

Glaciotectonic structure and genesis of the Herdla Moraines, western Norway

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At its type locality, glaciomarine sediment of the Herdla Moraines was deformed and consolidated by overriding ice. Two phases of moraine development took place in conjunction with ice movement coming first from the east–northeast and later from the southeast. The early phase is related to development of a local ice dome on Stølsheimen during the middle Younger Dryas (about 10,500 BP), when the ice sheet reached its late glacial maximum thickness in the region. The ice sheet then thinned rapidly, due to climatic amelioration and increased calving, and shifted to southeasterly movement at Herdla. Most glaciotectonic disturbance and consolidation of sediment at Herdla happened during this late phase, perhaps as a result of a glacier surge along Herdlafjord. The glaciotectonic structures at Herdla are local in character and cannot be used for regional correlation of glacial advances in western Norway.

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Younger Dryas moraines of western Norway

Aarseth & Mangerud (1974) defined the Herdla Moraines as a morphostratigraphic unit with a type locality at Herdla island (Fig. 1). Sediments of the Herdla Moraines are part of a conspicuous system of ice marginal deposits of Younger Dryas age (10,000–11,000 BP) that has been mapped around the periphery of southern and western Norway (Mangerud et al. 1979). These deposits define an irregular ice margin that was controlled by positions of fjords and mountains, by locations of ice domes and divides, and by climatic conditions (Mangerud 1980).

Younger Dryas deposits are thickest and morphologically most prominent below the contemporaneous marine limit. Large ice-contact submarine fans and deltas, many 10s to >100 m thick, are preserved in valleys and fjords both above and below present sea level. These fan deposits are often situated next to bedrock thresholds that controlled the positions of local ice calving. In many cases, fans were later overridden by ice and covered with till composed largely of material reworked from the fan deposits. Above the Younger Dryas marine limit, the moraines are thin and patchy.

The Younger Dryas deposits of western Norway include small glaciotectonic structures

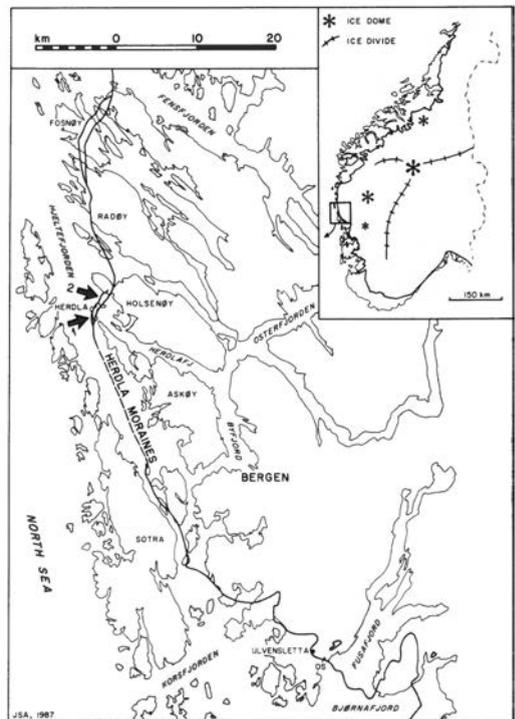


Fig. 1. Location map for the Herdla Moraines between Bjørnafjord and Fensfjord, western Norway. Position of moraine based on Aarseth & Mangerud (1974, Fig. 1). 1 = southern Herdla island (Fig. 2) and 2 = Herdlafaket (Fig. 8). Inset map shows Younger Dryas ice margin and reconstruction of ice domes and divides (adapted from Sollid & Reite 1983, Fig. 57).

that were produced by oscillations of the active ice margin (Sollid & Reite 1983). Anundsen (1972) mentioned folded glaciofluvial strata in the proximal part of an ice-front delta at Etne between Etnefjord and Stordalsvatn. Strongly disturbed moraine deposits are also present on Halsnøy and the adjacent mainland next to Hardangerfjord (Holtedahl 1975).

