Flow switching and large-scale deposition by ice streams draining former ice sheets

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ABSTRACT

Fast-flowing ice streams are responsible for the bulk of mass transfer through large ice sheets. We use extensive three-dimensional seismic data from the western Norwegian margin to explain how a several-hundred-kilometer-long ice stream has undergone major switching in flow direction from one glaciation to the next. The direction of ice flow is inferred from the pattern of buildup of thousands of cubic kilometers of glacier-derived debris and observations of large-scale streamlined landforms on former subglacial beds. We demonstrate that ice streams can undergo major changes in flow direction through modification of their large-scale topographic setting. Whereas ice-stream switching in modern ice sheets has been regarded mainly as a reflection of internal changes in ice-sheet dynamics, switching over successive 100 k.y. glacial cycles may in this case be a response to the effects of continuing sediment deposition and the large-scale development of ice-influenced continental margins.

Keywords: ice streams, Norwegian margin, three-dimensional seismic data, glacial lineations, glacial bedforms.

INTRODUCTION

Fast-flowing ice streams and outlet glaciers currently drain more than half of the mass from the Antarctic and Greenland ice sheets (e.g., Bentley, 1987; Thomas, 2004). Some temporal and spatial variability has been observed in the flow of modern West Antarctic ice streams (e.g., Retzlaff and Bentley, 1993; Jacobel et al., 1996; Conway et al., 2002), but major shifts in the location of ice streams flowing in deep channels have not been observed. We use an extensive three-dimensional seismic data set from the Norwegian margin to demonstrate and explain how an ice stream, >400 km long and draining an interior basin of >100,000 km^2, has undergone major switching in flow direction. We show that the ice stream flowed southwestward down a major depression during repeated glacial episodes, depositing a large accumulation of sediment eroded from beneath the parent ice sheet on the continental margin of mid-Norway. The direction and fast flow of ice are inferred from the pattern of buildup of glacier-derived debris and the observation of streamlined landforms on former subglacial beds. However, a change in orientation of these streamlined landforms indicates that the flow direction of the lowermost half of the ice stream switched dramatically by 90° between the penultimate (Saalian) and most recent (Weichselian) glacial periods, as the accumulation of sediments deposited during previous glacials progressively obstructed southwestward flow. The ice stream eroded a new 150-m-deep trough in underlying sediments, built a new depositional center ~100 km north of its previous terminus region, and became more than 100 km shorter. This demonstrates that ice streams can undergo major changes in flow direction as a result of long-term modification of their large-scale topographic setting over a number of glacial cycles, with considerable implications for the sedimentary architecture of the continental margin involved. Switching is linked to the long-term pattern of glacial erosion, transport, and deposition of several thousands of cubic kilometers of debris over a number of cycles of ice-sheet growth and decay since the first buildup of ice on Scandinavia ~2.5 m.y. ago (Jansen and Sjøholm, 1991).

ICE STREAMS: BACKGROUND

Ice streams are fast-flowing (10^2–10^3 m yr^{-1}) curvilinear elements within ice sheets that have sharp velocity gradients to slower-flowing (10 m yr^{-1}) ice beyond their crevassed margins (Bentley, 1987). They are usually tens of kilometers wide, as long as hundreds of kilometers, and ~500–2500 m thick. Fast motion is related to the deformation of their soft sedimentary beds, which, in contrast with slower-flowing ice, are at the melting point and probably water saturated. A component of basal sliding over the sediment surface may also be involved (Engelhardt and Kamb, 1998). Ice streams are important both glaciologically and geologically, because their rapid flow is responsible for the bulk of ice and sediment delivery to the margins of present and past ice sheets (e.g., Dowdeswell and Siegert, 1999; Pollard and DeConto, 2003). The successful reconstruction of past ice sheets requires that the locations and dimensions of former ice streams can be identified in the geological record and in numerical models of paleo–ice-sheet dynamics. In addition, knowledge of the nature and rate of sediment delivery to high-latitude glacier-influenced continental margins is also fundamental to understanding their evolution and sedimentary architecture.

SEISMIC EVIDENCE FOR LARGE-SCALE DEPOSITION ON THE MID-NORWEGIAN MARGIN

On the mid-Norwegian margin (Fig. 1A), seismic surveys have allowed detailed mapping of the three-dimensional form of the Naust Formation, which comprises a series of sedimentary units as thick as 2 km that make up the past 3 m.y. of predominately glacier-derived deposition (e.g., Dahlgren et al., 2002; Rise et al., 2005). Seismic reflectors within the formation that are often associated with past

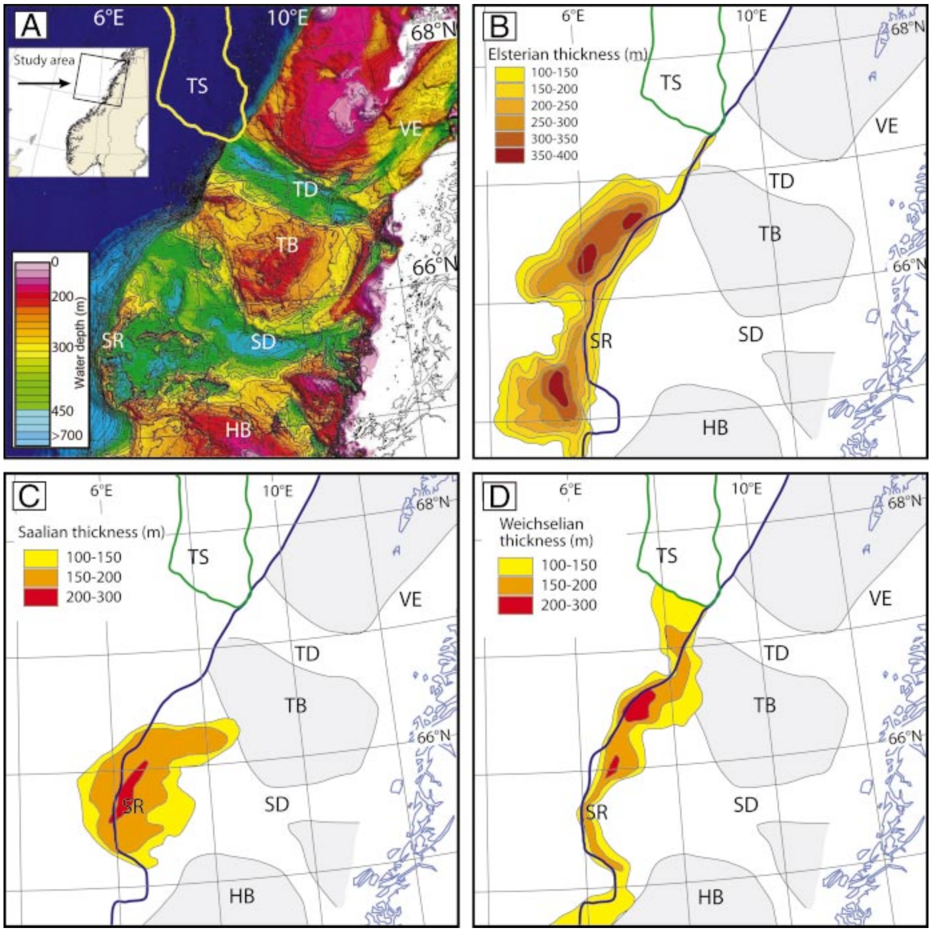

Figure 1. A: Bathymetric map of mid-Norwegian shelf. Study area is ~63,000 km^2. Isopach maps show sediment thickness for major glacial depositional centers along mid-Norwegian margin for Elsterian (B), Saalian (C), and Weichselian (D) glacial periods. Sediment thicknesses >100 m are plotted assuming sound velocity of 2000 m s^{-1}. Shelf edge at ~500 m water depth is shown in blue. Shaded areas in B–D are shallow banks. HB—Haltenbanken, SD—Sklinnadjupet, SR—Skjoldryggen, TB—Traenabanken, TD—Traenadjupet, TS—Traenadjupet Slide, VE—Vestfjorden.

ice-sheet erosion allow the thickness and volume of a number of units to be mapped. Maps have been constructed of sediment thickness along the mid-Norwegian margin between 65° and 67.5°N for each of the past 3 100-k.y.-long glacial cycles, known as the Elsterian, Saalian, and Weichselian (Figs. 1B–1D).

During the Elsterian and Saalian glaciations (~300 k.y.–135 k.y. ago) sedimentation on the mid-Norwegian margin was centered on the Skjoldryggen area between 65° and 66.5°N (Figs. 1B–1C), beyond the major cross-shelf trough of Sklinnadjupet. As much as 350 m of sediments were deposited, causing the building out of the continental shelf edge into deep water (Figs. 1B–1C). Thick sequences of glacial sediments were also deposited in the Traenabanken area during the Elsterian and Saalian glaciations. By contrast, ~100 km north of Skjoldryggen and Sklinnadjupet, offshore of what is now the cross-shelf trough of Traenadjupet (67°N), little sediment was delivered over these two glacial cycles (Figs. 1B–1C). During the last 100 k.y. of the Weichselian glacial cycle, however, sedimentation increased greatly beyond Traenadjupet to build a new depositional center, part of which was removed by more recent slides (Fig. 1D). The large-scale pattern of sediment delivery to the mid-Norwegian margin has therefore undergone a major shift during the Weichselian as compared with the two preceding glacial periods. In addition, the analysis of seismic data from the continental shelf around Traenadjupet also shows that a series of older reflectors has been truncated by a surface that forms the base of the Weichselian sedimentary unit (Fig. 2). Thus, the ~150 m overdeepened cross-shelf trough of Traenadjupet is a relatively recent morphological feature of the margin, which has developed mainly during the past ~100 k.y.

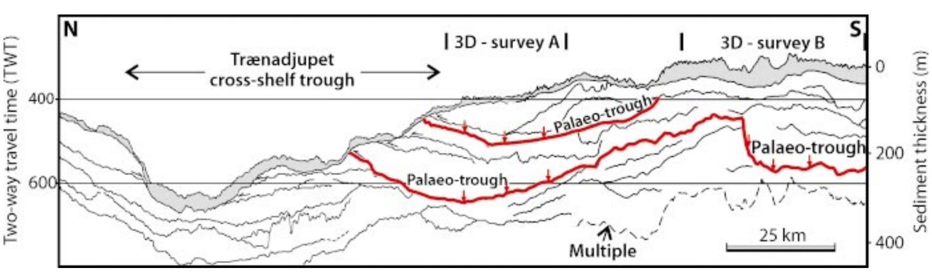

Figure 2. Seismic stratigraphy of upper ~0.5 m.y. of Naust Formation deposited on mid-Norwegian shelf. Sediment thickness is calculated using sound velocity of 2000 m s^{-1}. Seismic line (IKU B81–208) is located as N-S in Figure 3. Note erosion of underlying units by recent ice-stream flow in Traenadjupet and older ice-stream channels within sediments of Traenabanken to south (red lines labeled paleo-trough). Red arrows indicate presence of streamlined bedforms on buried paleotrough surfaces. Three-dimensional survey is NLGS-95 and survey B is NNE-2000.

PALEO–ICE STREAMS: FLOW DIRECTION AND SEDIMENT DELIVERY

Ice sheets have been the major source of sediment delivery to the entire Norwegian margin over the past ~2.5 m.y., since glaciers first formed on the Scandinavian mountains and extended outward to reach the coast and adjacent continental shelf (Jansen and Sjøholm, 1991). In addition, thick fast-moving ice streams, set within cold-based slower-flowing ice, provide the major source of icebergs, meltwater, and debris to the margin (e.g., Dowdeswell et al., 1996; Dowdeswell and Siegert, 1999; Siegert and Dowdeswell, 2002). These ice streams extended to the shelf edge during cold full-glacial conditions. The locations of past ice streams can be inferred from the presence of large-scale streamlined bedforms, known as megascale lineations, in continental shelf sediments. The lineations are formed at the base of fast-flowing ice streams by deformation processes affecting the upper few meters of sediment, which are often water saturated (Kamb, 2001; Clark et al., 2003; Dowdeswell et al., 2004). Individual lineations are as much as 15 m high and hundreds of meters wide. They are usually spaced hundreds of meters apart and are orientated in the direction of ice flow (Ó Cofaigh et al., 2002). Sets of lineations are well preserved and characteristic of modern cross-shelf troughs in both polar regions, where ice was present as recently as 10–15 k.y. ago (Shipp et al., 1999; Ó Cofaigh et al., 2002; Ottesen et al., 2005a, 2005b). They have been used to define the distribution of ice streams along the entire 2500-km-long western margin of the Scandinavian-Svalbard ice sheet at the last glacial maximum ~18 k.y. ago (Ottesen et al., 2005a).

