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The Bothnian Sea ice stream: early Holocene retreat dynamics of the south-central Fennoscandian Ice Sheet

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The Gulf of Bothnia hosted a variety of palaeo-glaciodynamic environments throughout the growth and decay of the last Fennoscandian Ice Sheet, from the main ice sheet divide to a major corridor of marine- and lacustrine-based deglaciation. Ice streaming through the Bothnian and Baltic basins has been widely assumed, and the damming and drainage of the huge proglacial Baltic Ice Lake has been implicated in major regional and hemispheric climate changes. However, the dynamics of palaeo-ice flow and retreat in this large marine sector have until now been inferred only indirectly, from terrestrial, peripheral evidence. Recent acquisition of high-resolution multibeam bathymetry opens these basins up, for the first time, to direct investigation of their glacial footprint and palaeo-ice sheet behaviour. Here we report on a rich glacial landform record: in particular, a palaeo-ice stream pathway, abundant traces of high subglacial meltwater volumes, and widespread basal crevasse squeeze ridges. The Bothnian Sea ice stream is a narrow flow corridor that was directed southward through the basin to a terminal zone in the south-central Bothnian Sea. It was activated after initial margin retreat across the Åland sill and into the Bothnian basin, and the exclusive association of the ice stream pathway with crevasse squeeze ridges leads us to interpret a short-lived stream event, under high extension, followed by rapid crevasse-triggered break-up. We link this event with a *c.* 150 year ice-rafted debris signal in peripheral varved records, at *c.* 10.67 cal. ka BP. Furthermore, the extensive glaciofluvial system throughout the Bothnian Sea calls for considerable input of surface meltwater. We interpret strongly atmospherically-driven retreat of this marine-based ice sheet sector.

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Marine-terminating glaciers and the sectors of ice sheets that are grounded below sea level are widely considered to be vulnerable to unstable retreat. Melting of ice shelves and retreat of the grounding line into deeper waters can, through removal of back-stress and increased ice flux, lead to enhanced calving, drawdown and thinning of the ice sheet interior. In the absence of topographic pinning this can lead to uninterrupted grounding line retreat (Weertman 1974; Thomas & Bentley 1978; Schoof 2007). Such sensitivities have been both modelled and reconstructed in contemporary and palaeo-ice sheet settings (e.g. Payne *et al.* 2004; Jamieson *et al.* 2012; Rignot *et al.* 2014; Jones *et al.* 2015; Pollard *et al.* 2015).

The southern sector of the retreating Fennoscandian Ice Sheet (FIS) comprised a large, aqueous-terminating ice sheet catchment grounded well below sea level throughout its deglaciation (De Geer 1940; Björck 1995; Andrén *et al.* 2011). However, the behaviour and timing of ice sheet retreat through the Baltic and Bothnian basins have thus far been inferred largely indirectly from peripheral, terrestrial-based geological archives (Kleman *et al.* 1997; Hughes *et al*. 2016; Stroeven *et al.* 2016). Glacial geological records from the Baltic Sea and Gulf of Bothnia are scarce, and virtually nothing is directly known about the palaeo-ice flow dynamics of these basins or the stability and pace of ice retreat.

The Baltic Sea and Gulf of Bothnia are shallow basins that presently host epicontinental seas and lie in the heart of the terrain formerly occupied by the FIS (Fig. 1). These basins are considered to have hosted many different ice dynamic regimes throughout the ice sheet's evolution, including the Last Glacial Maximum ice divide (Kleman *et al.* 1997), a major ice stream pathway that has been proposed to be responsible for maximum stage advances and early deglacial readvances at the southern FIS margin (Boulton *et al*. 2001; Kjær *et al*. 2003; Jørgensen & Piotrowski 2003; Kalm 2012), and a corridor of marine- and lacustrine-based retreat (De Geer 1940; Strömberg 1981; Lundqvist 2007). Late-deglacial Finnish ice lobes (Fig. 1), well-defined by morphological indicators, are usually assumed to cross the Gulf of Bothnia southeastwards (Punkari 1980; Johansson *et al*. 2011), while Lundqvist (2007) suggests that this configuration may have been interrupted by a late-stage surge through the Bothnian Sea. While the present-day terrestrial retreat pattern and timing is rather well-documented by extensive De Geer moraines and the Swedish and Finnish clay varve chronologies (e.g. Sauramo 1923; De Geer 1940; Aartolahti 1972; Zilliacus 1989; Strömberg 1989, 2005; Lindén & Möller 2005), the behaviour of the large marine portion of this sector, and the stability of and controls upon its deglaciation are entirely unknown. Numerical ice sheet models report difficulties in avoiding complete collapse of this sector early in deglaciation (Holmlund & Fastook 1995; Clason *et al*. 2014), suggesting a deficiency in understanding of local climate and/or ice dynamic processes governing ice flow here. The dearth of direct geological evidence from the present-day offshore terrain is a major hindrance in this regard.

Recent efforts to access the offshore glacial record are beginning to redress this balance. New, high-resolution multibeam bathymetric data reveal the direct geomorphological imprint of glaciation in the Gulf of Bothnia. Greenwood *et al*. (2015) identify the onset of a palaeo-ice stream in the northern Bothnian Sea from a combined multibeam and terrestrial LiDAR topographic dataset. Here we extend our mapping and interpretation of the glacial landform record to encompass the full acquired multibeam dataset, in order to inform palaeo-ice flow and retreat dynamics in this broad marine basin. We discuss the relationship between these offshore glacial landform assemblages and existing (terrestrial) chronologies for southern FIS retreat, and we consider mechanisms of retreat and possible controls on the dynamics and stability of marine ice sheet sectors.

Data and methods

Multibeam echo-sounding has revolutionised palaeo-glaciological research and understanding of glacial processes in marine sectors. The ability to image the landform imprint of past ice flow over continental shelves, at centimetre-metre scale resolution, has vastly improved our ability to reconstruct the spatial and temporal evolution of ice flow dynamics in environments considered critical to ice sheet stability, but previously limited to stratigraphic investigation and site-specific core sedimentology 'blind' to the regional ice dynamic context. Multibeam-surveyed landform assemblages proximal to contemporary ice sheet outlets have confirmed process-landform relationships previously only hypothesised from terrestrial mid-latitude palaeo-records (e.g. mega-scale glacial lineations and ice streams: Shipp *et al.* 1999; Canals *et al.* 2000). Furthermore, multibeam surveys have revealed previously unrecognised landform assemblages associated with marine grounding lines and the ocean–ice shelf–ice sheet transition zone (e.g. Jakobsson *et al*. 2011; Larter *et al*. 2012; Graham *et al*. 2013).

This study exploits a new, >15 300 km^2 high-resolution multibeam dataset (gridded at a cell-size of 5x5 m), recently acquired in the Gulf of Bothnia for the Swedish Maritime Administration (SMA), under the EU-MonaLisa hydrographic survey project (Fig. 1). Data were collected with a Kongsberg EM2040 and a Reson 7125SV multibeam echosounder (200-400 kHz, beam width 0.25-0.5° x 1°), and processed by the survey contractors Fugro and MMT (Marin Mätteknik) for SMA. This dataset provides areal coverage of approximately 15% of the Gulf of Bothnia and reveals, for the first time, the sub- and pro-glacial imprint of glaciation in this subaqueous sector of the FIS.

Multibeam data were integrated in a GIS with the regional, 500 m gridded Baltic Sea Bathymetry Database (Hell & Öiås 2014), and with Geological Survey of Sweden (SGU) marine datasets including interpreted seismic reflection profiles and seabed surficial geology maps (Lind 2016a, b; Nyberg 2016a, b). Terrestrial data supporting our work comprise SGU's digital surficial geology, sediment stratigraphy and striation databases, and the national LiDAR-based digital elevation model for Sweden, gridded with a cell-size of 2x2 m. Within this integrated data framework, glacial landforms imaged by the multibeam data were systematically and individually mapped by manual digitisation. Landforms were mapped and interpreted within three classes: subglacial bedforms, meltwater landforms and near-marginal landforms. Morphometric properties of the mapped vector features were routinely determined using basic geometric extraction tools in ArcGIS.

Glacial landforms of the Gulf of Bothnia

Multibeam data from the Gulf of Bothnia reveal an extensive, remarkably well-developed and well-preserved glacial geomorphological record (Figs 2-5). This record is dominated by *i)* abundant and distinct assemblages of subglacial bedforms, including drumlins, mega-scale glacial lineations (MSGLs), streamlined bedrock and ribbed moraine, and *ii)* a rich glacial meltwater landform population, comprising both eskers and meltwater channels across a range of spatial scales. There are noticeably fewer ice-marginal landforms: we find no grounding zones wedges, and recessional moraines are observed only closer to land in the northern sectors of the Bothnian Sea. The main central/southern sector of our dataset is instead overprinted by a large field of crevasse squeeze ridges.

Subglacial bedforms

We identify and map 11 182 glacial lineations in the Bothnian Sea (without assigning particular sub-classifications) and 538 ribbed moraines. The lineation population contains three end-member sub-groups: drumlins (Fig. 2A), MSGLs (Fig. 2B), and streamlined bedrock or 'rock drumlins', i.e. lineations with little or no apparent glacial sediment cover. Their lengths range from 50 m to over 20 km, with a mean length of 1036 m. These three lineation species typically form self-contained groups and are rarely intermixed, with the exception of flowset A (below).

