Glacial geology of the island Stord, west Norway

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Upland morphology suggests pre- or early-Weichselian glaciation prior to the inception of existing cirque basins formed when firn limit was 450 msl. Directional elements indicate complete ice cover probably during the main Weichselian ice advance. Cirque parameters suggest that 250–300 m of upland has been removed through a combination of glacial and fluvial erosion. Marine shells were dated by radiocarbon at 12,860 ± 250 yrs. BP designate an Older Dryas ice advance. Ice-cap conditions subsequent to the Older Dryas advance and a rising firn limit during deglaciation is postulated, with the probability of nivation processes occurring during Younger Dryas time. Isostatic adjustment of 134–138 m since Older Dryas time (12,000 yrs. BP) is calculated and a relative isostatic uplift of 12 m since Tapes transgression (6000 yrs. BP).

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The island Stord, situated some 75 km south of Bergen at the mouth of Hardangerfjorden (Fig. 1), is considered part of the strandflat region of Norway (Grønlie 1940, Andersen 1965). Recent studies on the western part of the Hardangervidda and surrounding areas (Mangerud, 1970a, Aarseth 1971, Follestad 1972) suggest that the island was probably immediately distal to the Younger Dryas ice margin. This study was undertaken to determine whether Stord was glaciated during the last glacial (Younger Dryas) ice advance and to determine the Pleistocene history of the island.

Brief descriptions of the Quaternary geology of Stord have been given by Reusch (1888), Holtedahl, H. (1967), and Mangerud (1970a). Some surficial deposits were described during bedrock mapping (Kvale 1937, Skordal 1948) and the glacial map of Norway (Holtedahl, O. & Andersen 1960) indicates some moraine deposits on Stord.

Bedrock

The bedrock consists of crystalline and volcanic rocks in the northern part of the island and metasediments of Cambrian to Silurian age in the southern part. The relationship of the igneous rocks to the metasediments has been described by Kolderup (1931) as a tectonic line which separates an upland igneous massif, in places exceeding elevations of 700 m, from low elevation metasediments metamorphosed during Caledonian time (Strand 1960), with elevations not exceeding 60 m.

The cirques

Most cirques are associated with the upland block in the northwestern segment of the island, but upland outliers with associated cirque-like forms occur at Midtfjell, Kinno, Mennene, and Store Melen (Fig. 2). Descriptions of Stord cirques are given elsewhere (Genes 1974), but some important points should be noted. First, the headwall of Svartavatn East has a polished, smooth appearance with crest ridges rounded to the west, indicating glacial modification. Secondly, most cirques have northwesterly aspect, and thirdly, moraines occur in several cirques. Fig. 2

Distribution

Cirque distribution with regard to quadrant location and cirque aspect is shown in the polar diagram (Fig. 3). The seemingly anomalous northwest location of most of the Stord cirques can be explained by preglacial valley incision along the Stord western coast resulting from the assumed Tertiary uplift (Holtedahl, H. 1967) of the landblock. The origin of cirques has been explained as resulting from primarily structural

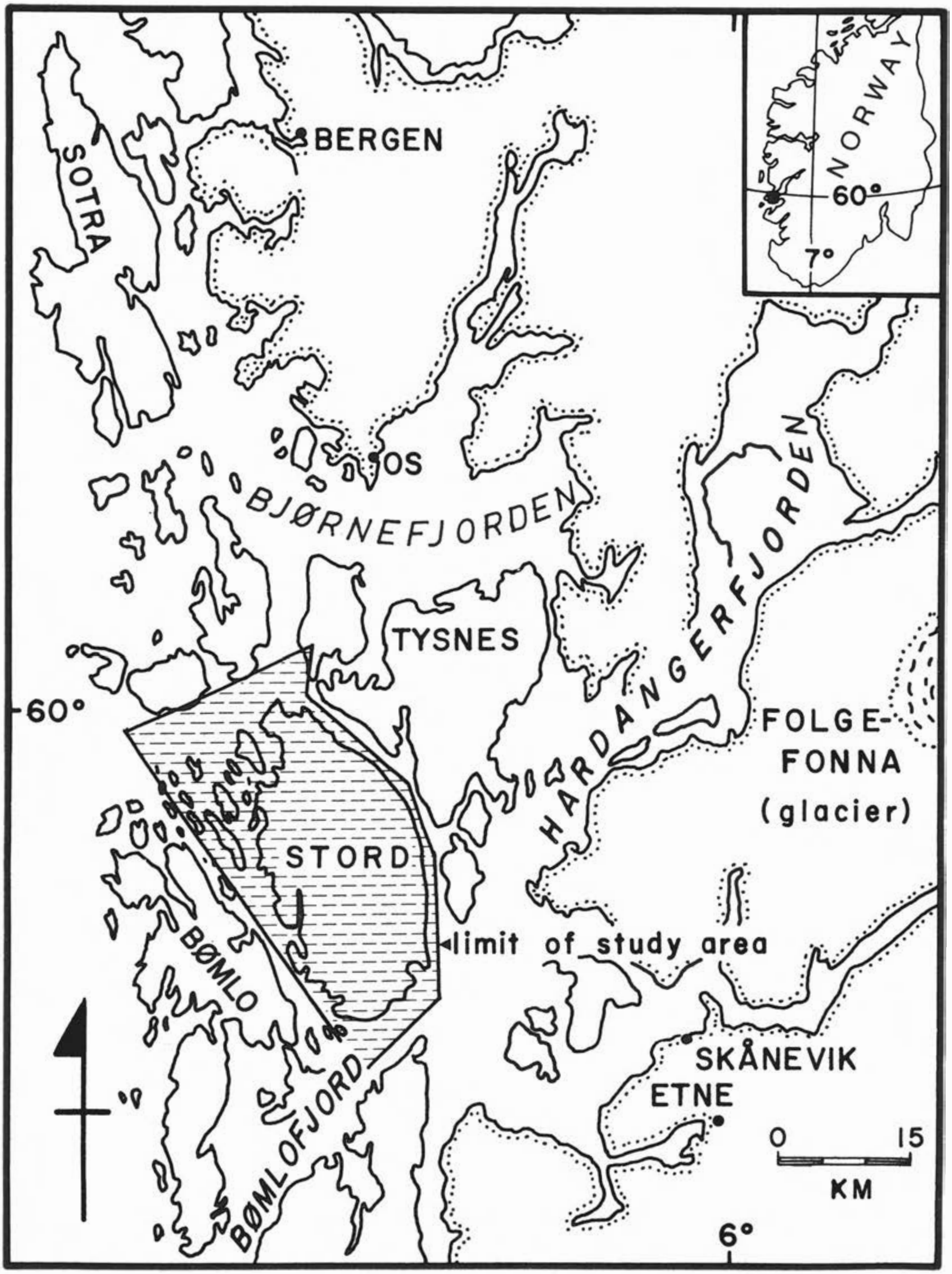


Fig. 1. Location map of Stord and vicinity, Norway.