We have used megascale glacial lineations (Clark, 1993) to define the locations and directions of ice-stream flow on the mid-Norwegian shelf over the past 300 k.y. Seismic reflectors, representing former glacier beds now buried hundreds of meters below the modern seafloor, have been identified and mapped (Figs. 3A–3B). These reflectors reveal detailed patterns of streamlined bedforms, identical in form to those from the latest Weichselian glaciation (Figs. 3C–3D). The changing pattern of ice flow inferred from the orientations of these sets of bedforms is shown in Figure 3E. During the Elsterian and Saalian glacial periods ice flowed southwest from the deep trough of Vestfjorden, across what is now Traenabanken, and into the Skjoldryggen area (Figs. 3A–3B). However, the present seafloor of Traenadjupet and Vestfjorden shows lineations extending down Vestfjorden and then turning almost 90° into Traenadjupet, where they extend ~100 km to the shelf edge (Ottesen et al., 2005a, 2005b) (Figs. 3C–3D). A major switch in ice-stream flow direction has therefore taken place (Fig. 3E).

The change in ice flow direction after the Saalian glaciation is supported by the growth of a major depositional center of at least 1500 km^3 at the mouth of Traenadjupet during the Weichselian period (Fig. 1D). The depocenter is composed mainly of stacked glacigenic debris flows, but ~900 km^3 has been removed by the 4-k.y.-old Traenadjupet Slide (Fig. 1D) (Henriksen and Vorren, 1996; Laberg et al., 2002) and an unknown but probably much smaller volume from the older Nyk Slide (Fig. 3E) (Lindberg et al., 2004). Rapid delivery of sediments from fast-flowing ice is required in order to build a depocenter of these dimensions (Dowdeswell and Siegert, 1999). In addition, seismic data show that these glacier-derived deposits are underlain by acoustically layered fine-grained sediments deposited over the Elsterian and Saalian periods (Dahlgren et al., 2002; Bryn et al., 2005). This material is interpreted as a contourite, deposited by northward-flowing ocean currents in the absence of large-scale sediment delivery from glaciers to the slope beyond the present Traenadjupet. The switch in flow direction is further supported through evidence of deep erosion by the Weichselian ice stream, which truncated earlier seismic reflectors and units at the flanks and base of Traenadjupet (Fig. 2). We calculate that the ice stream, ~100 km long, 50 km wide, and having an erosive depth of 0.15 km, evacuated at least 750 km^3 of sediment from this cross-shelf trough.

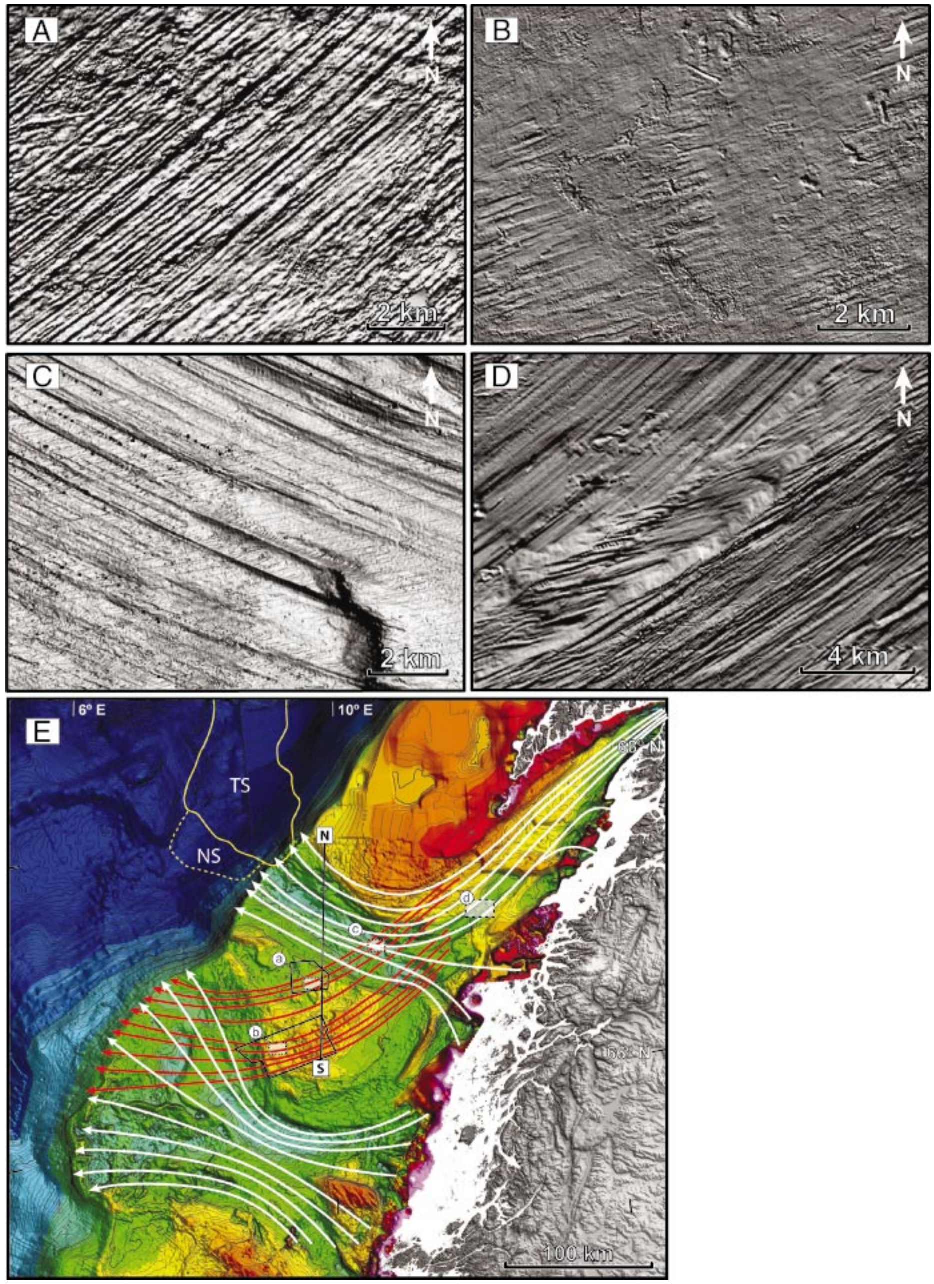

Figure 3. Streamlined sedimentary bedforms produced beneath paleo–ice streams on mid-Norwegian continental shelf. A: Buried surface ~100 m deep within upper Naust Formation (three-dimensional [3D] seismic block NLGS-95). B: ~200 m deep within upper Naust Formation (3D seismic block NNE-2000). C: Late Weichselian sediments at modern seafloor of Traenadjupet (3D seismic block ST-9404). D: Seafloor of Vestfjorden (EM1002 swath bathymetry). E: Map of changing ice-stream flow directions inferred from orientation of streamlined bedforms on mid-Norwegian shelf (located in Fig. 1A). Red lines are Elsterian–Saalian and white lines are Weichselian ice-stream flow directions, respectively. TS—Traenadjupet Slide, NS—Nyk Slide. Locations of surfaces in A–D are shown in E (labeled a–d), together with seismic line from Figure 2 (labeled N-S). Black boxes locate two 3D seismic blocks.

LARGE-SCALE ICE-STREAM SWITCHING

This evidence demonstrates large-scale switching of ice-stream flow within a major ice sheet from one glaciation to the next. The change in flow direction may have taken place because of continuing buildup of glacier-derived sediments on the mid-Norwegian continental shelf and the filling of the available accommodation space. When Weichselian ice began to flow down Vestfjorden, its easiest path to the shelf edge was no longer across Traenabanken, supplying ice and debris to the

Skjoldryggen area to the southwest. Instead, a new trough was excavated through sediments to produce the 150-m-deep depression now known as Traenadjupet (Figs. 2 and 3E). A large-scale geological consequence of this radical shift in ice-stream location was the development of a major new depositional center on the continental margin near 67°N (Fig. 1D). However, we cannot rule out the possibility that large-scale switching was a response to glaciological changes in, for example, the dimensions of the inland ice-sheet drainage basin or the thermal structure of the ice-sheet base. Ice-stream switching in modern ice sheets has previously been regarded mainly as a reflection of such internal changes in ice-sheet dynamics. We show that switching over longer periods may also be influenced by the effects of continuing sediment deposition and the large-scale topographic development of high-latitude continental margins.

ACKNOWLEDGMENTS

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Assemblages of submarine landforms produced by tidewater glaciers in Svalbard

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[1] High-resolution swath bathymetry from the marine margins of several Svalbard tidewater glaciers shows an assemblage of submarine landforms that is probably linked to glacier surging. These landforms are essentially unmodified since their initial deposition over the past hundred years or so because they have not been subjected to subaerial erosion or periglacial activity. Swath images comprise an assemblage of superimposed landforms, allowing reconstruction of relative age of deposition: (1) large transverse ridges, interpreted as recessional moraines overridden by a subsequent ice advance; (2) a series of curvilinear streamlined bedforms orientated parallel to former ice flow, interpreted as lineations formed subglacially during rapid advance; (3) large terminal ridges, marking the farthest extent of ice at the last advance, with flow lobes immediately beyond interpreted as submarine debris flows; (4) a series of interconnected rhombohedral ridges, interpreted as a product of soft sediment squeezing into crevasses formed at the glacier bed, probably formed during immediate post-surge stagnation; and (5) a series of fairly evenly spaced small transverse ridges, interpreted as push moraines produced annually at tidewater glacier termini during retreat. A simple descriptive landsystem model for tidewater glaciers of probable surge type is derived from these observations. We also show that megascale glacial lineations can form not only beneath large ice streams, but are also produced beneath surging tidewater glaciers lying on deforming sedimentary beds.

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1. Introduction

[2] The sedimentary processes taking place beneath modern glaciers and ice sheets are difficult to observe directly, yet they are important to an understanding of the nature and rate of glacier motion and, through this, of the responses of ice masses to environmental change. Settings in which glacier ice is no longer present, but the ice-bed interface, the landforms and the sediments lying immediately beneath it are preserved without significant modification, provide an important archive of past glacial activity [e.g., *Shipp et al.*, 1999; *Canals et al.*, 2000; *Stokes and Clark*, 2001; *Dowdeswell et al.*, 2004]. Seafloor sediments, when not reworked by mass wasting, currents or iceberg-keel ploughing, can provide such a setting, in which subaerial processes of landform modification and erosion do not operate. In addition, these marine environments provide ready access for ships equipped with geophysical tools to investigate the submarine sedimentary environment.

[3] In a number of parts of the Arctic, tidewater glaciers, often fed from large ice caps and upland icefields, terminate in marine waters [*Clarke*, 1987; *Meier and Post*, 1987]. In Svalbard, many of these tidewater glaciers are of surge type [e.g., *Liestøl*, 1969; *Schytt*, 1969; *Dowdeswell et al.*, 1991; *Hagen et al.*, 1993; *Hodgkins and Dowdeswell*, 1994; *Jiskoot et al.*, 2000], and are likely to advance over sedimentary beds of glacimarine fjord sediments, whose deformation is largely responsible for fast glacier flow [*Boulton et al.*, 1996]. During the active phase of the surge cycle, these tidewater glaciers advance rapidly for periods of up to a few years before undergoing much longer intervals of thinning and retreat lasting decades or more [e.g., *Meier and Post*, 1969; *Dowdeswell et al.*, 1991]. Mechanisms for triggering the active phase of the surge cycle are related to the internal behavior of glaciers and their beds rather than to climate forcing [*Clarke et al.*, 1984; *Kamb et al.*, 1985], although the length of the quiescent period between surges is linked to glacier mass balance [*Dowdeswell et al.*, 1995].

