Glacial Geology of the Area between Jostedalsbreen and Jotunheimen, South Norway

TORE O. VORREN

Vorren, T. O. 1973: Glacial geology of the area between Jostedalsbreen and Jotunheimen, South Norway. *Norges geol. Unders. 291*, 1–46.

Ice movement and the course of deglaciation are reconstructed. The age of and the climate during the deglaciation are discussed in the light of equilibrium line displacement, shoreline positions, pollen analysis and radiocarbon datings. The oldest regional ice movement was towards and out of Lusterfjord. Deglaciation is divided into five climatostratigraphic units: the Luster Interstadial (10200 ± 200 to 9800 ± 200 B.P.). a period of extensive glacial recession; the Gaupne Stadial (9800 ± 200 to 9500 ± 200 B.P.), a period of glacial advance and stagnation; a relatively short period of retreat follows this before the glaciers stagnate in the Høgemo Stadial; and finally (after 9100 ± 200 B.P.), a period of rapidly receding valley glaciers and downwasting of mountain glaciers.

Tore O. Vorren, Geologisk Institutt, avd. B, Universitetet i Bergen, 5014 Bergen – Universitetet, Norway

CONTENTS

Introduction	2
The oldest ice movement	3
Methods	3
Interpretation and discussion	3
Conclusion	5
The Luster Interstadial	5
Areal description	5
Conclusion	8
The Gaupne Stadial	8
Introduction	8
Areal description	10
Ice movement in the area between Fanaråken and Hestbrepiggan	21
Discussion and conclusion	23
The Høgemo Stadial	27
Areal description	27
Discussion and conclusion	29
The final deglaciation	30
Areal description	30
Conclusion	32
The deglaciation: age and climate	33
Equilibrium line displacements	33
Shoreline positions	34
Pollen analysis	35
Discussion and conclusion	37
Some trends in the deglaciation of Central West Norway	40
Main conclusions	42
Acknowledgements	43
References	43

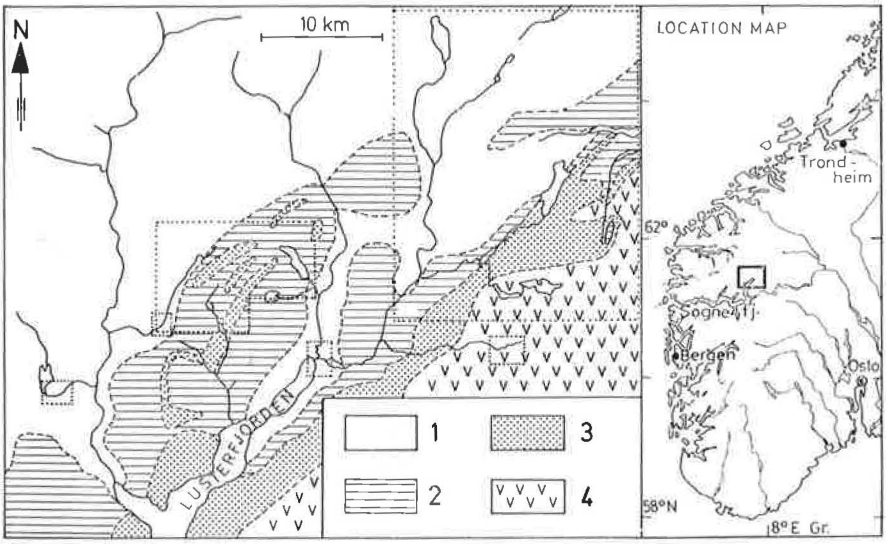

Fig. 1. Location map and simplified bedrock geology map of the investigated area. 1: Basal gneiss, 2: Mainly mica schist, 3: Mainly metasandstone, 4: Mainly basic plutonic rock. Dotted frames show location of figures.

Introduction

Regional bedrock geology (Fig. 1) of the investigated area has been mapped mainly by Rekstad (1914), Landmark (1949) and Banham (1968). The bedrock consists of three main units: rocks of the basal gneiss complex are mainly of granitic and quartz dioritic composition; the Eocambrian–Silurian metasediments are predominantly mica schists and metamorphosed sandstones, with limestones and conglomerates occurring in minor amounts; the overthrust Jotun eruptive rocks include basic and intermediate types.

Dividing the area into three elevation zones, below 800, 800–1600, and above 1600 m a.s.l., areal distributions of 8.9, 69.1, and 22.0%, respectively are obtained. The area can then be described as an undulating 'plateau surface' at approximately 1200 m a.s.l., above which the higher mountains rise, and into which the narrow valleys and fjords are eroded.

Because of pronounced relief large climatic variations exist within the area, with the climate in the lower-lying area being relatively continental (Bruun 1967, Det norske meteorologiske inst., in prep.).

The aim of the present paper is to reconstruct the history of ice movement and deglaciation in this area. Previous Quaternary geological studies have been concerned mainly with investigation of marine terraces (Kjerulf 1871, 1879, Helland 1876, Holmström 1880, Hansen 1891, H. H. Reusch (Brøgger 1901), Monckton 1913, Kaldhol 1941, Brathole 1951). Some marginal moraines have been identified in the mountain districts by Rekstad (1914), Ahlmann (1922), Liestøl (O. Holtedahl 1953), and Eriksson (1958). These observations indicate that deglaciation here was interrupted by periods of stagnation and/or advance of the glaciers. As a result of my investigations I have divided the deglaciation sequence climatostratigraphically into the following thermomers (relative warm periods) and kryomers (relative cold periods) (Lüttig 1965): Luster Interstadial (oldest), Gaupne Stadial, Gaupne/Høgemo Interstadial, Høgemo Stadial, and the final period of deglaciation.

The reader should note that the Norwegian words *dal*, *elv*, *vatn*, *bre*, and *fjell* mean, respectively, valley, river, lake, glacier, and mountain.

The oldest ice movement

Methods

Ice movement direction is determined by analysis of *orientation elements*, here defined as: 'Oriented form-elements produced by the glacier's own effects upon the surface over which it moves'. Johnson (1956) uses the same term, but in a much wider sense. The orientation elements are displayed graphically in Pl. 1. The orientation element which I term 'fluted rock' is best observed on aerial photographs, where it appears as a clear striation. Fluted rock occurs mainly in mica schists. Ridges and furrows are 2–3 m wide, with a height difference of ca 0.5 m between the bottom of the furrows and the top of the ridges. The furrows generally contain some weathered material, which supports the vegetation responsible for color variations in the aerial photographs. Breaks occur in places at right angles to the ridge axes, so that isolated roches moutonnées are formed.

In addition to relative age determinations, which are possible where several generations of striae are found, a relative age can also be obtained from the position of orientation elements in relation to other glacial features, especially marginal deposits. The interpretation of the ice movement is thus based on a total evaluation of glaciogeological phenomena.

Interpretation and discussion

The oldest regional ice movement is sketched in Fig. 2. Movements indicated in the northwestern area have been obtained by extrapolation and possible inaccuracies will thus be largest here.

West of Mørkrisdal and Inner Lusterfjord ice from the NW–NNW joined ice coming from the NNE and NE, and was deflected to the latter direction. At the mouth of the Jostedal and Gaupnefjord the ice flow from the NNW seems to have undergone less deflection from its original direction. This can be explained by an increased resistance to deflection owing to the greater amount of ice that moved through this low-lying area. The pattern of ice movement is more complex east of Fortunsdal. The striae at the mouth of Ringsdal indicate an old, westward movement. An old movement to the southwest seen on Nosafjell fits in well with the southwestward ice flow farther west. The above observations, plus the fact that there is a bastion at the mouth of Bergsdal, lead to the conclusion that there was westward ice flow in the northern parts of

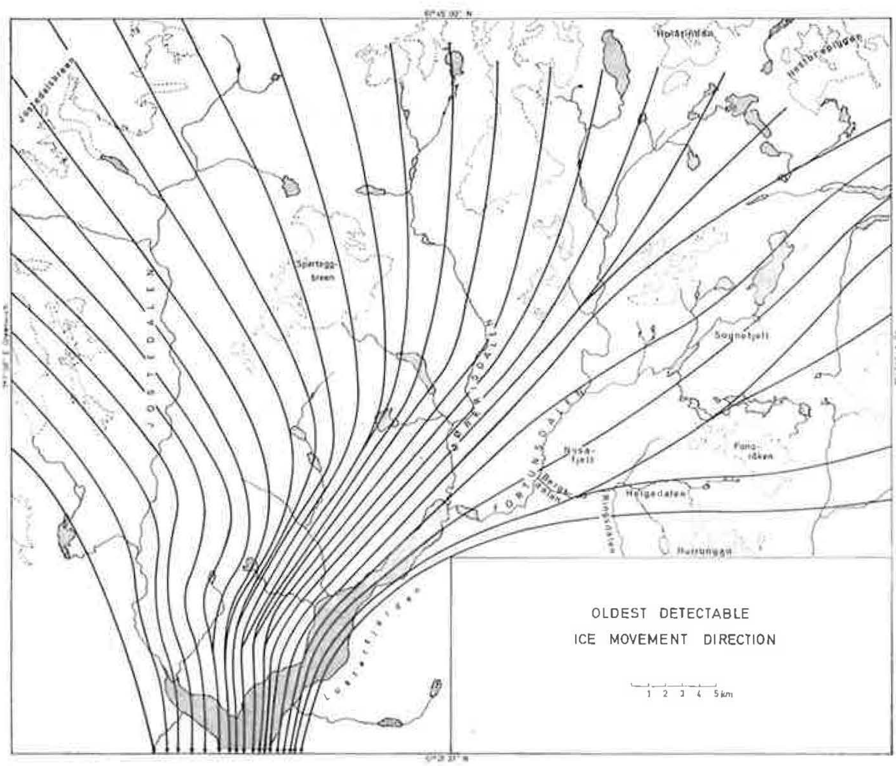

Fig. 2. Oldest regional ice movement inferred from observed orientation elements. Broken lines shows margins of present glaciers.

Hurrungane and the Helgedal–Bergsdal area. Farther west towards Lusterfjord, this ice flow was forced to the southwest both by topographical conditions and the southwestward-flowing ice in the north. Most of the orientation elements on Sognefjell are presumably of younger age, but all elements to which a pre-deglaciation origin can be assigned strike WSW and SSW.

The fact that none of the older orientation elements deviates from the direction sketched in Fig. 2 indicates that ice movement direction during Weichsel maximum might have coincided with the oldest ice movement direction found.

Studies in Antarctica and Southern Greenland (Aseev 1968, p. 112) indicate that subsurface topography influences the ice surface, and movement, if the relative relief over fairly large areas exceeds about ⅓ of the glacier's thickness. The relative relief in the Sognefjord–Lusterfjord area is 1500–2000 m; accordingly, 4500–6000 m far exceeds all reasonable estimates of the thickness of the inland ice during Weichselian time (Flint 1971, p. 596). One would thus expect convergence of drainage towards these fjords, as investigations at Lusterfjord indicate. With regard to the location of the ice divide, it is logical to assume that the high relief of the Jostedal plateau and of Jotunheimen resulted in individual ice culminations (Ljungner 1949) in these areas (O. Holtedahl 1953, p. 625, Strøm 1956). This assumption is supported in the case of Jostedal plateau by Fareth's (1970) observations on the northwestern side of the Jostedalsbre, which give a northwestward movement for the inland ice. To the north, in the Skjåk area, Tollan (1963) reports that in addition to a set of younger striae striking NE, there is a prominent set of striae oriented NW. Observations thus indicate that the ice divide extended from the northern parts of the Jostedal plateau, eastwards towards Jotunheimen.

Conclusion

It seems that during the Weichsel maximum ice drained from this area approximately as indicated in Fig. 2. The ice divide most likely extended from the Jostedal plateau to the Holåtinden–Hestbrepiggan area, and eastward to Jotunheimen. Ice culminations probably existed above the Jostedal plateau and Jotunheimen. The ice flow pattern sketched in Fig. 2 persisted until deglaciation began.

The Luster Interstadial

Areal description

The area west of Lusterfjord–Mørkrisdal. – Several lateral and subglacial meltwater channels are found in the Stordal area, at altitudes between 1050 and 760 m a.s.l. The channels are cut both in bedrock (Fig. 3) and in till.

Striae at the head of Dalsdal show that the ice moved down-valley. Similarly, an eastward ice movement is shown for the region around Opsarvatn. According to the orientation elements, the southern limit of this glacier must have been a few hundred meters south of Opsarvatn.

A widespread thinning of the ice sheet is thus indicated, as a result of which the highest areas near the fjord were deglaciated first. Ice drainage from the surrounding low-lying areas was channeled towards the fjord and valleys as the glacier surface in the Lusterfjord–Mørkrisdal area sank to the mountain-plateau level about 1000 m a.s.l.

The area east of Lusterfjord–Mørkrisdal. – Several meltwater channels cut in till are found on the south side of Helgedal. The 1–3 m deep channels run from a height of about 1100 m diagonally down the valley side to the NW. Younger almost horizontal lateral channels at about 1000 m a.s.l. are also found. These channels provided westward drainage for an ice-dammed lake at the mouth of Ringsdal, where glaciolacustrine sediments are found.

A terrace-like feature at 1280 m a.s.l. on the eastern side of Skagastølsdal has been interpreted by Eriksson (1958 p. 18) as a lateral moraine. In my opinion the feature itself, plus its location and extent, indicate that it was formed by lateral drainage. On the outer portions of the valley's eastern side are some decimeter-high furrows, which Eriksson (1958 p. 17) interprets as shore-lines or lateral channels. I have interpreted these as solifuction phenomena. At the mouth of the valley, a 4 m high ridge of sorted material is interpreted as an esker.

Fig. 3. Snow-filled meltwater channel eroded in bedrock south of Stordalen. Width 2–3 m.

Stratified and sorted material, somewhat disturbed, occurs on the northern side of Helgedal opposite the mouth of Ringsdal. These sediments were probably deposited laterally to a shrinking glacier that lay in Helgedal. The glacier in Helgedal stagnated briefly after it had retreated eastwards to the valley threshold found just west of the mouth of Skagastølsdal. Marginal deposits from this halt take the form of a 2 m high ridge farthest to the north, which grades into a terrace-form farther in along the valley. The nature and extent of the marginal deposit indicate that there was no ice supply from Skagastølsdal during the time of deposition. The barely noticeable rise in elevation of the deposits towards the east may indicate that there was little dynamic activity by the glacier at this time.

Remains of lateral moraines deposited by a glacier in Fortunsdal are found at elevations of 960 and 1015–1065 m, on two small spurs on the eastern side of the valley, just north of Bergsdal. Smaller lateral moraines, presumably corresponding to these, are found at about 900 m a.s.l. on the western side of Fortunsdal, opposite Bergsdal. These moraines, deposited before the Gaupne Stadial, have no parallel in the rest of the area except possibly the marginal deposit in Helgedal. Assuming a reasonable gradient for the glacier surface in Fortunsdal, a corresponding frontal deposit should have been located somewhere in the outer portion of the inner basin in Lusterfjord (Fig. 4), but no frontal deposit has been found there (see below).

The above phenomena indicate that the mountains between the tributaries to Helgedal were deglaciated first, with the glaciers in Helgedal and Ringsdal eventually becoming separated. Later, the connection between the glaciers in Helgedal and Bergsdal was also broken, so that the latter became a diffluent

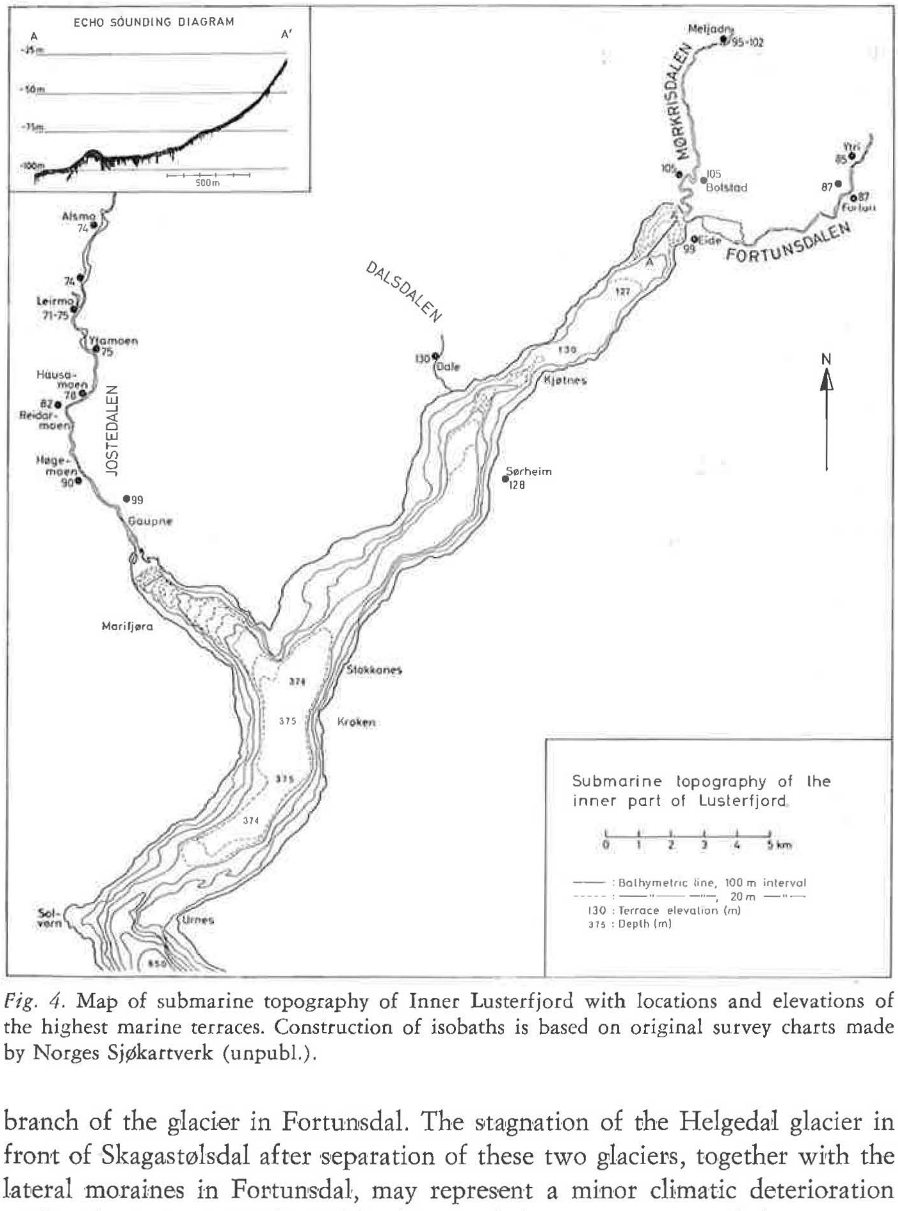

Fig. 4. Map of submarine topography of Inner Lusterfjord with locations and elevations of the highest marine terraces. Construction of isobaths is based on original survey charts made by Norges Sjøkartverk (unpubl.).

branch of the glacier in Fortunsdal. The stagnation of the Helgedal glacier in front of Skagastølsdal after separation of these two glaciers, together with the lateral moraines in Fortunsdal, may represent a minor climatic deterioration within the main, more favorable climate of the Luster Interstadial.

