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Younger Dryas glaciomarine sedimentation, push-moraine formation and ice-margin behavior in the Middle Swedish end-moraine zone west of Billingen, central Sweden

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A R T I C L E I N F O

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A B S T R A C T

New highway exposures and drilling reveal the stratigraphy and structure of the Middle Swedish end-moraine zone west of Billingen, Sweden. The material in the end moraines is primarily glaciomarine clay of Younger Dryas age that was deposited as varved clay in front of the retreating glacier and then pushed glaciotectonically to form push moraines during minor ice-margin oscillations during overall retreat during the Younger Dryas cold event. The moraines are composed of deformed and remobilized clay with some clayey diamicton and penecontemporaneously deposited and deformed sand. Between the moraines lie ‘intermoraine flats,’ composed of undeformed varved clay of Younger Dryas age and surface sands of Younger Dryas to early Holocene age. Based on estimations of moraine volume, sedimentation rate and ice-margin retreat rates, we calculate the overall ice-margin retreat and end-moraine construction to span 350–800 years within the Younger Dryas. Because the number of moraines in the Middle Swedish end-moraine zone varies across Sweden, we regard the individual oscillations west of Billingen to be driven by local physical and glaciologic factors rather than ice-sheet wide climate drivers. The study area is also the location of the early and final drainages of the Baltic Ice Lake. The final drainage of the Baltic Ice Lake took place several decades after the youngest moraine was formed. We consider it likely that the earlier, Allerød drainage of the Baltic Ice Lake (BIL) also took place at Billingen, despite the lack of clear local stratigraphic evidence. However, based on our model, a retreat driven solely by climate would not have exposed the outlet at Billingen, and we propose a dynamic break-up of the ice-margin likely centered on Valle Härad that was driven by the head difference between the BIL and the sea.

1. Introduction

The Middle Swedish end-moraine zone (MSEMZ) represents one of the most prominent moraine belts in northern Europe. It was constructed during the Younger Dryas cold event (YD). Our purpose here is to describe the internal structure of the moraine ridges and present a model of contemporaneous sedimentation and push-moraine formation in the MSEMZ west of Billingen, and thereby illustrate the dynamic behavior of the Scandinavian Ice Sheet margin during this YD interval.

Additionally, the behavior of the ice-sheet margin west of Billingen during deglaciation is critical to understand because this area is well-known as the location of the outlet for the final drainage of the Baltic Ice Lake (Munthe, 1910; Björck and Digerfeldt, 1984; Strömberg, 1992, 1994; Stroeven et al., 2015). The drainage occurred 11,620 cal yr BP, 30–40 years before the end of the Greenland Stadial I (GS1)(the Younger Dryas cold event in Greenland) (Andrén et al., 2002; Stroeven et al., 2015). Johnson et al. (2010, 2013a) were able to show stratigraphically that the ice-margin was north of Götene (Fig. 1) at the time of drainage. This means that the formation of the MSEMZ, at least in the study area, was completed before the final drainage.

Importantly, a suggested earlier drainage, at the end of the Allerød interstadial (Björck, 1995), has recently been supported by new studies of varved lake sediments in Lake Vättern and in eastern Sweden (Swärd et al., 2016; Muschitiello et al., 2015) and the presence of terrestrial organic sediments in the Arkona basin in the southern Baltic (Bennike and Jensen, 2013). This drainage occurred

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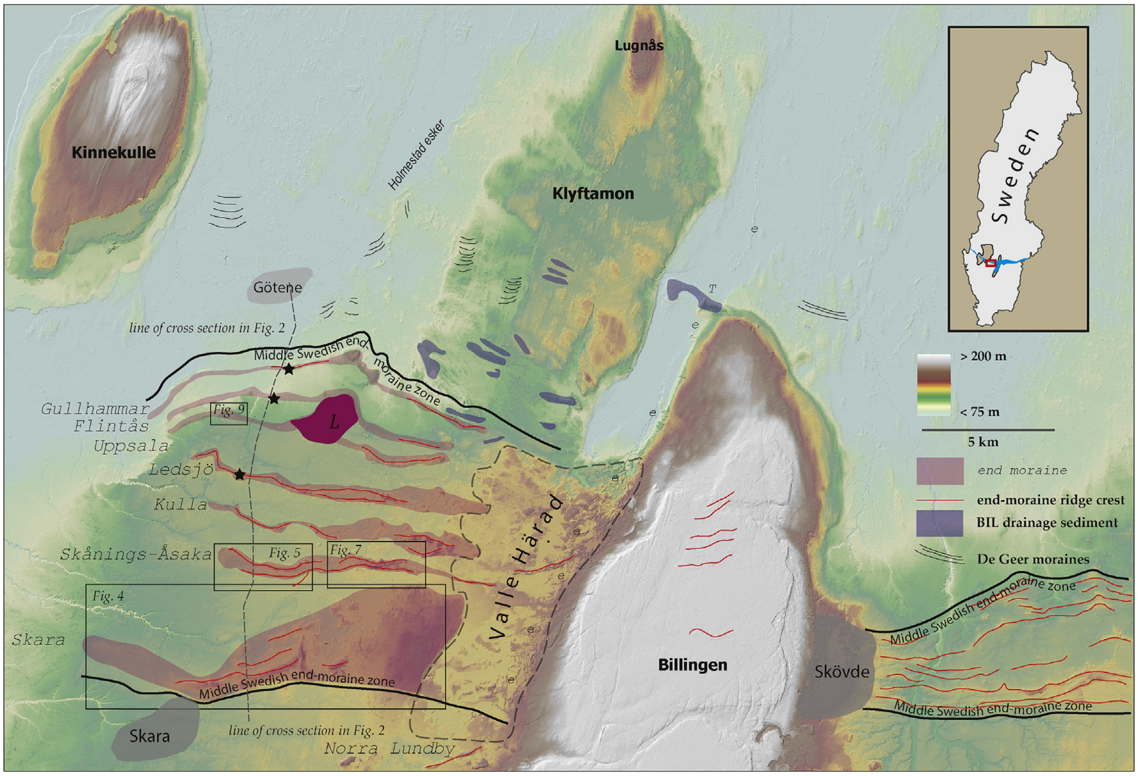

Fig. 1. Map of the study area showing the Middle Swedish end-moraine zone (MSEMZ). Inset shows map location between Lake Vänern (to the west) and Lake Vättern (to the east) as well as the position of MSEMZ across Sweden (in blue). The names of the end moraines are shown west of Billingen. 'e' = prominent eskers and esker nets. 'T' = Timmersdala ridge. 'L' = Ledsjömo, an ice-contact delta built up at the Uppsala and Flintås ice-margin positions. Black stars show locations of cuts through the Ledsjö (Fig. 8), Flintås (Fig. 10) and Gullhammar (Fig. 11) end moraines. Dashed line shows location of cross section in Fig. 2. Not all Baltic Ice Lake drainage deposits shown, nor are all De Geer moraines. Valle Härad contains thick outwash sediment in a bedrock trough with numerous hummocks, kettles, kettle lakes and eskers, but also uncollapsed outwash surfaces and end moraines. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

about 13.0 ka cal yr BP coincident with the start of GSI (Swärd et al., 2016; Muschitiello et al., 2015). This first drainage likely occurred at Billingen as well (Björck, 1995 and references therein, Swärd et al., 2016), although clear evidence for drainage deposits is lacking in the outlet area. Any drainage at this time would require a significant ice-margin retreat or ice-margin collapse, but evidence for this is also lacking. This study will provide a characterization of the ice retreat necessary to allow for this suggested early drainage.

This current study has been made possible due to four significant recent activities in the area. First, geologic drilling has helped reveal the subsurface stratigraphy (Johnson and Ståhl, 2008). Second, the Swedish traffic agency rebuilt the highway E20 starting in 2007, which resulted in large exposures in four of the MSEMZ ridges. The exposure at Ledsjö was described and interpreted in Johnson et al. (2013b), and the remaining moraines are presented here. Third, the Swedish Geological Survey recently completed making new surficial-geologic maps (Påsse and Pile, 2016). Finally, the availability of LiDAR based digital-elevation models (DEMs) of the area in the last few years has allowed the surface geomorphology to be seen in greater detail to allow for more precise mapping.

2. Study area

The study area (Fig. 1) is in south-central Sweden, and the overall physiography is characterized by the residual Paleozoic-cored hills of Kinnekulle, Billingen and Lungnås lying on the extensive, exhumed sub-Cambrian peneplain (Lidmar-Bergström, 1996). The Precambrian surface is essentially flat in the study area except for a prominent fault-line scarp at the west edge of Valle Härad, which in turn forms the edge of the bedrock upland called Klyftamon (Fig. 1).

The expression of the MSEMZ (the presence, number and morphology of moraines) varies across Sweden, but seven prominent end moraines make up the zone in this study area north of Skara and west of Billingen (Fig. 1). Previous work has shown that the ridges here are made predominantly of glaciotectonized glaciomarine clay (Ahlmann, 1910; Munthe, 1910; Lundqvist et al., 1931; Johansson, 1937; Björck and Digerfeldt, 1984; Johnson et al., 2013b) although closer to Billingen, the ridges have greater proportions of coarser sediment (Påsse and Pile, 2016; author field observations).

The seven prominent moraines in the study area are named after local place names (Johnson and Ståhl, 2010). The area defined as the MSEMZ in Fig. 1 is based on the distribution of prominent end moraines; it is a physiographic definition. These moraines have appeared on maps and in publications for over 100 years (Munthe, 1910; Munthe et al., 1928; Lundqvist et al., 1931; Johansson, 1937; Strömberg, 1969; Björck and Digerfeldt, 1984; Johnson et al., 2013a, b; Stroeven et al., 2016; Påsse and Pile, 2016), and their distribution appears basically the same in all these maps.

The moraines in the study area are of Younger Dryas age based on radiocarbon dating (Johnson and Ståhl, 2010) and the fact that Baltic Ice Lake drainage deposits, dated to have occurred 30–40 years prior to the end of GS1 (Andrén et al., 2002; Stroeven et al., 2015), are younger than the northernmost ridge at Gullhammar. The age of the southernmost Skara ridge is not clear but is likely younger than the start of the Younger Dryas (Björck and Digerfeldt, 1989).

In addition to these seven prominent moraines, a smaller moraine ('Norra Lundby,' Fig. 1) occurs south of the Skara line and is geomorphically older. Ahlmann (1910) describes exposures in the Norra Lundby ridge to consist of varved clay that was deformed by ice, indicating this ridge has a push-moraine genesis similar to the other MSEMZ ridges. This ridge is not further discussed here. Based on the elevation of an outwash fan associated with it is, likely of Younger Dryas age.

The end moraines mapped in Fig. 1 are elongate, linear swaths of higher topography that rise above the adjacent flats. In places the moraines are ridge shaped with a distinct crest, but in other places they occur as a line of irregular hills. The Skara moraine is the most complex topographically. It is broad and consists of three separate ridges in the central part. The next three ridges to the north (Skånings-Åsaka, Kulla and Ledsjö) are mostly simple ridges consisting of one or two crests. The northernmost three (Uppsala, Flintås and Gullhammar) appear as 'stair-steps' in the landscape along the line of cross section in Fig. 2. The composition of all these ridges is discussed in detail later in the paper, but there is a significant lateral (west to east) change in the composition. This is due to the different source materials that lay in front of the ice. In the west, the moraines are composed predominantly of deformed glaciomarine clay. To the east, the moraines contain a greater amount of fluvial sand and gravel derived from outwash streams headed in Valle Härad.

Much of the region west of Billingen lay submerged below sea level during deglaciation with the relative sea level between 120 and 130 m a.s.l. (Björck and Digerfeldt, 1984; Strömberg, 1992, and references therein). The crests of the largest end moraines exceed these elevations in places especially to the east, but much of the moraine construction occurred subaqueously along the line of section we have investigated (Figs. 1 and 2).

Between the moraines lie 'intermoraine flats.' These are plains composed of sand and some gravel at the surface (Fig. 2) but that are underlain by glaciomarine varved clay (Johnson and Ståhl, 2010). Thirteen drill holes along the line of cross section in these flats reveal the stratigraphy, which contains up to 25 m of clay (Johnson and Ståhl, 2008). Two holes (at Kollbogården and Mellemarken, Fig. 2) were continuously cored and revealed rhythmic bedding in the clay that was interpreted as varved sediment (Johnson and Ståhl, 2010). A description of the character and geochemistry of the varved clay can be found in Johnson and Ståhl (2010) and Johnson et al. (2013a), but the varves have grayish-red summer layers with 0–4% sand, 35–75% clay, and the remainder silt. These grade gradually upward to red winter layers with 0–1% sand, 50–85% clay with rest silt. Couplet thickness ranges from 1 to 74 cm with an average at Kollbogården and Mellemarken of 22 cm. Beneath the clay, bore holes reveal less than a meter of coarse sediment (sand, gravel or diamicton) overlying bedrock; the regional basal diamicton is thin. This is also shown in surface exposures north of the MSEMZ where sediment thickness is 0–5 m, and highway excavation revealed bedrock and the overlying sediment. In these exposures, the diamicton and any overlying coarse sediment is 0–1 m in thickness beneath the clay.

The intermoraine flats grade upward to the east where they merge with outwash surfaces in Valle Härad. (Fig. 1). The flats are thus distal portions of these outwash plains. In places, the flats show fluvial channels and dead-ice collapse pits. Near Valle Härad, the surface sediment in the flats is pebbly to cobbly, but this sediment fines westward; the surface sediment is predominantly sand along the line of cross section. Exposures and boreholes reveal that the sand and gravel on the flats is the top of coarsening-upward sequences. The sand on the flats is interpreted as marine and fluvial in origin, deposited during sea-level regression that occasioned westward shoreline migration; the age of the surface is younger to the west.

The ridges and the intermoraine flats and their underlying sediment form a regionally thick wedge that marks the YD zone; sediment further south and further north of the MSEMZ is thinner due to the faster ice-retreat rates before and after the YD (Fig. 3). The ice margin remained in this rather narrow zone for over 1000 years and this allowed large amounts of sediment to accumulate.

We consider the Gullhammar ridge to be the northernmost ridge in the MSEMZ (Fig. 1). North of this ridge, surface deposits of clay and till are thin (<5m), and the crystalline bedrock of the subcambrian peneplain crops out in many places. Prominent end moraines are lacking, but numerous smaller De Geer moraines occur here (Frödin, 1916; Johnson et al., 2013a; Bouvier et al., 2015; Påsse and Pile, 2016) as well as the prominent Holmestad esker. Finally, scoured bedrock and cobble-boulder sheets and bars occur on Klyftamon and represent primary features and deposits from the final drainage of the Baltic Ice Lake (Fig. 1) (Lundqvist et al., 1931; Strömberg, 1992; Stroeven et al., 2015).