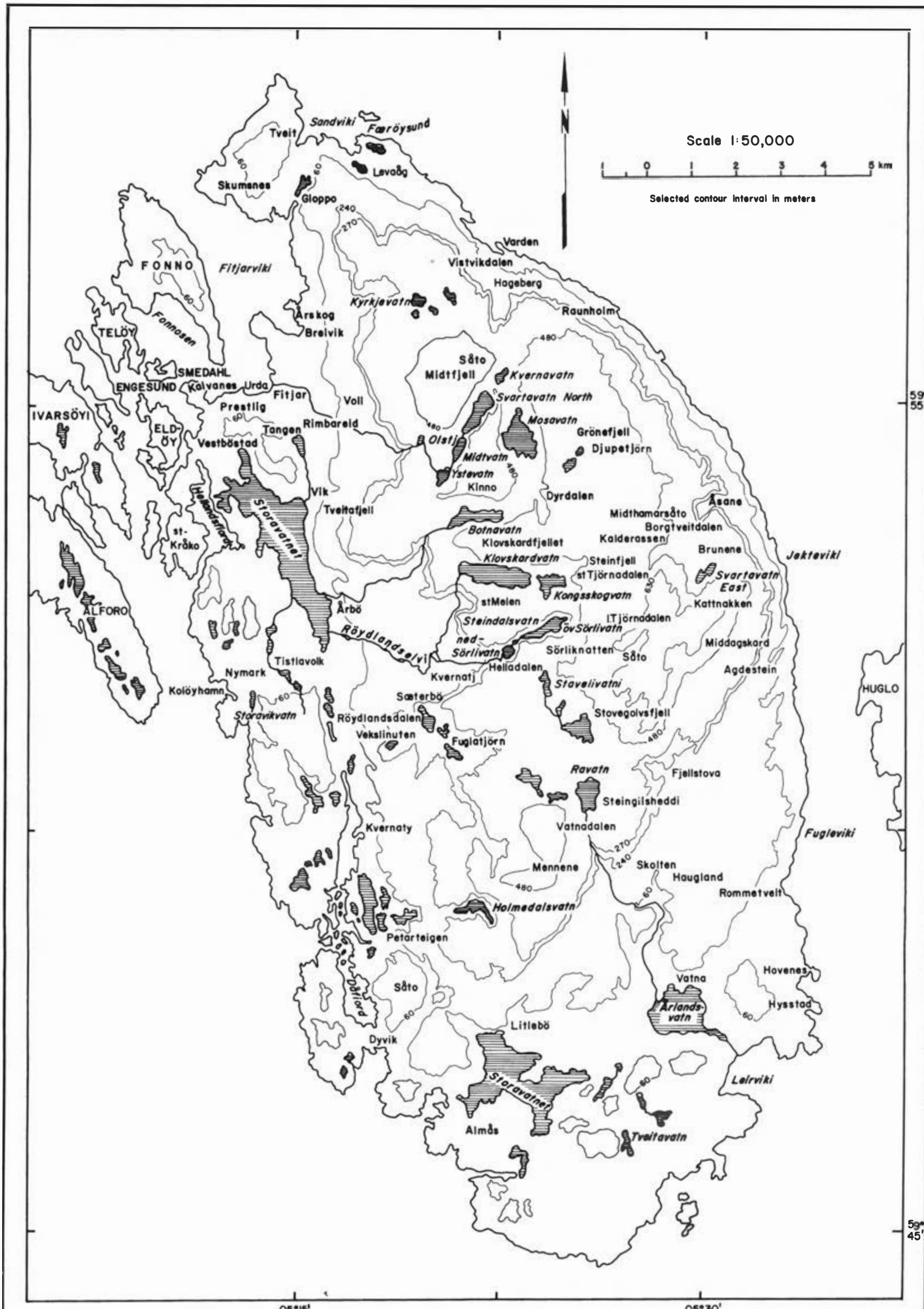
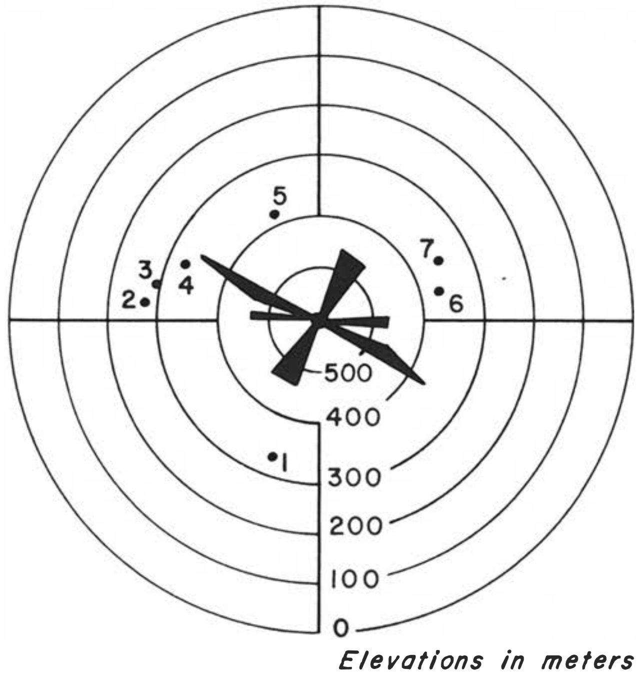


Fig. 2. Stord location map showing place names.



QUADRANT	NE	SE	NW	SW
NUMBER of CIRQUES	2	0	4	1
CIRQUE ASPECT	2	0	2	3

CIRQUES

1. MENNENE
2. STEINDALSVATN
3. KLOVSKARDVATN
4. BOTNAVATN
5. SVARTAVATN NORTH
6. SVARTAVATN EAST
7. ÅSANE

Fig. 3. Polar diagram (after Temple 1965) showing cirque aspects and quadrant locations.

(Lewis 1938, McCabe 1939, Temple 1965, Flint 1971) or meteorological (Battey 1960, Andrews 1965) considerations.

Clearly Stord cirques have originated from previously developed stream eroded sites. A possible exception is Svartavatn East, which is favorably located for accumulation of drifting snow. Structural control is even more evidenced by the aspect of Botnatvatn and Klovskardvatn where they cut along the east-west rock strike contact between granite and effusive rocks rather than having an insolation protected northwest aspect.

Morphology

Manley (1959) gives a ratio between 2.8 and 3.1 to 1 for cirque valley length to headwall summit height, whereas Andrews (1965) gives a ratio closer to 2.1 to 1. Embleton & King (1968) suggest that the latter ratio reflects immaturity of cirque development. Svartavatn East conforms to cirque morphology in all respects. However, the length to height ratio using the 660 msl contour is 0.98 to 1, considerably lower than the values given by Manley or Andrews. The cirques have been overridden, as evidenced by grooves, chattermarks, and faceting. Therefore it is not possible to determine accurately the original headwall elevation.

By using the 700 msl or 750 msl contour rather than the existing 480 msl contour which defines most of the present cirque headwalls, the length to height ratios of other cirques at Stord are in close agreement with that of Svartavatn East, as are the angles between tarn lake lips and headwall elevations.

Comparing these ratios and angles with length/height ratios of eight cirques at Rondane, the average length/height ratio of Rondane cirques is 1.2 to 1, and the lake lip to headwall angle is 41.0°. Thus it seems that Stord and Rondane cirques have smaller length/height ratios than cirques investigated by Manley & Andrews.