[4] Not all tidewater glaciers surge [*Meier and Post*, 1987]. Many tidewater glaciers flow between constricting valley walls, are fed from large interior drainage basins, and have high velocities related to a large mass flux [*Clarke*, 1987]. Tidewater glaciers may also undergo dramatic collapse if they retreat into the deep water that is often present between sills that are a typical component of fjord morphology [*Syvitski et al.*, 1987; *Warren*, 1991]. It is sometimes difficult to distinguish between glacier terminus retreat due to postsurge stagnation and that associated with enhanced iceberg calving rates during tidewater terminus retreat from a shallow sill into deeper water. Similarly, it is

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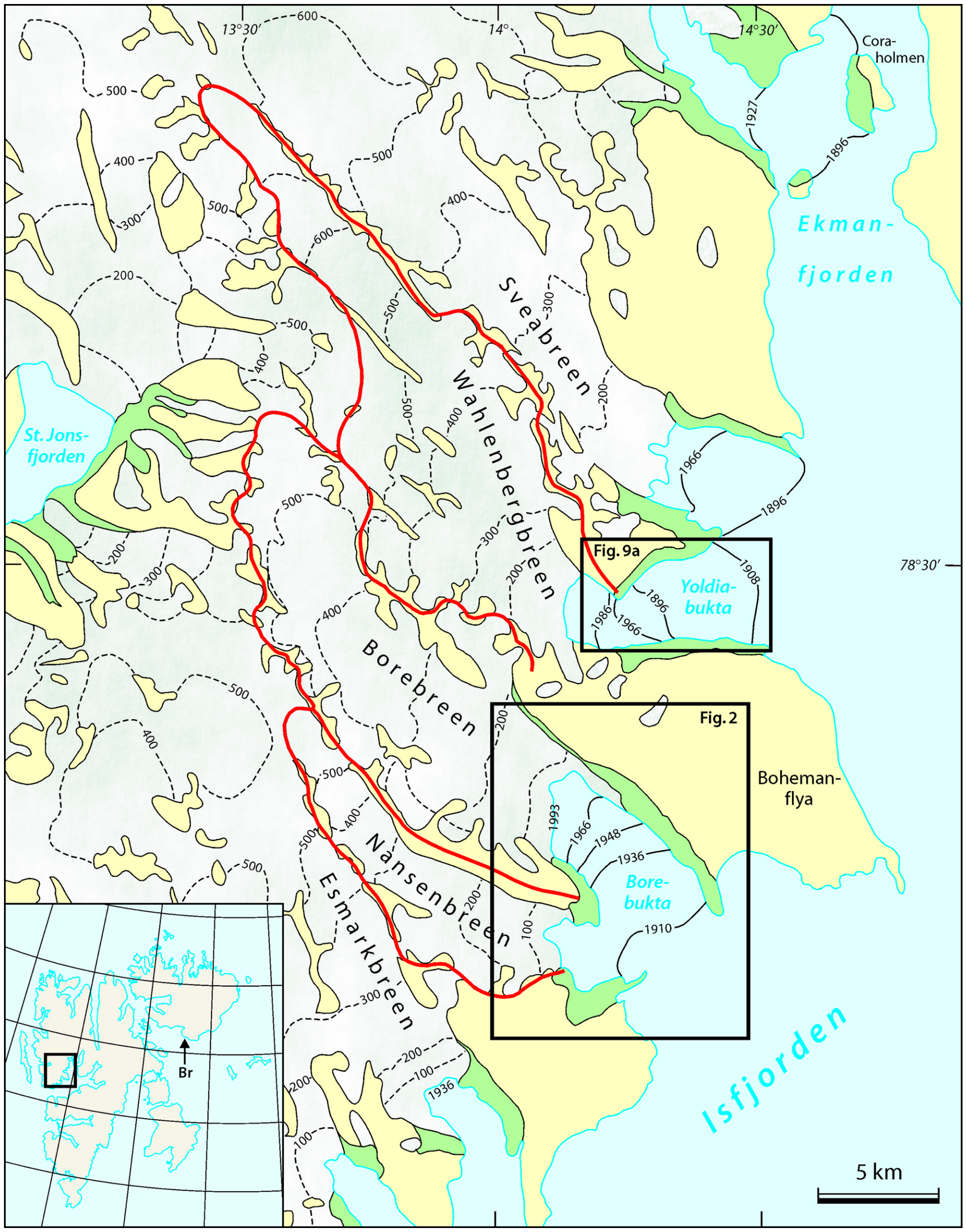

Figure 1. Location map of the study area, showing glaciers on the north side of Isfjorden, Spitsbergen, and the locations of our swath-bathymetric data in Borebukta and Yoldiabukta. The drainage basins of Borebreen, Nansenbreen, and Wahlenbergbreen are indicated (red lines), and past positions of the termini of several tidewater glaciers are shown. The study area is located within Svalbard (inset), where the location of swath-bathymetric data from Bråsvellbreen in Nordaustlandet is also shown (labeled Br). Green areas are marginal moraines, and yellow areas are ice-free bedrock.

not always easy to distinguish high velocities and ice advance linked to the active phase of the surge cycle from the rapid flow of tidewater glaciers between constraining valley walls and fed from extensive interior drainage basins.

[5] Many Svalbard glaciers and ice-cap drainage basins have been observed to surge into the surrounding seas [e.g., *Schytt*, 1969; *Dowdeswell*, 1986; *Hagen et al.*, 1993; *Dowdeswell and Benham*, 2003; *Murray et al.*, 2003]. In this paper, we describe a well-preserved suite of submarine landforms exposed recently on the seafloor at the margins of several Svalbard tidewater ice masses, and in particular a small Svalbard inlet, Borebukta in Isfjorden (78°N, 14°E), during the 10-km retreat of a tidewater glacier: Borebreen (Figure 1). We interpret this landform assemblage, or land-system [*Evans and Rea*, 2004], in the context of basal processes beneath a tidewater glacier, probably of surge type, overlying a sedimentary bed. We also provide a relative chronology for the deposition of each observed landform element and present a simple morphological model of these landform elements comprising the geomorphology of the bed of a tidewater glacier. Several of these landform elements have been reported from other Svalbard tidewater glaciers known to be of surge type [e.g., *Solheim and Pfirmann*, 1985; *Boulton et al.*, 1996], and we also

present seafloor imagery from offshore of Wahlenbergbreen and Bråsvellbreen in Svalbard which exhibit similar submarine landforms to those from Borebukta (Figure 1). It is the remarkable preservation of the assemblage of landforms imaged using high-resolution swath bathymetry that makes these marine-geophysical data sets particularly informative (Figure 2).

2. Glaciological Background

[6] Some authors have suggested that about 90% of the glaciers and ice-cap drainage basins on Svalbard are of surge type, although only about 100 or so ice masses have been observed to surge over the limited geographical coverage and time period for which observations have been made [*Lefauconnier and Hagen*, 1991; *Hagen et al.*, 1993]. More conservative estimates indicate that 30–50% of glaciers have either been observed to surge or exhibit ice-surface deformation structures that point to past instabilities [*Dowdeswell et al.*, 1991; *Hamilton and Dowdeswell*, 1996; *Jiskoot et al.*, 2000]. For tidewater glaciers in particular, where much mass is lost by iceberg calving as well as by surface melting and runoff, the ice-deformation structures and looped medial moraines characteristic of past surge activity [*Lawson et al.*, 1994] may be lost soon after the active phase has ended through the calving process, which allows glacier termini to retreat relatively rapidly during the quiescent phase of the surge cycle [*Hodgkins and Dowdeswell*, 1994].

[7] Geological and geophysical investigations at the marine margins of several Svalbard glaciers that have been observed to surge have shown the presence of submarine landforms that appear to be particularly characteristic of surging glaciers. *Solheim* [1986, 1991] and *Boulton et al.* [1996] have reported distinctive rhombohedral or polygonal ridge structures on the now-exposed seafloor close to the maximum surge extent of two drainage basins of the 8200 km^2 Austfonna ice cap in Nordaustlandet and offshore of Sefströmbreen in Isfjorden, Spitsbergen [*De Geer*, 1910; *Schytt*, 1969; *Solheim and Pfirmann*, 1985; *Dowdeswell*, 1986; *Hagen et al.*, 1993].

[8] Our main study area is Borebukta, a 10-km-long inlet on the north side of Isfjorden in Svalbard (Figure 1). The fjord is approximately 1 km wide and 50 m deep at the present glacier front, and widens and deepens to a maximum depth and width of 70 m and 4 km, respectively, about 5 km from the present ice front (Figure 2). Farther out it shallows once more to a 10-m-deep threshold at the fjord entrance. The tidewater glaciers Borebreen and Nansenbreen drain into Borebukta. Borebreen is about 22 km long and has a drainage area of 120 km^2. Nansenbreen is 14 km long and has an area of 45 km^2 [*Hagen et al.*, 1993]. The position of the terminus of Borebreen has been mapped for several intervals since its initial observation by *De Geer* [1910] from oblique and vertical aerial photographs acquired by the Norsk Polarinstitutt since the 1930s (Figure 2).

[9] We observe a zone of the seafloor covered with rhombohedral ridges offshore of Borebreen in Borebukta (Figure 3e). These distinctive ridges form a zone about 3 km wide located just inside a prominent terminal moraine ridge about 10 km from the present ice front. They are almost identical in form and scale to those described above from offshore of surging basins of Austfonna and Sefströmbreen [*Solheim*, 1986, 1991; *Boulton et al.*, 1996], suggesting that Borebreen may be a surge-type glacier. It should also be noted that Borebreen's neighbor, the smaller Nansenbreen, was observed to surge in 1947 and has retreated about 3 km from this maximum position (Figure 1) [*Liestøl*, 1969].

[10] In addition, many of the other tidewater glaciers on the north side of Isfjorden are also known to be of surge type [*Liestøl*, 1969; *Hagen et al.*, 1993]. In particular, Wahlenbergbreen is a surge-type glacier for which we also have swath-bathymetric data showing the former glacier bed in the adjacent Yoldiabukta (Figure 1). It last surged around 1908 [*Hagen et al.*, 1993]. Prior to this surge, in 1896, *De Geer* [1910] had noted that the tidewater margin of Wahlenbergbreen was some kilometers behind its 1908 post-surge position.

3. Geophysical Methods

[11] The Norwegian Hydrographic Service has collected detailed bathymetric data from the waters around Svalbard from the research vessel *Harald Sverdrup*. The geophysical equipment used was a Kongsberg Simrad EM1002 multibeam echo-sounder or swath-bathymetry system. The system has a frequency of 97 kHz and emits 111 individual beams.

[12] The whole seafloor of Borebukta, approximately 30 km^2, has been mapped using this equipment, except those areas which were, at less than 10 m, too shallow for the ship to traverse (Figure 2). We also present and discuss a 15 km^2 area of swath-bathymetric imagery from offshore of Wahlenbergbreen in Yoldiabukta, about 10 km north of Borebukta, and a 4 km^2 area of seafloor imagery offshore of the Austfonna ice cap in Nordaustlandet (Figure 1).

[13] The water depths in these areas are everywhere less than about 100 m. This gives a very close spacing of beams on the seafloor, allowing the absolute elevation data to be gridded with a horizontal cell size of 1 m. Errors in vertical elevation of the seafloor are about ±50 cm. The raw multibeam data were processed using Kongsberg Simrad Neptune software, and corrections for tidal variations and the sound-velocity structure of the water column were also applied.