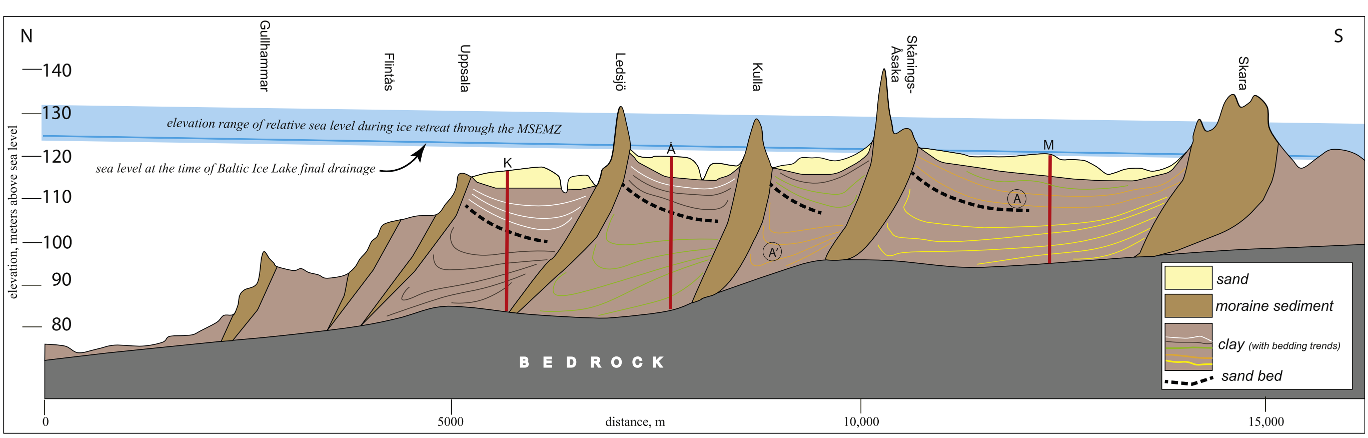

Fig. 2. North-south topographic profile and geological cross section through the study area. Location shown in Fig. 1. The geology is based on our drill holes and observations, but also represents our conceptual model for contemporaneous clay sedimentation and moraine formation. Bedrock surface is generalized but based on information from numerous boreholes recorded by the Geological Survey of Sweden as well as our boreholes and surface exposures. K (Kollbogården), Å (Åängen) and M (Mellemarken) refer to three boreholes described in the text. The colored lines in the clay represent bedding, and each color represents a set of contemporaneous deposits, for example, beds at A are the same age as beds at A'. The black dashed line represents the horizon of sedimentation when the adjacent moraine was being constructed. Vertical exaggeration 55X. See text for explanation. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

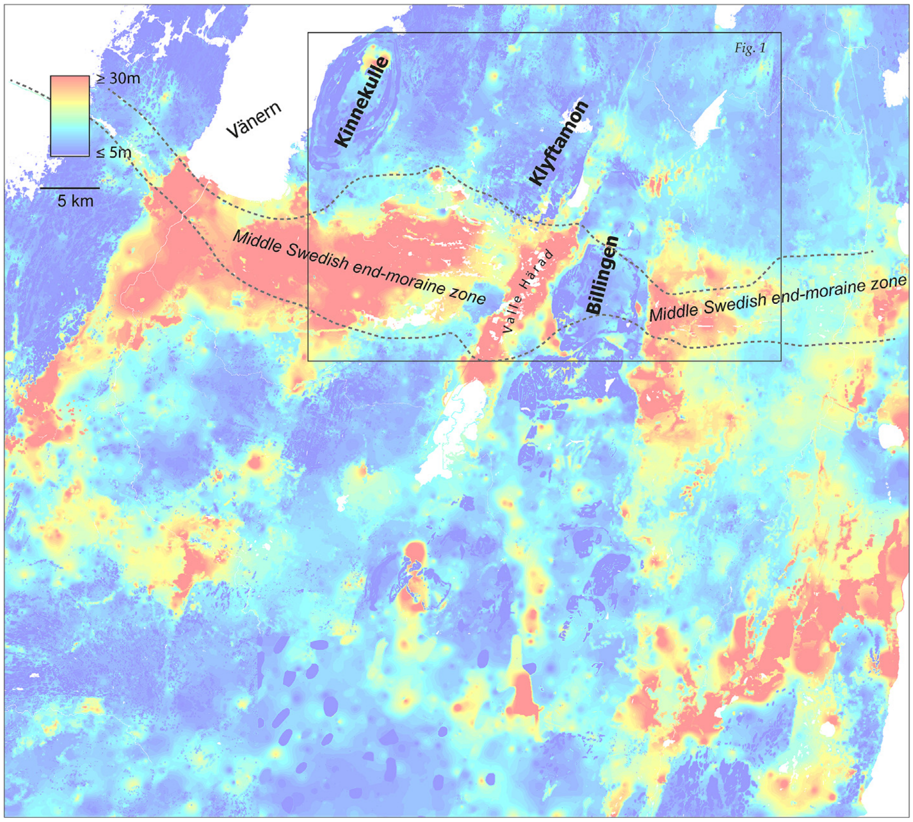

Fig. 3. Isopach map of Quaternary surface sediment from the Geological Survey of Sweden database. Compare to Fig. 1 for location. The east-west zone of thick sediment in the center of the map is broadly coincident with the Middle Swedish end-moraine zone, shown by the dashed lines.

3. Methods

Excavations in new road cuts by the highway department as well as in gravel pits allowed first-hand description of sedimentology and structure. Sediments were described and sketched using standard field techniques. Structural measurements were made with a Brunton compass. Cross sections were documented by laying out a grid and mapping on graph paper, as well as by using photomosaics. Most of the road cuts were documented during road construction, 2007 to 2008.

For drilling, we used a Geotech 604D to sample with auger and Shelby-tube cores at 13 localities in the study area. Continuous core was collected at Kollbogården and Mellemarken (Fig. 2). Continuous cone-penetration-test drilling (CPT; Geotech Probe No 3068) was performed at 11 locations from 2006 to 2008, and we present data from a CPT measurement at Åängen (Fig. 2). We interpret the bottom of each of our cored holes to be at or near bedrock rather than a boulder, because the bottom elevation in each hole was found to be at a similar elevation to bedrock in nearby well records on file at the Geological Survey of Sweden (Seger, 2011). A description of these cores can be found in Johnson and Ståhl (2008).

We present electrical resistivity measurements from the Skara and Skånings-Åsaka ridges. Resisitivity was measured using Continuous Vertical Electrical Sounding (CVES) using a Wenner 4-electrode array with spacing of 2.5 and 5-m between electrodes. The data was analysed using the RES2DINV program, version 3.3 (Loke, 1999).

3.1. Moraine composition, structure and genesis

In this section, we present the geomorphology, internal composition and interpreted genesis of the seven ridges in the MSEMZ west of Billingen.

3.2. Skara

The Skara moraine is the largest end moraine in the MSEMZ, up to 3 km wide and rising up to 35 m above the surrounding plains. It is geomorphically and sedimentologically more complicated than the other ridges (Fig. 4). It is a broad zone that varies from west to east in the field area. Two to three subtle ridges are apparent in the DEM in the western part, and the faint ridge visible in the DEM northwest of Skara (Fig. 4) is the westward extension of this moraine; further west, it is buried beneath clay and sand. The ridge in the town of Skara and to the east consists of reddish clay where observed in roadside exposures. In the eastern extension of this moraine, closest to Valle Härad, the ridges are replaced by a broad upland composed predominantly of coarse, sandy till, but there are also abundant outwash deposits, eskers and even isolated bedrock outcrops (Påsse, 2006) (Fig. 4). Clear moraine ridges do not exist in this part of the Skara zone.

Few exposures were available in the Skara ridge during field work. We present here the resistivity profile taken across the southern-most ridge as well as a short core (Fig. 4). The resistivity profile is over 500 m long and crosses the crest of the southernmost ridge. Resistivity values range from 10 to 1000 Ω. The bottom of the resistivity profile is characterized by material with high resistivity (over 800 Ω), above which there is a gradational change to low resistivity at the top. This general pattern is disturbed in the central part of the profile, at 500–800 m, where a south-verging structure of high resistivity extends almost to the surface. Based on resistivity values, we interpret the material with low resistivity to be

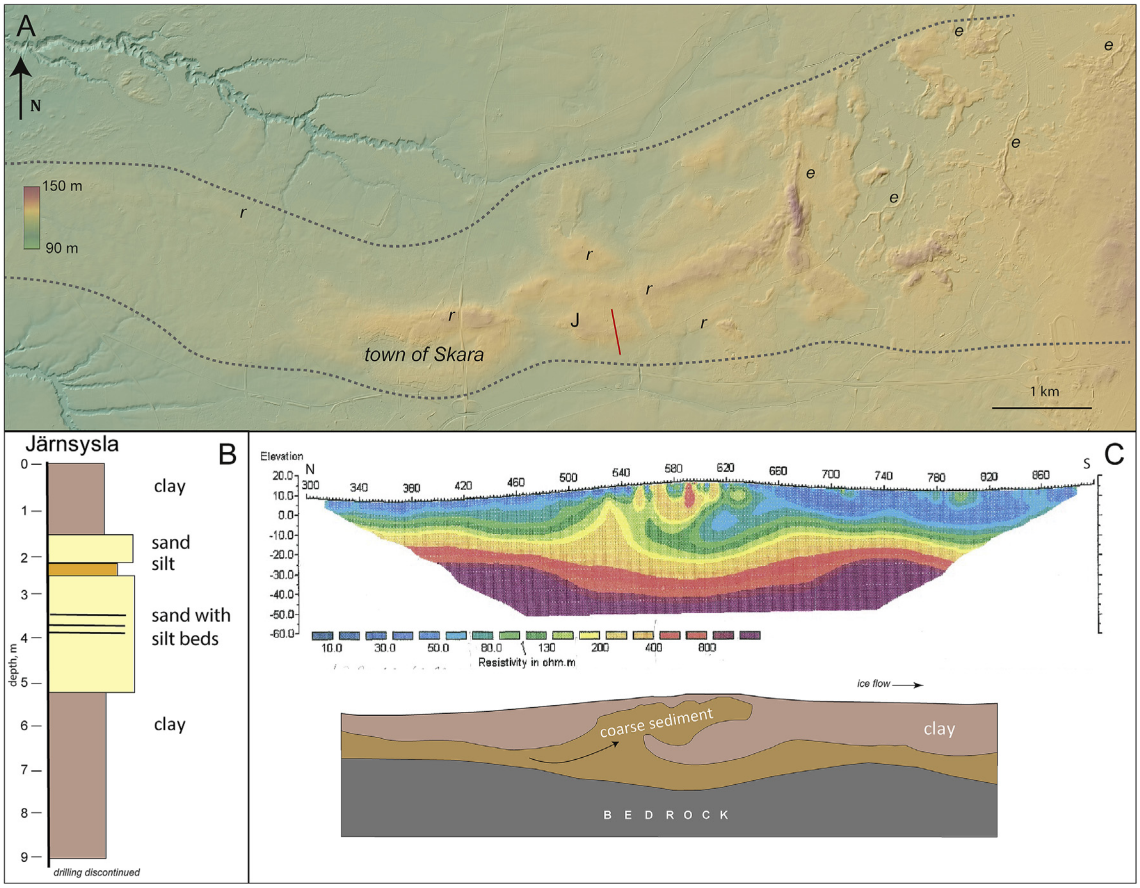

Fig. 4. The Skara end-moraine zone. A. LiDAR DEM with limits of moraine shown by dashed line, location shown in Fig. 1. Red line shows location of resistivity profile. 'J' = location of drill hole in 4B at Järnsysla, 'r' = ridges within the Skara moraine zone, 'e' = eskers. B. Drill hole log from Järnsysla. C. Resistivity profile and interpretation. Horizontal and vertical scale on interpretation is the same as in the resistivity profile. We interpret the pattern to indicate glacial thrusting of the coarser material up over the clay. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

predominantly clay and other fine sediment (Mussett and Khan, 2000; Lowrie, 2007). This is corroborated by a drill hole taken near the line of section at Järnsysla ('J' Fig. 4A) that contains clay, silt and fine sand (Fig. 4B). Reasonable estimates of material velocity suggest the high-resistivity material at the bottom to be bedrock (Mussett and Khan, 2000; Lowrie, 2007). Additionally, this level in the profile occurs at about the elevation for bedrock observed in nearby wells. Based on observations in outcrops further north, it is common that till and coarse, sorted material occur on top of bedrock underneath the thick clay, thus the zone of gradational change from high to low resistivity values likely includes coarse sediment. We interpret the sediment body in the middle of the section to be coarser material (likely sand and gravel, based on exposures in other ridges and the drill hole at Järnsysla, but perhaps gravel and diamicton).

Based on the structure seen and the regional stratigraphy, we suggest that this coarse material in the center of the resistivity profile was originally lower in the stratigraphic sequence and thrust up into the clay during a readvance of the ice (Fig. 4C). The top of the suggested thrust mass coincides also with the topographic high point of the ridge, suggesting that it was this thrusting that also produced the ridge. This is common for both modern and Pleistocene push moraines (e.g. Bennett, 2001; Johnson et al., 2013b; Benediktsson et al., 2015). We therefore interpret at least this part of the Skara ridge to be a push moraine. The interpreted stratigraphy and sedimentology requires that the ice margin retreated north of the Skara ridge position prior to readvance in order to allow sedimentation of material later pushed into the moraine.

3.3. Skånings-Åsaka

The Skånings-Åsaka ice margin is marked by a double ridge; a more northerly higher ridge, where Skånings-Åsaka church stands, and a southerly lower ridge with strikingly abundant boulders on the surface. An additional minor, more southerly third ridge is present but is for the most part buried beneath the sediment in the intermoraine flats (Figs. 1 and 5). Based on the evidence we present below, we interpret these ridges to be push moraines each preceded by a retreat of the ice margin.