Lewis (1949), using the gradient of present-day glaciers in Jotunheimen, calculated an 18° average value for the slope of glacier ice contained within valley walls extending from the headwall intercept to cirque thresholds. If the present 480 msl contour is taken as the headwall elevation, an extension of this average 18° ice-surface gradient from the cirque basin lip intercepts an elevation above the present headwall in the majority of cirques at Stord (Fig. 4).

The use of the 660 msl or higher contour as the headwall elevation at the time of cirque inception agrees with calculated cirque geometry. Use of higher initial headwall elevations brings the length/height ratios and cirque basin lip to headwall angles into better agreement with cirque morphology as expressed by Svartavatn East and Rondane cirques.

Interpretation

Characteristics of the Stord cirques clearly demonstrate that they were formed early in, or prior to, the last glaciation. They were active prior to the main Weichselian ice advance, were subsequently overridden, and served as nivation hollows during the early period of deglaciation.

Morphologically, the readily apparent monuments such as Kinno, Klovskardfjellet, Store Melen, Mennene, and Midtfjell, the total lack of arêtes and horns, and the almost complete disintegration of most headwalls, characterize a monumented-upland stage. With the exception of Svartavatn East, the cirques exhibit characteristics of an advanced stage of modification in that Ahlmann's rounded-head appearance of old age has been destroyed by prolonged weathering and erosion, or by glacial overriding.

Leveled places of large areal extent around Sæterbø, Tveitafjell, and Midtfjell are interpreted as 'bastions' representing old cirque floors, in the same manner that the Brunene flat has been developed at Svartavatn East (Fig. 2). The Brunene flat exhibits a mammillated, denuded surface, perhaps the 'glacial peneplain' to which Cotton (1942) alluded. Bowl-shaped incisions, as at Fuglatjoen, Kvernatjoen, Kyrkeveien, and Olstj., may represent almost completely disintegrated cirques. Alternatively, some of these may represent plunge pools, later modified by nivation processes. The origin of Kartreppen at Svartavatn North and in the valley south from Steindalsvatn may be ascribed to compressing flow at the base of small ice falls.

If the present-day relationship of precipitation between Borgtveitdalen and the remainder of Stord is assumed to have held in the past, this relationship could explain the intact cirque morphology at Svartavatn East, either by affording advantageous conditions for prolonged cirque glacier activity or subsequent nivation processes. The rounded cirque headwall and crest attest to overriding by an ice sheet with subsequent nivation processes responsible for widening of the cirque valley. An oceanic climate prevails at Stord with mild winters, cool summers, and high precipitation, especially in winter. Mean annual precipitation (1900–1949) is slightly higher in the south than in the north, being 1669 mm at Leirvik (30 msl) and 1490 mm at Fitjar (7 msl). Locally, precipitation can be

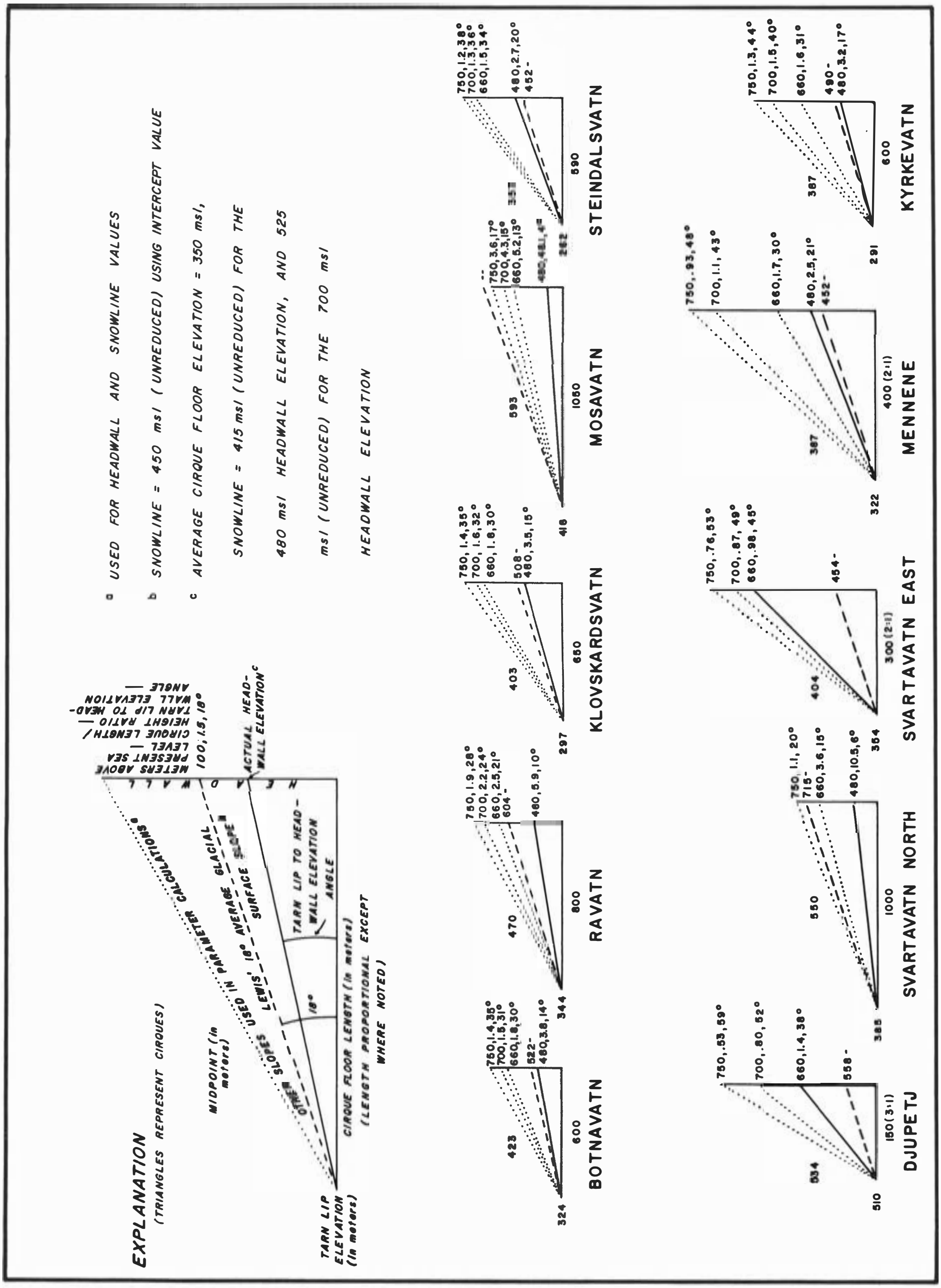


Fig. 4. Stord cirque geometry.