Sønstegaard (1979) emphasized the extremely local character of small ice-pushed folds from a site near Os, south of Bergen (Fig. 1). Nearby at Ulvensletta, a small ridge mapped as an end moraine by Aarseth & Mangerud (1974, Fig. 5) actually consists of thrust and folded sand and gravel that was shoved onto an ice-contact delta. Zones of suspected glaciotectionic disturbances have also been identified on seismic sections from portions of the moraine submerged in Nordfjord (Aarseth 1987).

Aside from the detailed study at Os (Sønstegaard 1979), none of these glaciotectionic examples have been carefully investigated. Our study on Herdla island was undertaken to determine if detailed analysis of small glaciotectionic features could provide additional information for interpreting the genesis of the Herdla Moraines at its stratotype.

The Younger Dryas glaciation experienced a major readvance in the Bergen region, where shelly tills containing reworked marine Allerød sediments are found. Up to 100 m of presumed Allerød marine strata are preserved in protected positions in Byfjord north of Bergen. A radiocarbon date of $10,540 \pm 130$ BP (Aarseth & Mangerud 1974), from whole *Mya truncata* shells directly below deformed and overridden ice-contact deposits on Herdla island, provides a maximum date for the farthest advance of the Younger Dryas ice sheet in this area.

Striations associated with the Younger Dryas glaciation exhibit a diverging pattern that reflects development of an ice dome located northeast of Bergen (Fig. 1) on Stølsheimen mountain plateau (Aa & Mangerud 1981; Hamborg & Mangerud 1981). In the Bergen area these striations trend toward the southwest. Still younger striations are more variable in orientation, showing a closer correspondence with fjords and other local topographic features.

Herdla island

The moraine on Herdla island is morphologically

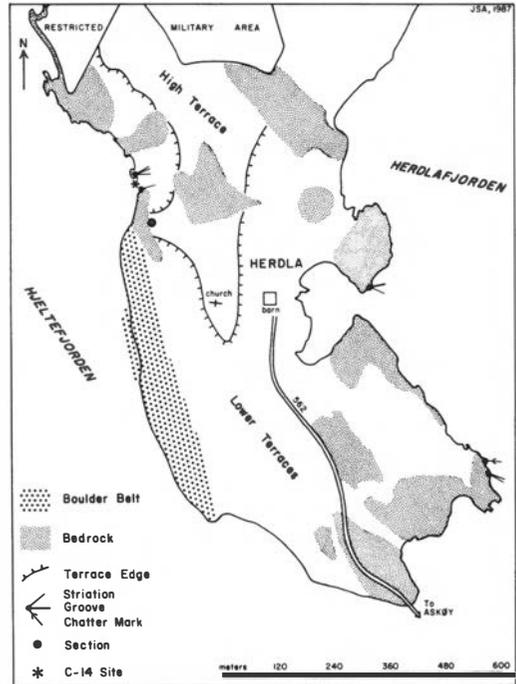


Fig. 2. Map of southern Herdla island showing geologic and glacial features. Based on interpretation of large-scale aerial photographs and ground observations.



Fig. 3. Photograph of large groove with internal meander-scroll P-forms at entrance to Herdla harbor. Groove trends 260°; scale pole is 2 m long.

the most conspicuous in the Bergen region. The island lies across the western end of Herdla fjord and includes several bedrock knobs, between and around which the ice marginal deposits are found (Fig. 2). The moraine surface has been modified into several marine terraces, the highest of which exceeds 30 m elevation. A distinct beach ridge on this terrace at 32 m elevation presumably represents contemporaneous sea level (Aarseth & Mangerud 1974).

Till is present on the proximal (eastern) side of the island, and a prominent boulder belt lies along the western edge. The belt contains hundreds of large boulders, which mark the outer limit reached by the Younger Dryas ice sheet on Herdla (Aarseth & Mangerud 1974). Clay- and silt-rich layers within the moraine are strongly consolidated. This degree of consolidation could only be caused by compaction beneath overriding ice,

because the sediments have never been deeply buried otherwise or subjected to chemical diagenesis below the shallow zone of oxidation.