The Lusterfjord area. – The only possible lateral deposit on the steep walls of the fjord is located just outside Stordal at an elevation of 440 m. The deposit consists of till and occurs partly in the form of a ridge and partly in the form of a terrace. It can be either a lateral moraine or a glaciofluvial erosional form.

Three well-defined basins, with depths of 650, 375 and 130 m, exist in Lusterfjord (Fig. 4). The near horizontal nature of the basin floors results from filling by large amounts of sediments. Echo-soundings across the innermost sill, near Kjøtnes, indicate that this sill consists of hard rock. The two outermost sills may represent frontal deposits. If so, they need not necessarily represent a glacial advance caused by a change to a colder climate, as topography alone could cause a halt here.

Only in a few places is it possible to determine the highest relative sea level after the retreat of the ice from Inner Lusterfjord. At Kroken (Fig. 4) there is a basin with a threshold elevation of 124 m that contains marine sediments (p. 35). The marine limit must therefore be above 124 m. At Sørheim there is a flat surface nearly 100 m wide cut in unconsolidated material. A steep slope extends from the outer edge of the flat, at 128 m, down to the fjord. The surface rises gently inward towards the side of the fjord with a gradual transition. Sand and gravel are found on top of the flat, while deeper sections reveal till. This flat is probably a marine abrasion surface cut when sea level stood a little more than 128 m higher than at present.

At the mouth of Dalsdal is a terrace with an elevation of 130 m. The glaciofluvial material (Fig. 5) of which the terrace is constructed must have been deposited in the fjord after the Lusterfjord glacier had retreated beyond Dale.

Kjerulf (1871) and A. M. Hansen (1891) describe what they term a terrace 131 m a.s.l. at Kroken, and Hansen reports another at 138 m near Stokkanes. My investigations lead to the conclusion that no terraces exist at either of these localities and a marine limit in the Inner Lusterfjord of ca. 130 m above present sea level can therefore be assumed.

Conclusion

The oldest deglaciation forms in the area bear witness to a period when climatic activity of glaciers was small. Dynamic inactivity may have occurred locally, but the high relief would tend to prevent stagnation. Relative sea level stood ca. 130 m higher than at present when the glacier receded from the Inner Lusterfjord. Except for a possible climatic deterioration in the later part of the interstadial, the climate was generally favorable.

The Gaupne Stadial

Introduction

The Gaupne and Høgemo Stadials are characterized by several periods of glacial advance and stagnation. Discussion of these stadials will therefore center on the resultant marginal phenomena. A good deal of variation exists in terminology relating to marginal phenomena. The terminology used here is modified from Andersen (1960).

Often two or more marginal moraines occur within a limited area or elevation zone. These moraines are said to constitute a moraine belt. Synchronous moraine belts and marginal deposits from different localities belong to the same marginal zone.

Frontal deposits in the large valleys will be treated separately. The following

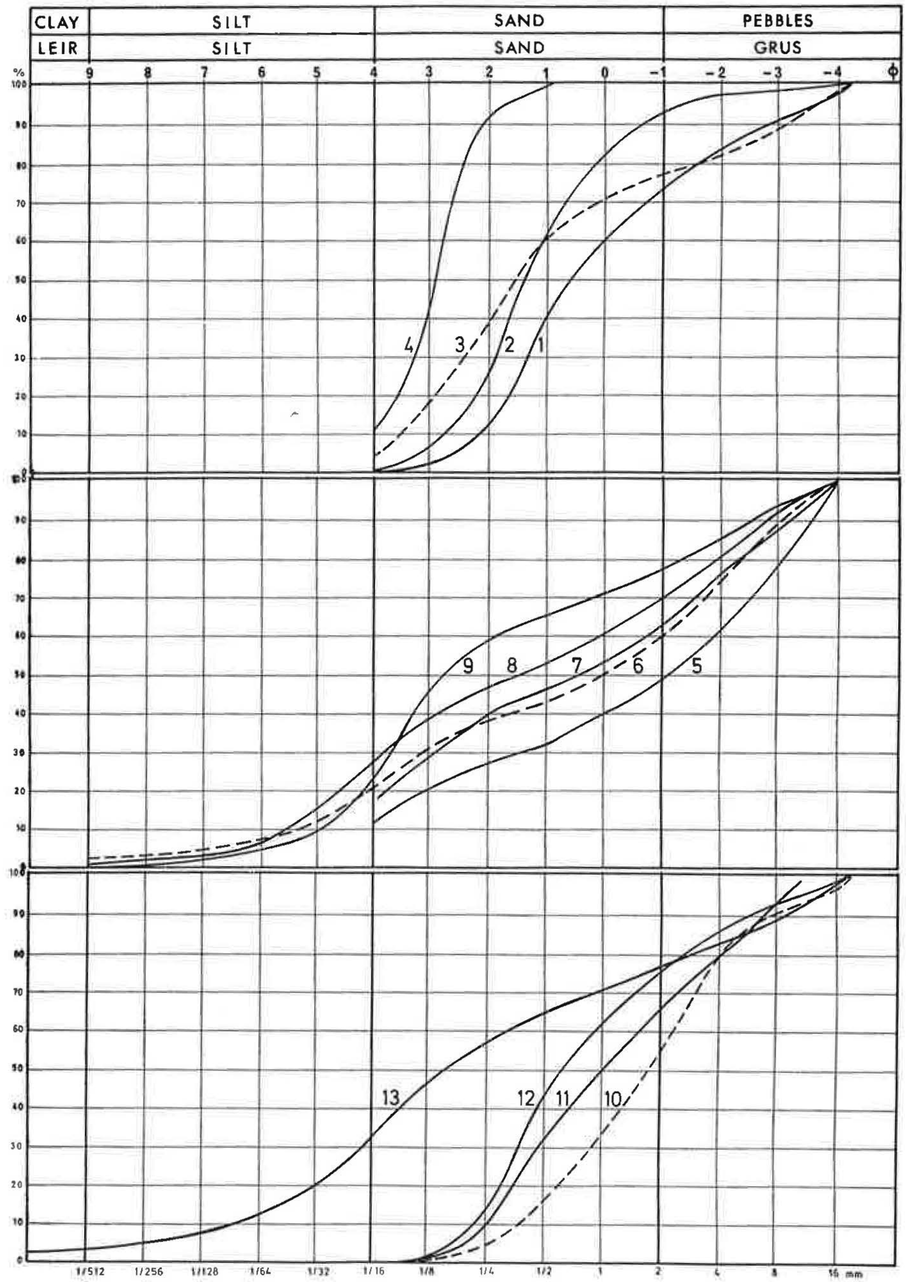

Fig. 5. Grain-size distributions. Glaciofluvial sediments from Høgemoen (1 and 2), Dalsdalen (3) and Eide (10, 11 and 12). Samples from marginal moraines: S side of Leirdalen (5), W of Fanaråken (6), N of Granfasta (7), S of Granfasta (8). From the outer series at Styggedalsbreen (9), Eide (13). Glaciolacustrine sediments in Leirdalen (4).

general description can be given of the other marginal deposits. The majority of these deposits are frontal and lateral moraines which vary in height from 1–16 m, with heights between 3 and 6 m most common. The material is normally unsorted (Fig. 5), but can, in some instances, consist entirely of boulders. The petrographic composition shows that the material is mostly derived from nearby bedrock and is thus of parautochthonous character.

The marginal deposits are displayed graphically in Pl. 1. Because of the small scale of Pl. 1, one symbol often represents two or more marginal moraines.

Areal description

The Tunsbergdal–Leirdal area. – Lateral moraines from the glacier occupying Tunsbergdal during the Gaupne Stadial are found on the mountain ridge between Leirdal and Tunsbergdal. The most conspicuous of these is a 1.5 km long ridge which lies 950 m a.s.l. in the north, and drops to 850 m a.s.l. in the south. The presence of a terrace-like lateral moraine on the western side of Tunsbergdal, lying just below the Hestbre Glacier, demonstrates that the latter has not extended down to the glacier in the Tunsbergdal. In the southernmost part of Tunsbergdal, the corresponding moraine belt lies between 800 and 870 m a.s.l. and consists of up to 12 ridges.

On the northern side of Leirdal a 100 m long lateral moraine at 800 m a.s.l. drops westward. This was clearly deposited by the glacier which existed in Jostedal.

The Jostedal area. – With the exception of a small break, caused by the steepness of the terrain, the moraine belt from Leirdal continues farther south into Jostedal. The proximal lateral moraines descend steadily from about 600 m a.s.l. in the north to 385 m a.s.l. just north of Kvernelvi, an average drop of 65–70 m/km. The distal moraines seldom lie more than 50 m above the elevation of the proximal moraines. The vertical extent of the moraine belt is smaller here than in Leirdal, owing to the very steep drop of the most distal moraines from Leirdal and out into Jostedal. Although direct evidence, such as intersecting moraines, is lacking, it appears that the moraines representing the maximum extent of the glaciers during the Gaupne Stadial are younger in Jostedal than in Leirdal. In other words, the intermittent advances and stagnations of the glacier in Leirdal resulted in deposition of lateral moraines almost always in a more proximal direction, while the oldest moraines in Jostedal were overrun by subsequent advances before an intermittent shrinkage began.

The southernmost lateral moraine in Jostedal, lying in the Kvernelvis Valley at an elevation of 260 m, indicates a very steep slope for the short tongue which branched into this valley. It seems to be common for small diffluent tongues to obtain high gradients (Anundsen & Simonsen 1968, p. 26, Fareth 1970, p. 88).

The Gaupne deposit is a remnant of the frontal deposits laid down by the glacier in Jostedal during the Gaupne Stadial (Fig. 6). It lies against the

Fig. 6. Gaupne and the mouth of Jostedalen, showing the Gaupne frontal deposit, G, and the Høgemoen frontal deposit, H.

eastern side at a point where the valley widens out. The terminus of the glacier would naturally stop at this widened point. The deposit has the form of a half-cone terminated at the top by a nearly horizontal surface with an elevation of 99 m. This represents sea level at the time of its formation. Glacial sediments are found up to approximately the same elevation on the opposite side of the valley. Unfortunately, there are at present no exposures in these deposits from which more detailed information can be obtained.

Scattered moraine belts are found along the eastern side of Jostedal at an elevation corresponding to the moraine belts on the western side. The northernmost of these, just south of Vigdal, is at 870 m a.s.l.

Marginal deposits upon which it is possible to base a reconstruction of several phases of the Gaupne Stadial are found in Vigdal near Storhaug (Fig. 7, A). During the earliest stadial phases (Fig. 7, B) a glacier flowing down Vigdal from the north joined another stream of ice from the west, which was a diffluent branch of the glacier in Jostedal. Later these ice flows separated, and frontal moraines were deposited by both (Fig. 7, C). The surface of the tongue from Jostedal was higher after the separation than during the earlier phases (compare Figs. 7, B and 7, C). This is a further indication that the glacier in Jostedal underwent a period of advance, culminating in the Storhaug Phase (Fig. 7, C), before shrinkage began.

There are no Gaupne Stadial lateral moraines found on the west side of Jostedal, north of Vigdal. There are, however, isolated lateral moraines on the eastern side of Jostedal. The northernmost and highest of these lies in the pass between Vanndal and the adjacent valley to the south. The moraine ridge rises

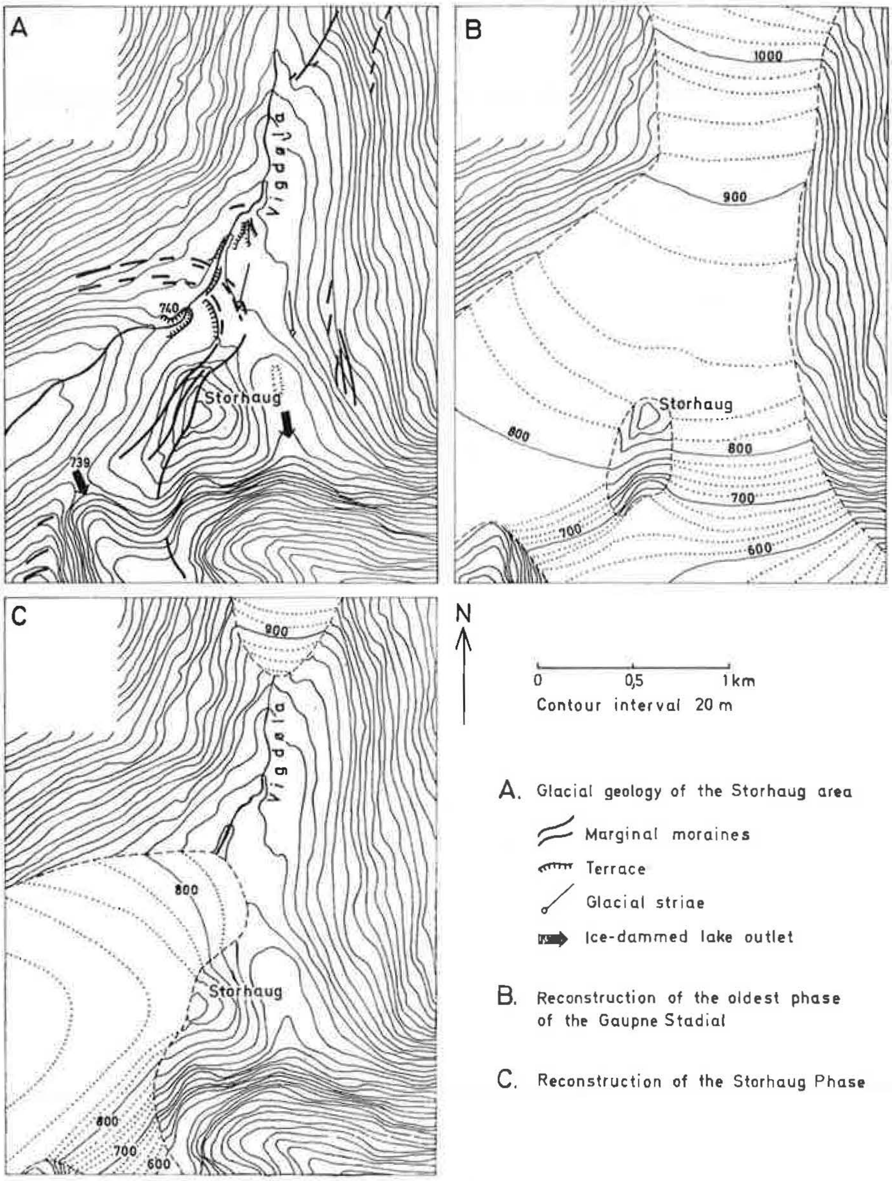

Fig. 7. Glacial geology sketch-map, and maps showing glacial extent during two phases of the Gaupne Stadial.

to 1250 m a.s.l. to the east and west. Clearly, these moraines were deposited by a glacier supplied from both Jostedal and the inner parts of Vanndal.

