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TILL AND MORaine EMPLACEMENT IN A DEFORMING BED SURGE — AN EXAMPLE FROM A MARINE ENVIRONMENT

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Abstract — The glacier Sefstrombreen in Spitsbergen surged across an arm of the sea between 1882 and 1886 and rode up onto the island Coraholmen. Marine and terrestrial geological observations and archive records show that the glacier advanced on a deforming carpet of marine mud which was eroded from its original location, transported, and smeared over the sea bed and Coraholmen as a deformation till. The glacier emplaced about 210^8 m^3 (0.2 km^3) of drift in the terminal 2 km of its advance in a maximum of 14 years, leaving a thickness of up to 20 m on Coraholmen, which was doubled in size as a result.

During the surge, subglacial muds were characterised by high water pressures, low effective pressures and low frictional resistance to glacier movement. Original sedimentary inhomogeneities permit fold structures to be identified, but repeated refolding and progressive remoulding produce mixing and homogenisation of deformation tills.

The surge was probably shortlived, and as the heavily crevassed glacier stagnated, underlying water saturated muds were intruded into crevasses and then extruded on the glacier surface. Reticulate “crevasse-intrusion” ridges on Coraholmen and the sea floor reflect the orientation of surge generated crevasses. Water and sediment was also extruded beyond the glacier at its maximum extent, to form extensive flows producing “till tongues” both on Coraholmen and the sea floor extending over 1.3 km from the glacier.

It is argued that subglacial deformation of pre-existing sediment will almost invariably be associated with glaciation of marine areas and that this process will not only produce deformation tills through remoulding of pre-existing sediments, but will also play a fundamental role in glacier dynamics. Criteria which permit glacial tills produced by such events from marine and glaciomarine muds are discussed. Copyright © 1996 Elsevier Science Ltd

INTRODUCTION

In recent years, the sedimentological and glaciological importance of deformation of sediments beneath glaciers has been widely recognised (e.g. Boulton *et al.*, 1974; Boulton and Jones, 1979; Alley *et al.*, 1986; Boulton and Hindmarsh, 1987; Clarke, 1987; Nesje and Sejrup, 1988). The process can clearly play a fundamental role in influencing the stability and dynamics of glaciers (Boulton and Jones, 1979; Alley *et al.*, 1986; Nesje and Sejrup, 1988) and can also produce considerable thicknesses of deformation till by remoulding pre-existing subglacial sediments (Boulton, 1987).

Glaciers which extend over arms of the sea, continental shelves and lake basins are particularly prone to generate deformation till by this mechanism. Normally consoli-

dated, fine-grained sediments (which may be proximal glaciomarine or glaciolacustrine sediments produced by sedimentation from suspension in the glaciers own outwash) when overridden by an advancing glacier, will be subject to an immediate additional load. Porewater pressures will increase instantaneously, but porewater expulsion will be slow, as a consequence of the relatively low permeability of fine-grained sediments (Boulton and Dobbie, 1993). Moreover, the glacier may be melting basally, thus recharging the sediment's porewater and further inhibiting consolidation. As a consequence, the effective pressure in the sediment will tend to be low, its shear resistance correspondingly low, and there will be a strong tendency for the sediment to undergo subglacial shear deformation in response to the drag force exerted by the overriding glacier.

Deformation will occur in sediments immediately beneath the glacier when

$$\tau_b \geq (p_i - p_w)\tan\phi + C \quad (1)$$

where (τ_b) is the shear stress at the base of the glacier, p_i the ice pressure, p_w the water pressure, ϕ the angle of internal friction and C sediment cohesion. (τ_b) will generally lie in the range 30–100 kPa for a typical glaciomarine mud where $\phi = 12^\circ$ and $C = 10$ –30 kPa. Thus, for deformation to occur, the effective pressure ($p_i - p_w$) must be less than about 100–400 kPa, equivalent to a load of 11 to 44 m of ice. It is clear from work in Antarctica however (Alley *et al.*, 1986) and from valley glaciers (e.g. Fountain, 1994) that even under ice pressures of the order of 8000 kPa, effective pressures in subglacial sediments can be as little as 50 kPa because of inhibition of water drainage. It has been shown (Boulton, 1979; Boulton and Hindmarsh, 1987) that where subglacial deformation does occur, it does so by simple shear in a horizon in which a very high void ratio and low-effective pressure is sustained by poor drainage and dilatation of the deforming mass. Boulton and Hindmarsh termed this the tectonic A-horizon in which sediment deformation can be approximated as that of a non-linearly viscous or a Bingham fluid in which very high horizontal discharge rates can occur ($200 \text{ m}^3 \text{ m}^{-1} \text{ a}^{-1}$ Breidamerkurjökull, Iceland — Boulton, 1987; 100 – $1000 \text{ m}^3 \text{ m}^{-1} \text{ a}^{-1}$ — Ice stream B, Antarctica — Alley *et al.*, 1986). This may be underlain by an unyielding lithified

stratum, or an unlithified B-horizon in which void ratios are much lower, effective pressures much higher, and in which only very slow elastic-plastic strains occur along shear joints parallel to the A/B interface.

The thickness of the A-horizon (t_A) is given (Boulton and Dobbie, 1993) by

$$t_A = \frac{\left(\frac{\tau_b - C}{\tan\phi}\right) - p'_0}{\left(\frac{\delta p'}{\delta z}\right)} \quad (2)$$

where p'_0 is the effective pressure ($p_i - p_w$) at the glacier sole, and $(\delta p'/\delta z)$ is the vertical effective pressure gradient, normally equal to about 10 kPa m^{-1} .

The horizontal flux of sediment in the deforming horizon (Q_A) will be

$$Q_A = \int U \delta z \quad (3)$$

(Boulton, 1996) where U is the mean velocity in the deforming horizon.

Where the glacier undergoes extensional flow, the deforming horizon will tend to thin, but as the thickness of the A-horizon must be maintained in order to sustain the velocity of the glacier, the A/B interface will descend, producing erosion by incorporating hitherto undeforming material into the deforming sediment (Fig. 1c). The rate

FIG. 1. (a–b) Subglacial stress, strength and strain conditions, (a) shows increasing sediment strength with depth, which is a linear function of effective pressure, and which eventually exceeds the constant value of shear stress at depth t_A , below which no further deformation will occur, (b) shows the distribution of strain rate in the deforming, dilating A-horizon, overlying a stable B-horizon. Strain rate depends on shear stress and sediment strength. (c–d) The conditions for erosion and deposition on a deforming bed. In (c), longitudinally or temporally increasing ice flux is associated with higher shear stress and increasing sediment discharge, causing material from the stable B-horizon to be added to the mobile A-horizon, thereby lowering the A/B interface and causing erosion, (d) shows the converse case of deposition resulting from longitudinal compression.

of erosion (E) will be:

$$\frac{E = \delta Q_A}{\delta x} \quad (4)$$

(Boulton, 1996).

Conversely, in a zone of compressive flow, there will be a rise in the A/B interface, and strongly deforming dilating A-horizon sediment will consolidate as it becomes part of the B-horizon and thereby deposited (Fig. 1d).

In view of the low permeability of marine or lacustrine muds, it is difficult to envisage how such deformation could be other than an almost universal phenomenon and fail to generate substantial masses of deformation till in zones of deposition, where glaciers extend over such sequences, although their recognition may pose problems (but see van der Meer, 1993). Indeed, Boulton (1990) has drawn attention to the way in which a glacier flowing along Kongsfjorden, Spitsbergen, produced tills with a mean thickness of 10 m along the centre-line of the fjord trough, where thick fine mud sequences were overridden and remoulded by the advancing glacier, and of only 3–5 m along the flanking strandflat where the glacier advanced over a sand and gravel sequence.

In this article we document the sedimentary and geomorphic consequences of a glacier advance over a marine mud sequence and suggest general implications for the interpretation of diamictos on continental shelves.

THE 1882/1896 SEFSTRÖMBREEN SURGE ONTO CORAHOLMEN

Coraholmen is a small island of about 8 km² composed of Carboniferous limestone bedrock lying at the confluence of northerly and northwesterly arms of Ekmanfjorden in central Spitsbergen (Fig. 2). The surface of bedrock does not rise above 10–20 m a.s.l. Bedding dips south at low angles, and a series of well-defined limestone scarps trend east–west (Fig. 3).

In the latter part of the last century, Sefströmbreen, a calving, tidewater glacier flowing into the northwestern arm of Ekmanfjorden (Fig. 2), advanced by 6–7 km between 1882 and 1896 onto the western side of Coraholmen and Flinholmen, another smaller island to the southwest (De Geer, 1910).