Glacial lineations are distributed in distinct clusters or flowsets (cf. Clark 1999; Greenwood & Clark 2009; Fig. 3A), defined by local consistency of individual lineation orientation and morphology. The most striking of these is an assemblage dominated by MSGLs (flowset A; Fig. 3A), which weaves approximately southwards through the central Bothnian Sea. This flow assemblage is an extension of that identified by Greenwood *et al.* (2015) from a more limited dataset, and it forms a continuous set of drumlins and MSGLs (Fig. 2B) that stretch from their onset over the north Swedish coast over 300 km southward. These lineations increase in length in a downstream direction (Fig. 3A), with peak length approximately two-thirds of the distance downstream in the flowset (i.e. ~ 200-230 km along-flow); lineations also increase in length towards the (lateral) centre of the assemblage. In the upper, northern part of flowset A lineations converge towards a main trunk, in which MSGLs possess high parallel conformity (e.g. Fig. 2B), while at the southern (distal) end of the group lineations splay out and display a complex and varied sequence of superimposition and cross-cutting. This flowset ceases abruptly in the central-southern Bothnian Sea (Fig. 3E). Here, the seafloor surface is extremely smooth and interpreted as post-glacial infill of a local basin (Nyberg 2016a, b), though MSGLs are initially still visible beneath the cover (Fig. 3E) and it is likely more are buried by thick post-glacial sedimentation. 10-15 km distal to the MSGL limit there is a group of similarly oriented, S-SSW lineations (flowset A´). However, these lineations are considerably shorter (e.g. 75-1650 m) than the distal MSGLs (100-9000 m), and immediately beyond flowset A´ lies a distinctly different flowset (D) oriented SE. We are therefore confident that flowset A terminates in the central-southern Bothnian Sea, approximately where we observe the MSGLs to dissipate.

West of flowset A, and separated by a gap in multibeam data coverage, assemblage B comprises a suite of SSE-directed drumlins (lengths 245-2680 m; n = 968). In the central-west of this group, a sub-cluster of drumlins take a due S orientation, but overall the drumlins of flowset B are oriented in a tight 30° window between 156-186°. Individual drumlins appear to be locally clustered on larger bodies of till (Figs 2A, 4A, 5A), which are raised above post-glacial sediment infill of the inter-drumlin depressions (cf. Nyberg 2016a, b). Seismo-acoustic profiles show that a number of the drumlins of flowset B are anchored on local bedrock bumps, i.e. the till bodies and/or drumlins contain bedrock 'cores' (Fig. 2A). The bedrock relief is, however, amplified by a greater till thickness over the highs than in the inter-drumlin lows, and the shape and orientation of drumlins is ultimately a reflection of glacial processes rather than bedrock morphology. At the distal end of the assemblage, across a data gap, flowsets B´ and B´´ are broadly aligned with B, though B´´ (SE) cross-cuts B´ (S) (Fig. 3B). These three assemblages are all truncated by flowset A (Fig. 3D): the distal SW-directed MSGLs superimpose SSE-directed drumlins.

Lineation flowsets C and D lie in the southern Bothnian Sea, separated from those described above by post-glacial sediments and by gaps in multibeam coverage (Fig. 3A). Group C comprises lineations of lengths ~ 60-2025 m, oriented SSW towards the northern Uppland coast. To the east, flowset D comprises much smaller crag and tails and streamlined bedrock forms (55-700 m) that are oriented to the SSE with slight convergence towards the Åland Sea. It is not possible to securely identify the relative chronology between these two groups. They are potentially contemporaneous, their contrasting orientations reflecting ice flow divergence around Öregrund-Gräsö and local drawdown through the deep, narrow Åland Sea. Alternatively, potential lineation superimposition would lead us to interpret that flowset D post-dates flowset C.

Towards the west coast of the Bothnian Sea, offshore from Söderhamn, we identify two groups of ribbed moraine (Fig. 2C). These fields of curvilinear ridges indicate ice flow to the SW. The larger forms are typically 200-300 m wide, have 500-1000 m spacing and amplitudes of ~ 8-10 m, whilst a smaller class of ribbed or ridge-like forms lies within and on top of the larger features. Sediment thicknesses in this area derived from Geological Survey seismo-acoustic interpretations (Nyberg 2016a, b) range from 3-50 m; one profile line directly crossing the bedforms indicates a typical sediment thickness of ~ 15 m in inter-rib zones and ~ 23m under rib features, consistent with their morphological interpretation as glacial bedforms shaped in unconsolidated sediments. There is weak drumlinisation or streamlining of the ridge tops, similarly indicating a SW-ward ice flow direction; we group this whole assemblage as flowset E.

Finally, on the north-west flank of flowset A, a small group of lineations with lengths ~ 400 m overprint the underlying MSGLs (typically ~ 1500 m in this area; Fig. 3C). These small lineations are oriented SE, in contrast to flowset A which is directed S-SSW here. This large offset in orientation and in morphology, in contrast to convergent drumlins and lineations elsewhere in this upper region of flowset A, leads us to group the overprinting landforms separately (flowset F).

Meltwater landforms

Eskers and meltwater channels are abundant throughout the Bothnian Sea, and possess a variety of forms, scales, orientations and spatial relationships to other glacial landforms (Fig. 4). We qualitatively recognise three different categories, and provide quantitative descriptions of their typical morphologies in Table 1. The smallest channels (Fig. 4A, B), referred to here as "streams", are often found as isolated (unconnected) features with lengths up to ~ 5 km. They occur largely within areas where post-glacial sediments bury the underlying glacial morphology, but several such streams transition into more pronounced subglacial meltwater conduits (e.g. Fig. 4A, C and E are part of the same system), and we interpret the streams as glacial in origin. Their subdued and fragmented appearance is likely due to post-glacial draping of a more pronounced original morphology and, potentially, a more connected network of subglacial drainage conduits. These smallest meltwater streams typically follow inter-drumlin topographic lows (Fig. 4A), but also occasionally run directly alongside MSGLs in flowset A (Fig. 4B).

Eskers and meltwater channels of comparable sizes form interconnected systems (Fig. 4C), marking alternating dominance of erosive and depositional processes within a single meltwater pathway. Where associated with subglacial bedform fields, these conduits almost exclusively incise (channels) or drape (eskers) underlying forms (Fig. 4C, E) and must therefore post-date the final shaping of the bedform. Both eskers and channels display a braided topology, with no systematic dendritic arrangement. Rather, series of braided channel and esker segments are connected into long (30-80 km), single drainage pathways. Within MSGL flowset A, one single, focussed pathway that is approximately 1-2 km wide comprises several series of locally braiding channels and eskers that together are connected over ~ 50 km (see Fig. 2B). More fragmented, poorly connected segments are dispersed through the rest of this MSGL assemblage. In contrast, meltwater within drumlin flowset B drains through several coherent pathways spaced ~ 3-5 km apart. In the south-west Bothnian Sea, there are broad patches of braiding meltwater channels which cannot be linked together due to gaps in multibeam data coverage. We do, however, identify esker and channel segments that, guided by a pronounced sinuous ridge in the regional 500 m gridded bathymetry model, we interpret as extensions or branches of the Uppsala esker (cf. Hoppe 1961; Nyberg & Bergman 2012; Nyberg 2016a). We do not have sufficient data coverage to trace its headward limit, but this hydrological conduit appears to become more fragmented and dispersed towards the north.

Up to an order of magnitude wider than individual channel-esker conduits, we identify several different forms of large, erosional corridors (Fig. 4F, G). Some of the braided esker-channel pathways could be described as such (e.g. Fig. 2B); others clearly reflect a different scale of erosional process. In the southern Bothnian Sea, amongst lineation flowset D, slightly sinuous conduits exploit crystalline bedrock fracture zones (Fig. 4G). Seismo-acoustic cross-profiles (Nyberg 2016a, b) reveal that beneath an infill of post-glacial and glaciolacustrine clays, glacial silts record use of deep bedrock incisions by glacial meltwater. While it is not possible to determine the degree to which recent subglacial meltwater has created channel relief (as opposed to multi-glaciation or pre-glacial excavation), it is clear that these deep canyons provided pathways for subglacial meltwater drainage and, in places, localised basins for meltwater storage (Fig. 4G).

In some contrast, in the central-northern Bothnian Sea, immediately to the SE of the Härnösand Deep, a large erosional corridor cuts through recent glacial sediments. Its scale can therefore be linked solely to meltwater erosion during the last deglaciation (Fig. 4F). Seismo-acoustic profiles (Lind 2016a, b) indicate that its position is guided by a bedrock step along its eastern flank, but the opposing bank has no such constraint and an erosional corridor has removed both till and underlying lacustrine deposits. The corridor is lined with glaciofluvial sediments, indicating that a larger amplitude valley existed prior to sedimentation; esker fragments and small meltwater channels in the floor of the corridor suggest later use of the corridor by meltwater conduits smaller than bankfull. Several large tributary channels with a braided/chaotic topology feed the corridor from the north. These tributaries are variably hanging, and therefore pre-date a phase of corridor erosion, or are graded to the level of the corridor floor. At the northern end of the corridor a large (~ 5 x 5 km) glaciofluvial deposit is deeply incised (e.g. 30-60 m deep, 300-1000 m wide) by sinuous channels. The corridor additionally both truncates MSGLs and appears to seed lineations off the western/down-ice flow flank. These morphological relationships collectively suggest that this system experienced multiple erosional and depositional events of different magnitudes; we do not rule out bankfull subglacial meltwater flow at some stage.

Grounding line and near-marginal landforms

Moraines and grounding zone wedges (GZWs) are virtually absent from the Bothnian Sea. There is, most notably, no GZW associated with the terminal zone of MSGL flowset A. A small cluster of moraines is found at the northern limit of flowset B, most commonly positioned draping or stacked against the proximal (up-ice) side of the till bodies and drumlins which sit above the post-glacial sediment infill (Fig. 5A). These are difficult to connect into coherent ice-margin positions, but nonetheless point to a zone in which the retreating grounding line is pausing or sticking against its substrate. A second potential zone of stability is identified on the central-upper lateral flank of flowset A, where asymmetric ridges oriented transverse to underlying MSGLs appear to be stacked and draped over the lineations (Fig. 5B). The asymmetric profile and stacked appearance strongly resemble GZWs (e.g. Batchelor & Dowdeswell 2015), albeit an order of magnitude smaller than typically considered (see, however, Halberstadt *et al.* 2016). However, in the same vicinity MSGLs appear to have been 'broken' by active reshaping or fracture, which elsewhere has been used to indicate subglacial 'ribbing' by ice stream sticky spots (Stokes *et al*. 2008).