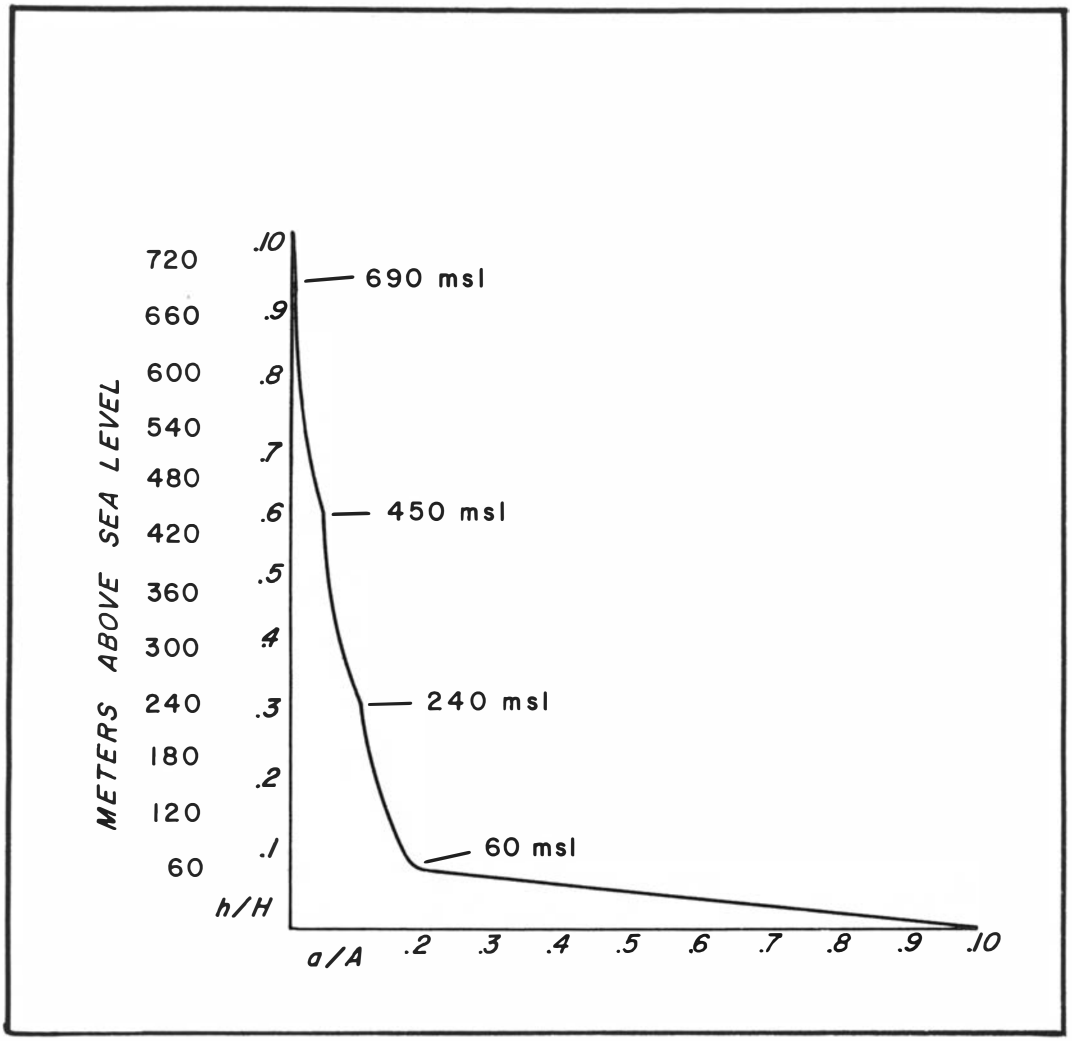


Fig. 5. Hypsographic curve for Stord. 450 msl elevation corresponds approximately to maximum elevation of upland outliers and cirque headwalls. 240 msl elevation corresponds to supposed 'bastions' of earlier cirque floors. Area determined by polar planimeter and plotted according to Strahler (1952: 1117–1142).

higher, as at Borgtveit (350 msl), midway between Leirvik and Fitjar, with a mean annual precipitation of 2833 mm (Hafsten 1965).

Glacial debris in the form of either subdued moraines, widespread glacial drift, or concentrated block deposits conforming to cirque valley mouths, suggests cirque activity after ice-sheet cover. At Midthamarsåto and Kalderassen, the highest elevations at Stord, there occur abundant chattermarks, quartzite and nordmarkite erratics, in addition to faceted surfaces and several large glacial grooves directed to 300°. These occurrences indicate complete overtopping by thin, topographically controlled ice, probably not associated with the main ice direction when ice moved independently of topography.

Using only the previously described cirques (Fig. 3) the average cirque floor is calculated to be 320 msl. The value for the snow line based on the 480 msl headwall elevation is 380 msl (unreduced).

Triangles (Fig. 4) representing most cirques and cirque-like forms on Stord show midpoint elevations between tarn lips and intercepts of the average 18° glacier ice-surface gradient. Calculation of snowline (Embleton & King 1968, Flint 1971) suggests that the average midpoint of these intercepts should approximate snowline at time of cirque formation. Unreduced snowline values for the 480 msl and 700 msl headwall elevations are 415 msl and 525 msl respectively. These values, calculated by using the average cirque floor value (350 msl) and the applicable headwall elevations, suggest the possible range of snowline elevation. The 450 msl snowline elevation value (unreduced) is obtained through the use of the average midpoint and the 18° glacier ice-surface intercept. The 450 msl snowline value is in accordance with the pronounced areal surface in Fig. 5. The previously mentioned bastions are at lower elevations than the existing cirque floors.

It is hypothesized that these cirques, of which only the bastions remain, were initiated when the snowline was considerably below 450 msl. After cirque coalescence and the development of the mammillated bastion surface, snowline rose initiating the 450 msl snowline phase of cirque formation, resulting in the present cirque configuration at Stord.

The existence of a plateau ice cap is postulated for the final phases of glacial activity, where cirques became heads of glacial valleys containing ice streams of all types which overflowed cols and eroded other cirque-generated forms such as the arêtes and horns.

The Store Tjørnadalen area is drift covered and has associated kettled lakes and large patches of till on bedrock. In many places no bedrock protrudes and moraine ponding is evident. This entire area and most other parts of the uplands have characteristic 'dead-ice' appearance suggestive of snowline elevation above the island preventing nourishment of the ice cap. The lower ends of the valleys leading from the cirques have shattered valley walls, and weathering and erosion are evident. However, the relative freshness and smoothness of the upper parts of the valley walls and the existence of protalus ramparts near most hollows attest to nivation processes during the last phases of glacial activity when the snowline became elevated above Stord.

Firn limit

Østrem & Liestøl (1964) calculated a present-day firn limit of 1400 msl for the east side of the Folgefonna ice cap. Firn limits of 1300 msl and 1200 msl were calculated for the Melderskin and Ulvenos areas respectively, as western extensions of the ice cap. Pytte & Østrem (1965) also calculated firn-limit values on Folgefonna of 1410 msl on the west and 1350 msl on the east sides for 1964. Values of 1370 msl on the west and 1390 msl on the east sides were calculated for 1968 (Pytte 1969). The intervening years had similar but fluctuating relationships.

The glaciation limit at the central-southern part of Folgefonna was determined to be 1500 msl (Østrem & Liestøl 1964:325). Follestad (1972) determined the glaciation limit to be 1420 msl on Melderskin with both the firn limit and glaciation limit depressing to the west, but with the glaciation limit approximately 100 m above the firn limit.

Similar relationships of firn and glaciation limits were determined from Ålfotbreen to Gråsubreen through Nigardsbreen. The difference between this firn limit and glaciation limit was 70–100 m within 50–70 km of the coast (Østrem & Liestøl 1964).