The geological result of this surge was to plaster against the western side of Coraholmen a mass of drift of up to 10–20 m in thickness which more than doubled the size of the island (Fig. 3). A similar mass was plastered onto the western side of Flinholmen. Bathymetric and seismic profiles between Coraholmen and Flinholmen, and between Coraholmen and the mainland to the west, show boundaries between highly irregular sea floor to the west and a smoothly rising surface to the east in precisely the positions mapped by De Geer (1910) for the glacier's maximum extent just prior to 1896 (Fig. 20). We presume this to reflect a thick (10–20 m) drift mass resting on a

FIG. 2. Map showing ice marginal positions of Sefströmbreen during the 1882–1896 surge and during recent years and the extent of morainic deposits on Coraholmen, Flinholmen and the nearby mainland. It shows how the limestone island of Coraholmen (shaded) has doubled in area as a consequence of the smearing of till (triangular ornament) against its western shore. Data mainly from de Geer (1910); Dineley and Waters (1960). Heavy arrows show the line of the section shown in Fig. 4.

FIG. 3. Air photograph of Coraholmen showing the contrast between the limestone scarps on the eastern (right hand) side of the island and the till on the western side. Within the till area, the zone of parallel ridges is particularly well seen at the eastern extremity of the till mass in the centre of the island (marked R). The zone of basins to the west is marked by lakes. The narrow, rectilinear ridges in this area are believed to be crevasse fillings produced during stagnation. The ice contact slopes at the boundary of the two zones is marked by IC.

smooth bedrock surface. At a rough estimate the glacier emplaced about $2\text{--}10^8\text{m}^3$ (0.2 km^3) of drift in the terminal 2 km of its advance in a maximum of 14 years.

De Geer documented the surge carefully. In 1882 the northern terrestrial margin of Sefströmbreen was a gentle slope with an outwash plain beyond it, and the marine margin was a steep ice cliff (Figs 2 and 4). Bardebreen seems to have been the most active glacier stream flowing into this ice cliff, though Sefströmbreen lay at a higher elevation. De Geer describes Bardebreen as having a looped medial moraine, suggesting that it had surged at some time previously.

By 1896 the front of Sefströmbreen had advanced by 6.5 km from the 1882 ice cliff location and retreated by 1 km from its maximum extent (Figs 2, 4 and 5a). From the pattern of medial moraines, it was clear that Sefströmbreen itself had surged by that time, pushing Bardebreen to one side. The higher elevation of Sefströmbreen in 1882 may have reflected the down-glacier passage of a wave front which initiated the surge when it reached the terminus.

By 1898 there had been relatively little recession. The recession in subsequent years is shown in Figs 2 and 4.

The slow speed of this recession probably reflects the shallowness of the Sefström Bay (Fig. 4) which does not permit calving of large icebergs.

De Geer described the surface of Sefströmbreen in 1896 as rugged and full of fissures. Trevor-Battye (in Conway, 1897) described the overriding of Coraholmen in the same year: "More than half the island is now overspread. Without moraine, without dirt or discoloration, the glacier is pouring over it, and great seracs lie there, separated only, or barely separated, from the flowers and grasses by the clear stream their drip has formed". By 1908, a channel had appeared between Sefströmbreen proper and a large mass of dead glacier ice left grounded on Coraholmen and the shallow shoal to west (Figs 2, 4 and 5c), with a channel of jagged outline between the two, bordered by ice cliffs on both sides.

By 1910 the visible front of the stagnant glacier on Coraholmen had retreated by 600 m to the west of its maximum position, but its surface had changed. It had clearly melted down sufficiently to reveal morainic debris in the lower part of the glacier (Fig. 5c). Lamplugh (1911) described it as showing "rude confused ridges running in general parallelism with the outer margin of the moraine, but sharply broken by many cauldron-shaped hollows... ranging from a few yards to 200 to 300 yards in diameter; their sides, sometimes 30 or 40 feet high, are steep and crumbling... and the presence of oozing muddy pools in them confirmed the supposition that they were due to the melting of patches of ice concealed under the moraine".

A photograph by Strahan (in Lamplugh 1911, plate 24) of the dead glacier shows a vertical, crevassed margin in which drift appears to fill the open bases of crevasses (cf. Fig. 5a). Strahan writes of this that "the ice was clean from top to bottom, but that it was underlain by boulder-clay at this spot, and overlain by it close by was clear". On the glacier surface he wrote "this ice... presented a gently undulating surface traversed by irregular crevasses. Each crevasse had been filled up with boulder-clay and the melting of the surface had left these casts of crevasses projecting like raised walls, or still more like igneous dikes". Such crevasses can be observed in photographs in the De Geer archive in Stockholm (van der Meer, 1992).

There are strong similarities with Gripp's (1929) descriptions from other Spitsbergen glaciers of "clay walls" (Lehmmauern) which he believed to have been subglacial debris pressed up crevasses which reached to the glacier sole.

Although Lamplugh believed that the drift emplaced on Coraholmen was transported englacially, the observations of Trevor-Battye, Lamplugh and Strahan, together with modern knowledge of the patterns of interaction between drift and melting ice, lead us to believe that the drift was largely transported subglacially onto Coraholmen and that ice became isolated or buried within drift after the drift had been squeezed up crevasses onto the glacier surface.

When Gripp (1929) studied the island in 1927, ice was no longer visible at any point, but because of impassable muddy areas (in spite of a dry summer)

FIG. 4. Flow-parallel, longitudinal section through successive stages of the Sefströmbreen surge. It advanced by six kilometres between the years 1882 and 1896, a minimum frontal advance rate of 460 m/year. Note the channel which had opened up between a stagnant ice mass on Coraholmen and the main body of the glacier by 1908. The sea bed profile was obtained by echo sounding. Note the rapid retreat of the glacier since 1958 as it withdrew from the shoal. Sources of data are referred to in the text. The 1921 position is derived from van Oordt (1923).

he assumed that buried glacier ice still existed. Apart from those areas inundated by lakes, the drift on Coraholmen is now dry, although some buried ice might still be present.

LANDFORMS

The drift landscape now seen on Coraholmen (Figs 6 and 7) is very similar to that described by Lamplugh (1911), Cole (1911–1912), Ahlmann (1912) and Gripp (1929). These descriptions, taken together with Trevor-Battye's description (in Conway, 1896) of the clean glacier advancing over Coraholmen, strongly suggest that the landscape is essentially a subglacial one reflecting the final form of the glacier/bed interface, and that any subsequent dead ice melting has had a minor effect on its evolution. The stability of the drift landscape on Coraholmen is remarkable in a maritime arctic area where typical stable slopes in muddy sediments are, at most, a few degrees. On Coraholmen, drift ridges, which may be the remnants of Gripp's *Lehmmauern*, still survive with slopes of 40°. Freshly-dug sections show overconsolidated sediment with undrained shear strengths between 100 and 220 kPa. This may partly explain the high slopes but the crystallisation of salts due to drying out of glacially transported marine sediments may also give them unusual cohesion.

If this landscape is a subglacial landscape, it is a striking contrast to typical drift landscapes produced beneath active ice, which are generally strongly lineated and streamlined.

Two morphologically distinctive zones are apparent in the drift covered area of Coraholmen (Figs 3, 8, 9 and 10). At the eastern extremity, a zone 200–300 m wide comprises a series of up to 12 low amplitude ridges (<2 m) parallel to the drift margin. To the west of this zone, there is a series of basins bounded by rectilinear ridges.

The Zone of Basins Bounded by Rectilinear Ridges

These occur on two scales, one in a series of about 1 km in diameter and another between about 100 and 250 m in diameter (Figs 3 and 8). They are up to 10 m in depth and the smaller ones are clearly the "cauldron-shaped hollows" of Lamplugh (1911). Within the larger depressions, and on the ridges which separate them, are a series of narrow (<20 m), steep sided (up to 40°) ridges up to 6 m high. They are straight in plan or are made up of series of straight segments. They frequently form branched systems and often entirely surround the smaller (<250 m) basins. They are the ridges termed "*Lehmmauern*" by Gripp and subsequently described by Sharp (1985) from Iceland. From their rectilinear plan and interrelationships, and from Strahan's and Lamplugh's description of the glacier margin and surface on Coraholmen, we suggest that they represent originally subglacial material pressed up crevasses which penetrated the surged lobe of Sefströmbreen to its base. We suggest that the small-scale basins which they frequently enclose represent areas from which sediment has been squeezed towards the crevasses.

