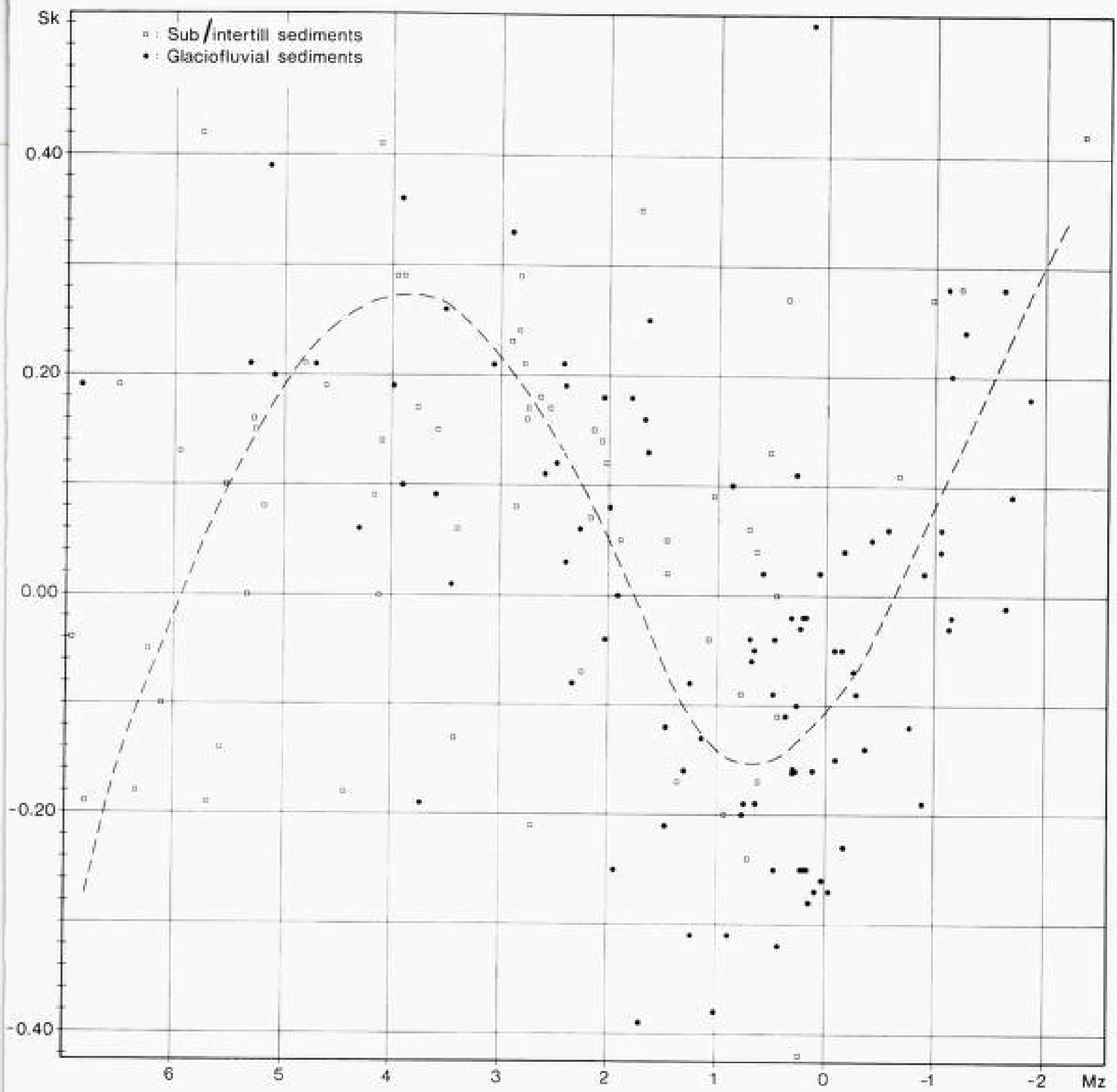


Fig. 34. Scatter plot of mean size (Mz) versus skewness (Sk) for 67 sub/intertill samples and 96 glaciofluvial samples. The dashed line gives the main trend visually determined.

from suspension. In the latter group the gravel portion of a gravel and sand mixture was dominant.

Table 6. Relationship between Mz and Sk for 163 sorted sediments from Hardangervidda

Mz:	7		6		5		4		3		2		1		0		−1		−2		−3
Sk > 0		2		9		7		12		21		11		9		6		9		1	
Sk = 0				1		1						1		1		1					
Sk < 0		6		2		1		2		4		11		31		11		3			

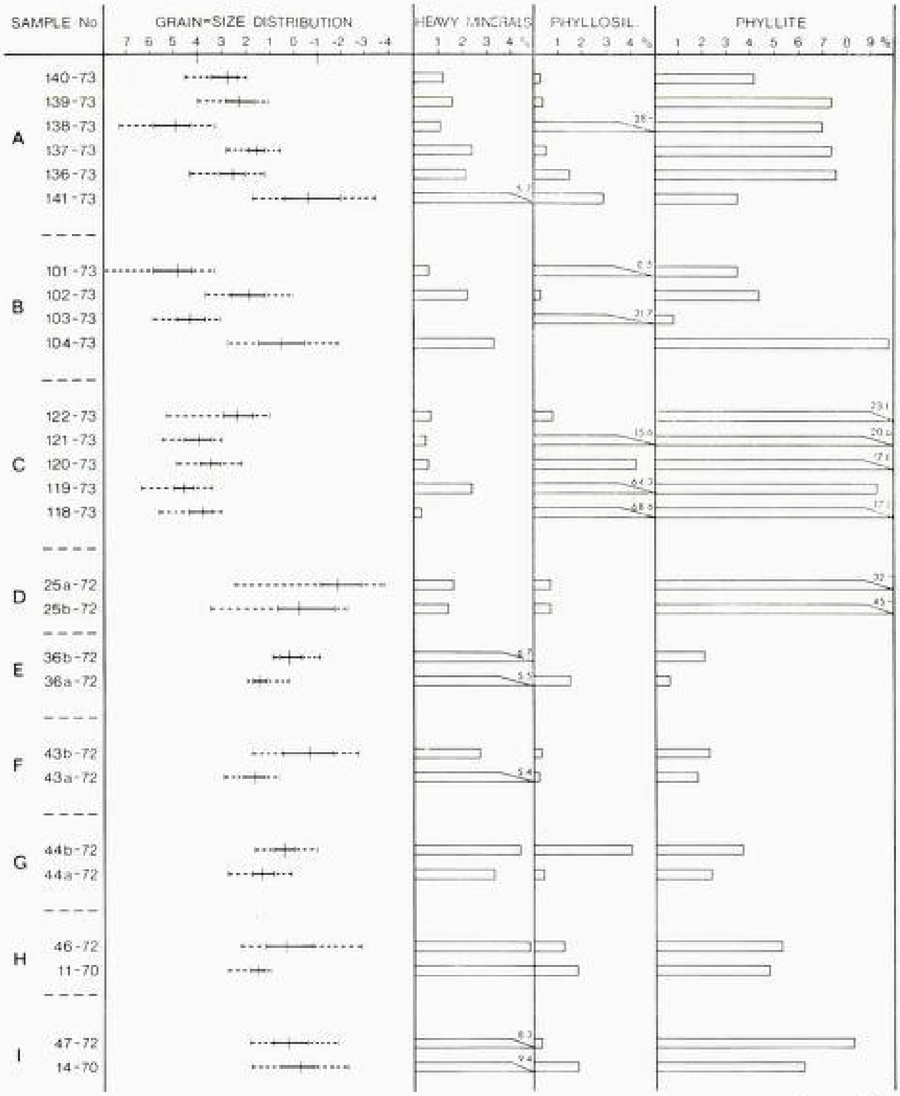


Fig. 35. Relationship between grain-size distribution and content of heavy minerals (weight percent), coloured phyllosilicates (number percent) and phyllite (number percent) in the 3 to 2 phi fraction. The samples are from glaciofluvial sediments at 9 different localities, A to I. Cf. Fig. 43 for explanation of grain-size distribution.

Samples whose mean is in the 2 to –1 phi range display a tendency to symmetrical grain-size distribution. Sediments with Mz around 2 phi are, as already shown, the best sorted and consist of nearly pure sand.

Petrography and grain-size distribution

Some of the petrographical analyses are presented in Fig. 35 together with grain-size distribution. The samples are from different sedimentation units in glaciofluvial deposits from 9 localities.

The heavy mineral content clearly varies not only from locality to locality but also between different sedimentation units from the same locality. For localities A, B, D, E, G, H and I, i.e. a majority of them, there is a plain correlation; increasing heavy mineral content with coarser sediments. Sample 119–73 at locality C is anomalous, and localities F and H also fall outside the general pattern.

The phyllosilicate content also displays both inter- and intra-locality variation, though samples from the same locality show a clear tendency for finer sediments to contain more phyllosilicates than the coarser ones. There are exceptions here, the most notable being at localities G and I.

The phyllite content varies from locality to locality, but the variation between sedimentation units from a single locality does not seem to be related to the grain-size distribution.

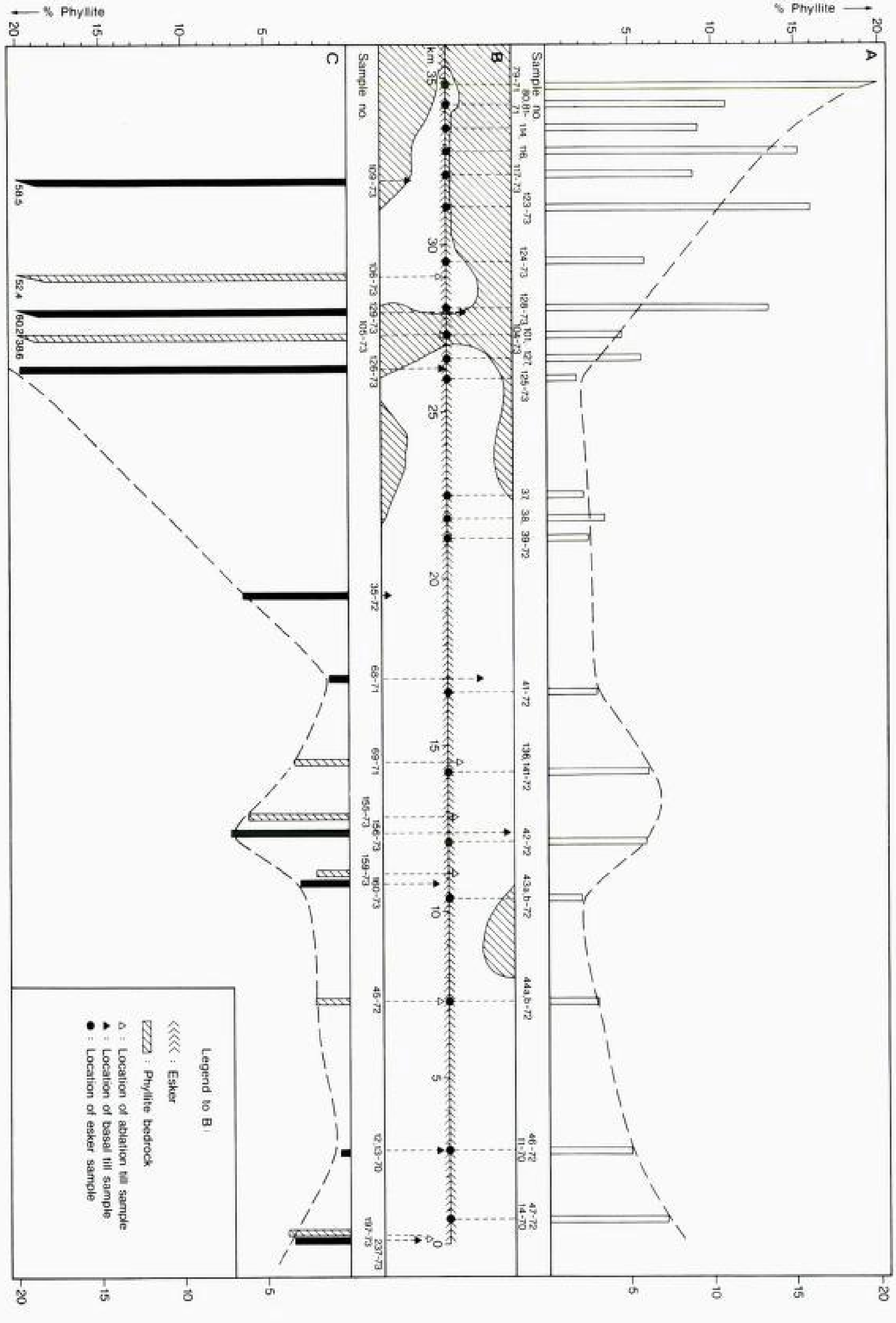
The established relations between heavy minerals and phyllosilicates and the grain-size distribution are explicable in terms of their hydraulic equivalents (Pettijohn 1957). The heavy minerals are deposited along with larger felsic minerals, a well-documented phenomenon, e.g. Rittenhouse (1947), Briggs (1965) and Hand (1967). The phyllosilicates are deposited in the company of smaller felsic minerals because of their platy shape (Pettijohn 1957, p. 565).

Since phyllite grains have no systematic relation to the grain-size distribution, one can assume that they have an hydraulic equivalent similar to the felsic minerals; this is supported by their specific weight (p. 39) and shape. This relationship also may explain some of the heavy mineral and mica content anomalies. Sample 119–73 with abnormally low heavy mineral content has very little phyllite compared with locality C's other samples. This indicates that the material was derived from deposits of a different petrographical composition, i.e. with more heavy minerals and less phyllite. The anomalously high phyllosilicate content in sample 141–73 can ,in principle, be explained the same way.

Glaciofluvial transport

The dependence of the content of both heavy minerals and mica on grain-size distribution shows that neither group can be employed in a study of transport distances. Phyllite seems much more suitable, though it too is limited — by some variation within one and the same locality. The average of a number of samples should be used, or sampling must be dense. The samples collected from the large Halne–Eitro esker more or less meet these demands. This esker is almost continuous from the southern end of Halnefjorden to Eitro (Pl. 2). In Fig. 36B the esker is drawn out into a straight line and the extent of phyllitic bedrock within a 2 km-broad zone on either side is shown, together with sampling localities and their phyllite content (Fig. 36A and C). Some of the per cent values cited are averages of two or more samples.

The average curves for phyllite content (Fig. 36) are rather similar for both till and glaciofluvial samples. However, the phyllite curve for esker material is displaced in the direction of transport relative to the till curve. From 0 to 10 km there is more phyllite in the glaciofluvial sediments, which must be due


% Phyllite
A
B
C
Sample no.
km
Legend to B
: Esker
: Phyllite bedrock
: Location of ablation till sample
: Location of basal till sample
: Location of esker sample

to phyllite being brought in from more phyllite-rich areas farther north around Halnefjorden. At c. 12 km there is a phyllite maximum in the till samples, while the esker's corresponding maximum is displaced c. 1.5 km downstream. In the phyllite area to the west the phyllite content of the till rises quickly to > 30% at c. 27 km, but the associated esker high does not appear until 35 km, a phase displacement of 8 km. This displacement in fact reflects the transport distance not of phyllite but of the other components (mostly quartz and feldspar) within the 3 to 2 phi fraction.

Hellaakoski (1930, p. 32) discovered a c. 4 km phase displacement for fine gravel from Rapakivi granite and other crystalline rocks. Lee (1965) found that gravel-sized fragments of trachyte, dunite and magnetite attained maximum frequencies in esker material after 5–13 km transport from their bedrock sources (till was not examined). Gilleberg (1968), however, reported that silt, sand and gravel particles from Cambro– Silurian bedrock in south Swedish eskers was of '. . . extremely local origin'.

The present results suggest transport distances for fine sand material varying from 1.5 to 10 km. The esker material's composition will be dependent on several factors including degree of comminution, supply of local material (dilution effect), the nature of the esker system (sub-, supra- or englacial). In the area from 0–10 km the higher phyllite values in the esker indicate a fairly long transport. This might be due to the dilution effect from local material being weak here. The displacement of the maximum at 12 km shows a short transport, 1–2 km, which may reflect a large supply of local material, especially as there is little reason to expect that comminution was more active here than in other parts of the esker. The relatively low phyllite content in the western end of the esker indicates that the majority of the material there was supplied distally. Nevertheless a sudden increase in the phyllite frequency shows that some of the material must be of rather local origin. The Halne–Eitro esker seems to show that the esker drainage was not a closed system, but the number and capacity of tributary channels has varied from place to place. Furthermore, the esker petrography's overall agreement with the surrounding till points to a subglacial formation of most parts of the esker.

Stratigraphy

Several localities with subtill and intertill sediments have been found on Hardangervidda. Two with interstill sediments have been described previously (Vorren & Roaldset 1977). In the following section, localities with sub- and intertill sediments and with double tills will be described which throw some light on Hardangervidda's Quaternary stratigraphy. Fig. 3 shows the localities which will be discussed.

Fig. 36. Phyllite percentages in the Halne–Eitro esker, A, and in nearby tills, C. Diagram B givet the distance scale along the esker axis and the bedrock in a strip two km either side of the esker. The esker drainage was from right to left.

Pollen and spore samples analysed were prepared according to the method described by Vorren (1972, p. 237), while the actual identification was done by the help of Fægri & Iversen (1966) and the reference collections at the universities in Bergen and Tromsø.

