Glaciodynamics, Deglacial Landforms and Isostatic Uplift during the last Deglaciation of Norrbotten, Sweden

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Mattias Lindén

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Glaciodynamics, deglacial landforms and isostatic uplift during the last deglaciation of Norrbotten, Sweden

Abstract

The aim of this thesis was to reconstruct the glaciodynamics, deglacial landforms, isostatic uplift and to date the deglaciation. Glaciodynamics and deglacial landforms were focused on to provide a process and depositional model for De Geer moraine and Niemisel moraine and to reveal their internal and spatial relationship, based on detailed sedimentological and structural investigations. The isostatic uplift and deglacial chronology was reconstructed from age measurements on the initial organic production in two lakes above the highest shoreline and isolation of lake basins at different altitude below the highest shoreline.

The results show that the deglaciation of Norrbotten occurred earlier than previously thought, at 10 500 cal. yr BP, which implies that the deglaciation after the Younger Dryas re-advance, from the Skövde-Billingen and Salpausselkä moraines, was more rapid than previously thought. Within c. 1 000 years the ice sheet retreated c. 600 km, or in the order of 600 m/yr given an even recessional rate. This rapid ice sheet retreat is supported by the glaciodynamic conditions, i.e. the deforming bed. As the margins of the late-glacial Scandinavian ice sheet became wet-based, ice-flow velocities increased resulting in thinning of the ice. As the ice margin was subaqueous, situated in the Gulf of Bothnia, calving was enhanced promoting the rapid deglaciation.

The De Geer moraine ridges was formed due to subglacial sediment advection to the ice margin during temporary halts in grounding-line retreat, forming gradually thickening sediment wedges. The proximal part of the moraines were built up in submarginal position through continuous stacked sequences of deforming bed diamictons, intercalated with glaciofluvial canal-infill sediments, whereas the distal parts were built up from the grounding line by prograding sediment gravity-flow deposits, distally interfingering with glaciolacustrine sediments.

The Niemisel moraine ridges was formed due to subglacial folding/thrust stacking of pre-deposited sediments contemporaneously with lee-side cavity deposition, forming vertically and distally prograding moraine ridges transverse to ice-flow. The proximal part of the moraines were built up by subglacial folding and thrust stacking sequences of pre-ridge formation sediments, whereas the distal parts were built up by glaciofluvially and gravity-flow deposited sediment in lee-side cavities.

Key words: glaciodynamics, deglaciation, isostatic uplift, De Geer moraine, Niemisel moraine, ribbed moraine, deforming bed, thrust stacking, glaciotectonic

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by

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This thesis is based on three papers listed below as Appendix I-III. The papers are reproduced with permission from John Wiley & Sons Limited (Appendix II) and Taylor & Francis AS (Appendix III).

Appendix I: Lindén, M., Möller, P. and Adrielsson, L. manuscript: Niemisel moraines: ribbed moraine formed by subglacial folding, thrust stacking and lee-side cavity infill. Submitted to *Quaternary Science Reviews* (2005-12-08).

Appendix II: Lindén, M. and Möller, P. 2005: Marginal formation of De Geer moraines and their implications to the dynamics of grounding-line recession. *Journal of Quaternary Science 20*, 113-133.

Appendix III: Lindén, M., Möller, P., Björck, S. and Sandgren, P. (2006): Holocene shore displacement and deglaciation chronology in Norrbotten, Sweden. *Boreas 35,* 1-22.

Introduction

Three quarters of the surface of Sweden is covered by till (Fredén, 1994). Most of this area consists of cover moraine, draping the underlying bedrock. Occasionally the till forms distinct depositional landforms such as transverse moraine ridges, oriented perpendicularly to the former ice flow (e.g. *De Geer* and *Niemisel moraines*), and longitudinal moraines, parallel to the ice flow (drumlins etc), but also hummocky moraines with no preferred orientation. Investigations of glacial sediments and landforms are crucial for the understanding and reconstruction of the glacial history of Sweden. By studying the glacial sediments it is possible to reconstruct the formation of a landform, whether formed in a marginal, subglacial or supraglacial position, by accumulation, erosion or deformation, or a combination of these. It is even possible to bring the understanding one step further by reconstructing the glaciodynamic and hydrologic conditions during the formation of the landform.