If this is so, the pattern of "crevasse fillings" must reflect the pattern of crevasse of Sefströmbreen during

FIG. 5. (a) Photograph from the de Geer archive, Stockholm, showing the margin of Sefströmbreen in 1896, with Coraholmen to the right. The ice cliff shows till dykes penetrating from the bed to the surface of the glacier (An example is indicated by the arrow). Down-melting of the glacier surface has already released much till onto the glacier surface, suggesting that the culmination of the surge significantly predates 1896. (b) Photograph from the de Geer archive, Stockholm, showing till ridges emerging from the surface of the stagnating ice mass on Nansenbreen in 1908 or 1910 (termed *Lehm-mauern* by Gripp, 1928) and clearly equivalent to those on Coraholmen. It is suggested that they represent subglacial deformation till masses which were squeezed into crevasses during post-surge stagnation of the glacier, and released onto the glacier surface as this subsequently melted down. (c) Photograph taken in 1908, from the eastern side of Coraholmen across the channel which had opened up between the stagnant ice mass and the main glacier (see Fig. 4). The slope apparent on the surface of the glacier suggests that it was active rather than stagnant. The ridges on Coraholmen are till dykes projecting above the glacier surface.

the final stage of its surge onto Coraholmen, and the pattern of fracturing and fissuring on the glacier surface referred to by De Geer and Trevor-Battye.

Figure 9b shows rose diagrams from each of five major lobe-shaped depressions on Coraholmen (Fig. 9a). These are bounded to the north and the south by relatively high ridges and plateaux, and to the east by the zone of parallel ridges. Lobes 2 and 3 show crevasse fillings oriented predominantly in an east–west direction, whilst lobes 4 and 5 show a distinct northeast–southwest orientation. During a surge, we expect basal friction to decrease to near zero. As the surging glacier overrode Coraholmen, we would expect longitudinal compression to develop, in which circumstances, crevasses would develop parallel to the principal compressive stress direction. In view of the known disposition of the glacier on Coraholmen at the culmination of the surge, it seems most likely that the crevasses reflect contemporary directions of longitudinal compression and transverse extension, suggesting dominantly west–east flow, but with a strong northwest–southeast-component in lobe 4.

The lobes shown in Fig. 9a show well-defined re-entrant angles at the eastern margin of the zone of basins. From these re-entrants stem the ridges and plateaux which laterally separate the lobes. It will be suggested in the next section that the steep scarp at the eastern margin of the zone of basins is an ice contact slope which formed at the maximum extent of the Sefströmbreen surge, and thus that the lobes it defines reflect glacier lobes. Lobed frontal margins tend to develop as a result of subglacial interlobe ridges, and thus we must conclude that the interlobe ridges on Coraholmen were not produced during the phase of stagnation and extrusion which formed the crevasse fillings, but that they existed beneath the active glacier and were responsible for the development of a lobate glacier margin. There are two obvious possible origins for the interlobe ridges: that they are cored by extensions of the prominent limestone scarps which occur beyond the drift mass on Coraholmen (Fig. 3); or that they represent subglacial drumlins which coincided with low velocity zones (Boulton, 1987), which may have been deformed slightly during the last stage of the surge. In both cases, enhanced drag along the line of the ridges would make these relatively low velocity zones and produce co-linear re-entrants in the ice margin.

Figure 10 shows measured profiles across Coraholmen in the zone of basins, normal to the direction of glacier flow over the island. The distinction between the zone of basins and the zone of parallel ridges is very clear (Fig. 8). If we assume that the original subglacial surface produced during the surge was relatively smooth and streamlined on a scale larger than the largest boulders, as most sediment surfaces beneath active glaciers are, and that the dominant roughness is due to local patterns of sinking and extrusion, we can estimate that a total mass of $2 \times 10^8 \text{ m}^3$ of mud was extruded along crevasses on Coraholmen.

The Zone of Parallel Ridges

The zone of parallel ridges at the easterly extremity of the glacially-emplaced mass is up to 300 m wide (Figs 3, 8 and 9a). The ridges are up to 2 m in height and in most places are closely spaced, with inter-crest distances of no more than 5 m, although in some cases up to 20 m. They are disposed in lobate groups in which the lobes often become more pronounced distally. They are generally bounded at their western extremity by a westerly facing scarp up to 10 m in height, which separates them from the zone of dead ice basins.

An obvious possibility is that the ridges are series of individual frontal push moraines produced at the extremity of the Sefströmbreen surge. There are reasons why this is unlikely. They are unlike individual push moraines produced during small, winter readvances of a glacier undergoing general retreat. These tend to be discrete, asymmetric, mutually cross-cutting ridges superimposed on a fluted surface (Boulton, 1986). On Coraholmen, sequences of contiguous ridges make up the whole surface in many places. They tend to have symmetrical and cross-cutting relationships and flutes are absent.

We therefore suggest that the prominent scarp at the western extremity of this zone represents an ice-contact scarp which marked the maximum extent of the glacier, and that the parallel-ridged drift mass beyond it was emplaced by some other means. The only mechanisms which we are able to suggest are that the whole mass was pushed in front of the surging glacier, or that it was extruded from beneath the glacier at the end of the surge as it sank into the water soaked-sediment beneath it, and that the ridges formed during flow or pushing of the mass. Evidence for the precise process of emplacement will be presented below. The tendency for the small ridges on the surface in this zone to bow more acutely away from the glacier with distance from it is explained by gravitational creep at the surface away from the ice contact zone.

The conclusion that the glacier terminus at its maximum extent was located along an ice contact slope at the western margin of the zone of parallel ridges is supported by the fact that a series of dry, stream-cut valleys originate at or near this scarp (Fig. 8). These presumably carried meltwater from the glacier front at the maximum extent of the surge when a continuous glacier surface sloped eastwards from the accumulation area of Sefströmbreen and the pressure gradient in the glacier maintained an eastward subglacial water flow. At this stage, the rate of water expulsion from the extruding till in the earliest, most rapid, phase of consolidation of water-soaked subglacial till, would be at a maximum.

Subsequently, as Sefströmbreen broke up into isolated stagnant ice masses on Coraholmen, the drainage catchments became very much smaller, and local drainage basins which were only able to feed small streams developed on the stagnating surface of the glacier. Small washed gravel and sand stripes which are oriented as they would be if they produced by drainage

FIG. 6. A view towards the northwest from the ice contact slope on Coraholmen across the zone of basins. Note the steep slopes.

from stagnant ice in the centres of basins are ubiquitous in the zone of basins.

SEDIMENTOLOGY

A detailed study was made of a series of sections on the south and west coasts of Coraholmen where wave erosion and cliff falls have exposed sections through the drift deposits. At no point was bedrock observed beneath the sediments. Several distinctive facies occur:

- (a) Red silt- and clay-rich diamicton with a relatively small clast frequency (Fig. 11a). Clasts are subangular to subrounded, many of the latter being striated. Fragmental or complete mollusc shells occur.
- (b) Red, silty-sandy diamicton with a high clast frequency, which is sometimes sand-rich (Fig. 11b).
- (c) Thin layers and lenses of red-brown sand with mollusc fragments.
- (d) Green silty clay.
- (e) Green sand and gravel which frequently contain large quantities of the alga *Lithothamnium* (Fig. 11c). Molluscs (dominantly *Mya truncata*) are often abundant. Lithologies (d) and (e) are usually intimately associated.

Grab samples taken from the sea to the east of Coraholmen show a lithology similar to (a) above, though

stoneless. This probably reflects the fact that access to this area by Sefströmbreen icebergs is limited by the shoal across the Sefström Bay, where maximum water depths are about 10 m. Inside this shoal, significant clast frequencies are found in grab samples, presumably as a result of iceberg transport.

The clast-rich diamicton (b) is very similar to the till currently accumulating on the surface of the northern arm of Sefströmbreen although the silt/clay content is frequently higher on Coraholmen than found on Sefströmbreen (Boulton and van der Meer, 1989).

We presume that the red sands, often containing molluscs, reflect the action of strong bottom currents.

The green sediments presumably owe their colour to the reducing conditions produced by high concentrations of *Lithothamnium* and the molluscs with which they are associated. A limited grab sampling programme failed to find similar *Lithothamnium*-rich sediments on the modern fjord floor. It seems most likely that they formed on the limestone-bedrock shallows around Coraholmen before the glacier surge imported large quantities of mud into the area.

The "green" facies is concentrated on the eastern side of the drift belt and at the extreme eastern end of the sections exposed on the south coast, where "green facies" comprise about one fifth of the total sediment exposed (Fig. 11c). They were only found in very minor quantities beyond 400–500 m west of the eastern margin of the drift mass. We presume that

FIG. 7. A narrow, steep-sided 'crevasse-filling' in the zone of basins, Coraholmen. It is a former till dyke which had intruded through the ice along the line of a crevasse (cf. Fig. 5c). The distribution of such forms is shown in Figs 8 and 9.