DESCRIPTION OF LOCALITIES

The Møsvatn area

Førnes and Hovden. (Fig. 3, locs. 1 and 2). These are the two localities described in Vorren & Roaldset (1977). Based on various litological characteristics, palynomorphs and the deposits' relation to ice movement phases, the following informal lithostratigraphical correlations and their equivalent climatostratigraphical units were proposed:

Førnes upper till	→ Hovden upper till	→ Førnes kryomer
Førnes intertill sediments	→ Hovden sand	→ Førnes thermomer
Førnes lower till	→ Hovden lower till	→ Hovden kryomer
	→ Hovden clay	→ Hovden thermomer

During the Hovden thermomer, which is oldest, the vegetation in the dictrict around Møsvatn was characterized first by spruce and then later by birch. Soil profile development was at an advanced stage with the formation of dioctahedral illite.

In contrast, the Førnes thermomer produced only a sparse vegetation dominated by grasses, and it is doubtful whether any trees were growing in the Møsvatn area. Soil profile development was slight.

Trengsle. (Fig. 3, loc. 3). The following stratigraphy was visible in a c. 15 m-high section section on the shore at Trengsle:

Top — a thin (0.3–0.4 m) layer of diamicton, interpreted as basal till.

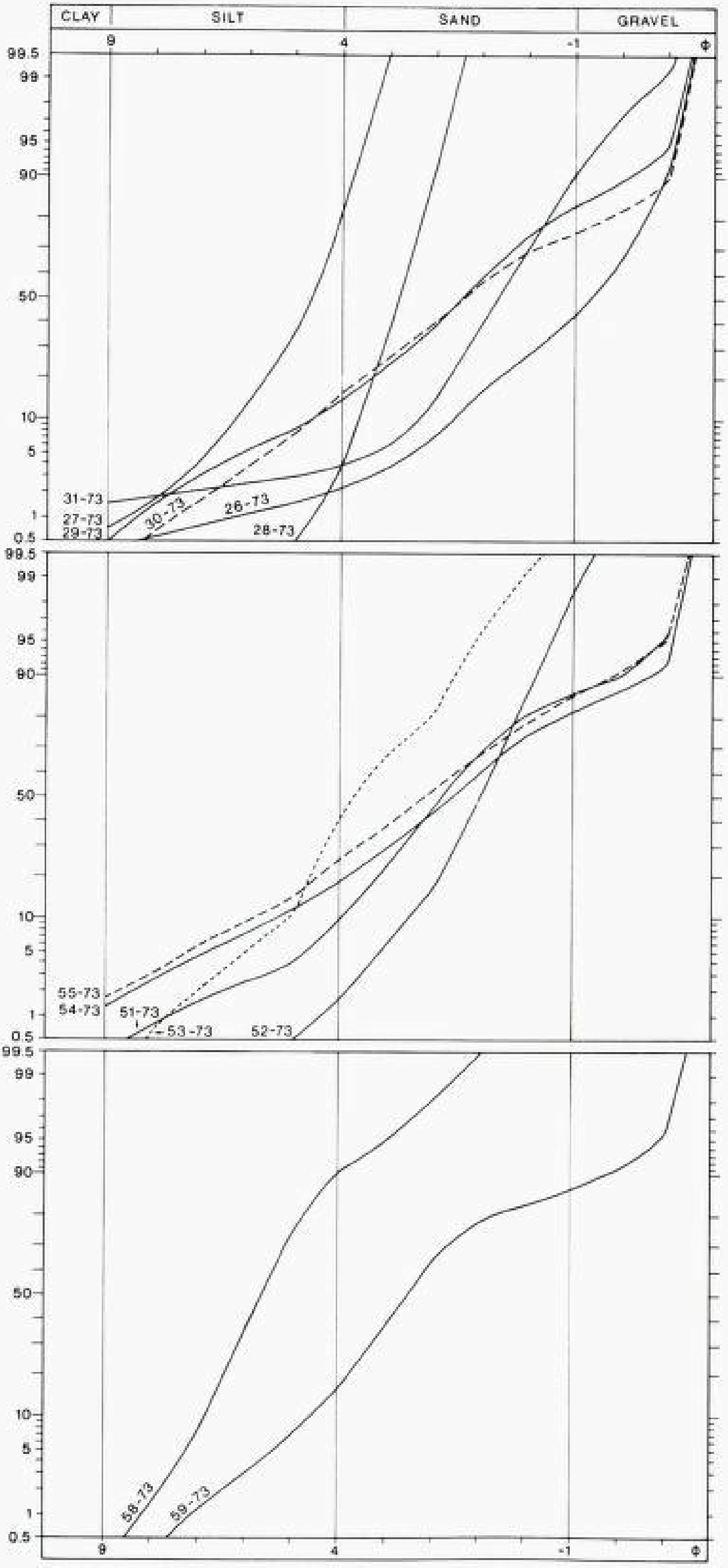
Middle — a 3–4 m-thick sequence of strongly folded sorted sediments (Fig. 38). The folds are asymmetrical with their axes aligned roughly E–W. The sediments consist partly of gravel, partly of sand and partly of silt beds (Fig. 37); samples 26–, 27– and 31–73.

Base — separated from the overlying folded strata by a tectonic discordance, a unit of vertically inclined beds of medium to fine sand; see Fig. 37, sample 28–73. The beds strike approximately E–W.

Two more sections have been observed, north of the Trengsle locality, with till on top of sorted sediments. The till is fairly sandy and contains a lot of rounded gravel and cobbles (Fig. 37); samples 29– and 30–73. The heavy mineral (HM), phyllite (P) and phyllosilicate (M) contents of the respective samples in the 3 to 2 phi fraction are presented below.

Sample No.	HM	P	M	
30–73	6.1	0.5	2.7	
29–73	5.7	1.0	2.8	till
31–73	6.0	4.4	4.2	
26–73	5.9	4.0	8.1	subtill
27–73	1.0	1.2	69.1	sediments
28–73	1.7	0.4	2.1	

Fig. 37. Cumulative grain-size distribution of sediments from the Møsvatn area drawn on probability percent ordinate. Upper diagram shows sediments from Trengsle; 29–73 and 30–73 represent basal tills and the rest subtill sediments. Middle diagram shows sediments from Laksastøl; 51–, 54– and 55–773 represent basal tills while 52– and 53–73 are intertill sediments. Lower diagram shows basal till, 59–73, and subtill sediment, 58–73, from Falkeriset.


CLAY
SILT
SAND
GRAVEL
9
4
-1
Φ
99.5
99
95
90
50
10
5
1
0.5
31-73
27-73
29-73
30-73
26-73
28-73
55-73
54-73
51-73
53-73
52-73
58-73
59-73

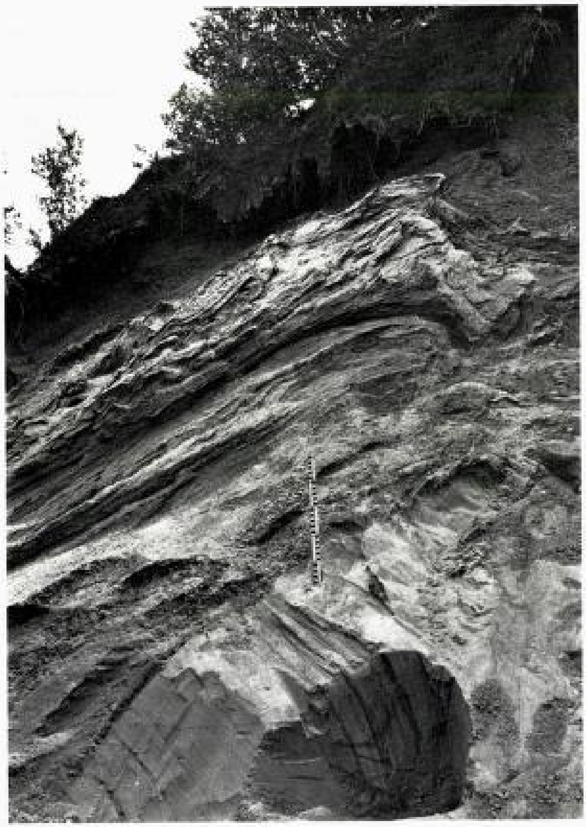


Fig. 38. Deformed subtill sediments at Trengsle. Nearly upright beds can be seen lowermost in the picture. An asymmetric fold can be seen above these beds bounded by a thrust plane lying near the top of the scale. The scale is 0.5 m long.

The extremely high phyllosilicate content in sample 27–73 is due to hydraulic separation; compare the grain-size distribution curve and the low content of heavy minerals. One should in addition note that the coarsest subtill sediments have the highest phyllite contents.

The Trengsle stratigraphy is interpreted as follows. The subtill sediments were originally laid down as part of a fluvial delta or fan at the northern end of Møsvatn. The comparatively high phyllite content indicates that the river Kvenna, rising in the phyllite area, supplied the material. That the phyllite content is greatest in the gravelly samples is reasonable enough since erosion and transport capacity must have been at their peak during maximum discharge. Later these deposits were deformed. Both the strike of the lower, vertically inclined beds and the fold axis orientation of the sequence above them witness N–S compressive deformation forces. The major fold (Fig. 38), interpreted as a drag fold, show that this movement was southwards. The sediments have evidently been thrusted and folded by a glacier moving southwards.

The till deposited here during this (and subsequent) ice movement is shown by the grain-size analyses to have incorporated a good deal of these sediments; the roundness of the material supports this too.

There are two alternative correlations of the Trengsle subtill sediments possible with the Førnes and Hovden stratigraphy. Either they are equivalent to the Hovden clay or to the Førnes intertill sediments/Hovden sand. It is difficult to draw a correlation on a lithological basis. Both the lithostratigraphical units have similar phyllite contents. Other lithological characteristics such as texture are unsuitable for correlation purposes since they are facies dependent. Overall considerations suggest that since the Trengsle subtill sediments are

Fig. 39. Diamicton overlying strongly disturbed sand in Bitdal. The diamicton is probably glacially resedimented sorted material.

overlain by just one till bed, they are most likely contemporary with the youngest subtill sediments, namely the Førnes intertill sediment/Hovden sand. This implies a climatostratigraphical correlation with the Førnes thermomer.

Laksastøl. (Fig. 3, loc. 4). There is a c. 20 m-high section at the shore, and most of the exposed sediment is till; Fig. 37, samples 51–, 54– and 55–73. Approximately 1 m below the top there is an elongated lens, up to 1 m thick, of disturbed stratified sediments; Fig. 37, samples 52– and 53–73. An interesting feature of the sediments in the lens is that they have the same yellow colour characteristics of certain beds in the Hovden sand. The adjacent till shows signs of having incorporated some sorted sediments, notably sample 51–73. Of the two till types found at Hovden (Vorren & Roaldset 1977) the Laksastøl one most resembles the Hovden upper till.

The intertill sand was most likely glacially thrust into its present position when the surrounding till then being correlatable with the Hovden upped till.

Falkeriset and Bitdal. (Fig. 3, locs. 5 and 6). Subtill sediments have been found at Falkeriset. They can be seen to be at least 1 m thick and are essentially silty; Fig. 37, sample 58–73.

At Bitdal there are strongly disturbed sand-silt sediments beneath a diamicton (Fig. 39). The latter deposit includes mainly rounded/well-rounded cobbles and is obviously reworked fluvial/glaciofluvial material for a large part.

Neither site has produced definitive evidence of their ages, though general considerations suggest that the subtill/subdiamicton sediments belong to the Førnes thermomer.

The Mår area

There have been two well-defined ice movements across the tracts around the northern end of Mår (Fig. 3, loc. 7 and Fig. 10). The sediments around Mår (Fig. 40) consist of a locally rather thick till cover. Scattered minor glacio-

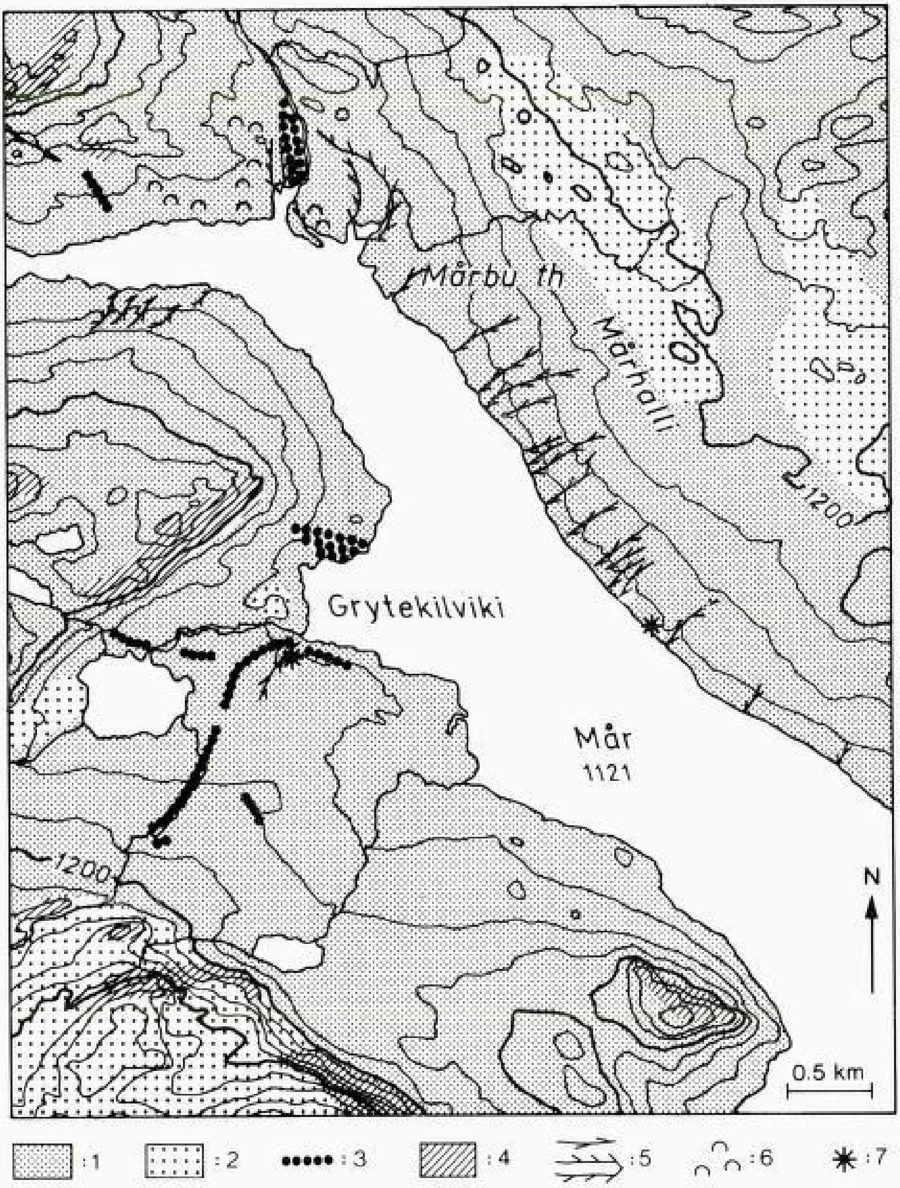


Fig. 40. Glacial geology map of part of the Mår lake area. 1: Till, 2: Thin and/or discontinuous cover of till, 3: Esker, 4: Exposed bedrock, 5: Glaciofluvial drainage channels, 6: Hummocky moraine, 7: Localities with subtill sediments at Grytekilviki and double tills at Mårhalli.

fluvial deposits, mostly eskers, are found too. Particularly along the eastern side of Mår there are numerous glacial drainage channels.

Below, two sites will be described, Mårhalli on the eastern side of the lake and Grytekilviki on the western.

Mårhalli. There are several sections up to 6–8 m high in the superficial deposits on the shore at Mårhalli, cut by wave erosion during high water levels in the regulated lake. At and close to the locality plotted in Fig. 40 there are two till beds exposed. Some of their

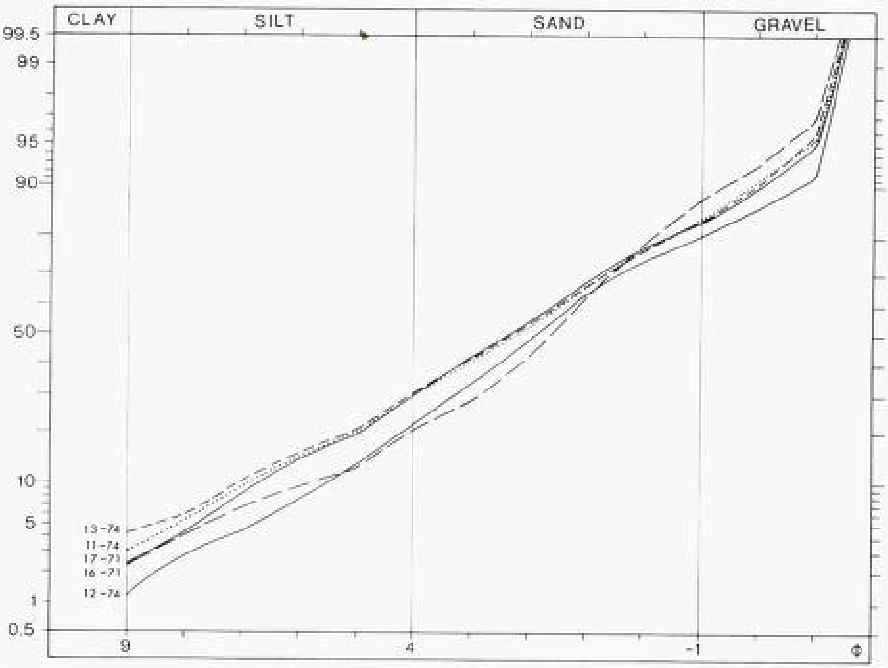


Fig. 41. Cumulative grain-size distribution drawn on probability percent ordinate; 17–71, 11 and 13–74 represent Mår lower till; 16–71 and 12–74 represent Mår upper till. The samples were collected on the eastern bank of Mår Lake.

lithological characteristics are listed below, the upper till being represented by samples 12–74 and 16–71 and the lower one by samples 17–71, 11– and 13–74.