Follestad (1972) extrapolated a probable glaciation limit of 1300–1350 msl for Ulvenos. He calculated a glaciation limit of 900–950 msl with firn limit at approximately 800 msl for the plateau area west of Ulvenos and correlated these limits with the Fjordbrestade of Younger Dryas time. Follestad thus assumes a firn- and glaciation-limit depression of 400±50 m for Younger Dryas time in that region.

A climatic snowline depression of 450–600 m at Lysefjorden in Ryfylke (Andersen 1968) and a 600 m depression of the firn limit on Sunnmøre (Reite 1966) suggest the magnitude of firn limit depression in southwestern Norway during Younger Dryas time.

The average westerly depression gradient of the glaciation limit calculated by Follestad & Pytte is 2.6 m/km.

Extrapolation of the glaciation limit from Folgefonna to Stord using this average 2.6 m/km depression gradient yields a present-day theoretical glaciation limit of 1350 msl and a firn limit of approximately 1250 msl. By using the average firn limit of 1310 msl on the east side of Folgefonna in conjunction with the previously mentioned 2.6 m/km glaciation limit-depression gradient, a present-day firn limit of 1140 msl over Stord is obtained. An average of the two firn limits thus obtained yields a present-day theoretical firn limit over Stord of 1200 msl at Midthamarsåto. A firn limit of 800±50 msl is obtained for Younger Dryas time over Stord.

As will be shown later, the calculated isostatic adjustment of Stord suggests that the firn limit could not readily intercept the highest parts of the island, Midthamarsåto, Kaldersassen, and Kattnakken during Younger Dryas time, all lying within a reduced range of 660–700 msl during Older Dryas time.

Table 1. Pollen and spore count, S14-8, Beivik.

	Determined	Questionable	Total
Betula	10	2	12
Salix	8	2	10
Pinus	6.5	–	6.5
Other AP	3	3	6
Artemesia	7–	–	7
Gramineae	5	1	6
Koenigia islandica	1	–	1
Other NAP	7	3	10
Total pollen	47.5	11	58.5
Hystrix			261
Spores			18

The pollen count performed by cand. real. Lotte Selsing, and the identification of *Koenigia islandica* by cand. real. Dagfinn Moe, both of Universitetet i Bergen.

Stratigraphy

The late Quaternary environment of Stord was not favorable for the preservation of an extensive Pleistocene stratigraphic record. The oldest Pleistocene sediments on Stord are dark grey, silty-sand to sandy-granule tills which overlie bedrock. The angularity of clasts within this unit increases northwestward from Fugleviki to Breivik. At Breivik, marine shell fragments are concentrated in the lower 0.8 m of this till. Shells from this zone have been identified as *Chlamys islandicus, Mytilus edulis,* and *Macoma calcaria,* and were dated by radiocarbon at 12,860 ± 250 yrs. BP (DF 445 T-1171, Trondheim Radiological Laboratory ref. no.). The presence of *Hystrex* and Foraminifera in the pollen analysis spectrum (Table 1) of the fossiliferous till indicates that the sediment was initially deposited in a marine environment. The pollen in the sediment was probably redeposited, and is suggestive of vegetation occurring during either Oldest Dryas or Bølling time.

Till fabric analyses of this unit from widely separated localities yield mean azimuths which range from 272°–345°. Sediment parameters and directional elements of tills are consistent with the view that all the tills on Stord were deposited by the same ice advance.

Above the till and bedrock is a 0–17 m thick zone of well-sorted and interbedded sand and gravel with minor amounts of silt. At Breivik, Fugleviki and Vik, where the sand and gravel overlie till, the contact is sharp. The areal distribution of these deposits is patchy and intermittent throughout the island, occurring primarily at elevations from 44–55 msl. The sediments at Vatna and Vik are deltaic in origin.

A massive blue to bluish-grey clay, 0.5 m thick and containing numerous pebbles, comprises the upper unit of the Stord section. This clay, overlying only bedrock, occurs intermittently as patches along the southeastern coast. At Borgtveit whole shells in fine condition are randomly distributed throughout a 1-m section. Shells identified as *Mya truncata, Hyatella arctica* and *Chlamys islandicus* were collected from approximately 0.3 m above the base of the section and dated by radiocarbon at 10,600 ± 140 yrs. BP (DF 546 T-1370, Trondheim Radiological Laboratory ref. no.).

Ice-recessional topography

Constructional and depositional features are in close accordance with iceflow indicators and afford good evidence of deglaciation chronology (Fig. 6).

The topography north of Årskog is predominantly erosional except for isolated patches of subdued ground moraine on the northeastern segment of Tveit. Stoss-and-lee forms at Kalvanes, sparseness of drift, and the dominance of westerly striae suggest that the northern segment of Stord was essentially ice free during the last phase of Older Dryas glaciation.

In marked contrast, the Fitjar area is characterized by hummocky moraine topography extending from Breivik to Vest-Bostad and south to Vik. A large end moraine approximately 1500 m wide, herein named the Fitjar