GLACIAL GEOMORPHOLOGY OF CORA ISLAND, WEST-SPITSBERGEN

FIG. 8. The glacial geomorphology of Coraholmen.

FIG. 9. (a) A map showing the inferred distribution of glacier lobes on Coraholmen. It differentiates between the zone of parallel ridges (referred to as squeeze ridges), the zone of basins (chequered ornament), and the plateau-like forms which tend to separate lobes. The arrows show the preferred orientations of crevasse fillings within each lobe. (b) Rose diagrams showing the orientations of crevasse fillings in each of the lobes.

this reflects an original location of green facies on the shallows around Coraholmen which are thus concentrated in the most easterly of the displaced drift masses.

STRUCTURAL GEOLOGY

The southern coast of Coraholmen gives a relatively well-exposed section across the whole of the Coraholmen sequence, roughly parallel to the direction of tectonic transport (Figs 12 and 13). At its eastern extremity (0–100 m) the coast intersects the zone of parallel ridges. The rest of the south coast intersects the zone of basins. Although most basins truncated by the coast have their bases at about beach level, the structures beneath some are exposed. Several intervening crevasse fillings intersect the coast, and their internal structures have been studied.

There are three principal deformational styles in the south coast sections:

- A Flat-lying, highly attenuated multi-phase isoclinal folds (Fig. 12, sections 1 and 4; Fig. 13). These appear to be the primary fold set, and have been refolded by B_1 and B_2 folds.
- B_1 A-folds have been refolded in crevasse fill ridges into anticlinal folds with near vertical axial planes (Fig. 12, sections 2 and 3).
- B_2 A-folds have been refolded in the zone of parallel ridges into "mushroom folds" with near vertical axial planes (Fig. 12, section 6; Fig. 11c).

We suggest that A-folds are the primary folds generated during subglacial shearing when the surging glacier smeared sediment onto Coraholmen. B_1 -folds are regarded as the product of subsequent extrusion of subglacial material into crevasses, whilst B_2 folds

FIG. 10. Topographic transect across the surface of Coraholmen showing the cross sectional form of crevasse fillings.

represent extrusion of subglacial material into the proglacial zone which was then pushed in front of the advancing glacier.

A-Folds — The Product of Subglacial Shear Deformation

Section 4 (Fig. 12)

At this site, deformation style can be reconstructed from the distribution of relatively fine-grained clayey-silt and stoney sandy-silty diamicton (facies b) horizons within a less stoney and finer grained silty diamicton (facies a). Careful observation was required to establish structure, as the lithological distinctions, though real, are not prominent. Several fold closures were plotted, and individual horizons lensed out locally. Even the limited information available however, suggests very strongly the occurrence of a series of flat-lying nappe structures.

The folds are highly attenuated and indicate strong stretching. Fold closures were infrequently found. Although the sediments at the right-hand end of the section only show planar laminae, we suspect that they reflect attenuated fold limits.

Section 5 (Fig. 13)

The most complete reconstruction of A-folds was made at Section 5. This coastal section exposed sediments in the middle of a basin. Being relatively far from the sites of intrusion into basal crevasses, only A-folds are exposed. Although folding is complex, fold components are relatively easily identified by *Lithothamnium*-rich horizons (facies e). These always appear to separate facies (b) from facies (a). Moreover, small boulders within this horizon tend to be encrusted on one side only. We take this to be the upper side, which, always coinciding with the facies (b) side, is an excellent means of establishing the direction in which the sequence becomes younger. Using this evidence it has been possible to reconstruct the topology of folding and faulting in such a way as to identify several phases of overfolding and overthrusting, with intervening extensional episodes have contributed to the exposed structures (Fig. 14).

- (1) (Fig. 14a) Sefströmbreen advances over a *Lithothamnium* (facies e) surface, which is underlain by earlier glaciomarine mud (facies a). Transported beneath the

glacier in a continuously deforming layer is a till (facies b) comprising far-travelled remobilised glaciomarine muds.

- (2) (Fig. 14b) Deformation beneath the overriding glacier produced a large fold (C_1) involving all three sediment units (a, b, e).
- (3) (Fig. 14c) A phase of longitudinal stretching and vertical thinning occurs as a result of simple shear strain.
- (4) (Fig. 14d) A small thrust plane develops (C_2).
- (5) (Fig. 14e) A large fold (C_3) refolds the earlier fold. This probably developed immediately after C_2 .
- (6) (Fig. 14f) A further phase of longitudinal attenuation occurs.
- (7) (Fig. 14g) Another fold (C_4) develops, which re-folds earlier folds (C_1 and C_3). We suggest that this slightly pre-dates the overthrust fault (C_5) after which the entire fold/fault stack is partially carried over a till mass (facies b).

A further till mass which overlies the fold/fault stack is much thicker than the till units folded within the sequence. It is therefore suggested that much of it was emplaced by subglacial deformation after formation of the underlying folds, and after the fold/fault stack had come to rest. We suggest that the *decollement* surface rose until it was within the upper till unit.

It is possible to estimate the stretching which took place during tectonic transport if we assume that stretching was achieved by simple biaxial deformation with longitudinal extension and vertical thinning, then, $t - dt = l + dl$, where l is the length of a given element and t its thickness.

If some parts of a bed of initially uniform thickness are vertical and other parts horizontal, the vertical part will appear to thicken and the horizontal part to become thinner. The original thickness (t_0) will be:

$$\frac{t_0 = t_v + (t_v - t_h)}{2} \quad (5)$$

where t_v and t_h are the thickness of vertical and horizontal elements.

If we take the last fold, C_4 in Fig. 14g, and measure the *Lithothamnium* bed thickness in the hinge and on the limbs, we can estimate the thickness of beds (t_3) before

FIG. 11. Deformation till facies exposed along the southern shore of Coraholmen. (a) Lithofacies a: a red, silt- and clay-rich diamicton with a relatively small clast frequency. Clasts are subangular to subrounded; many of the latter being striated. Fragmental or complete mollusc shells occur. By comparison with sediments currently accumulating in the basin in front of Sefströmbreen, it is suggested that this is a remoulded glaciomarine mud. The shear fractures at this site reflect the last stages of ice movement. (b) This shows lithofacies b overlying lithofacies a. Facies b is interpreted as reflecting mixing of glaciomarine mud with debris derived from glacial erosion of bedrock and released by melting of the basal ice within which it was contained. Note the shell at the top of facies a. (c) Mixture of facies e and d (light tones) comprising *Lithothamnium* and mollusc-rich sand and gravel, and green silty clay, respectively, overlain by the red diamicton of facies a. All were originally marine sediments, transported by subglacial shearing and deposited as deformation till. Facies e formed on shoals immediately west of Coraholmen and facies a was deposited in relatively deep water further west. This site lies in the zone of parallel ridges, largely composed of sediment extruded from beneath the glacier at the end of the surge, which then flowed over the island. The folds at this site show early, isoclinal folds formed subglacially, and later broader folds which formed during proglacial flow.

any post C_4 extension, and therefore the post C_4 -change in thickness. By subtracting this change from the thickness of *Lithothamnium* beds in the hinge and limbs of the C_3 folds, any extension between C_3 and C_4 can be established, thus:

$$\frac{t_3 = t_4 + t_{4v} - t_{4h}}{2} \quad (6)$$

$$\frac{t_2 = t_3 + [t_{4v} - (t_3 - t_4)] - [t_{4h} + (t_3 - t_4)]}{2} \quad (7)$$

The principal phases of extension, E_1 and E_2 , occurred between the compressional folding phases C_1 and C_2 , and C_3 and C_4 respectively.

The results are shown in Fig. 15, which shows the inferred shortening associated with compressive phases and the extension associated with phases of simple shear. It also shows the changing length of section occupied by the marker *Lithothamnium* horizon between the initiation of folding and its emplacement in its present position. Sliding along a based *decollement* is assumed not to occur. The length of the bed before folding, assumed to be the original length, was 14 m, and it was located entirely west of the present section. It was initially shortened in a phase of folding (C_1) then lengthened in a phase of extension (E_1) to 22 m. Two phases of local compression (C_2 and C_3), possibly related to a zone of irregularity or roughness on the underlying *decollement*, then reduced the bed's extent. A major phase of extension (E_2), considerably increased its extent from 17 to 47 m, before two final phases of compression, both probably associated with riding up of the fold/fault stack over a wedge of till, shortened the bed to 33 m.

We suggest that longitudinal extension and vertical thinning will be associated with generalised shear strain in sediments beneath the glacier. Moreover, a surge of a glacier terminus is associated with loss of frictional resistance, and because of restraint at the proximal extremity of the surged mass, longitudinal tensile forces will dominate within it. We suggest that longitudinal compressive strain, reflected in folding and overthrusting

is likely to result from local bed irregularities or inhomogeneities (Boulton, 1987). Note that where compressive structures are paired (C_2 - C_3 , C_4 - C_5) the distal structure forms first.