Besides these two instances, grounding line landforms only become part of the glacial landform assemblage once retreat has progressed onto the present-day land area in the far north (Strömberg 1989; Hättestrand 1998; Bouvier *et al*. 2015). However, a central tract of the offshore area is dominated by a dense network of angular, criss-crossing, narrow (~ 20-50 m wide) and low-amplitude (<1 m) ridges that we interpret as basal crevasse squeeze ridges (Fig. 5D, E). These are exclusively located within the lower ~ 160 km of lineation flowset A (see Fig. 5C), arranged in broad swathes in the lower parts of the flowset and in narrower corridors upstream. Their orientations and distribution reflect a densely crevassed ice body, with fractures along-, across- and oblique to the ice flow direction.

Finally, there are abundant iceberg scours on the Bothnian Sea floor (observed, but not individually mapped). In most areas they take the form of shallow and brief scrapes or pits (e.g. see the drumlin tops in Figs 2A, 4A-B). There are relatively few extended, linear or criss-crossing scours (such as cutting through Fig. 5D), which are typical in continental shelf settings. Only in some distal zones of flowset A are they sufficiently dense or deep that the underlying primary glacial relief is obscured.

A reconstruction of ice flow and retreat dynamics

Based on the distribution of landform assemblages, their cross-cutting relationships and their compositional make-up, we are able to reconstruct aspects of the geometry and behaviour of ice flow during deglaciation of the Bothnian Sea. Given our still limited data coverage, any detailed reconstruction is inevitably fragmentary. However, we recognise three ice flow/retreat stages (Fig. 6): Stage 1) flow from the southern Bothnian Sea to an ice margin south of Åland-Uppland, for example during the Younger Dryas or soon after; Stage 2) an ice stream event in the central-southern Bothnian Sea; and Stage 3) ice confined to the northern coastal zones, with local oscillation in response to loss of the main Bothnian Sea ice body.

Stage 1: post-Younger Dryas margin retreat

This stage comprises bedform flowsets C and D, which depict convergence of flow into the Åland Sea and more divergent flow towards the Uppland coast. The relative chronology of these flow paths is unclear, mirroring conflicting data from the terrestrial realm in which the youngest striae appear to correspond variably to either a more Åland-convergent or an Uppland-directed trend. Based on terrestrial striae and glacial landform assemblages, Kleman *et al.* (1997) treat this sector as a broad, single divergent flowset across the whole southern Bothnian Sea (cf. Punkari 1980, 1994), while several earlier studies (e.g. Hoppe 1961; Strömberg 1981, 1989) reconcile locally conflicting striation directions by recourse to an intricate calving ice margin whose local embayments stimulate flow drawdown in variable directions. Offshore, there is a marked offset between the orientations of two, internally coherent groups of lineations (e.g. around 18°25' E, see Fig. 3). Whether these lineation sets are contemporaneous or mark a sequential development of flow, we suggest this marks the up-ice reach (50-60 km) of a drawdown effect through the Åland Deep. It is likely that as margin retreat progressed from the Younger Dryas position south of the Åland sill and encountered the greater water depths of the Deep, the increase in calving would enhance drawdown and convergence of ice flow into this passage, manifest in the lineations of flowset D.

Stage 2: Bothnian Sea ice stream event

This stage comprises flowsets A, B and subsidiaries (A´, B´, B´´). We interpret the highly elongate and parallel MSGLs of flowset A as the imprint of a palaeo-ice stream. The ice stream trunk extends from an onset zone marked by convergence of flowlines and extension of lineations in the north Bothnian Sea (cf. Greenwood *et al.* 2015), where a lateral transition from drumlins to elongate lineations marks the bounds of a rather narrow ice stream (~ 40 km wide). The distribution of MSGLs indicates the ice stream terminated in the south-central Bothnian Sea; it did not reach the present-day coast. At its distal end, the complex sequence of MSGL cross-cutting and the splayed geometry of the assemblage suggest the ice stream terminus experienced local shifts in outflow close behind the grounding line, while a stable geometry was maintained upstream in the main ice stream trunk.

In its trunk, the ice stream pathway is aligned with flowset B, collectively indicating flow to the SSE. These pathways are, however, separated by both a data coverage gap and by contrasting dynamic regimes, while flowing across comparable terrain and the same bedrock substrate (Ordovician limestone, Figs 3, 7). The drumlinised flowset B is indicative of warm-based ice, but likely not streaming. Its meltwater system exhibits a systematic downstream evolution from meltwater streams up-ice (Fig. 4A), to channel-esker conduits (Fig. 4C), to large eskers with broader 'beads' or fans (Fig. 4E), indicating an increasingly ice-marginal setting favouring deposition of glaciofluvial material. At the head of the flowset, however, up-ice of the small streams, eskers reappear accompanied by moraines (e.g. Fig. 5A), marking a new marginal zone. This arrangement of meltwater and ice-marginal landforms suggests a back-stepping hydrological system (Hebrand & Åmark 1989; Mäkinen 2003), whose close correspondence with underlying drumlins leads us to interpret the whole assemblage as a retreat assemblage (cf. Kleman & Borgström 1996).

Consistency in ice flow direction between flowsets A and B would suggest, in the simplest interpretation, that these two assemblages belong to the same phase of ice flow/retreat. However, they represent different glaciodynamic environments and the toe of flowset A bends to cross-cut that of B (Fig. 3D). One reconstruction scenario therefore encompasses *i)* a phase of drumlinisation accompanying ice margin retreat into the southern Bothnian Sea, *ii)* margin retreat proceeding across flowset B, followed by *iii)* a readvancing pulse of a newly triggered ice stream (flowset A). This ice stream pulse would erase (overprint) the former toe of flowset B, while preserving the upper drumlin-meltwater assemblage. An alternative scenario does not invoke significant margin retreat and readvance. Rather, the two flowsets record broadly simultaneous flow, with streaming and non-streaming regimes separated by a lateral shear margin located within the data coverage gap. The distal overprinting could be explained by temporary lateral encroachment of fast flow in this zone (e.g. Stearns *et al.* 2005; Catania *et al.* 2012). This is consistent with abundant crevasse squeeze ridges in this zone that are sub-parallel to ice flow, indicating strong lateral tensile stress (extension). With such weak lateral constraint within the ice body, the dominant flow path may have locally shifted. Lineation cross-cutting within the toe of flowset A points to such a shifting flow direction close to the margin.

This Bothnian Sea ice stream event underwent rapid retreat. The first ice margin position encountered in flowset B lies ~ 80 km up-ice, indicating a large magnitude step back in margin position, while crevasse squeeze ridges dominate the lower 160 km of the ice stream trunk. These point to a highly fractured ice body, which was yet capable of preserving a delicate imprint of <1 m amplitude landforms. We infer a rapid break-up of the lower and central portions of the Bothnian Sea ice stream, over an area on the order of 10-20 000 km^2. Extended and criss-crossing iceberg scours in the toe of the ice stream pathway suggest that a large volume of icebergs was released via full-depth calving in this zone. The grounding line likely re-stabilised on the downstream flanks of the Härnösand Deep, where a suite of stacked, transverse wedges are located (Fig. 5B). Tentatively connecting this margin position with that in upper flowset B (Fig. 5A) leads us to reconstruct a post-ice stream, deep calving embayment in the central Bothnian Sea (Fig. 6).

Stage 3: local margin oscillation following collapse of central Bothnian Sea ice stream

This stage comprises flowset F. Small drumlins and lineations encroach (E)SE-ward over the ice stream pathway in its upper sector, though limited to the western flank of flowset A. This configuration must postdate the ice stream and the rapid retreat which ensues. We suggest that this is a manifestation of a deep calving bay opening in the northern Bothnian Sea, and/or the large-scale and ongoing ice divide shift from the Bothnian Bay westwards towards the Scandinavian mountains (Kleman *et al*. 1997; Lundqvist 2007). In either case, it is likely that the loss of ice from the central Bothnian Sea triggered local ice margin oscillations (overprinting) in response to this loss of offshore buttressing (cf. De Angelis & Skvarca 2003; Rignot *et al.* 2004).

Misfit flow geometry?: flowset E.

This assemblage of weakly streamlined ribbed moraine indicates SW flow towards the Swedish coast as far north as Söderhamn. Terrestrial stratigraphic and striation evidence of shore-ward encroaching ice has long been known from this south Norrland coast, and has typically been associated with the final deglacial events (e.g. Sandegren 1929; De Geer 1940; Strömberg 1981, 1989; Lundqvist 2007; see below). However, the flow direction is entirely inconsistent with the drumlinisation and ice streaming of Stage 2, and since we interpret rapid margin retreat and break-up of the central Bothnian Sea ice body to have followed, it is difficult to relate this SW-ward flow off Söderhamn to the final retreat stage. This flow direction is more consistent with Stage 1 (e.g. flowset C), in which ice reaching southern coasts of the basin diverges over land (cf. Punkari 1980, 1994; Kleman *et al.* 1997), implying somewhat radial outflow from the north-eastern Bothnian Sea. Alternatively, these landforms belong to an older phase of ice flow.

Discussion

Regional framework for a Bothnian Sea ice stream

The glacial geomorphological assemblages revealed by presently available multibeam coverage provide important fragments of information on the palaeo-flow dynamics and ice sheet retreat history of the Bothnian Sea. The most striking aspect of the offshore retreat record is the operation of an ice stream, confined to the central parts of the Bothnian Sea. This indicates that the ice stream was triggered during the retreat sequence, and does not appear to be active earlier in deglaciation.