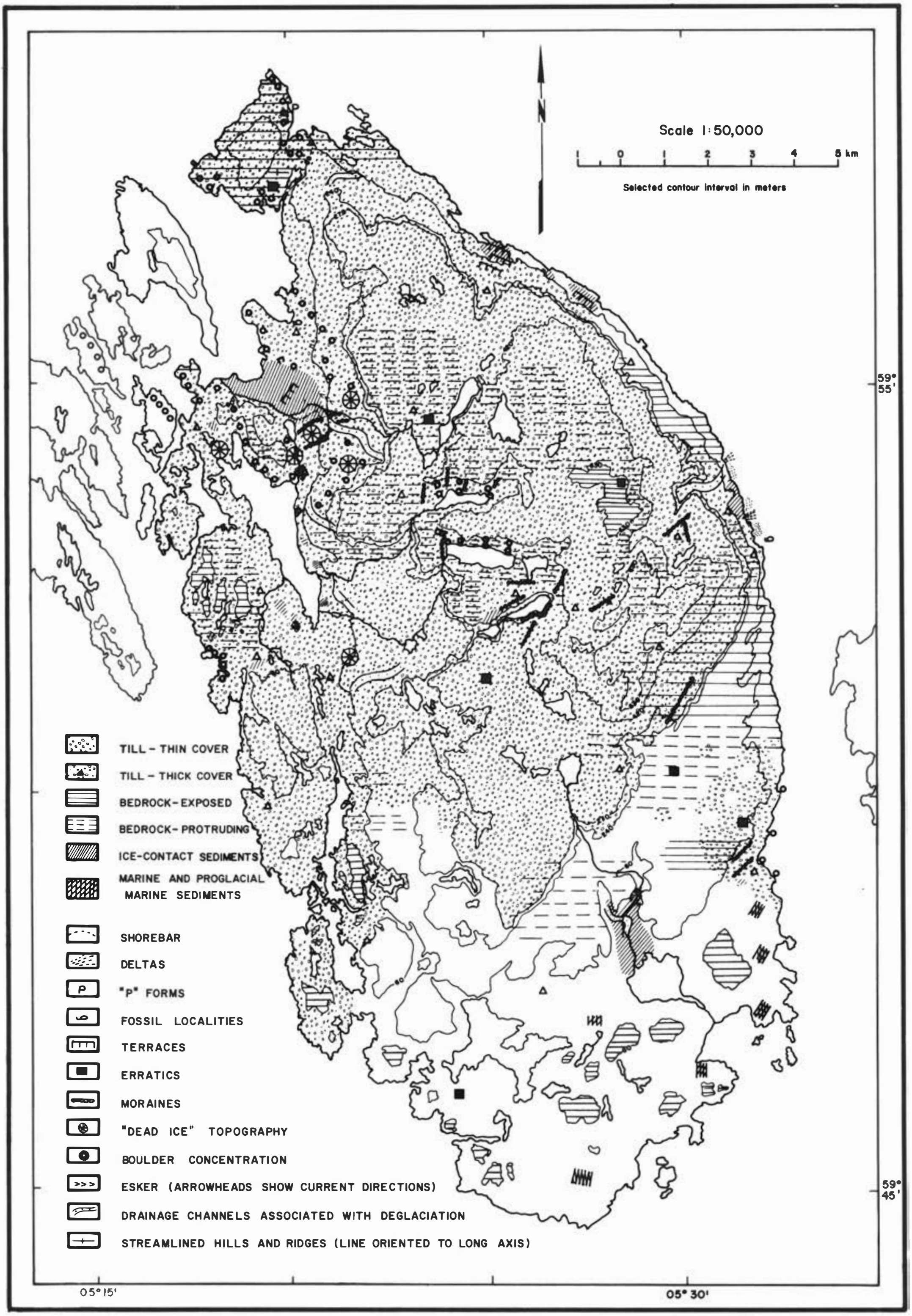


Fig. 6. Surficial geology of Stord.

Moraine, extends from the Midtfjell western slope west-southwesterly to Tangen near the northern end of Storavatnet. The Fitjar Moraine ends at Voll, at a ridge interpreted as a shore bar. The surface of low relief on which Fitjar is located is topographically expressed by fluvially incised and terraced flatland.

Hummocky moraine surface characterizes the Stord western coast to Almås in the south, although bedrock control is obvious in many instances. Boulder accumulations are associated with the Fitjar Moraine and are prominent in nearly every bay and coastal segment along the western coast.

Gravel pits occur along the Storavatnet shore and extensive delta deposits at Vik and Nymark Church both have northwesterly dipping foreset beds. Associated with the Fitjar Moraine complex are patches of stratified glaciofluvial deposits, as at Nymark, Bostad, Dyvik, Kvernaty, and the valley leading from Nymark Church to Storevatnet.

A delta at Vatna with southeasterly dipping foreset beds and poorly stratified deposits dipping northeasterly along the Borgtveit coast are the only glaciofluvial deposits along the east-southeastern parts of Stord. Terraces at Varden and Raunholm in the northeastern section are comprised of sandy to gravelly sediment. Large boulders are common along the Stord eastern coast, particularly in the Fugleviki area. The boulder pattern suggests an ice-rafted deposit. This is interpreted as deposition by 'Ra' ice during the Younger Dryas ice advance to Huglo.

Echo soundings interpreted from echograms utilizing a Simrad Scientific Counter EK, Simonsen Radio A/S, Oslo, Norway, revealed extensive submarine deposits around the Stord coast, which, with deposits on some islands off the northwestern coast, are considered to be associated with the Fitjar Moraine.

Sediment is lodged along bedrock slopes and on the inner portion of the bay, approximately 75–80 m below sea level at Fitjarviki. The morphology of the submarine sediment is expressed as a wide subdued ridge. Smoothness of the ridge surface and depth of echo penetration of the deposit suggest loosely consolidated, fine-grained to gravelly sediment. Marine sediment of similar characteristics occurs along the coast at Fonnosen, Kobeviki and Hellandsjøen at the Ålforo coast. Locally, as at the Teløy area, the irregular surface suggests a bouldery moraine. The submarine ridge extends as an arcuate form from Fitjarviki to Koløyhamn. Sediment accumulation is also evident at Færøysund on the northeastern coast.

Most islands comprising the archipelago on the northwestern coast are barren of deposits. Those islands, or parts of islands, corresponding with the trend of the submarine ridge are either drift covered or have large boulder accumulations along inlets or coasts.

Widespread, thick submarine sediment occurs in Leirvik. These deposits are interpreted as interbedded silts and clays deposited during the Younger Dryas ice advance which reached Huglo and surrounding areas.

A linear, symmetrical ridge, 14 m high, is superposed on the marine sediment at Fitjarviki. This ridge, interpreted as an esker, extends from the approximate midpoint of Fonno to the Fitjarvik coast at Urda.

Large drainage channels radially distributed around Stord imply greater erosional capability by early postglacial streams than by the present streams occupying these valleys.

Late Pleistocene shorelines

Evidence of former sea-level stands on Stord include the shorebar at Voll, the terraces at Varden, Årskog and Fitjar, and the Vik and Vatna deltas (Fig. 7).

Glacial-isostatic uplift along the southwestern coast of Norway is considered to have been discontinuous (Marthinussen 1960, Hoppe et al. 1968, Aarseth 1971, Follestad 1972) because of the effect of high elevations and deeply incised fiords upon calving and ice recession. The discontinuity of uplift is attributed to differential release from ice load (Holtedahl, O. 1960) or to discontinuity of isostatic recovery itself (Fægri 1943).

Andersen (1960) has assigned an Older Dryas/Bølling age to an approximate 46-msl strandline at Lillesand, southern Norway. This strandline corresponds closely with sea level delimited by the Fitjar Moraine, the shorebar at Voll, and the surface of low relief distal to the moraine.

Assuming contemporaneity of formation, marine limits based on barometric measurements of 44 m at Voll (Fig. 7, no. 4), 48 m at Varden (Fig. 7, no. 3) and 52 m at Vatna (Fig. 7, no. 1) suggest an 0.83 m/km isostatic uplift gra-