The mountain area between Jostedal and Mørkrisdal. – Marginal deposits from several smaller outlets are found in this area. An almost continuous series of deposits exists between the distal and proximal marginal moraine ridges. The

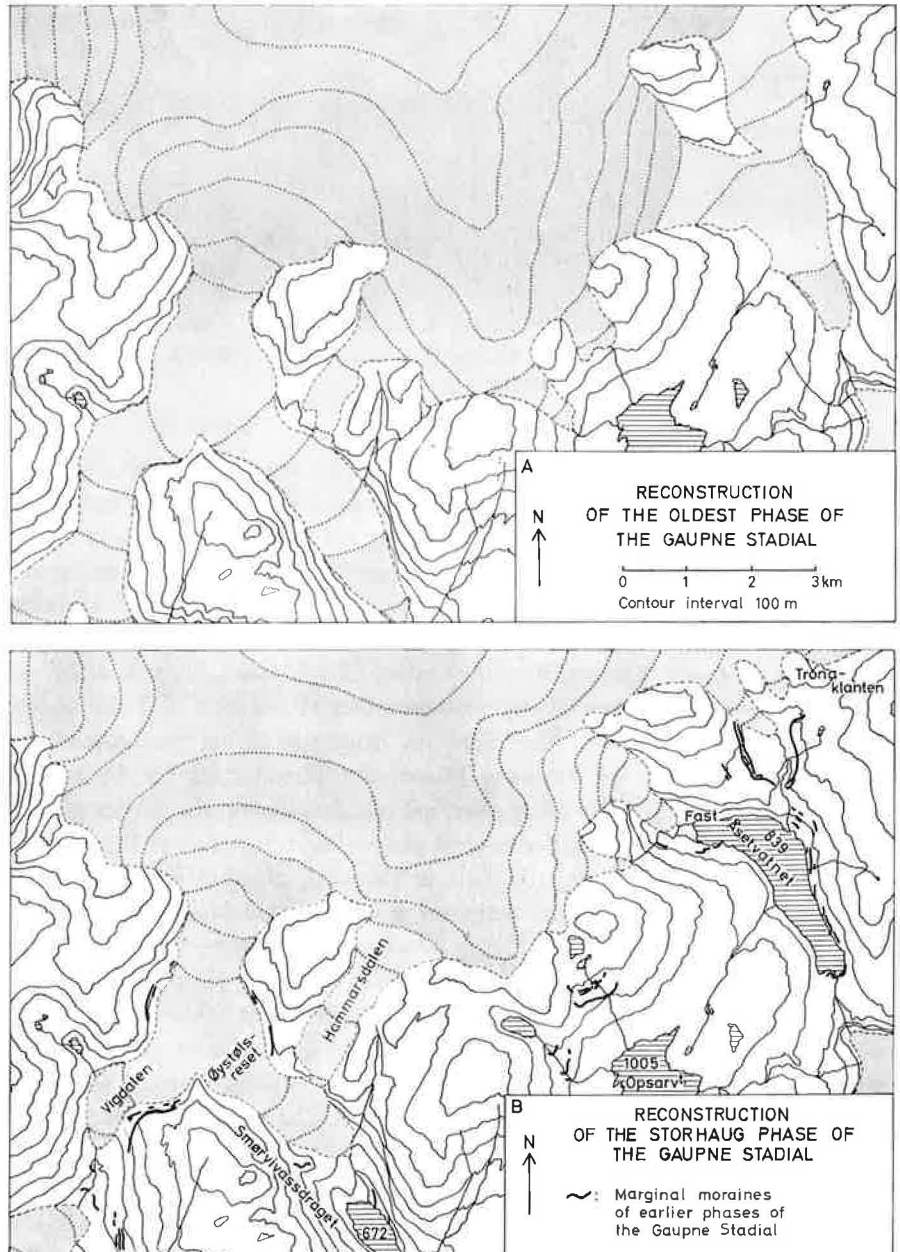

Fig. 8. Reconstruction of glacial extent in the mountain area between Jostedal and Mørkrisdal, during two phases of the Gaupne Stadial.

extent of the glacier during the earliest phase plus the Storhaug Phase of the Gaupne Stadial can be reconstructed with a good deal of certainty (Fig. 8).

With a few exceptions, the oldest phase (Fig. 8 A) is represented by the most distal moraines. The exceptions are found in Hammarsdal and east of

Fig. 9. Frontal moraine of the Storhaug Phase at Fast viewed towards NW. Ridge crest dotted. Arrow indicates direction of flow of glacier.

Tronklanten. North of Øystølsreset the oldest of the lateral moraines deposited by the glacier which occupied Smørvivassdrag are preserved, demonstrating that the glacier in Hammarsdal did not completely fill the southwest parts of the valley during the earliest phase. The terminus of the glacier in Hammarsdal thus must have been proximal to the outermost 7–8 m high frontal moraine, which I believe was deposited during the Storhaug Phase.

There is only one marginal moraine east of Tronklanten. This moraine can be correlated with what are clearly push-moraines to the west of Tronklanten. The latter in turn intersect the moraines from the earlier phases and are believed to date from the Storhaug Phase. The glacial flank to the east of Tronklanten probably had a more proximal position during the earlier phases.

The glacier surfaces during the earliest phases had relatively small gradients, indicating that only small reactivations of stagnant, ablating ice masses took place. This is illustrated by the marginal moraines from phases between the earliest phase and the Storhaug Phase, which lie in a depression to the north of Åsevatn. The location of these moraines shows that the glacier lay 20–40 m higher up on the western than on the eastern slope, even taking into account the most proximally lying moraines on the western slope. The explanation must be that during the preceding thermomer, the ice had ablated faster on the steep, heat-absorbing west-facing slope of the depression. The subsequent reactivation of the ice mass, during which the marginal deposits were formed, was of such modest extent that it was not possible to re-establish stability in the glacier.

The paleogeographic reconstruction of the Storhaug Phase (Fig. 8 B), is based on conclusions arrived at in the type area around Storhaug. This indicates a period of marked glacial advance. There is a particularly clear group of marginal deposits found in this area. These include frontal moraines up to 7 m high in Vigdal and in front of Smørvivatn, the 7–8 m high frontal moraine in Hammarsdal, the 8–10 m high frontal moraine at Fast (Fig.9) and the clear push-moraines west of Tronklanten. All of these are assumed to be from the Storhaug Phase.

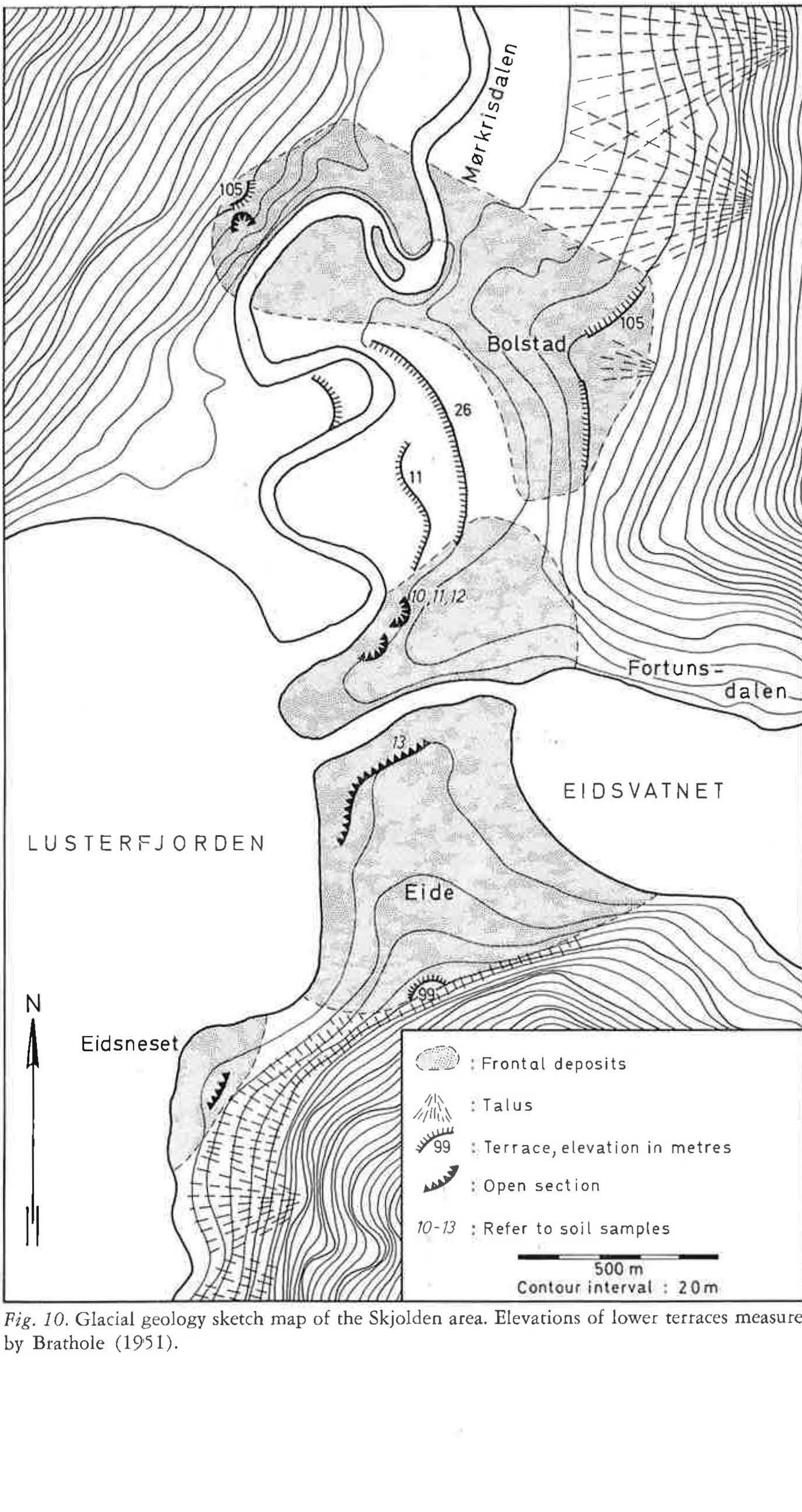

Fig. 10. Glacial geology sketch map of the Skjolden area. Elevations of lower terraces measured by Brathole (1951).

Fig. 11. Aerial photo of the Skjolden area. B: Bolstad, E: Eide, En: Eidsneset. In the background Fortun frontal deposit, F, Fanaråken, Fa, and Granfasta valley, G (photo by Widerøe).

The Mørkrisdal area. – Two frontal features were deposited by the glacier that occupied Mørkrisdal during the Gaupne Stadial: the Eidsnes deposit, deposited in front of a glacier that was probably also supplied from Fortunsdal, and the Bolstad deposit (Figs. 10 and 11).

The following sequence is observed in an approximately 100 m long open section in the Eidsnes deposit, from south to north: ca. 60 m of poorly bedded sand and gravel, containing subrounded pebbles and cobbles. These layers dip south, and are bounded on the north by a 15 m long sequence of stratified silt. The stratification in the silt is somewhat disturbed, with fold axes striking at right angles to the fjord. Farthest to the north is a 30 m long sequence of layered pebble-gravel that dips north.

The following depositional history is proposed: the silt layers were deposited first, and later disturbed by a glacier moving out to the fjord. The sand and gravel were deposited at this later stage by meltwater rivers. After the glacier had receded, relative sea level sank and marine processes redeposited material on both sides. That the Eidsnes deposit really is a frontal deposit seems to be confirmed by echo soundings, which show a definite ridge-form off Eidsnes (Fig. 4). The ridge is presently 5 m high on the proximal side and 10–15 m high on the distal side, with a width of about 150 m. It must be considered, however, that sedimentation after the formation of the ridge has reduced its relief; the weak echo penetration on the south side presumably shows the distal slope beneath the younger, finer sediments.

The Bolstad deposit extends completely across Mørkrisdal (Figs. 10 and 11). The elevation of the terrace on the east slope of the valley is 105 m a.s.l., and a somewhat higher-lying talus cone is found near the middle of the terrace. The western terrace, with an area of 150 m^2, also has an elevation of 105 m a.s.l. Thus it is reasonable to assume that this was the sea level at the time of deposition. An open section 60–80 m a.s.l. in the western portion of the deposit reveals sand, gravel and small boulders. Brathole (1951) has described an exposure near the eastern terrace where, beneath 1.5 m of sorted material, he found 'hard-packed clay'. This clay could be a clayey, silty moraine, or, alternatively, a marine clay incorporated into the deposit. The latter possibility is supported by Reusch's (Brøgger 1901, p. 545) statement that marine shells have been found in the vicinity.

No lateral phenomena which can be correlated with the Eidsnes or Bolstad deposits are found on the western side of the valley. On the eastern slope the southernmost lateral moraines are found at Rebnis Li, where 3–4 ridges curve around the valley spur from 530 up to 600 m a.s.l. One of these ridges dams a 7 m deep bog. Barely 3 km farther north, a 5 m thick moraine cover extends to an elevation of 880 m a.s.l., where it is terminated by nearly horizontal lateral moraines. Correlating this with the moraines at Rebnis Li, it seems that the glacier surface had dropped 280 m over a distance of barely 3 km. However, a high gradient is to be expected here, as the valley floor shows an equally large descent over the same distance. Outside of Haugs–Sveigdal is a 15 m thick, terraced lateral moraine, 920 m a.s.l. Within the valley are frontal moraines deposited by a cirque glacier, presumably during the Gaupne Stadial. Three smaller lateral moraines on the south side of Bolstad–Sveigdal, about 1050 m a.s.l., indicate that the glacier which occupied the Mørkrisdal was in contact with the glacier in this valley. The elevation of these moraines corresponds to the height one would expect for the Mørkrisdal glacier surface on the basis of the lateral moraine west of Tronklanten. Their correlation with the Gaupne Stadial moraines is thus clear. It is difficult to say whether the lateral zone in Mørkrisdal should be correlated with the Bolstad deposit or the Eidsnes deposit, or both. It seems probable, however, that the Mørkrisdal glacier, like the glacier in Jostedal, was thinner during the earliest phases, and that the lateral zone in question, with its small vertical extent, was formed during later phases, contemporaneous with the deposits at Bolstad.

The Fortunsdal area. – During the Gaupne Stadial the Eidsnes and Eide deposits were laid down in front of a glacier in Fortunsdal. The Eide deposit lies at the mouth of the valley, and dams Eidsvatn (Figs. 10 and 11). It is terraced in the south at an elevation of 99 m. The terrace dips gently east and is probably not a primary surface (Monckton 1913). The terrace does not, therefore, correspond to sea level at the time of the formation of the Eide deposit. Open sections in the Eide deposit north of the river reveal glaciofluvial material (Fig. 5). The layers dip to the west and northwest at angles of up to 35°. Till

Fig. 12: Marginal moraines on the southern side of the Granfasta Valley seen towards NE. Sheep on the nearest ridge indicate scale.

overlain by horizontal layers of sand and silt (Fig. 5) was observed in the large open section south of the river. The Eide frontal deposit is thus primarily constructed of both sorted and non-sorted sediments. The horizontal layers of sediment, which contain shells of mainly boreal molluscs (Reusch, in Brøgger 1901), together with eastward-dipping layers on the northern proximal slope (Monckton 1913), indicate that modification of the material by marine processes occurred after the terminus had retreated.

Lateral glacial phenomena corresponding to the Eide deposit are found at several places on the western side of the valley. There are widely spaced moraine belts which descend from 1080–1180 m 2 km north of Bjørkenosi, to approximately 600 m a.s.l. across from Bergsdal. The moraine belts are composed of up to five ridges, 2–5 m high. The topography along the eastern side of the valley limits the possibilities of the existence of lateral moraines to Bergsdal. Large concentrations of till between 450 and 500 m, with a steep, westward-facing slope, may mark the proximal border of the ice which lay here during the Gaupne Stadial.

The Granfasta Valley. – The glacier in Fortunsdal was nourished not only from the north but also through the Granfasta Valley. The Gaupne Stadial's marginal zone is represented on the south side of the valley by a moraine belt consisting of up to 6 parallel ridges, 2–10 m high (Fig. 12). Metasediments are the dominant constituents, but there is also a considerable amount of Jotun Gabbro, comprising 15–20% of the pebble fraction in the west, and increasing to 25–

35% in the east. The heavy mineral content of the fine-sand fraction in the west is 28.1%.

On the northern side of the valley the distal ridge of the correlated moraine belt lies between 1220 m a.s.l. in the west, and 1300 m a.s.l., in the east. The till from the lateral moraines here shows a grain-size distribution similar to that of material from the south of the valley (Fig. 5), but distinguishes itself petrographically by the absence of Jotun Gabbro. The difference also appears in the fine sand fraction, the heavy-mineral content here being only 13.0%. Evidently the stream of ice from the east which transported the gabbro material was deflected into the down-valley direction of flow of the Granfasta glacier, so that no gabbro was transported to the northern flank.

The Rya Valley. – The composition of the material in the Gaupne Stadial's marginal belt on the south of the Granfasta Valley, together with orientation elements, shows that an ice source existed in the area north of Fanaråken. This same area also nourished a glacier flowing down the Rya Valley. Marked marginal moraines show the extent of the glacier to the east, but no trace of marginal deposits is found along the western side. The eastern marginal belt begins in the northeast at an elevation of 1550 m a.s.l. and can be traced, with minor breaks, down to 1050 m a.s.l. Four moraines, with a maximum height of 8 m, constitute the belt. The belt lies in front of the small cirque glacier west of Fanaråken; the latter thus cannot have reached the Rya Valley during the Gaupne Stadial. There is, moreover, nothing that suggests that its extent during this period was greater than that indicated by the sub-Recent marginal moraines.

The Styggedalsbre Glacier and vicinity. – Scattered remains of marginal moraines (Fig. 13) to the west and north of Skautetjern Lake indicate that a standstill or very small advance of the glacial margin took place immediately after it had retreated to a position opposite the abrupt end of Helgedal. A small frontal moraine remnant is found on the southern bank of the river that flows from Illvatn. The natural continuation of this moraine is a high-angle slope in the profile of the talus cone found on the north side of the river. When the glacier lay here, talus must have accumulated at its margin. The high-angle slope formed during glacier recession.