Simon Johansson (1926, p. 248), through a series of borings 150 m west of the church, was able to show that the northern ridge consists of a surficial unit of clay with some surface boulders overlying thick (7.5 m), uniform sand which, in turn overlies clay. Johansson interpreted the thick sand as being deposited as an ice-contact delta built out over clay of glaciomarine origin (Fig. 5C). This interpretation requires an ice-margin retreat followed by a readvance of the ice. An exposure described by Johansson also showed that the ice eventually 'slid' over and incorporated the sand, producing the thin clayey diamicton with boulders at the surface on the proximal side; these boulders are visible today. The abundant boulders on the southern ridge were explained by Johansson (1926) in part as being concentrated by wave-current washing of till (Fig. 6).

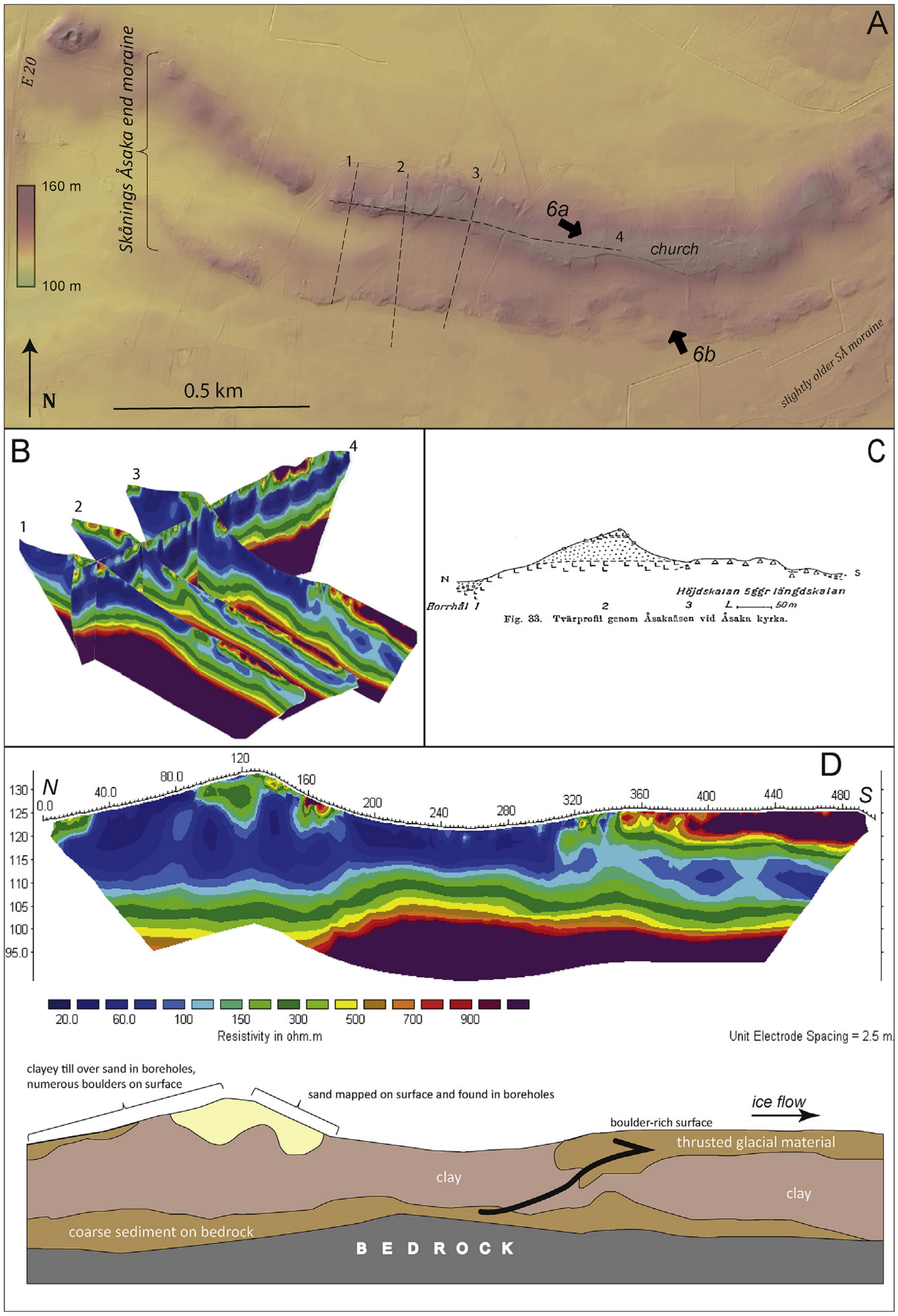

Fig. 5. The Skånings-Åsaka end moraine near the Skånings-Åsaka church. A. LiDAR DEM showing the two prominent ridges; location shown in Fig. 1. Dashed lines show location of resistivity profiles 1–4 in B and D. Lettered arrows are locations and view-directions of photos in Fig. 6. B. Fence diagram of the four resistivity profiles. C. Profile and cross-section from Johansson, 1926). L = clay (Sv. *lera*), triangles = boulder rich area, stippled pattern = sand. (Sv. *Tvärprofil genom Åsakaåsen vid Åsaka kyrkan* = 'Cross section through Åsaka ridge at Åsaka church,' *Höjdskalamn 5 ggr längslana* = Vertical scale 5X horizontal scale) D. Resistivity-profile 3 and interpretation based on profile, borings and surface observations. Arrow represents interpreted thrust movement of ice, see text for details.

Johnson and Ståhl (2008) show borings in the northern ridge that agree with Johansson's observations including the presence of a clayey diamicton (up to 5 m thick) on the proximal side of the northern ridge overlying sand mixed with some clay. On the south side of this ridge, Johnson and Ståhl (2008) report medium sand with minor beds of clay in boreholes (up to 10 m) similar to Johansson's (1926) observations. Bedrock is found around 80 m a.s.l. according to local well logs indicating that the clay in the flats north and south of the ridges is about 40 m thick. During ice retreat and ridge formation, it is difficult to know precisely the relative sea level, but we assume it to have been between 125 and 130 m above present sea level (Fig. 2); the ridge crest is up to 140 m in places.

To compliment Johansson's and our observations, we present resistivity profiles that show the character of sediment above bedrock. Three profiles cross both ridges and one runs parallel to the crest (Fig. 5). High-resistivity material occurs at the bottom of all profiles. The southern ridge has high-resistivity material at the surface overlying low-resistivity material in the middle, while the northern ridge comprises intermediate- and high-resistivity material at the proximal and distal slopes, respectively. Similar to the

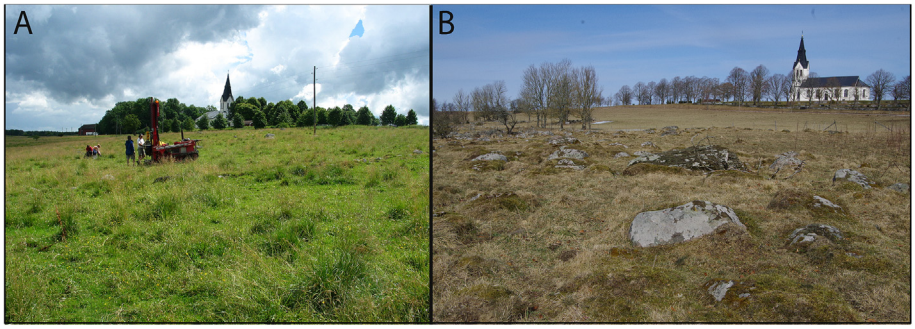

Fig. 6. Photos of the Skånings-Åsaka end moraine, locations shown in Fig. 5. A. On the proximal side of the north ridge looking towards the church. B. On the distal side of the southern ridge looking towards the church on the north ridge. The southern ridge is covered with abundant boulders.

Skara profile, we interpret the results to indicate glacier-shoved sediment composed primarily of clay and coarser sediment overlying bedrock. We interpret the high-resistivity material at the surface of the southern ridge to correspond to the boulder material seen in the field (Fig. 6). The low-resistivity material below is interpreted to be clay. In the northern ridge, the coarse, high-resistivity material at the bottom is overlain by more clay, which in-turn is overlain by coarser material (sand) at the surface. The margin must have retreated north of location of the boulder ridge to allow for clay deposition, followed by advance and thrusting up of basal and subglacial boulder-rich debris over the clay to form the southern crest of the ridge (Fig. 5). Alternatively, the pattern in the southern crest of the main ridge can be interpreted to show that the ice itself rode up over the clay and deposited boulder-rich sediment directly upon the clay (as opposed to proglacial thrusting). The ice margin then retreated allowing glaciomarine clay (and perhaps sand) to accumulate, but then readvanced again to form the northern ridge. The sand in the northern ridge was either deposited in place at the ice margin as a prograding ice-contact delta (as suggested by Johansson (1926)), or deposited a short distance to the north and moved into the ridge glaciotectonically. The fact that the clay in the resistivity profile is ridge-shaped indicates that glaciotectonism was involved to push up the clay. Finally, the ice deposited the layer of clayey diamicton on the proximal side, as seen in borings by Johansson (1926) and Johnson and Ståhl (2008). We interpret this diamicton to be subglacial traction till.

Four kilometers further east of the Skånings-Åsaka church, near Bränningsholm, an isolated outcrop in the Skånings-Åsaka ridge occurs on the proximal side of the northern ridge (Figs. 1 and 7). Here the ridge is composed of an overturned fold of sand with thin beds of red clay; the deformation orientation is from the west-northwest (Fig. 7). Several low-angle reverse faults and several steep normal faults occur in the section striking roughly parallel to the local proximal slope and to the axis of the fold. The sand is medium grained, bedded and contains clasts that include redeposited red glaciomarine clay and Cambrian shale, the latter of which indicates sediment derived not directly from the adjacent ice to the north but from westward-flowing outwash streams sourced in Valle Härad area to the east (Fig. 1). This sediment was deposited as outwash in shallow, marine water where both sand and thin beds of clay could accumulate, and this occurred just north of the Skånings-Åsaka line during a brief ice-margin recession. This also indicates that this sediment accumulated relatively close to and slightly below the former sea level. This outwash was then pushed during readvance into the fold observed here. The Skånings-Åsaka ice margin is oriented predominatly east-west, but the local moraine morphology along with this particular stress orientation shows that the ice margin included some embayments, and we consider the northwest structural orientation to be expected from such a geometry (Fig. 7). This exposure shows that ice-margin oscillations were occurring simultaneously with subaerial outwash deposition in Valle Härad and westward shoreline progradation.

3.4. Kulla

The Kulla ice-margin position is characterized by a discontinuous ridge and series of hills. No significant exposures were found in this ridge during our study. Carlsson (2002) described several small outcrops in the moraine that consisted of sand and clay, but not diamicton. Some sand layers displayed steep dips and small faults, indicating deformation, and a resistivity profile showed primarily low-resistivity material in the core of the ridge. He interpreted these findings to indicate the ridge is composed mostly of clay and was emplaced by ice shove.

3.5. Ledsjö

The Ledsjö moraine is the largest ridge that was excavated during the construction of E20. The composition, structure and genesis of this ridge as exposed along the highway was reported on by Johnson et al. (2013b), and detailed sketches and structural data can be found there as well as an explanation of the ridge's push-moraine origin. Here, we provide a summary of their findings and interpretations as well as a summary figure (Fig. 8).

According to Johnson et al. (2013b), the Ledsjö ridge is a push moraine formed sub-aquatically and composed mostly of glaciomarine clay, but also with pods and beds of fine to medium sand. The geomorphology, structure and sediment architecture of the exposed ridge reveal that it is the product of two advances of the ice margin. The main ridge (from 200 to 450 m in Fig. 8) was formed during the first advance to the Ledsjö position, and is composed of deformed clay with pods of sand. The color banding associated with the source varved clay (Johnson and Ståhl, 2010) is present only in a few places, and Johnson et al. (2013b) suggested that, as the ice advanced, the clay was mobilized as flows down the distal side of the ridge as well as sliding along slump surfaces (Fig. 8). This interpretation is similar to that proposed for subaqueous push moraines in Svalbard, where Kristensen et al. (2009) refer to them as ‘continuously failing, mobile push moraines.’

The sand beds and pods in Fig. 8 are interpreted to represent penecomtemporaneous deposition near the ice margin by sea-bed reaction currents generated by lofted meltwater currents coming from the ice that were carrying sand and fine sediment (for example, Powell, 1990). As rising meltwater plumes lost their competence, the lofted sediment rained down and was redeposited by reaction currents flowing at the bed towards the ice margin (see

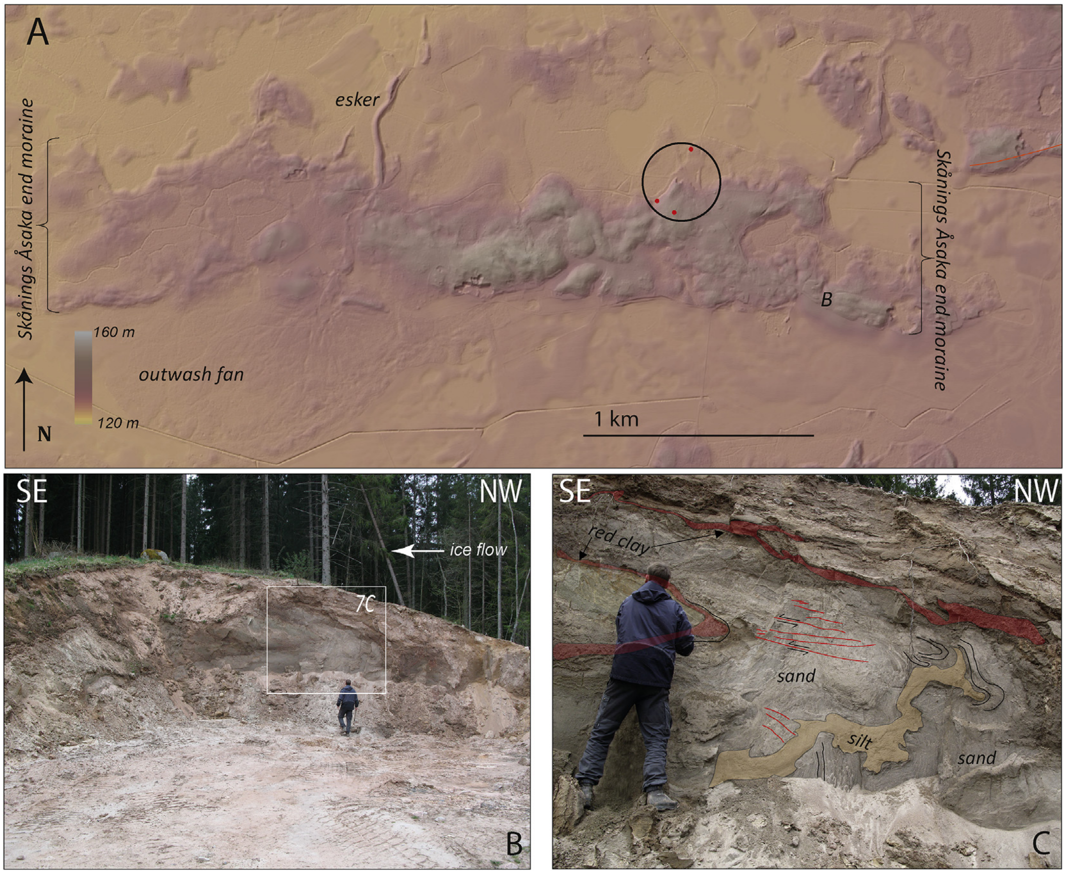

Fig. 7. Outcrop in the Skånings-Åsaka end moraine near the farm of Bränningsholm. A. LiDAR DEM of the east-west Skånings-Åsaka ridge, outcrop in middle of circle. B = Bränningsholm. The circle is also a stereonet that shows the orientation of three fold axes (red dots) including the prominent one shown in C. Also shown is an esker and outwash-fan pair created while the ice margin stood at Skånings-Åsaka moraine. Note that the proximal side of the moraine at the outcrop is oriented NE-SW. B. Photo of exposure looking towards the southwest; ice flow from the right. C. Exposure of sand, red clay and silt in an overturned fold. Red lines indicate minor thrust faults with arrows showing sense of motion; black lines trace some representative bedding. Glaciotectonic stress was from the right. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