Extension occurs by relative advance of the front of the bed and compression by relative advance of the rear. It shows a snake-like forward motion (Fig. 15).

Section I (Fig. 12)

The most obvious structure at this site is a flat-lying isoclinal overturned synform showing tectonic transport towards the east. In the core there are green facies, gravels which include algal material, sands and muds. The limbs appear to be sheathed by diamicton, whilst a silty clay occurs in an antiformal core to the west of the synform. We suggest that the silty clay has behaved in a highly incompetent fashion and has been squeezed out of the antiformal core. The sandy-gravelly beds in the synform core would have provided a relatively stiff obstacle to deformation of the subglacial mass and as such have been responsible for the large overfold in which shortening of the order of at least 40 m is represented. This is a typical subglacial fold (Boulton, 1987), a highly attenuated antiformal fold with a more open underlying synformal fold facing up-glacier.

There are several signs however of earlier phases of folding and deformation, such as the plications of the green mud/diamicton interface in the synformal nose, repetition of the gravel horizon in the same nose, and tectonic lamination of sand and clay in the synformal core. The lamination of the silty clay in the antiformal core may also have been produced tectonically (i.e. repetitive interfolding of the beds).

The fine-grained horizons at this site, as elsewhere, are well-jointed (see also Fig. 11b). In lower horizons the joints dip towards the east at 20° - 30° , but generally steepen and become overturned upwards. The joints at lower horizons have orientations similar to those expected from plastic slip-line fields under simple shear. This may indicate that at a late stage during shear displacements of the sediment, and at places where coarse, permeable sediments permitted sufficient water to be squeezed from the deforming mass, significant consolidation occurred, which caused discrete failure planes to develop. During the

FIG. 13. A fold complex exposed on the southern shore of Coraholmen within the plateau zone between lobes 4 and 5 (Fig. 12). A distinctive *Lithothamnium* horizon of facies e provides a valuable strain marker.

FIG. 14. A schematic reconstruction of the evolution of the fold illustrated in Fig. 13. Phases of compression (b, d, e, g) and extension (c, f) occur.

very last stages of shear deformation these pre-existing planes were rotated in the upper part of the sediment.

B₁ Deformation Produced by Intrusion into Crevasses of Sub-Glacial Sediments

Section 3 (Fig. 12)

At this site a well-defined crevasse intrusion between 6 and 7 m in height intersects the coast and permits its internal structure to be studied. The section trends at 70° to the crest line of the ridge. The ridge contains two lithologies; a red silty diamicton (lithology a) and a red and green sandy horizon with occasional *Lithothamnium* skeletons (lithology e). The mud is unusual in being relatively well-stratified. Some of this is due to thin sandy partings.

The predominant structure comprises two high angle anticlines with an intervening syncline. The axial planes of the anticlines dip outwards and away from the ridge crest. It appears to reflect intrusion into the base of crevasses of subglacial sediment which had already suffered strong shear deformation. The existence of two counterpoised folds may reflect flow towards and into the crevasse from both its flanks.

In the centre of the feature there are bivalve shells

which show evidence of shear between the two valves (Fig. 16). This occurred during primary, A-folding stage.

Section 2 (Fig. 12)

This section shows a muddy diamicton (lithofacies a) and a sandy diamicton (lithofacies c). The deformation during the A-folding stage at this site is very similar to section 1. There is small scale A_i-folding that has been refolded by a larger A_{ii}-fold. This early folding has then been refolded during the B₁-folding stage when material was squeezed up into a crevasse-fill ridge.

Pushing and Extrusion in Front of the Surging Glacier

Section 6 (Figs 12 and 11c)

The western end of section 6 is formed by the prominent scarp interpreted as an ice-contact front which marks the easternmost extension of Sefströmbreen during its surge. Sediments beneath the lower part of this slope show a strong tectonic lamination parallel to the slope. *Lithothamnium* facies (e) delineate mushroom-folds adjacent to the upper slope. As the fold limbs are traced downwards they become attenuated (Fig. 11c), suggesting strong longitudinal compression at depth, which produced upward flow into an unconfined position.

Well-defined overturned folds, their lower limbs cut by thrusts, occur between 20 and 30 m. Given that the glacier front did not extend further than the westernmost end of the section, we interpret these folds as evidence of pushing of this drift mass by the glacier.

A little further to the east, beyond a large wedge of diamicton, another mushroom-shaped box-fold occurs, which we interpret as indicating the drift mass's opportunity to expand upwards under longitudinal compression in the absence of a restraining ice roof (Fig. 11c).

We suggest that as Sefströmbreen surged over the soft glaciomarine muds in the bay to the west of Coraholmen, it extruded mud beyond it and pushed this forward as it advanced, and that much of section 6 comprises such a proglacial push moraine. However, as the surge stagnated, and hitherto dilatant, underconsolidated, subglacial muds were able to consolidate, extrusion not only drove more sediment up any crevasses which still remained open, but also extruded more material into the proglacial zone. Thus, although the mushroom structures adjacent to this ice-contact slope could reflect upward relief of horizontal compression produced by forward movement of the glacier, it could equally well reflect late-stage extrusion due to the weight of the glacier alone. We suspect that both processes are likely to have been at work.

Although some parallel ridges on the surface coincide with deep-rooted folds beneath, many are unrelated to the deeper structure of this zone, and are interpreted as gravity flow features formed during and/or after emplacement of the push moraine. This explains the tendency to a distal increase in the curvature of ridges in many places, which we suggest are flow lobes (Fig. 8).

THE CORAHOLMEN SEDIMENTS AS DEFORMATION TILLS

Although only facies (b) along the south coast of Coraholmen is identical to a sediment which is currently accumulated elsewhere as a till, we suggest that all the sediments exposed along the south coast facies (a)–(e) are tills. All show the consequences of pervasive shear deformation in which the geometrical interrelationships of the original sedimentary components has been become so highly distorted as a consequence of frequent folding events and intervening attenuation, that the material has become a new sediment through this mixing process. We apply the term deformation till to it. Although mixing in many parts of the mass is advanced, many original lithological distinctions are retained. This is particularly true of coarser-grained, less competent sediments.

Supporting evidence that existing lithologies have undergone major remoulding comes from studies of thin sections. Microstructures can be used to distinguish primary glaciomarine sediments from glacially remoulded glaciomarine sediments through structures which reflect their different stress histories. Glaciomarine sediments have been deposited under low or zero stress conditions, so that clay minerals tend to be randomly oriented and consequently fail to show birefringence under the microscope (e.g. Fig. 17a). Where such sediments have been overridden and remoulded by subglacial shear stresses, even under conditions of very high water content (Fig. 17b), clays are reoriented to show highly birefringent plasmic fabrics (van der Meer, 1993). Structures, such as small-scale folds, fractures, fissility and dewatering structures are also common in these sediments.

RESULTS OF THE SEFSTRÖMBREEN SURGE ON THE SEA-BED

During summer 1986 the bay to the east of Sefströmbreen and the passage between Coraholmen and Flinholmen was investigated from research vessel G.A. Reay. A transit sonar was used to examine the sea bed morphology produced by the surge and pinger and air-gun used to establish the thickness of any glacially-remoulded sediments emplaced by the surge. The latter was particularly important as the base of the surge deposits is not exposed on Coraholmen and Flinholmen.

Figure 18 shows part of the transit sonar record. It reveals a well-defined reticulate pattern of ridges and intervening hollows. As longitudinal and transverse scales of the sonographs do not match, tracings of the crest lines of ridges were digitised to produce the scale-corrected interpretations shown in Figure 19, which also shows the ridge patterns on Coraholmen itself. The sea floor ridges form closed, polygonal patterns in which individual polygons are as much as 150 m and as little as 20 m in diameter. Particularly long ridges may form parts of many polygons. They clearly show a scale and pattern similar to the ridges on Coraholmen, although the terrestrial pattern is much less complete. We presume this to result from two factors:

- (1) The ablation of the stagnant glacier ice from the sea area by calving, rather than surface melting as on land, would be less destructive, avoiding the erosional activity of meltwater streams, the perennial wetting and drying of the surface and the resultant instability associated with decay of a stagnant ice mass.
- (2) Periglacial processes, which lead to instability, such as seasonal freezing and thawing, wetting and drying, have continued to effect the terrestrial areas since disappearance of the glacier ice.