Several concentric moraine ridges, 2–6 m high, lie beyond the terminus of Styggedalsbre. On the basis of vegetation cover and soil profiles, two series can be distinguished. As Ahlmann (1922 p. 46) also concludes, there can be no doubt that the inner series is sub-Recent/Recent, i.e. from the A.D. 1750 advance or younger, while the outer series is older. In contrast to the inner series, which was deposited by Styggedalsbre alone, the outer series was deposited by both Styggedalsbre and a second glacier that flowed out of Styggedal and received supply from an inland ice to the east. Proof of this is seen where one of the outer moraines can be followed to the southeast up Styggedal (Fig. 14). Three parallel ridges are found here, the most distal of which can be

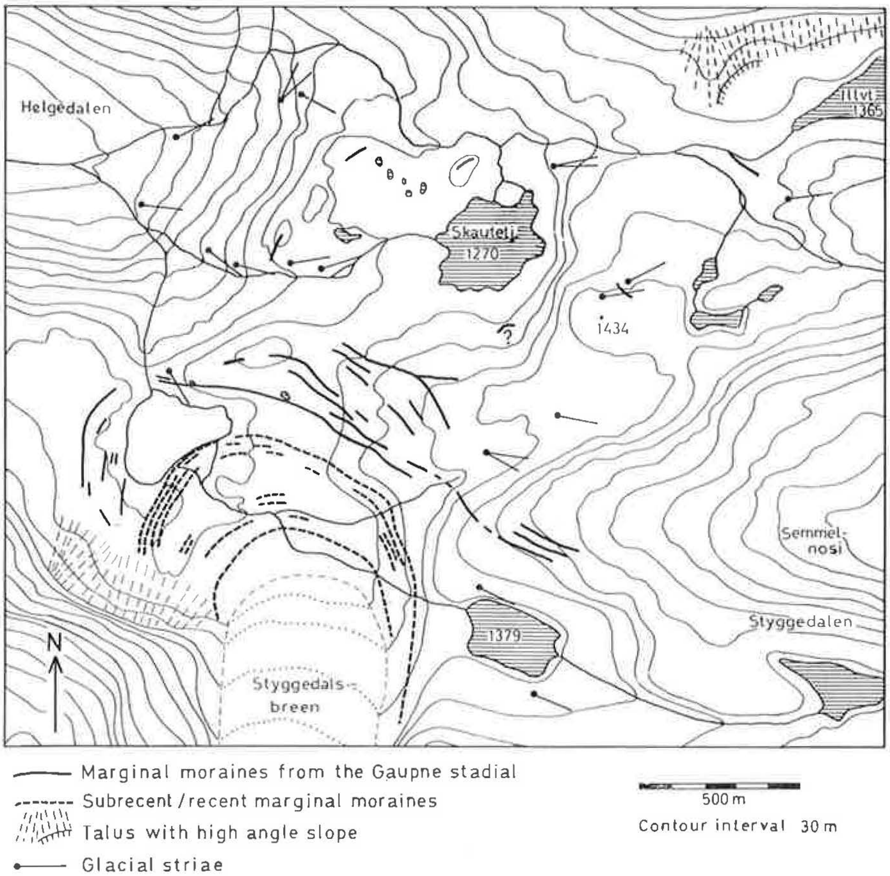

Fig. 13. Glacial geology sketch map of the Styggedalsbre area.

followed up to 1480 m a.s.l. As the outer series of moraines was deposited in front of a glacier receiving supply from the inland ice, it seems clear that, for morphostratigraphic reasons, it must be assigned to the Gaupne and also possibly the Høgemo Stadials.

The Skagastølsdal area. – Eriksson (1958) describes several marginal moraines in Skagastølsdal. In the middle of the lower parts of the valley a low, wide mound arcs across the valley floor, its convex side pointing down-valley. Eriksson interpreted this as a transverse moraine deposited by a glacier which extended down the western wall of the valley. As the mound can be followed up the west side of the valley until it disappears beneath the talus, it is in my opinion a moraine deposited in front of a glacier moving down Skagastølsdal itself. It may be that this moraine can be correlated with the marginal deposit in Helgedal, while the marginal moraines farther up the valley date from the Gaupne and possibly also the Høgemo Stadials.

Fig. 14. Aerial photo looking south showing Styggedalsbreen to the right. Arrow indicates a marginal moraine of the outer series which can be traced to Styggedalen. In the foreground Fanaråken, 2069 m a.s.l., with meteorological station. (Photo by Widerøe, Sept. 1936.)

Ringsdal. – No marginal deposits have been found on the distal side of the Recent/sub-Recent frontal foraines outside the terminus of the Ringsdal glacier (south of the area shown in Pl. 1). The Ringsdal glacier is thus the third glacier which was of lesser extent during the Gaupne and Høgemo Stadials than during sub-Recent time.

Ice movements in the area between Fanaråken and Hestbrepiggan

The youngest ice movements as deduced from observations of orientation elements, are indicated in Fig. 15. The main orientation elements observed are striae, but they are frequently accompanied by crescentic gouges and fractures indicating actively flowing glaciers. The movements in the several local areas in Fig. 15 seem to me clearly interrelated, and I thus assume them to be synchronous. This means that there was an ice culmination above the Hestbrepiggan–Holåtindan area connected by a saddle-like ridge extending SSE, east of Storevatnet to the Smørstabbreen area, where the ice again domed up. In addition, there must have been a smaller NE-trending ridge on the north side of Fanaråken and probably a third ridge above the Liabre massif. In this connection, it is worth mentioning Tollan's (1963) observation of drumlins at the mouth of Tundradal, 12–12 km north of the map edge, which indicate that the youngest ice movement was to the NE, parallel to the valley. The drumlins are most likely synchronous with the movements outlined here, indicating that there was an ice divide to the north in the vicinity of Hestbrepiggan–Holåtindan, i.e. approximately where the regional water divide is now located.

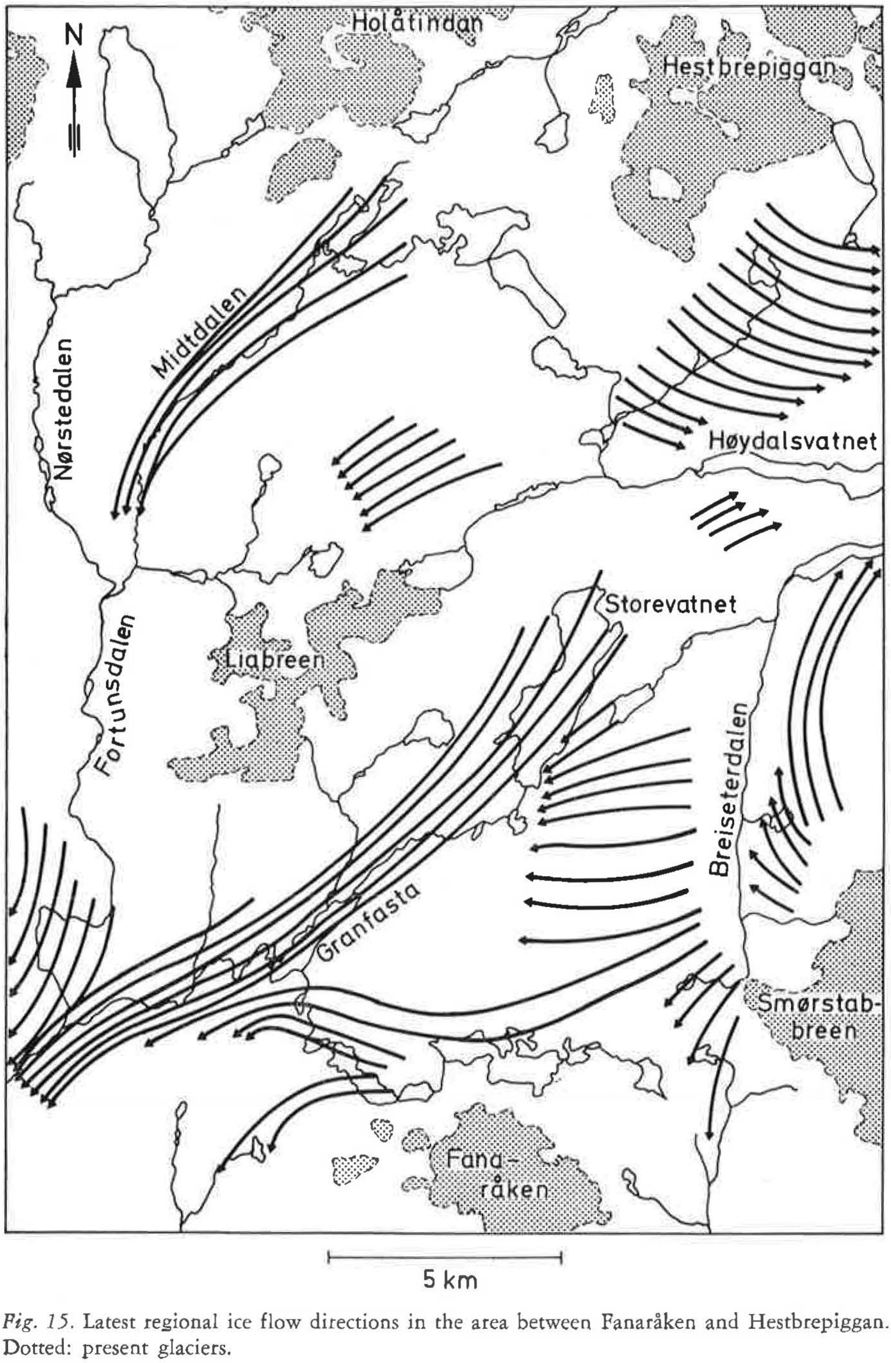

Fig. 15. Latest regional ice flow directions in the area between Fanaråken and Hestbrepiggan. Dotted: present glaciers.

This pattern of glacier movement can generally be considered to mark the final regional movement of active glaciers. As will be demonstrated shortly, the period following the Høgemo Stadial is marked by climatically dead glaciers. This must also have caused a retardation in the dynamic aspects. There were, no doubt, minor movements of the ice until it had almost completely disappeared. Still it seems probable that the indicated movements took place during the last active stages of deglaciation, the Gaupne and Høgemo Stadials. Such an interpretation is strengthened by the nature and regional occurrence of the orientation elements, and the fact that they can be followed to the marginal deposits from this period.

If one accepts that this was the pattern of movement during the Gaupne Stadial, then it is possible to obtain some idea of the glacier's absolute relief at this time. The surface of the ice could not have been much more than 1900 m a.s.l. to the north of Fanaråken, or there would have been ice supplied from here to the Helgedal and Styggedalsbre areas. The contours on the glacier surface were trending almost perfectly north-south, normal to the direction of movement, from Fanaråken northward to the southern end of Storevatnet. This gives an elevation of 1800–1900 m a.s.l., and a somewhat higher elevation at the ice divide north of Storevatnet. These heights have been used in the reconstruction of the ice sheet of the Gaupne Stadial shown in Fig. 17.

Discussion and conclusion

The development of the Gaupne Stadial. – The regional and continuous extent of the marginal deposits, plus clear evidence that some of them were formed during periods of glacial advance, demonstrate conclusively that they are the result of the glacial response to a climatic deterioration. The various glaciers have, however, reacted differently to the climate fluctuations, a phenomenon also observed in Recent glaciers. This can be explained by the relationship between the mean equilibrium line and the steady-state equilibrium line. The *mean equilibrium line* is the average height over a given number of years of the equilibrium line. The *equilibrium line* is the height of the line connecting points were the net balance is zero at the end of a balance year (Anonymous 1969). The *steady-state equilibrium* line is the elevation of the equilibrium line when the total accumulation on a glacier is equal to total ablation. The steady-state line always tends to approach the mean equilibrium line, the rate of this approach being controlled by the size of the glacier, the topography, and other factors.

The histories of the glacier in Tunsbergdal/Leirdal versus that in Jostedal provide an illustration of this. In addition to the several aspects of ablation common to the two, calving must have been important to the glacier in Jostedal. As a result, it was possible for this glacier to make a much quicker adjustment of its steady-state equilibrium line to the climate during the Luster Interstadial. The lowering of the mean equilibrium line during the Gaupne Stadial made it necessary for the glacier in Jostedal to grow in order to re-establish equilibrium.

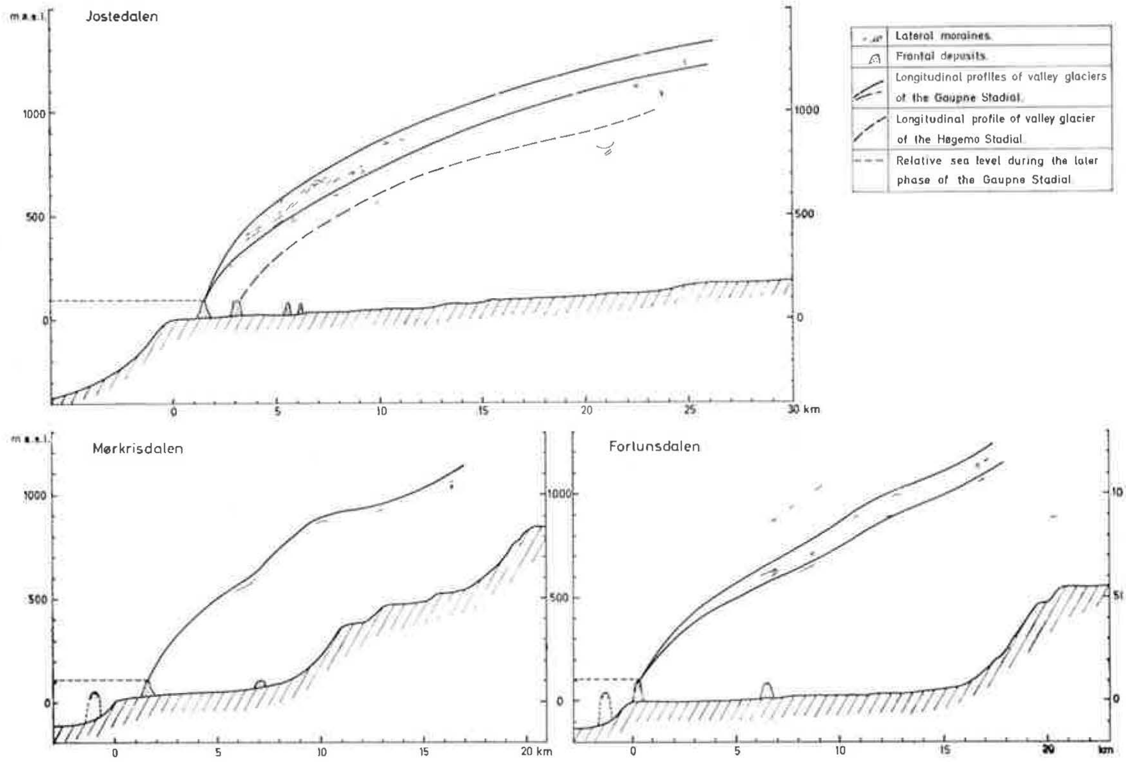

Fig. 16. Longitudinal profiles of the valleys and valley-glaciers in Jostedal, Mørkrisdal and Fortunsdal.

The Tunsbergdal–Leirdal glacier had not had the same possibility for adjustment to the climate of the Luster Interstadial, and therefore it reacted to the climatic deteriorations during Gaupne Stadial with smaller intermittent advances. The resulting marginal features were deposited in a steadily more proximal direction, except perhaps in one case.

Common features, however, also exist in the development of the different glaciers. Where the moraine stratigraphy is most clearly developed, it points to a larger, marked climatic deterioration, the Storhaug Phase, in the middle of the Gaupne Stadial. It is difficult to say exactly how many climatic deteriorations occurred during the Gaupne Stadial, but an indication is given by the greatest number of lateral moraines deposited by a single glacier, i.e., twelve.

Glacier profiles. – The longitudinal profiles of the large valley glaciers are constructed by projecting the marginal moraines onto a vertical plane along the valley axis. Lateral deposits normally lie a little lower than the central axis of the glacier. Taking this into account, the glacier profiles shown in Fig. 16 are obtained, from which the average gradients given in Table 1 have been cal-

Table 1.

Ice-surface gradient in m/km during the Gaupne Stadial in a) Jostedal, b) Mørkrisdal and c) Fortunsdal. d) gives the ice-surface gradient in Jostedal during the Høgemo Stadial. Figures in parentheses are valley-floor gradients. Gradients are given over intervals of 5 km, e.g. the gradient of the Mørkrisdal glacier between 0 and 15 km from its terminus is 68 m/km while the gradient between 5 and 15 km is 52 m/km

a)

Km from terminus	5	10	15	20	25
0	90–114 (4)	70–83 (5)	60–67 (5.2)	52–58 (5.5)	40–50 (6.4)
5		51 (6)	45 (5.8)	39 (6)	35 (7)
10			37 (5.5)	33 (6)	30 (7.3)
15				28 (6.5)	25 (8.3)
20					22 (10)

b)

Km from terminus	5	10	15
0	100 (5)	82 (38)	68 (35)
5		64 (64)	52 (47)
10			40 (30)

c)

Km from terminus	5	10	15
0	86–100 (3)	81–86 (4.5)	63–70 (6)
5		50–62 (6)	51–55 (7.5)
10			50 (9)

d)

Km from terminus	5	10	15	20
0	84 (5)	64 (5.5)	51 (6)	45 (5.8)
5		44 (6)	35 (6.5)	32 (6)
10			26 (7)	26 (6)
15				26 (5)

culated. As is apparent from Fig. 16 and Table 1, the termini of the great valley glaciers of the Gaupne Stadial had rather steep slopes. Up-valley the glacier in Jostedal flattens out a great deal, while the slope of the Mørkrisdal Glacier slightly exceeds that of the bedrock throughout its entire course. The rather poor agreement between the relief of the valley and the relief of the glacier in the case of Fortunsdal must be seen in light of the large ice supply from the Granfasta Valley, approximately 15 km up Fortunsdal. The large degree of control which topography exerts on the location of the termini is clearly seen. The termini stop just before a widening of the valley or a large depression. This phenomenon also helps to explain why the termini had a relatively stable position during the various phases even though there were vertical fluctuations in the proximal part of the glacier, as indicated for the glaciers in Jostedal and Fortunsdal.