The Finnish deglacial pattern and conventional ice margin reconstructions, widely based on recessional moraines, eskers and the long-standing clay varve chronologies, suggest SE-ward encroachment of a 'Baltic Sea lobe' across southern Finland, and project a NE-SW oriented ice margin retreating over the Bothnian Sea (e.g. Aartolahti 1972; Punkari 1980, 1994; Boulton *et al.* 2001, 2009; Strömberg 2005; Stroeven *et al.* 2016). The Bothnian Sea retreat sequence that we identify here is not entirely consistent with this conventional view. In the southernmost realm, flow is funnelled S/SSE towards Åland Sea while the SW Finland terrestrial pattern records SE or even ESE flow (Punkari 1994; see Fig. 1). The Bothnian Sea ice stream possesses a SSE/SSW-ward flow direction and has limited extent both laterally and in length, indicating that the Bothnian Sea did not host a basin-wide stream or surge event (e.g. Kleman & Applegate

2014), at least during its final deglaciation. Rather, any large ice lobe that may have existed in the Bothnian Sea was dissected by spatially and temporally variable flow behaviour (cf. 'sub-lobes', Ahokangas & Mäkinen 2014). While the early retreat geometry of the Bothnian Sea may have encompassed a broadly divergent ice sheet sector, the ice stream event recorded by flowset A post-dates the 'Baltic Sea lobe' configuration (*sensu* Punkari 1980, 1994; Boulton *et al.* 2001, 2009), and mobilised only a local corridor within the Bothnian Sea, rather than the whole basin.

There has long been a discussion regarding a late-deglacial Bothnian Sea readvance onto the Swedish east coast: the so-called Gävle oscillation. SW-ward striae and drumlins widely indicate land-ward motion of ice along the stretch of coast from Uppland to Hudiksvall, in contrast to inland of this coastal zone where the terrestrial retreat pattern is clearly to the NW (De Geer 1940; Hoppe 1961; Strömberg 1989; Lundqvist 2007). Sandegren (1929) interpreted till layers within the glacial lake varve sequence as evidence for a readvance of grounded, Bothnian Sea ice. However, these units have been reinterpreted as products of iceberg re-grounding and melt-out distal to the grounding line, and a number of authors reconcile the observations around Gävle with an intricate ice margin, punctuated by calving bays and local flow direction offsets (De Geer 1940; Hoppe 1961; Strömberg 1981, 1989). Nonetheless, along the Västerbotten coast and across Norra Kvarken, striae indicate shore-parallel (SW) flow, perpendicular to the De Geer moraine-demarcated retreat direction here (Lundqvist 2007). On this basis, Lundqvist (2007) reconstructs a large-scale retreat-readvance sequence in the Bothnian Sea, with a surge drawing down flow from Norra Kvarken towards the SW, and terminating just offshore from the Uppland – southern Norrland coast.

We show here that, while Bothnian Sea ice did form a substantial flow path onto the Uppland and southern Norrland coast, and while the Bothnian Sea did experience a deglacial streaming event, the two are not likely to be linked. Bothnian Sea-sourced ice flowed over Uppland as deglaciation of the north Baltic and southern Bothnian Sea began (Stage 1; Fig. 6), possibly as part of a wider, heterogeneous 'Baltic Sea lobe' (as above). The central Bothnian Sea ice stream, triggered during retreat, drew down ice from Norra Kvarken and likely accounts for the aligned flow traces observed here by Lundqvist (2007). A notable implication of this flow geometry, in the early Holocene, is that an ice divide branch over the Bothnian Bay must have either persisted throughout deglaciation or have been renewed by this stream event. However, since the ice stream event was confined to the offshore sector, it must be distinct from the SW-flow over Uppland/Gävle; we find no reason to invoke a margin readvance to account for these terrestrial traces, nor do we find sufficient reason to invoke complete retreat of Bothnian Sea ice prior to a readvance (e.g. Lundqvist 2007).

While a fit with the terrestrial-based ice flow and retreat pattern remains somewhat uncertain, the existing varve-chronology for deglaciation places rather tight constraint on Bothnian Sea retreat and on the Bothnian Sea ice stream event. Evacuation of ice from the Uppsala – Åland region to the Västerbotten coast (i.e. Fig. 6) must be accomplished within ~ 600-700 years (Stroeven *et al.* 2016, after Strömberg 1989, 2005), which implies an average retreat rate of ~ 600-650 ma^{-1}. Strömberg (1989, 2005) does not identify any readvances within this time frame. From numerous sites across Uppland, however, varved sequences contain a ~ 150 year period with a high content of limestone clasts (Strömberg 1989). This so-called 'spot-zone' is taken to correspond to increased iceberg transport of debris entrained over the plateau of Ordovician limestone in the central-western Bothnian Sea (Fig. 7). Strömberg (1989) interprets the interval as an abrupt switch to intense calving of the retreating ice margin, which Kleman & Applegate (2014) suggest is indicative of large-scale collapse of the Bothnian Sea sector. It seems likely that the spot-zone corresponds, in some manner, to the behaviour of the Bothnian Sea ice stream that we have reconstructed here. We can envisage three possible explanations: *i)* the spot-zone records the sudden activation and duration of the ice stream, whose path crossed the limestone plateau and which terminated in a calving margin approximately 60 km from the Uppland coast; *ii)* it records only the break-up of the highly crevassed ice stream trunk; or *iii)* it encapsulates both of these events. If the spot-zone records only the break-up and release of a huge iceberg volume, we question why there would be no distinct signature of the ice stream itself in the proglacial sediment record, prior to collapse; conversely, one might expect a distinct peak in limestone rafting at the point of collapse. The spot-zone contains the highest concentration of clasts in the first 90-110 years, after which follows a 40-50 year gradual decrease in limestone content. It is possible, therefore, that the first interval represents the phase of ice streaming, with high entrainment and delivery of limestone clasts to the calving margin. It culminated in ice stream collapse after ~ 100 years, with mass release of icebergs and gradual melt-out of their debris content thereafter.

On Strömberg's (1989) revised De Geer timescale, the spot-zone occurs from varves -580 to -424 (9818-9662 varve years BP (AD 1950); 10.67-10.514 cal. ka BP based on connection to ice core chronology GICC05 by Stroeven *et al.* 2015). During this time, the ice margin over present-day land retreated from south of Söderhamn to Sundsvall (Strömberg 1989), consistent with the latitude of the Bothnian Sea ice stream and its likely retreat geometry. The Bothnian Sea ice stream event, however, has no corollary in the Swedish terrestrial retreat pattern, where varve-based reconstructions show steady retreat (Strömberg 1989). To the east, the Central Finnish Ice Marginal Formation (CFIMF) has been interpreted as the terminus of an early Holocene readvance (Rainio 1986) though its timing and dynamics (e.g. uniform lobe or a heterogeneous sector) are not clear. Saarnisto & Saarinen's (2001) chronology places the Formation at c. 11.1 ka BP, while Strömberg's (2005) correlation of the Finnish varve chronology with the Swedish Time Scale places it at or after -800 varve years: 10.89 cal. ka BP after Stroeven *et al.* (2015). If we take the limestone spot-zone to date the activation and the collapse of the Bothnian Sea ice stream, then the CFIMF would appear to pre-date this event. This is consistent with Lunkka & Gibbard (1996), who find that Bothnian Sea ice responsible for the Pohjankangas ice-marginal formation west of the CFIMF post-dates the CFIMF itself, though the requisite ice flow direction (E or ESE) is not easy to reconcile with the trajectory of the Bothnian Sea ice stream. Collectively, these asynchronous deposits and ice flow trajectories suggest that a succession of ice stream, surge or readvance events and margin oscillations occurred across the south-central sector of the FIS throughout its deglaciation (Fig. 8).

Implications for forcing mechanisms of ice sheet retreat behaviour

Efforts to examine the behaviour and stability of marine-based ice sheet sectors have typically been confined to a particular ice dynamic setting, namely topographically-funnelled ice streams that drain deep interior basins and/or occupied continental shelf troughs during glacial maxima. In such settings, basal topography (e.g. direction of slope, trough width, relative relief) is known to be an important control (e.g. Thomas & Bentley 1978; Schoof 2007; Jamieson *et al.* 2012; Gudmundsson *et al.* 2012), while bed geology (e.g. crystalline or sedimentary substrates and structural elements) has been widely invoked as a control on ice stream evolution (e.g. Ó Cofaigh *et al.* 2002; Wellner *et al.* 2006; Graham *et al.* 2009). The Bothnian Sea represents a contrasting setting to such prior investigations: a broad basin, underlain by a variable geology and lacking a mature trough (Figs. 1, 7).

Aspects of the landform record indicate some correspondence with both the substrate topography and geology. Ice flow is drawn down through the Åland Deep (flowset D) and in this region the crystalline substrate (Fig. 7) and its fractured bedrock topography have a clear impact on lineation form (e.g. small crag and tails) and meltwater drainage routes. In contrast, the sedimentary substrate of the central Bothnian Sea supports a thicker sediment (till) cover and well-developed bedforms and meltwater assemblages. While much of this meltwater record (e.g. pathways, type or connectivity of landforms) exhibits little correspondence with the substrate, the Härnösand erosional corridor (Fig. 4F) and the channel-esker corridor through flowset A (Fig. 2B) are both flanked on one side by a bedrock step (albeit under a thick till cover). In these cases the underlying bedrock topography may control the position of major corridors of meltwater drainage, though not their form.

We find no significant regional substrate explanation for the existence or the position of the Bothnian Sea ice stream. Ice is not funnelled into the eastern trough of the Bothnian Basin;

rather, the ice stream bypasses the head of the trough and flows onto the higher ground of Eystrasaltbanken, with no perturbation to either the direction or the morphology of its component MSGLs. This absence of topographic control on ice flow direction is surprising given the late stage and, one would expect, low surface profile associated with the event. This implies significant driving flow from the source region over the Bothnian Bay. Substrate geology similarly displays no clear relationship with the ice stream position. While a downstream elongation of MSGLs broadly corresponds to the transition from Proterozoic clastic sedimentary bedrock to Ordovician limestone, this limestone plateau underlies much of the central-western Bothnian Sea and supports a range of bedform (e.g. flowsets A, B and E) and meltwater landform assemblages. At the distal end of the ice stream, subdued MSGLs continue southward of the limestone-sandstone boundary. The substrate therefore does not dictate the landform type or the style of ice flow; rather, the contrasting landform types reflect spatially variable glaciological regimes. It appears that, in this case, the basal conditions exert little control on the primary location or operation of the ice stream.