4. Submarine Landforms in Borebukta

4.1. Assemblage of Landforms and Relative Age of Deposition

[14] The full swath-bathymetric data set showing the seafloor morphology of Borebukta is illustrated in Figure 2. The image comprises a series of several styles of landform, many of them superimposed upon one another. The superimposition of these landform elements allows the reconstruction of the relative age of deposition of each element, with the oldest successively overlain and cross-cut by younger landforms. We have analyzed the swath imagery of Borebukta in order to extract and map a number of specific landform elements in Figures 3b to 3f. The gross form, spatial distribution and relative timing of deposition of each element are shown in this figure. The suite of landforms is also well dated by reference to

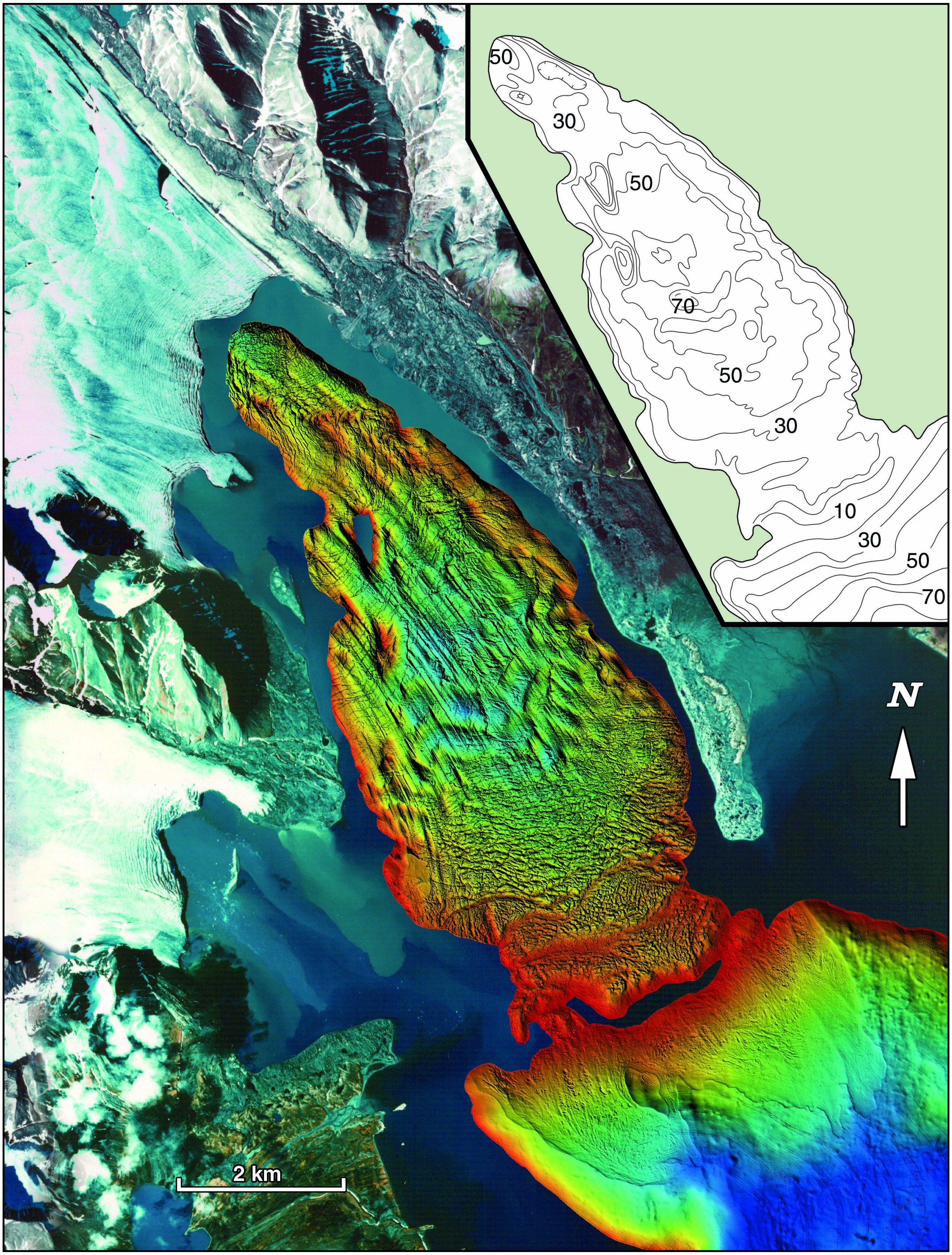

Figure 2. Full swath-bathymetric data set for the 30 km^2 seafloor of Borebukta. The main diagram shows a shaded relief image which highlights the submarine landforms. In the inset, the bathymetry of Borebukta is contoured at 10-m intervals. The glaciers and adjacent land are from Norsk Polarinstitutt false-color aerial photographs acquired in 1990.

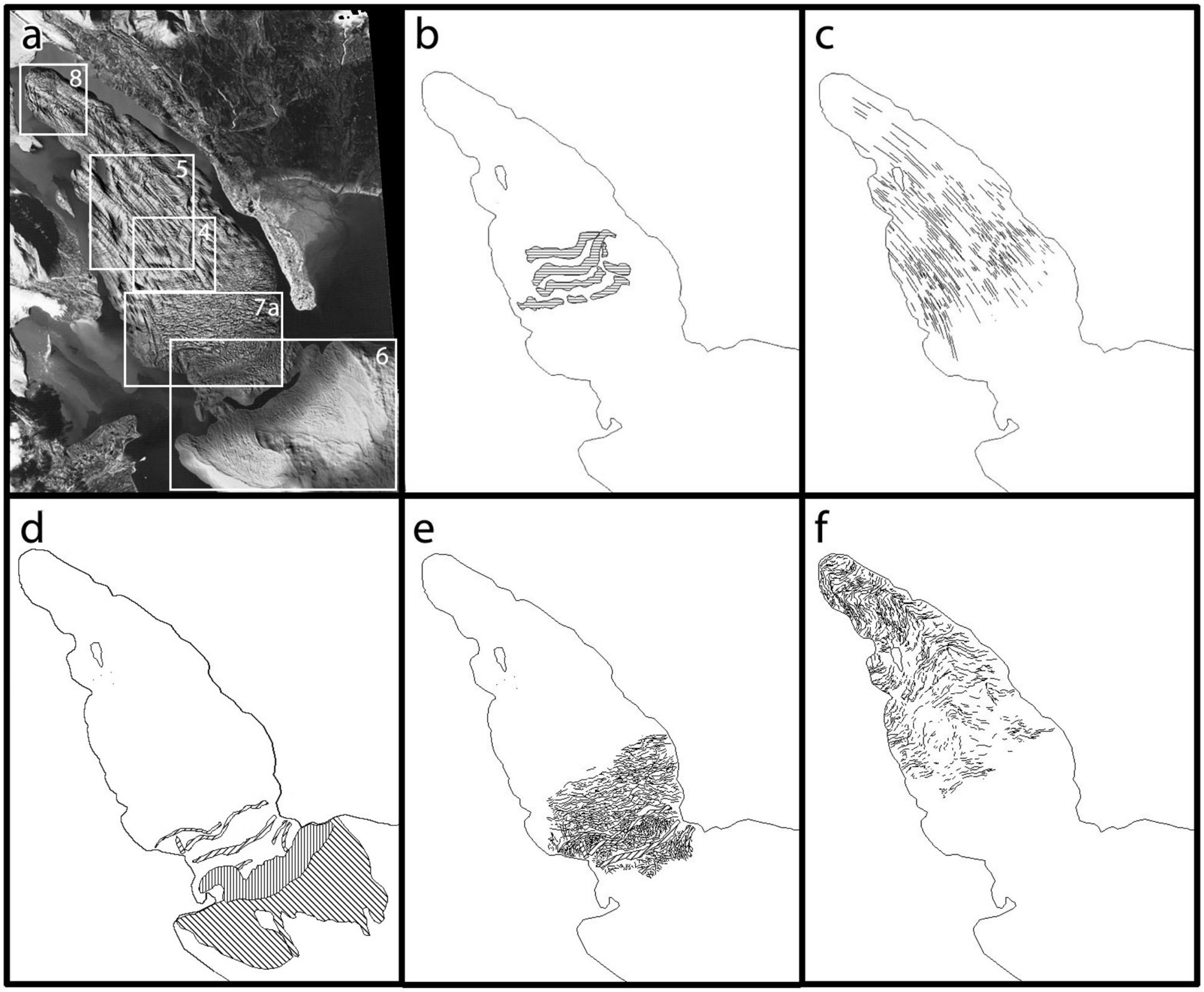

Figure 3. A spatial and temporal analysis of the individual submarine landform elements making up the landform assemblage or landsystem in Borebukta. The landform elements (Figures 3b–3f) are shown in the stratigraphic order in which they where formed, inferred from cross-cutting relationships and superposition to distinguish the relative age of each element. (a) Swath-bathymetric shaded relief image of Borebukta showing the suite of superimposed landforms and location of subsequent figures. (b) Large transverse ridges. (c) Streamlined bedforms. (d) Terminal ridges, together with flow lobes beyond. (e) Rhombohedral ridges and associated larger transverse ridges. (f) Regularly spaced transverse ridges, the youngest individual landform element.

aerial photographs showing former positions of the terminus of Borebreen (Figure 1).

[15] The main landforms observed in the sediments forming the present sea bed of Borebukta, and therefore the former subglacial bed of Borebreen, are, from oldest to youngest in relative age of deposition: (1) a set of four large transverse ridges (Figure 3b); (2) a series of curvilinear streamlined bedforms oriented parallel to the former direction of ice flow in the fjord (Figure 3c); (3) large terminal ridges, together with flow lobes immediately beyond the terminal ridge which appear to have been deposited contemporaneously with the ridge (Figure 3d); (4) a series of interconnected rhombohedral ridges (Figure 3e); and (5) a series of fairly evenly spaced small transverse ridges (Figure 3f). Each of these landform elements is now described and illustrated in detail, followed by genetic interpretation in terms of depositional processes.

4.2. Large Transverse Ridges: Recessional Moraines Overridden by an Advancing Glacier

4.2.1. Observations

[16] In the central and deepest parts of Borebukta, four slightly sinuous ridges occur, orientated approximately perpendicular to the main fjord axis (Figure 4). The longest ridge is 2 km in length from northeast to southwest and crosses the deepest part of Borebukta. The width of the individual ridges is about 300 m and the average distance between the ridges is around 400 m. The average ridge

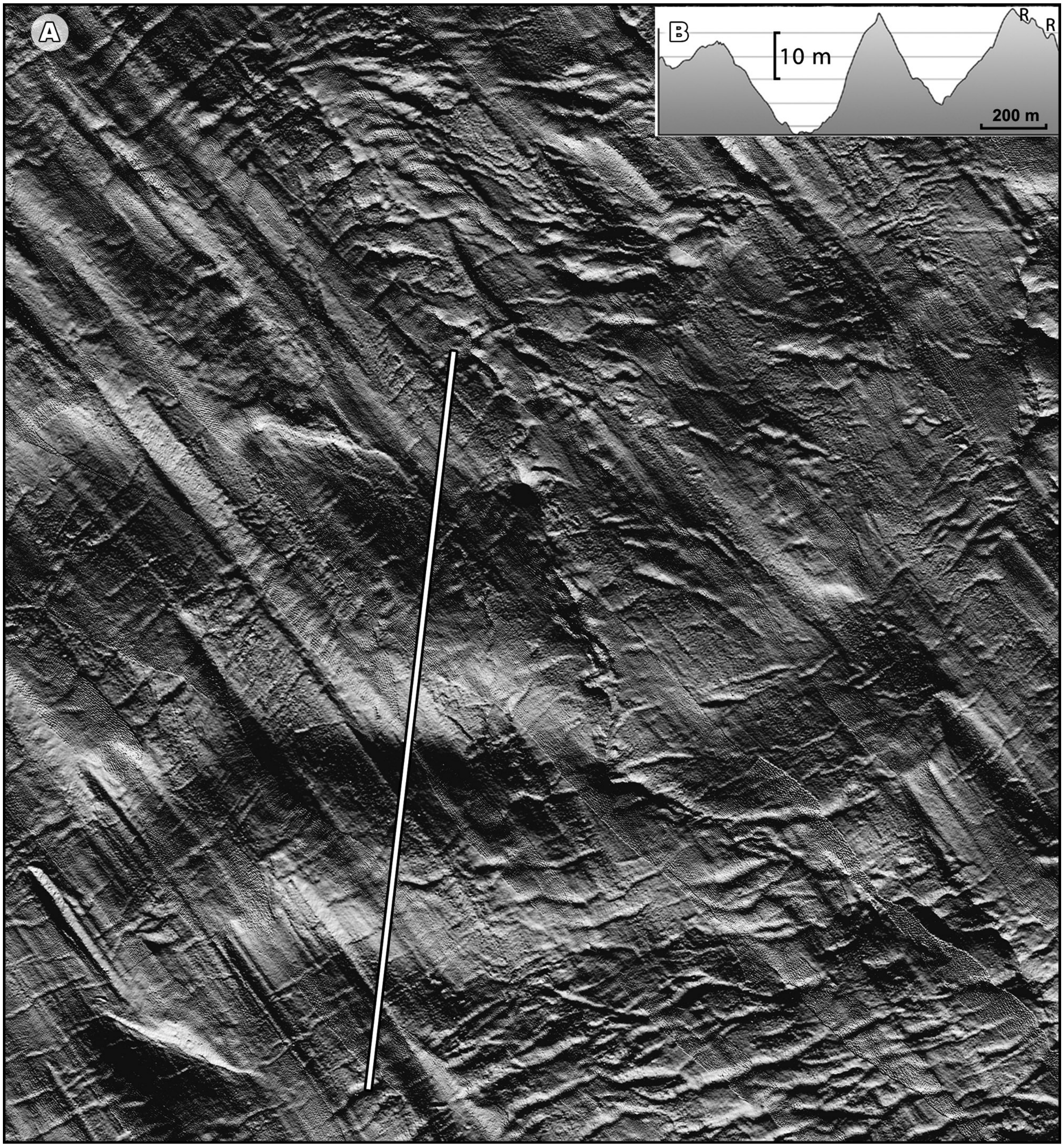

Figure 4. (a) Detailed swath-bathymetric shaded relief image of large transverse ridges with streamlined bedforms and small transverse ridges superimposed. The image is located in Figure 3a. (b) Vertical profile across three transverse ridges. Note the two small recessional moraine ridges on top of the southernmost large ridge (marked R).