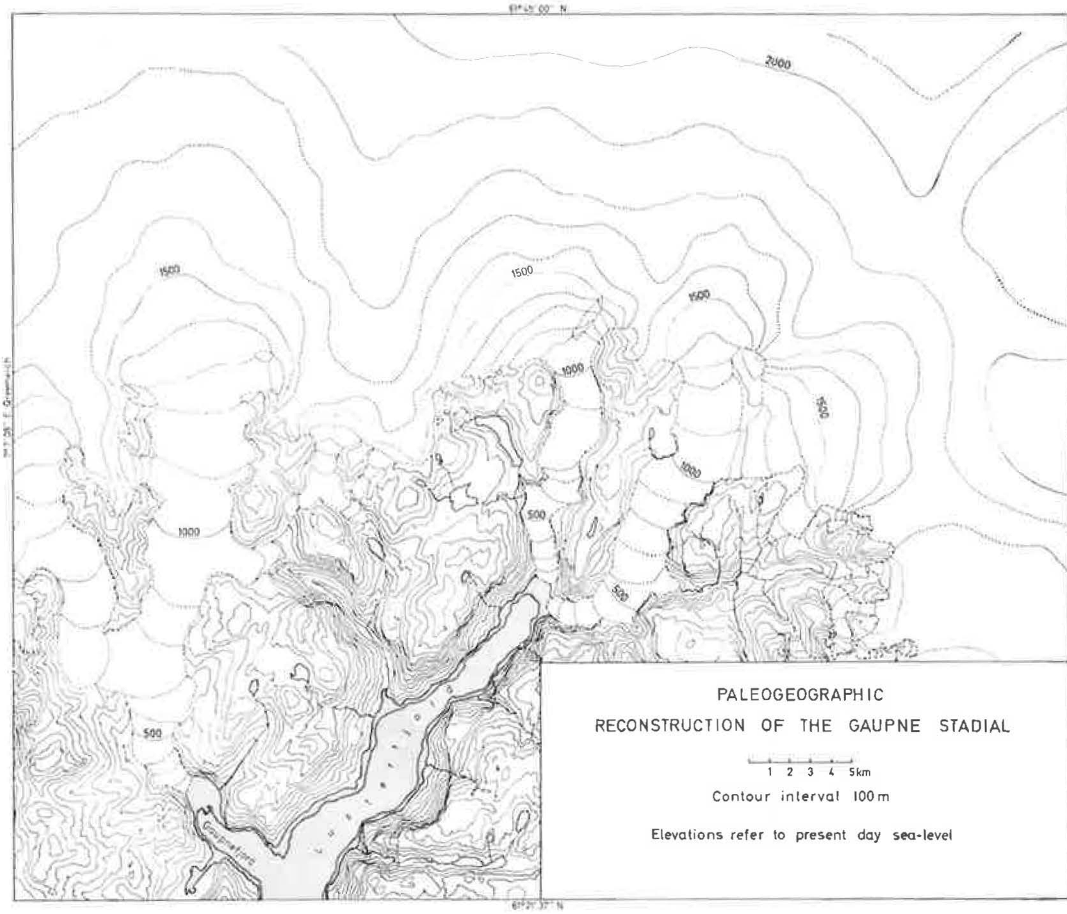

Fig. 17. Paleogeographical reconstruction of the Gaupne Stadial.

It is difficult to make a direct comparison between gradients found here and those which have been obtained for Norwegian Late–Glacial and Early–Holocene fjord and valley glaciers elsewhere (Andersen 1954, 1960, 1968, Anundsen & Simonsen 1968, Fareth 1970, Follestad 1972, Anundsen 1972) because of differences in bedrock relief. It is, however, clear that the gradients found here are among the highest. This may be taken as an indication of relatively large total exchange (Anonymous 1969) of the glaciers' mass.

Reconstruction. – The large number of marginal deposits allow for a complete reconstruction of the areal extent of the glaciers and an approximate reconstruction of their relief (Fig. 17). The margin of the glacier in the mountains between Jostedal and Mørkrisdal is based on the Storhaug Phase marginal deposits; otherwise the most marked deposit, where such can be distinguished, has guided the reconstruction. The contours in the eastern areas of accumulation have been discussed above. Contours on most of the western areas are obtained by extrapolations and probability considerations. The errors in the reconstruction of the areas of ablation is probably rarely in excess of 50 m vertically and 500 m horizontally. The deviation may be a little greater in the area of accumulation, especially with respect to the elevations indicated.

The Høgemo Stadial

After a period of general retreat, the Gaupne–Høgemo Interstadial, there was a pause in the recession, the Høgemo Stadial.

Areal description

The Tunsbergdal–Leirdal area. – In Tverrdal, the southernmost tributary on the western side of Tunsbergdal, a moraine belt extends from 920–940 m a.s.l., in front of the small valley that branches off to the south. The path of the marginal moraines shows that they were deposited by a glacier in Tverrdal connected with the glacier in Tunsbergdal. At the southern end of Tunsbergdal is a remnant of a small ridge, 560 m a.s.l. There is a 1.5 km long lateral moraine at the same elevation on the south-side of Leirdal.

The Jostedal area. – Lateral phenomena from the Høgemo Stadial are found in only two places. In Vigdal, there is a minor remnant of a lateral moraine on the northern side of the valley, at 760 m a.s.l. A terrace found on the proximal side of the Gaupne Stadial deposits, with an elevation of 740 m a.s.l., indicates the existence of an ice-dammed lake beyond the glacier terminus during the Høgemo Stadial. The elevation of the terrace was determined by the elevation of the pass to the south of Storhaug, 739 m a.s.l. (Fig. 7, A). The other locality in which lateral moraines are observed occurs on the western side of Jostedal, in Vassdal. Here a small, high-gradient, diffluent tongue of the glacier occupying Jostedal deposited a series of lateral moraines between 760 and 800 m a.s.l.

Three frontal deposits can be correlated with the lateral moraines described above (Fig. 4). The southernmost of these, the Høgemo frontal delta, affords the most reasonable figures for the gradients on the surface of the ice. The frontal delta lies on the west side of the river, about 1 km inside the Gaupne deposit, and is approximately 400 m long and 200 m wide. Only a few, lower-lying, erosional terraces occur on the eastern side of the valley, indicating that the marginal deposit at one time extended completely across the valley. The Høgemo delta has a gently undulating surface at an elevation of 90 m a.s.l. It is composed of stratified silt, sand and gravel (Fig. 5) with isolated cobbles and boulders. A 1 m thick topset bed with a somewhat higher content of cobbles and boulders, indicates that the terrace was built up to the then existing sea level, which, therefore, must have been ca. 90 m.

The Reidarmoen frontal delta, on the western side of the river two km north of the Høgemo delta, also has lower-lying terraces associated with it on the eastern side of the valley. The terrace dips gently into the valley from an elevation of 82 m a.s.l. at the juncture with the valley wall.

About 1 km to the north of the Reidarmoen delta is the Hausamoen frontal delta; this is approximately 150 m in length and 75 m in width. Here again the terrace dips gently into the valley from an elevation at the inner edge of 78 m a.s.l. Small open sections in both this and the Reidarmoen delta reveal that they are composed of sorted silt, sand and gravel.

The mountains between Jostedal and Mørkrisdal. – Marginal moraines, most often of modest size, are found proximal to the Storhaug Phase's marginal moraines throughout this area. Some of these certainly belong to the later phases of the Gaupne Stadial. Whether or not there are also some which date from the Høgemo Stadial is difficult to say. It is possible that the valley glaciers in this area melted away during the Gaupne/Høgemo Interstadial, so that the margin of the ice during the Høgemo Stadial lay to the north of the Spørteggbreen glacier's present southern margin.

The Mørkrisdal area. – A sandur delta at Meljadn, 5 km up-valley from Bolstad, is thought to be the frontal deposit from the Mørkrisdal glacier during the Høgemo Stadial. The sandur delta is about 500 m long and 50–100 m wide, and lies along the inner side of the sharp curve in the valley. The surface rises from 95 to 102 m in a proximal direction (Brathole 1951), indicating a relative sea level of 95–97 m at the time of its formation (Andersen 1960, p. 88). The difference between the Gaupne and Høgemo Stadials' sea level is of the same order of magnitude in Jostedal as in Mørkrisdal, thus affirming the correlations.

Lateral moraines deposited by the Mørkrisdal glacier during the Høgemo Stadial are not found, but indications of the relief of this valley glacier are found at two locations. After the surface of the Mørkrisdal glacier had sunk below the mouth of Bolstad–Sveigdal, 980 m a.s.l., a glacier moved down this valley and deposited two frontal moraines, presumably during the Høgemo Stadial. The surface of the Mørkrisdal glacier must therefore have been below the mouth of Bolstad–Sveigdal. Striae east of the Spørteggbre glacier indicate that an ice lobe flowed eastwards from the Spørtegg Plateau during the final phases of deglaciation. An ill-defined ridge of boulder-rich till at the east end of Grånosi, 1240–1200 m a.s.l., probably represents a lateral moraine formed along the northern flank of the lobe. If this was deposited during the Høgemo Stadial, then the surface of the Mørkrisdal glacier here at that time must have been lower than 1200 m.

The Fortunsdal area. – As is the case in Mørkrisdal, the Høgemo Stadial frontal deposits in Fortunsdal are found where the valley first narrows about 6 km upstream at Fortun (Fig. 11). The deposit lies on the eastern side of the valley and is about 300 × 300 m. Small exposures reveal sorted material of sand- to boulder-size. The proximal slope is a secondary feature produced by fluvial erosion. Talus covers the distal slope and part of the top of the deposit. In the north the top is terminated by a small terrace 87 m a.s.l. rising in a southerly direction to about 100 m a.s.l. Because of the talus material on the top, it is difficult to identify the upper limit of the glaciofluvial material and thereby the shore level at the time of deposition. The correlation of the Fortun frontal deposit with the Høgemo Stadial is thus based solely on morphostratigraphy.

Lateral moraines of probable Høgemo Stadial origin are found only on one spur, 20 km above the mouth of the valley: three parallel ridges, 3–4 m high, are around a small crevice 890 m a.s.l.

The mountain area east of Fortunsdal. – Ice supply from the Granfasta Valley to Fortunsdal also existed during the Høgemo Stadial. A few small remnants of lateral moraines in the outer part of the Granfasta Valley show that the surface of the ice was ca. 150–200 m lower then than during the Gaupne Stadial. A 7 m high lateral moraine-ridge found between 1230 and 1260 m a.s.l. near the river entering Skålvatn from the south indicates that the supply from the east to Granfasta Valley had ceased by this time.

Some of the small lateral moraines on the proximal side of the Gaupne Stadial moraine belt in Rya Valley may date from the Høgemo Stadial. As mentioned earlier, it is also possible that some of the most proximal pre-sub-Recent moraines in the Styggedalsbre area and Skagastølsdal were deposited during the Høgemo Stadial.

Discussion and conclusions

The correlations of the various marginal deposits assumed to belong to the Høgemo Stadial are based mainly on morphostratigraphic factors. That these are valid may be shown by the good agreement between local correlations made on the basis of gradients, as on the glacier in Jostedal, and sea levels, as between Jostedal and Mørkrisdal.

Although the representation is scarce, the marginal deposits have a regional extent which cannot be attributed to anything other than the common effects exerted by the climate on the glaciers of this area. The frontal deposits from the Høgemo Stadial, in contrast to those of the Gaupne Stadial, seem to consist exclusively of glaciofluvial material. This indicates that the Høgemo Stadial is a stagnation phase, whereas the Gaupne Stadial was a period of active advance. The difference is probably due to the fact that the climate was less depressed in the Høgemo Stadial than in the Gaupne Stadial.

It has only been possible to reconstruct a longitudinal profile for the glacier in the Jostedal (Fig. 16, Table 1). The gentler frontal slope indicates that the glacier's total exchange was smaller during the Høgemo Stadial than during the Gaupne Stadial. The frontal deposits proximal to the Høgemo deposit have no parallel in the other valleys. They probably represent local events in the final deglaciation that were the result of local glaciological and topographical conditions in Jostedal.

The marginal deposits from the Høgemo Stadial are too sparse to allow a full reconstruction of the glaciers during this period.

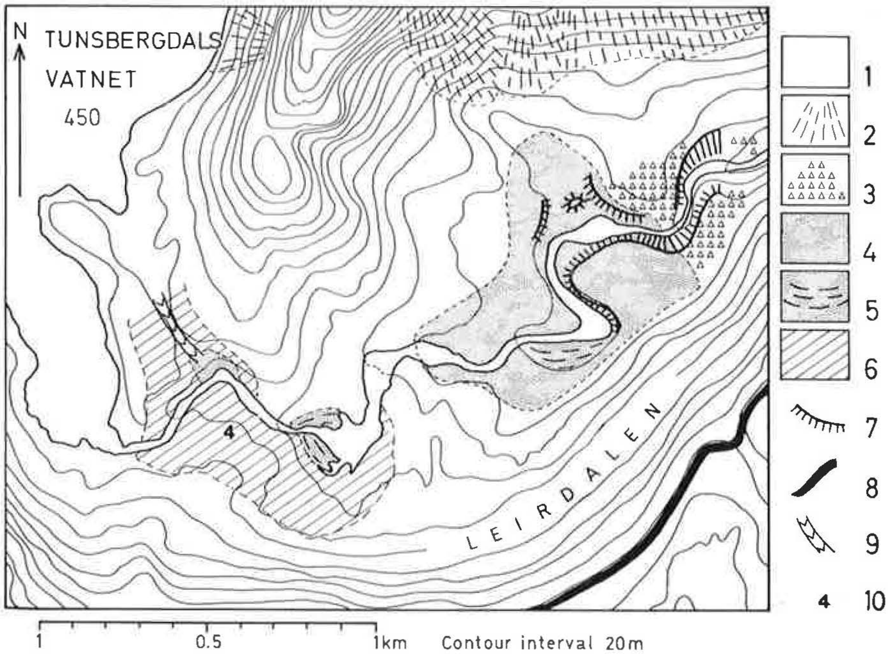

Fig. 18. Glacial geology sketch map of Leirdal. 1: Discontinuous till cover, 2: Talus, 3: Lodgement till, 4: Moraine-dammed lacustrine sediments, 5: Recent fluvial sediments, 6: Glacio-lacustrine sediments, 7: Terrace, 8: Lateral moraine, 9: Meltwater channel, 10: Sediment sample 4 in Fig. 5.

The final deglaciation

Areal description

The Tunsbergdal–Leirdal area. – The sediments on the valley floor of Leirdal can be divided into three units (Fig. 18). Farthest up the valley, east of Tunsbergdalsvatn, are glacio-lacustrine fine-sands (Fig. 5) with isolated cobbles, grading into laminated silt and clay to the east. The upper boundary of this unit lies approximately 470 m a.s.l. To the east it is bounded by a bedrock threshold. In the center of the valley are stratified sediments, mainly sand, overlying lodgement till. The beds dip in varying directions but mostly down-valley. The third unit, a lodgement till, is exposed to the east. On the southern side of the river, the till is terraced at about the same level as the sediments in the center of the valley, 370 m a.s.l.

The sequence of sediments is interpreted as follows: the glacier underwent downwasting after the Høgemo Stadial, the area to the east of Tunsbergdalvatn emerging first. An ice-dammed lake formed, with a minimum surface elevation of 470 m a.s.l. Fine sand and silt/clay were deposited in the shallower and deepers parts of the lake, respectively. Boulders were frequently dropped from icebergs. After the glacier had receded from the central part of Leirdal, huge masses of sediment were transported to this area. These were deposited in a

Fig. 19. To the left the Early Holocene delta, and to the right the sub-Recent/Recent fan of the Leirdøla river (L). J is Jostedalen river running from right to left.

basin created by a barrier of till that had accumulated as the result of previous cross-valley ice movements (Fig. 2) and, later, converging glaciers from Tunsbergdal and Jostedal. Drainage from this sedimentation basin caused a terracing of till and later eroded completely through it.

The Jostedal area. – After the ice retreated from Hausamoen, the sea filled the lower portions of Jostedal, and smaller marine terraces were formed. At Ytamoen, 300–400 m north of Hausamoen (Fig. 4), a terrace is present at 75 m a.s.l.; this is composed of stratified sand and silt dipping up-valley at an angle of ca. 20°. This deposit is clearly the northern distal part of a delta constructed by the Rydøla River, a tributary from the east.

The Leirdøla River also deposited a delta in the postglacial 'Jostedal Fjord' (Fig. 19). The surface of this lies between 71 and 75 m a.s.l. (Brathole 1951, p. 20). Just north of the Leirdøla's Recent river fan is a terrace 75 m a.s.l. Open-sections in the terrace reveal, lowest, layers of cobbles and boulders in a southward-dipping matrix of sand and silt. These lowest sediments are probably glaciofluvial material deposited by the glacier occupying Jostedal. The upper 2 m, consisting of horizontal layers of silt and fine-grained sand, are the distal sediments deposited by the Leirdøla River in the narrow Jostedal Fjord.

At Alsmo, a small terrace, with an elevation of 74 m a.s.l., is mainly built up of sand, with some gravel and cobbles.

As all of the highest terraces upstream of Hausamoen lie 74–75 m a.s.l., it is most probable that one sea level controlled their elevation. This indicates a rapid recession of the terminus. At its greatest postglacial extent, the sea in Jostedal reached north almost to where the Vigdøla tributary joins Jostedal River.

The Fortunsdal area. – A small 10–15 m wide abrasion surface was cut into the proximal side of the Fortun frontal deposit after the ice had retreated. The elevation of the surface is 87 m a.s.l. On the opposite side of the valley, about two hundred meters farther north, is a bedrock spur. A thick gullied cover of lodgement till north of the spur is terminated at the top by a terrace, also at 87 m a.s.l. Approximately 1.5 km farther north, on the west side of the valley, there is yet another terrace, elevation 85 m a.s.l. The highest traces of postglacial marine activity upstream of the Fortun deposit all have elevations around 85–87 m, which probably represents the marine limit here. This indicates rapid retreat of the glacier terminus with the sea reaching its maximum extent just north of where the Granfasta River joins Fortunsdal.