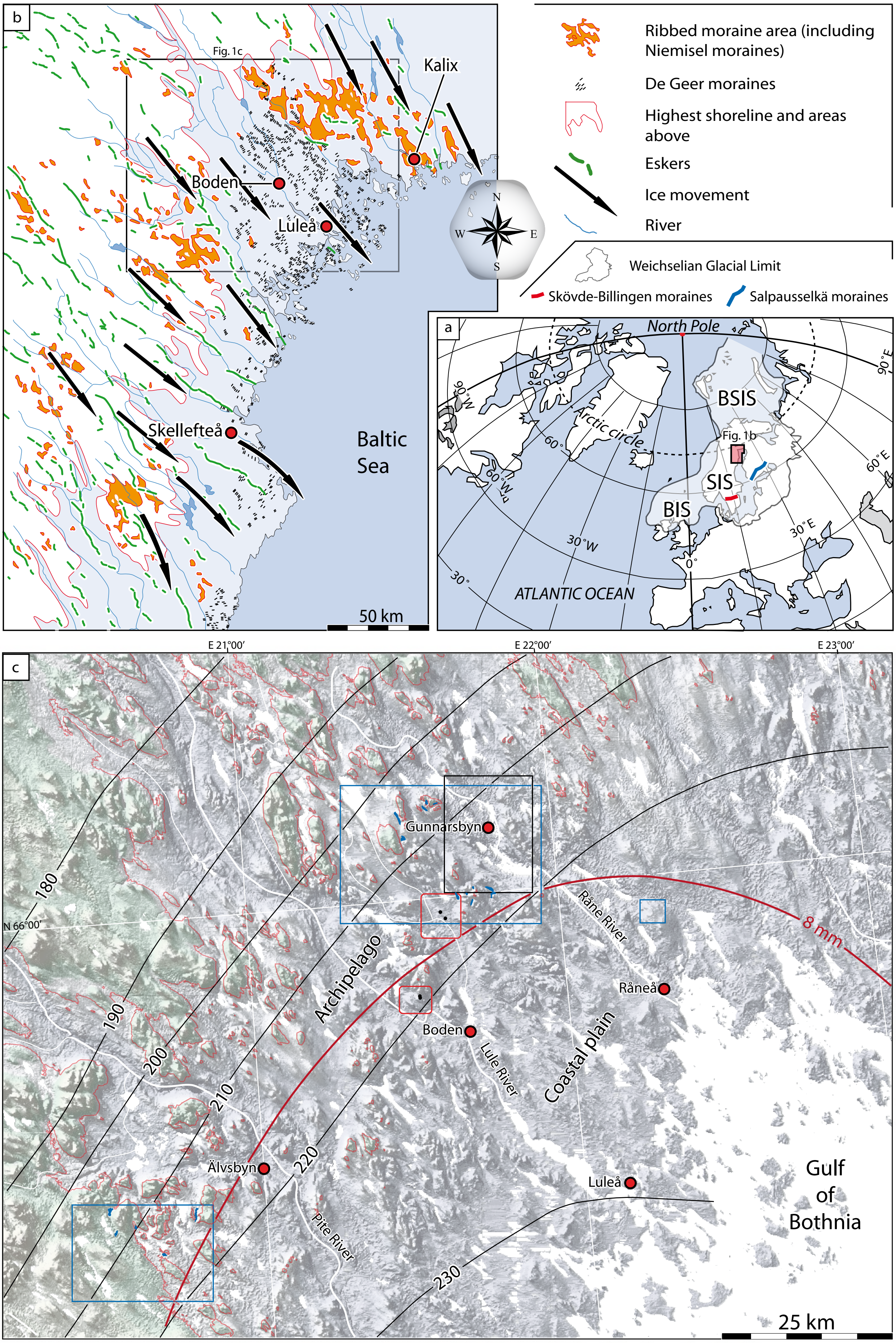
A rapid decay of the last Scandinavian Ice Sheet succeeded the Younger Dryas cold-event re-advance from the Skövde-Billingen Moraines in Sweden (e.g. J. Lundqvist and Wohlfarth, 2001; Fig. 1a) and the Salpausselkä Moraines in Finland (e.g. Donner, 1995; Fig. 1a). Along a flow-line from the Salpausselkä Moraines towards the last ice remnants in northern Sweden, the ice margin retreated 600 km within c. 1000 years, 11 500-10 500 cal yr BP (Berglund, 2004; Lindén *et al.*, 2006), equal to a mean recession rate of c. 600 m/yr. Climate probably provided a strongly negative mass balance on the ice sheet driving this rapid deglaciation (Siegert and Dowdeswell, 2002). Glaciodynamics and the deglacial environment were also important factors; longitudinal ice sheet lineation patterns north of the Younger Dryas Moraines indicate the existence of major ice-streams over Finland and the Gulf of Bothnia during ice sheet decay (Boulton *et al.*, 2001), enhancing the evacuation of large ice masses. The ice margin was also situated in deep water (up to c. 330 m in the deepest part of the Bothnian Bay), promoting a retreat dominantly forced by calving. It is, however, unknown whether climate or glacier physics was the main trigger behind this rapid decay of the large ice volume.

It is in this context questions on the deglacial paleo-environment and glaciodynamics arise. The rapid deglaciation of coastal Norrbotten, northern Sweden (Fig. 1), left vast areas of landforms such as *De Geer moraines* and *Niemisel moraines* as archives of glaciodynamic conditions. The emphasis in this thesis is on *De Geer moraines* (Hoppe, 1959; in North America wash-board or cross-valley moraines, e.g. Mawdsley, 1936; Norman, 1938; Elson, 1957; Andrews, 1963a, 1963b; Prest, 1968) and *Niemisel moraines* (Beskow, 1935; Hoppe, 1948, 1959; Fromm, 1965; J. Lundqvist, 1981), two types of transverse moraines often abundant in areas with a subaqueous environment during the deglaciation. In this thesis special focus is on the formation of these landforms, with a main aim to reconstruct the sedimentological and landforming processes, and from this deduce the deglaciation pattern and glaciodynamics of the area. This work aims to reconstruct and understand the rapid decay of the Scandinavian Ice Sheet.

Through dating the isolation of a number of lake basins, i.e. when a basin was uplifted above the surface of the Ancylus Lake and became isolated, the shore displacement and the timing of deglaciation can be reconstructed. The chronology of the shore displacement is constructed by ^{14}C dating of identified isolation levels in lake sediments from lake basins situated at gradually lower altitudes below the highest shoreline. The isolation ages of the highest situated basins, complemented by dating of the first organic sedimentation in two lakes slightly above the highest shoreline, address the deglaciation age issue and form age control on the present clay varve chronology of the area. Thirteen OSL-dated beach ridges and terraces, in a sequence of 36 (at Heden, Figs. 1 and 2f-g), are used to independently test parts of the chronology of the lake-isolation based shore displacement curve.

The detailed sedimentological and structural investigations of *De Geer* and *Niemisel moraines* are combined with geomorphologic expression and spatial distribution patterns, and form the basis for process/facies models of *Niemisel* and *De Geer moraine* formation and the ages of lake basin isolation form the basis for establishing a shore displacement curve and dating the deglaciation in Norrbotten. The proposed process/facies models for *De Geer* and *Niemisel moraines* in Norrbotten have wider implications for the understanding of formation of moraine ridges in similar environments in Scandinavia and North America.