No.	Colour	Clay	Silt	Sand	Gravel	HM	P
12–74	5Y7/2	1.2	21.0	58.2	19.6	4.0	1.6
16–71	5Y7/1	2.3	18.2	66.5	13.0	2.4	1.4
13–74	5Y5/1	4.3	26.4	53.3	16.0	3.2	4.1
11–74	5Y6/1	3.0	27.2	54.1	15.7	2.4	3.4
17–71	5Y6/1	2.3	29.0	52.6	16.1	2.0	5.6

In this table the heavy mineral contents (HM) are in weight per cent of the 3 to 2 phi fraction and the phyllite contents (P) in number per cent of the same fraction. The colour codes are from the Munsell Rock Color Chart and refer to room-dried samples.

The lower till is well consolidated, sometimes to the extent that samples must be hacked out. Locally it possesses a fissile structure (Fig. 21). It is dark grey when wet and grey after room-drying. It is relatively rich in silt and clay, with an homogeneous textural composition (Fig. 41). The phyllite content is unusually high in relation to the transport distance (p. 54).

The upper till is less consolidated and is light grey, both when wet and dried. It contains frequent lenses of sorted sediments (Fig. 20). The grain-size distribution is more variable than that of the lower till and on average it has less silt (Fig. 41) and a markedly lower phyllite content.

Pollen analyses were made of the respective samples. No pollen at all was found in 16–71, 12– or 11–74. Sample 13–74 revealed a *Betula* and a couple of unidentifiable grains. Sample 17–71 contained relatively many pollen, of which 80–90% could be identified. The results are shown in Table 7. The variapollen is *Polygonum* sect. *Bistorta*.

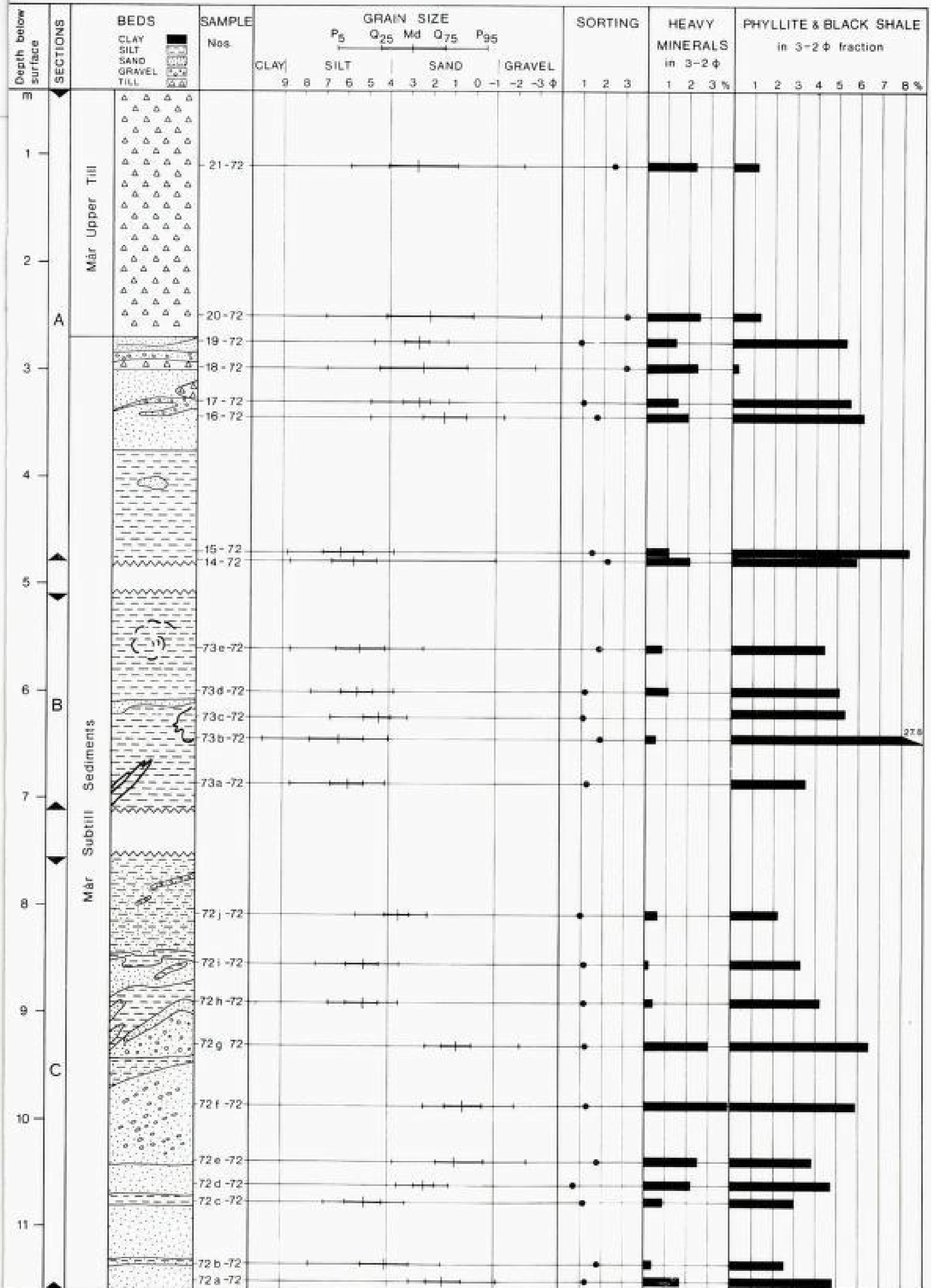
Grytekilviki. In the southwest side of Grytekilviki the meltwater drainage eroded a 10–15 m deep channel in the superficial deposits. In the southern channel side three exposures were

Fig. 42. Photograph, looking south, showing the locality of the subtill sediments at Grytekilviki, Mâr. The three sections investigated (cf. Fig. 43) are shown. The scale is given by the person (circled) at the lowermost section.

Table 7. Palynomorphs from sediment samples collected in the Mār area

	14–72	15–72	73f–72	73b–72	72h–72	72c–72	72b–72	17–71	X
Alnus	–	0.2	–	–	–	3.4	0.8	–	2.7
Betula	6.1	15.3	6.6	3.3	6.4	17.2	19.7	18.0	27.3
Corylus	1.5	1.1	–	0.7	2.1	6.9	3.9	1.0	1.8
Picea	0.2	0.4	–	–	–	–	–	–	–
Pinus	0.2	3.5	–	–	4.2	3.4	7.9	2.0	7.3
Ulmus	–	–	–	–	–	–	1.6	–	–
Σ AP	8.0	20.5	6.6	4.0	12.7	30.9	33.9	21.0	39.1
Ameria type	0.7	0.4	0.7	0.7	2.1	–	–	2.0	–
Artemisia	–	0.6	–	1.6	–	–	–	1.0	0.9
Caryophyllaceae	3.4	7.7	2.0	1.3	6.4	3.4	2.4	5.0	0.9
Compsitae	4.4	3.3	2.0	3.9	8.5	3.4	4.7	5.0	3.7
Cyperaceae	0.2	2.2	1.3	1.0	–	–	–	4.0	5.5
Epilobium	0.7	–	–	–	–	3.4	–	2.0	–
Ericales	–	1.8	–	0.3	–	–	–	6.0	1.8
Gramineae	80.7	60.0	83.6	84.0	70.2	58.6	59.1	49.0	39.1
Ranunculus	1.5	1.8	4.0	2.6	–	–	–	4.0	2.7
Thalictrum	0.2	0.6	–	–	–	–	–	–	1.8
Varia	–	1.1	–	0.7	–	–	–	1.0	3.6
Σ NAP	91.8	79.5	93.6	96.1	87.2	68.8	66.2	79.0	60.9
Dryopteris typ.	0.2	0.2	–	0.6	7.8	75.2	3.0	10.9	7.3
Lycopodium	0.2	–	–	–	–	–	–	2.5	1.6
Sphagnum	0.2	0.4	–	0.6	–	0.8	2.2	2.5	0.8
Spores	0.6	0.6	–	1.2	7.8	80.0	5.2	15.9	9.7
TOTAL (NO)	412	545	152	311	51	121	134	119	123

Fig. 43. Stratigraphy and some lithological parameters of the sediments in Grytekilviki (cf. Fig. 42).


Depth below surface
m
SECTIONS
BEDS
CLAY
SILT
SAND
GRAVEL
TILL
SAMPLE Nos.
GRAIN SIZE
P5 Q25 Md Q75 P95
CLAY
SILT
SAND
GRAVEL
9 8 7 6 5 4 3 2 1 0 -1 -2 -3 Φ
SORTING
1 2 3
HEAVY MINERALS in 3–2 Φ
1 2 3 %
PHYLLITE & BLACK SHALE in 3–2 Φ fraction
1 2 3 4 5 6 7 8 %
Mår Upper Till
Mår Subtill Sediments
A
B
C
1
2
3
4
5
6
7
8
9
10
11
21-72
20-72
19-72
18-72
17-72
16-72
15-72
14-72
73e-72
73d-72
73c-72
73b-72
73a-72
72j-72
72i-72
72h-72
72g 72
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72b-72
72a-72
27.8

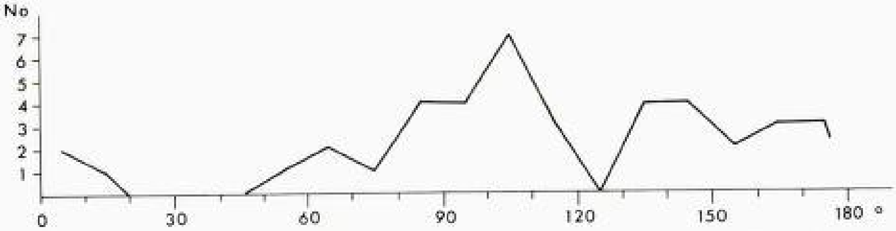


Fig. 44. Fabric analysis of 40 cobbles in Mår upper till at Grytekilviki. The values are grouped and represented in 10°-intervals.

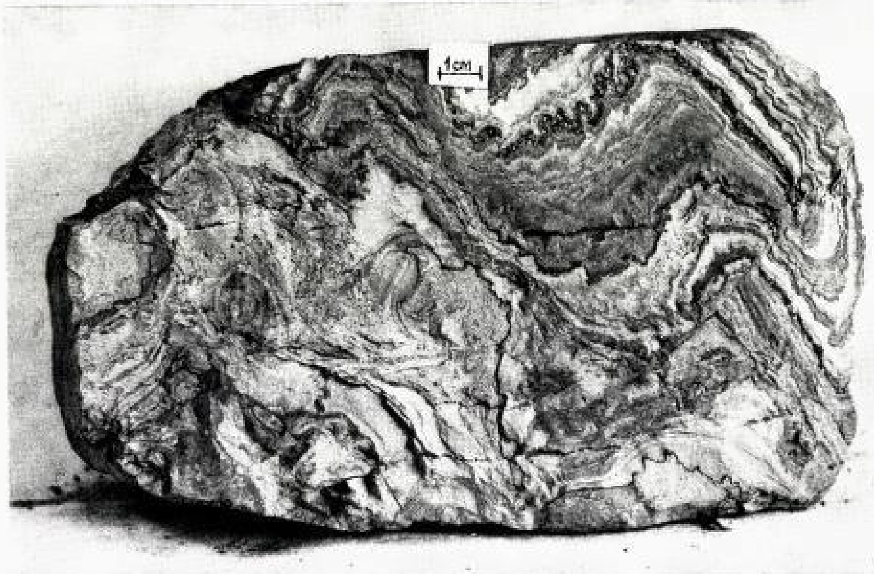


Fig. 45. Sample of strongly disturbed subtill sediments from Grytekilviki, Mår. The fold axes trend approximately towards N (into the picture). The fault plane to the right strikes N–S and dips towards W. Deformation directed towards the east is indicated.

cut at successively lower levels (Fig. 42). The stratigraphy and lithology reveal an upper till (Mår upper till) with underlying sediments (Mår subtill sediments) of varying composition (Fig. 43).

The moraine surface here is even, though from aerial photographs flutings, oriented towards c. 140°, can be observed just south of the sections (Fig. 10). The till is almost 3 m thick, normally consolidated and light grey, with a varying but chiefly sandy texture. The phyllite content is comparable with the upper till at Mårhalli. A fabric analysis was made 1–2 m below the surface, on stones with long axes of 3 to 12 cm. The analysis was complicated by the equidimensional nature of many of the stones. The result of 40 measurements, on stones having their longest axis at least twice as long as their median axis, indicates a rather wide dispersion, although two minor maxima may be discerned around 100–110° and 140° (Fig. 44). These indicate that the till was deposited by an ice movement towards the east, while a partial reorientation occurred during the youngest movement.

The subtill sediments have been strongly disturbed and display intricate structure patterns. At the 73a–72 sampling point there are partly sheared folds (Fig. 45) indicative of pressure from the west. A section immediately below the till, to the west of section A, where the till cover is thinner, reveals drag folds with axes oriented 240°–60°, indicating pressure towards c. 150°. The sediments therefore appear to have been deformed by both an eastward and the youngest ice movements. There are a whole series of structures

Fig. 46. Deformed subtill sediments from Grytekilviki, Mår. Clastic dykes of clayey composition (black) and isolated grains of the same material can be clearly seen.

present in addition to the folds and faults. Thin veins (clastic dykes) of dark clayey material have been found cutting in part across the primary bedding at several places. Elsewhere there are isolated lenses of the same material (Fig. 46). Presumedly the clayey material was mobilised and sequeezed up into frost cracks developed in the surrounding, more frost-sensitive, silty material. Some of these structures may be associated with glaciotectonic movements.

The subtill sediments in the lower and uppermost parts of the sequence are predominantly sand with some gravel, but the remainder is medium to coarse silt. The heavy mineral content is generally low, in common with the other sediments on Hardangervidda. The variations between individual samples appear to be entirely due to hydraulic separation, the finest sediments having lower heavy mineral contents (Fig. 43). The phyllite content is consistenly high, without exception greater than in the overlying till. Of the 20 samples analysed, 14 had between 3% and 6% which corresponds to the lower till values at Mårhalli.

8 samples from the Mår subtill sediments were examined for pollen and spores (Table 7). The 'total' counts represent the pollen content of one slide and thus gives an approximation to the absolute pollen frequency in a sample, since the samples were of roughly the same size. Samples 73f–72 and × were taken from clayey material collected c. 6 and 4 metres below the surface, respectively. In the varia group, sample 15–72 had 1 *Plantago* cf. *maritima*, 1 *Campanula*, 1 cf. Rosaceae and 3 *Polygonum* sect. *Bistorta;* sample 73b–72 had 2 Rosaceae, 1 *Salix* and 1 cf. *Juniperus*. Samples 73f–72 and × revealed 3 and 2 specimens, respectively, of *Pediastrum*.

Discussion. With regard to the relative ages of the various stratigraphical units, both their high phyllite content and the palynomorphs present indicate a connection between the Mår subtill sediments and the lower till at Mårhalli. The question is which unit was derived from the other. One could postulate that the Mår subtill sediment was oldest and that ice flowing eastwards had transported the material across to the eastern side of Mår and redeposid it as lower till there. However, it is very unlikely that the till could become so well homogenised in such a short distance; one would rather expect inclusions of sorted material. On the other hand such inclusions are found in the upper till at Mårhalli (Fig. 20). The phyllite contents of these inclusions correspond to the Mår subtill sediments. It seems clear that the subtill sediments must be older than the upper till at Mårhalli (which lithologically corresponds to the upper till at Grytekilviki) and younger than the lower till at Mårhalli. Thus the oldest unit is the lower till at Mårhalli (hereafter informally called Mår lower till), followed by the Mår subtill sediments and with the Mår upper till being youngest.

Turning to the correlations with ice flow phases, the fabric analysis in the Mår upper till and the deformation structures within the subtill sediments show that this till was originally deposited by eastward-flowing ice. This interpretation is further supported by the sediment lenses from the subtill sediments which were discovered at Mårhalli and also indicate eastward transport. The source material of the Mår lower till has been discussed earlier (pp. 53 ff.). and revealed a similar transport direction. Thus, there have been two phases of ice movement towards the east, separated by an ice-free period during which the Mår subtill sediments were deposited. These phases must be Phase I (Mår lower till) and Phase II (Mår upper till).

Palynologically the Mår lower till and Mår subtill sediments display major similarities, as mentioned above. All the spectra are dominated by herbaceous pollen with Gramineae commonest. *Betula* dominates the AP counts. The lithological conditions show that a significant proportion of the subtill sediments are derived from material corresponding to the Mår lower till. It is reasonablet to believe, therefore, that the same is true for the palynomorphs. This is supported by the fact that the highest absolute frequency of palynomorphs is found in samples rich in medium silt, i.e. 14–, 15– and 73b–72. Donner & Gardemeister (1971) have shown that this grain size is equivalent to the sedimentation rate of pollen grains.