The potential timing of the Bothnian Sea ice stream event is several hundred years after the rapid warming that followed the Pre-Boreal Oscillation (11.47-11.35 ka: Stroeven *et al.* 2015; Fig. 8), and the possibility of staggered southern FIS fast flow or readvance events would argue against a direct climate trigger. We do, however, invoke Bothnian Sea retreat under high surface melting conditions. The abundance of meltwater landforms in the Bothnian Sea, and their connectivity into coherent drainage systems, suggests that subglacial meltwater volumes were likely considerable under both streaming and non-streaming flow conditions. Basal melt rates are typically calculated to be on the order of mm to 10s cm a^{-1} (Fahnestock *et al.* 2001; Joughin *et al.* 2003, 2009). Based on such rates (e.g. 20 mm – 20 cm), and the typical meltwater channel sizes observed in the Bothnian Sea (e.g. Table 1), a single conduit could drain an entire year's basal melt (from the ~ 80 000 km^2 basin) over a timescale of hours to weeks. To account for the meltwater-rich subglacial geomorphology, we appeal to abundant surface melting (on the order of m a^{-1}) and delivery of meltwater to the bed during these retreat stages. This is consistent with independent interpretations of atmospherically driven ice sheet retreat at this time (Andrén *et al.* 2002; Cuzzone *et al.* 2016), and comparable with early Holocene retreat of the southern Laurentide Ice Sheet margins where surface melting was on the order of 5-9 m a^{-1} (Carlson *et al.* 2009) and when esker abundance is highest (Storrar *et al.* 2014).

The precise role of increased subglacial meltwater drainage is, however, equivocal. While increased basal water may be expected to enhance flow or even trigger ice streaming (Alley *et al.* 1986; Zwally *et al.* 2002; Bell *et al.* 2007; Stearns *et al.* 2008), contemporary observations (Sundal *et al.* 2011; Sole *et al.* 2013) and palaeo-interpretations (Jørgensen & Piotrowski 2003; Storrar *et al.* 2014) suggest that the enhanced efficiency of meltwater drainage arising from channelisation may ultimately reduce basal water pressures, such that ice flow instabilities cease. In the Bothnian Sea, a few instances within the ice stream landsystem of meltwater streams running alongside MSGLs (e.g. Fig. 4B), and repeated drainage events through the Härnösand erosional corridor that both pre- and post-date MSGL formation, point to the presence of abundant basal meltwater during ice stream operation. Indeed, the Härnösand Deep offers potential for considerable meltwater storage and episodic release of meltwater, independent of any climate forcing. The episodic nature of high magnitude flow here suggests that the Deep may have acted as a subglacial lake, periodically draining stored meltwater from a basin with a potential maximum volume on the order of ~ 15-65 km^3 (based on a horizontal upper surface and minimum/maximum drainage thresholds into the erosional corridor). Such outburst events in Antarctica have been observed to locally enhance ice flow velocities (Bell *et al.* 2007; Stearns *et al.* 2008), and it is possible that here fast ice flow velocities were similarly influenced. Nonetheless, the great majority of channelised meltwater landforms across the Bothnian Sea clearly drape or incise bedforms, both within and beyond the ice stream pathway (Fig. 2B, 4C, 4E), suggesting that widespread channelisation of meltwater drainage immediately post-dates the cessation of bedform shaping. Channelisation of meltwater may therefore be part of a suite of processes that shutdown ice stream operation (Jørgensen & Piotrowski 2003; Sole *et al.* 2013; Storrar *et al.* 2014).

Finally, crevasse squeeze ridges are a dominant component of the lower ice stream landform assemblage. Their preservation implies rapid retreat without regrounding of the ice body. Crevasse squeeze ridges are intermixed with iceberg pits and scours which, in the most distal portion of the ice stream trunk, have a linear geometry and indicate the drift of deep-draught icebergs calved from close to the grounding line. We propose that the Bothnian Sea ice stream comprised, in its final stage of operation, a highly fractured ice body that underwent mass melting. We invoke rapid grounding line retreat driven by large-scale hydrofracture and the mass release of icebergs.

Conclusions

High-resolution multibeam data reveal, for the first time, the direct imprint of the ice sheet retreat sequence in the Bothnian Sea. A rich glacial landform record comprises subglacial bedforms, abundant meltwater products and widespread evidence of basal crevassing. These data lead us to conclude:

- A Bothnian Sea ice stream was activated during ongoing retreat of ice through the marine basin after the Younger Dryas. It mobilised a rather narrow corridor of fast flow within the broader basin, and extended into the south-central Bothnian Sea but did not reach onto present-day land.
- The ice stream event underwent high extension, producing a highly crevassed ice body which likely precipitated the demise of the ice stream via mass iceberg calving.
- The ice stream pathway lies across a topographic high rather than occupying the shallow trough along the eastern flank of the basin, indicating little topographic forcing of flow. Rather, a rich meltwater landform record throughout all areas where we have multibeam data coverage points to a significant supply of surface meltwater to the ice sheet bed, and atmospherically-driven retreat of this marine ice sheet sector.
- The ice stream event likely accounts for the so-called 'spot zone' of limestone ice-rafted

debris previously observed in sediment records across Uppland and southern Norrland. This would constrain the duration of ice stream flow and collapse to approximately 150 years, and its timing to c. 10.6 ka BP.

Our reconstruction fills an important gap in understanding of the retreat behaviour of the south-central sector of the Fennoscandian Ice Sheet. It suggests that the Bothnian Sea sector rapidly pulsed and then collapsed, possibly the culmination of a succession of stream, surge or readvance events along the southern ice sheet margin. Our observations point towards a strong surface (atmospheric) control on the final retreat of ice in the Bothnian Sea. There is a clear need to better link the offshore record with the terrestrial-based retreat pattern and chronology. Indeed, there is an opportunity to improve chronological correlations across the basin using the marine landform information as a correlation tool and as a guide for future strategic coring of new, offshore chronological sequences. The annually-resolved regional chronology makes the Bothnian Sea an unrivalled catchment in which to more widely inform questions of the stability of and controls upon the behaviour of marine ice sheet sectors.

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Figure Captions

Table 1: Typical morphometric properties of subglacial meltwater landforms (individual conduit segments).

Figure 1: The Bothnian and Baltic basins. ~ 85% of the offshore terrain is shallower than 100 m. Area of multibeam coverage is outlined in dark blue; Ordovician limestone plateau in dashed green. Locations referred to in the text: Su = Sundsvall; H = Hudiksvall; Sö = Söderhamn; G = Gävle; Ör = Öregrund-Gräsö; Å = Åland; U = Uppsala; E = Eystrasaltbanken. Ice sheet retreat isochrons from Hughes *et al.* (2016) (their “most credible” positions). Finnish ice lobe flowlines after Punkari (1994), Johansson *et al.* (2011). Inset: the Bothnian and Baltic basins lie in the heart of the Last Glacial Maximum Eurasian ice sheet complex (21 ka extent from Hughes *et al.* 2016).

Figure 2: Examples of subglacial bedforms. **A.** Drumlins, often clustered on larger till bodies, are here oriented SSE. Interpreted acoustic stratigraphy across profile X-X′ shows that drumlins are anchored on bedrock highs, while post-glacial clays infill inter-drumlin lows. IB marks iceberg pits and scrapes across drumlin tops. **B.** MSGLs 15-20 km in length are here oriented SSE. They are buried in the SE corner of this area by post-glacial infill. C-E marks a braiding channel - esker corridor which cuts obliquely across the MSGL population. **C.** Broad, curvilinear ridges in thick till are interpreted as ribbed moraine, facing SW.

Figure 3: **A.** Mapped subglacial lineations are shown on a red-green colour scale according to their length. Ribbed moraines are shown in white. Subglacial bedforms are grouped into distinct flowsets labelled A-F (thick white/grey/black lines; shades used only to clarify separate sets), on the basis of local conformity of orientation and morphology, and superimposition relationships. Discs mark the order of superimposition where lineation cross-cutting is observed; the line through the disc denotes uppermost (youngest) flow direction. Location labels as in Fig. 1. **B-D.** Examples of cross-cutting lineations that inform the relative chronology of flowsets. **E.** The terminal zone of flowset A, in which lineations splay and cross-cut (cross-cutting not visible at this scale). MSGLs are buried by post-glacial basin infill, beyond which flowset D marks a distinct change in morphology and orientation; the dissipation of MSGLs is interpreted as the terminus of flowset A, and the palaeo-ice stream.

Figure 4: Examples of meltwater landforms. **A.** Small-scale “streams” are chaotically arranged and poorly connected. They appear and disappear in the lows between drumlins and till bodies, often in areas infilled by post-glacial sediments. **B.** Streams occasionally follow (and define?) MSGLs within flowset A, suggesting the meltwater stream and the lineation formed contemporaneously. **C.** Connected channel-esker conduit incises and drapes drumlins in a sinuous path. This conduit is downstream of the meltwater streams in panel A. IB marks iceberg pits and scrapes across drumlin tops in A and C. **D.** Distribution of meltwater channels (blue) and eskers (green) relative to the bedform population in grey; locations of other figure panels indicated (A-G). **E.** Eskers drape underlying drumlins and swell

downstream into broader deposits. **F.** Meltwater erosional corridor, ~ 4 km wide, cuts SE-ward from the Härnösand Deep in the NW. **G.** Meltwater corridors, arrowed, exploit large-scale bedrock fracture pattern. Local overdeepenings may have provided pockets of meltwater storage.