height is 15 m, and the maximum is about 20 m. These ridges are crosscut by several of the other landform elements observed in Borebukta that are described subsequently, implying that they are the oldest morphological features in the assemblage that we have imaged.

4.2.2. Interpretation

[17] The ridges are interpreted to be terminal or recessional moraine ridges from a previous ice advance. If Borebreen is of surge type, they also mark the extent of the previous surge of Borebreen. No information about the absolute age of the ridges exists. Both streamlined glacial lineations and small recessional moraines are found on top of the ridges (Figure 4). On the basis of their morphology, it appears that parts of the ridges have been modified and incorporated into the streamlined bedforms that overlie them. Such a process presumably took place during the most recent advance of Borebreen.

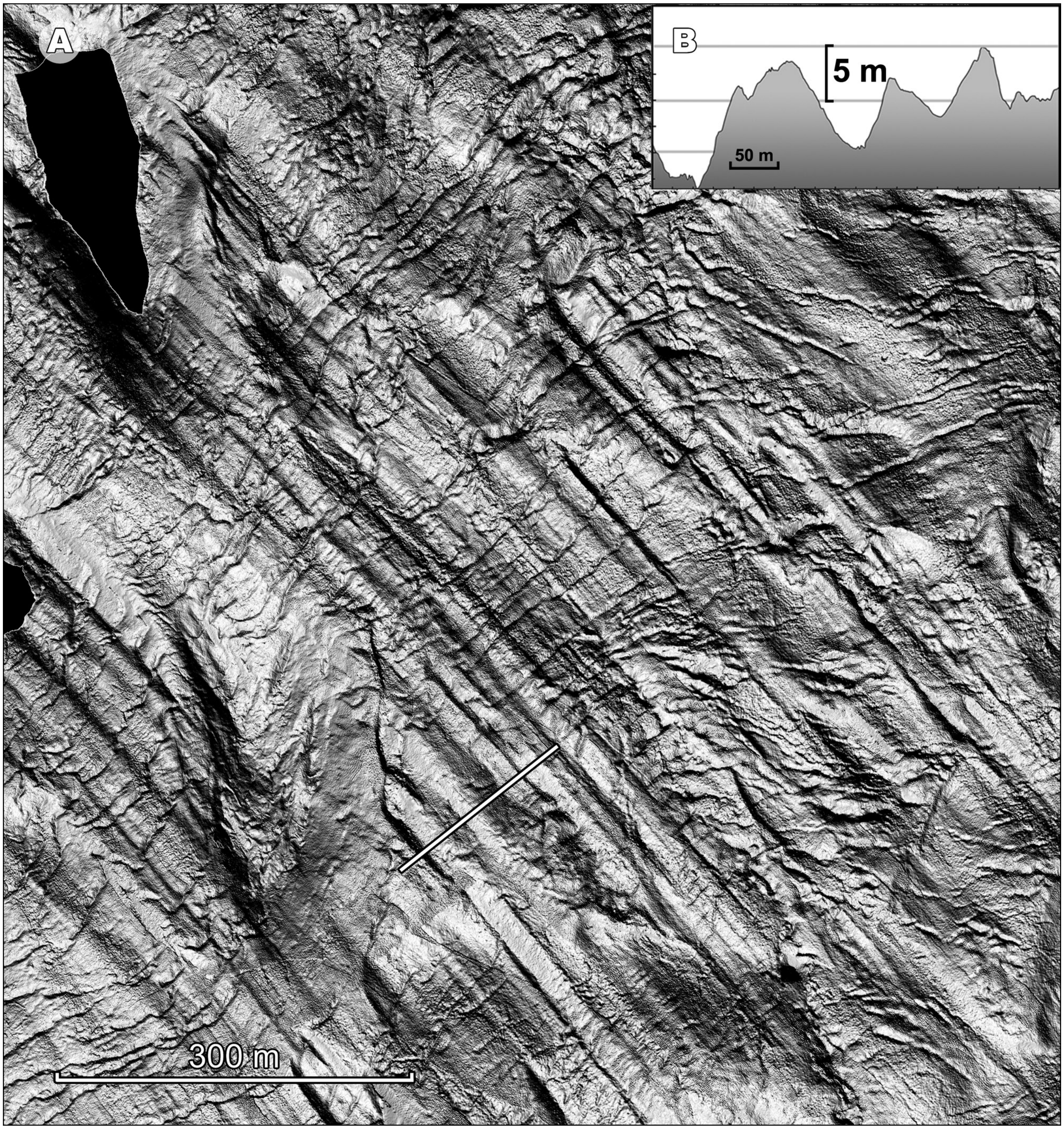

Figure 5. (a) Detailed swath-bathymetric shaded relief image showing streamlined sedimentary bedforms orientated along the long axis of Borebukta. The image is located in Figure 3a. (b) Vertical profile across streamlined bedforms.

4.3. Streamlined Bedforms: Glacial Lineations

4.3.1. Observations

[18] Streamlined sedimentary bedforms, with individual lineations of up to several kilometers in length, are imaged on the sedimentary sea bed of Borebukta (Figures 4 and 5). The bedforms are aligned parallel to one another and are oriented in a curvilinear pattern along the main axis of the fjord (Figure 2). The lineations are present over the bulk of the seafloor between the present tidewater glacier front and the large moraine ridge that marks the limits of recent advance (Figures 2 and 3c). Within 2 km of this large ridge, we observe no lineations and they may have been obscured by subsequently deposited landforms (Figures 2 and 3e). Elsewhere, the streamlined lineations clearly overprint older transverse ridges (Figure 4). In some areas it appears that the underlying ridges may have been in part eroded and modified to form the lineations. The height of the lineations occasionally reaches 10 m, but amplitudes are generally less

than 5 m (Figure 5b). The distance between each ridge crest is usually between 20 and 80 m, sometimes reaching 100 m. The continuous length of individual bedforms is up to about 2.5 km.

4.3.2. Interpretation

[19] The streamlined glacial lineations were formed subglacially during an ice advance about 100 years ago. At the simplest level of explanation, the movement of ice over a sedimentary substrate often produces streamlined bedforms aligned parallel to the ice flow on a variety of scales [*Clark*, 1994]. During the period of high velocities associated with tidewater glacier advance and, possibly, the active phase of the surge cycle, the upper surface of the till was streamlined into glacial lineations, at least to a depth defined by the amplitudes of the lineations (locally up to 10 m, but normally less than 5 m). These amplitudes are similar to those observed for well-preserved lineations on Antarctic shelves [*Dowdeswell et al.*, 2004]. Streamlined subglacial bedforms, often referred to as 'mega-scale glacial lineations' [*Clark*, 1993, 1994], have been used to infer the former presence of fast-flowing ice streams on continental shelves in both the Arctic and Antarctic [e.g., *Shipp et al.*, 1999, 2002; *Canals et al.*, 2000; *Wellner et al.*, 2001; *Ó Cofaigh et al.*, 2002; *Ottesen et al.*, 2002, 2005].

4.4. Large Ridges: Terminal Moraines

4.4.1. Observations

[20] Four large transverse ridges mark the shallow threshold of Borebukta and occur between 8 and 10 km from the present tidewater glacier front (Figures 2 and 3d). The ridges are subparallel to each other, and perpendicular to the fjord axis. The outermost of the four ridges is the largest, at about 600 m wide, and up to 15 m high (Figure 6a). Its cross-sectional shape is asymmetrical, with a smooth outer or distal slope and a steeper inner or proximal slope. The crest of the ridge is slightly arcuate and is located in water depths of between 8 m and 15 m. This explains the lack of swath imagery over the ridge, since the survey ship could not cross this very shallow area (Figure 2).

[21] A second ridge is located about 700 m farther inside with its crest at a water depth between 15 and 20 m (Figures 6a and 7a). The ridge is 300 to 400 m wide and up to 8 m high. The third ridge is located 300–700 m behind the second ridge and is about 150 m wide and up to 8 m high and is located in water depths between 15 and 28 m. This ridge is arcuate and crosses the whole fjord. A fourth ridge covers the western half of the fjord and is located 100–400 m inside the third ridge. It is 60–80 m wide, up to 8 m high and with an arcuate form. The top of this ridge is between 18 and 26 m below sea level.

4.4.2. Interpretation

[22] These large submarine ridges are interpreted as a terminal moraine complex marking the recent extent of ice in Borebukta. The outermost ridge marks the maximum extent of the ice advance. We have no acoustic evidence on the structure of the submarine moraines in Borebukta, but penetration would probably be difficult owing to their likely heterogeneous grain size [e.g., *Sexton et al.*, 1992].

[23] Terminal moraine complexes have been observed in many inner fjord locations in Svalbard. They are associated with either recent surge activity or the maximum ice extent reached at the end of the cold "Little Ice Age" in the archipelago [e.g., *Liestøl*, 1976; *Elverhøi et al.*, 1983; *Boulton*, 1986; *Sexton et al.*, 1992]. On land, the internal structure of surge moraines has been examined in some detail. An origin for the surge moraine at Sefströmbreen has been inferred to involve the extrusion and pushing of soft muds during ice advance and subsequent settling during stagnation [*Boulton et al.*, 1996]. Large thrust blocks and sheets have been reported from sections through some surge moraines in Svalbard and Iceland [e.g., *Sharp*, 1985b; *Bennett et al.*, 1996] although some of these features have also been interpreted as a product of debris squeezing into crevasses [*Evans and Rea*, 2004].

4.5. Lobe-Shaped Landforms: Glacigenic Debris Flows

4.5.1. Observations

[24] Beyond the outermost large transverse ridge that divides Borebukta from Isfjorden (Figure 2), several sediment lobes are found (Figure 6). They appear to have sources on the distal flank of the major ridge marking the farthest recent ice advance into the fjord. The ice-distal side of the terminal ridge has a slope of about 1.5°. The lobes extend up to 2.5 km downslope from about 15 m to a maximum of 90 m in water depth. Several downslope-oriented flow structures are imaged on the seafloor, and a number of blocks a few meters wide can be seen in the western lobe in particular. On the basis of several 3.5 kHz profiles, *Plassen et al.* [2004] have observed two overlying sediment lobes with maximum thicknesses of 25 m and 15 m (Figure 6b).

4.5.2. Interpretation

[25] The lobe-shaped forms are interpreted as debris flows deposited beyond the outermost ridge formed when Borebreen last advanced. They originate on the outer slope of the terminal ridge and are probably made up of debris which has been extruded from the tidewater glacier terminus. Curvilinear features on the debris flow surfaces suggest the direction of flowage during lobe emplacement (Figure 6a).