Fig. 1. (a) Overview map showing coastal Norrbotten in relation to the Late Weichselian glaciation maximum over Fennoscandia. Ice Sheet boundaries according to Svendsen et al. (1999). BIS = British Ice Sheet; SIS = Scandinavian Ice Sheet; BSIS = Barents Sea Ice Sheet. (b) Spatial distribution of glacial landforms and last ice-flow directions along the coast of Norrbotten, northern Sweden (modified from Hättestrand, 1997b). The position of the highest coastline is from J. Lundqvist (1994). (c) DEM-based map of coastal Norrbotten showing the spatial distribution of the highest shoreline (thin red line), the highest shoreline isobases (black lines, J. Lundqvist 1994), present day isostatic uplift isobase (thick red line), and areas above and below the highest shoreline (green and grey, respectively). The regional distribution of Niemisel moraine can be seen as transverse elements in all major River valleys. Elevation and highest shoreline data was provided by the Swedish Geological Survey (SGU). Frames mark the investigation areas of Niemisel moraine (black), De Geer moraine (red) and shoreline studies (blue).


b
Fig. 1c
Kalix
Boden
Luleå
Skellefteå
Baltic
Sea
50 km
N
W
E
S
Ribbed moraine area (including Niemisel moraines)
De Geer moraines
Highest shoreline and areas above
Eskers
Ice movement
River
Weichselian Glacial Limit
Skövde-Billingen moraines
Salpausselkä moraines
a
North Pole
90°E
90°W
BSIS
Arctic circle
60°
60°W
Fig. 1b
60°E
SIS
BIS
30°W
30°E
0°
30°
ATLANTIC OCEAN
c
E 21°00'
E 22°00'
E 23°00'
N 66°00'
180
190
200
210
220
230
Gunnarsbyn
Råne River
8 mm
Archipelago
Råneå
Boden
Lule River
Coastal plain
Älvsbyn
Pite River
Luleå
Gulf
of
Bothnia
25 km

Research rationale

De Geer and *Niemisel moraines* (i.e. ribbed moraines) have generated great attention during the years, resulting in a large number of investigations in order to reveal their depositional history. In spite of this, the active processes responsible for landform generation seem to be poorly understood. The issue on *De Geer* and *Niemisel moraine* formation in Norrbotten is a complex of sedimentological, morphological, environmental and chronological problems. Beside this there are theoretical considerations of the physical conditions related to glaciodynamics and subglacial hydrology. The aim of my studies in coastal Norrbotten was: a) to provide a process and depositional model for *De Geer moraine* and *Niemisel moraine* and to reveal their internal and spatial relationship, based on detailed sedimentological and structural investigations; b) to provide age control of the deglaciation, based on age measurements on the initial organic production in two lakes above the highest shoreline and two lake basins below, and to provide an isostatic uplift curve for the area, based of age measurements of isolated lake basins at different altitude.

Previous studies of internal composition of *Niemisel moraine* (e.g. Hoppe, 1948, 1959; Fromm, 1965) were conducted in the once numerously occurring 'gravels pits', opened up to compensate for the regional lack of coarse glaciofluvial deposits, e.g. eskers. The gravel pits where typically located in the distal part of the ridge where sorted sediments predominate. This is possibly the basis for the concept that *Niemisel moraine* dominantly consist of sorted sediment, and that previous depositional models have (glacio)fluvial activity as the primary sediment source and landform builder. Previous studies on *De Geer moraines*, especially the old ones (Hoppe, 1948, 1959; H. Möller, 1962; Bergström, 1968; Zillacus, 1987, 1989) were focused on spatial distribution, morphologic expression and occasional test pits. Only few *De Geer moraine* investigations were based on sedimentological and structural studies (Sollid and Carlson, 1984; Beaudry and Prichonnet, 1991, 1995; Larsen *et al.*, 1991; Blake, 2000) With this background our investigation strategy is to focus on the total picture of sediment facies and structural variation and to obtain continuous cross-sections from the proximal to the distal side of *Niemisel moraines* (five) and *De Geer moraines* (four). Equally important, in the *Niemisel moraine* context, is to get information from the whole form spectra, i.e. both from distinctly transverse ridges and from the smaller-scale irregular forms. For practical reasons the highest ridges were excluded as it were more important to excavate the whole ridges than just 'scratch the surface'.

The first aim of this thesis is to identify and distinguish the processes involved in the formation of *De Geer* and *Niemisel moraine* and to put up depositional models. The second aim is to relate the two depositional models to each other and to establish the internal relationship between these two types of moraine ridges, with special attention to environment, time and space. The final aim is to put the landform system in a broader regional context and to reconstruct the deglaciation history, the glaciodynamics and the isostatic uplift.

Geological setting of the investigation area – regional introduction

The large-scale geomorphology of the investigation area is characterised by southeast-trending broad valleys and bedrock hills, the latter reaching altitudes of 600 m a.s.l. (Fig. 1). Streamlined bedrock knobs, drumlins and lee-side moraines, all preferentially on higher ground, indicate a predominating ice movement towards southeast, possibly repeated through several glacial cycles (Fromm, 1965; J. Lundqvist, 1981; Hättestrand, 1997b; Kleman, 1990, 1992; Kleman *et al.*, 1997).

Valleys are predominantly occupied by till and glaciolacustrine/lacustrine silt, while glaciofluvial sediments seem to have a restricted occurrence. Often the thickness of glaciolacustrine deposits increases towards the present coast as the valleys become wider. Most of the till below the highest shoreline is represented by the local "Kalixpinnmo" till, most typical seen as bimodal, poorly sorted sediment predominated by fine sand with dispersed coarse clasts (Beskow, 1935; Fromm, 1965; Hoppe, 1948, 1952, 1959; J. Lundqvist, 1981). Till surfaces are significantly wave washed below the highest shoreline (200-220 m a.s.l.) and beach sediments are frequently occurring on slopes exposed to wave action during regression.