Tributary valleys and mountain areas. – Glacial deposits are scarce in the mountainous areas. Some areas, such as the floor of Granfasta Valley, are completely bare. Numerous pot-holes and other plastic forms attest to intense sub-glacial drainage (Dahl 1965).

In other places, such as Krondal, Rausdal, Greindal, Nørstedal, Midtdal, Høydal and Breiseterdal, the terrain is spotted by mounds and ridges, some consisting of till and some composed of glaciofluvial material. The latter has accumulated locally as terraces and small eskers. Some of the moraines have previously been considered frontal moraines (Holmström 1880, Rekstad 1914). In my opinion they were formed in connection with stagnating glaciers, and must be regarded as ablation till and, in a few places, as erosional remnants of lodgement till. That the deposits extend all the way to the areas of recent glaciers indicates that these, too, were partially or completely deglaciated at this time.

Conclusion

Marine levels in Jostedal and Fortunsdal attest to rapid retreat of the ice front. This probably occurred when the ice had been reduced by downwasting to such a thickness that a quick recession by calving could take place. The glacial sediments in the tributary valleys and mountain areas were deposited in connection with stagnating glaciers. The erosional forms are of glaciofluvial origin and were mainly produced subglacially. The overall picture gives evidence of a climatic amelioration following the Høgemo Stadial.

The deglaciation: age and climate

Equilibrium line displacements

Equilibrium line displacements refer to how much lower (–) or higher (+) former equilibrium lines lay in relation to the present line. As the equilibrium line displacement is used to indicate the difference between climates of the past and present, it is important to use the mean equilibrium line as an expression of the present climate (p. 23). The average height of this line over the last thirty-year (meteorological) period is, in my opinion, best suited as a reference value or datum. Accurate isohypses for the mean equilibrium line in this area can be obtained by using Liestøl's (1967) map of steady-state equilibrium isohypses together with his data from the Storbre Glacier to the east. For this glacier Liestøl has determined an elevation of 1785 m a.s.l. for the mean equilibrium line in the period 1931–60, while the steady-state equilibrium line lies at 1690 m a.s.l., a difference of 95 m. For the specific area in question then, no significant errors will result from placing the mean equilibrium isohypses in conformity to, and 100 m higher than, the steady-state equilibrium isohypses. The mean equilibrium isohypses thus obtained vary in elevation from 1600 m a.s.l. in the southwest to 1800 m a.s.l. in the northeast.

The highest lateral moraines give a minimum altitude for the equilibrium line during previous kryomers (e.g. Andersen 1968). The highest clearly non-topographically controlled lateral moraines from the Gaupne Stadial are those west of Fanaråken and in Styggedal. These give an equilibrium line displacement of –260 and –330 when the relative sea level is subtracted. These figures agree very well with the results obtained by comparing the three glaciers, Hestbreen, Ringsbreen, and the cirque glacier west of Fanaråken, which did not extend beyond the A.D. 1750 advance limit during the Gaupne Stadial. The equilibrium line displacement during the 1750 advance was ca. –250 m (Liestøl in O. Holtedahl 1960, Fig. 485), which means that it is rather unlikely that the Gaupne Stadial displacement exceeded –250 m in addition to the approximate 100 m shore level displacement. The equilibrium line displacement during Gaupne time was, accordingly, –300 ± 50 m.

It is difficult to give exact figures for the equilibrium line displacements during the other periods of deglaciation, because there are so few marginal phenomena. The far-advanced deglaciation of the above-mentioned local glaciers indicates, however, that the displacement has not been lower than –200 m in the Luster Interstadial. Furthermore, the character of the Høgemo Stadial makes it probable that the displacement during this period was less negative than during the Gaupne Stadial. Something between –300 m and –100 m seems probable. The dead-ice phenomena from the final deglaciation found up to presently glaciated areas indicate that the mean equilibrium line during the end of this period was at least as high as it is today.

Equilibrium line displacements from Younger Dryas time vary between –350 and –575 m in different parts of Norway as at Central Rogaland: –525 ± 50

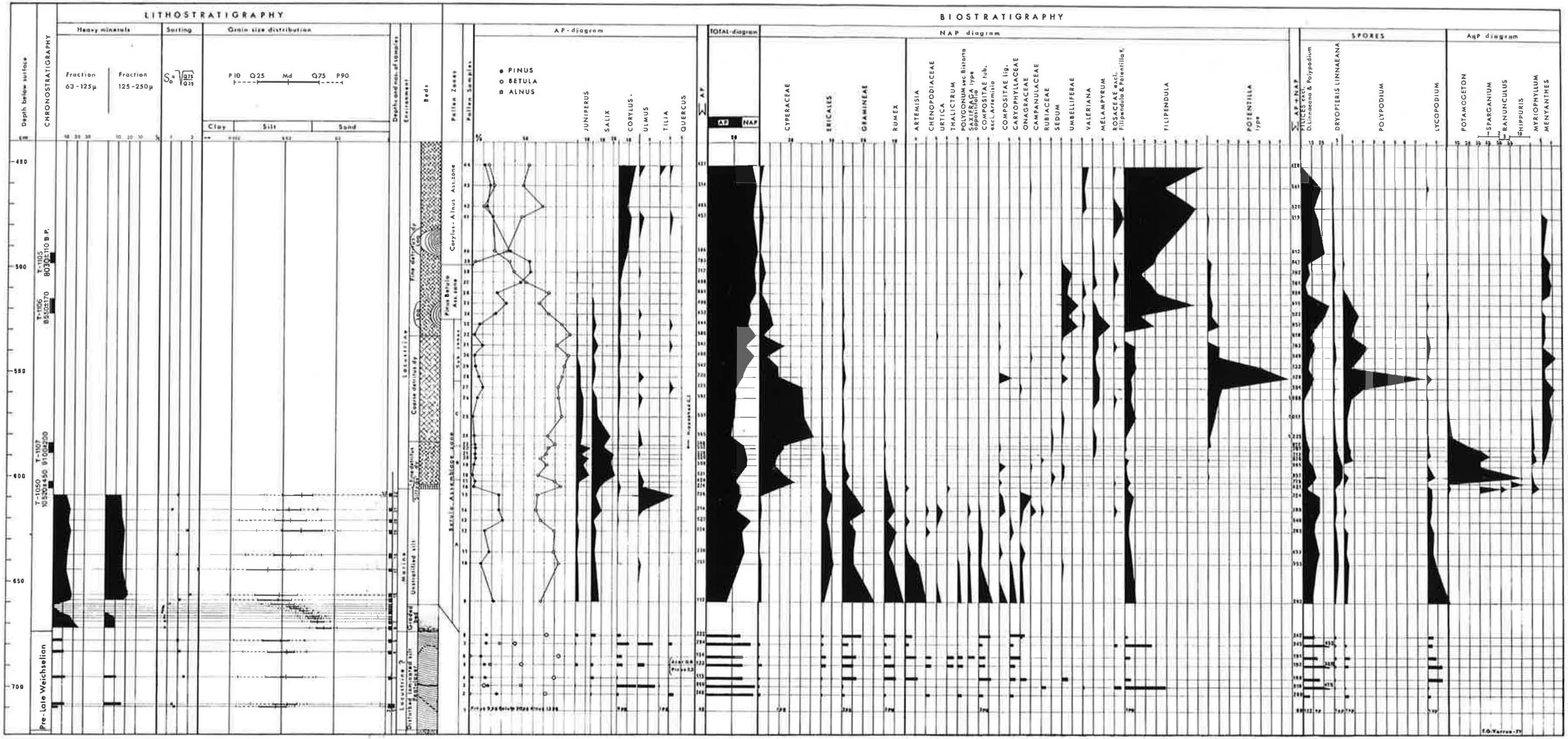

Fig. 20. Stratigraphy and pollen diagram from the basin at Kroken.

(Andersen 1968), North Rogaland and Sunnhordland: –375 ± 25 (Anundsen 1972), Folgefonn Peninsula: –400 ± 50 (Follestad 1972), Nordfjord: –400 (Fareth 1970), and Troms: –475 ± 50 (Andersen 1968). Displacements for Early Holocene glacial events lie between –150 and –400 as at North Rogaland and Sunnhordland: –375 ± 25 and –350 (Anundsen 1972), Inner Hardanger: –200 (Liestøl 1963), –350 (Anundsen & Simonsen 1968), Aurland area in Inner Sogn: –300 ± 25 (Bergstrøm 1971), and Troms: –200 ± 50 (Andersen 1968).

Not all of these results can be directly compared with the present ones: with the exception of Liestøl's (1963) and Anundsen & Simonsen's (1968) results, it seems that the reference value for the calculations has been the steady-state equilibrium line. As glaciers in Norway have generally been receding in the past few decades, this line will be lower than the mean equilibrium line (e.g., Storbreen, where there is a difference of 95 m). As a result, the negative displacement should probably be larger than indicated in the other calculations given above.

On the basis of this type of comparative study of equilibrium line displacements it can be concluded that the Gaupne and Høgemo Stadials are probably not of Younger Dryas age; an Early Holocene age is more probable.

Shore-line positions

Only the highest marine levels have been investigated (Fig. 4). Especially striking is the large shore-level displacement, ca. 50 m, which has occurred between the time the ice calved in Lusterfjord (M.L. ca. 130 m) and the time it disappeared from the outer parts of the large valleys (M.L. 75 m in the west, 85 m in the east). It is well known that shore-level displacements occur very quickly during periods immediately following deglaciation, and, consequently, this displacement may not represent such a long period of time. Examples of this rapid displacement can be taken from several places, e.g. in Inner and Outer Oslofjord the displacement has been found to be 11 and 9 m/century, respectively (Hafsten 1956, Danielsen 1970); in Southern Sweden it was ca. 5 m/century for the first 1500 years after deglaciation (Mörner 1969, Fig. 143); in Spitsbergen it was ca. 3 m/century (Hoppe 1971). The many unknown factors involved make direct comparisons with other areas impossible, but a shore-level displacement of about 5 m/century or more during the earliest centuries after deglaciation does not seem unreasonable.

With regard to the shore-level displacement during the various deglaciation periods it must be kept in mind that the marine levels indicated by the frontal deposits stem from the later phases of deposition. Thus, when the stadials began, the shore-levels were probably higher than the level marked by the frontal deposits. The displacements are distributed throughout intervals of ca. 25 m from the middle of the Luster Interstadial to the end of the Gaupne Stadial ca. 10 m from the Gaupne to the end of the Høgemo Stadial; and ca. 15 m from the Høgemo Stadial to a point in the period of final deglaciation.

Great uncertainties surround the dating of the deglaciation obtained from shore-line diagrams for the Sognefjord, owing to both the uncertainty involved with the applied N-S isobase direction and the paucity of dates for the supposedly correlated levels. The level which Carlsson (Nydal et al. 1964) believes to belong to Younger Dryas time is found by extrapolation to lie 150–160 m a.s.l. in Inner Lusterfjord (Brathole 1951, Pl. XI). Surprisingly, roughly the same result is obtained upon extrapolation of Fareth's (1970) Younger Dryas isobases from Nordfjord, which give an isobase direction of N 30° E in Luster. Part of the reason for this may be that there is an area just north of Sognefjord where the isobase direction gradually shifts, becoming more N-S farther south. In any event, both of the above results indicate that the deglaciation of the investigated area is younger than Younger Dryas.

Pollen analysis

To obtain a more exact date for the deglaciation, pollen analysis and radiocarbon dating were performed on material from a now regrown basin at Kroken (Figs. 4 and 20). A paper dealing with the bottom sediments from the basin includes a description of the field and laboratory methods, diagram construction and interpretation of the genesis of the minerogenic sediments (Vorren 1972). Some of these topics, however, will be expanded here.

Environment. – The basin threshold is 124 m a.s.l., i.e., below the marine limit, which here is ca. 130 m a.s.l. The finding of Hystrix and marine diatoms proves that the minerogenic sediments were deposited in a marine environment. The marine benthonic, euryhaline *Navicula digitoradiata* was dominant 2 cm below the silty dy. Other forms observed here were: *Amphora* cf. *ovalis*, *A. ovalis* var. *lybica*, *Diploneis interrupta*, *Gomphonema intricatum*, *Nitzschia* spp., *Rhopalodia gibba*, *R. musculus*, *Pinnularia mesolepta*, *Synedra ulne* and *Tabellaria flocculosa.* The following diatoms were observed in the silty dy: *Cymbella ventricosa*, *C. aspera*, *Epithemia zebra*, *Fragilaria pinnata*, *F. construens*, *F. brevistriata*, *Stauronesis phoenicentron*, *Tabellaria fenestrata*, and *T. flocculosa.* The *Fragilaria* species dominated on the whole, especially *F. spinnata.* This, together with a tremendous bloom of *Pediastrum,* indicates that eutrophic conditions obtained. Two cm above the silty dy, considerably less *Pediastrum* was found along with the following diatoms: *Cymbella turgida*, *C. gracilis*, *Eunotia diodon*, *Fragilaria* spp., *Frustulia rhomboides*, *Gomphonema gracile*, *G. longiceps*, *Nitzschia fonticola*, *Pinnularia interrupta*, *P. mesolepta*, *P. viridis*, *Stauroneis phoenicentron*, *Tabellaria fenestrata* and *T. flocculosa.* Less eutrophic conditions are thus indicated. The transition from salt to fresh water seems to have been rapid, indicating rapid shore-level displacement. Moreover, there was a relatively rapid transition from a more to a less eutrophic environment in the beginning of the lacustrine facies.

Vegetation; description and interpretation. – Subzone A in the *Betula* Assemblage zone, is characterized by high contents of Gramineae, Ericales and *Rumex*. The *Corylus*, *Ulmus* and *Pinus* pollen in these spectra must be due to long-distance transport, and the relatively high *Pinus* frequencies are probably due to marine over-representation (Florin 1945). The lowest spectrum in Subzone A represents a very early phase in the plant-colonization of the area, as it is taken from directly above the graded bedding which was deposited just after the deglaciation of the basin (Vorren 1972). A typical pioneer flora is reflected here by Gramineae, *Rumex*, *Artemisia*, *Polygonum* and *Saxifraga*. Pollen frequencies of the most typical pioneers, however, soon decrease. An idea of the time span involved is obtained by noting that if a shore-level displacement of a magnitude about 5 m/century is accepted, the marine silt would be deposited within a hundred years or so. After the basin became isolated there was a marked increase in Cyperaceae. Near spectrum 16, K.-D. Vorren has found, by size analysis, that 36% of the *Betula* pollen is *B. nana*, and the rest is *B. alba* coll. Tree birch has, therefore, probably grown here at least in the later part of the time interval represented by Subzone A.

Subzone B is characterized by high contents of *Salix* and *Juniperus*. Seeds and catkin scales of *Betula alba* coll. found here show that tree birch grew in the immediate vicinity of the basin. Among the hydrophytes, *Potamogeton natans* (seed-determined) is completely dominant. At present *P. natans* normally grows up to the pine limit, but also in the birch region (Hultén 1970). Accordingly, the vegetation in this subzone parallels that in the lower sub-alpine belt, or lower.

Subzone C is characterized by a very high Cyperaceae content. Macrofossil analysis of a sample between spectra 25 and 26 showed *Carex rostrata* seeds, *Betula alba* coll. seeds, *Sphagnum teres* and *Drephanocladus exannulatus*. This plus the decrease in *Potamogeton* point clearly towards a partial regrowth of the basin. *Salix* undergoes a marked decline after a small increase at the beginning of this subzone. In conjunction with the decline of *Juniperus*, this can be interpreted as the result of an increased density of birch forest.

As the isolation contact represents a relative sea level of 124 m, and the Gaupne and Høgemo Stadials represent lower sea level, it is clear that the sediments corresponding to these stadials must lie above the marine/lacustrine contact. In my opinion, the vegetation in Subzone B is evidence of a colder and more humid period corresponding to the Gaupne and Høgemo Stadials. The contemporaneous decline of Cyperaceae and rise of *Potamogeton* are, in my opinion, due to a higher water level as a result of more humid conditions, causing transgression of the shallower portions of the basin. The apparent increase of AP in Subzone B is not an expression of the then prevailing conditions; it primarily reflects the decrease in Cyperaceae. The high content of *Salix* and the heliophile *Juniperus* indicates a forest of rather low density. The rise of *Juniperus*, together with the decrease in Ericales from Subzone A indicate a heavier snow cover (Iversen 1954) during the time represented by Subzone B than during the time represented by Subzone A. The increase in *Salix* also indicates more humid conditions during the Subzone B time.

Concerning the other parts of the diagram attention is drawn to the date of 8500 ± 170 B.P. for the rise of *Pinus*. The *Pinus* maximum coincides with a decline in *Filipendula* and probably represents a dry climatic period. The rational and empirical limit for *Alnus* is dated at 830 ± 110 B.P. The rise of *Corylus* is more or less contemporaneous with this. Five *Alnus* seeds were found around spectra 42–43, of which one was surely from *A. glutinosa* and four were from *A.* cf. *glutinosa*. The rise of *Alnus* is thus probably represented by *A. glutinosa*.