[26] The acoustic stratigraphy of *Plassen et al.* [2004] records two superimposed debris lobes. The older lobe may possibly be related to the recessional moraine ridges observed in Borebukta that were overridden and modified by the last ice advance (Figure 4). However, we have no age control on these features and the two flows could simply represent multiple mass-wasting events.

4.6. Rhombohedral Ridge Systems: Basal Crevasse-Fill Ridges

4.6.1. Observations

[27] Between and inside the four large transverse ridges, a series of much smaller intersecting linear ridges oriented in several directions forms a reticulate or rhombohedral pattern (Figures 3e and 7a). The ridges have vertical relief on the order of 5 m (Figure 7a). Ridge directions vary, but most often the directions are subparallel or subperpendicular to the larger terminal ridges. These rhombohedral ridges extend in a zone about 3 km wide inshore from the bathymetric threshold delimiting the outer margin of Borebukta (Figures 2 and 3e).

4.6.2. Interpretation

[28] We interpret these rhombohedral ridges to be a product of soft sediment squeezing into crevasses formed at the glacier bed [*Van der Veen*, 1998]. Tidewater glaciers,

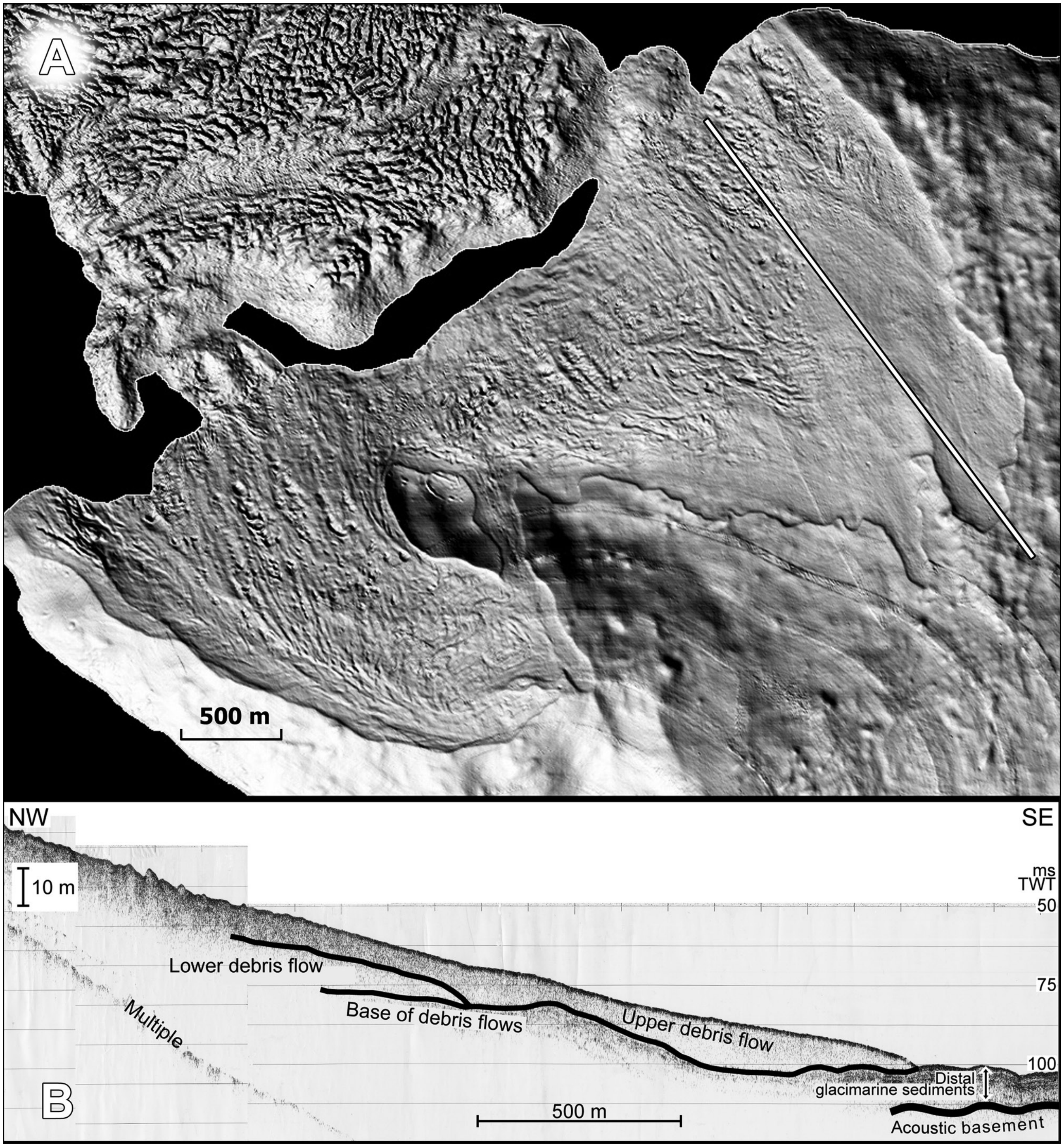

Figure 6. (a) Detailed swath-bathymetric shaded relief image showing terminal transverse ridge complex and flow lobes beyond. The image is located in Figure 3a. (b) The 3.5 kHz penetration echo-sounder profile section through lobes (adapted from *Plassen et al.* [2004]).

especially those in the active phase of the surge cycle, are well known to be heavily crevassed [*Meier and Post*, 1969]. In addition, fast flow is often associated with the motion of a deforming bed of water-saturated sediments, where processes of debris squeezing are likely to be operating [*Boulton et al.*, 1996]. The presence of a deformable bed is particularly likely where tidewater glaciers are surging across fjord floors made up of glacimarine sediments.

[29] Similar landforms have been reported from the exposed beds of known surge-type glaciers in Iceland and Svalbard, and the same conclusions have been drawn concerning their origin. Crevasse-fill ridges were described from the retreating margin of a surging Icelandic glacier by *Sharp* [1985a]. *Solheim and Pfirmann* [1985] have reported a similar rhombohedral ridge pattern from beyond the present margin of Bråsvellbreen on Nordaustlandet, which surged in the 1930s. We have obtained recent high-

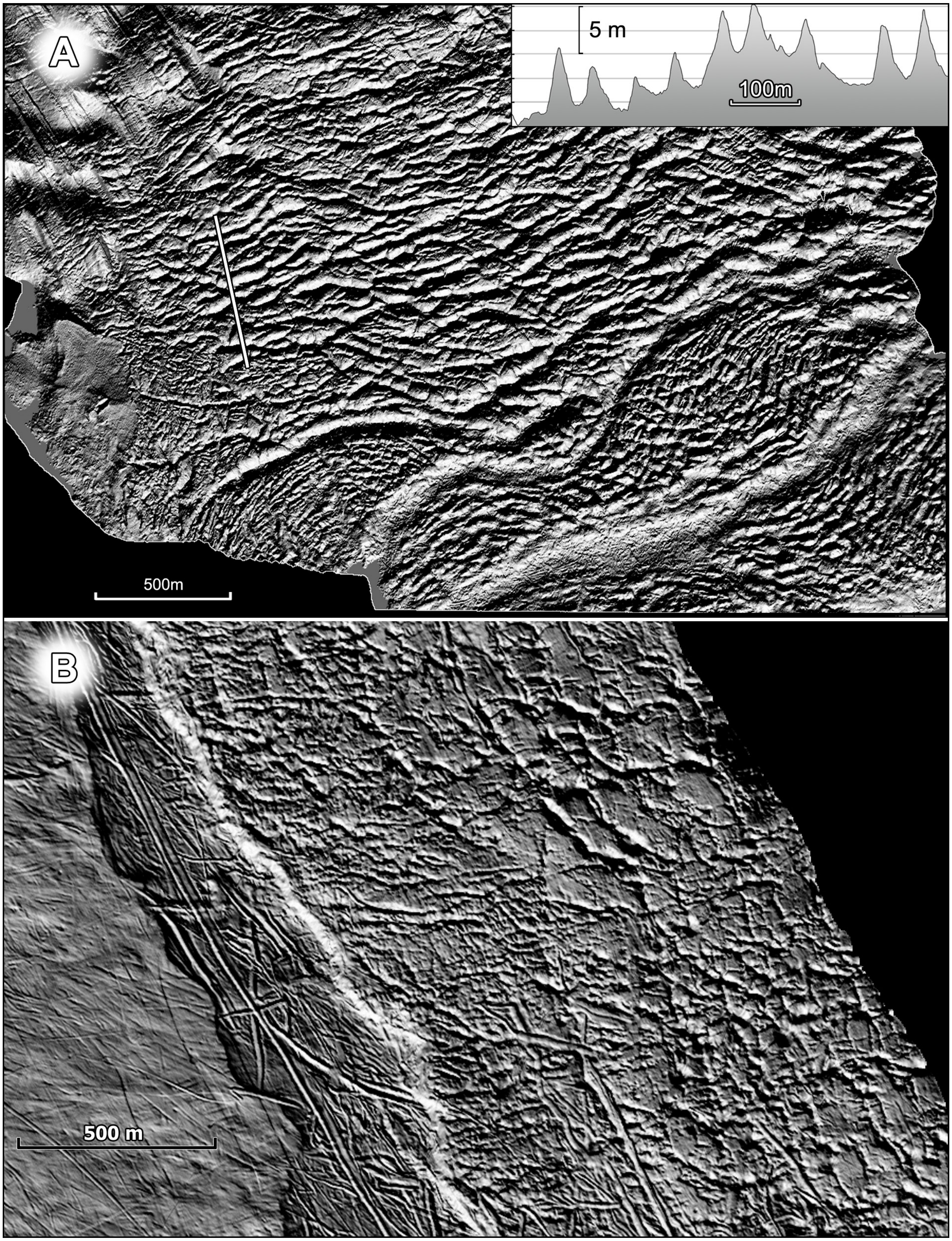

Figure 7. Swath-bathymetric shaded relief images of submarine rhombohedral ridges. (a) Rhombohedral ridges in Borebukta, with vertical profile inset. The image is located in Figure 3a. (b) Similar rhombohedral ridges off Bråsvellbreen in Nordaustlandet, Svalbard (Figure 1). Note the linear and curvilinear scour marks produced by iceberg keels, which are more common to the left of the moraine ridge marking the limit of the last surge in the 1930s. Rhombohedral ridges are present to the right, on the ice-proximal side of this ridge.