The palynomorphs referred to in Table 7 mostly represent, therefore, an ice-free period prior to the deposition of the Mår lower till. There may possibly have been some additional supply of Gramineae pollen when the Mår subtill sediment was accumulating since the latter (except for sample ×) has a higher content of these than the Mår lower till.

By comparison with the established stratigraphy of Førnes/Hovden, the pollen spectra from Mår possess an AP–NAP ratio which most resembles the spectra from the Førnes thermomer. However, one must take into account the more than 300 metre difference in elevation separating the two areas. Furthermore the Mår spectra have more types of trees represented, as well as a greater frequency of *Pinus,* than the Førnes thermomer spectra. Considering these facts a correlation with the spectra from the Hovden thermomer may be more probable. Such a correlation is also supported by the fact that both the Hovden and Mår lower tills appear to represent the very first Weichselian deposits. This implies that the Hovden kryomer corresponds to the Phase I ice movements. Thus the most likely correlation for the Mår lower till is also with the Hovden kryomer, while the Mår subtill sediments would then have been deposited during the Førnes thermomer and the Mår upper till during the Førnes kryomer.

The stratigraphy at Mår may be summarised in the following events:

a) During an eastward glacial advance across Hardangervidda (Phase I, Hovden kryomer), palynomorphs from the preceding ice-free period (Hovden thermomed) were incorporated in the till.
b) Later on, the Mår area was deglaciated during the Førnes thermomer, and the Mår subtill sediments were deposited. The general textural development of these from coarse to fine material may mean that the basal beds were deposited in connection with the deglaciation. The finer silty sediments higher up the sequence are probably lacustrine, having settled in the then-existing Mårvatn. The top of the sequence lies some 15–20 m above the present lake's outlet, which in the complete absence of evidence to suggest that the sediments were moved en masse, implies a lowering of Mårvatn's threshold during later glaciation. The palynomorphs in the Mår subtill sediments are essentially derived from the preceding ice-free period, the Hovden thermomer, via the Mår lower till. If there was any vegetation at all when the Mår subtill sediments were being deposited, it must have a sparse herbaceous community dominated by Gramineae. *Pediastrum* may have been living in the water then.
c) After the Mår subtill sediments had been deposited there was a new glaciation of the area, once again with ice flowing eastwards (Phase II).
d) Younger ice movements have later crossed the area.

The Gøyst area

The river Gøystdalvassåi, draining Øvre Prestetjern (Fig. 47), follows an up to 15 m-deep channel in superficial deposits. The sediments adjacent to the channel have a relatively even surface with flutings directed towards 105°–110°, which corresponds with Phase II ice movements as deduced from scouring observations.

The sections in the channel's west side, A, B and C in Fig. 48, were cleaned up. These revealed that below a 3 m-thick till bed, the Gøyst till, there is a sequence of more than 11 m of sorted sediments, the Gøyst sand.

The Gøyst till is sandy with layered structures (Fig. 22), which were discussed earlier (p. 44). A fabric analysis of stones with 3 to 10 cm-long major axes showed a primary maximum between 100° and 110° and a secondary one between 170° and 180° (Fig. 49). This confirms that the till was deposited by the Phase II ice movement.

The Gøyst sand consists chiefly of medium to coarse sand with a clear tendency for the basal beds to contain more silt and the upper beds more gravel. There are several structures present. The top, gravelly sand bed displays disturbed bedding which must be due to glacial tectonisation. Below that are

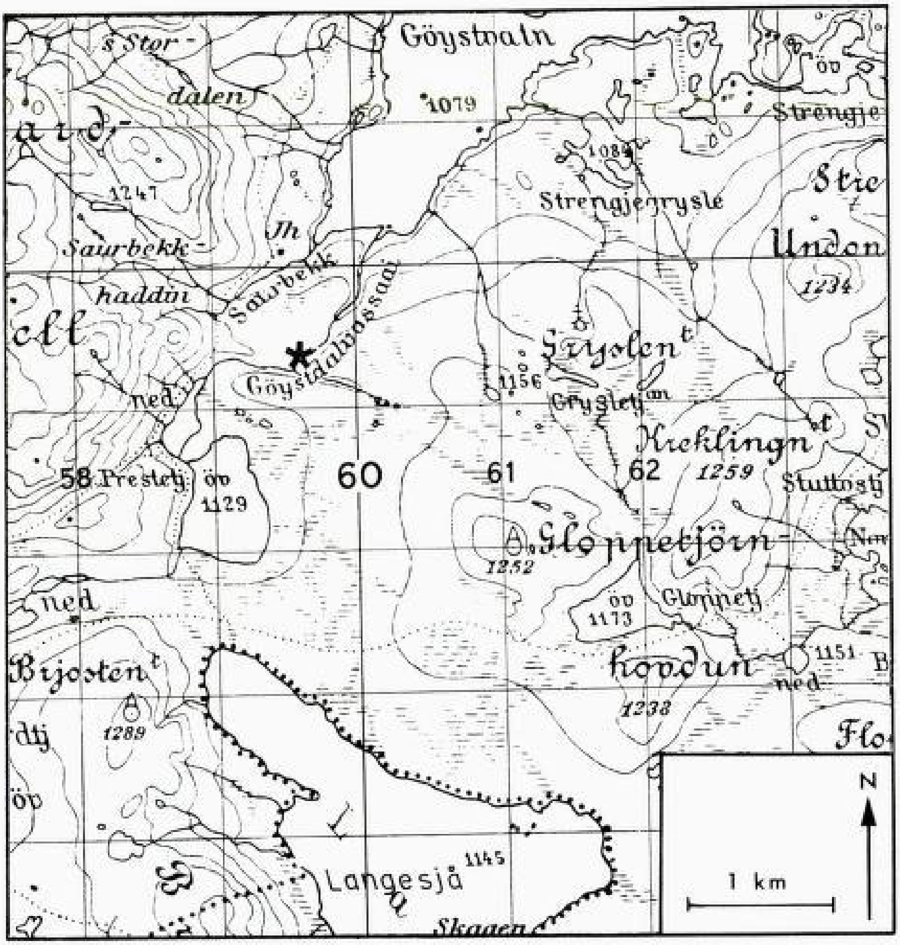


Fig. 47. Map showing locality of subtill sediments at Gøystdalvassåi (asterisk) and surroundings.

horizontal beds and laminae of sand and gravel. Section B has horizontal beds and laminae, load structures, scour and fill and climbing ripple lamination (see Figs. 48 and 50).

The uppermost structures in section B, described as load structures, have a flame-like appearance highlighted by iron precipitation. The overlying coarse sand seems to have been pressed down into fine sand beds. The lower load structures (Fig. 50) are considerably more irregular, but also here a coarse sand layer has been pressed down into underlying fine sand. Certain parts of the coarse sand are isolated, producing ball and pillow structures. The scour and fill structures have asymmetrical troughs with the steepest side towards the northeast. The troughs are about 15 cm across. The climbing ripples are Reineck & Singh's (1976, p. 96) 'type 1 ripple laminae in drift'. Their textural composition, coarse silt/fine sand, is somewhat finer than the neighbouring

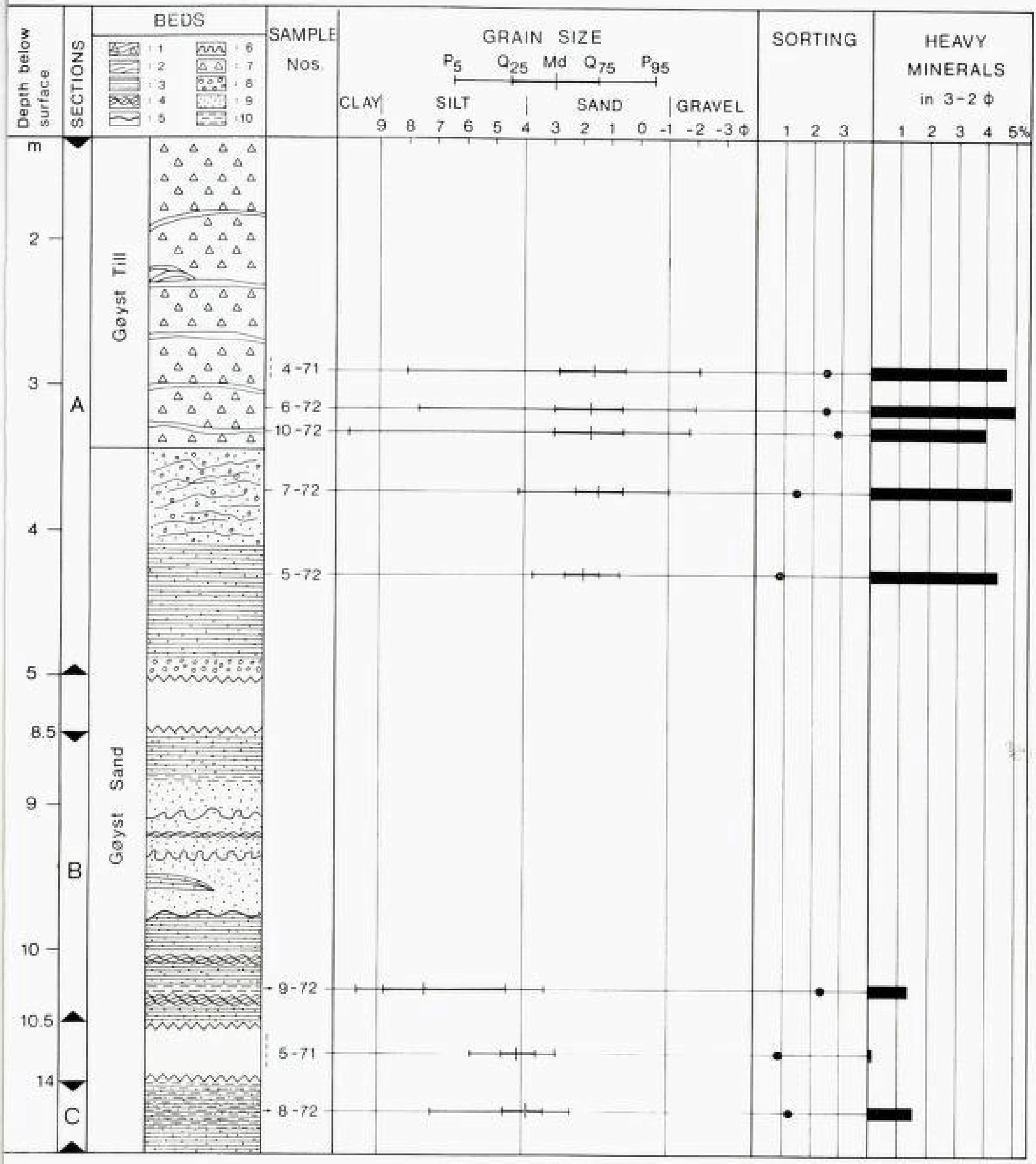


Fig. 48. Stratigraphy, structures and lithology of the sediments at Gøystdalvassâi. Structures; 1: Layering in till, 2: Deformed bedding planes, 3: Horizontal laminae, 4: Climbing ripples, 5: Scour and fill, 6: Load structures. Textures; 7: Unsorted material, till, 8: Gravel, 9: Sand, 10: Silt.

sediments. The ripple wavelength is c. 10 cm and the crests are straight and aligned NW–SE.

In section C the bedding is dominated by wavy parallel laminae.

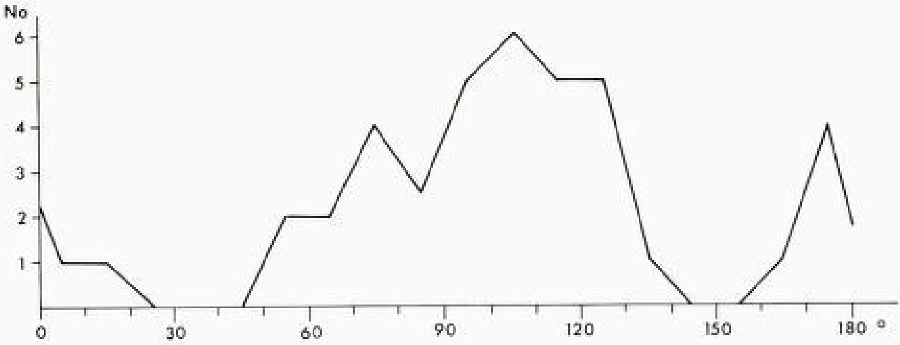


Fig. 49. Fabric analysis of 40 cobbles in Gøyst till. The values are grouped and represented in 10°-intervals.

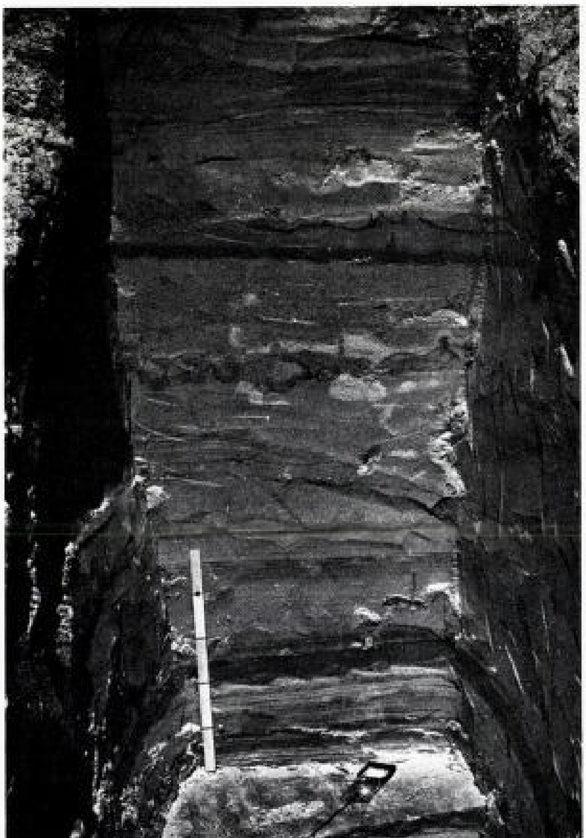


Fig. 50. Photograph showing part of Gøyst sand, equivalent to section B in Fig. 48 (see that figure for description of structures). The compass is lying on straight crested ripples trending NW-SE.

None of the samples from either the Gøyst till or sand contain phyllite. The heavy mineral content in the till is 4% to 5%. In the subtill sediments the heavy mineral content in the 3 to 2 phi fraction also shows a reduction as the sediment becomes more fine-grained, comparable with the Mår subtill sediments. The relatively high heavy mineral content in sample 9–72, despite the low median value, is due to the bimodal grain-size distribution with maxima in the 9 to 8 phi and 5 to 4 phi fractions, representing the fine and coarse laminae, respectively.

Samples 8–, 9–, and 10–72 were examined for pollen and spores, without success.

Discussion. The depositional environment of the Gøyst sand can to a large extent be deduced from its texture and structure.

The presence of climbing ripples and scour and fill structures points to a fluvial environment. The type of climbing ripples found here suggests an abundant supply of material both in suspension and as bed load. The relation between grain-size and structures shows an alternation of rather quietly flowing water (straight-crested climbing ripples) and high velocity flow (plane beds) (Allen 1970, p. 79; Reineck & Singh 1973, p. 11). Together, this implies deposition in braided streams. The transport direction as deduced from the ripples was towards the S and SW. The steeper side of the scour and fill structures, towards the NE, supports this, as it should normally face downstream according to Reineck & Singh (1973, p. 62).

The question of when the Gøyst sand was deposited is probably answered by the fabric of the Gøyst till and flutings on the surface which have been correlated with Phase II. It is then tempting to attribute the sand to the same ice-free period as the Mår subtill sediments, i.e. the Førnes thermomer.

Since the area's present-day drainage is northeastwards (Fig. 47), the indicated flow in the opposite direction for the Gøyst sand requires explanation. The outlet of Gøystvatn, which Gøystdalsvassåi flows into, lies at 1079 m a.s.l., while the watershed to the south to Langesjå is at 1150 m a.s.l., i.e. an elevation difference of c. 70 metres. Even if the watershed, consisting of moraine, may have been raised by subsequent accumulation, the uppermost layer of the Gøyst sand lies at c. 1120 m a.s.l. and implies that the main reason for the southerly draining must be a change in Gøystvatn's outlet level. It may reasonably be assumed that it was lowered by glacial erosion during Phase II and later glaciations. An alternative possibility is that the outlet was blocked by ice occupying the Gøyst basin. The coarsening upwards sequence within the Gøyst sand may be due to the approach of an advancing ice front. If so, the sand must then belong to the close of the ice-free period preceding Phase II, as opposed to the Mår subtill sediments which are from an earlier period.

Sterra

A tributary to the Sterra river from the east, the Starraslekja, has eroded a channel up to 15 m deep in superficial deposits. Originally the incision may have occurred through glaciofluvial erosion. In the erosion bluff 1200 m a.s.l. there are two distinctly different tills. The exposed thickness of the lower one (Sterra lower till) is 1.5 m, whereas the other one (Sterra upper till) is c. 5 m thick here. Both are well consolidated and the surface bears flutings. Clearly these are two basal tills.