Most of the bedrock striations and grooves on Herdla trend west or southwest between 240° and 275° (Fig. 2). A smaller group trends toward the northwest between 290° and 300°. The age relationship is demonstrated at the entrance to Herdla harbor on the eastern side of the island, where an older set of grooves trends 260° (Fig. 3). Younger striations trend across these grooves at 295°. These two directions indicate respectively the average older and younger ice movements on Herdla.

Herdla section: sediments

A new road was built during late 1985 or early 1986 on the western side of Herdla island. The

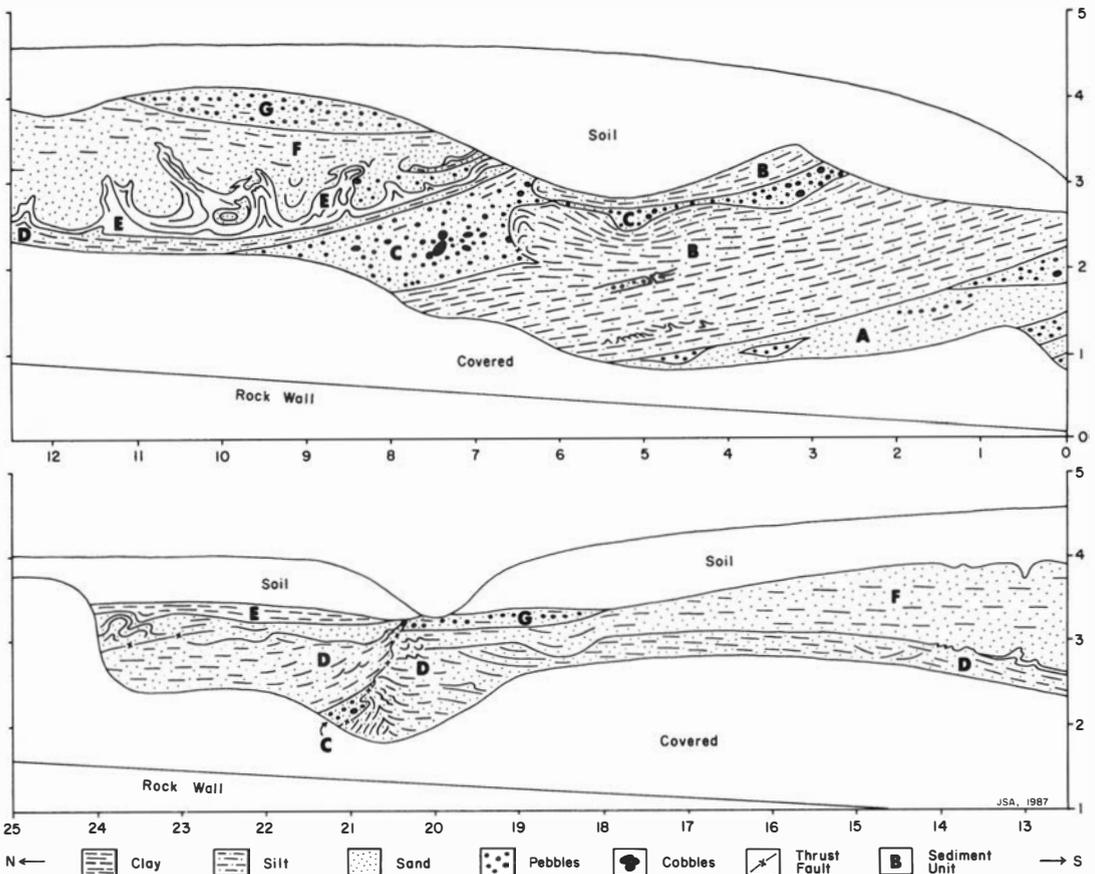


Fig. 4. Measured section on western Herdla island (see Fig. 2 for location) showing internal structure of glaciomarine deposits. Base of section is approximately 18 m elevation; scale in meters.

road cuts across the western edge of the high terrace above a bedrock knob. The internal structure is visible in a section 25 m long and up to 5 m high that strikes due north–south. Beds of stratified clay, silt, sand, and gravel up to cobble size are exposed and display a variety of deformations.