Discussion and conclusion

Age. – The deposit in the basin at Kroken, representing Subzone B in the pollen diagram and correlated with the Gaupne and Høgemo Stadials has been found to have a radiocarbon age of between 10520 ± 450 (T–1050) and 9100 ± 200 (T–1107 years B.P. There are several indications that the former age is too high: the Younger–Dryas/Preboreal boundary is considered by most to fall between 10300 and 10000 years B.P. (Mangerud 1970. Fig. 11). If the 10520-year figure is correct, one would expect to find evidence in the pollen composition in Subzone B. There is, however, no such evidence. Pollen spectra of Younger Dryas age from other parts of South Norway (Fægri 1936, 1940, 1944, Hafsten 1963, Chanda 1965, Mangerud 1970) show much larger NAP values, and tree birch has, at most, existed only in especially favorable places. These spectra are from areas with pronounced maritime climates. The same conditions, however, seem to have prevailed in Blekinge in South Sweden (Berglund 1966), which more closely resembles Inner Sogn in the maritime character of the climate. Thus the vegetation in Subzone B appears to be too thermophile to be of Younger Dryas age. Glaciological considerations also indicate that 10520 years is too high a figure. The first pronounced marginal event proximal of Kroken, the marginal event of the Gaupne Stadial, should in this case be of Late Younger Dryas age. As has been shown above, neither the equilibrium line displacements nor the shoreline positions agree with such an interpretation. There is, however, no reason to believe that the T–1050 dating itself is erroneous. The standard deviation here is so large that just one additional standard deviation will give an age of ca. 10000 years, which is considered more probable. Figure 21 shows the C^{14} chronology for deglaciation which this implies.

Climate. – On the basis of the equilibrium line displacements, the following conclusions can be drawn. Assuming a vertical temperature gradient of 0.65°C/100 m (Andersen 1968) and the same precipitation pattern as we have today, the ablation season or summer temperature during the Gaupne Stadial must have been 1.6 – 2.1° lower than at present (1931–60). During the Luster Interstadial the summer temperature was just a little lower or perhaps the

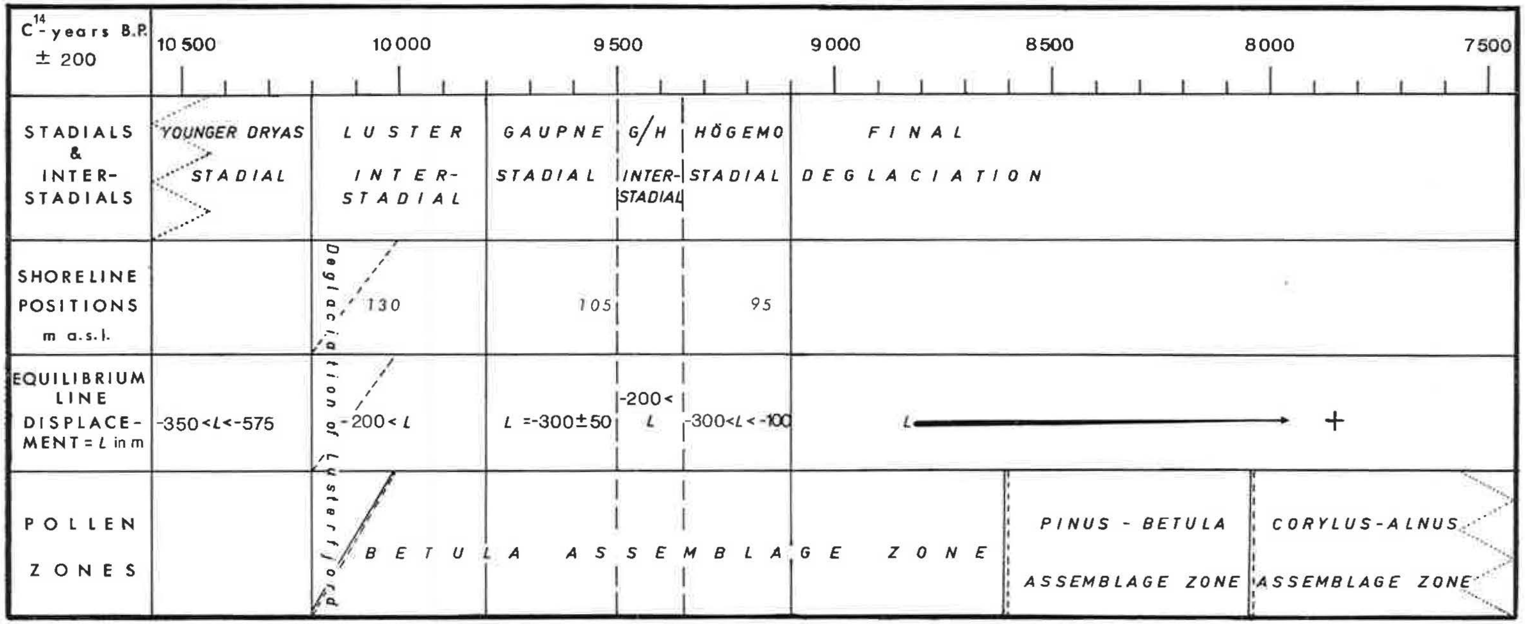

Fig. 21. Correlation chart. The Younger Dryas equilibrium line displacement is based on various authors' results elsewhere in Norway (see text). The shoreline positions refer to Inner Lusterfjord and Mørkrisdalen. The pollen zones are based on data given in Fig. 20.

same as it is today. The Høgemo Stadial summer temperature was between the Gaupne Stadial summer temperature and present summer temperatures, while the summer temperature during parts of the final deglaciation was probably higher than at present. It is, however, unlikely that the winter precipitation during the various stadial and interstadial periods was the same as it is at present. The vegetation as reflected by the pollen rain indicates higher precipitation during the Gaupne and Høgemo Stadials than during the Luster Interstadial. This implies a smaller reciprocal temperature difference during the summer season.

Correlations. – Many more or less well-dated phenomena indicate the occurrence of one or more kryomers in the time interval which the Gaupne and Høgemo Stadials cover.

Glacial marginal events of similar age are observed at many places in Norway. In the Oslo area the Ås–Ski event, the Aker event and the Romerike event are known. Dating of sediments correlated with the Ås–Ski event yields an age of 9750 ± 250 B.P., while 9450 ± 250, 9250 ± 250 and 9850 ± 350 B.P. have been obtained for the Aker event (O. Holtedahl 1960). In central Rogaland the Trollgaren event, which is younger than Younger Dryas (Andersen 1954), is known. Anundsen (1972) correlates this with the Eidfjord–Osa event in Hardanger (Liestøl 1962, Anundsen & Simonsen 1968). The younger portions of the latter must be less than 9680 ± 90 B.P. (Rye 1970). In Northern Rogaland and Sunnhordland, Anundsen (1972) has found a marginal event, the Blåfjell event, which is younger than the Trollgaren event, perhaps of Boreal age. On the Folgefonn Peninsula, Follestad (1972) reports a marginal event which he thinks is of Preboreal age. There are three marginal events younger than Younger Dryas in Aurland, Inner Sogn. The outermost of these is older than 9790 ± 160 years, while the two innermost are probably younger (Bergstrøm 1971). In Nordfjord, Fareth (1970) has found two marginal events younger than Younger Dryas which are both older than 9420 ± 200 years. Andersen (1968) has found at least three Early Holocene marginal events in Troms with ages of 9880 ± 240, 9700–9400 and 9300–9000 B.P. Marginal events of similar age are also observed in other parts of the world: the Alps (Heuberger 1968), Alaska (Goldthwait 1966), Ontario (Zoltai 1967) and Baffin Island (Andrews et al. 1970). Pollen analyses also indicate climatic depressions in the time interval under discussion, among others 'Piottino–Schwankung' ca. 10050–9650 in Switzerland (Zoller 1960), in Friesland (Behre 1966, 1967) and in Denmark (Iversen 1967). Oxygen–isotope analyses from the Greenland Ice-sheet (Dansgaard et al. 1970) indicate a relatively unfavorable climate 9600–9000 years ago. Mörner's (1969) investigations of eustatic sea-level changes point towards a climate depression around 9700 years ago and a relatively unfavorable period up until 9280 years B.P. The climate depressions thus demonstrated for the Early Holocene have a wide regional extent. Most of the dates for the depressions fall between 9800 and 9200 years B.P. with, however, a concentration in the first half of this interval.

At the present time, it is difficult to determine how many climate deteriorations have occurred in this time interval. In Norway even one of the most sensitive climate detectors, the glaciers, give different results, depending on which area is considered. This can be the result of many factors, but glacial and topographical conditions are certainly of importance. A period of unfavorable climate can thus cause one marginal event in one area but two or more in other areas (compare the Ra, Central Swedish and Salpausselkä moraines). Regional climate variations cannot be ignored either.

The above problems notwithstanding, it is possible to correlate the Luster Interstadial with the thermomer after Younger Dryas, the so-called Friesland Interstadial (Behre 1966, 1967). Behre gives the age of this thermomer as 10250–10050 B.P., while Mörner (1969) dates it as 10000–9750 B.P. The Gaupne Stadial is very likely correlated with the Eidfjord–Osa event, around 9700 years B.P., while the Høgemo Stadial represents a climate depression a little more than 9100 ± 200 years ago.

Some trends in the deglaciation of Central West Norway

Figure 22 has been prepared to place the present investigations into a somewhat larger regional perspective and to illustrate some deglaciation problems of Central West Norway.

Younger Dryas Stadial. – The reconstruction of the glaciers' maximum extent during this period is based on Mangerud (1970) and Follestad (1972). The ice-sheet in the southern parts of the Bergen area and the Hardangerfjord first reached the indicated position late in this period (Holtedahl 1964, 1967, Mangerud 1970). Fareth (1970) reports that the same is true in the Nordfjord area. The position of the marginal zone in the Sognefjord area is uncertain. Undås (1963) thinks the marginal zone extends across the mouth of Sognefjord, while Carlsson (Nydal et al. 1964) thinks that frontal deltas at the mouths of small tributary valleys in the outer and central portions of Sognefjord mark its extent during this period.

The period from Younger Dryas to 9800 ± 200 B.P. – A marked glacial recession characterizes this period. The recession was especially large in the fjords, where calving played an important role in ablation. Accordingly, the entire Hardangerfjord was deglaciated (Holtedahl 1967, Anundsen & Simonsen 1968), along with the portions of the Sognefjord and its tributaries which were not already ice-free.

The period from 9800 ± 200 to 9100 ± 200 B.P. – Several kryomers occurred in this interval during which the glacial terminus advanced or stagnated. The recessions during the intervening thermomers were most often of a relatively modest extent because of decreased possibilities for calving. As a result, the

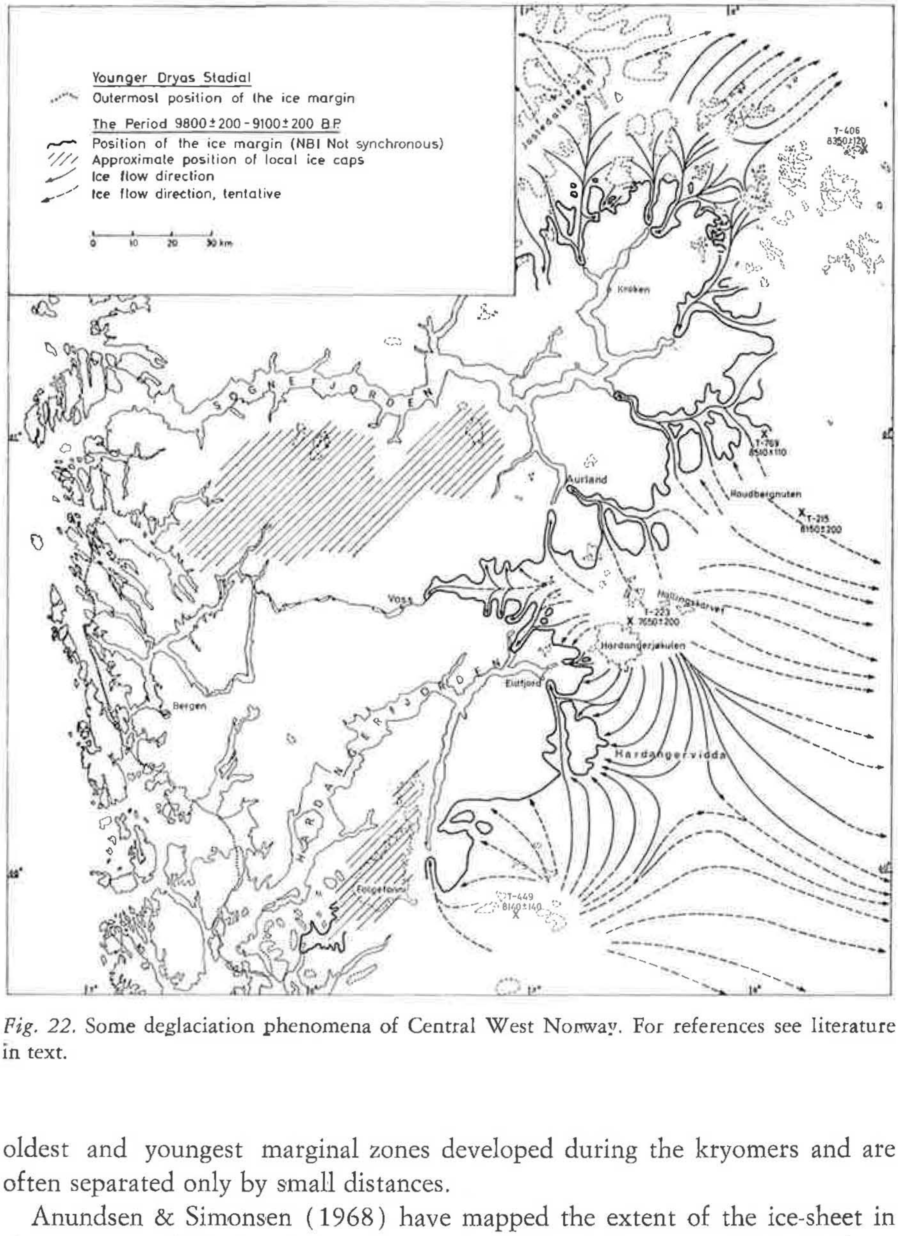

Fig. 22. Some deglaciation phenomena of Central West Norway. For references see literature in text.

oldest and youngest marginal zones developed during the kryomers and are often separated only by small distances.

Anundsen & Simonsen (1968) have mapped the extent of the ice-sheet in the area around Hardangerfjord during the Eidfjord–Osa event. Basically from studies of aerial photographs they have also reconstructed the marginal zone as far north as Sognefjord. The marginal deposits upon which the reconstruction in the latter area is based are not all synchronous. Bergstrøm (1971) has shown for the Aurland area that they actually come from three marginal zones. The distance between the zones, however, is small, so that the general picture given in Fig. 25 is not changed. Possibly a more serious error concerns the valley glacier they show flowing towards Voss (see below).

The reconstruction of the ice movement outside the area dealt with in this paper is based on Tollan's (1963) observations in the northernmost area. In the area between Hallingskarvet and Raubergskarvet, Liestøl (1963) has observed a late ice divide which here is correlated with this time interval. The ice movement displayed in the Hardangervidda area is based on personal observations (see also Liestøl 1963, Fig. 3). Otherwise the reconstruction is tentative. It is clear, however, from the existing observational data that the ice culminations during this time were largely topographically determined.

Evidence of the existence of local ice-caps during this period separate from the inland ice comes from the Folgefonn Peninsula, where Follestad (1972) has mapped the southern extent of such a glacier. In the area south of Sognefjord, Kolderup (1908) described several frontal deltas east of the Younger Dryas marginal zone which probably belong to this deglaciation period. They must have been deposited in front of outlet-glaciers supplied by an ice-cap in the mountain areas south of Sognefjord (Anundsen & Simonsen 1968). The extent of this ice-cap is unclear. Kolderup's (1908) observations of frontal deposits at the mouth of the valley tract which extends out from Voss may indicate a connection between the ice-cap and the inland ice. If such is the case, then either Anundsen & Simonsen's reconstruction of the valley glacier flowing towards Voss is incorrect, or this glacier existed in later phases of the period in question.

The period after 9100 ± 200. – A minimum age for the deglaciation of the accumulation areas of the preceding period's glaciers is obtained by archaeological dating, indicated in Fig. 25, of charcoal: T–223 (Nydal et al. 1970), T–215 (Nydal et al. 1964), T–769 (Johansen in prep.), of humus: T–406 (Østrem 1965) and of gyttja: T–449 (Nydal et al. 1970) (cf. Mangerud 1970).

Main conclusions

1. The oldest regional ice movement, as it can be deduced from orientation elements, is found to have converged towards, and to have proceeded out of, Lusterfjord (Fig. 2). A similar movement is probable for the Weichselian maximum. Furthermore, it is probable that an ice divide extended along the Jostedal Plateau towards Hestbrepiggan and on to Jotunheimen, and that ice culmination existed above the Jostedal Plateau and Jotunheimen during this time.

2. The first period of deglaciation, the Luster Interstadial, represents a time of marked glacial retreat. Inner Lusterfjord became ice-free during this period, and the marine limit here is ca. 130 m a.s.l. The equilibrium line displacement can hardly have been less than –200 m. Winter precipitation, at least in the later parts of this period, seems to have been relatively light. The Luster Interstadial is a thermomer which follows the Younger Dryas period but is older than 9800 ± 200 years.

3. The Gaupne Stadial is characterized by advances and stagnations of the glaciers caused by several climate depressions. The Storhaug Phase is an especially conspicuous kryomer roughly mid-way through the Gaupne Stadial. The longitudinal profiles of the valley glaciers (Fig. 18) indicate a large total exchange. The equilibrium line displacement during the Gaupne Stadial was –300 ± 50 m, which means summer temperature was ca. 1.6 – 2.1 °C lower than it is today, assuming no difference from the present day precipitation pattern. Sea level at the end of the stadial was ca. 99 m higher at the head of Gaupnefjord and ca. 105 m higher at the head of Lusterfjord. The extent and relief of the Gaupne Stadial glaciers are shown in Fig. 17. The Gaupne Stadial encompasses the time interval 9800 ± 200 to 9500 ± 200 B.P.