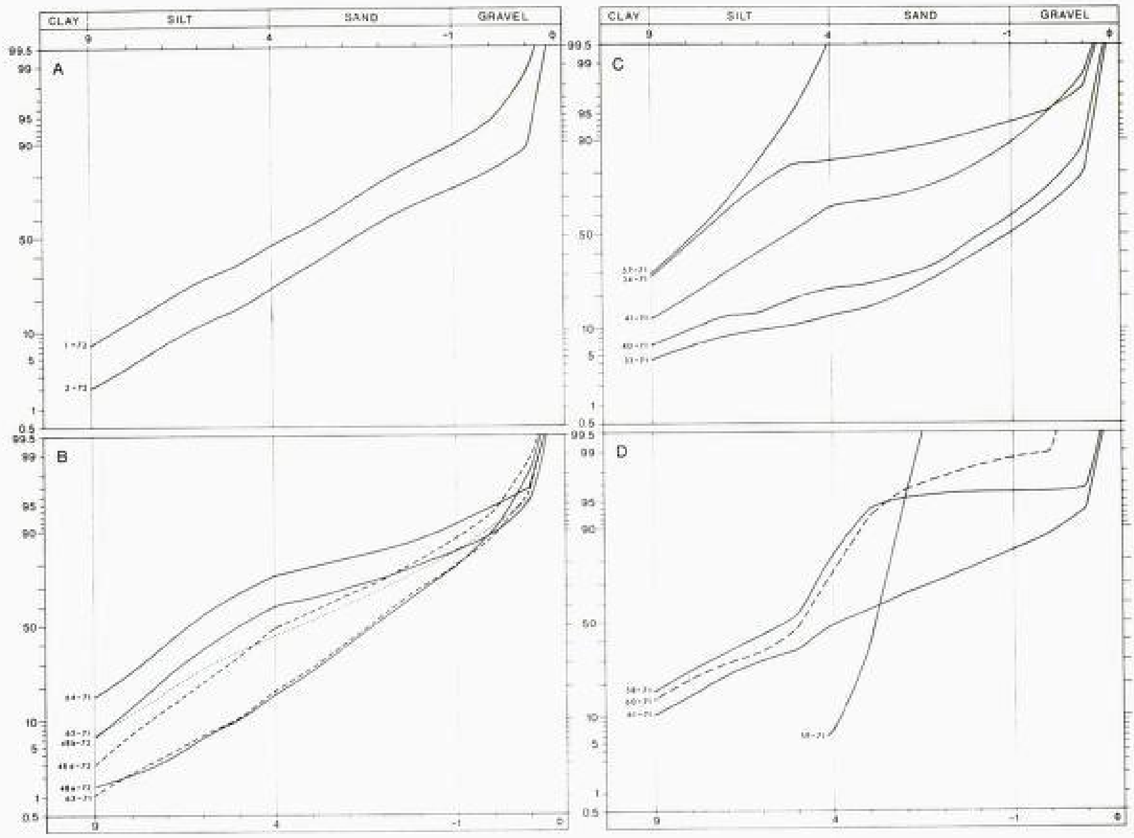


Fig. 51. Cumulative grain-size distribution curves drawn on probability percent ordinate. A: Sterra upper till (1–72) and Sterra lower till (2–72). B: Samples from Holsbu. Genetic interpretation are given in the text. C and D: Samples from the Hansbu area. For description of type and localities, see text and Fig. 52.

Their lithological differences are summarised below:

	Lower till	Upper till
Colour	grey 5Y5/1	yellowish grey 5Y7/2
No–% blue quartz	25.1	5.0
» phyllite	7.8	0.5
» phyllosilicates	1.1	5.0
Weight–% heavy minerals	2.1	3.4
Clay ($<$ 9 phi)	7.4	2.1
Silt (9 to 4 phi)	38.6	23.1
Sand (4 to –1 phi)	43.4	48.8
Gravel (–1 to –4 phi)	10.6	26.0

The colours referred to are for dried material. When wet, the lower till is blue-black. The blue quartz contents cited are the average number per cent of the three phi-fractions, 0 to –3 phi. Phyllite, mica and heavy minerals refer to the 3 to 2 phi fraction. Fig. 51 shows the grain-size distribution curves.

The tills have also been examined palynologically. No palynomorphs were found in the upper one but there were plenty in the lower one. The 383 specimens identified were assigned as shown below. The percentages AP and NAP are based on the sum AP + NAP, while the spore percentages are calculated from sum AP + NAP + spores.

Alnus	4.2	Caryophyllaceae	0.7
Betula	21.2	Comp. lig.	0.7
Corylus	1.3	Comp. tub.	7.2
Betula/Corylus	5.2	Cyperaceae	6.8
Picea	3.6	Ericales	10.4
Pinus	5.5	Graminea	30.9
		Polygonum sect. Bistorta	1.0
$\sum$ AP	41.–	Ranunculus	0.3
		cf. Rosaceae	0.7
Filices	11.7	cf. Valeriana	0.3
Lycopodium sp.	2.9		
Lycopodium selago	0.3	$\sum$ NAP	59.–
Sphagnum	5.0		
$\sum$ Spores	19.9		

Discussion. The fluting direction on the surface of the upper till has alread been attributed to Phase II ice flow. The upper till is therefore probaby of the same age, so the underlying one must date from either an earlier subphase of Phase II or a quite separate, earlier glaciation phase.

The petrographical composition of the lower till with its high content of Cambro–Ordovician materiel, i.e. phyllite and blue quartz, shows that this till come from the west too. As mentioned earlier (p. 53) the phyllite content is anomalously high, seen in relation to the distance from phyllitic bedrock. The most likely explanation for this is transport as thrust slices (p. 57), which is certainly supported by the high contents of palynomorphs and clay/silt, indicating that much of the material is derived from lacustrine deposits.

Compared with present conditions, the palynomorphs indicate a favourable climate, taking into account the elevation. The relatively high content of Ericales and *Spagnum* witnesses a well-established plant community with advanced paludification and soil profile development. There are basic similarities with the pollen spectra from the Hovden thermomer (Vorren & Roaldset 1977) with fairly high *Picea* and spore contents.

It seems reasonable to conclude that the palynomorphs in the Sterra lower till are derived from deposits of the Hovden thermomer. The Sterra lower till is thus Hovden kryomer age, correlated with Phase I while the Sterra upper till dates from Phase II.

The Hansbu area

There are three localities within the Hansbu area (Fig. 3, loc. 10, Fig. 52) which will be discussed in detail.

Locality a, (Fig. 52) This lies in a stream ravine. There is a c. 3 m-thick sequence of laminated silty–clay sediments (Fig. 51, sample 57–71) which are faulted and folded. These are overlain by boulders and cobbles but there is no proper till cover. Beneath the sediments there is a boulder-rich diamicton of varying textural composition (Fig. 51), samples 40– and

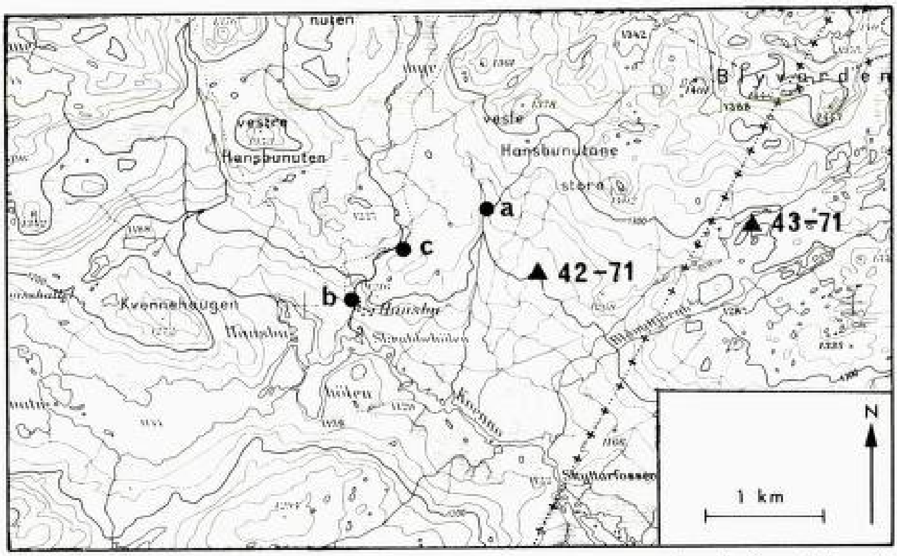


Fig. 52. Map showing localities a, b, c of described sections with subtill and intertill sediments. The sites of the till samples 42– and 43–71 are also shown.

41–71). The grain-size distributions reveal that the diamicton consists of two populations separated at 4 phi. Since the laminated sediments consist of silt/cay, the diamicton is probably a mixture of these sediments with coarser ones. Thus the whole sediment sequence at this locality has been glacio-tectonically moved into its present position.

The petrographical composition of the diamicton's 3 to 2 phi fraction is, for samples 40— and 41—71, respectively:

heavy minerals	10.4% and 7.5%
phyllite	3.7% and 7.4%
phyllosilicates	26.9% and 12.6%

Seen in the regional context they are poor in phyllite (Fig. 25) and rich in heavy minerals (Fig. 30) and phyllosilicates.

Samples 41– and 57–71 were examined for palynomorphs. In sample 41–71, 1 *Alnus*, 1 *Betula*, 3 *Pinus*, 5 *Dryopteris* type and 1 *Lycopodium* were found; in sample 57–71, 1 *Betula* and 6 *Dryopteris* type.

Locality b, Fig. 52. A section c. 1 m deep at Hansbu revealed a wellconsolidated diamicton in the lower c. 0.4 metres overlain by strongly deformed stratified sediments. The latter's grain-size distributions (Fig. 51) indicate beds of pure sand (sample 59–71) and of laminated silt/clay sediments (samples 58– and 60–71). These last two have trimodal distributions with the primary mode in the 5 to 4 phi fraction, a secondary one in the clay fraction and a minor one in the gravel fraction. The first two reflect coarser and finer laminae, respectively, while the last one probably represents secondary incorporation. The diamicton below, sample 61–71, also possesses a primary mode in the 5 to 4 phi fraction and a secondary one in the clay fraction. The deposit is therefore most likely a basal till which in part contains material derived from the overlying sediment type. The entire sequence has thus been glacially transported and deformed.

The heavy mineral, phyllite and phyllosilicate contents of the 3 to 2 phi fraction in the respective samples are as follows:

Sample No.	HM	P	M
58–71	2.4	13.0	not analysed
59–71	3.4	20.3	2.2
60–71	3.0	13.5	not analysed
61–71	5.5	8.5	2.5

Samples 61– and 60–71 were examined for palynomorphs. In sample 61–71, 1 *Betula*, 2 Gramineae, 2 *Dryopteris* type and 4 *Lycopodium* were found; in 60–71 1 *Alnus*, 4 *Betula*, 1 *Pinus*, and 4 *Dryopteris* type.

Locality c, Fig. 52. 1 km north of Hansbu there is a gravel-rich till (Fig. 51, sample 33–71) with inclusions of clay sediments, sample 34–71. The petrographical composition of the 3 to 2 phi fraction in samples 33– and 34–71 is, respectively:

heavy minerals	1.5% and 6.5%
phyllite	14.3% and 8.2%
phyllosilicates	12.5% – (not analysed)

An analysis of the palynomorphs in 34–71 howed 7 *Betula*, 1 *Pinus*, 5 Gramineae, 1 Caryophyllaceae, 1 Ericales and 20 *Dryopteris* type.

Discussion. The sediments from localities a and c are rather similar petrographically, with high heavy mineral and phyllosilicates contents but relatively little phyllite. The same relationship is shared by two other till samples in the area, 42– and 43–71, Fig. 52. All these samples are located along a zone stretching E–W, and it seems reasonable to interpret this in terms of the tills and sorted sediments at localities a and c having been transported by an eastward-flowing ice stream. This means transport by Phase II ice, or possibly Phase I.

The sediments at locality b do not have the same high contents of heavy minerals and mica, but they are relatively poor in phyllite, especially sample 61–71, which suggests that also these experienced an eastward transport.

The palynomorph assemblages in the different samples show significant similarities and indicate a correlation with the Hovden thermomer. Nevertheless the absolute frequencies are extremly small, suggesting resedimentation, in which case the sediments must date from a younger ice-free period. Since the sediments were transported from the west, this ice-free period must have preceded Phase II, so a correlation with the Førnes thermomer is more likely. The grain-size analyses imply a lacustrine origin for the sediments, which will be referred to from now on as the Hansbu silt.

Holsbu

Holsbu, Fig. 3, loc. 11, lies by the northeastern shore of Nordmannslågen (1244 m a.s.l.). The ground here is characterised by earth hummocks. A section c. 1 m deep showed:

a) At the base, weathered light granite bedrock.
b) 0–20 cm, brown gravelly sand bed (Fig. 51, samples 63–71 and 48a–72.
c) 0–5 cm silty clayey sediment (samples 64–71 and 48b–72) with gravel and sand horizons.
d) Upper part, a diamicton (samples 65–71 and 48d–72).

Petrographically there are only relatively small differences between the sediments. In the 3 to 2 phi fraction they contain from 4.7% to 9.4% phyllite, the gravelly sand bed having least. They heavy mineral content varies between 2.0% and 4.4%, being highest in the gravelly sand bed. All the sediments are characterised by high contents of biotite, from 5.1% to 16.1%. The majority of biotite grains show signs of weathering.

Sample 48b–72 was examined palynologically by D. Moe. 101 palynomorphs were counted and found to be as follows: 1 *Betula*, 1 *Pinus*, 1 Umbelliferae, 6 Gramineae, 41 *Dryopteris* type, 5 *Dryopteris linneaneae*, 16 *Lycopodium clavatum* and/or *L. alpinum*, 16

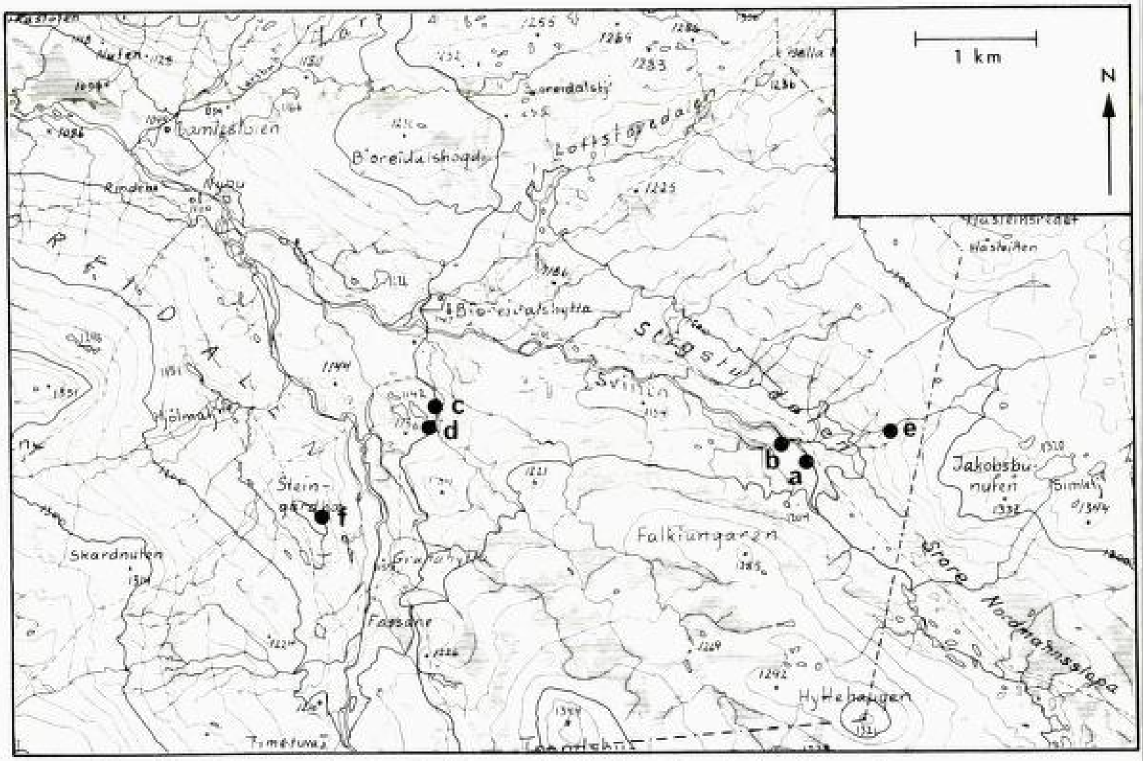


Fig. 53. Map of the Svinto area showing localities of described stratigraphical sections, a–f.

L. selago, 5 cf. *Lycopodium* sp., 4 *Sphagnum* and 5 unidentified specimens. A sample taken from 64–71 and counted by the writer revealed almost the same proportions though with more *Betula* and *Pinus*, 6 and 5 grains respectively out of a count of 102. The silty clayey sediments are thus dominated by spores, in particular *Dryopteris* and *Lycopodium selago*, together with *L. clavatum* and/or *L. alpinum*

The individual deposits are far from easy to interpret genetically. The underlying gravelly sand bed might be either a till or a poorly sorted glaciofluvial deposit. The grain-size distribution curve of the silty clay sediment shows a marked break at 4 phi, indicating that two populations have become mixed together. This could be a glaciolacustrine sediment with icedrop material, but it may be more likely that the silty clay portion was originally lacustrine and later on had coarser material introduced through glacio-tectonic processes. The overlying diamicton seems to be rather heterogeneous which probably is due to picking up material from formations below, partly by glacio-tectonic movements and partly by cryoturbation.