Sedimentary units are given informal letter designations on the measured section (Fig. 4). Fine-grained sedimentary units (B and E) are moderately to strongly consolidated; coarser beds are unconsolidated; and the uppermost unit (G) is cemented with iron oxide. With the exception of a few minor unconformities and small channels, this entire sequence appears to have originally been conformable.

Given the marine limit of 32 m on Herdla, the sediment revealed in the section was presumably deposited in water some 10–15 m deep in front of the Younger Dryas ice sheet. A rapidly fluctuating sedimentary environment is indicated, and the inclusion of cobbles (but not boulders) suggests that the ice margin was a short distance away at the time of deposition. The sediments are interpreted as part of a submarine fan laid down when the ice margin was located along the eastern edge of Herdla island.

Herdla section: structures

The sediments are disturbed by several secondary structures, the largest of which is a broad syncline (0–15 m, Fig. 4). The axis of the syncline trends/plunges 45/06 NE, and the limbs define an interlimb angle of about 150°. In the southern limb of the syncline, units B and C display structures typical of soft-sediment deformation (Fig. 5). Coarse gravel of unit C appears to have settled into fine sand of unit B, and bedding in the sand is compressed below thicker knots of gravel. Small, cm-scale folds are visible in the lower portion of unit B. These deformations happened penecontemporaneously with deposition of units B and C sediment, before synclinal folding, and the units above and below are not involved.

Toward the center of the syncline, between 7 and 14 m, unit E forms a series of small, but elegant diapirs. The biggest diapirs, up to 1 m high, occupy the syncline trough, and progressively smaller diapirs and overturned folds are found on the syncline flanks. The smaller flank diapirs are overturned inward (toward the trough), whereas bigger trough diapirs bend out-



Fig. 5. Photograph of Herdla section (5–15 m, Fig. 4) showing soft-sediment deformation of units B and C in southern limb and small diapirs of unit E in trough of ice-pushed syncline. Scale pole is 2 m long.

ward. The sand of unit F above the diapirs is foliated and wraps smoothly around each diapir. Individual diapirs are thin, only 10–20 cm thick, and have a curved shape in the third dimension.

The diapirs appear to be thin-walled partitions forming a bent, three-dimensional network or grid of intrusions (Fig. 6). The partition network is bent according to its position within the broad syncline. The clayey silt sediment forming the diapirs thickens toward the center of the syncline, as if sediment had flowed toward the fold axis during intrusion of the diapirs. The overall geometry of the diapir network implies that the diapirs were created at the same time as the syncline was folded. Flow and intrusion of unit

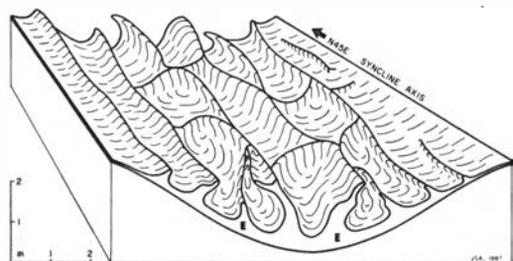


Fig. 6. Cut-away block diagram showing idealized geometry of unit E diapirs within the broad syncline at Herdla section. Diapirs form a network of bent partitions.

E sediment caused slight thickening in the trough portion of the syncline.

The northern portion of the section is disrupted by thrust faults and associated folds. At 20.3–21.4 m, a thrust cuts through the entire exposed sequence along an irregular, stepped fault (Fig. 7). The fault plane strikes/dips 65/28 NW. Toward the bottom of the fault, unit C is involved in a small fold, whose axis trends/plunges 55/25 NE. The orientation of this fold indicates primarily dip-slip displacement by the fault. Below the fault in unit D, a chevron fold with a disrupted core is visible. This fault and related folds are subparallel with the broad syncline axis and presumably formed at the same time as the syncline. Together they produce overall structural shortening of the section in a SE–NW direction.