4. The Gaupne/Høgemo Interstadial is characterized by glacial retreat. The surface of the ice-sheet sank ca. 150–200 m near the accumulation area.

5. The Høgemo Stadial is characterized by a stagnation of the glacier margins. The termini in Jostedal, Mørkrisdal and Fortunsdal lay at Høgemoen, Meljadn and Fortun, respectively. Marginal deposits are so sparse that further reconstruction of the ice cover is difficult. Sea level during the Høgemo Stadial was 9 m lower than during the Gaupne Stadial. The climate did not deteriorate as much during this stadial as it did during the Gaupne Stadial. The Høgemo Stadial ended 9100 ± 200 years ago.

6. The character and location of the deglaciation phenomena from the period of final deglaciation indicate that there were climatically inactive masses of ice, and an equilibrium line displacement which must have been positive, at least in the later stages of the period.

7. The date of the expansion of pine in the Lusterfjord area is 8550 ± 170 B.P. The empirical and rational limit for black alder is 8030 ± 110 B.P. Hazel expansion occurred contemporaneously or just prior to this (Fig. 20).

Acknowledgements. – Statsgeolog P. Holmsen and Universitetslektor O. Fr. Bergersen introduced me to the field area. The field project was supported financially by Norges geologiske undersøkelse. The echo-profiling was carried out by Professor H. Holtedahl and his assistants. Dr. M.-B. Florin supervised the diatom analysis. P. Tallantire, B. Sc.. identified the seed of alder, and the remaining macro-fossils were analyzed by Forskningsstipendiat K.-D. Vorren. The illustrations were prepared by Miss E. Irgens, Miss T. Nagelsett and Mr. L. Tangedal. Cand. real. O. W. Fareth critically read the thesis (1970) upon which most of the present paper is based. The present manuscript was read critically by Førsteamanuensis J. Mangerud and doctoral cand. A. Genes, translated by Miss K. Sullivan, and typewritten by Miss V. Meyer. To all of these people, and many others with whom I have had valuable discussions, I extend my most sincere thanks.

REFERENCES

Ahlmann, H. W. 1922: Glaciers in Jotunheimen and their physiography. *Geogr. Annaler 4*, 1–57.

Andersen, B. G. 1954: Randmorener i Sørvest-Norge. *Norsk geogr. Tidsskr. 14*, 274–342.

Andersen, B. G. 1960: Sørlandet i Sen- og Postglacial tid. *Norges geol. Unders. 210*, 142 pp.

Andersen, B. G. 1968: Glacial geology of Western Troms, North Norway. *Norges geol. Unders. 256*, 160 pp.

Andrews, J. T., Buckley, J. T. & England, J. H. 1970: Late-Glacial chronology and glacio-isostatic recovery, Home Bay, East Baffin Island, Canada. *Geol. Soc. Am. Bull. 81*, 1123–1148.

[Anonymous] 1969: Mass-balance terms. *J. Glac. 8*, 3–7.

Anundsen K. 1972: Glacial chronology in parts of Southwestern Norway. *Norges geol. Unders. 280.*

Anundsen, K. & Simonsen, A. 1968: Et preborealt breframstøt på Hardangervidda og i området mellom Bergensbanen og Jotunheimen. *Univ. Bergen. Årb., Mat.-Naturvit. Ser. 1967*, 7, 42 pp.

Aseev, A. A. 1968: Dynamik und geomorphologische Wirkung der europäischen Eisschilde. *Petermanns geogr. Mitt. 112*, 112–115...

Banham, P. H. 1968: The basal gniesses and basement contact of the Hestbrepiggan area, North Jotunheimen, Norway. *Norges geol. Unders. 252*, 77 pp.

Behre, K.-E. 1966: Untersuchungen zur spätglazialer und frühpostglazialen Vegetationsgeschichte Ostfrieslands. *Eiszeitalter Gegenwart 17*, 69–84.

Behre, K.-E. 1967: The Late Glacial and early postglacial history of vegetation and climate in northwestern Germany. *Rev. Paleobotany Palynology 4*, 149–161.

Berglund, B. E. 1966: Late Quaternary vegetation in Eastern Blekinge, South-Eastern Sweden. I. Late-Glacial time. *Opera Botanica 12*, 1, Lund, 180 pp.

Bergstrøm, B. 1971: Deglaciasjonsforløpet i Aurlandsdalen og områdene omkring. *Thesis. Univ. of Bergen.* (Unpublished.)

Brathole, A. 1951: Kvartærgeologiske undersøkelser i Indre Sogn. *Thesis. Univ. of Oslo.* (Unpublished.)

Brøgger, W. C. 1901: Om de senglaciale og postglaciale nivåforandringer i Kristianiafeltet (Molluskfaunan). *Norges geol. Unders. 31*, 731 pp.

Bruun, I. 1967: *Standard Normals 1931–60 of the Air Temperature in Norway.* 270 pp. Det Norske Meteorologiske Inst. Oslo.

Chanda, S. 1965: The history of vegetation of Brøndmyra. A Late-Glacial and early Post-Glacial deposit in Jæren, South Norway. *Univ. Bergen. Årb., Mat.-Naturv. Ser. 1965, 1*, 17 pp.

Dahl, R. 1965: Plastically sculptured detail forms on rock surfaces in northern Nordland, Norway. *Geogr. Annaler Ser. A, 47*, 55–85.

Danielsen, A. 1970: Pollen-analytical Late Quaternary studies in the Ra district of Østfold, Southeast Norway. *Univ. Bergen. Årb. Mat.-Naturv. Ser., 1969, 14*, 146 pp.

Dansgaard, W., Johnsen, S. J., Clausen, H. B. and Langway, C. C. Jr. 1970: Ice cores and paleoclimatology, pp. 337–348 *in* Olsson, I. U. (ed.): *Radiocarbon Variations and Absolute Chronology.* 652 pp. Proc. XII Nobel Symp. Uppsala Univ. Almqvist & Wicksell, Wiley Interscience Division.

Det Norske Meteorologiske Institutt (in prep.): *Nedbøren i Norge 1930–61.* Oslo.

Eriksson, B. E. 1958: Glaciological investigations in Jotunheimen and Sarek in the years 1955 to 1957. 208 pp. *Geographica 23*, Uppsala.

Fægri, K. 1936: Quartärgeologische Untersuchungen im Westlichen Norwegen. I. Über zwei präboreale Klimaschwankungen im südwestlichsten Teil. *Bergens Mus. Årb. 1935. Naturvit. Rekke, 8*, 40 pp.

Fægri, K. 1940: Quartärgeologische untersuchungen im westlischen Norwegen. II. Zur spätquartären Geschichte Jærens. *Bergens Mus. Årb. 1939–40, Naturvit. Rekke.* 7, 201 pp.

Fægri, K. 1944: Studies on the Pleistocene of Western Norway. III. Bømlo. *Bergens Mus. Årb. 1943. Naturv. Rekke, 8*, 100 pp.

Fareth, O. W. 1970: Brerandstadier i midtre og indre Nordfjord. Thesis. Univ. Bergen (Unpublished).

Flint, R. J. 1971: *Glacial and Quaternary Geology.* 892 pp. John Wiley & Sons, Inc.

Florin, M.-B. 1945: Skärgårdstall och 'strandskog' i västra Södermanlands pollendiagramm. *Geol. Fören. Förh. 67*, 511–533. Stockholm.

Follestad, B. A. 1972: The glaciation of the south-western part of the Folgefonn Peninsula, Hordaland. *Norges geol. Unders. 280.*

Goldthwait, R. P. 1966: Evidence from Alaskan glaciers of major climatic changes. *In* World climate from 8000 to 0 B.C. *Int. Symp. Lond. 1966 Proc.* Royal Meteorological Society, London.

Hafsten, U. 1956: Pollen-analytic investigations on the late Quaternary development in the inner Oslofjord area. *Univ. Bergen. Årb. 1956. Naturv. rekke 8*, 161 pp.

Hafsten, U. 1963: A late-Glacial pollen profile from Lista, South Norway. *Grana Palynologica 4*, 326–337.

Hansen, A. M. 1891: Strandlinje-studier. *Arkiv Mat.-Naturvit. 15, (1)*, 96 pp.

Helland, A. 1876: Om beliggenheden af moræner og terrasser foran mange indsøer. *Øfvers. Kgl. Vit. Akad. Forhand. 32*, 53–82.

Heuberger, A. 1968: Die Alpengletscher im Spät- und Postglacial. *Eiszeitalter Gegenwart 19*, 270–275.

Holmstrøm, L. 1880: Om moräner och terrasser. *Øfvers. Kgl. Vit. Akad. Forhandl. 36*, 5–47.

Holtedahl, H. 1964: An Allerød fauna at Os, near Bergen, Norway. *Norsk geol. Tidsskr. 44*, 315–322.

Holtedahl, H. 1967: Notes on the formation of fjords and fjord-valleys. *Geogr. Annaler Ser. A 49*, 188–203.

Holtedahl, O. 1953: Norges geologi. *Norges geol. Unders. 164*, 1118 pp.

Holtedahl, O. 1960: Geology of Norway. *Norges geol. Unders. 208*, 540 pp.

Hoppe, G. 1971: Nordvästeuropas inlandsisar under den siste istiden – några glimtar från et forskningsprogram. *Svensk Naturvetenskap 24*, 31–40.

Hultén, E. 1971: *Atlas över växternas utbredning i Norden.* 531 pp. Generalstabens litografiska Anstalts Förlag. Stockholm.

Iversen, Johs. 1954: The Late-Glacial flora of Denmark and its relation to climate and soil. *Danmarks geol. Unders. II Række, 80*, 87–119.

Iversen, Johs. 1967: Naturens udvikling siden sidste istid. pp. 345–445. *In* Nørrevang, A. & Meyer, T. J. (eds.): *Danmarks Natur I.* Politikens Forlag.

Johansen, A. B. (in prep.): Høyfjellsfunn ved Lærdalsvassdraget II.

Johnson, G. 1956: Glaciomorfologiske studier i södra Sverige. *Medd. Lunds Univ. Geogr. Inst. Avhandl. 30*, 407 pp.

Kaldhol, H. 1941: *Terrasse- og strandlinjemålinger fra Sunnfjord til Rogaland.* 206 pp. Hellesylt.

Kjerulf, Th. 1871: Om skuringsmærker, glacialformation og terrasser samt om grundfjeldets mæktighed i Norge. I Grundfjeldet. 101 pp. *Univ. program første Halvår 1870.* Kristiania (Oslo).

Kjerulf, Th. 1879: *Udsigt over det sydlige Norges geologi.* 262 pp. Kristiania (Oslo).

Kolderup, C. F. 1908: Bergensfeltet og tilstødende trakter i senglacial og postglacial tid. *Bergens Museum Aarb. 1907, 14*, 266 pp.

Landmark, K. 1949: Geologiske undersøkelser Luster–Bøverdalen. *Univ. Bergen Årb. 1948. Nat.vit. rekke 1*, 57 pp.

Liestøl, O. 1963: Et senglacialt breframstøt ved Hardangerjøkulen. *Norsk Polarinst. Årb. 1962*, 132–139.

Liestøl, O. 1967: Storbreen glacier in Jotunheimen, Norway. *Norsk Polarinst. Skr. 141.* 63 pp.

Ljungner, E. 1949: East–West balance of the Quaternary ice caps in Patagonia and Scandinavia. *Bull. geol. Inst. Uppsala 33*, 11–96.

Lüttig, G. 1965: Interglacial and interstadial periods. *J. Geol. 73*, 597–591.

Mangerud, J. 1970: Late Weichselian vegetation and icefront oscillations in the Bergen district, Western Norway. *Norsk geogr. Tidsskr. 24*, 121–148.

Mörner, N. A. 1969: The Late Quaternary history of the Kattegat Sea and the Swedish west coast. *Sveriges geol. Unders. Ser. C 640*, 487 pp.

Monckton, H. W. 1913: On some valleys, terraces and moraines in the Bergen district, Norway. *Proc. Geol. Ass. 24*, 33–52.

Nydal, R., Løvseth, K., Skullerud, K. E. & Holm, M. 1964: Trondheim natural radiocarbon measurements IV. *Radiocarbon 6*, 280–290.

Nydal, R., Løvseth, K. & Syrstad, O. 1970: Trondheim natural radiocarbon measurements. *Radiocarbon 12*, 205–237.

Østrem, G. 1965: Problems of dating ice-cored moraines. *Geogr. Annaler Ser. A 47*, 1–38.

Rekstad, J. 1914: Fjeldstrøket mellom Lyster og Bøverdalen. *Norges geol. Unders. 32*, 124–214.

Rye, N. 1970: Einergrein av Preboreal alder funnet i israndavsetninger i Eidfjord, Vest-Norge. *Norges geol. Unders. 266*, 246–251.

Strøm, K. 1956: The disappearance of the last ice sheet from central Norway. *J. Glac. 2*, 747–755.

Tollan, A. 1963: Trekk av isbevegelsen og isavsmeltningen i Nordre Gudbrandsdalens fjelltrakter. *Norges geol. Unders. 223*, 328–345.

Undås, I. 1963: *Ra-morenen i Vest-Norge.* 78 pp. J. W. Eide, Bergen.

Vorren, T. O. 1970: Deglaciasjonsforløpet i strøket mellom Jostedalsbreen og Jotunheimen. Thesis. Univ. of Bergen. (Unpublished).

Vorren, T. O. 1972: Interstadial sediments with rebedded interglacial pollen from Inner Sogn, West Norway. *Norsk geol. Tidsskr. 52*, 229–240.

Zoller, H. 1960: Pollenanalytische Untersuchungen zur Vegetationsgeschichte der insubrischen Schweiz. *Denkschr. Schweiz. Nat. Ges. 83, No. 2*, 45–156.

Zoltai, S. L. 1967: Glacial features of the north-central Lake Superior region, Ontario. *Canad. Earth Sciences 4*, 515–528.

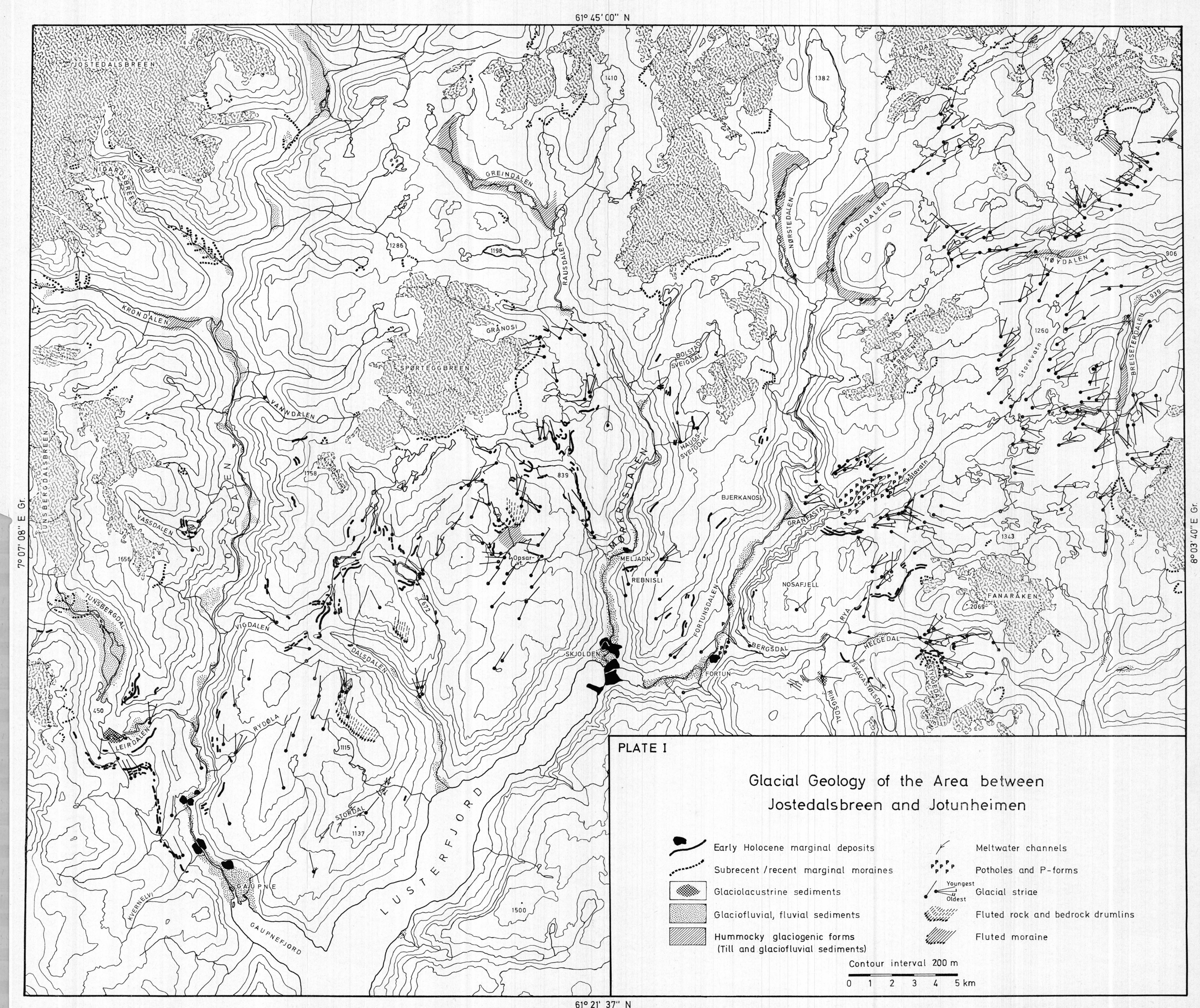

PLATE I

Glacial Geology of the Area between Jostedalsbreen and Jotunheimen