A possible explanation of the stratigraphy may be proposed. During an ice-free period the bedrock and the gravelly sand bed were weathered. At the same time the silty clay sediments were accumulating in a nearby lake. During a subsequent glaciation they were thrust up onto the gravelly sand bed, some material from which become incorporated within the silty clay. In addition some till was deposited on top of these beds — the upper diamicton, which itself had

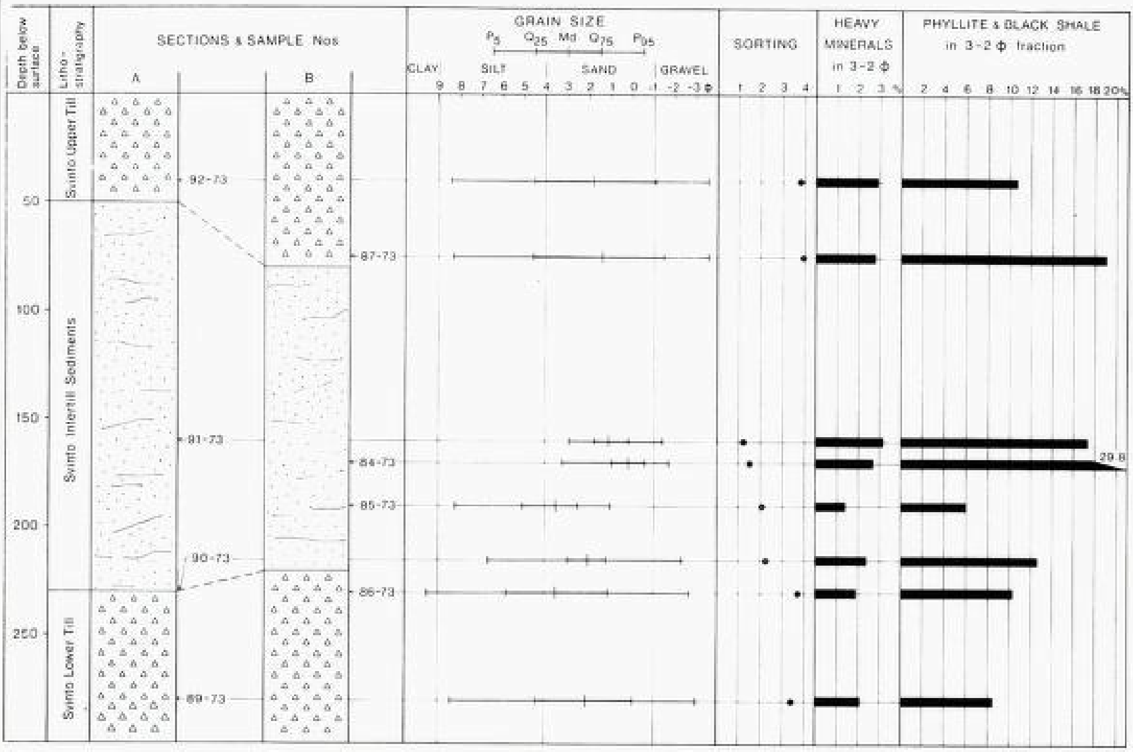


Fig. 54. Stratigraphy and lithology of two sections at the river Svinto, a and b on Fig. 53.

incorporated some of the underlying sediments. After the last glaciation the upper parts of the sequence have suffered cryoturbation.

The high spore content and the presence of *Pinus* pollen suggest that the palynomorphs in the silty clay (Holsbu silt) were derived from the Hovden thermomer. To what extent they are reworked from other sediments is difficult to determine. The Holsbu silt could therefore date from either the Hovden or the Førnes thermomer.

The Svinto area

Six localities (Fig. 53 a-f) will be described from the area around the Svinto (Fig. 3, loc. 12), a tributary to the river Bjoreio. The first five localities, a-e, have a number of features in common, so after individual description they will be discussed together. Locality f will be described and evaluated separately.

Localities a and b, Fig. 53. At c. 1180 m a.s.l. by the river Svinto, two till sheets have been found, the Svinto upper and lower tills, separated by the Svinto intertill sediments. Two sections on the west side of the river were examined (Fig. 54).

The Svinto lower till has an exposed thickness of at least 1.5 m, is well consolidated and dark grey. A thin-section of an oriented sample retrieved from section a, 3.1 m below the surface, revealed a marked grain orientation towards 305–315° (Fig. 55). Grain-size analysis showed a relatively large clay fraction (Figs. 27 and 54) and a comparatively low heavy mineral content.

The Svinto intertill sediments vary in thickness, being 1.8 and 1.4 m in the two sections.

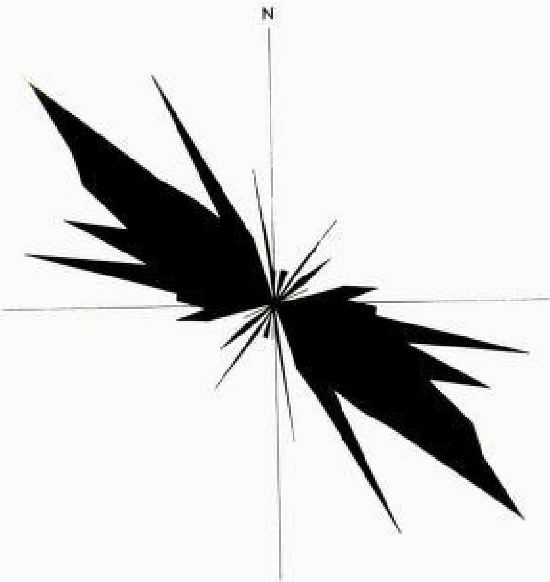


Fig. 55. Fabric diagram of Svinto lower till. The diagram represents 106 measurements on grains in thin-sections.

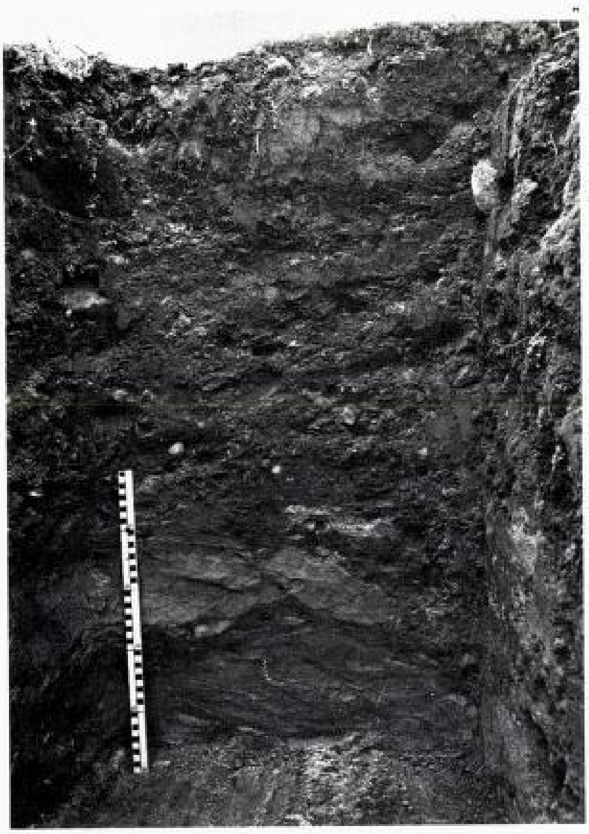


Fig. 56. Svinto upper till overlying strongly disturbed Svinto intertill sediments. The scale is 0.5 m long.

They are considerably disturbed, but primary bedding can be seen in section a, along with centimetre-thick veins of dark clayey silt which transect this bedding. Texturally the sediments consist mostly of sand (Fig. 54). The phyllite content varies markedly, from 6% to

29.1% in the 3to 2 phi fraction. Once again the heavy mineral content tends to be highest in the coarsest sediments on account of hydraulic separation. Sample 85–73 was examined with a view to counting the palynomorphs, but none was found.

The Svinto upper till (Fig. 56) has an observed thickness of less than 0.5 m up to 0.8 m. It is dark grey and normally consolidated. The grain-size distribution (Figs. 27 and 54) is platykurtic — Kg 0.91 and 0.88 — due to a more even distribution throughout the individual fractions than in the lower till which is mesokurtic — Kg 1.06 and 1.05. The heavy mineral content is about 1% higher than in the lower till. The phyllite content varies somewhat (Fig. 27), but for both samples there is a marked increase in the finer gravel fractions (p. 50).

Localities c and d, Fig. 53. At locality c there are two till beds with a c. 1 m-thick intervening sequence of strongly disturbed sorted sediments. Two till beds can also be discerned at locality d. Grain-size analyses are presented in Fig. 57. The lower till samples 33– and 35–70 are clay-rich and mesokurtic. The upper ones, samples 34– and 39–70, are clay-deficient and platykurtic. The intertill sediments at locality c have a varying grain-size composition, being partly silty (e.g. sample 37–70) and partly sandy/gravelly (e.g. sample 38—70).

The contents of heavy minerals (HM), phyllite (P) and phyllosilicates (M) in the 3 to 2 phi fraction of the different samples are as follows:

	HM	P	M	
39–70	2.7	21.3	0.8	upper till
34–70	2.3	7.4	2.0	
38–70	1.1	14.7	2.1	intertill sediments
37–70	3.1	27.8	3.3	
35–70	2.3	11.0	0.3	lower till
33–70	2.3	8.5	1.1	

As can be seen, the petrographical composition is more homogeneous in the lower till than in the upper one, especially with respect to phyllite.

Locality e, Fig. 53. At locality e which is situated in a stream ravine, there is an upper, darker till separated from a lower, lighter one by a sharp boundary. The lower till has irregular inclusions of light sand lenses. Fig. 57, sample 95–73, shows this till's grain-size distribution wich is characterised by low clay and high silt contents. Sample 96–73 was taken from one of the sand lenses.

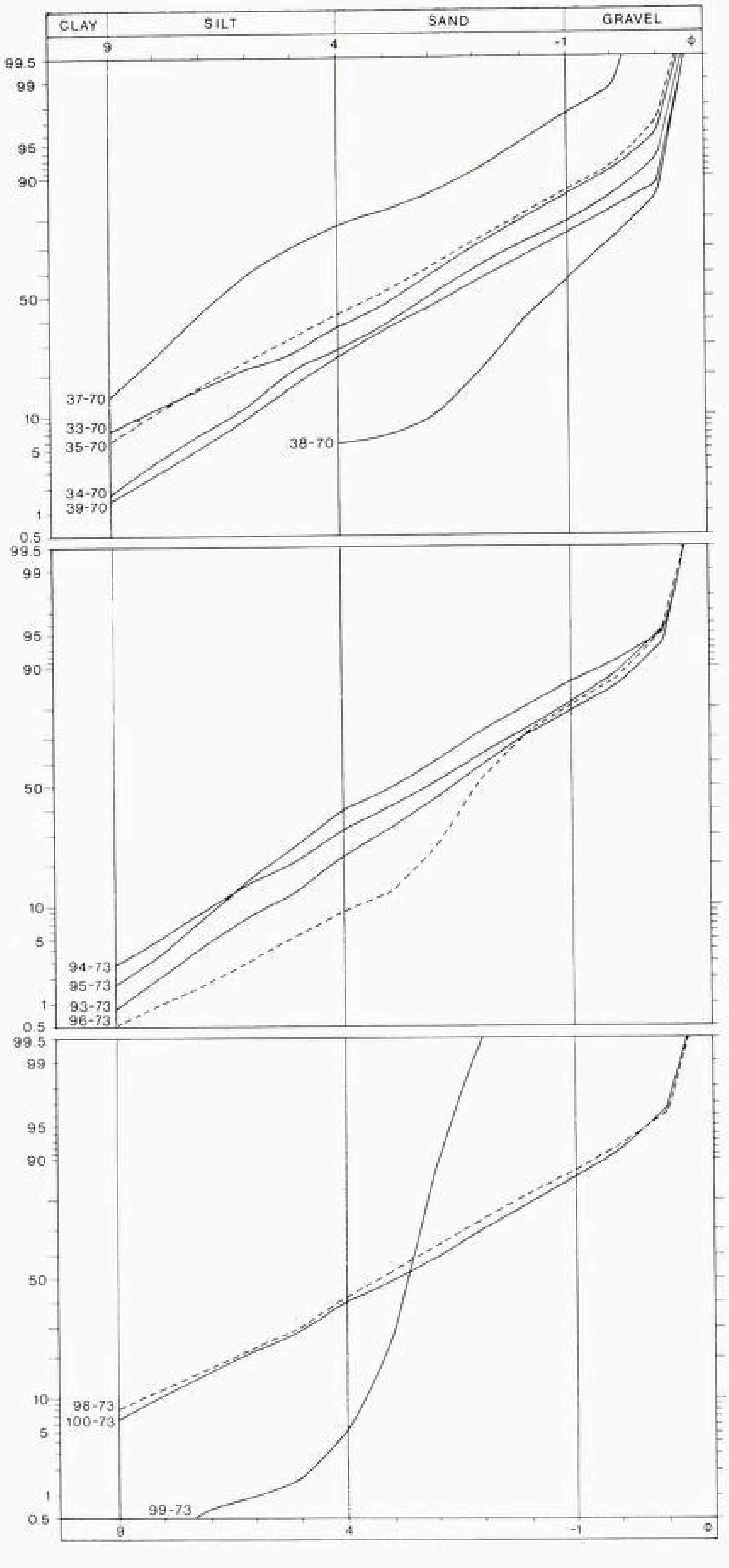
The upper till is 1.5 m thick with an increasing sand content towards the top. In Fig. 57 the grain-size distributions are presented from samples recovered 0.1 m (sample 94–73) and 0.8 m (sample 93–73) above the bonudary with the lower till.

The heavy mineral (HM), phyllite (P) and phyllosilicate (M) contents in the 3–2 phi fractions were found to be:

No.	HM	P	M	
93–73	2.9	2.4	2.1	upper till
94–73	2.3	7.9	1.2	
95–73	1.9	6.8	0.9	lower till
96–73	2.8	8.7	2.6	sand lense

A relatively large petrographical variation in the upper till is seen here too.

Comparison and interpretation of localities a-e. All of these localities reveal two till beds, and there seem to be consistent inter-locality features typical for each till. The lower till has more clay and less gravel than the upper one, though for clay content sample 95–73 provides an exception. The gravel content (–1 to –4 phi) of the lower till is in the range 12.8 to 17.5% while that for the upper till is 18.6 to 29.1%, a significant difference. The mica content also tends to be greater in the upper till. A fairly even phyllite content is seen for



CLAY
SILT
SAND
GRAVEL
9
4
-1
Φ
99.5
99
95
90
50
10
5
1
0.5
37-70
33-70
35-70
34-70
39-70
38-70
94-73
95-73
93-73
96-73
98-73
100-73
99-73

the lower till, between 6.8% and 11.0% while the upper one shows very large variations, from 2.4% to 21.3%. The heavy mineral content is 1.9 to 2.3% and 2.3 to 2.9%, respectively. These lithological features, viewed collectively, indicate that the two till beds may be correlated at all five sites and that the informal stratigraphical designations proposed for sites a and b may be extended to c, d and e.

With regard to correlation with specific ice movement phases, a fabric analysis (Fig. 55) of the Svinto lower till at locality a revealed a dominant orientation corresponding to Phase III. The Svinto upper till, being the stratigraphically youngest till in the area, is thus allocated to Phase IV.

One is left with the question of the genesis of the Svinto intertill sediments. They were clearly deposited during or after Phase III, but prior to Phase IV. Texturally they are of varying composition, the most likely explanation being that they were laid down subglacially. The moraine topography between Halnefjorden and Svinto is interesting in connection with this, being irregular with E-W aligned ridges. These ridges are composed of basal till and crossed by eskers from the final deglaciation. The ridges themselves are probably erosion remnants left by Phase III subglacial drainage. The Svinto intertill sediments are presumedly glaciofluvial deposits resulting from that subglacial drainage. This interpretation implies a temperature at the pressure melting point in the basal parts of the ice sheet at that time.

To summarise, localities a–e reveal a lower, relatively clay-rich till in the Svinto area, the Svinto lower till. It is overlain locally by sorted sediments, of varying petrographical and textural composition, interpreted as subglacial deposits. On top is a gravelly till dating from Phase IV, the Svinto upper till. An attempt was made earlier to explain the high gravel content (p. 50), while the phyllite content variations in the 3 to 2 phi fractions can probably be accounted for in part by the incorporation of varying amount of the petrographically varying intertill sediments.

Locality f, Fig. 53. The following stratigraphy was observed in the uppermost part of a landslide scar at locality f, Steingardhø:

Top – 4 m-thick till.

Middle – 0.1–0.5 m-thick layer of fine sand. The sand layer was seen to extent the length of the exposure, almost 100 metres.

Bottom – till of unknown thickness. However, it is several tens of metres thick and probably make up the bulk of Steingardhø which is 50 to 60 m high.

Grain-size distributions for samples from the upper (100–73) and lower (98–73) tills are shown on Fig. 57. These are very similar and resemble the

Fig. 57. Cumulative grain-size distribution curves drawn on probability percent ordinate. For locality references, see Fig. 53. 33– and 34–70 are basal till samples from loc. d, 35– and 39–70 (basal tills) and 37– and 38–70 (intertill sediments) are from loc. d, 93–, 94– and 95–73 (basal tills) and 96–73 (intertill sediment) are from loc. e, 98– and 100–73 (basal till) and 99–73 (intertill sediment) are from loc. f.