Another thrust fault can be seen at the far northern end of the section, 22.6–24 m, entirely within unit D. This fault strikes/dips 350/30 NE. Above the thrust, overturned drag folds are developed with trend/plunge of 310/04 NW. The lateral extent of these structures to the north could not be determined due to slumping of the section. These features indicate fault displacement in a NE–SW direction.

The structural features indicate deformation of the sediment sequence in two general directions: (1) SE–NW and (2) NE–SW. However, structures of one direction do not cut across structures of the other direction anywhere in the section. Therefore, the age relationship between the two directions of deformation cannot be determined

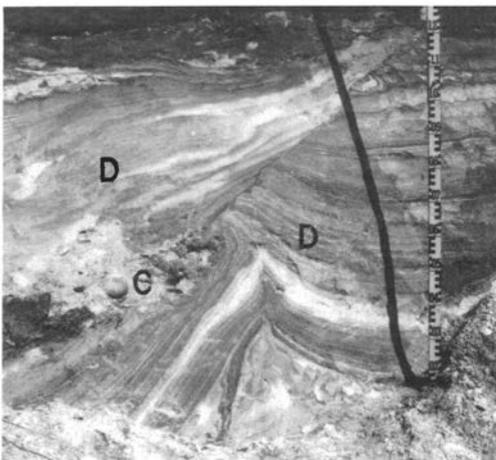


Fig. 7. Photograph of thrust fault and associated folds at 20–21.5 m of section (Fig. 4). Scale pole marked in 10 cm intervals.

from the structural features alone. This situation is not uncommon in glacial sequences, particularly for exposures of such small size.

Creation of the broad syncline and thrust faults with associated folds is interpreted as the result of two directions of ice pushing. However, the possibility of submarine slumping due to soft-sediment deformation must also be considered. Slumping or collapse of ice marginal sediment typically involves normal faulting; no normal faults were observed in the section.

Reverse and thrust faults can develop above the zone where a buried ice mass melts away (McDonald & Shilts 1975) or in the distal end of a slump mass. In view of the geological setting for the section on the upper edge of a marine terrace, the possibility of an ice mass becoming buried by sediment is considered remote. It is also unlikely that the section cuts through the toe of a slump mass. Furthermore, genesis of the structures due to slumping does not explain strong consolidation of fine-grained layers. We conclude, therefore, that the major structures in the Herdla section are glaciotectionic in origin.

Herdlaflaket

Herdlaflaket is a shallow sill, <10 m deep, that extends underwater north of Herdla island toward Holsenøy (Fig. 1). Detailed seismic-reflection investigation of Herdlaflaket was undertaken in connection with plans to dig a canal for offshore drilling platforms. Herdlaflaket is underlain by moraine sediment up to 100 m thick (Fig. 8), and both the inner and outer sides of Herdlaflaket are marked by narrow, boulder-covered ridges of the Herdla Moraines.

Much of the moraine sediment is strongly consolidated. The acoustic velocity of the consolidated sediment is estimated to be 1,800 m/s, compared with 1,600 m/s for unconsolidated sediment. Consolidated sediment includes scattered boulders and makes up much of the proximal (eastern) side of Herdlaflaket. Conversely, the distal side is underlain by unconsolidated sand and silt containing few boulders. These sediments were supposedly deposited in an ice-contact submarine fan or delta, much like the sediment exposed in the Herdla section, and compacted by overriding ice.