Moraine landforms are abundant in low-lying terrain, especially *De Geer moraines*, but also large, continuous areas of *Niemisel moraines* (Hoppe, 1948; Fromm, 1965; Figs. 2 and 3). *De Geer* and *Niemisel moraine* ridges are usually spatially separated, but in some areas these moraines occur together and *De Geer moraine* ridges are at several places seen to superimpose *Niemisel moraine* ridges (Figs. 2c and 3).

Niemisel moraines come in a large size spectrum, from small hummocks 50-100 m in length and width and thus with planforms hardly relatable to any paleo-iceflow, to large moraine ridges up to 30 m high, 200-300 m wide and 1000-1500 m long, all with well defined ridge axes being dominantly orientated perpendicular to the former ice flow direction (Fig. 2a-b). However, it is observed that

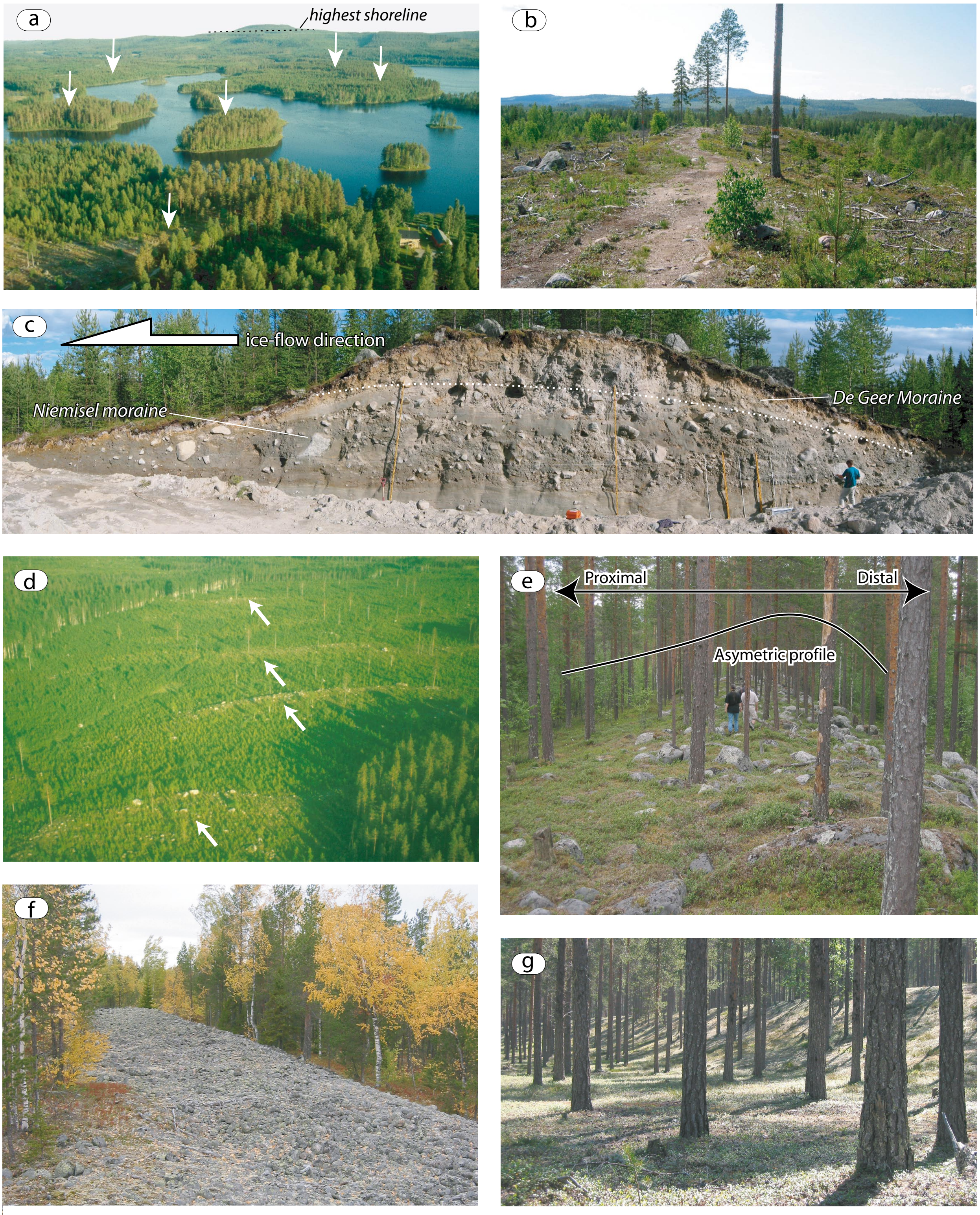


Fig. 2. (a) Morphology of Niemisel moraines at the eastern end of Lake Degerselet oblique aerial view from NNW. White arrows mark positions of individual Niemisel ridges. The height in the far background is a till-capped hill (in Swedish: kalottberg, in this case mount Snöberget demarcated with red circle at the map base, Fig. 3b), an island in the deglacial archipelago, indicated with preserved till above the highest shoreline, and with wave-washed bare bedrock and beach terraces below. (b) Longitudinal (SW) view of a Niemisel ridge. (c) Cross-sectional view at the Degerselet site, showing a De Geer moraine on top of a 90 m wide Niemisel ridge. (d) Oblique aerial photograph showing four narrow and slightly curved De Geer moraine ridges marked by arrows, also indicating the ice-flow direction. (e) Typical De Geer moraine with asymmetric cross profile and boulder concentration. Ice-flow was from left to right. (f) Cobble beach terrace close to the summit of the Sandkölen hill (c. 145 m a.s.l., Heden. (g) Sandy beach terraces (c. 70 m a.s.l.) at Heden.

sets of moraine hummocks in an, at first sight, irregular planform pattern often are aligned along iceflow-transverse axes. No consistent occurrence of downflow-oriented ridge 'horns' like in Rogen moraines (e.g. J. Lundqvist, 1969; Hättestrand, 1997a; P. Möller, in press) can be seen, though an irregular, both up- and down-flow bent curvature frequently occur (Fig. 3). The dominant moraine ridge-axis orientation is NE-SW, i.e. perpendicular to the regional paleo-iceflow direction as indicated by streamlined rock spurs, striae and drumlins. However, when the dominant NW-SE valley axis direction locally changes, then the orientation of *Niemisel moraine* ridges usually also changes to become transverse to the changed valley axis. This implies that the transverse pattern of the moraines is not related to the regional paleo-ice flow but rather to a near-marginal topographically constrained ice flow.