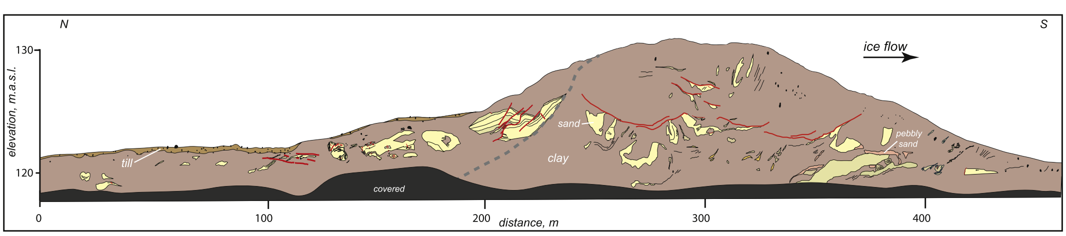

Fig. 8. Outcrop sketch of the Ledsjö end moraine based on field sketches presented in Johnson et al. (2013b) and other field notes and photographs. The sketch is a composite of exposures on the east- and west-facing sides of the highway excavation. Red lines are faults; black lines represent bedding. Gray dashed-line is contact between first-phase and second-phase sediment; the first advance to the Ledsjö position created the crest and deposited the sediment to the right of the dashed line. Sand beds to the left (north) of the dashed line were during a small retreat and then deformed during the second advance to the Ledsjö position. Vertical exaggeration 5X. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

Johnson et al., 2013b). Some sand pods were subsequently deformed as they became covered by mobilized clay and further pushed by the advancing ice, and these are present as irregular pods in the clay displaying both brittle and fluidized deformation. Thus, some of the most distal (and youngest) sand beds (near the base of the outcrop, 350–420 m Fig. 8) are only slightly deformed and buried by flowed clay. The orientation of the ridge crest and structural elements shows ice-shove from the north during this phase (Johnson et al., 2013b).

On the proximal side lies a younger stratigraphic and tectonic unit composed of clay and sand that is less deformed. This sand was deposited during ice-margin retreat from the ridge crest (Fig. 8). The second advance to the Ledsjö position deformed these sand beds (at 190–240 m, Fig. 8) and they display normal and thrust faults as well as intruded clay. The sand here is also interpreted as reaction-current sediment because of 'up-ice' ripple-cross-bed orientations. The orientation of faults and drag folds, as well as a younger ridge crest (running west-northwest from the cut to where Ledsjö church sits, see Johnson et al., 2013b) indicates ice flow during this second advance to be from the north-northeast.

Up-ice, between 125 and 190 m (Fig. 8), these same sand beds are further disrupted into 1–5 m boudin-like pods surrounded by clay, which is interpreted to have been mobilized and to have flowed around the pods. Internally, the pods are undeformed to brittley deformed.

Further up-ice, at 0–120 m (Fig. 8), an entirely different tectonic style is present with shallow, down-ice dipping normal faults associated with smaller up-ice dipping reverse faults. This represents a subglacial, extensional environment, similar to that found submarginally at Eyjabakkajökull (Croot, 1988; Benediktsson et al., 2010), and it is this zone that we consider to be the source of remobilized clay that is moved around the sand boudins and to the ice margin. It is only in this proximal part of the moraine (from 0 to 200 m (Fig. 8)) that diamicton is present, a thin layer of clayey till at the top of the moraine, which shows the deformation is this section to be subglacial/submarginal.

3.6. Uppsala

The Uppsala, Flintås and Gullhammar ice-margin positions (IMP) appear as 'stair steps' in the landscape along the line of cross section; the total sediment thickness south of each ridge is thicker than north of each ridge (Fig. 2). Further to the east, the stair-step nature is replaced by clear moraine ridges (Fig. 1). An additional difference from the moraines to the south in the MSEMZ is that these three ridges contain significant amounts of clayey diamicton instead of primarily glaciotectonized clay.

A small portion of the Uppsala ridge is exposed in the gravel pit at Stenåsen south of Götene (Fig. 9). The pit contains sand and gravel, but much of this has been removed, and, as of 2017, the pit is closed. The gravel removed was up to cobble size and, together with subaqueous outwash material mapped to the south, this suggests that the Stenåsen deposit was a submarine tunnel-mouth deposit.

Excavation of coarse material revealed a stroke of massive clay with boulders, which we interpret as clayey diamicton making up a buried moraine ridge (Fig. 9). This buried ridge aligns with the topographic expression of the Uppsala ridge, and thus we consider it to be an exposure of the sediment in the ridge. The clayey diamicton forming the ridge here has numerous large boulders, many of which match the lithology of the local fine-grained gneissic bedrock. Additionally, there are small exposures that occur in the gravel pit of bedded and folded clay (Fig. 9). Deformation of clay is also documented by Påsse and Pile (2016, their Fig. 42). All of these deformations occur in locations on the distal, down-ice side of the clay-diamicton ridge. Strike-and-dip measurements of these folds (Fig. 9C) indicate that they have been folded by a force from the northwest, and we interpret them to have been formed through push by the advancing ice.

We suggest that following initial ice-margin retreat north of this position, the ice re-advanced through the recently deposited glaciomarine clay, pushing and deforming it. Because the clay was thin, the ice also was able to incorporate local bedrock (and/or coarse basal debris), producing the clast-rich clayey diamicton, which we interpret to be till, and pushing it into the ridge form as well as deforming the bedded clay in front (Fig. 9, unit 2).

While the ice stood at the Uppsala position, the ridge was buried by outwash material largely deposited subaquatically at the ice margin (there is no evidence the outwash reached the sea surface) (Fig. 9, unit 3). The clay till and the outwash are in turn covered by bedded glaciomarine clay deposited in the sea following retreat (Fig. 9, unit 4), and wave-washed sand deposited during emergence (Fig. 9, unit 5).

3.7. Flintås

During highway construction, a small exposure was also cut through the Flintås moraine, the second of 'stair-step' ice-margin positions (Figs. 2 and 10). Here, too, the glacier deposited clayey diamicton, interpreted as till, which overlies proglacial varved clay that has been deformed. Strike-and-dip measurements show a deformation of the varved bedding from the northeast. The diamicton unit ends at the location of the ridge, indicating this to be the former front edge of the glacier and the glacier advance. Sand, interbedded in clay, is present a few meters distal of the southernmost edge of the diamicton unit. The sand lens is slightly folded in a U-shape (Fig. 10) and can be interpreted as also being deformed by the ice. Road exposures further south (distal) of the moraine show the proglacial clay at the surface to be underformed. We interpret the Flintås ridge as a push moraine.

3.8. Gullhammar

The northernmost ridge in the MSEMZ is the ridge at Gullhammar, exposed by excavation during highway construction (Figs. 2 and 11). The stratigraphy and structure of this ridge perhaps shows best the oscillatory nature of the ice-margin retreat and the push-moraine origin of the ridges. The new overpass constructed over E20 rests on the actual crest of the moraine. The exposure reveals diamicton and clay; the former interpreted as till and the latter in various states of deformation (Fig. 11). The crest of the moraine and the proximal side is composed of coarse till. This till is somewhat variable in composition, with lenses of sorted sediment, but is much coarser than the till found in the more southerly moraines. Bedrock outcrops of the subcambrian peneplain occur shortly north of the moraine and indicate that the base of the exposed section is not far above bedrock.

On the distal side of the moraine crest, three units of clay are present; (1) undisturbed varved clay, (2) folded and thrusted varved clay, and (3) remobilized, flowed clay (Fig. 11). The varved clay is of the gray-up-to-red type described by Johnson and Ståhl (2010), and in the outcrop each couplet is on the order of 15–20 cm thick (Fig. 11). The undisturbed varved clay occurs in the middle and distal parts of the exposure. Importantly, at the far distal (southerly) end of the cut, the varved clay drapes a small outcrop of till that has the geometry of a De Geer moraine. Numerous De Geer moraines occur immediately north of this ridge (Fig. 1), and these have a similar relief and composition as the till in the outcrop, As suggested for other De Geer moraines, this minor till ridge was likely made by pushing during a small oscillation (winter advance?) of the ice during overall retreat (Bouvier et al., 2015; Ojala et al., 2015, 2016). Following and during ice-margin retreat, the minor ridge was draped with the varved clay. About 20 undisturbed varves overlie the minor ridge.

Closer to the center of the cut, and immediately abutting the diamicton, the varved clay is folded and thrusted. Structural data on faults indicate force from the north (Fig. 11). The graded nature of the varves provides a way-up indicator to show that some of the meter-scale folds are overturned (Fig. 11C).

At the top of the cut and distal from the ridge crest is clay that we interpret to be flowed clay that was varved but then remobilized during ice advance. The clay overlies the folded thrust clay near the ridge, and it overlies the undisturbed varves further to the south. The clay has similar color variations and grain size as the varved clay, but has been completely deformed; the color and grain-size variations appear as small irregular wisps and lenses in the overall massive clay (Fig. 11B). As opposed to the folded clay below, we interpret this clay to have been transported by shoving of the ice a longer distance, during which time the varved structure was completely destroyed. It is this mobilized clay that covers the ~20 undeformed varves. This indicates that it took at least (or as little as) 20 years for the ice margin to retreat from the De Geer

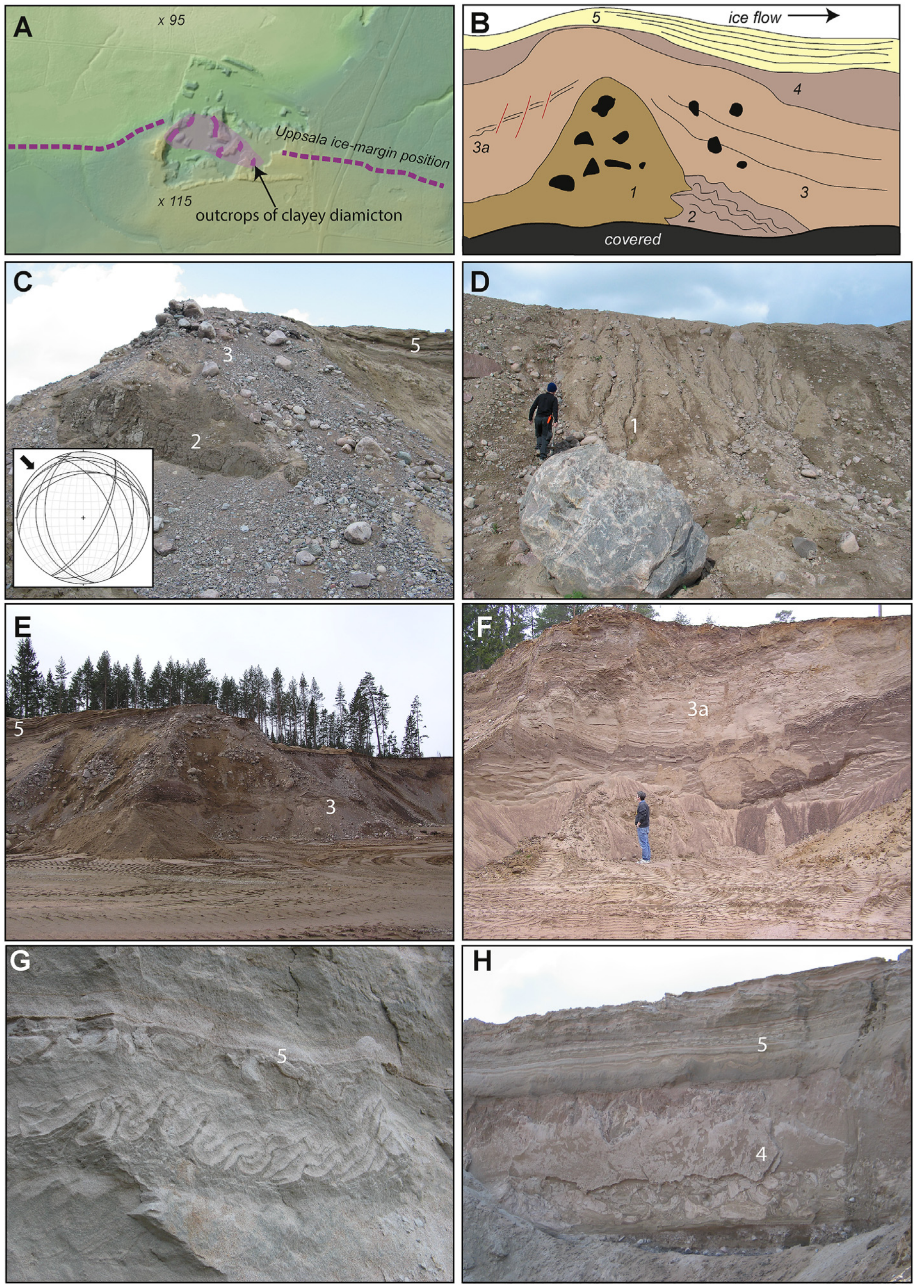

Fig. 9. Uppsala end moraine exposed in the Stenåsen gravel pit south of Götene (Fig. 1). A. LiDAR DEM showing Stenåsen gravel pit, outcrops of clayey diamicton (dark pink lines) and interpreted extent of buried moraine ridge (light pink). B. Cartoon sketch of stratigraphic relationship of units shown in the pit. Orientation roughly north-south with ice flow to the south. C–H. Photos of stratigraphic units. 1 = clayey diamicton (D), 2 = bedded and folded glaciomarine clay (C), 3 = sand and gravel (up to cobble size) (C, E, F), 3a = sand and gravel with normal faults (F), 4 = marine clay (H), 5 = wave-washed sand (C, E, G, H). In C, stereonet shows attitudes of clay fold limbs from unit 2. Folds indicate NW compression. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

moraine and readvance to the Gullhammar position.