Boulder-rich moraine ridges are clearly defined in the seismic profiles. The outer (western) ridge is present along all of Herdlaflaket and appears

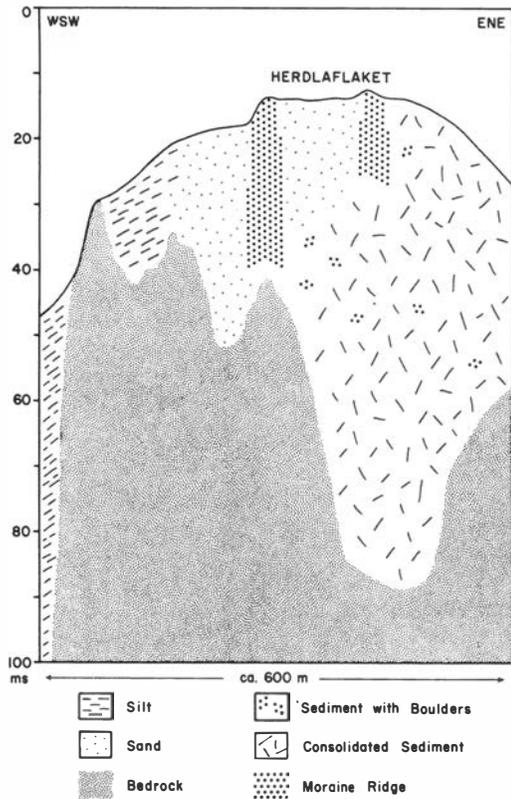


Fig. 8. Interpretation of representative seismic profile across Herdlaflaket. Depth scale in milliseconds two-way travel time for acoustic signal. Depth scale varies according to acoustic velocity of sediments and bedrock; approximate vertical exaggeration = 10×.

to be the larger of the two ridges. Its western face has an unusually steep slope, up to 15°. This slope is maintained by a boulder cover and possibly by the consolidated nature of the sediments. The smaller inner ridge is not visible in all profiles across Herdlaflaket. Where present, it is normally separated from the outer ridge by 100–200 m. The moraine ridges are interpreted as ice-margin accumulations, similar in origin to the boulder belt on western Herdla island.

Genesis of the Herdla Moraines

The major portion of the Herdla Moraines was deposited as submarine fans or deltas along the calving margin of the Younger Dryas ice sheet. The active ice margin was unstable and experienced local oscillations. Where such oscillations

took place, glaciotectonic disturbance and strong consolidation of the overridden sediment may be expected.

At Herdla and Herdlaflaket, two phases of moraine development took place corresponding to local changes in movement of the Younger Dryas ice sheet (Fig. 9). During the early phase, the ice margin reached eastern Herdla and Herdlaflaket, at which time a submarine fan was constructed. Ice movement, as shown by older striations, was from the east–northeast. Deformations at the northern end of the section may have occurred during this phase when the ice margin oscillated slightly.

The direction of ice movement subsequently changed to southeasterly, essentially parallel to Herdlaflaket, and the ice margin advanced to the western edge of Herdla and Herdlaflaket. Most of the deformation in the section on western Herdla, including folding and thrusting of the broad syncline, occurred during this late phase. Consolidation of the fine-grained sediment and intrusion of diapirs in a fluidized state also took place at this time due to increased loading by overriding ice. The boulder belt on western Herdla island and Herdlaflaket was deposited along the ice margin.

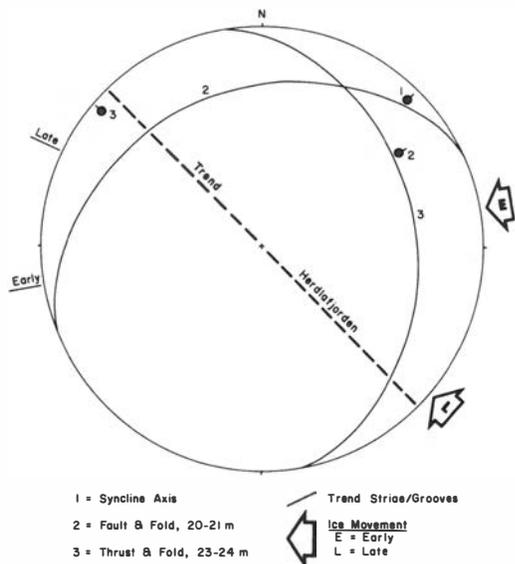


Fig. 9. Summary equal-area stereonet plot of directional data for various features of the Herdla vicinity. Early phase of ice movement from east–northeast; late phase from southeast subparallel to Herdlaflaket.