De Geer moraines are found below the highest shoreline of the last glacial event and extend well below the present sea level in the Gulf of Bothnia. Thus *De Geer moraines* are abundant in a 30-100 km wide zone along the coast of Norrbotten (Fig. 1b). In general, the height of *De Geer moraines* in Norrbotten varies between 1-3 m, although heights up to 6-7 m have been recorded (Hoppe, 1948; Fromm, 1965). The average width is c. 30 m and the length is highly variable; some ridges are only 100 m long while others can be traced continuously over distances of 2-3 km. The cross-sectional shape varies, but most commonly shows a gently sloping proximal side and a somewhat steeper distal side (Fig. 2e). The typical morphology is therefore a rather low, narrow and elongated moraine ridge with variable length (Fig. 2d). The distance between adjacent ridges is usually between 50 and 200 m, occasionally up to 500 m. However, the average distance is c. 100-150 m. With respect to ice-flow direction *De Geer moraine* tracts form a transverse lineation pattern being slightly up-flow concave in valleys (embayed) and topographic lows and slightly convex over, or close to, elevated ground (Fig. 3; see also Fig. 3 in Appendix II). Even though the frequency of *De Geer moraines* is smaller on higher ground and close to the highest shoreline they are often found on lower topographic heights and half-way up on valley sides.

The accepted concept for northern Sweden is that, prior to deglaciation, the ice divide moved from the Bothnian basin to the mountain region, from east to west (e.g. J. Lundqvist, 1994; Kleman *et al.*, 1997). The Ancylus fresh-water Lake gradually inundated the area as the ice margin retreated westwards and most of the investigated area was below water at deglaciation. However, the most elevated parts of the terrain formed an archipelago at the time of deglaciation with island and peninsulas along which the highest shoreline was developed at c. 200-230 m a.s.l. (Fig. 1). The highest shoreline is well developed on the higher bedrock knobs, sometimes resulting in till-capped hills (Fig. 2a). The altitude of the highest shoreline decreases from the present coast and further inland due to higher isostatic uplift within the Bothnian Bay, where the ice sheet was at its thickest during the maximum glaciation (e.g. J. Lundqvist, 1994; Ekman, 1996). A secondary cause is that the landscape already had begun to rise faster than the sea level rise when the ice margin retreated across the area, gradually exposing new areas to the Bothnian Bay and subsequent wave action.

Methods

Morphological mapping of moraines was conducted by means of aerial photographic interpretation. Most of the aerial photographs used were grey monochromatic photographs at a 1:30.000 scale. If available, false-coloured infrared photographs (1:60.000 scale) were also used in order to cover larger areas. Partly, the mapping was done as a supplement to the sediment/landform maps at a scale of 1:50.000, published by the Swedish Geological Survey (Dahlberg, 1993a, 1993b, 1994; Dahlberg and Grånäs, 1994; Grånäs, 1990; Rodhe and Svedlund, 1990; Svedlund, 1991, 1993; Rohde, 1994a, 1994b) and to construct high-resolution maps of the area with a higher control of distribution and spacing of the moraine ridges.

Digital topographic data was put at our disposal by the Swedish Geological Survey to produce Digital Elevation Models (DEM's). ArcView served as a platform from which basic map calculations were performed and a map of the highest shoreline was generated. The digital data was also manipulated for shadowing (sun angle) effects in order to highlight minor topographic lineations, which sometimes proved useful in connection with the aerial photographic interpretation.

Sedimentological investigations were carried out in excavated trenches, placed perpendicular to the crest lines. Lithologic units were recognized based on lithofacies classification (Eyles *et al.* 1983). The boundaries between lithofacies units, their internal structures and deformation structures, were measured and documented at a scale of 1:20. Sediment samples where taken for grain size analysis.

Clast fabric analyses were carried out on prolate pebbles, excavated from 50x30 cm horizontal benches in the sections and with vertical sampling less than 20 cm. Each fabric set comprises 30 pebbles with the longest axis (a-axis) ranging between 2 and 12 cm, and only clasts with an a/b-axis ratio of ≥ 2 were accepted. Pebbles with close contacts to boulders were discarded due to possible orientation interference. The orientation data were statistically evaluated according to the eigenvalue method of Mark (1973, 1974) and graphically manipulated