FIG. 15. A reconstruction of the net longitudinal extension (which must have been accompanied by vertical thinning) of the *Lithothamnium* bed illustrated in Figs 12 and 13. From an arbitrary original length shown at the top left of the Figure, extensional (E) events and compressional (C) events ultimately produce the net extension shown at the bottom right. Simple shear deformation is essentially an extensional process, although the results of compression are visually most obvious.

Compression is assumed to occur because of local obstacles to simple shear flow.

FIG. 16. Molluscan bivalve shells within deformation till of lithofacies a. In each of the two molluscan specimens, both valves remain associated. In the smaller example, they are still hinged. In the larger example, the upper, ventral valve has moved to the right in relation to the lower, dorsal valve. In this case, hinge and integument have become detached. We suggest that both examples remained hinged whilst being transported in the deforming sediment mass for at least 2–3 km from their nearest probable original locus of primary deposition as glacial marine mud. Shear displacement between the valves of the larger specimen is suggested to have occurred as the till drained and consolidated at the end of the surge event.

Figure 20 shows part of the pinger record across the submarine moraine belt south of Coraholmen and Fig. 21 its interpretation. Individual ridges which form part of the polygonal network shown in Figs 18 and 19 are up to 12 m in height, generally higher than those on Coraholmen. We attribute this to the degradation processes referred to above which cause ridge collapse and basin filling on land.

A prominent reflector underlies the submarine moraine ridges, some 20 m below their mean surface level. We presume this to be a major *decollement* horizon separating glacially-remoulded marine mud which has been through a phase in the tectonic A-horizon phase, from relatively undisturbed mud beneath which has only been in the B-horizon. The air-gun records reveal a further, deeper interface, which we assume to represent the surface of the limestone of which Coraholmen is constructed.

The zone of reticulate ridges is bounded to the east, in the sea as on Coraholmen, by a steep 15 m slope shown clearly both on sonographs and seismic records (Figs 20

and 21), and which can be connected to the ice contact scarp on Coraholmen by seismic and echo-sounder surveys (Fig. 19). We suggest therefore that this is an ice-contact slope, and that it marks the eastward extent of the Sefströmbreen surge. Well-defined westerly-inclined reflectors (a, b, c, d; Fig. 21) occur just east of the ice-contact scarp. We suggest that these may be shear planes formed during the last phase of movement against the ice contact front.

Solheim and Pfirman (1985) reported sonographs of the sea floor in an area covered by the 1936–1938 surge of Bråsvælbreen in Nordaustlandet, Spitsbergen. They show reticulate networks of ridges of precisely the same type as those described here, and suggested that the ridges reflected bottom topography of the glacier and might be analogous to terrestrial features explained as crevasse-fill features. In view of their close morphological similarity to the features produced on land and in the sea by Sefströmbreen, we have no doubt that they originated in a similar way, by intrusion of a soft, low-strength “surge-carpet” of subglacially-deformed sediment into a network of crevasses typical of a surging glacier.

PROGLACIAL TILL FLOWS AND THE ORIGIN OF TILL TONGUES

Much of the drift mass just to the east of the ice-contact front (Figs 20 and 21) is presumed to be analogous to the pushed mass shown in section 6 (Fig. 12) on Coraholmen. The reflectors marked a–d (Fig. 21) are presumed to be thrust planes analogous to those in section 6 at 30 m (Fig. 12). However, this drift mass extends some 2 km beyond the ice-contact front (Figs 21 and 22), whereas the analogous mass on land extends for a maximum of 400 m beyond the ice-contact front. We suggest that this occurs because, whereas on Coraholmen the pushed and extruded mass lies against a counter-slope, in the sea to the south it was pushed and extruded onto the crest of the shoal from which it has flowed into the basin to the east. Indeed, transit sonar records over the distal part of this mass show lobate flow forms which appear to coincide with the crenulations on the surface of the mass between 2000 and 2600 m from datum (Fig. 21), and may be analogous to the small ridges in the outer zone of parallel ridges on Coraholmen.

The margin of the till mass to the east abuts against a rising bedrock surface which forms the eastern flank of a submarine valley (Fig. 21). It is possible that the till flows have been diverted to the south to flow for a greater distance along the axis of the valley. Similar, extended submarine flows occur to the north of Coraholmen (Fig. 19).

King and Fader (1986) reported relatively thin till units which are rooted in major submarine till moraines on the Nova Scotia continental shelf. They preferentially occur on the distal flank of the moraine, and were interpreted as products of ice contact deposition and a reflection of the movement of the grounding line as relative sea levels changed due to glacier isostasy. We suggest however that

FIG. 17. (a) Detail of a sample of undeformed glacial marine sediment. Note absence of birefringence and frequency of microfossils. The elongate black void in centre right is artificial, it is a shrinkage crack. Crossed polarizers, width of view 18.0 mm. (b) Detail of sample O.701 (for position see Fig. 12, section 3) showing deformed, fissile, glacial marine mud viewed under crossed polarisers. Within an overall clayey sediment, fine sandy bands have become boudinaged. Shears are visible as bright lines of high birefringence. Fissility follows discrete shears. Note calcareous microfossil. Width of the field of view is 18.0 mm.

these till tongues, and analogous features which King *et al.* (1991) have reported from the Norwegian continental slope, are the product of mass flow of sediment pushed up before and extruded from beneath a glacier which has remoulded large volumes of pre-existing marine sediment to form a thick deforming sediment carpet sheared along beneath the glacier.

DISCHARGE OF MELT-WATER FROM THE ICE MARGIN

Figures 3 and 8 show the locations of the courses of former streams on Coraholmen. They emanate from the

ice contact slope and extend as far as the coast. There are a series of such channels in the southern part of the proglacial zone and a single channel at the northern end of the margin, where drainage appears to have been parallel to the margin. They may have been produced by meltwater from the glacier surface but there is another potential source. At the time of emplacement, the deforming sediment is likely to have had a void ratio of the order of about 0.4 (Boulton, 1979), compared with current values of about 0.3. The total volume of till on Coraholmen is about $2.5\text{--}5 \times 10^7 \text{ m}^3$. Dewatering of this mass from the higher to the lower void ratio would involve expulsion of about $2.5\text{--}5 \times 10^6 \text{ m}^3$ of water. This would be driven along the potential gradient determined

FIG. 18. The reticulate pattern of ridges on the sea bed to the south west of Coraholmen shown on transit sonar traces. Positions are shown in relation to point B in Fig. 19. The width of the swathe (top to bottom) is approximately 200 m. The top of the image represents the sea bed profile beneath the ship, and shows a maximum relief of about 10–15 m. The ridges are believed to have been intruded into the bases of crevasses.

by the ice surface slope onto Coraholmen (Fig. 4). It is not unlikely therefore that water pressed out from consolidating subglacial tills at the end of the surge played at least a partial role in producing the fluvial features reflecting drainage from the ice front and shown in Figs 3 and 8.

CONCLUSIONS — EMPLACEMENT OF THE DEFORMATION TILLS ON CORAHOLMEN AND IN THE SURROUNDING SEA AREA

We suggest that prior to 1882 glaciomarine muds had accumulated between Coraholmen and the shoal shown at 5 km in Fig. 4 (see also Fig. 22A). Shortly after 1882, Sefströmbreen advanced beyond this rocky shoal over the glaciomarine muds. Because of rapid loading, and poor drainage, they were unable to mobilise additional frictional resistance to glacier movement. As a result the weak sediments deformed rapidly at low effective stress and the reduction of basal drag suffered by the glacier in advancing onto soft sediments beyond the rocky shoal led to an acceleration of the glacier and a rapid advance under a predominantly extensional spreading regime (Fig. 22B). All pre-existing marine sediments were remoulded to form the till which currently appears to rest directly on bedrock.

In non-surging glaciers on land, shear stresses are significantly greater than longitudinal stresses, and they strongly influence the orientation of crevasses (Nye, 1952). If bed friction is very small, as in a surge of the type envisaged here, longitudinal tensile or compressive stresses may dominate the stress field. In the longitudinally extensional regime inferred for Sefströmbreen during its surge in a confined fjord channel, we would expect the maximum stress (σ_1) to be transverse to flow, the intermediate stress (σ_2) to be vertical and the minimum, tensile stress (σ_3) to be

parallel to flow. From this we would expect conjugate fracture planes to develop with vertical intersections, progressively attenuated by longitudinal extension and with orientation peaks occurring at acute angles on either side of the flow direction; precisely the orientation seen in the ridges shown on sonographs from the bay to the west of Coraholmen. The pattern persists onto Coraholmen (Fig. 9) even though we would expect longitudinally-compressive stresses to have developed as the glacier mounted the Island to generate longitudinal crevasses. We suggest that this lack of reorientation reflects rapid passive transport of crevassed ice onto Coraholmen, with inadequate time for major transverse extensional strains to develop which might have re-oriented crevasses.