Regional implications for Younger Dryas glaciation

The overall position of the Herdla Moraines and pattern of older striations were related to radial outflow from an ice dome to the northeast during the early phase. The ice sheet must have been fairly thick, as ice flow was largely independent of local topography. The shift to fjord-parallel ice movement in the Herdla vicinity occurred as the ice sheet was becoming thinner, and ice flow was thus more responsive to local topography.

A marine transgression took place along the entire western coast of Norway during the Younger Dryas as a result of crustal depression by the expanded ice sheet. This transgression culminated on the island of Sotra (Fig. 1) around 10,400 BP after which sea level fell precipitously to levels beneath its previous low (Anundsen 1985). Most of this fluctuation in sea level is probably due to glacio-isostasy, although small geoid effects are also involved (Fjeldskaar & Kanestrøm 1980). All fjords in the Bergen region were ice free by 10,000 BP (Aa & Mangerud 1981).

Rapid reduction of the late Younger Dryas ice sheet in western Norway resulted from climatic change; however, whether an ice sheet of that size could respond so quickly to changing climate alone is open to question. The role of sea level may also be significant. Rising sea level could have caused an increase in the rate of ice calving, and the sea may have broken through certain deep fjord thresholds, such as the entrance to Fensfjord (350 m deep) and others. Calving bays may have formed in these deeper fjords, so that the ice surface was lowered and movement directions altered in surrounding areas. Where fjord thresholds are shallow, either because of bedrock bars or moraine buildup, fjord glaciers were able to remain grounded against the thresholds.

Due to a time lag between ice-sheet growth and marine transgression, the increase in calving did not happen until after the ice sheet had reached its maximum size. According to this scenario, an increased rate of calving and resulting rapid drawdown of the ice surface may have caused the shift along the Herdla Moraines from east-northeasterly ice movement to fjord-parallel ice movement. This shift in ice movement may have taken place quite quickly, perhaps in less than a century, due to the efficiency of calving in fjords with deep thresholds.

Another possible consequence of the marine transgression was the occurrence of late Younger Dryas glacier surges along the fjords. Rising sea level would necessarily cause an increase in hydrostatic pressure beneath the ice and could trigger surges. Where surging ice was grounded on shallow thresholds, as at Herdla, ice marginal sediments could be deformed and consolidated. Many similar push-moraines were formed in the shallow fjords of western Spitsbergen by glacier surges during the Little Ice Ages (Elverhøi et al. 1983).

Ice-push deformation of sediment on Herdla island was related mainly to activity of the glacier in Herdla fjord and may not, therefore, correspond in direction or exact age to glaciotectonic structures in other portions of the Herdla Moraines system. Thus, ice-push structures are of limited stratigraphic usefulness in western Norway and cannot be used for regional correlation of glacial deposits. Such structures may provide valuable information for interpreting the genesis of deposits within a restricted area.

Conclusions

Glaciotectonic structures exposed in the section on Herdla were created when the Younger Dryas ice sheet advanced over a previously deposited ice-contact submarine fan. Two phases of moraine development were related to ice movement, which came initially from the east-northeast and later from the southeast. The shift in ice movement reflected increasing influence of local topography on thinner ice during the latter part of the Younger Dryas glaciation. The Younger Dryas marine transgression may have caused increased calving and thus rapidly drawn down the ice sheet. Most glaciotectonic structures at Herdla were created during this latter phase, perhaps as a result of a glacier surge along Herdla fjord. Glaciotectonic deformations are of limited extent and cannot be used for regional correlation of glacial advances in western Norway.

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