The following sequence of events for the formation of the Gullhammar ridge is clear. The ice formed the small De Geer like ridge (water depth ~60 m) and then retreated north of the eventual position of the Gullhammar ridge. Throughout this time, varved clay was deposited, draping the newly deglaciated sea-bed, but as the ice readvanced, till was deposited and the clay was folded, thrusted and eventually thoroughly deformed during continued ice advance. The folded facies would have been formed at the advancing ice margin, but continued advance would raise this material and cause further deformation, not unlike how a snow-plow blade picks up undisturbed snow, and then raises it and pushes it forward in a deformed mass. The folded facies of the clay represents the initial stage of the last clay deformed, and the overlying mobilized clay represents clay entrained earlier, from further north.

3.9. Summary of moraine composition and genesis

We interpret the seven ridges described above as formed as push moraines (Chamberlin, 1894; Boulton et al., 1999; Bennett, 2001). The composition of the moraines is predominantly proglacial glaciomarine clay that was deposited during ice-margin retreat and then pushed and deformed during an oscillation of the overall,

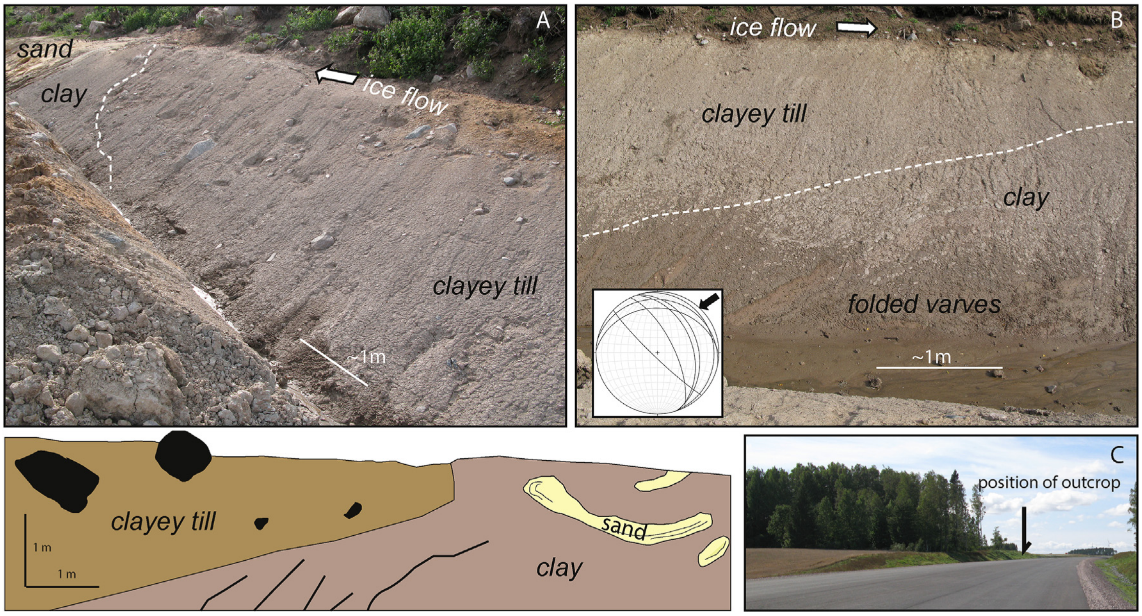

Fig. 10. Photos and sketch of the Flintås end moraine exposed during road excavation with clayey till, folded varved clay and slightly deformed sand pods (location in Fig. 1). The sketch is from the west-facing side of the road cut; ice flow to the south, which is to the right in the sketch. A. Photo of east-facing cut looking south. B. Photo in west-facing cut of clayey till over folded varved beds (reddish layers). Bed orientation indicates stress from the northeast (arrow). C. Photo looking south towards the outcrop position. Line of trees is on the crest of the end moraine.

retreating ice margin. Bodies of sand are seen in outcrops, interpreted from resistivity profiles, and reported in the literature, and these are interpreted as being deposited at the ice margin either as fan/delta deposits or reaction-current rain-out sediment. Diamicton is more prevalent in the three northern most ice-margin ridges, with clayey diamicton from reworked glaciomarine clay in the Uppsala and Flintås ridges and a sandier diamicton in the Gullhammar ridge. The south-to-north evolution of sediment composition (clay (and sand) to diamicton) is seen as a result of thinner sediment sequences and greater proximity to bedrock during ice-margin retreat.

3.10. Stratigraphy of intermoraine flats

The stratigraphy of the intermoraine flats as described below provides important additional information to understand the overall genesis of the moraine ridges and ice-margin behavior in the MSEMZ. The surface unit of sand (pebbly in places, especially to the east on each flat) is 2–8 m thick and is in all cases a coarsening-upward sequence overlying the glaciomarine varved clay, which is 20–30 m thick (Johnson and Ståhl, 2008). There is only minor deformation in the clay revealed in the Kollbogården and Mellmarken boreholes (Fig. 2) (slightly tilted or irregular varves above and below undeformed varves), which is attributed to minor mass movements during deposition, as opposed to pervasive glaciotectonism. One surface exposure formed during construction also revealed the undeformed varve bedding (Fig. 12).

What we regard as most significant in comparing the intermoraine-flat stratigraphy and the moraines is that the moraines are a clear product of glaciotectonism whereas the clay in the flats is essentially undisturbed. This relationship is a critical observation that undergirds the model we discuss below. Additionally, the age of the glaciomarine clay under the flat had been interpreted to be of ‘last interglacial age’ by Johansson (1937, p. 427–428) and suggested to be of Allerød age by Björck and Digerfeldt (1984, 1986), but Johnson and Ståhl (2010) showed that the clay was of Younger Dryas age by radiocarbon dating. Tests of foraminifera and ostracods from glaciomarine clay, cored at Kollbogården (Fig. 2) between the Ledsjö and Uppsala ridges have an age of 12,150 cal yrs BP (Johnson and Ståhl, 2010). The dated varves from this core were deposited shortly after (a few decades) the construction of the Ledsjö ridge (Fig. 1) and prior to the Uppsala ridge. Thus the age of the varved clay is from the Younger Dryas cold interval indicating that clay deposition was occurring during the same time interval as end-moraine formation.

Additionally, there are distinct sandy horizons at depth in the varved sequence that we regard to indicate proximity of the ice margin during a subsequent oscillation (Fig. 2). At Kollbogården, north of the Ledsjö ridge, Johnson and Ståhl (2010) described about 30 muddy sand layers within the varved sequence at 20.5 m in the core as being deposited when the ice had readvanced to and occupied the younger Uppsala ice margin. At that time, an ice-marginal fan (‘L’, Fig. 1) formed, and Johnson and Ståhl (2010) interpreted the sand beds to represent distal sand flows formed when the ice was at the Uppsala margin. Similarly, a continuous cone-penetration test (CPT) at Åängen, 800 m south of the Ledsjö ridge, reveals distinct intervals at 15–16 m depth that are interpreted as sandy beds within the clay (Fig. 13). These sand beds are interpreted to have formed when the ice readvanced to and was at the Ledsjö ridge. The sedimentology, stratigraphy and structure of the sand pods in the Ledsjö push moraine (Fig. 8) show that proglacial, marine sand was being deposited at the ice margin during moraine construction, and we interpret the sand in the Åängen core to have been deposited at the same time but more distally. The significance of these beds as a stratigraphic marker is discussed below.

4. Discussion

4.1. A conceptual model for penecontemporaneous clay deposition and push-moraine formation

The cross section in Fig. 2 is used to illustrate a conceptual model

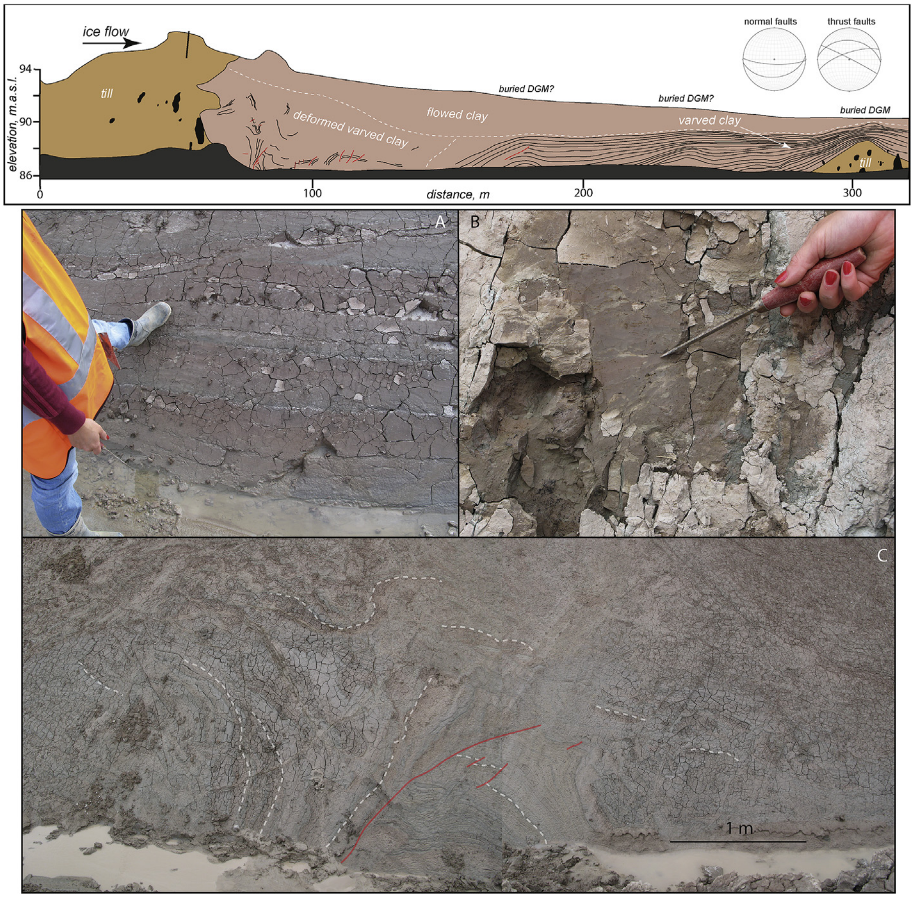

Fig. 11. Gullhammar end-moraine sketch and photos. Dashed white lines in the sketch are approximate boundaries between the varved clay, deformed clay and flowed clay. Measurements of thrust and normal faults taken within the deformed clay. DGM = De Geer moraine. Vertical exaggeration 5X. Photos: A. Undeformed, varved clay. The reddish part of the varve is the 'top' of the couplet. B. Flowed clay. Wisps and patches of red, gray and tan represent remobilized, flowed varved clay. C. 2-photo mosaic of folded and thrusted deformed varved clay at the bottom of the cut. White-dashed lines added for help in seeing bedding. Red lines are thrust faults. View looking east; ice flow left-to-right. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

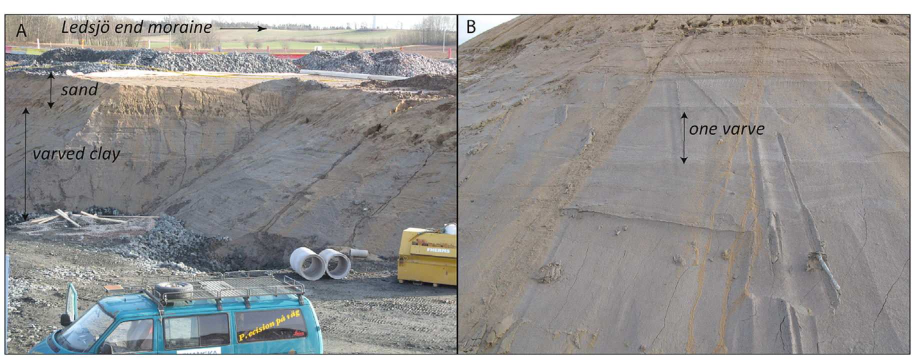

Fig. 12. Photos from Kollbogården. A. View to the southeast, Ledsjö moraine in the background, Kollbogården farm immediately to the left. Sand overlies varved clay. B. Undisturbed varves in the Kollbogården exposures; shovel handle 55 cm. The varves have a gray-up-to-red color change and at the top of the varve section, are up to 50 cm thick. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

for penecontemporaneous clay deposition and push-moraine formation. The cross section shows the actual topography, and the bedrock level is based on borehole information on file at the Swedish Geological Survey. The thickness of clay and sand are based on our own drill holes, but the shapes of the moraine-sediment bodies are hypothetical.