Figure 5: Examples of near-marginal landforms. **A.** Moraines oriented WSW-ENE drape the up-ice flanks of drumlins. **B.** Asymmetric wedges approximately 1-2 m high and 100 m wide drape underlying MSGLs. **C.** Distribution of ice-marginal landforms (purple) and crevasse-squeeze ridges (pink), relative to other landforms in grey tones; locations of other figure panels indicated (A-E). **D** and **E.** Angular and irregular, low-amplitude (~ 1 m) ridges interpreted as basal crevasse squeeze ridges criss-cross each other in dense networks, overlying subdued MSGLs (oriented SSW in D, SSE in E). IB marks large, linear iceberg scour cutting through all other landforms in D.

Figure 6: Reconstruction of ice retreat in the Bothnian Sea. Stage 1 groups spatially variable, local flow structures a few tens of km behind an ice margin retreating from the Younger Dryas position, including landward flow onto the Swedish coast from the SW Bothnian Sea. During Stage 2, the Bothnian Sea ice stream activates. It undergoes lateral migration in its most distal parts, before collapsing into a large calving embayment (pale blue, dotted). The ice margin experiences a local oscillation following collapse of the main offshore sector in Stage 3, before final retreat. Terrestrial varve-based margin isochrons, in ka BP, are from Stroeven *et al.* (2016). CFIMF = Central Finnish Ice Marginal Formation. Location labels as in Fig. 1.

Figure 7: **A.** Bedrock geology of the Bothnian Sea, summarised from Geological Survey of Sweden 1:1million marine bedrock database. Landform mapping overlaid: lineations (white), ribbed moraine (grey), eskers (green), meltwater channels (blue; channels using crystalline structures in dark blue), moraines (purple) and crevasse squeeze ridges (pink). Location labels as in Fig. 1. **B.** Sediment thickness (Lind 2016a, b; Nyberg 2016a, b).

Figure 8: Retreat events of the south-central sector of the FIS. General trend (dotted) and chronology of ice retreat taken from Stroeven *et al.* (2016). We superimpose and approximately scale ice margin oscillations and dynamic events. BIL = Baltic Ice Lake; MSEMZ = Middle Swedish End Moraine Zone; PBO = Pre-Boreal Oscillation.

Table 1: Typical morphometric properties of subglacial meltwater landforms (individual conduit segments)

Form	Width (m)	Depth (m)	Length (km)
Streams	~ 30 – 80	~ 0.4 - 3.5	~ 0.5 - 6
Linked esker-channel conduits	~ 60 – 250	~ 4 – 10	~ 1.5 - 20
Erosional corridors	~ 600 – 3800	~ 8 – 80	~ 20 - 40

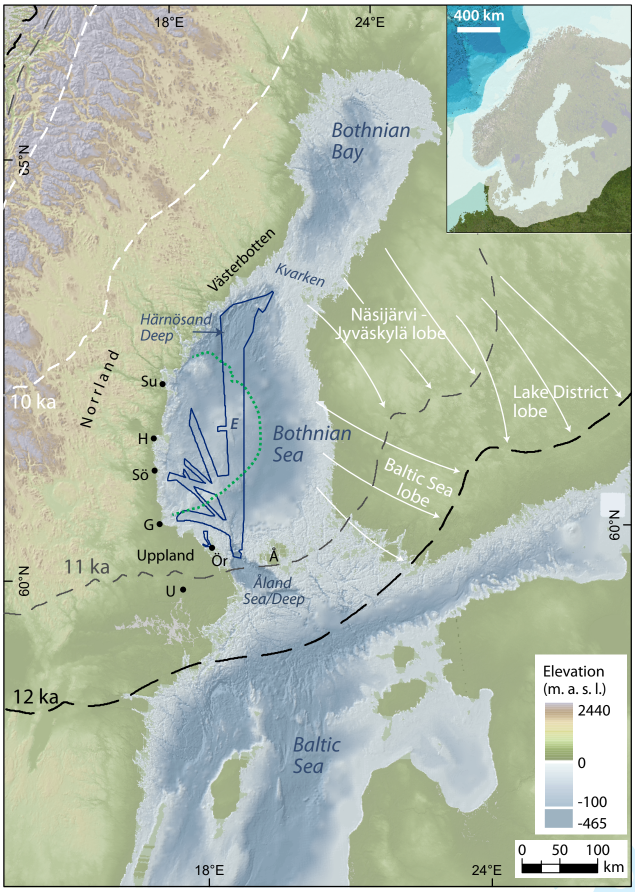

400 km
18°E
24°E
65°N
Bothnian Bay
Västerbotten
Kvarken
Näsijärvi - Jyväskylä lobe
Härnösand Deep
Norrland
10 ka
Su
H
Sö
G
E
Bothnian Sea
Lake District lobe
Baltic Sea lobe
Uppland
Ör
Å
11 ka
U
Åland Sea/Deep
60°N
60°N
12 ka
Baltic Sea
Elevation (m. a. s. l.)
2440
0
-100
-465
0 50 100 km
18°E
24°E

Figure 1

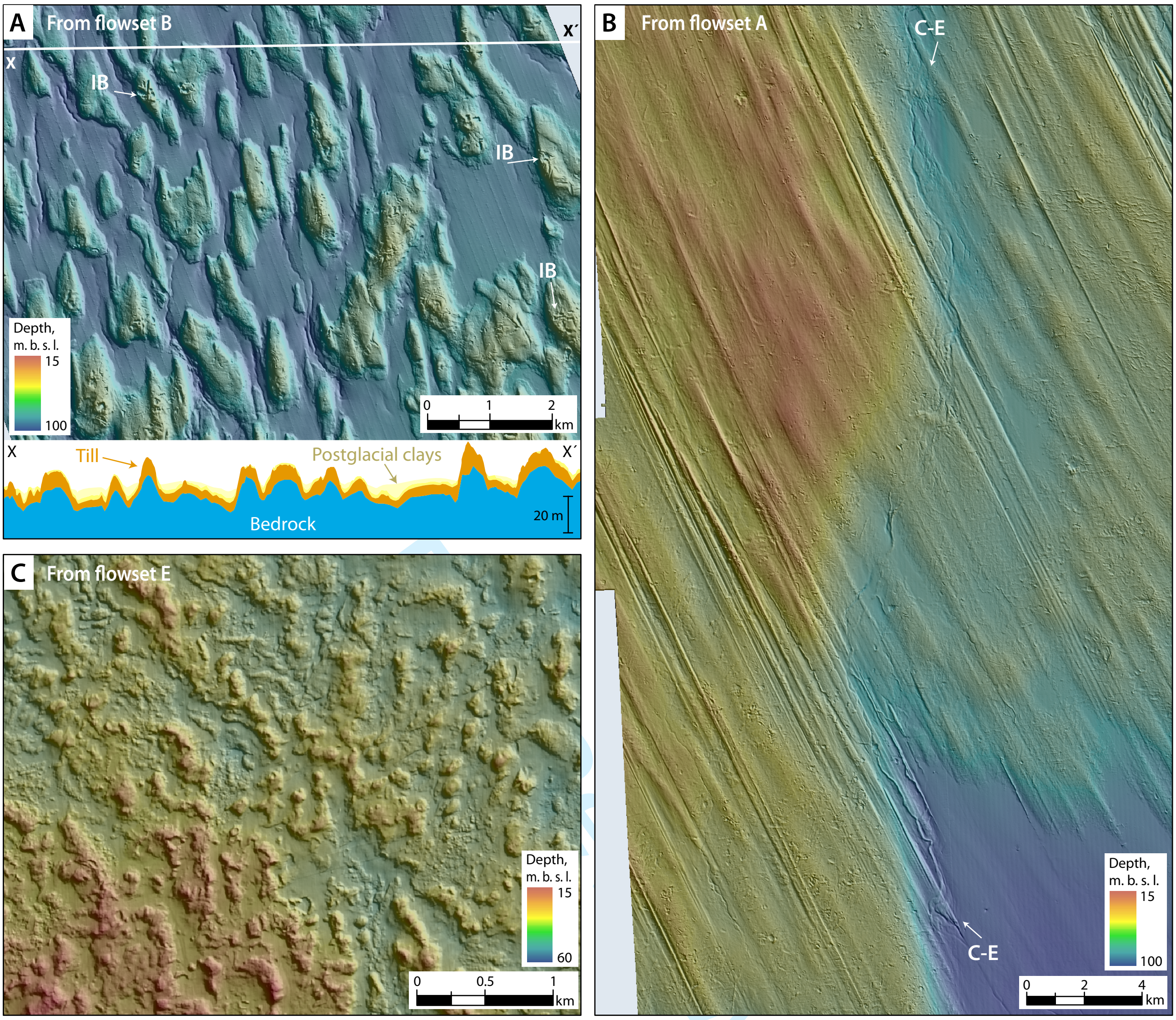

A
From flowset B
X
X´
IB
IB
IB
Depth, m. b. s. l.
15
100
0
1
2
km
X
X´
Till
Postglacial clays
Bedrock
20 m
B
From flowset A
C-E
C-E
Depth, m. b. s. l.
15
100
0
2
4
km
C
From flowset E
Depth, m. b. s. l.
15
60
0
0.5
1
km