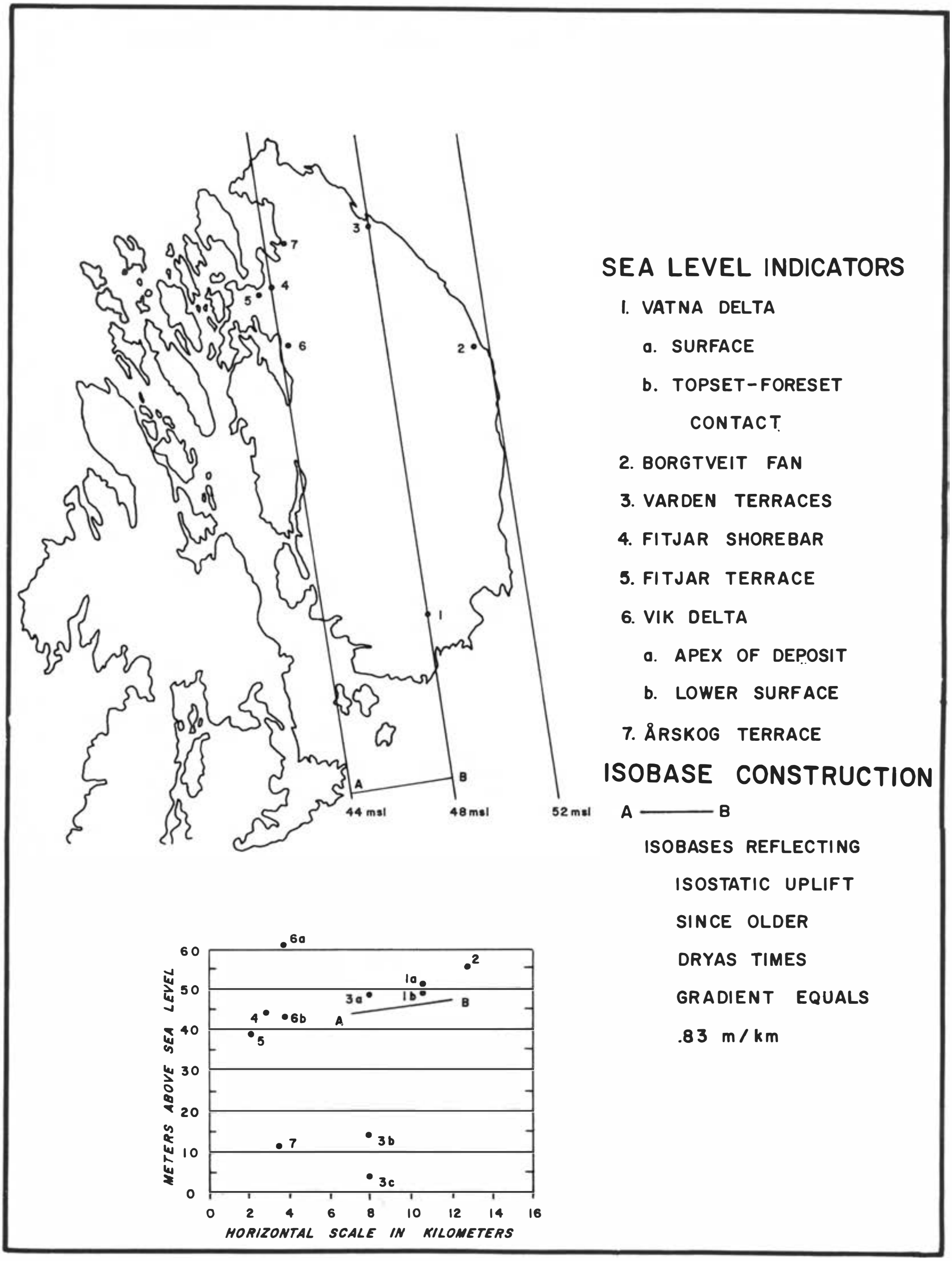


Fig. 7. Reconstructed isobases on Stord showing uplift since Older Dryas time.

dient. Isobases constructed using these measurements trend N. 6° E. Using the Stord topographic map elevation for Vatna, 48 msl, isobases trend N. 9° W. The 0.83 m/km isostatic gradient compares with Rønnevik's (1971) 0.77 m/km isostatic uplift gradient in Haugesund for isostatic compensation since Older Dryas time.

It is possible that isobase construction could have more northwesterly orientation than N. 6° E. and would include or be near all plotted points except fan apices and lower elevation terraces. The Vatna delta surface encompasses a zone rather than a point; therefore, a value between the calculated isobases is probably more correct. Isobases were constructed based on topographic map elevations because of the uncertainty of barometer readings.

Tapes transgression on Haugesund has been demonstrated by Rønnevik (1971), who correlated an east-west 10–14 msl terrace level of 0.2 m/km gradient with Fægri's (1943) highest Tapes limit on Bømlo. The Årskog terrace and the middle terrace at Varden correspond with the Tapes transgression limit on Bømlo calculated by Fægri.

The gradient of the assumed Tapes shorelines between Årskog and Varden is 0.6 m/km. Fægri (1943) used Tapes measurements at Etne and Halsnøy, inland from Stord, to correlate Tapes transgression on Bømlo to those areas and found that a 0.6 m/km gradient corresponded well at each locality, although Fægri thought the gradient slightly high. Extrapolation of Tapes shorelines from Stord to Etne and Halsnøy using the 0.6 m/km gradient correlates within 1 m at Etne and 2 m at Halsnøy with Tapes shorelines in these areas. Extrapolation to Bømlo is within 3 m of Fægri's 10 m Tapes line.

Complexity and inaccuracy of Holocene sea-level limits (Curray & Shepard 1972, Bloom 1972) prohibit an absolute isostatic compensation value. However the Fitjar Moraine delimits sea level at time of deposition and yields a relative isostatic-eustatic rise of 44 m since Older Dryas time (12,000 yrs. BP). Using Curray's 40 msl rise value for the past 9000 years (in Curray & Shepard 1972:16), an isostatic compensation of 81–85 m is calculated for Stord and a possible 14 m isostatic compensation through the last 6000 years. Applying the 44 m isostatic-eustatic rise of Stord to Milliman and Emery's curve (Milliman & Emery 1968), a relative isostatic uplift of 134–138 m since Older Dryas time (12,000 yrs. BP) is calculated. An uplift of 12 m also is calculated since Tapes transgression, based on Milliman & Emery's curve.

Discussion

The presented observations do not support the concept that Stord was glaciated during Younger Dryas time. Also Younger Dryas ice would have had to override the Huglo-Halsnøy-Sandvoll island chain at the mouth of Hardangerfjorden and project an ice tongue at least 1.5 km across the body of water, presently 320 m deep, separating Huglo from Stord, an improbable model of ice advance. It is more probable that ice flow was inhibited by topography in agreement with H. Holtedahl's (1967) ice dam effect of the Huglo-Halsnøy-Sandvoll island chain.

The 800±50 msl firn limit value calculated for Stord for Younger Dryas time combined with the isostatically reduced 665–700 msl upland elevations argue against the probability of active cirque glaciation on Stord during this time. However, the existence of nivation hollows on Stord during Younger Dryas time is probable.

The interpretation of late Pleistocene history on Stord is based on the relationship between the structure and form of glacial features on Stord with those of surrounding Sunnhordaland. The westerly striae on the Stord uplands in free position and the oldest westerly striae on the northwestern coast associated with the erosional features at Kalvanes indicate a westerly ice movement and correlate with that movement attributed to the maximum extent of Weichselian glaciation (Bøe 1949, Holtedahl, H. 1967, Mangerud 1970a, Aarseth 1971, Rønnevik 1971, Follestad 1972, Holtedahl, H. 1975).

Ice movement on Stord subsequent to the Weichselian maximum ice advance is suggested by predominant northwesterly striae on the northwestern coast, and southwestern striae on the southern coast. The striae pattern in conjunction with whalebacks at Årby and Vest-Bostad conform to coastal configuration and reflect topographic control of the ice advance. (A complete description of the striation localities is given in Genes 1974.)

Till fabrics with westerly component at Fugleviki and Sandvikvåg and till fabrics with northwesterly component along the western coast of Stord parallel striae directions. Clast inclusions with higher percentages of subround characteristics at eastern coast localities than along the western coast are interpreted as displaced beach deposits which have undergone little transport corresponding to Bergesen's (1970) allochthonous moraine classification. Tills on the western coast of Stord have higher percentages of angular and subangular clasts, and increase in subangular characteristics toward Fitjar. The difference in roundness characteristics is attributed to longer distance transport than those tills farther south, whereas the occurrence of inherited forms in all tills attest to short-transport, autochthonous conditions.

The close correspondence of till fabric and glacial striae azimuths, alignment of erosional features, and pattern of deglaciation sediments support the view that a single ice advance, the Older Dryas, occurred subsequent to the main Weichselian ice advance.