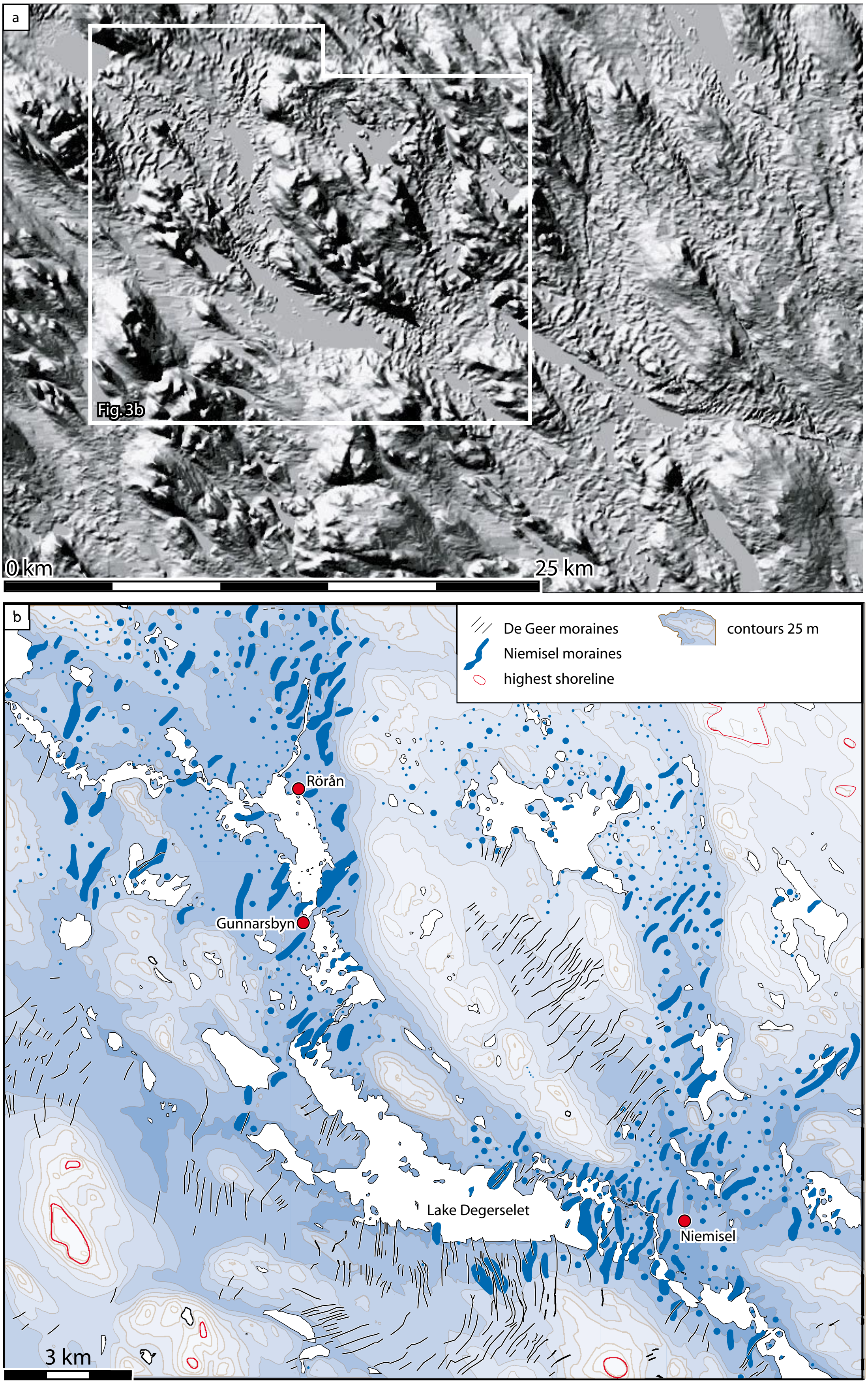


Fig. 3. (a) DEM-based relief map over the Råne River valley area. Note the dense occurrence of Niemisel moraine ridges in all low positions in the terrain, lying transverse to valley trends. Frame marks the position of Fig. 3b. (b) Detailed map over one of the investigation areas, showing the spatial distribution of De Geer and Niemisel moraine ridges. Note the occasional superimposing of De Geer moraine ridges on Niemisel moraine ridges.

with StereoNet© 1.01 for Windows.

Glaciotectonics measurements and analyses were carried out on deformation structures, fold axes, fault planes and thrust planes, as well as measurements on translocations along these. The orientation data were statistically evaluated according to the eigenvalue method of Mark (1973, 1974) and graphically manipulated with StereoNet© 1.01 for Windows.

Lake Coring – Sediment cores were recovered from 15 lake basins during winter fieldwork in 2003 (Gunnarsbyn area) and 2004 (Älvsbyn area) for shore displacement reconstruction. In the Gunnarsbyn area both a rod-operated piston corer (Aaby and Digerfeldt, 1986) and a Russian corer (Jowsey, 1966) were used. The piston corer provided continuous sediment sequences, enclosed in plastic liners (diameter 7.6 cm). The Russian corer was used to collect several 1 meter long (diameter 7.5 cm) and overlapping cores of the lowermost meters around the isolation level. In the Älvsbyn area only the Russian corer was used. To ensure that the entire lake sediment sequence was penetrated and the underlying till was reached a hammer-operated rod was used with both equipments. All cores were collected in the centre of the lakes.

Identification of isolation levels, sediment analysis – Classification of the sediments was mainly based on the lithological composition obtained from visual observation and dissection of the cores, both in the field and in the laboratory. Theoretically the terms gyttja clay, clay gyttja and gyttja refer to sediments with more than 3, 6 and 30 % dry weight of organic content, respectively (according to the Swedish Geological Survey); however, for practical reasons these differentiations are here based on visual estimations and are thus approximate.

The isolation level – was identified on the basis of lithological changes, evident as a shorter or longer transition zone of lithological change from pure minerogenic to more organic sediments. The lithologically identified isolation level is supported with measurements of magnetic susceptibility and determination of loss on ignition. The first appearance of organic matter, often reflected in the colour (a change from grey to brownish) and accompanied by decreasing minerogenic material and grain-size, are considered to reflect the initial basin isolation from a larger water body (in this case, the Gulf of Bothnia). However, the appearance of the isolation sequence varies slightly from lake to lake. In cases of gradual or transitional isolation sequences, the upper part was used as the isolation level as it marks the timing of complete isolation from the Gulf of Bothnia.

Magnetic susceptibility – the extent to which material can be magnetized (Walden *et al.*, 1999), was determined across the identified isolation zone in all cores. A Bartington Instruments MS2E1 magnetic susceptibility high-resolution surface scanning censor, coupled to a TAMISCAN automatic logging conveyor system, were used to measure the bulk magnetic susceptibility every 4 mm on the cleaned surface cores. The measurements were carried out immediately after fieldwork to eliminate any storage influence, i.e. oxidation, which can affect the stability of ferromagnetic iron oxides and iron sulphides (Snowball and Thompson, 1988; Oldfield *et al.,* 1992).