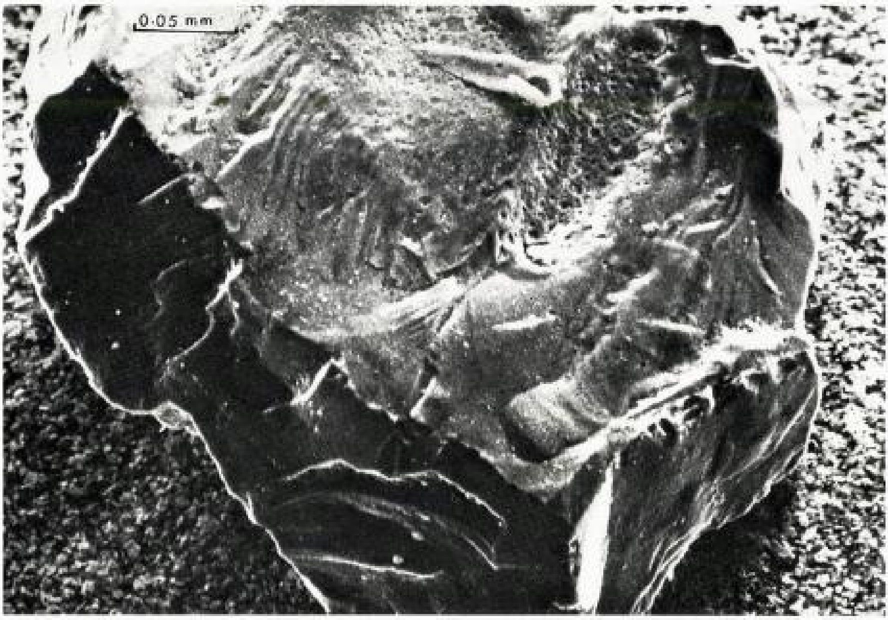


Fig. 58. SEM-photograph showing quartz grain from intertill sediments at Steingardhæ (loc. f on Fig. 53). The grain shows conchoidal fractures of glacial origin and area with etched surfaces.

Svinto lower till. The intertill sand (sample 99–73) has the following grain-size parameters – Mz = 2.76 phi, σ = 0.76, Sk = 0.16 and Kg = 1.29.

The heavy mineral (HM), phyllite (P) and phyllosilicate (M) contents in the 3 to 2 phi fractions of these three samples are:

No.	HM	P	M	
100–73	2.3	9.9	0.8	upper till
99–73	1.3	1.2	9.3	intertill sand
98–73	1.8	6.5	1.9	lower till

The low phyllite content of the fine sand bed is striking. The bed's extent and grain-size distribution can indicate an eolian deposit. Kukal (1971) found that the majority of eolian sands possess the following grain-size distribution characteristics: Md betwen 0.15 and 0.25 mm, sorting (Trask's So) usually less than 1.25 and skewness (Inman) between 0.13 and 0.30. The corresponding values for sample 98–73 are: 0.15 mm, 1.36 and 0.13. Apart from the insignificantly higher sorting value, the parameters fall within the limits for normal eolian sand.

To investigate the origin of this sand more fully, the surface textures of quartz grains were analysed by SEM (Krinsley & Doornkamp, 1973). This was kindly carried out by I. F. Strass, who examined 23 and 21 grains, respectively, in the 1 to 0 and 2 to 1 phi fractions. About half of the grains displayed advanced solution and precipitation features. Apart from this, virtually all the

Table 8. Stratigraphical correlation chart

Climato-stratigraphy	Ice movement phase	LITHOSTRATIGRAPHY Hovden	Førnes	Mår	Gøyst	Steira	Hansbu	Holsbu	Svinto loc. f	Svinto loc. a-e
	IV									Svinto upper till
Førnes kryomer	III	upper till	upper till	upper till						Svinto intertill sediments Svinto lower till
	II	Hovden	Førnes	Mår	Gøyst upper till	Steira upper till			upper till	
Førnes thermomer		Hovden sand	Førnes inter-till sedim.	Mår sub-till sedim.	Gøyst sand		Hansbu silt	Holsbu silt	fine sand	
Hovden kryomer	I	Hovden lower till	Førnes lower till	Mår lower till		Steira lower till			lower till	
Hovden thermomer		Hovden clay						?		

mechanically produced relief forms were essentially glacial (Fig. 58). Possible eolian fractures (upturned plates) were discerned on 3 or 4 grains form each fraction while textures acquired by water abrasion were not found at all. The analysis was not conclusive, therefore, but of the two agencies which could have deposited the fine sand bed, only an eolian one is indicated. An eolian origin therefore seems most likely.

The question immediately arises of where the ice-free period, necessary for the eolian deposition, belongs in the regional stratigraphy. As mentioned, both the upper and lower tills here at Steingardhø lithologically resemble the Phase III till. One may thus postulate an ice-free period during Phase III itself or between Phases II and III. So far, though, there are no other signs of this on Hardangervidda. The writer thinks that it is more reasonable that the sand would have been deposited between Phases I and II, i.e. during the Førnes thermomer. The lithological similarity of the two tills is then readily attributable to the fact that both Phases I and II had roughly the same flow direction and the same bedrock sources for their debris. This also account for the unusually thick accumulations here; the locality lies in a valley which was transverse to these ice movements (cf. Fig. 12) and acted as a sediment trap (cf. Mangerud 1965, Nobles & Weertman 1971).

DISCUSSION

Recapitulation. The above section presented proposed correlations betwen the lithostratigraphy of the different localities and the ice movement phases and climatostratigraphical units. These are summarised in Table 8. In addition to the lithostratigraphical units listed there, there are several localities near Møsvatn (Trengsle, Laksastøl, Falkeriset and perhaps also Bitdal) with sediments possibly from the Førnes thermomer. If the correlations are correct, they imply two periods when most or all of Hardangervidda has been ice-free, namely the Hovden thermomer and Førnes thermomer.

The Hovden thermomer is lithostratigraphically represented by the Hovden clay and possibly also by the Holsbu silt. However, resedimented pollen from the Hovden thermomer is present in the tills of the Hovden kryomer at Hovden, Førnes, Mår and Sterra and in younger subtill sediments at Mår, Hansbu and possibly Holsbu (see above).

The palynomorph assemblage from the Hovden thermomer is variable. The spectra from the Hovden clay and the lower tills at Hovden, Førnes and Sterra indicate that during one period *Picea* flourished, at least in the lower-lying part of Hardangervidda, and *Alnus* may possibly have grown there too. The relatively high *Picea* content in the Sterra lower till sample, which comes from areas to the west, is of interest here. Even if the pollen frequency is perhaps insufficient to justify concluding that *Picea* was growing in the same area where the sediments were originally deposited, it does suggest there were stands nearby. This implies that *Picea* has thrived in the western areas during the Hovden thermomer, in contrast to the Holocene (Moe 1970, Fig. 1).

In a later period within the Hovden thermomer, *Betula* became the dominant tree type, according to the Hovden clay spectra. The resedimented palynomorphs in the Mår lower till and subtill sediments, which are chiefly NAP-pollen, probably came from the youngest part of the Hovden thermomer when the climate deteriorated during the transition to the Hovden kryomer. At times the Hovden thermomer seems to have had a rich growth of brackens, *Lycopodium* and *Sphagnum*, which implies rather humid conditions. The advanced soil profile development indicated by the clay mineralogy in the Hovden clay points to the Hovden thermomer having been of long duration.

Palynomorphs attributable with any degree of certainty to the Førnes thermomer have only been found in the Hovden sand and the Førnes intertill sediments. Some of the herbaceous pollen (Gramineae) in the Mår subtill sediments may possibly belong to the same period. The pollen spectra from the Møsvatn localities have been discussed by Vorren & Roaldset (1971) who concluded that they represented a herbaceous-dominated community with the tree line situated at least 200 metres lower than today's. The sediments from the Førnes thermomer on the Hardangervidda plateau itself imply that it was virtually bare of vegetation. If the interpretation of the fine sand bed at locality f, Svinto, is right, there were favourable conditions for eolian erosion and sedimentation.

Correlations with other Nordic areas. Ljunger (1943, 1945, 1946 and 1949) has reconstructed the ice movement conditions of the Weichselian ice sheet in northern and central Sweden. His conclusions are summarised in Lundqvist (1974). Both authors claim an early glaciation (Ljungner's 'Prime Glaciation', Lundqvist's 'Weichsel I') followed by a period when most of Fennoscandia, except possibly the mountain areas, was deglaciated (Ljungner's 'The Interval', Lundqvist's 'Jämtland Interstadial'). During the early glaciation the ice divide lay near the watershed or perhaps even farther west. Ljungner asserts that during a later phase of this glaciation a northerly situated dome developed east of the watershed, but Lundqvist is sceptical.

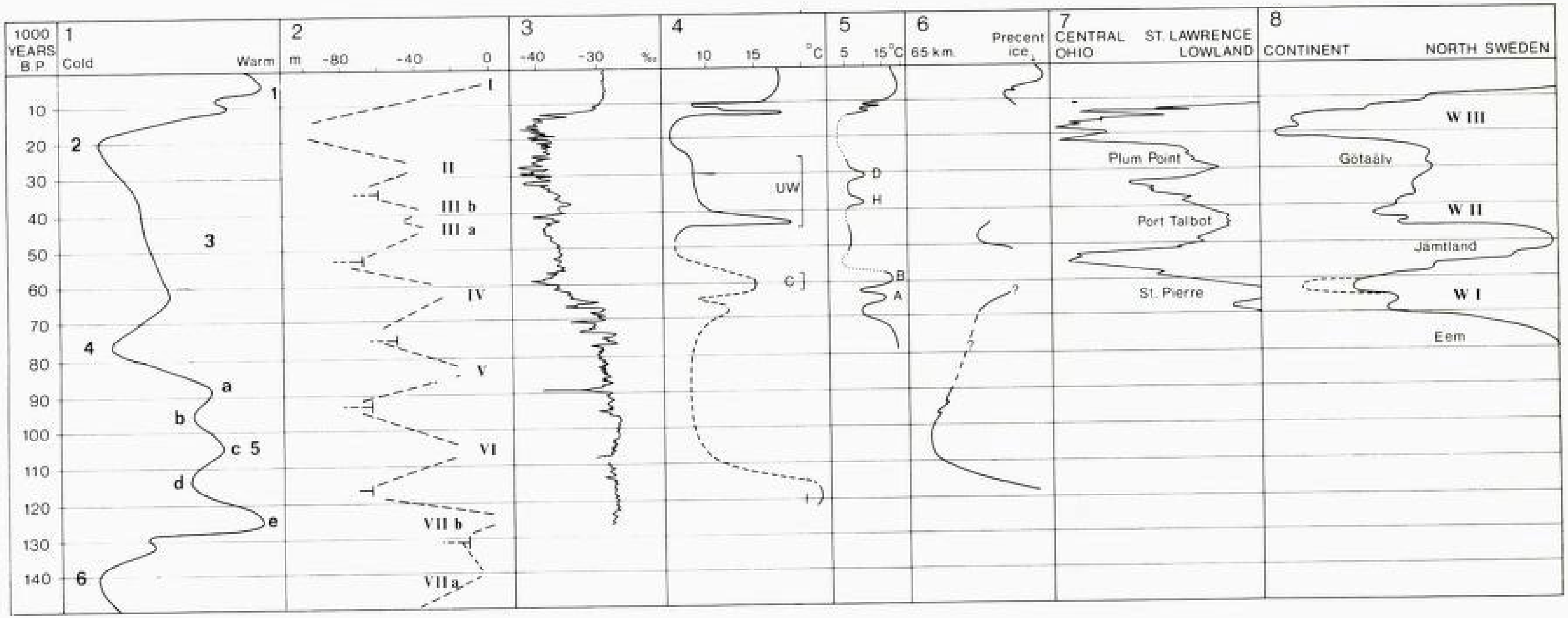
During the opening events of the glaciation which succeeded the ice-free period ('Posterior Glaciation', 'Weichsel II and III'), the ice divide once again lay close to the watershed in the mountain tracts. Later it was displaced eastwards; according to Ljungner right to the Gulf of Bothnia, and as far as eastern Jämtland according to Lundqvist. Finally, the ice divide migrated westwards back to the mountains.

Lundqvist (1969) gives a detailed picture of these events for the Jämtland region. He distinguishes between palaeoscandian (ice divide in the west), mesoscandian (ice divide in the east) and neoscandian (ice divide in the west) ice movement phases. When compared with Ljungner's and Lundqvist's results, the main trends in the ice flow phases on Hardangervidda are similar. Phase I can be correlated with Ljungner's 'Prime Glaciation' and Lundqvist's 'Weichsel I'. Phases II, III and IV reveal a similar sequence, with the ice divide migrating to the east and back again to the west, to that shown by Ljungner's and Lundqvist's 'Posterior Glaciation'/'Weichsel II and III'. Compared with the terminology applied to the Jämtland region (Lundqvist 1969), Phase I corresponds to the older palaeoscandian ice movements. Phase II is correlated with palaeoscandian movements younger than the Jämtland interstadial, Phase III with the mesoscandian and Phase IV with the neoscandian.

The Hovden thermomer, as already discussed, was a relatively long and sometimes relatively warm ice-free period. Palynological comparisons with known Eemian deposits in Denmark (Andersen 1965) and North Sweden (Robertson 1971) led Vorren & Roaldset (1977) to conclude that the Hovden clay belonged to the later part of the Eemian. Indications that *Picea* was growing within or near to the western part of Hardangervidda also support a correlation with the Eemian in that Mangerud (1970) found evidence of *Picea* flourishing in West Norway during this period.

The Hovden kryomer has been correlated with the Phase I ice movement, which in turn corresponds to the 'Prime Glaciation' and 'Weichsel I' of Ljungner and Lundqvist. The Hovden kryomer was therefore the first glaciation period in the Weichselian.

The Førnes thermomer clearly represents an interstadial, and the stratigraphical evidence shows that it existed during the Weichselian. Mangerud (1972) has provided a summary of interstadial and interglacial finds in South Norway, and subsequent discoveries have been made in the Numedal area too (Roaldset 1973b, Rosenqvist 1973). Of particular relevance in this connection are occurrences which reveal that large parts of central Norway must have been deglaciated. This is the case with the fossiliferous (mammoth bones) subtill sediments in Gudbrandsdalen (Mangerud 1965, Bergersen & Garnes 1971, 1972, Heinz 1971, 1974). Four of the finds have been dated to 19,000 ± 1,200 BP (Toten), 24,400 ± 900 BP (Kvam), 22,370 ± 980 BP and 20,000 ± 250 BP



1000 YEARS B.P.
1 Cold Warm
2 m -80 -40 0
3 -40 -30 ‰
4 10 15 °C
5 5 15°C
6 65 km. Precent ice
7 CENTRAL OHIO ST. LAWRENCE LOWLAND
8 CONTINENT NORTH SWEDEN
Plum Point
Port Talbot
St. Pierre
Götaälv
Jämtland
Eem
W III
W II
W I
UW

(Fåvang), 46,000 ± 2,000 BP and 45,000 ± 1,500 BP (Lillehammer) (Heintz 1974).

Relying on the Kvam dating, Bergersen & Garnes (1971, p. 106) believe that the subtill sediments were laid down sometime in the period between 30,000 and 24,000 BP. Lundqvist (1974 has adopted this conclusion, which allows correlation between the Gudbrandsdal interstadial and the Göta älv interstadial. Since ^{14}C-dating of bone material is fraught with problems (Olsson 1974, p. 314) and several of the datings are close to the supposed glaciation maximum of the Weichselian, the proposed chronological placing is uncertain.

If the proposed correlation between Hardangervidda and Jämtland of the ice movement is correct, the Førnes thermomer may be correlated with the Jämtland interstadial. The latter was in turn probably contemporary with the Peräpohjola interstadial in North Finland (Korpela 1969), the Karuküla interstadial in Estland (Serebryanny et al. 1970) and the Brørup interstadial in Denmark (Andersen 1961) – cf. Lundqvist (1974).

The Førnes kryomer corresponds to Phase II, III and IV which can presumedly be correlated with Ljungner's 'Posterior Glaciation' and Lundqvist's 'Weichsel II and III'.

Chronology. Based on different investigations, several curves have been constructed in connection with climatic changes dring the last interglacial and glacial. Some of the more recent attempts are reproduced in Fig. 59. A whole series of palaeotemperature curve has been made, based on deep sea sediment cores, e.g. Rona & Emiliani (1969), Sancetta et al. (1972, 1973), Hays & Peruzza (1972), Shackleton & Opdyke (1973). The oxygen isotope curve shown in Fig. 59, 1 is a generalised curve for the Carribean, originally calculated by Emiliani (1971) and fitted to Broecker & van Donks (1970) time scale by Matthews (1973).

The palaeo sea-level curve (Fig. 59, 2) is based on the study of raised coral reef terraces on Barbados, the Ryukyu Islands and New Guinea and was put together by Bloom et al. (1974). The next curve (Fig. 59, 3) represents the well-known isotope curve from Camp Century on Greenland (Dansgaard et al. 1971 and 1972). Coope's (1975) and van der Hammen's (1967) curves (Fig. 59, 4 and 5) chow the average July temperature curves for England and the Netherlands, respectively. The last three curves in Fig. 59 are glaciation curves based on litho- and biostratigraphical data. The curves are for Baffin Island (Fig. 59, 6) (Andrews et al. 1975); southern parts of the Laurentide ice sheet's area (Fig. 59, 7) (Dreimanis & Goldthwait 1973); and the Weichselian ice sheet of Northern Europe (Fig. 59, 8) (Lundqvist 1974).