Submarine sediments interpreted as till at Fitjarviki and the associated scattered till deposits on surrounding islands are considered part of an end-moraine complex, the outer series of moraines designating the assumed extent of advance.

The fossiliferous clay till at Breivik containing broken shells dated to 12,860±250 (T-1171) places the ice advance as having occurred during the Older Dryas Stadial.

The pollen analyses (Table 1) of the fossil bearing till suggest an open vegetation cover predominantly of herb community. 63 percent of the AP consists of *Salix* and *Betula* with the rest of the pollen probably indicative of long-distance transport.

Chanda's (1965) pollen diagram of Brøndmyra at Jæren has Gramineae, Cyperaceae, *Artemisia,* and *Rumex* species extending from Oldest Dryas to Preboreal time.

The high percentage of *Salix* in the Breivik spectrum is abnormal and is associated with an oceanic climate (L. Selsing, pers. comm. 1973). The presence of *Koenigia islandica* indicates heliophilous, non-littoral, bare-soil, arctic-alpine conditions (Hulten 1950, Lid 1963, Danielsen 1969, Florin 1969). The *Artemisia* finds, also heliophilous, support this conclusion.

Although the pollen spectrum results cannot be interpreted with certainty, an arctic or tundra pioneer phase of vegetation is suggested, with vegetation placed probably in either zone 1a (Oldest Dryas) or zone 1b (Bølling) time.

Glacial drift in Svartavatn East, extending into Borgtveitdalen, suggests cirque activity postdating general deglaciation during Older Dryas time. Glacial striae at the tarn lip, implying ice flow down valley, support this view.

The areal distribution of 'dead ice' deposits in upland areas is suggestive of a combination of a rising firn limit and general deglaciation.

The difference in elevation of firn limits during Older and Younger Dryas times must have been minimal. Andersen (1965) subdivided the Trømso-Lyngen substage in the north into two glacial phases corresponding to Older and Younger Dryas times, with an intervening phase of less glacial activity. He was unable to determine any difference in firn limit elevation for the two time periods. The Younger Dryas firn limit value calculated for Stord is probably too high to account for cirque activity in Svartavatn East, whereas a slightly lower firn limit during Older Dryas time would account for cirque activity. This view is consistent with the interpretation of a rising firn limit and normal deglaciation to explain the general deglaciation of Stord during Older Dryas time.

Stord was completely covered with ice during maximum extension of the main Weichselian ice advance, approximately 20,000 yrs. BP, as inferred from faceting, grooves, and erratics on Midthamarsåto. The predominance of westerly striae along the northwestern coast of Stord, the whaleback field at Kalvanes, and the 'P' forms on the northern coast of Tveit are also considered to reflect this ice movement.

Studies of sediment characteristics between outer and inner Norwegian shelf areas (H. Holtedahl 1965), occurrences of striae and erratics in the Møre-Romsdal area (H. Holtedahl 1960), and seismic profiling of assumed terminal moraines on the Norwegian shelf (H. Holtedahl & Sellevoll 1971) suggest that the last ice sheet extended to the edge of the continental shelf and support the view that Stord was completely ice covered during the main Weichselian ice advance.

Anomalous striae with north-south component at Kalvanes and in the Leirvik area suggest an ice movement which possibly predates the main Weichselian westerly ice movement. Striae at Rekstern with north-south component subsequently crossed by striae with westerly component and the questionable striae with north-south component at Fonno also could represent this movement. These observations are in accordance with Undås (1949), who suggested an old, north-directed ice movement along the southwestern coast of Norway, but Andersen (1964) has disputed this ice movement.

Subsequent to the main Weichselian ice advance the ice retreated during the Bølling Interstadial to the head of Hardangerfjorden (Mangerud 1970a). The Fitjar substage ensued during the Older Dryas ice advance which was responsible for the deposition of the Fitjar Moraine.

Investigations of fossiliferous tills and subtill sediments around the Bergen area indicate that they have originated from widely different times (Mangerud 1970a, 1970b, Mangerud & Skreden 1971) and that marine sediments have been overridden or displaced by ice (Mangerud 1970b).

Mangerud (1970a), Aarseth (1971) Aarseth & Mangerud (1974) and Follestad (1972), mapping 'Ra' end moraines deposited during the Younger Dryas Stadial, show that these moraines around the Stord region extend from Herdla north of Bergen, follow mainland coastal configuration generally around Bjørnafjorden, and continue along the eastern coasts of Tysnes and Huglo and the northern part of Halsnøy. The glacial striations on Stord formed subsequent to the main Weichselian ice advance are assumed to have been formed prior to the Younger Dryas ice advance, as no evidence is available to indicate that the inland ice reached Stord during the Younger Dryas Stadial.

Sediments at Blomvåg, dated by radiocarbon at 12,700 ± 350 yrs. BP and 12,200 ± 350 yrs. BP, have been correlated with fossiliferous clay, dated by radiocarbon at 12,470 ± 150 yrs. BP (Mangerud 1970a), at Sandviken near Bergen. Mangerud described the movement of the glacier front as having retreated landward past Blomvåg prior to 12,700 yrs. BP (Bølling) with subsequent advance after 12,200 yrs. BP. during the Older Dryas Stadial.

The Fitjar Moraine is the first demonstrated morphological evidence of Older Dryas inland-ice advance in the Hordaland district, if the Breivik deposit is properly correlated with the Blomvåg and Sandviken deposits.

Conclusion

Stord retains morphological evidence of at least five phases of late Pleistocene history:

Evidence of the main Weichselian ice advance when ice was not influenced by topographic control manifested in the Tveit area where all roches moutonées trend 270°; the well-preserved 'P' forms north of Sandvikvåg and along the Fugleviki coast; bedrock knobs at Olstjøni overridden by westward-flowing ice; the roche moutonée field at Kalvanes, and the overrun cirque headwall at Svartavatn East.

The Fitjar Moraine complex in connection with dissected till deposits at Fugleviki, Kalvanes, and Vik associated with the Older Dryas ice advance.

Ice recession following Older Dryas glaciation manifested in stratified outwash deposits at Dafjorden, Nymark and Vik, the marine-outwash plain at Fitjar, kettle lakes, and outwash deposits along both coasts.

The final phase, difficult to distinguish from general ice-recession characteristics of Older Dryas glaciation, concerns evidence for the existence of a plateau ice cap. Radially distributed drainage channels, moraines in proximity to cirque mouths, protalus ramparts as at Tveitafjellet and Sæterbo, glaciated valleys with rock steps as at Steingilsheddi, the delta at Vatna, marine outwash at Jektivik and Borgtveit, and the 'dead-ice' topography on upland areas are all suggestive of ice-cap conditions on Stord following general deglaciation of Older Dryas time.

The 240 m.a.s.l. and 450 m.a.s.l. tablelands correlate with snowlines that fluctuated throughout Weichselian and probably Preweichselian time.

Evidence of ice flow and ice recession on and around Stord, in conjunction with radiocarbon dates and interpretation of submarine deposits, indicate that the Fitjar moraine is of Older Dryas age. Marine limits, correlations of surrounding area deposits, and estimated depression of the firn limit and glaciation limit all support this interpretation and also indicate that the Younger Dryas advance did not reach Stord.

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