Loss on ignition – was also measured across the identified isolation zone, mainly to verify and strengthen the visual and lithological observations and magnetic susceptibility analyses. A minimum of 20 samples (1 cm thick) were analysed from each core. The samples were dried (at 105°C) and weighed before ignition for four hours at 550°C, and then finally weighed to obtain the dry mass carbon concentration (Bengtsson and Enell, 1986; Heiri *et al.*, 2001). Additional burning at 925°C was carried out on the cores from Mörttjärn and Stor-Knuttjärn to obtain the carbonate content.

Radiocarbon datings – A total of 67 samples were ^{14}C dated from the 15 lakes (Table 1 in Appendix III). Macrofossils were too sparse and therefore all age measurements were performed on bulk sediment samples (1 cm thick). The lowermost sample was collected at or just below the visually determined, fully completed isolation level, followed by discontinuous sampling upwards for 4-33 cm (2-7 samples for each core) into the more organic-rich lacustrine sediments. All samples were AMS dated at the Poznan Radiocarbon Laboratory in Poland. The acquired ^{14}C ages are calibrated into cal. yr BP, i.e. AD 1950, according to the Oxcal curve (Stuiver *et al.*, 1998) and displayed in time/depth diagrams (Fig. 4 in Appendix III). Isolation ages predominantly used are the dates immediately above the identified isolation, calculated as the mean age within the 2σ standard deviation. However, in a number of cases there are indications that the date immediately above the isolation is 'incorrect'. In these cases the apparently 'incorrect' age is excluded from the time/depth curve and the isolation age is based on extrapolation.

Investigations and OSL dating of beach deposits – At Heden (Fig. 1) a profile, set out for *c.* 1100 m over a series of 36 beach terraces and ridges (Fig. 7 in Appendix III), was levelled with a Topcon GTS-226 precision levelling instrument. The absolute altitude for the profile was tied to a nearby lake, the surface given with 1 m accuracy on the ordnance map. A shorter part of the profile (*c.* 280 m) was surveyed with Ground Penetration Radar (GPR), a pulseEKKO IV system with a pulser voltage of 400 V.

Thirteen of the 36 beach terraces and ridges, evenly distributed along the profile, were chosen for dating. Holes were dug to *c.* 50 cm depth, and plastic tubes with a diameter of 60 mm were hammered into the pit walls, extracted and sealed. Dating of the beach terraces were carried out with the Optically Stimulated Luminescence method (OSL) at the Nordic Laboratory for Luminescence Dating (Risø, Denmark) using the single aliquot regenerative dose (SAR) procedure to quartz grains (Murray and Wintle, 2000; Wallinga, 2002; Duller, 2004).

Summaries of papers

During the course of this project, several researchers have contributed with fieldwork, analyses, discussions, and have also been involved as authors (Table 1). The head supervisor, Per Möller, initiated the project, participated in all field seasons and has contributed during all phases of the project. However, the main fieldwork, analyses in field and in the laboratory has been carried out by the author of this thesis.

Paper I

Lindén M., Möller P. and Adrielsson L., manuscript: **Niemisel moraine: ribbed moraine formed by subglacial folding, thrust stacking and lee-side cavity infill**. Submitted to *Quaternary Science Reviews* (2005-12-08).

Niemisel moraine occurs in abundance in the costal zone of northern Sweden, preferentially in areas below the highest shoreline (200-220 m a.s.l.) but occasionally also slightly above. Based on detailed sedimentological and structural analyses of exposed sediments a depositional model for *Niemisel moraine* formation is suggested. For this purpose in total five *Niemisel moraine* ridges were excavated, with documented section lengths between 36 and 117 m and heights between 5 and 10 m. From these studies it is concluded that the moraines formed due to subglacial folding/thrust stacking of pre-deposited sediments contemporaneously with lee-side cavity deposition, forming vertically and distally prograding moraine ridges transverse to ice-flow. The proximal part of the moraines were built up by subglacial folding and thrust stacking sequences of pre-ridge formation sediments, whereas the distal parts were built up by glaciofluvially and gravity-flow deposited sediment in lee-side cavities. The initial thrusting and folding is suggested to be a result of near-marginal differences in bed rheology during the early or late melt season, generating a transverse to ice-flow compressive zone as a result of an up-glacier more mobile bed and a down-glacier less mobile bed. The lee-side cavities are considered as a result of ice-bed separation on the distal slope of the created obstruction, forming an integrated part of a subglacial linked-cavity drainage network, regulated in their degree of interconnection, size and shape by fluctuations in basal meltwater pressure/discharge and basal ice-flow velocity. Four of the five investigated *Niemisel moraines* are erosively cut and/or draped with a consistently more homogeneous deforming bed till associated with *De Geer moraine* formation and the approaching (retreating) ice margin.

Paper II

M. Lindén and P. Möller, 2005: **Marginal formation of De Geer moraines and their implications to the dynamics of grounding-line recession**. *Journal of Quaternary Science 20*, 113-133.

The aim of this paper was to, through caterpillar excavated trenches, do detailed sedimentological and structural analyses of the exposed sediments and to put up a depositional model for De Geer moraine formation. For this purpose in total four *De Geer moraine* ridges were excavated, documented and analysed. In one of the *De Geer moraines* both the section walls were documented to obtain the sediment distribution in three dimensions. It is concluded that *De Geer moraine* ridges formed due to subglacial sediment advection to the ice margin during temporary halts in grounding-line retreat, forming gradually thickening sediment wedges. The proximal part of the moraines were built up in submarginal position through continuous stacked sequences of deforming bed diamictons, intercalated with glaciofluvial canal-infill sediments, whereas the distal parts were built up from the grounding line by prograding sediment gravity-flow deposits, distally interfingering with glaciolacustrine sediments. Consequently, the implications of the reconstructed glaciodynamic and paleo-environmental conditions where examined on both local and regional scale. The rapid grounding-line retreat (c. 400 m/yr) was driven by rapid calving, in turn enhanced by fast ice flow and marginal thinning of ice due to deforming bed conditions. The spatial distribution of the moraine ridges indicates step-wise retreat of the grounding line. It is suggested that this is due to slab and flake calving of the ice cliff above the waterline, forming a gradually widening subaqueous ice ledge which eventually breaks off to a new grounding line, followed by regained sediment delivery and ridge build-up.