The sedimentology, structure and stratigraphy of the moraine ridges and the intermoraine flats indicates that glaciomarine clay was deposited continuously during overall ice-margin retreat through the MSEMZ and that oscillations (minor readvances) of the ice margin deformed the clay into the push moraines. Fig. 14 provides a simplified sketch of the sequence of events that took place to form the push moraines. For example, in order to produce the Skara moraine, the ice margin must have retreated north of the Skara position to allow for accumulation of proglacial, glaciomarine clay. The readvance pushed this recently deposited sediment into

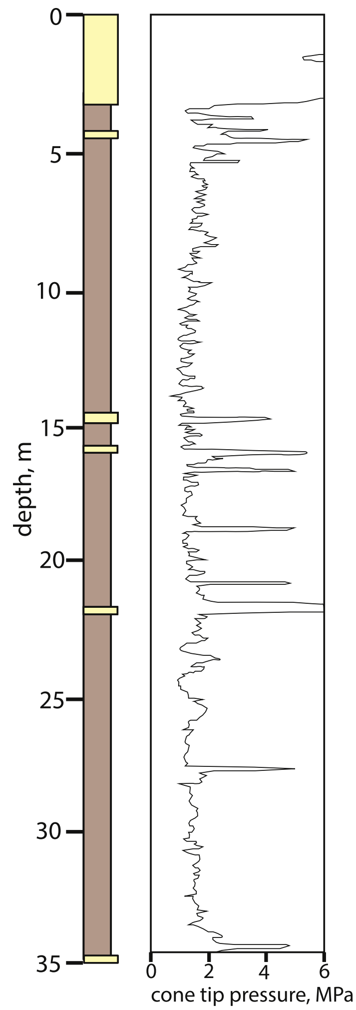

Fig. 13. Stratigraphy and cone-penetrometer measurements from Åängen (see Fig. 2 for location). The core is mostly clay (purple color) with sand beds (yellow). Stratigraphy predicted from cone-penetration test according to the Geotech 604D software package. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

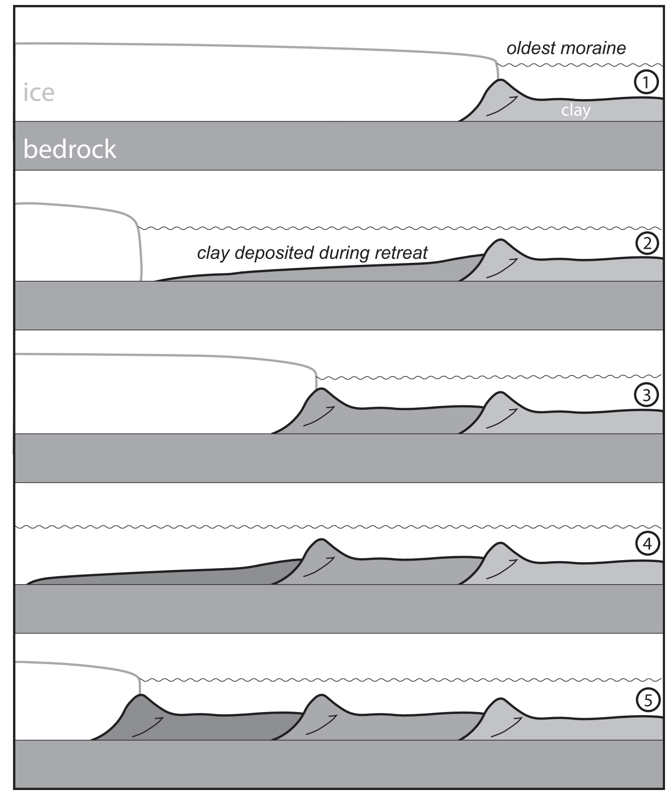

Fig. 14. Cartoon showing the sequence of events in push-moraine formation according to our conceptual model. Following the formation of the first moraine, ice retreats, and clay accumulates. The ice margin readvances and deforms the clay into a push moraine. This sequence is then repeated.

the ridge. Significantly, the bore hole at Mellemarken (Fig. 2) in the flat north of the Skara ridge has undisturbed varves overlying almost directly on bedrock (Johnson and Ståhl, 2010). The undisturbed nature of the varves almost directly overlying bedrock indicates four things; (1) the glacier was resting on bedrock at the Mellemarken site when the ice margin was at the Skara ridge, (2) during a readvance to the Skara ridge, the ice removed all sediment that had accumulated during the initial minor retreat, (3) the varves (and overlying sand) at Mellemarken was deposited only after the ice margin retreated from the Skara ridge to a position north of the Mellemarken location, and (4) sediment older than the sediment in the Skara ridge (older than the Younger Dryas) is not present in the core.

The same explanation is used for the stratigraphic relationships seen at Kollbogården, which lies between the Ledsjö and Uppsala ridges. Because ice has removed older sediment down to bedrock, the Ledsjö ridge must be made up of sediment deposited during a preceding interval when the ice margin lay farther north, followed by readvance and deformation to form the ridge. The core at Kollbogården also indicates that no sediment (or very little) lay between the ice bed and bedrock when the margin was at the Ledsjö ridge. The clay preserved in the Kollbogården core accumulated only after the retreat from the Ledsjö moraine.

The moraines consist of primarily of deformed clay with minor amounts of other sediment, but in Fig. 2 they are grouped together and referred to as 'moraine sediment.' The clay under the intermoraine flats is shown with colored lines indicating the orientation of stratification (varves). These beds are undisturbed save for those immediately in front of the next moraine to the north. In Fig. 2, a distinction is made between clay strata deposited prior to and after the adjacent ridge was formed, and this is indicated by a color difference. For example, varved clay deposited directly after the Skånings-Åsaka ridge was formed is shown in orange. South of the ridge it is undeformed and at the top of the stratigraphy (A in Fig. 2); north of the ridge, clay of the same age overlies bedrock and is deformed against the Kulla moraine sediment (A′ in Fig. 2).

The level of the sandy beds in the clay sequence in the Kollbogården core and in the Åängen CPT record mentioned above are shown as black dashed lines in Fig. 2, and these sandy layers are interpreted to have been deposited when the ice margin was located at the adjacent moraine to the north of each site (Uppsala and Ledsjö respectively). These sand beds are a stratigraphic marker signifying when the ice-margin was at the adjacent ice-margin position to the north, and the position of this layer is shown as a black-dashed line in Fig. 2. The varves above this marker bed were deposited after the ice retreated from the adjacent ice margin.

4.2. Sediment budget and the timing of moraine construction

Because the moraine ridges consist primarily of varved, glaciomarine clay and were formed when this clay was tectonized, and because the varve thickness yields a sedimentation rate, we can estimate the amount of time it took to create the ridges. That is, we can determine (1) the time it took for the ice to retreat north of each moraine position and then return, and (2) the distance the ice margin must have retreated prior to each readvance. Not only does this provide an indication of the time needed to construct such push moraines, it also serves as a simple test to see whether the amount of time represented in the sedimentary record broadly accords with the duration of the Younger Dryas.

To determine the time needed and the retreat distance, we estimated (1) the volume (or cross sectional area along a flow line) of the moraine sediment, (2) the sedimentation rate, (3) the ice-retreat rate and the (4) ice-advance rate. We realize that this exercise provides only a rough estimate of the time needed and therefore we provide a wide range of values for each parameter.

An estimate of the cross-sectional area of each moraine along an ice-flow path was made by calculating the cross-sectional area of each ridge above the level of the surrounding flats as a simple triangle (Fig. 15). The cross-sectional area is calculated along the dashed line shown in Fig. 2. Except for the Skara ridge, there is little significant variation in moraine width laterally in the study area, and we consider the widths along the flow-line representative. For the Skara ridge, further to the west, moraine ridges are likely buried, and further to the east, the ice-margin positions are difficult to trace across the in-part bedrock upland. Therefore, we consider the width along the dashed line appropriate for our calculations.

The area of the deformed ridge below the flats down to bedrock was estimated in three ways in order to get a reasonable range of cross-sectional area for each ridge. For the first, we extended the sides of the triangle to bedrock creating a large triangle cross section. For the second, we treated the ‘root’ to be an extension of the width of the ridge at the surface as the same width all the way to bedrock. And for the third, we assumed that the root would taper to the bedrock contact (Fig. 15). The first would give a large cross-sectional area, the third a small, and the second intermediate (Table 1). Although these three models may not represent the true or even realistic shape of the subsurface moraine-sediment body, we assume these geometries to bracket the true moraine cross-sectional area.

The sedimentation rate was determined using the two cores at Kollbogården and Mellmarken. In these two cores, the varve thickness ranges from 6 to 56 cm (Johnson and Ståhl, 2010), and the average varve thickness is 22 cm. We use 20 cm in our calculations. We note that the moraines contain some coarse material as well, and this means use of just the clay overestimates the amount of time needed to make each ridge. We also note that some clay is likely eroded and removed during ice advance, but we ignore this in the calculations.

Rates of ice-margin retreat were determined as follows. Björck et al. (2001), using Strömberg's varve chronology, show that the average ice-retreat rate through the MSEMZ on the east side of Billingen (Fig. 1), was about 15 m/year, which we calculated from

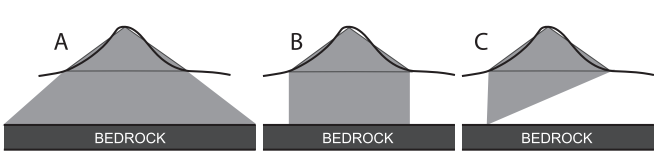

Fig. 15. Sketch showing how cross-sectional area was determined. Black line is the surface profile of a ridge. The area of a triangle representing the moraine was determined from the relief and width of the moraine. A maximum (A), intermediate (B) and minimum (C) cross-sectional area was determined by extending the sides of the triangle to bedrock (A), by projecting the width vertically downward (B) and by tapering the width to a point on bedrock (C).

their maps that display 100-year ice-recession lines. Strömberg (1977) stated that it took 650 years for the ice to retreat through the moraine belt east of Billingen, indicating a retreat rate of 18 m/yr. However, west of Billingen, the equivalent band of moraines is 2,5 to 3 times as wide, implying that the ice retreat rate was 2,5 to 3 times faster here, or up to 45 m/year. Additionally, these estimates are net ice-retreat rates and do not account for the oscillations shown here in the study area or those known to have occurred east of Billingen in the MSEMZ (Strömberg, 1977, 1994). Therefore, these estimates are less than the actual retreat rate. We therefore adopt ice-retreat rates using what we consider a low estimate of 45 m/year and a high of 75 m/year (for comparison, after the Younger Dryas, the retreat rate in the field area increased to 120 m/yr based on De Geer moraine spacing, assuming annual De Geer moraine formation (Bouvier et al., 2015; Ojala et al., 2015; Ojala, 2016)).

We have two independent checks that our ice-retreat rates are reasonable. For the first, the cross section at Gullhammar shows the buried De Geer moraine ridge, but also two others revealed by the draping pattern of the varved sediment (Fig. 11). The spacing between these is about 60 m, and, if annual, represent a retreat rate of 60 m/yr, which is well within the range we are applying. The second independent check of the retreat rate is provided by an interpretation of the Kollbogården varve core (Johnson and Ståhl, 2010). In Fig. 12 of Johnson and Ståhl (2010), they suggest, based on varve thickness and sedimentology, that it took 50–60 years for the ice to (1) retreat from Kollbogården, (2) retreat north of the Uppsala position, and then (3) readvance back to the Uppsala position, forming the ridge. Considering the distances necessary, the ice retreat and advance rate can be calculated to lie between 40 and 70 m/yr (see Table SM1 in supplementary material for supporting calculations).

Another source of uncertainty is the rate of ice sheet readvance during the oscillations. . Therefore, we calculate two scenarios; one in which the advance rate is the same as the retreat rate, and one in which the advance rate is ‘instantaneous’, or much faster in relation to the retreat. This latter is unrealistic and we do not believe the moraines are necessarily surge moraines (Johnson et al., 2013b), but this serves to put an upper limit of the retreat distance prior to the formation of each ridge.

Table 1 shows the values that we calculated for the duration and retreat distance required for each ridge to have formed (see Table SM2 in supplementary materials for a more complete table of calculations). To form the large Skara ridge, we estimate that the ice retreated 2–8 km, depending on the scenario. The small Gullhammar required only 0.3–0.5 km retreat prior to formation. We calculate that each ridge was formed in 15–130 years, depending on the ridge, and represents: (1) the time required to retreat north of the subsequent moraine position, during which time sediment accumulated in front of the ice, and (2) the time required for the ice margin to readvance to the moraine position. For the Skara ridge, 70–130 years is needed. The small Gullhammar ridge is estimated to need 15–24 years to form, which agrees with the ~20 years of undisturbed varves overlying the buried DGM and underlying the flowed clay deposited during the readvance (Fig. 11).

The total time required for the ridges to form according to our calculations, and the ice to finally retreat from the MSEMZ in the study area, ranges from about 350 to 800 years, a duration that fits within the time span of the Younger Dryas cold interval. Fig. 16 is a time-distance diagram based on one set of these calculations using a 60 m/yr retreat rate, an advance-rate that equals the retreat-rate, and a moraine area of type 2 (Fig. 15). Choosing the values shown in Fig. 16 as an example yields a result that lies in the middle of the range of our estimations.

However, neither our calculation nor Fig. 16 includes an estimate

Table 1
Estimation of moraine cross-sectional area, time of construction and oscillation retreat distances.

	Skara	Skånings-Åsaka	Kulla	Ledsjö	Uppsala	Flintås	Gullhammar	Total time of moraine construction[b]	Total time of moraine construction + retreat between moraines[c]
Moraine relief, m	19	19	19	20	9	9	8		
Moraine width, m	1950	450	650	400	250	250	250		
Depth to bedrock from base, m	20	25	25	35	30	20	5		
Area Max, m^2	78051	22926	33116	30250	21125	11681	2641		
Area Intermediate, m^2	57525	15525	22425	18000	8625	6125	2250		
Area Min, m^2	38025	9900	14300	11000	4875	3625	1625		
Time needed for construction, yr[a]	71–132	36–71	44–86	38–82	25–69	22–51	15–24	178–515	338–781
Distance of retreat needed prior to readvance, km	1.7–7.6	0.9–4.1	1.0–5.0	0.9–4.8	0.6–4.0	0.5–3.0	0.3–0.5		

[a] Time for initial retreat from given ice-margin position, clay sedimentation, and readvance to given ice-margin position. The result is based on a sedimentation rate of 20 cm/yr, ice-margin advance rates of 45–75 m/yr, the range of moraine areas depicted in Fig. 15, and ice-readvance rates either equal to the retreat rate or instantaneous, as described in the text.
[b] The time range is the sum of the time needed for each of the 7 moraines.
[c] The time range is the time of moraine construction plus the time of ice-margin retreat between the moraines using the range of 45–75 m/yr.