Figure 2

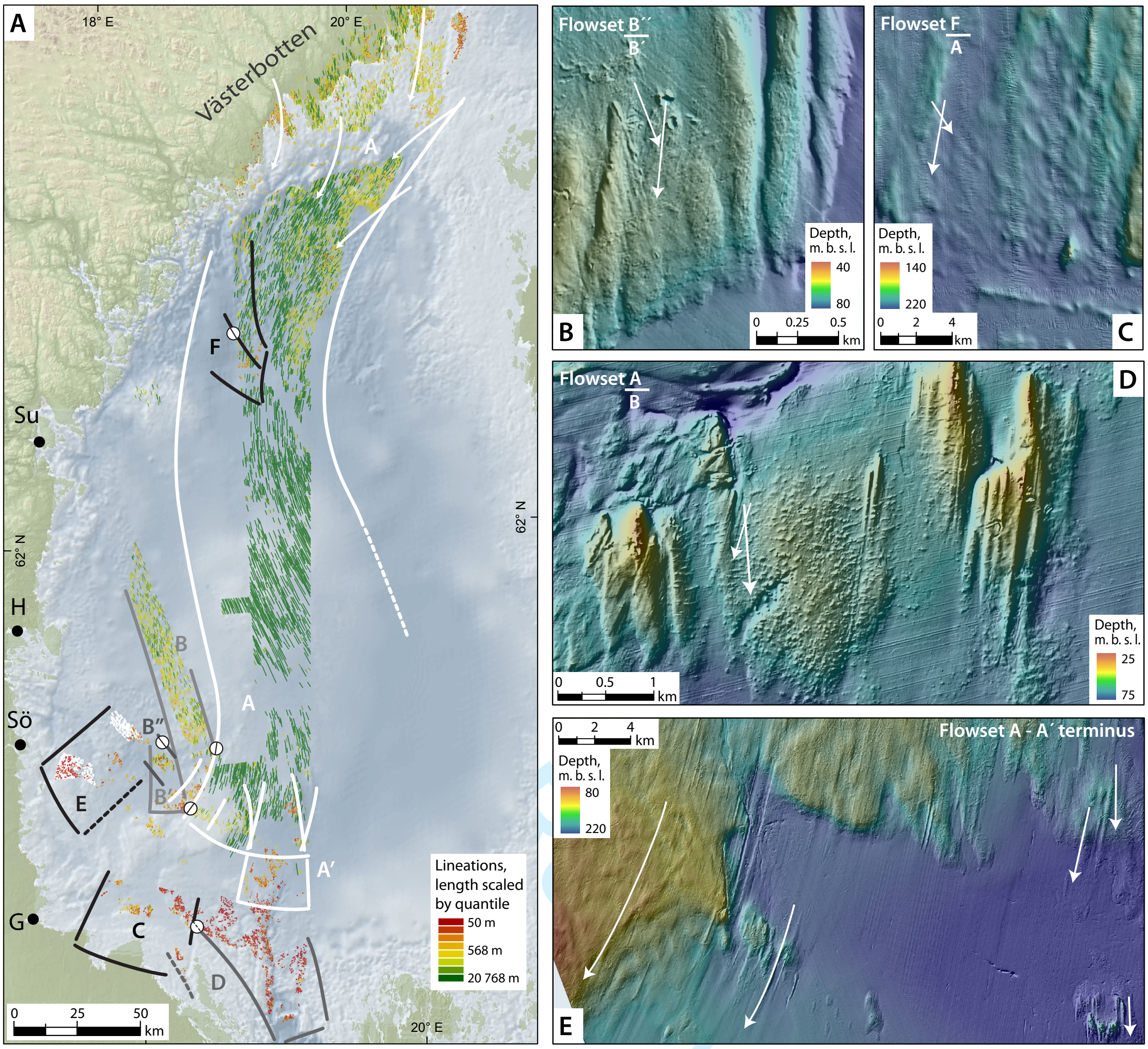

A
Västerbotten
18° E
20° E
62° N
Su
H
Sö
G
F
A
B
B″
B′
E
C
D
A′
Lineations, length scaled by quantile
50 m
568 m
20 768 m
0 25 50 km
Flowset B″/B′
Depth, m. b. s. l.
40
80
0 0.25 0.5 km
B
Flowset F/A
Depth, m. b. s. l.
140
220
0 2 4 km
C
Flowset A/B
D
Depth, m. b. s. l.
25
75
0 0.5 1 km
Flowset A - A′ terminus
0 2 4 km
Depth, m. b. s. l.
80
220
E

Figure 3

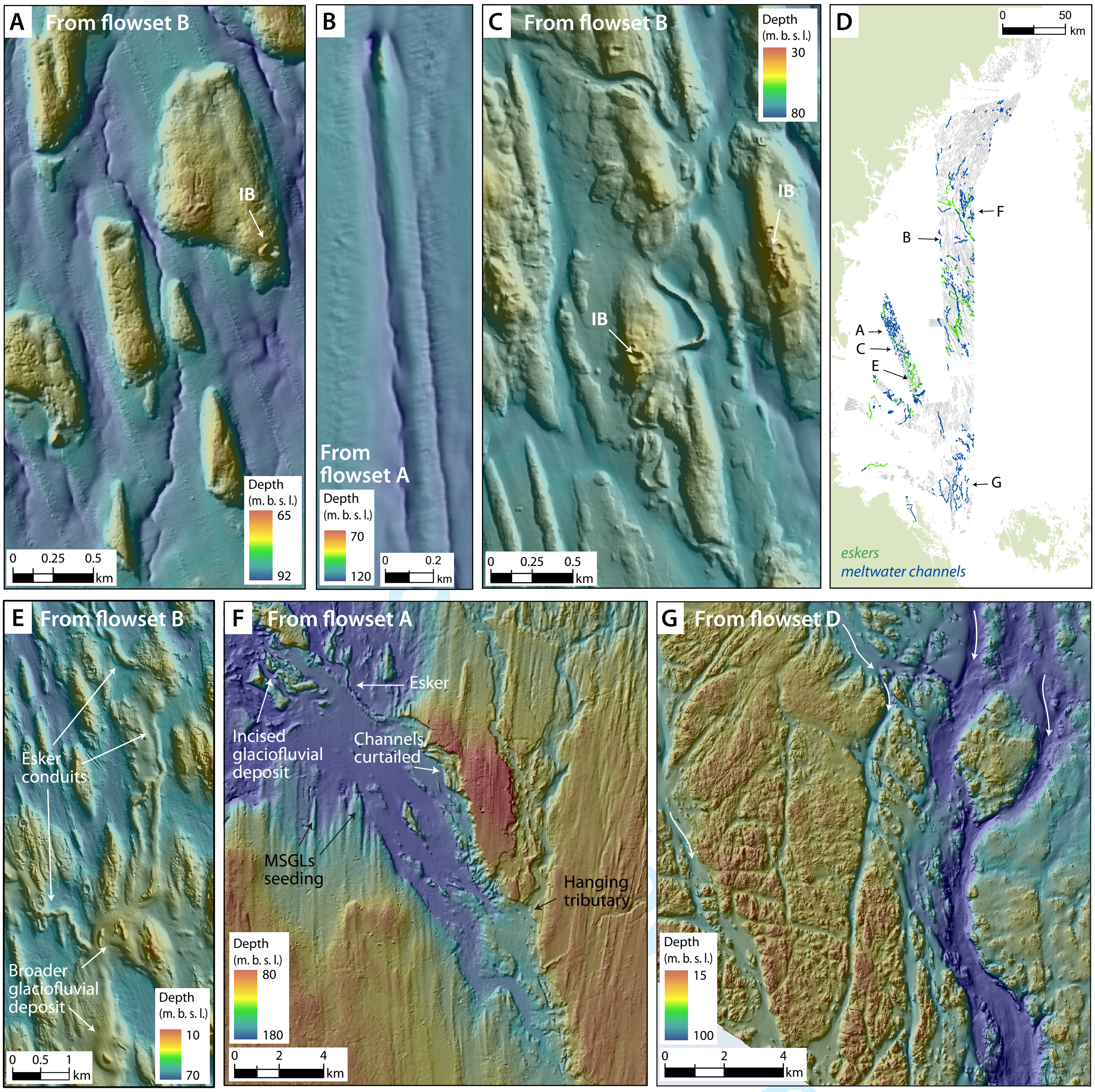

A From flowset B
IB
Depth (m. b. s. l.) 65 92
0 0.25 0.5 km
B
From flowset A
Depth (m. b. s. l.) 70 120
0 0.2 km
C From flowset B
Depth (m. b. s. l.) 30 80
IB
IB
0 0.25 0.5 km
D
0 50 km
F
B
A
C
E
G
eskers
meltwater channels
E From flowset B
Esker conduits
Broader glaciofluvial deposit
Depth (m. b. s. l.) 10 70
0 0.5 1 km
F From flowset A
Esker
Incised glaciofluvial deposit
Channels curtailed
MSGLs seeding
Hanging tributary
Depth (m. b. s. l.) 80 180
0 2 4 km
G From flowset D
Depth (m. b. s. l.) 15 100
0 2 4 km