During the extensional phase of the surge, we expect shear deformation of subglacial sediments to have been associated with longitudinal attenuation of existing structures (Fig. 22B–C). Folds will tend to become rod-like, with longitudinal axial planes, except where small shoals, irregularities on the bed or coarser-grained sediments (such as the *Lithothamnium* gravels) occur. A qualitative index of the consistency of the flowing subglacial sediment is reflected by the fact that effective stresses during glacier overriding were inadequate to break mollusc shells, and shear forces were inadequate to break the ligament holding the valves together (Fig. 16).

We suggest that the well-defined reflector some 20 m below the sea bed in the seismic profile east and south of Coraholmen (Figs 20 and 21) reflects the lowest position of the *decollement* plane between the rapidly flowing A-horizon and the underlying B-horizon of very slow elastic/plastic deformation found in subglacially-deforming sediment masses (Boulton and Hindmarsh, 1987). The A-horizon is one of strong mixing and remoulding, where deformation till is

FIG. 19. Map showing: scale corrected interpretation of the transit sonar images along two transects to the south west of Coraholmen; the reticulate pattern of crevasse-fill ridges; the distribution of crevasse-fill ridges on Coraholmen itself; track lines along which pinger and echo-sounder surveys were made; and the estimated position of the ice contact scarp which marks the maximum extent of the Sefströmbreen at the end of its surge, based on intersections along survey lines. It also shows the estimated extent of the submarine, proglacial flowed till zone, marked on land by the zone of parallel ridges. The eastern margin of this zone in the southeast appears to be determined by the eastern flank of a submarine valley which prevents the flow from extending further east (Fig. 20), but diverts it towards the south.

FIG. 20. Pinger profile across the maximum extent of the Sefströmbreen surge along the line A-C shown in Fig. 19. Distances are shown in relation to point B (Figure 19). The features from 200 m to 1400 m are crevasse-fill ridges. The scarp at 1600 m is the extension of the ice contact slope on Coraholmen and coincides with the mapped maximum extent of Sefstrombreen during the surge shown in Figs 2 and 19. The stratum above reflector X in the zone of crevasse fill ridges is suggested to be the base of the deformation till mass, to the right (east) of the ice contact slope, it is interpreted as underlying a sediment flow unit consisting of material pushed before and extruded from beneath the surging glacier margin. It appears to pass laterally into the distal zone of parallel ridges on Coraholmen.

produced. This appears to have had a thickness of about 20 m, similar to the thickness of deforming sediment inferred by Alley *et al.* (1986) to exist beneath ice stream B in Antarctica.

Crevasse opening during extensional flow of Sefströmbreen led to intrusion of deformation till up into crevasses (Fig. 22C). The maximum height to which we would expect sediment to rise in crevasses is given by

$$\frac{t_s = t_i \rho_i}{\rho_s} \quad (8)$$

where t_i and t_s and ρ_i and ρ_s are the thickness and density of ice and sediment respectively.

Given values of $\rho_i=0.9$, $\rho_s=1.9$ from the measured bulk density of actively deforming fine-grained till at Breida-

FIG. 21. Interpretation of pinger records along the transect A-B-C in Fig. 19. Distances along the line are shown in relation to point B (Fig. 19). Three zones can be defined, from left to right as: the zone of crevasse-fill ridges (0–1600 m), an ice contact zone (1600 m) and a zone of sediment flows (1600–1900 m). Reflector X is interpreted as the pre-surge surface. To the west (left) of the ice contact front, it is overlain by deformation till whose upper part has intruded into open crevasses during post surge stagnation. In the ice contact zone, the overlying till contains reflectors (a, b, c, d) which are interpreted as shear planes in sediment pushed by the glacier as the surge died. To the east of the ice contact zone, reflector X is overlain by sediments interpreted as flows derived from sediment extruded from the beneath the surging glacier and pushed beyond it. The crenulations just beyond the ice contact zone are analogous to those in the zone of parallel ridges on Coraholmen, and of similar size. The flowed till mass is prevented from flowing further east by the rising valley wall at 2900 m. The flowed mass extends some 1.3 km to the east of the surge terminus. It is suggested that it is equivalent to a 'till tongue' (King *et al.*, 1991).

FIG. 22. Summary of the processes of emplacement of the Coraholmen sediments. (A) The position of Sefströmbreen and the character of sea floor sediments prior to the surge. The modern position of the south-western peninsula of Coraholmen (Fig. 19) lies below sea level. (B) Sefströmbreen in a late stage of the surge. Glacialine muds are completely remoulded to form deformation till and transported by shearing beneath the glacier whose surging they facilitate. Transverse crevasses penetrate through the glacier in a zone of extension in the deep water south-west of Coraholmen and deformation till is intruded into them. As the glacier rides up onto the Coraholmen shoal, transverse crevasses close around the intruded sediment, but the dominant longitudinal crevasses (Fig. 19) remain open. *Note the change of section scale in C–E.* (C) As the surge reaches its maximum, sands and algae on the shoal (facies e) are sheared along with muddier sediments; sediment is pushed beyond the glacier and material is extruded from beneath it. This produces extensive till flows beyond the terminal ice contact scarp. (D) After the surge, the glacier stagnates and a large ice mass is left on the Coraholmen shoal, separated from the main body of the glacier. As the surface melts down, till intrusions in crevasses are exposed at surface and spread out over the glacier, thereby burying large ice masses. The underlying till consolidates under the weight of ice, water is expelled from it and further extrusion occurs from beneath the ice. (E) As the ice melts further and disappears from between the crevasse fill ridges, till slumps from them into the resultant depressions. Coraholmen has been greatly extended by glacial deposition.

merkurjökull, Iceland) and $t_i = 85$ m in the vicinity of the West Coast as suggested by De Geer's survey, and assuming a parallel-sided crevasse, we calculate the maximum height of intrusion as 40 m. This would explain why the ice on Coraholmen at the end of the surge in 1896 was clean (Fig. 22C), whereas by 1910 the surface had melted down sufficiently to expose the till in crevasse intrusions at surface (Figs 5 and 22D); a till which then flowed over and covered the stagnant glacier ice.

We expect a strong vertical velocity gradient to have existed in the upper part of the subglacial deforming sediment (cf. Boulton and Hindmarsh, 1987). As a consequence the till intruded into crevasses, with a predominantly vertical fabric, will tend to be separated by a *décollement* zone from the underlying till in which planes of shear will be roughly horizontal (Fig. 22D). Such an expected contrast was not seen in Coraholmen sections, possibly because a low enough level of exposure was not seen, and possibly because late stage settling and consolidation (see later) obscures the contrast. Note, however, that strong reflectors pass beneath crevasse intrusion ridges between 1500 and 2000 m on the seismic traverse (Figs 20 and 21).

As the glacier stagnated, deglaciation proceeded differently in the sea and land areas. In the former, progressive melting would eventually permit crevasse-bounded ice blocks to float away from the fringing crevasse intrusion ridges. In the latter, ice blocks melted *in situ*.

Deposition associated with compression (Fig. 1c) is to be expected as a consequence of the glacier riding up onto Coraholmen and the shoals to the south. The inclined reflectors, a–d in Fig. 21 may reflect this as a consequence of reduction in longitudinal strain rate in the till and progressive displacement towards the west of major *décollement* planes. It may also be reflected by a change in fold style from the highly attenuated folds of the southwestern extremity of Coraholmen to the more frequent, shorter-limbed folds further east. The alternating phases of compression and extension illustrated in Fig. 15 indicate a style of stick-slip movement, whilst some sequences of stacked folds could reflect successive lodgement of folded masses.

As the glacier moved over the soft fjord muds, we would expect some consolidation and extrusion of mud and water from beneath the glacier into the proglacial zone. Such extruded mud would then be pushed forward as a push moraine, finally to be deposited beyond the limit of the advance (Fig. 22D). We suggest that the structures shown in Figs 11c and 12 (Section 6) reflect this pushing, and that the major ice-contact front seen both on land and in the sea represents the pushing front.

As the surge died, the shear force, which through dilation in the deforming mass (Boulton and Hindmarsh, 1987; Clarke, 1987) had helped to sustain very high void ratios, also decayed, permitting consolidation to occur through water expulsion and drainage (Fig. 22D). Assuming a maximum thickness of ice on the western side of Coraholmen and in the adjacent sea area of 60–80

m, consolidation tests on the Coraholmen muds suggest that a void ratio of about 0.4 would have been an equilibrium with this load. Assuming a void ratio during deformation of about 0.7 (Boulton and Hindmarsh, 1987), then consolidation of the muds to equilibrium would cause a vertical shrinkage due to water loss.