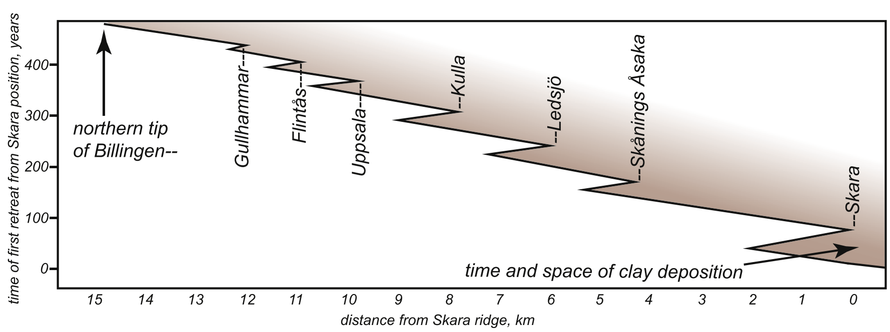

Fig. 16. Time-distance diagram of ice-margin behavior and clay sedimentation. The length of time for moraine formation by our calculations ranges from 350 to 800 years, but here the time scale is based on just one scenario with a 60 m/year ice-retreat rate, an intermediate moraine cross-sectional value (B in Fig. 15) and assuming the rate of ice-margin advance was equivalent to the rate of ice-margin retreat. These values produce a curve and ages that are in the middle of our range of estimates. See text for explanation.

for how much time the ice margin stayed at each ridge when the moraines were formed. It is possible that push moraines are made ‘instantly,’ as described for a push moraine at Bruarjökull, Iceland (Benediktsson et al., 2008), but it is more likely in the MSEMZ that the ice occupied each ice-margin position for a number of years prior to subsequent ice-margin retreat, which would add an unknown number of years to our estimates. For example, the crest section at Ledsjö (Fig. 8) shows a sequence of slightly to very deformed sand lenses that were deposited over time and deformed while the ice margin was present; it is not clear how much time is represented. Also at Ledsjö, up-ice of the moraine crest, a younger, less-deformed sequence records a second advance to the Ledsjö ice-margin position. A palinspastic reconstruction of these thrust-faulted sand beds at 190–240 m (Fig. 8) show a 29% shortening and, with a 45–70 m/year advance rate, indicates a deformation time of 34–66 days. If this duration is transferred to the entire moraine, two to four years are required. We think it is likely that this would represent a minimum duration for the ice-margin to be located at this moraine.

Another indication of the duration the ice stood at each IMP is the delta deposits that accumulated in the prominent triangular indentation of the Uppsala and Flintås moraines. Locally referred to as Ledsjömo (L in Fig. 1), this geomorphic feature is an outwash surface with braided fluvial channels and an ice-contact face at the northern edge. It is an ice-contact delta built up while the ice was at the Uppsala and later the Flintås IMPs. Similar ice-contact deltas in Norway and Sweden have been calculated to have formed in less than 70–100 years based on local ice-retreat rates (Tuttle et al., 1997; Plink-Björklund and Ronnert, 1999). However, these durations likely represent the maximum amount of time, and we consider it likely that the deltas were formed more quickly, but they still indicate occupation of these two IMPs for a period of time.

Our estimation of the time for sedimentation and push-moraine formation also cannot account for additional till and sorted sediment added to the moraines, nor does it account for sediment lost by erosion from the area (primarily suspended clay). Johnson and Ståhl (2010) argue that increased varve thickness at Kollbogården is due to erosion of clay during the advance to the Uppsala ice-margin position. Lastly, our estimate also does not include other potential moraines made by oscillations that were later completely buried by clay.

A comparison with earlier work on the MSEMZ shows that the estimated amount of time available for the moraines to have formed is similar to the duration we have estimated. For example, Strömberg (1977), based on varve chronology, measured the duration of the MSEMZ east of Billingen to be 650 years. A second estimate is based on the work of Björck and Digerfedt (1984, 1986, 1989). The duration of Greenland Stadial I (the event stratigraphic term for the Younger Dryas cold interval) on Greenland is 1150 years (Björck et al., 1998), and we assume that the length of the cold interval in western Sweden was essentially the same. However it is likely that the Skara moraine was formed well after the start of the GS1 and the Gullhammar well before its end, based on the work of Björck and Digerfeldt (1984, 1986, 1989). Björck and Digerfeldt (1989) assumed an Allerød ice-margin retreat to the north of Billingen and suggest that the ice re-closed the outlet at Billingen 120 years after the start of the Younger Dryas. Add to that an unknown number of years needed for the ice margin to reach the Skara position, and the Skara ridge may have formed 200–300 years after the start of the GS1.

Additionally, the youngest ridge, Gullhammar was formed before the end of the GS1. Based on our ice-retreat rates and the location of deposits of the final Baltic Ice Lake drainage (Johnson et al., 2010), the Gullhammar ridge was formed 50–150 years prior to the drainage, which itself is dated to have occurred 30–40 years prior to the end of GS1 (Andrén et al., 2002; Stroeven et al., 2015). If the Skara ridge formed 200–300 years after the start of GS1, and the Gullhammar ridge 80–200 years prior to the end of GS1, the seven moraines were made in 650–870 years.

Our estimate of 350–800 years agrees well with these previous estimates, especially if we add some additional years to account for time spent at each moraine. These calculations support our conceptual model of contemporaneous push-moraine and glaciomarine clay sedimentation. They indicate our oscillation model is reasonable and possible, and that our clay sedimentation rates are sufficient to account for the size/volume of the moraines.

4.3. Reconstruction of ice dynamics locally and along the southern margin of the Scandinavian Ice Sheet

Based on our observations and analysis, the SIS readvanced during the Younger Dryas cold event to form the ridges of the MSEMZ west of Billingen. The ice advanced first to the Skara moraine, and likely occupied this IMP for a long time as shown by its multiple ridges and large volume. The six other push moraines to the north indicate a dynamic ice-margin behavior during the subsequent 300–800 years with several ice-margin oscillations. There is no evidence in the field area of moraines being overridden during the oscillations.

The configuration of the MSEMZ in the study area is in many ways similar to the Melasveit area in western Iceland where Sigfúsdóttir et al. (2018) described a series of glaciotectonic moraines and undisturbed inter-moraine sediments formed by a marine-terminating glacier during overall retreat, most likely during the Younger Dryas. The results from these two different areas suggest that marine-terminating glaciers were more dynamic during the Younger Dryas than previously thought and raise questions about whether ice-margin oscillations across the North Atlantic region were climatically or dynamically driven during this cold interval.

Multiple end moraines mark the Younger Dryas zone elsewhere along the southern margin of the Scandinavian Ice Sheet. The Younger Dryas moraines in the Oslo area are strikingly similar to those in the study area, with the largest moraine, the Ra moraine, also at the southern edge of a band of multiple moraines formed in a marine setting, with the other five moraines (Ås 1, Ås 2, Ås 3, Ski 1 and Ski 2) consisting of smaller ridges formed during active retreat (Mangerud et al., 2011).

A geomorphic correlation of the moraines in our study area and Oslo is difficult because between the two sites, the number of ridges varies considerably. From the Norwegian border to the west side of the field area, only one to two ridges are traceable (Stroeven et al., 2016). Directly east of Billingen (Fig. 1), 10 to 15 ridges are present, but east of Vättern (Fig. 1), the number of moraines varies from 1 to 6 (Stroeven et al., 2016). This area east of Vättern also includes areas where no moraines exist at all, presumably due to the lack of sediment availability. Finally, in Finland the Younger Dryas zone is represented by the prominent Salpauseselkä moraines formed in the Baltic Ice Lake, but only the first two of these three ridges was formed during the Younger Dryas, with the third formed immediately following the Younger Dryas (Johansson et al., 2011).

Although there is evidence for climatic variation within the Younger Dryas (for example, Alley, 2000; Muscheler, R., et al., 2000; Baldini et al., 2015; Aichner et al., 2018), it is difficult to tie these variations directly to the end-moraine record of the SIS. We therefore argue here that the oscillations and ridges along the southern margin of the SIS cannot represent the result of a simple regional climate signal but rather indicate a complex response due to local conditions. These variables likely include relative difference in sea level, varying lake level, subglacial topography, sediment availability and ice dynamics on top of climate forcing.

4.4. Allerød-Younger Dryas ice-margin behavior and the early drainage of the Baltic Ice Lake

Because the outlet for the early drainage of the Baltic Ice Lake at the end of the Allerød interstadial has been proposed to be at Billingen (Björck and Möller, 1987; Björck, 1995; Muschitiello et al., 2015; Swärd et al., 2016) it is important to examine what is known about the ice-margin behavior prior to the Younger Dryas when this first drainage occurred.

How far did the ice retreat during the Allerød warm interval? It is apparent that the Scandinavian ice sheet responded differently to Allerød warming and Younger Dryas cooling in the different sectors of the ice sheet. In Norway, the Younger Dryas moraines were formed during a readvance (Mangerud et al., 2011; Hughes et al., 2016) following a significant retreat of the ice margin during the Alleröd; up to 40 km in western Norway (Mangerud et al., 2011) and 10 km in southern Norway (Romundset et al, 2019). But in western Sweden, Johansson (1982) concluded that the ice margin retreated only 2–4 km prior to Younger Dryas advance at Dals Ed, an area that lies about 100 km west northwest of this paper's study area. At the west edge of the field area, Dennegård (1984) documented overconsolidated clays that he interpreted to be due to an Allerød retreat and compaction during the Younger Dryas readvance, but he does not indicate how large this retreat was. East from Billingen to the Baltic, strong evidence for a significant Allerød ice-margin retreat is lacking (Lundqvist and Wohlfarth et al., 2008). Based on varve chronology, Strömberg (1994) concluded the ice margin at the end of the Allerød directly east of Billingen (at Svekhult, Sweden), was about 20 km south of the northern tip of Billingen. In eastern Sweden, varve correlations have produced ice-retreat estimations that indicate that the ice-margin only slowed during its retreat during the Younger Dryas, and that no post-Allerød readvance is apparent (Brunnberg, 1995). However, in Finland, the ice margin may have retreated up to 50 km north of the Salpausselkä I in southern Finland during the Alleröd as suggested by Rainio (1993) and Rainio et al. (1995). This retreat, called the Heinola deglaciation, is based on till-covered clays near Salpausselkä I and a change in striation orientations (Rainio et al., 1995) but also on geomorphic considerations (Eronen and Vesajoki, 1988). However, the Finnish part of the ice sheet was much more dynamic than in central Sweden and was broken into distinct lobes that at times acted independently (Johansson et al., 2011). For example, the activity of the Lake District Lobe, where the ice retreat of Rainio et al. (1995) occurred, would not be representative of the ice sheet as a whole (Eronen and Vesajoki, 1988).

An interpolation among the western and eastern Swedish sites mentioned above would imply little to no Allerød retreat in our field area. This conclusion agrees somewhat with our estimation of the amount of retreat prior to the readvance that formed the Skara ridge, which is on the order of 2–8 km (Table 1). However, the northern tip of Billingen is 15 km from the Skara moraine (Fig. 1), and the ice would had to have retreated that far in order to uncover the lower elevations necessary to allow the early drainage of the

Baltic Ice Lake. Based on our calculations as well as the field evidence cited above, the ice did not retreat enough to uncover the outlet at Billingen.

Nonetheless, the geologic evidence for the early drainage has been greatly strengthened in recent years. Notably, two of the five 'arguments against a drainage' listed by Björck (1995) have been falsified by recent work. Muschitiello et al. (2015) have shown a distinct change in varve character in Baltic Ice Lake varves at the time suggested to be the early drainage in the Baltic basin. They concluded that the change in character is due to a drainage at Billingen. Furthermore, Swärd et al. (2016) reported on a core from Lake Vättern where increased chlorine content in the clays are interpreted to represent a marine signal of incursion of Atlantic sea water from northern Billingen. This chlorine signal is also dated around 13 ka BP. Additionally, Bennike and Jensen (2013) reported the presence of terrestrial organic matter dated 12.7–13.1 cal ka BP in a core from the Arkona basin in the southern Baltic that requires a low stand of water in the Baltic basin at that time.

Based on these arguments, we conclude that the early Allerød drainage likely occurred, and the outlet, indeed, was at Billingen. The studies cited above shows little evidence for a regional ice-margin retreat that would have been sufficient to uncover northern Billingen. so we suggest, as a working model, a dynamic, local collapse of the ice margin to allow for the early drainage. This model suggests that at a critical stage of the retreating Allerød ice margin, a subglacial drainage began, driven by the head difference between the Baltic Ice Lake and the western sea (~10 m, Björck, 1995 or as much as 20 m, Bennike and Jensen, 2013)) leading to the collapse of the ice dam. This collapse may have been aided and localized by the bedrock trough that lies along the west edge of Billingen, where Valle Härad now lies (Figs. 1 and 3). Assuming that the ice during the late Weichselian had removed sediment in this trough, a calving-bay may have developed in this deep, bedrock valley. Subglacial drainage of ice lakes has been observed in Alaska and Iceland (Björnsson, 2002; Post and Mayo, 1971; Mayo, 1989), and an initial drainage under the Scandinavian ice sheet has been suggested to be a part of the character of the final drainage of the Baltic Ice Lake (Strömberg, 1977, 1992; Björck, 1995; Påsse and Pile, 2016). This scenario for the initial drainage also gives support to the suggestion by Björk and Digerfeldt (1984, 1986) that the thick subsurface sediment in Valle Härad (Fig. 1) was deposited during this event.

5. Conclusions

The seven prominent ridges that make up the Middle Swedish end-moraine zone west of Billingen are push moraines, and they were constructed during the Younger Dryas cold interval by ice-margin oscillations during overall retreat. The southern-most moraine, the Skara moraine, is the oldest, and the moraines to the north successively younger.

The material in the end moraines is primarily glaciomarine clay of Younger Dryas age that was deposited in the sea in front of the retreating glacier and then pushed glaciotectonically to form the end moraines during ice-margin oscillations. Each ice-margin oscillation allowed accommodation space for the accumulation of proglacial, marine clay during ice-margin retreat, and the clay was in-turn bulldozed forward during each minor readvance to form the push moraines.

Based on estimations of moraine volume, sedimentation rate and ice-margin retreat rates, the overall ice-margin retreat and end-moraine construction took 350–800 years, well within the time-span of the Younger Dryas.

Because the number of moraines in the Younger Dryas zone varies across the southern margin of the Scandinavian Ice Sheet, we regard the seven individual oscillations west of Billingen to be strongly affected by local factors. There likely are climate forcings influencing the ice-margin behavior along the southern margin of the Scandinavian Ice Sheet during the Younger Dryas interval, but these are difficult to separate from local controlling factors such as water depth, sea level, lake levels, sediment availability and bedrock topography.

Finally, we consider it likely that the Allerød drainage of the Baltic Ice Lake took place at Billingen and passed through the field area although we have found no field evidence for it., We suggest the drainage was associated with an ice-margin collapse initiated by subglacial drainage driven by the head difference between the Baltic Lake and the sea and likely centered on Valle Härad.