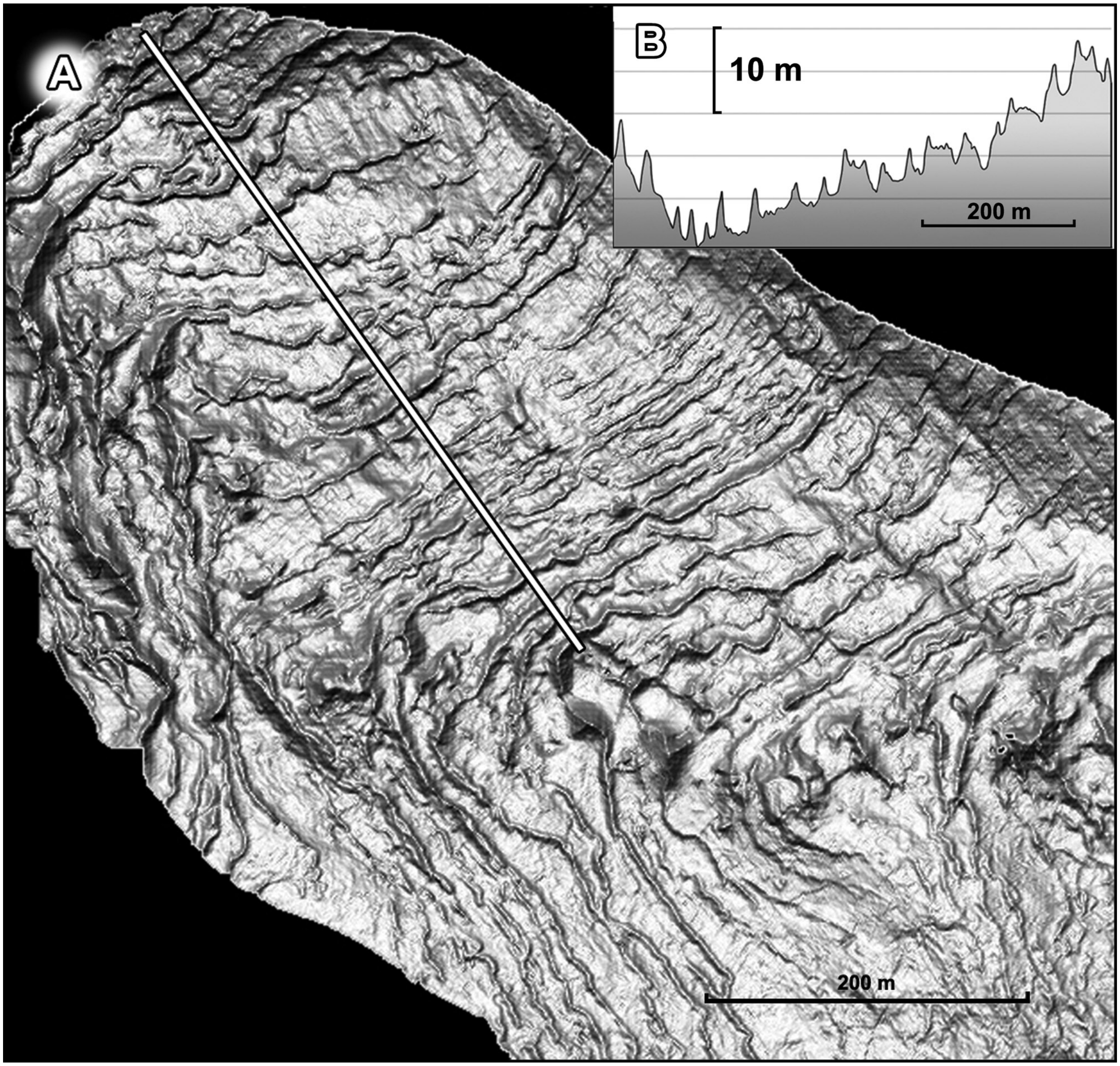

Figure 8. (a) Detailed swath-bathymetric shaded relief image showing small transverse ridges interpreted as annual retreat moraines in the inner part of Borebukta. The image is located in Figure 3a. (b) Vertical profile across the ridges.

resolution swath-bathymetric imagery of these features (Figure 7b). Aerial photographs taken at the time of the surge show a similar pattern of surface crevasses on Bråsvellbreen [*Schytt*, 1969]. *Boulton et al.* [1996] have mapped an almost identical reticulate pattern of ridges and intervening hollows near Coraholmen in inner Ekmanfjorden, northeast of Borebukta (Figure 1). They infer a similar process of formation beneath the surging glacier Sefströmbreen.

[30] If rhombohedral moraines are associated with the filling of basal crevasses during initial postsurge ice stagnation, then they would provide a landform indicative of surge-type glaciers rather than tidewater glaciers that retreat due to enhanced calving into deepening water inshore. Enhanced calving takes place due to increased buoyancy and ungrounding, and rhombohedral moraines could not form by squeeze into basal crevasses in such circumstances.

4.7. Regularly Spaced Transverse Ridges: Annual Push Moraines Formed During Ice Retreat

4.7.1. Observations

[31] A series of small transverse ridges, of relatively regular spacing, is observed in Borebukta (Figures 3f and 8). The ridges are located subparallel to one another. They are generally perpendicular to the fjord axis, but near the fjord margins tend to turn towards the fjord walls and to follow the bathymetric contours. This gives a slightly arcuate form to most of the ridges across the fjord. The ridges crosscut the streamlined linear bedforms described

above (Figure 5), and are thus the youngest landforms on the seafloor of Borebukta (Figure 3).

[32] The transverse ridges are most common in the inner part of the bay, closest to the modern margin of Borebreen (Figure 2). Here nearly continuous ridges can be traced right across the fjord (Figure 8a). In the innermost, ice-proximal 600 m of our imagery, about 20 ridges are present, with an average spacing of 30 m (Figure 8). The average width of the transverse ridges is about 15 m, and mean height is 3 m (Figure 8b). In the central and deepest part of Borebukta, where water depths are between 40 m and 70 m, the distance between the ridges is an average of 90 m, and the width and height of each ridge are 25 m and 2 m, respectively.

4.7.2. Interpretation

[33] The transverse ridges are interpreted as a series of small push moraines that are produced annually when a tidewater glacier terminus undergoes minor readvance each winter during more general retreat [*Boulton*, 1986]. Winter readvance in tidewater glaciers is a result of the suppression of iceberg calving due to the presence of winter shorefast sea ice. Indeed, minor deformation of the sea-ice cover is observed at the margins of many Svalbard tidewater glaciers in winter. An implication of this interpretation of ridge origin is that Borebreen is not completely stagnant today. This may also be why these ridges are not present in outer Borebukta, close to the moraine ridges marking the terminus of the last advance. Here ice was probably stagnant shortly after the last advance, especially if it was related to surging, producing instead the zone of rhombohedral ridges (Figure 7a).

[34] Evidence that individual transverse ridges are produced annually during general tidewater-glacier retreat comes from mapping ridge locations against the known positions of the ice front of Borebreen at a number of times in the last century. Ice front positions come from early reports and the analysis of oblique and vertical aerial photographs acquired by the Norsk Polarinstitutt. The front positions for Borebreen and a number of other glaciers in this part of Isfjorden are mapped in Figure 1. We observed 20 ridges between the ice front positions in 1966 and 1986, 15 ridges between 1948 and 1966, 12 from 1936 to 1948, and 41 from 1910 to 1936. This makes a total of 88 ridges in the period of 76 years from 1910, and 47 ridges in the 50 years from 1936. We suggest that the ice-front position is best know from 1936 onward, since aerial photographs exist only from this date. The 1910 ice front position was inferred from the sketch map of *De Geer* [1910]. These findings confirm that small transverse ridges are formed in most years at the margin of the retreating glacier.

[35] Series of small transverse ridges have been observed on the seafloor beyond a number of retreating Svalbard tidewater glaciers, some of which are known to have surged [e.g., *Boulton*, 1986; *Whittington et al.*, 1997]. However, this landform alone is not necessarily diagnostic of past surges, because sets of such ridges have also been observed at the margins of terrestrial glaciers with no history of surging [e.g., *Sharp*, 1984]. However, because they require grounded ice to form, the ridges could not form where tidewater glaciers retreat catastrophically owing to floatation and enhanced calving into deepening water.

5. Submarine Landforms in Yoldiabukta

[36] Swath-bathymetric imagery is also available from Yoldiabukta, which is about 10 km north of Borebukta (Figure 1). This bay was affected by a documented surge of Wahlenbergbreen in 1908 [*Hagen et al.*, 1993]. A mosaic of the swath data from Yoldiabukta is presented in Figure 9. The assemblage of submarine landforms observed on the seafloor of Yoldiabukta is now described in the order in which each landform element was deposited, which can be inferred from the crosscutting relationships between the landforms.

[37] The oldest features in the swath imagery are a set of ridges that are crosscut by streamlined linear bedforms oriented parallel to the fjord axis and to former ice-flow direction (Figure 9b). The streamlined bedforms appear to have partly remobilized material from the ridges beneath. A terminal moraine complex, coincident with the observed limit of the 1908 surge, is present about 7 km from the modern tidewater margin of Wahlenbergbreen (Figure 9a). Distal of the terminal moraine, several debris flows are imaged forming flow structures down its distal slope (Figure 9c). Immediately behind the terminal moraine, a series of rhombohedral ridges is present (Figure 9d). Closer to the present ice front, a large number of annual retreat ridges are observed, formed transverse to ice-flow direction and subparallel to one another (Figure 9e). These annual retreat ridges are superimposed upon the streamlined bedforms associated with ice advance, and are thus the most recently deposited landforms on the imagery (Figure 9b).

[38] The assemblage of submarine landforms observed in Yoldiabukta (Figure 9) is very similar indeed to that from Borebukta (Figure 2). The same six landform elements are present in each bay (Figures 3 and 9). Each landform is also in the same spatial and stratigraphical relationship in the two bays. Yoldiabukta has been affected by a recent surge of a tidewater glacier that has now retreated from its surge maximum to reveal the seafloor landforms that we have described and interpreted. The considerable spatial and temporal similarities between Yoldiabukta and Borebukta, together with existing descriptions of the landforms observed at the margins of other surge-type glaciers [e.g., *Sharp*, 1985b; *Solheim and Pfirmann*, 1985; *Boulton et al.*, 1996; *Evans and Rea*, 2004], imply that we can use these data sets to construct a model of the landsystem, or assemblage of landforms, associated with Svalbard tidewa-

Figure 9. (a) Swath-bathymetric shaded relief image of submarine landforms in Yoldiabukta, offshore of the tidewater glacier Wahlenbergbreen (Figure 1). Several landform elements are shown in more detail: (b) streamlined linear landforms superimposed on large transverse ridges; (c) submarine debris-flow lobes beyond a large surge terminal moraine shown in Figure 9a and located to the immediate left of this image; (d) rhombohedral moraines interpreted as crevasse-fill ridges formed during immediate postsurge ice stagnation; and (e) regularly spaced transverse ridges inferred to be annual retreat moraines.