Several proposals exist for dating the boundary between the last interglacial

Fig. 59. Curves with paleoclimatic relevance for the last interglacial – glacial cycle. U. W. = Upton Warren interstadial Complex, C = Chelford interstadial, I = Ipswichian interglacial, D, H, B and A = Denekamp, Hengelo, Brørup and Amersfort interstadials, respectively, W I, II and III = Weichsel I, II and III, respectively. For further explanation, see text.

(Eemian) and the last glacial (Weichselian) which, according to the proposed correlations, is equivalent to the Hovden thermomer/Hovden kryomer boundary. Essentially the various suggestions fall into three periods; around 70–75,000 BP, e.g. McIntyre & Ruddiman (1972, p. 350), Dansgaard et al. (1972, p. 396), Zagwijn (1974, p. 376), Suggate (1974, p. 251); around 90–97,000 BP, e.g. Mørner (1972, Fig. 1), van der Hammen et al. (1971, Fig. 1), Phillips (1974, p. 589); around 110–116,000 BP, e.g. Matthews (1973, Fig. 2), Kukla & Kukla (1972, p. 423), Fairbridge (1972, Fig. 7), Coope (1975).

There appears to be no conclusive evidence for any particular one of these alternatives but recent opinions favour the latter (cf. Bowen 1978).

If one accepts the Førnes thermomer — Jämland interstadial — Brørup interstadial correlation, it is rather important to clarify dating of these. The Hardangervidda material provides no absolute datings. Lundqvist (1974) placed the Jämtland interstadial at about 50,000 BP (cf. Fig. 59, 8). However, his datings are not in agreement with Wijmstra & van der Hammen's (1974) view that the Brørup Interstadial seems to have a minimum age of 57,000 BP. Wijmstra & van der Hammen (1974), furthermore, hold that oxygen isotope phase 4 corresponds to a cold period during Early Weichselian time, for example between Amsersfoort and Brørup of before Amsersfoort. This then implies that Brørup is younger than 75,000 years, and a correlation with the warm period marked by oxygen isotope phase 3's oldest part and by coral reef complex IV from New Guinea then seems likely. Around this time, i.e. from c. 65,000 BP to slightly younger than 60,000 BP, the other curves in Fig. 59 also register a climatic amelioration; The Chelford interstadial and St. Pierre interstadial, amongst others, are located here. However, an even older age was proposed for these continental interstadials (Terasmae & Dreimanis 1976, Bowen 1978). Clearly there are a great many unresolved questions regarding the datings of the Early Weichselian interstadials, but the available data appear to favour an age of 60,000 BP or more for the Brørup interstadial, which should also correspond to the age of the Jämtland interstadial and the Førnes thermomer.

Where the duration of the Førnes thermomer is concerned, it has probably been quite short. Korpela (1969, p. 97) implies for the Peräpohjola interstadial that it 'vielleicht dort nur eine Zeitspanne von etwa 2000 Jahren umfasste'. The more central location of Hardangervidda to the ice sheet's initial position in comparison to North Finland, suggests that the equivalent ice-free period was even shorter here.

According to the above, the Førnes kryomer must have lasted from at least about 60–65,000 BP until c. 9,000 BP. With respect to dating the ice movement phases of the Førnes kryomer, there are few clues. Possibly the displacement of the flow pattern of Phase II to that of Phase III is connected with the ice sheet's growth during the cold period around 20,000 BP, which is well documented (cf. Fig. 59). The gradual displacement from Phase III to Phase IV was probably associated with the waning of the ice sheet after ca. 15,000 BP. Phase IV has been correlated with the Preboreal Eidfjord–Osa advance of about 9,700 BP (Anundsen & Simonsen 1968).

Summary and main conclusions

Introduction

The present study comprises an investigation of ice movement, Quaternary sediments and stratigraphy on Hardangervidda. These investigations are covered in the respective chapters but emphasis is laid on studying the relationship between them.

Hardangervidda is a high plateau c. 9000 km², most of it lying about 1200 m a.s.l. It is situated around 60°N and 7°E (Fig. 1). The bedrock consists mainly of Precambrian rocks in the east and Cambro–Ordovician in the west (Fig. 2).

Ice movements

Reconstruction of the ice-movement direction has been carried out by analysis of glacial directional elements based on field data (more than 1000 localities) and on aerial photographs.

Glacial directional elements are defined as: 'Oriented forms created by the moulding and sculpturing effect which moving glacier ice and its incorporated debris has on the underlying surface.' Directional elements are classified morphologically as:

Transverse	*Drumlinoid*	*Linear*	
crescentic fracture	rock drumlin	scouring	scratch
crescentic gauge	roches moutonnées		striae
lunate fracture	crag and tail		furrow
			groove
chattermarks	drumlin	fluted surface	fluted rock
conchoidal fracture	knob and tail		fluted sediments

Directional elements should be analysed with regard to direction and chronology as well as to property. Scanning electron microscopic studies of scratches can provide valuable additional information about direction and relative chronology. Property analysis includes those factors (form, size and in the case of sediments also composition) which may give information about the physical characteristics (temperature, thickness, velocity etc.) of the glacier. At the present stage, however, too little is known about this relationship for a detailed analysis to be worthwhile.

The directional element analysis from Hardangervidda shows that the direction of movement of the ice sheet can vary greatly over short distances. This can occur by plastic diffluence or confluence in areas of reasonably marked relief, or in the border zone between competing ice streams from different regions. Directional elements formed in unconsolidated sediments may be preserved even though younger ice movements with different directions have crossed over them.

Three distinct ice movement phases, Phase II, III and IV, can be recognised on Hardangervidda on the basis of directional elements. These are reconstructed on Figs. 12, 13 and 14, respectively. Directional elements which belong to subphases are also present. In addition to all these there is scouring from a probably independent, older ice movement phase, Phase I.

The growth of the ice sheet to Phase I and II, both of which had a westerly situated ice divide, most likely occurred with the weather pattern basically similar to today's, though with higher precipitation and/or lower summer temperatures. The relative relief and relative velocities of the ice sheet during Phase II are indicated in Fig. 16. The temperature in the basal layers of the ice sheet during Phase II was probably at the pressure melting point. However, during the onset of Phase I and II, the ice sheet locally has probably been cold based.

Change in the position of the ice divide from west (Phase II) to east (Phase III) was probably caused by a lowering of the ice sheet surface in the west due to a glacier surge along the Hardangerfjord draining ice from these areas.

The relative relief and relative velocities dring Phase III are indicated in Fig. 16. The temperature of the basal layers during Phase III was probably at pressure melting point. On the assumptions given in Vorren (1977b), the ice thickness, mean annual temperature at the surface and net mass balance at the ice divide during Phase III were 2500–2800 m, c. –25°C and c. 150 mm (in ice), respectively.

The change in relief and ice movement between Phase III and IV was gradual, being due to a general shrinking of the ice sheet, both in thickness and extent.

The relative relief and relative velocities during Phase IV are indicated in Fig. 16. Phase IV represents the last regional ice movement prior to the Holocene deglaciation of Hardangervidda.

Sediments

A total of 487 sediment samples have been collected and analysed with regard to grain-size distribution, content of phyllite, heavy minerals and coloured phyllosilicates; in some also for blue quartz content.

Till. A total of 240 basal till samples and 48 ablation till samples have been analysed.

Analysis of the phyllite content in different phi size classes between 4 and –3 phi shows that comminution of phyllite seems to occur in two stages. In the first, comminution gives a mode in the 3 to 2 phi (0.125–0.250 mm) and –1 to –3 phi (2–8 mm) fractions. In the second stage, these modes disappear, apparently with a displacement of the 3 to 2 phi mode towards smaller grain-size. The transition between the stages occurs rapidly and generally requires less than 1 km transport. Further comminution involves a relatively even reduction of the fractions between 4 and –3 phi. There are, however, indications that the largest of these fractions disappear first.

It is further shown that the phyllite content is dependent upon a) the areal extent of source rock; b) the relative relief both on a large and small scale (the relative relief determines the erosion mechanism which in turn determines the amount of material produced); c) the degree of comminution and dilution during transport which is dependent upon the mode and distance of transport.

With a transport distance greater than 15 km, most till samples (74.8%) contain less than 1% phyllite in the 3 to 2 phi fraction. Positive anomalies from this do occur and are caused by a) interglacial/interstadial fluvial transport and later incorporation of these sediments in till; b) transport of till as a thrust slice or, less likely; c) a rich source of weathered phyllite after a relatively long ice-free period.

The average content of heavy minerals (s.g. $\geqslant$2.88) in the 3 to 2 phi fraction in till on Hardangervidda is 3.84 and 3.67% for basal till and ablation till, respectively. The regional distribution of heavy mineral content shows that basic rocks have contributed little to the till. The average content of coloured phyllosilicates (mainly biotite) in the 3 to 2 phi fraction is 1.25 and 1.60% for basal till and ablation till, respectively.

A textural classification of till based on grain-size distribution below 2 mm is proposed as follows:

sandy till $>$ 60% sand and $<$ 15% clay
silty till $<$ 60% sand and $<$ 15% clay
clayey till $>$ 15% clay

According to this classification, the basal till samples from Hardangervidda are 0.4% clayey, 33.8% tilty and 65.8% sandy. All ablation till samples with the exception of one are sandy types.

The grain-size distribution of basal tills on Hardangervidda shows clear regional variations (Fig. 31). This is due partly to the petrographic composition; e.g. phyllitic till is distinctive in its relatively high content of gravel and clay. It is also partly due to incorporation of older sorted sediments and, to a lesser degree, the transport distance.

Seen in relation to the ice movement phases, most of the till in the eastern, southern (and central?) areas of Hardangervidda was deposited during Phase II. Most of the till in the northern and south-western areas are from Phase III. Phase IV has deposited some material in the northern areas.

The till on Hardangervidda is probably chiefly the product of mechanical erosion during the Late Pleistocene together with a smaller amount of chemical-weathered material mostly of Quaternary age.

Sorted sediments. A total of 96 glaciofluvial samples, 77 subtill and intertill samples and 4 glaciolacustrine samples have been analysed. There are considerable textural variations of sub- and intertill sediments, indicating differences in their primary genesis. There is a relation between sorting and mean grain-size parameters, showing that fine and medium sand are the best sorted sediments. Both finer and coarser sediments exhibit poorer sorting in general. This phenomenon is best explained partly by hydrodynamic principles (cf. Inman 1949) and partly by different availabilities of the individual size grades (cf. Folk 1974).

The mean-skewness relation is defined by a sinusoidal curve. Sediments with $M_z > 6$ phi and between 1 and 0 phi mostly have negative skewness, while those with M_z between 6 and 2 phi and between –1 and –3 phi usually display positive skewness. The reason for this is briefly discussed.

The phyllosilicate and heavy mineral contents in the 3 to 2 phi fraction of the sorted sediments are related to the samples' grain-size distribution. With coarser sediments the heavy mineral content increases while the mica content decreases. The phyllite content is apparently independent of grain-size distribution. These conditions are expliable by the different hydraulic equivalents of these grains.

The transport distance of fine sand in the Halne–Eitro esker varies between 1.5 and 10 km. This variation is interpreted as a dilution effect reflecting variations in the number and capacity of tributary channels.

Stratigraphy

Several localities with subtill and intertill sediments together with till sheets are described. The following climato-stratigraphy from oldest to youngest has been established on the basis of lithological studies and palynological analysis: Hovden thermomer, Hovden kryomer, Førnes thermomer, Førnes kryomer.

The Hovden thermomer was a relatively long ice-free period. During part of this, the vegetation in the bordering areas of Hardangervidda at least, contained *Picea* both in the east and the west. *Betula* dominated the three-vegetation later. Periodically there was a rich growth of fern, *Lycopodium* and *Sphagnum*. Soil development was far advanced. The Hovden thermomer is correlated with the Eemian or with parts of it.

The Hovden kryomer corresponds with ice-movement Phase I and is correlated with Ljungner's (1949) Prime Glaciation and Lundqvist's (1974) Weichsel I.

The Førnes thermomer represents an ice-free period, probably of short duration. Sediments found from this time are partly lacustrine, partly fluvial. Aeolian sediments are probably also represented. The vegetation on Hardangervidda during this thermomer was very sparse or non-existent. The vegetation in the lower lying peripheral areas such as around Møsvatn, has been dominated by herbs, mainly Gramineae. The three limit is assumed to have been at least 200 m lower than today. The thermomer is correlated with the Jämtland interstadial and the Brørup interstadial. It probably occurred before c. 60,000 BP.

The Førnes kryomer corresponds with ice-movement Phases II, III and IV. It is correlated with Ljungner's (1949) Posterior Glaciation and Lundqvist's (1974) Weichsel I and II. A tentative chronology for the ice-movement phases is suggested: Phase II, end of Førnes thermomer to c. 25,000–20,000 B.P., Phase III, c. 25,000–20.000 to c. 15,000 B.P., Phase III/IV, c. 15,000 to c. 9,700 B.P., Phase IV, some few hundred years around 9,700 B.P. Strong indications of sub-glacial drainage are found for Phase III and Phase IV.

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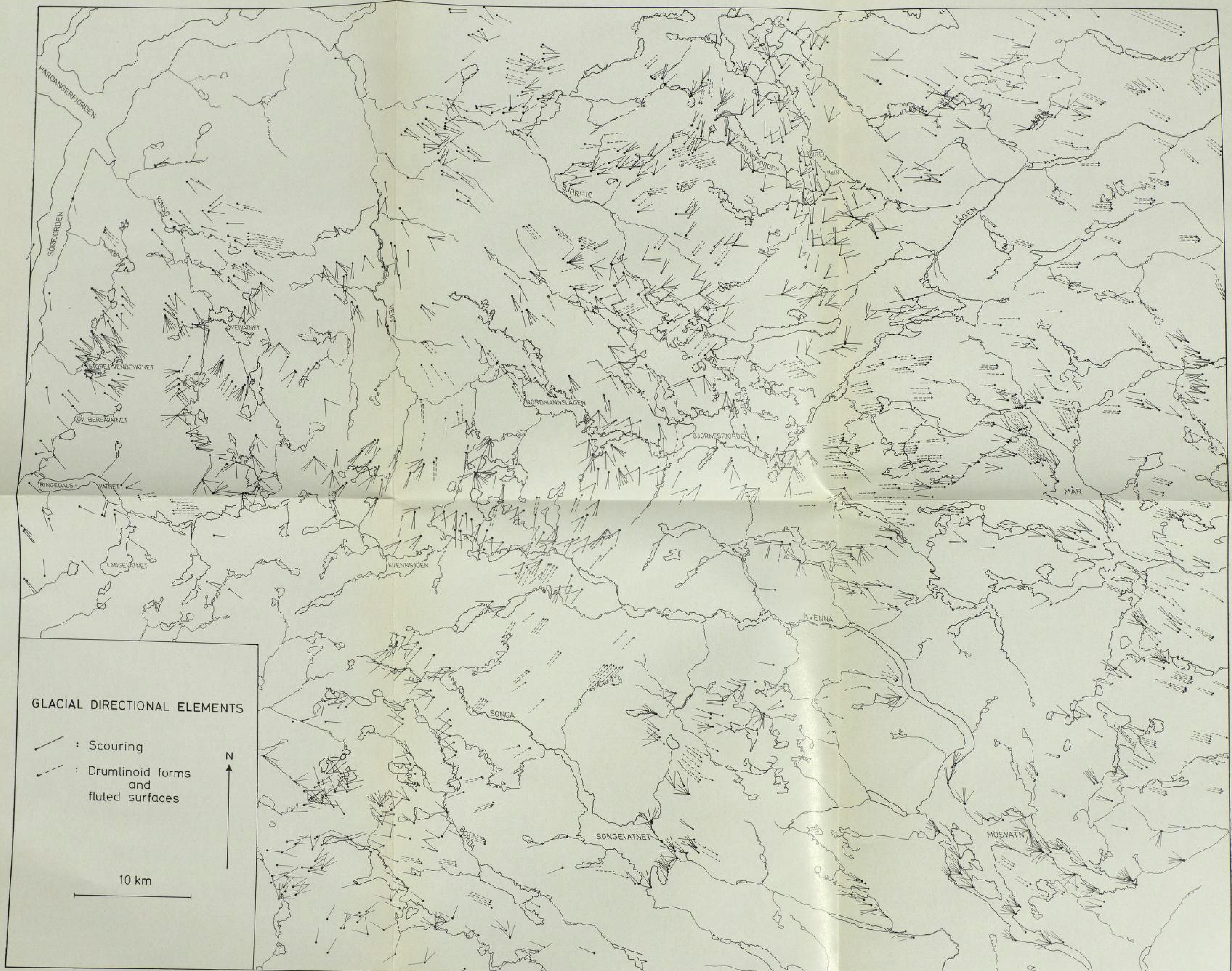
Pl. 2. Quaternary geological map of Hardangervidda.



Pl. 1. Map showing directional elements on Hardangervidda. The localities of the scouring observations are indicated by dots, and directions by lines towards the observation points. The direction of fluted surface and drumlinoid forms are indicated by dashed arrows.