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Appendix A. Supplementary data

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References

Ahlmann, H.W., 1910. Studier öfver de medelsvenska ändmoränerna. Arkiv för kemi, mineralogi och geologi, utgivet av K. Svenska Vetenskapsakademien I Stockholm. Band 3 (No. 29), 15.
Aichner, B., Ott, F., Słowiński, M., Noryśkiewicz, A.M., Brauer, A., Sachse, D., 2018. Leaf wax n-alkane distributions record ecological changes during the Younger Dryas at Trzechowskie paleolake (northern Poland) without temporal delay. Clim. Past 14 (11), 1607–1624.
Alley, R.B., 2000. The Younger Dryas cold interval as viewed from central Greenland. Quat. Sci. Rev. 19 (1–5), 213–226.
Andrén, T., Lindeberg, G., Andrén, E., 2002. Evidence of the final drainage of the baltic ice lake and the brackish phase of the yoldia sea in glacial varves from the baltic sea. Boreas 31, 226–238.
Baldini, L.M., McDermott, F., Baldini, J.U., Arias, P., Cueto, M., Fairchild, I.J., Ontañón, R., 2015. Regional temperature, atmospheric circulation, and sea-ice variability within the Younger Dryas Event constrained using a speleothem from northern Iberia. Earth Planet. Sci. Lett. 419, 101–110.
Benediktsson, Í.Ö., Möller, P., Ingólfsson, Ó., van der Meer, J.J.M., Kjær, K.H., Johannes Kruger, J., 2008. Instantaneous end moraine and sediment wedge formation during the1890 glacier surge of Bruarjokull, Iceland. Quat. Sci. Rev. 27, 209–234.
Benediktsson, Í.Ö., Schomacker, A., Lokrantz, H., Ingólfsson, Ó., 2010. The 1890 surge end moraine at Eyjabakkajökull, Iceland: a re-assessment of a classic glaciotectonic locality. Quat. Sci. Rev. 29, 484–506.
Benediktsson, Í.Ö., Schomacker, A., Johnson, M.D., Geiger, A.J., Ingólfsson, Ó., Guðmundsdóttir, E.R., 2015. Architecture and structural evolution of an early Little Ice Age terminal moraine at the surge-type glacier Múlajökull, Iceland. J. Geophys. Res.: Earth Surface 120, 1895–1910.
Bennett, M.R., 2001. The morphology, structural evolution and significance of push moraines. Earth Sci. Rev. 53, 197–236.
Bennike, O., Jensen, J.B., 2013. A baltic ice lake lowstand of latest Allerød age in the Arkona basin, southern baltic sea. Geol. Surv. Den. Greenl. Bull. 28, 17–20.
Björck, J., Possnert, G., Schoning, K., 2001. Early Holocene deglaciation chronology in Västragötland and Närke, southern Sweden–biostratigraphy, clay varve, 14C and calendar year chronology. Quat. Sci. Rev. 20, 1309–1326.
Björck, S., 1995. A review of the history of the Baltic Sea, 13.0-8.0 ka BP. Quat. Int. 27, 19–40.
Björck, S., Digerfeldt, G., 1984. Climatic changes at Pleistocene/Holocene boundary in the Middle Swedish End moraine zone, mainly inferred from stratigraphic indications. Dordrecht, D. In: Mörner, N.-A., Karlén, W. (Eds.), Climatic Changes on a Yearly to Millennial Basis. Reidel Publishing Company, pp. 37–56.

Björck, S., Digerfeldt, G., 1986. Late Weichselian–Early Holocene shore displacement west of Mt. Billingen, within the Middle Swedish end-moraine zone. Boreas 15, 1–18.
Björck, S., Digerfeldt, G., 1989. Lake Mullsjön–a key site for the understanding the final stage of the Baltic Ice Lake east of Mt. Billingen. Boreas 18, 209–219.
Björck, S., Möller, P., 1987. Late Weichselian environmental history in southeastern Sweden during the deglaciation of the Scandinavian ice sheet. Quat. Res. 28 (1), 1–37.
Björck, S., Walker, M.J.C., Cwynar, L.C., Johnsen, S., Knudsen, K.-L., Lowe, J.J., Wohlfarth, B., INTIMATE Members, 1998. An event stratigraphy for the Last Termination in the North Atlantic region based on the Greenland Ice-core record: a proposal by the INTIMATE group. J. Quarternary Sci. 13, 283–292.
Björnsson, H., 2002. Subglacial lakes and jökulhlaups in Iceland. Glob. Planet. Chang. 35, 255–271.
Boulton, G.S., Van der Meer, J.J.M., Beets, D.J., Hart, J.K., Ruegg, G.H.J., 1999. The sedimentary and structural evolution of a recent push moraine complex: holmstrømbreen, Spitsbergen. Quat. Sci. Rev. 18, 339–371.
Bouvier, V., Johnson, M.D., Påsse, T., 2015. Distribution, genesis and annual-origin of De Geer moraines in Sweden: insights revealed by LiDAR. GFF 137, 319–333.
Brunnberg, L., 1995. Clay-varve Chronology and Deglaciation during the Younger Dryas and Preboreal in the Easternmost Part of the Middle Swedish Ice Marginal Zone. Doctoral dissertation. Stockholm university.
Carlsson, C., 2002. Bildandet Och Uppbyggnaden Av Ryggar I Mellansvenska Israndzonen Väster Om Billingen, Kullalinjen. University of Gothenburg thesis, p. B334.
Chamberlin, T.C., 1894. Studies for students: proposed genetic classification of Pleistocene glacial formations. J. Geol. 2 (5), 517–538.
Croot, D.G., 1988. Morphological, structural and mechanical analysis of neoglacial ice-pushed ridges in Iceland. In: Croot, D.G. (Ed.), Glaciotectonics: Forms and Processes. Balkema, Rotterdam, pp. 33–47.
Dennegård, B., 1984. Late Weichselian and Early Holocene Stratigraphy in Southwestern Sweden with Emphasis on the Lake Vänern Area. Ph.D.-thesis. Chalmers University of Technology/University of Gothenburg, Department of Geology A48, p. 187.
Eronen, M., Vesajoki, H., 1988. Deglaciation pattern indicated by the ice-margin formations in Northern Karelia, eastern Finland. Boreas 17, 317–327.
Frödin, G., 1916. Über einige spätglaziale Kalbungsbuchten und fluvioglaziale Estuarien un mittleren Schweden. Bull. Geol. Inst. Univ. Upps 19, 159–174.
Hughes, A.L.C., Gyllenkreutz, R., Lohne, Ø.S., Mangerud, J., Svendsenm, J.I., 2016. The last Eurasian Ice Sheets - a chronological database and time-slice reconstruction, DATED-1. Boreas 45, 1–45.
Johansson, B.T., 1982. Deglaciationen av norra Bohuslän och södra Dalsland. Geologiska Institutionen, Chalmers Tekniska Högskola och Göteborgs Universitet, Dissertation, Publ. A 38, 180.
Johansson, P., Lunkka, J.P., Sarala, P., 2011. The glaciation of Finland. In: Developments in Quaternary Sciences, vol. 15. Elsevier, pp. 105–116.
Johansson, S., 1926. Den Baltiska issjöns tappning. GFF (Geol. Foren. Stockh. Forh.) 48, 186–263.
Johansson, S., 1937. Senglaciala och interglaciala avlagringar vid ändmoränstråket i Vätsergötland. GFF (Geol. Foren. Stockh. Forh.) 59, 379–454.
Johnson, M.D., Ståhl, Y., 2008. Sedimentology, Stratigraphy, Age, and Paleoecology of Marine Varved Clay in the Middle Swedish End-Moraine Zone-Aa Final Report to the Swedish Geological Survey (SGU). Earth Sciences Centre, C84. Göteborg University, p. 54.
Johnson, M.D., Ståhl, Y., 2010. Stratigraphy, sedimentology, age and palaeoenvironment of marine varved clay in the Middle Swedish end-moraine zone. Boreas 39, 199–215.
Johnson, M.D., Ståhl, Y., Larsson, O., Seger, S., 2010. New exposures of baltic ice lake drainage sediments, Götene, Sweden. GFF 132, 1–12.
Johnson, M.D., Kylander, M.E., Casserstedt, L., Wiborgh, H., Björck, S., 2013a. Varved glaciomarine clay in central Sweden before and after the Baltic Ice Lake drainage: a further clue to the drainage events at Mt Billingen. GFF 135, 293–307.
Johnson, M.D., Benediktsson, Í.Ö., Björklund, L., 2013b. The Ledsjö end moraine—a subaquatic push moraine composed of glaciomarine clay in central Sweden. Proc. Geologists' Assoc. 124, 738–752.
Kristensen, L., Benn, D.I., Hormes, A., Otesen, D., 2009. Mud aprons in front of Svalbard surge moraines: evidence of subglacial deforimng layers or proglacial glaciotectonics? Geomorphology 111, 206–221.
Lidmar-Bergström, K., 1996. Long term morphotectonic evolution in Sweden. Geomorphology 16, 33–59.
Loke, M.H., 1999. Electrical Imaging Surveys for Environmental and Engineering Studies. A Practical Guide to 2-D and 3-D Surveys (Published by the author).
Lowrie, W., 2007. Fundamentals of Geophysics. Cambridge University press.
Lundqvist, G., Högbom, A., Westergård, A.H., 1931. Beskrivning till kartbladet lugnås. Sveriges Geologiska Unbdersökning Ser. Aa No. 172, 185.
Mangerud, J., Gyllenkreutz, R., Lohne, O.S., Svendsen, J.I., 2011. Glacial history of Norway. Dev. Quat. Sci. 15, 279–298.
Mayo, L.R., 1989. Advance of hubbard glacier and 1986 outburst of Russell Fiord, Alaska, USA. Ann. Glaciol. 13, 189–194.
Munthe, H., 1910. Studies in the Late-Quaternary history of southern Sweden. Geol. Foren. Stockh. Forh. 32, 1197–1293.
Munthe, H., Westergård, A.H., Lundqvist, G., 1928. Beskrivning till kartbladet skövde. Sveriges Geologiska Undersökning, Series Aa 121, 182.
Muscheler, R., Beer, J., Wagner, G., Finkel, R.C., 2000. Changes in deep-water formation during the Younger Dryas event inferred from 10 Be and 14 C records. Nature 408 (6812), 567.
Muschitiello, F., Lea, J.M., Greenwood, S.L., Nick, F.M., Brunnberg, L., MacLeod, A., Wohlfarth, B., 2015. Timing of the first drainage of the baltic ice lake synchronous with the onset of Greenland stadial 1. Boreas 45, 322–334.
Mussett, A.E., Khan, M.A., 2000. Looking into the Earth: an Introduction to Geological Geophysics. Cambridge University Press.
Ojala, A.E., Putkinen, N., Palmu, J.P., Nenonen, K., 2015. Characterization of de geer moraines in Finland based on LiDAR DEM mapping. GFF 137, 304–318.
Ojala, A.E., 2016. Appearance of De Geer moraines in southern and western Finland—implications for reconstructing glacier retreat dynamics. Geomorphology 255, 16–25.
Påsse, T., 2006. Jordartskarta Skara NV. Sveriges Geologiska Undersöknings Databas.
Påsse, T., Pile, O., 2016. Beskrivning till jordartskartorna 8D Skara NV, NO, SV och SO och 9D Mareistad SV. Sveriges Geologiska Undersökning 534–538, 132.
Plink-Björklund, P., Ronnert, l., 1999. Depositional processes and internal architecture of Late Weichselian ice-margin submarine fan and delta settings, Swedish west coast. Sedimentology 46, 215–234.
Post, A., Mayo, L.R., 1971. Glacier Dammed Lakes and Outburst Floods in Alaska. US Geological Survey Hydrologic investigations atlas HA-4455, 10 p, Washington, DC, p. 10.
Powell, R., 1990. Glaci0marine Processes at Grounding-Line Fans and Their Growth to Ice-Contact Deltas, vol. 1990. Geological Society, London, Special Publications, pp. 53–73, 53.
Rainio, H., 1993. The Heinola deglaciation and Salpausselkä readvance as recorded in the lithostratigraphy of the distal area of Salpausselkä I at Ihalainen, Lappeenranta, Finland. Geol. Surv. Finl. Spec. Pap. 18, 53–62.
Rainio, H., Saarnisto, M., Ekman, I., 1995. Younger Dryas end moraines in Finland and NW Russia. Quat. Int. 28, 179–192.
Romundset, A., Lakeman, T.R., Høgaas, F., 2019. Coastal lake records add constraints to the age and magnitude of the Younger Dryas ice-front oscillation along the Skagerrak coastline in southern Norway. J. Quat. Sci. 34 (2), 112–124.
Seger, S., 2011. A 3d GIS Study of Stratigraphy and Late Stage Quaternary Development in the Skara-Götene Area, South-Central Sweden. University of Gothenburg thesis. Department of Earth Sciences, p. B648.
Sigfúsdóttir, T., Benediktsson, Í.Ö., Phillips, E., 2018. Active retreat of a Late Weichselian marine-terminating glacier: an example from Melasveit, western Iceland. Boreas 47, 813–836.
Stroeven, A.P., Heyman, J., Fabel, D., Björck, S., Caffee, M.W., Fredin, O., Harbor, J.M., 2015. A new Scandinavian reference 10 Be production rate. Quat. Geochronol. 29, 104–115.
Stroeven, A.P., Hättestrand, C., Kleman, J., Heyman, J., Fabel, D., Fredin, O., Caffee, M.W., 2016. Deglaciation of fennoscandia. Quat. Sci. Rev. 157, 91–121.
Strömberg, B., 1969. Den mellansvenska israndzonen. Stockholms Universitet. Naturgeografiska Institutionen Forskningsrapport 6, 56.
Strömberg, B., 1977. Deglaciationen vid Billingen och Baltiska issjöns tappning. GFF 99, 92–95.
Strömberg, B., 1992. The Final Stage of the Baltic Ice Lake, vol. 81. Sveriges Geologiska Undersökning, Series Ca, pp. 347–353.
Strömberg, B., 1994. Younger Dryas deglaciation at Mt. Billingen, and clay varve dating of the Younger Dryas/Preboreal transition. Boreas 23, 177–193.
Swärd, H., O'Regan, M., Ampel, L., Ananyev, R., Chernykh, D., Floden, T., Greenwood, S.L., Kylander, M.E., Mörth, C.M., Preto, P., Jakobsson, M., 2016. Regional deglaciation and postglacial lake development as reflected in a 74 m sedimentary record from Lake Vättern, southern Sweden. GFF 138 (2), 336–354.
Tuttle, K.J., Øustmo, S.R., Andersen, B.Ö.G., 1997. Quantitative study of the distributary braidplain of the Preboreal ice-contact Gardermoen delta complex, southeastern Norway. Boreas 26, 151–156.
Wohlfarth, B., Bjorck, S., Funder, S., Houmark-Nielsen, M., Ingólfsson, Ó., Lunkka, J.P., Mangerud, J., Saarnisto, M., Vorren, T., 2008. Quaternary of norden. Episodes 31, 73–81.