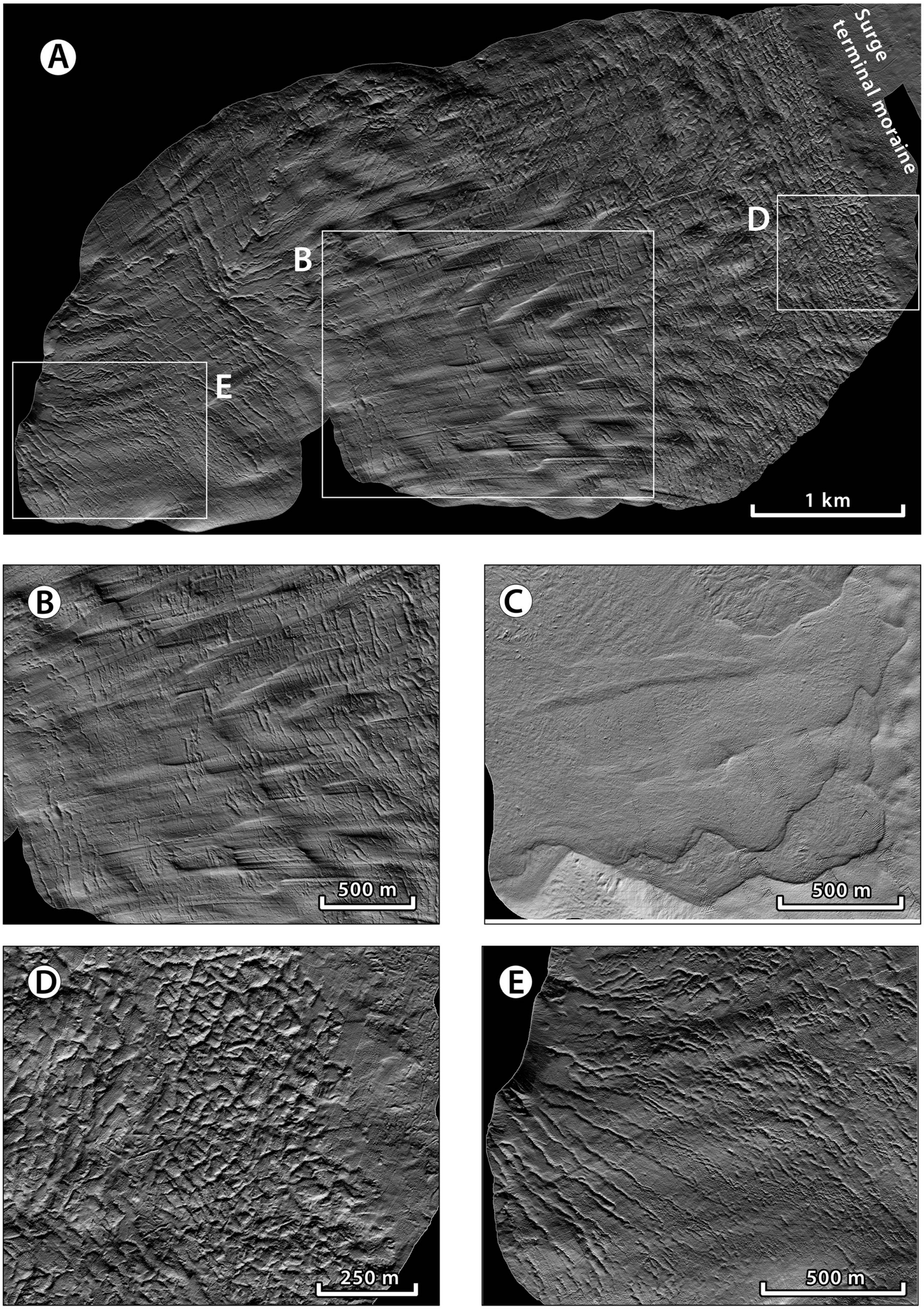

Figure 9

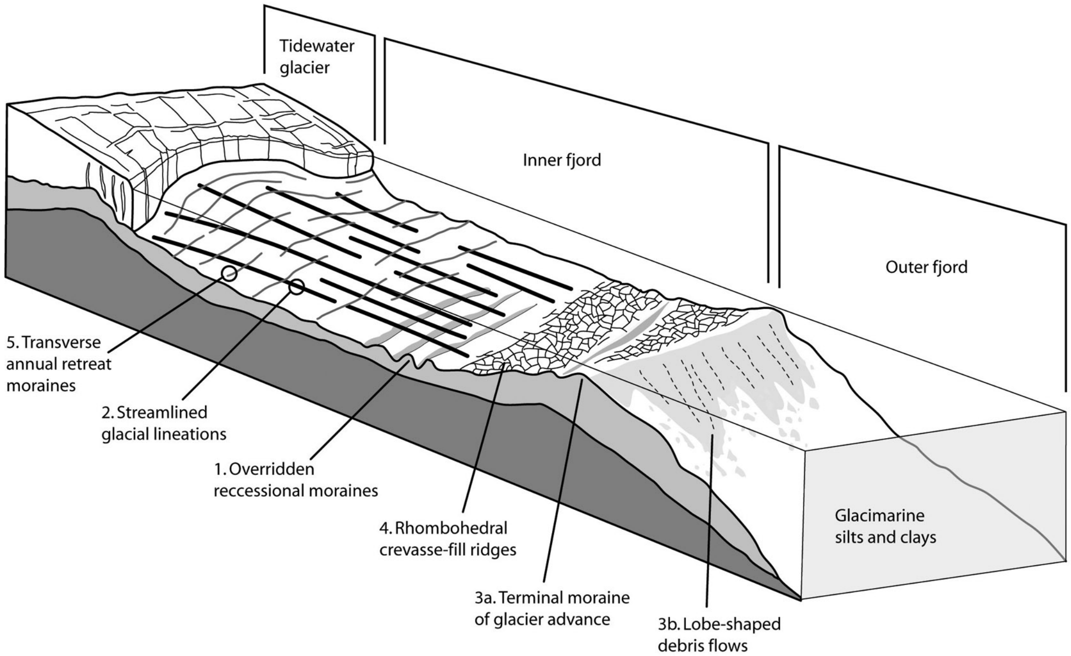

Figure 10. A landsystem model for Svalbard tidewater glaciers of probable surge type. The model is derived from the analysis and interpretation of high-resolution swath-bathymetric imagery from waters beyond several Svalbard glaciers in Borebukta and Yoldiabukta, Isfjorden, and offshore of Austfonna on Nordaustlandet (Figures 2, 3, and 9). The relative timing of deposition is indicated by the numbering of individual landform elements, with the lowest numbered deposited first and often crosscut by subsequent depositional landforms.

ter glaciers, several of which have a known history of surging.

6. Discussion: A Simple Landsystem Model for Tidewater Glaciers in Svalbard

[39] A simple descriptive landsystem model for Svalbard tidewater glaciers is set out in Figure 10. The model is derived from the observations and interpretations of the swath-bathymetric data presented from Borebukta and Yoldiabukta on the north side of Isfjorden in Svalbard (Figure 1). These submarine landforms, together with data collected offshore of Austfonna on Nordaustlandet (Figure 7b), record a very similar geomorphic assemblage. The submarine landforms beyond the present margins of Borebreen and Wahlenbergbreen have a particularly striking resemblance, suggesting that the landsystem model has some generality.

[40] Such marine morphological data are especially useful in the construction of landsystem models because of the lack of landscape modification subsequent to deposition at former glacier beds. This contrasts with the much more fragmentary and often reworked record observed at the margins of glaciers ending on land [*Evans and Rea*, 2004]. The retreat of tidewater glaciers, whether of surge type or otherwise, is likely to take place mainly by mass loss through iceberg calving. The subglacial landforms will, therefore, not be subject to the ice-stagnation and meltwater reworking that are so common at the retreating margins of glaciers ending on land.

[41] The submarine landforms that we have imaged in the marine waters around Svalbard are also preserved at or close to the seafloor on the timescale of the past century or so because the sedimentation rate here is relatively low, at a few millimeters to tens of millimeters per year [e.g., *Elverhøi et al.*, 1983; *Dowdeswell and Dowdeswell*, 1989; *Dowdeswell et al.*, 1998]. By contrast, in the fjords beyond the sometimes surging tidewater glaciers of southeast Alaska, sedimentation rates are 2 or 3 orders of magnitude higher [*Powell and Molnia*, 1989]. The sediment yield from ice-covered drainage basins in Svalbard and Alaska also varies by about two orders of magnitude [*Hallet et al.*, 1996; *Elverhøi et al.*, 1998]. Any landforms exposed at the seafloor by tidewater glacier retreat will be buried rapidly in an environment of such high sedimentation and the striking landforms we report from Svalbard will not be found offshore of Alaskan tidewater glaciers or in other areas where rates of glacimarine sedimentation are particularly high [*Dowdeswell et al.*, 1998].

[42] The landsystem model we present for Svalbard tidewater glaciers has not only a morphological but also a temporal component. This temporal component is identified in the sequence of numbers assigned to each landform element in Figure 10, with the lowest number deposited first. The crosscutting relationships between the various landform elements identified in Borebukta and Yoldiabukta provide the evidence on which this relative chronology for deposition is based. The sequence of depositional processes appears to be very similar in both bays (Figures 3 and 9).

[43] Earlier planform and landsystem models of the deposits characteristic of surge-type glaciers were based mainly on observations at the exposed beds of glaciers ending on land, where preservation is not as complete as in the marine environment. Nonetheless, the model we present, based on observations of the seafloor at the margins of several tidewater glaciers in Svalbard, has a number of clear similarities with those of earlier workers. *Sharp* [1985b] produced a planform model of several landform-sediment complexes at the margin of an Icelandic surge-type glacier. This included ridges of glacitectonically thrust sediments forming the surge limit, and inside this chaotic hummocky topography and fluted lodgement till together with crevasse-fill ridges. The surging glacier landsystem model of *Evans and Rea* [2004] also included thrust-block moraine, hummocky moraine, flutings and crevasse-squeeze ridges and was based on a number of glaciers ending on land in Iceland and Svalbard. Their thrust-block moraine is similar in gross form to our large terminal ridge or surge moraine, although we have no evidence on internal structure. Their hummocky moraine may be a subaerially degraded form of our rhombohedral or crevasse-fill ridges. Their flutings resemble the lineations or sedimentary bedforms of our study in terms of ice moulding although not in scale. Their crevasse-squeeze ridges, given the location away from the surge terminus, are most similar to the small transverse ridges that we interpret as annual moraines produced by winter readvance at the retreating ice margin. Our interpretation of the processes responsible for these landforms agrees largely with those of *Sharp* [1985a], *Boulton et al.* [1996] and *Evans and Rea* [2004], although we have no direct evidence that our terminal moraines are thrust-induced. We conclude that these earlier models have many landform elements in common with Figure 10. It is the exceptional subaqueous preservation of the landform elements we observe in Borebukta and Yoldiabukta (Figures 2 and 9) that allows a particularly clear view of the landforms associated with Svalbard tidewater glaciers, probably of surge type, and also their sequence of formation.

[44] Finally, it is important to stress that, if the tidewater glacier landsystem model in Figure 10 is used to assist in the identification of former surges in the geological record, care is needed. The individual landform elements, found in isolation, are not necessarily diagnostic of former surge activity. Terminal moraine complexes and annual retreat moraines, for example, are common to both surging and nonsurging glaciers that terminate on land, and streamlined linear bedforms are also associated with ice-stream flow. It may be that the rhombohedral moraines (Figure 7), first reported by *Solheim and Pfirmann* [1985] at the margins of the surging Bråsvellbreen in Nordaustlandet, eastern Svalbard, are most clearly indicative of past surge activity in tidewater glaciers. It has been suggested that these features may have formed by the squeezing of sediments into basal crevasses during initial ice stagnation after the termination of the active phase of the surge cycle [*Solheim and Pfirmann*, 1985]. This is a process that is unlikely to take place where nonsurging tidewater glaciers retreat catastrophically into deep water, because of the rapid ungrounding of the glacier from the underlying substrate during such retreat.

[45] The streamlined glacial lineations observed in Borebukta and Yoldiabukta, Svalbard (Figures 2 and 9), are similar in general morphology to the megascale glacial lineations of *Clark* [1993, 1994]. They are of relatively similar amplitude, of a few meters, but are generally smaller in length and width. Our study demonstrates that streamlined glacial lineations, and not simply smaller-scale flutes [*Evans and Rea*, 2004], are present not only beneath ice streams that are tens of kilometers wide and tens to hundreds of kilometers long, but are also found beneath smaller tidewater glaciers of probable surge type.

7. Conclusions

[46] 1. Marine-geophysical investigations at the marine margins of several Svalbard tidewater glaciers have shown the presence of very well-preserved submarine landforms (Figures 2, 7, and 9). These landforms are essentially unmodified since initial deposition because they have not been subjected to postdepositional subaerial processes of erosion and periglacial activity.

[47] 2. The major submarine landform elements are as follows (Figures 3 and 9): (1) several large transverse ridges, probably recessional moraines overridden by a subsequent glacier advance; (2) a series of streamlined bedforms orientated parallel to former ice flow and formed subglacially during fast flow and rapid ice advance; (3) a large terminal moraine complex marking the further recent extent of a tidewater glacier, together with debris flow lobes on the distal side of the terminal ridge (Figure 3d); (4) a series of interconnected rhombohedral ridges, thought to be a product of soft sediment squeezing into crevasses, probably produced during immediate postsurge ice stagnation; and (5) a series of fairly evenly spaced small push moraines, produced annually when a tidewater glacier terminus undergoes minor readvance each winter during more general postsurge retreat.

[48] 3. A simple descriptive landsystem model for tidewater glaciers of likely surge type is given in Figure 10. The submarine landforms beyond the present margins of Borebreen and Wahlenbergbreen have a particularly striking resemblance (Figures 2 and 9), suggesting that the landsystem model has some generality. Earlier planform and landsystem models of the deposits characteristic of surge-type glaciers were based mainly on observations at the exposed beds of glaciers ending on land, where preservation is less good than in the marine environment. Even so, the model we present has several clear similarities with those of earlier work in Iceland and Svalbard [*Sharp*, 1985b; *Evans and Rea*, 2004].

[49] 4. Our landsystem model for tidewater glaciers of probable surge-type has a temporal component (Figure 10). The crosscutting relationships between the various landform elements identified in Borebukta and Yoldiabukta provide the evidence on which this relative chronology for deposition is based. The sequence of depositional processes appears to be very similar in both bays (Figures 3 and 9).

[50] 5. Our study shows that streamlined megascale glacial lineations [*Clark*, 1993] are found beneath tidewater glaciers of probable and known surge type as well as beneath ice streams that are tens of kilometers wide and tens to hundreds of kilometers long [*Clark and Stokes*, 2003].

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