If the present thickness (t_1) is 20 m and the original thickness t_0 , shrinkage will be

$$t_0 - t_1 = 20 \left(\frac{1.7}{1.4} \right) - 1 = 4.3 \text{ m} \quad (9)$$

We presume that this settling process will cause further mud extrusion into crevasses, and in particular into the proglacial area. The concentric ribbing seen in the muds beyond the ice-contact front, and the flow structures seen on sidescan sonar traces are interpreted as the product of flow under gravity of the frontally-pushed mass and the extruded sediments which we suggest overlie them.

The deformation till bears evidence of flow through the folding which is apparent from place to place. These folds are frequently overprinted by fracture systems which, though in many places may be a product of passive consolidation, in other places appear to show evidence of rotation (Fig. 12, Section 1), suggesting that deformation due to glacier movement still continued during fracture development. If, for simplicity's sake, we assume that all fractures have a common origin, rotation of fractures implies that they formed beneath the still-active glacier, and were not the product of unloading alone. We would suggest that they are analogous to fractures in the B-horizon of subglacial tills in Iceland, explained by Boulton and Hindmarsh (1987) as a consequence of consolidation of till as it changed from being a rapidly deforming high void ratio A-horizon into a consolidated, slowly deforming, elastic-plastic B-horizon, which yields to stress by fracture along closely-spaced shear planes. There are then two possibilities for their origin:

- (1) As the glacier mounted over Coraholmen, the lower part of the deforming mass was strongly retarded, ceased dilating strongly, began to consolidate as a tectonic B-horizon, and began to yield by fracture rather than flow. Progressive thickening of the B-horizon occurred as more till lodged against the island and shoal and the A/B *décollement* moved upwards. Deformation still occurred within the B-horizon as a response to glacier-overriding, leading to rotation of earlier-formed fractures. This explanation is compatible with the suggestion that the inclined reflectors shown in Fig. 20 represent successive upward and westward shifts of the A/B *décollement*.
- (2) As the surge stagnated, subglacial sediments underwent general drainage and consolidation. A slight further forward movement at this stage might have led to deformation by fracture rather than flow in subglacial sediments.

As settlement and consolidation of till occurred, till

was extruded from beneath the glacier, and this material, together with that which had been pushed in front of the glacier, flowed freely into the proglacial area to produce very extensive mass flow deposits (Fig. 22D–E), particularly in the submarine zone where sediment did not dry out. Large masses of water were also expelled from the tills during the consolidation process, pressed out beyond the stagnant glacier margin and may have left sedimentary evidence of stream channel deposits in the proglacial area.

During final decay of the dead ice mass on Coraholmen, the ice between crevasse-fill ridges decayed, leaving these ridges as free standing forms, till slumped from their sides to accumulate in the inter-ridge basins (Fig. 22E). The prior consolidation of this till, possibly associated with interstitial salt, favoured the preservation of steep slopes.

IMPLICATIONS FOR THE INTERPRETATION OF ANCIENT GLACIAL SEDIMENTS

Many high latitude continental shelves and sea areas have been extensively and repeatedly glaciated during the Late Cenozoic, and much of the sediment on them is of glacial origin. Because of the potential similarity between glaciomarine diamictos and glacial tills which may have originated by glacial remoulding of pre-existing glaciomarine sediments, it is often difficult to determine the extent and frequency of glacial invasions of the continental shelves.

The available evidence comes from three sources: natural sections where glacio-isostatic uplift has carried marine sequences above sea level; marine cores and boreholes; continuous reflection seismic data. We review the extent to which evidence from each can be used to distinguish deformation till.

Natural Sections

The history of the Sefströmbreen surge documented by De Geer makes it quite clear that the thick mud sequence draping the western side of Coraholmen was emplaced as deformation till by the surge. Had that history not been known, a large part of it could still have been inferred from the fold structures apparent in section and from the island's geomorphology. However, the folds are generally only so readily distinguished because of lithologic contrasts, in particular the presence of facies (e), the *Lithothamnium* beds. If only facies (a) and (b) had been present, it would have been difficult to distinguish them from a relatively massive *in situ* glaciomarine deposit.

Although the crevasse-intrusion ridges are quite diagnostic of glacier overriding, they only survive because of stagnation of the glacier at the end of the surge. They do not survive where glaciers undergo active retreat where they would be deposited as frontal moraines.

The Coraholmen tills are slightly overconsolidated. However, we believe much of this to have been a late-

stage feature when drainage allowed them to consolidate. It is not difficult to conceive of situations (Boulton and Hindmarsh, 1987) in which inhibition of drainage to the moment of deglaciation will preserve a normally consolidated till. The inference of Josenhans and Fader (1989), that normally consolidated till on the eastern Canadian shelf could only have been deposited during retreat as the ice began to float, is not correct. Moreover, overconsolidation is only a criterion for glacier overriding for the uppermost overconsolidated unit in the sequence, as lower units will suffer the same overconsolidation as those forming the immediately subglacial bed, depending on drainage conditions.

Shear fracture patterns, particularly systematically rotated patterns, may be a strong index of glacier overriding, although a single phase of overriding may induce fractures in appropriate lithologies at different depths within a subglacial sequence. In theory at least, fractures resulting from unloading may also reflect unloading by non-glacial erosion.

The existence of undamaged molluscs in a sediment would normally, we believe, be taken erroneously by most investigators as evidence that a diamicton is not a till.

Cores and Boreholes

We have compared channel samples from the Coraholmen till with cores of modern glaciomarine sediment taken from the bay between Coraholmen and Sefströmbreen. From lithology and structure alone, only the chance of sampling the fractured facies is likely to yield a reasonably unequivocal diagnosis of glacier overriding. Only a large scale sampling programme would permit granulometry to be used as a diagnostic criterion; and deformation structures whose large scale style cannot be resolved could either be produced by glacier overriding or submarine slumping. Evidence of significant overconsolidation where it is clear that there has not been major non-glacial erosional unloading would be a good criterion of ice overriding. However, as with fracturing, it can only be applied to the uppermost glacially-overridden bed. Microstructures observed in thin sections may help in differentiation between till and glaciomarine sediment (see also van der Meer, 1993).

Seismic Evidence

Where till units are expected, seismo-stratigraphic units with acoustically "chaotic" signatures and an absence of coherent internal reflectors are often interpreted as till (e.g. Praeg *et al.*, 1986; Vorren *et al.*, 1989).

We presume that the absence of coherent reflectors reflects the absence of vertical contrasts in acoustic impedance derived either from granulometric layering or density layering, and that chaotic signatures are derived from interference patterns produced by strong point reflectors. On the other hand some acoustically layered sediments have been interpreted as till (e.g. Vorren *et al.*, 1989).

Granulometric layering can clearly be achieved in a

deformation till by deformational mixing of inhomogeneous source materials, as occurs to a certain extent on Coraholmen. On the other hand relatively massive and homogeneous source materials will tend to produce unlayered deformation tills. Boulton and Dobbie (1993) have argued that the propensity to develop density (consolidation) layering will depend upon the drainage regime during glacier overriding. Where drainage in a deforming till is downwards, into sub-till aquifers, horizontal layering may be produced through A/B horizon contrasts. Seismically identified layering in Coraholmen deformation till (Fig. 20) may thus reflect drainage into underlying fractured limestones.

Where the deformation till is underlain by an aquiclude, drainage will be horizontal and no horizontal consolidation contrasts will be expected.

Chaotic signatures may be due to boulders, or to complex fold patterns, or de-rooted fold closures. Deformation tills derived from homogeneous fine-grained muds will lack these features. We would suggest therefore that although a chaotic and incoherent acoustic signature may well be a good basis for interpreting a seismic unit as till, non-chaotic and coherently layered units cannot be dismissed as "non-till".

A persistent feature of many tills positively-identified from seismic data is a relatively smooth basal boundary and an irregular upper boundary. The lower boundary is likely to be dominated by streamlined shearing and erosional processes, and the upper by complex deglaciation processes, as in the case of the Sefströmbreen surge.

The difficulty of identifying tills from glaciomarine or glaciolacustrine units is reflected by many disagreements about diamicton genesis in marine or lacustrine environments. For instance Eyles and Eyles (1983) have interpreted sediments in the Toronto area which have long been interpreted as tills (Dreimanis, 1977) as lacustrine mass flow deposits. Similarly, many of the shelly fine-grained diamictons around the Irish Sea basin have been variously interpreted as tills or glaciomarine sediments.

We offer therefore, no single diagnostic criterion, and can only advocate, yet again, careful sifting of a variety of evidence, to which process we hope our observations and inferences will contribute.

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