Figure 4

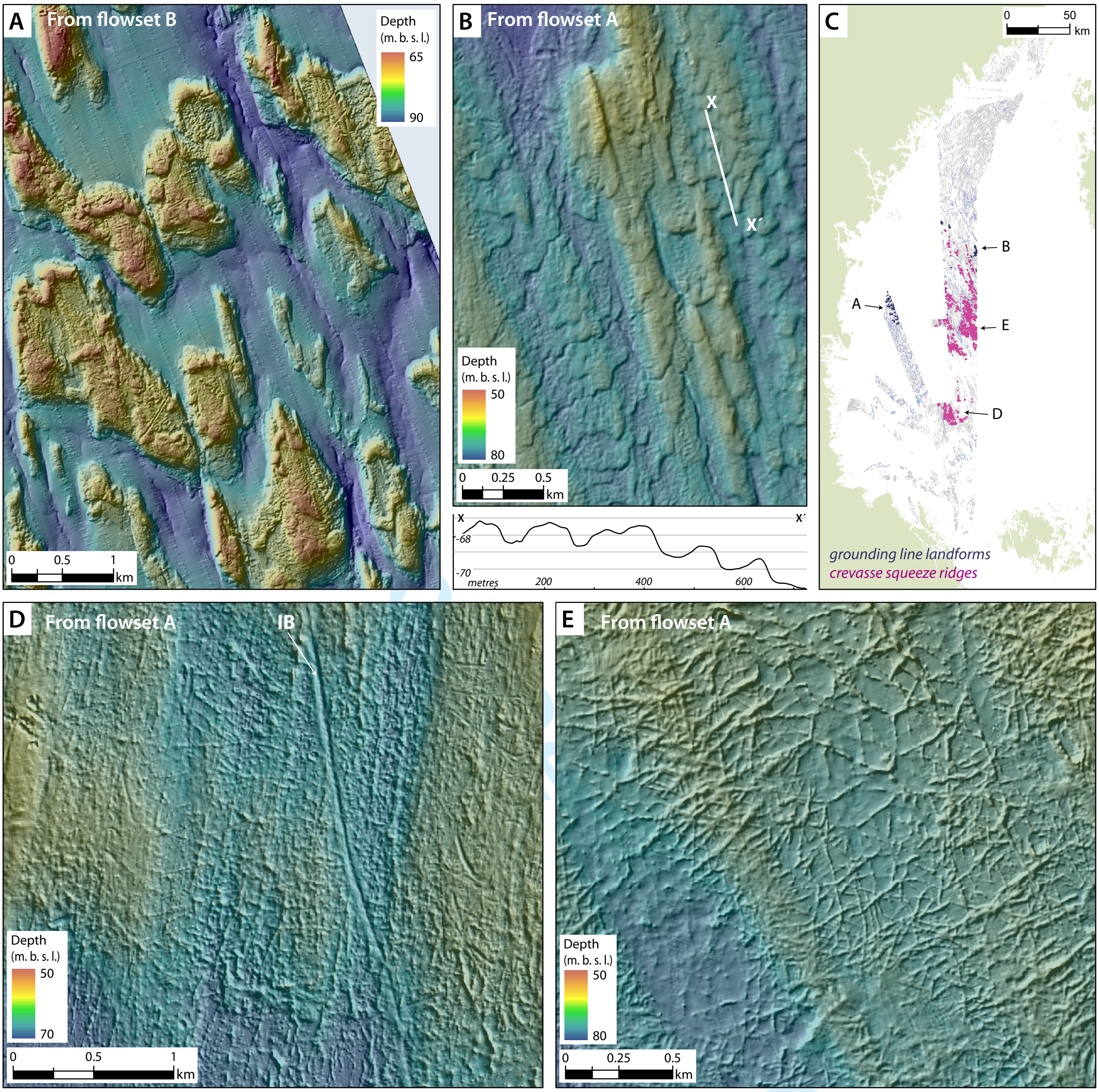
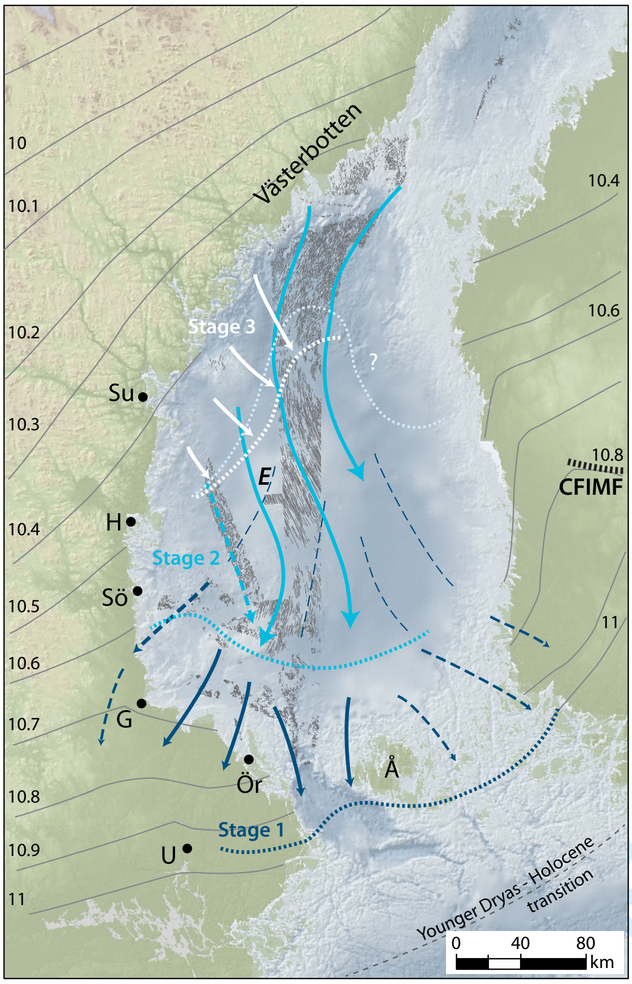

A From flowset B
Depth (m. b. s. l.)
65
90
0 0.5 1 km
B From flowset A
X
X´
Depth (m. b. s. l.)
50
80
0 0.25 0.5 km
-68
-70
metres 200 400 600
C
0 50 km
B
A
E
D
grounding line landforms
crevasse squeeze ridges
D From flowset A
IB
Depth (m. b. s. l.)
50
70
0 0.5 1 km
E From flowset A
Depth (m. b. s. l.)
50
80
0 0.25 0.5 km

Figure 5



Västerbotten
Stage 3
Stage 2
Stage 1
Su
H
Sö
G
Ör
U
Å
E
?
CFIMF
10
10.1
10.2
10.3
10.4
10.5
10.6
10.7
10.8
10.9
11
10.4
10.6
10.8
11
Younger Dryas - Holocene transition
0
40
80
km

Figure 6

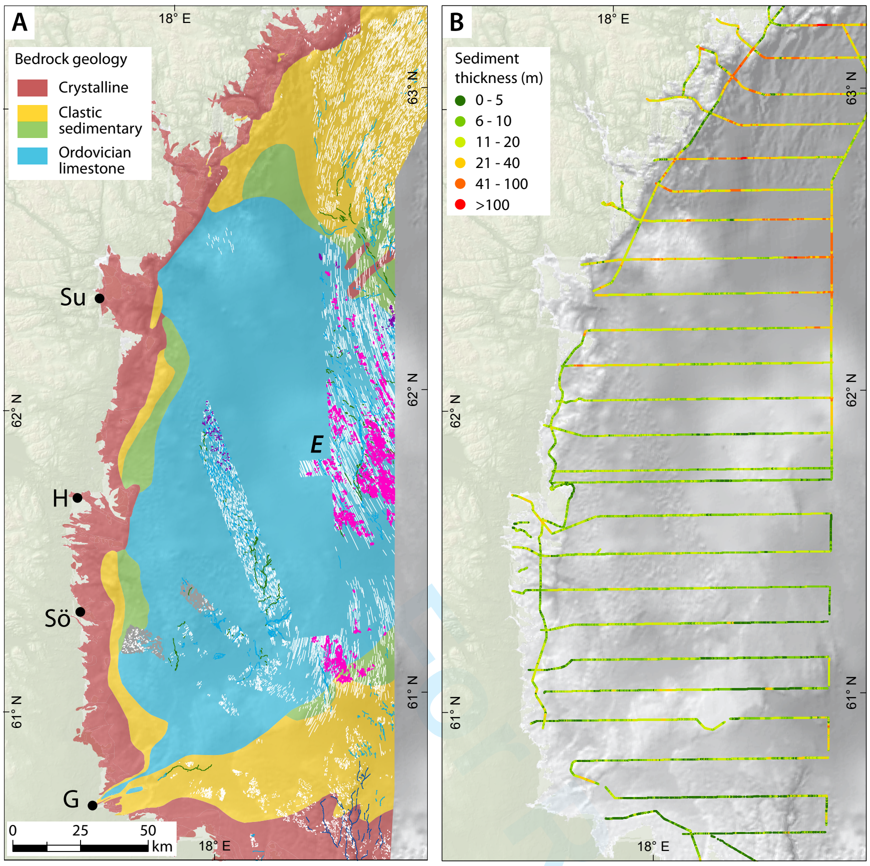

A
18° E
Bedrock geology
Crystalline
Clastic sedimentary
Ordovician limestone
63° N
Su
62° N
E
H
Sö
61° N
G
0 25 50 km
18° E
B
Sediment thickness (m)
0 - 5
6 - 10
11 - 20
21 - 40
41 - 100
>100

Figure 7

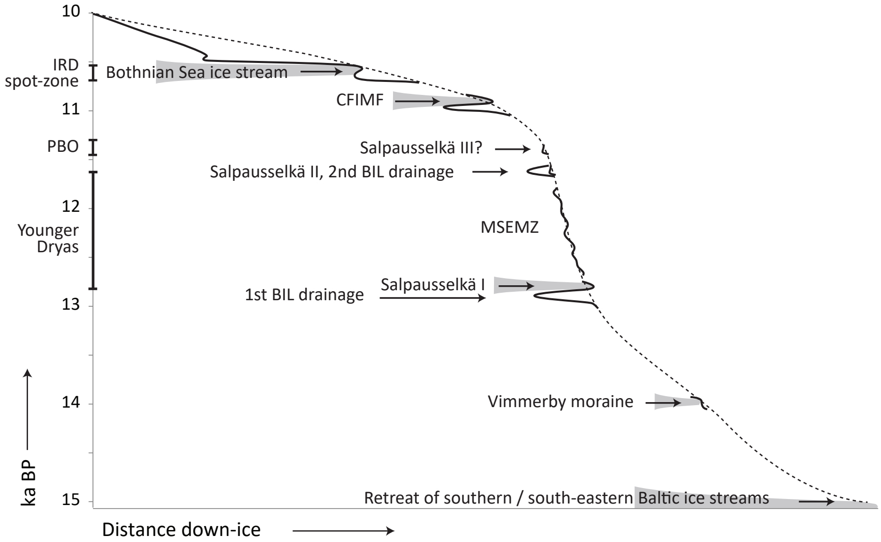

10
IRD spot-zone
11
PBO
12
Younger Dryas
13
14
15
ka BP
Bothnian Sea ice stream
CFIMF
Salpausselkä III?
Salpausselkä II, 2nd BIL drainage
MSEMZ
Salpausselkä I
1st BIL drainage
Vimmerby moraine
Retreat of southern / south-eastern Baltic ice streams
Distance down-ice

Figure 7