Paper III

Lindén, M., Möller, P., Björck, S. and Sandgren, P. 2006: **Holocene shore displacement and deglaciation chronology in Norrbotten, Sweden**. *Boreas* 35(1), 1-22.

The aim of this paper was to date the deglaciation and reconstruct the isostatic uplift since the deglaciation. This was done through dating of cored isolation sequences in twelve lake basins at different altitudes, thus reconstructing the Holocene shore displacement curve. Through subtraction of the rising global sea level and the damming and drainage effect of Lake Ancylus from the shore displacement curve an isostatic uplift curve was reconstructed. The coastal zone of Norrbotten was gradually inundated by the Ancylus Lake, following the retreating ice margin and forming a highest coastline approximately 210 m above present sea level. The highest lake basins, together with two basins above the highest shoreline, suggest ice free conditions already at 10 500 cal. yr BP. This is at least 500 years earlier than previously thought and implies rapid ice sheet break-up in the Gulf of Bothnia. The shore displacement (RSL) curve represents a forced regression of successively decreasing rate through the Holocene, from 9 m/100 yr to 0.8 m/100 yr. During the first 1000-1200 years the isostatic uplift was exponentially declining, followed by a constant uplift rate from c. 9500 cal. yr BP to 5500-5000 cal. yr BP. The last 5000 years seem to have been characterised by a low but constant rebound rate. The development of the Ancylus Lake stage of the Baltic may also be discerned in the Norrbotten RSL curve, suggesting that the chronology of the Ancylus Lake stages may have to be revised. The Littorina transgression is also reflected in the RSL curve shape.

In addition, a series of early to mid Holocene beach terraces were OSL dated to allow for comparison with the ^{14}C dated shore displacement curve. Interpretations of the ages of the OSL dated beach trerraces and their relation to former sea levels were clearly more problematic than the dating of the lake basin isolations and the precision of the ages retained sometimes poor.

	De Geer moraine	Niemisel moraine	Isostatic uplift
Fieldwork	M. Lindén, P. Möller *(SGU)*	M. Lindén, P. Möller, L. Adrielsson *(SGU)*	M. Lindén, P. Möller, P. Sandgren *(Crafoord)*
Loss of ignition			M. Lindén
OSL-dating			Nordic Laboratory for Luminescence Dating, Risö *(VR)*
Magnetic suseptibility			M. Lindén, P. Sandgren
^{14}C-dating			Poznan Radiocarbon Laboratory in Poland *(Royal Physiographical Society, Lund)*
Fabric analyses	M. Lindén, P. Möller	M. Lindén, P. Möller	
DEM-modelling	M. Lindén	M. Lindén	
Ground pentrating RADAR		M. Lindén, P. Möller, K. Kjær, A. Schoemacker	M. Lindén, P. Möller, K. Kjær, A. Schoemacker *(Lunds Geologiska Fältklubb)*

Table 1. Summary of contributors in different fields of research. Financial support in italic.

Results and discussion

Deglaciation chronology and shore displacement

The deglaciation of Norrbotten occurred earlier than previously thought, at 10 500 cal. yr BP (Lindén *et al.*, 2006; Appendix III). This was not the expected result from the investigation and nothing actually pointed against such an early deglaciation. However, this implies that the deglaciation after the Younger Dryas re-advance, from the Skövde-Billingen and Salpausselkä moraines, was very rapid – more rapid than previously thought. Within c. 1 000 years the ice sheet retreated c. 600 km, or in the order of 600 m/yr given an even recessional rate. This rapid ice sheet retreat is supported by the glaciodynamic conditions (see below), i.e. the deforming bed. As the margins of the late-glacial Scandinavian ice sheet became wet-based, ice-flow velocities increased resulting in thinning of the ice. As the ice margin was subaqueous, situated in the Gulf of Bothnia, calving was enhanced promoting the rapid deglaciation.

At deglaciation significant wave-washing occurred along the shores of the Ancylus Lake, resulting in a well developed highest shoreline and formation of till-capped hills on former islands (Fig. 4). During the subsequent isostatic uplift large areas of coastal Norrbotten were exposed to wave action (Fig. 4), forming wave-washed till, beach terraces of various types (Fig. 2f-g), as well as wave-washing of *De Geer* and *Niemisel moraines*, often giving exposed slopes an apparent increased boulder content (Fig. 2e). The dated isolation sequences of lake basins reveal that the early Holocene isostatic uplift rate was very high, and still is with a present uplift rate of c. 8 mm/yr (Fig. 5).

Shore displacement curve

A shore displacement curve is compiled from ^{14}C-dated lake basin isolations with all sites projected onto the 210 m HS isobase. The isolation dates are plotted against the corrected threshold altitudes of the lakes (Fig. 5). As noted in appendix III the lake surface altitudes according to existing ordnance maps are used as threshold altitudes. The error is estimated to be less than 1 m and these uncertainties are assumed to be of no importance for the outline of the curve. The shore displacement or relative sea level curve (RSL) can be divided into three phases, A-C, as follows (Fig. 5):

A. The deglaciation of coastal Norrbotten, which was more or less completed at *c.* 10 500 cal. yr BP, coincided with the Ancylus transgression in the southern Baltic Sea (Björck, 1995). The uplift rate was high for the first 1500 years after deglaciation, although declining. However, the detailed outline of the curve may imply that the Ancylus transgression and regression (drainage) can explain the slightly irregular

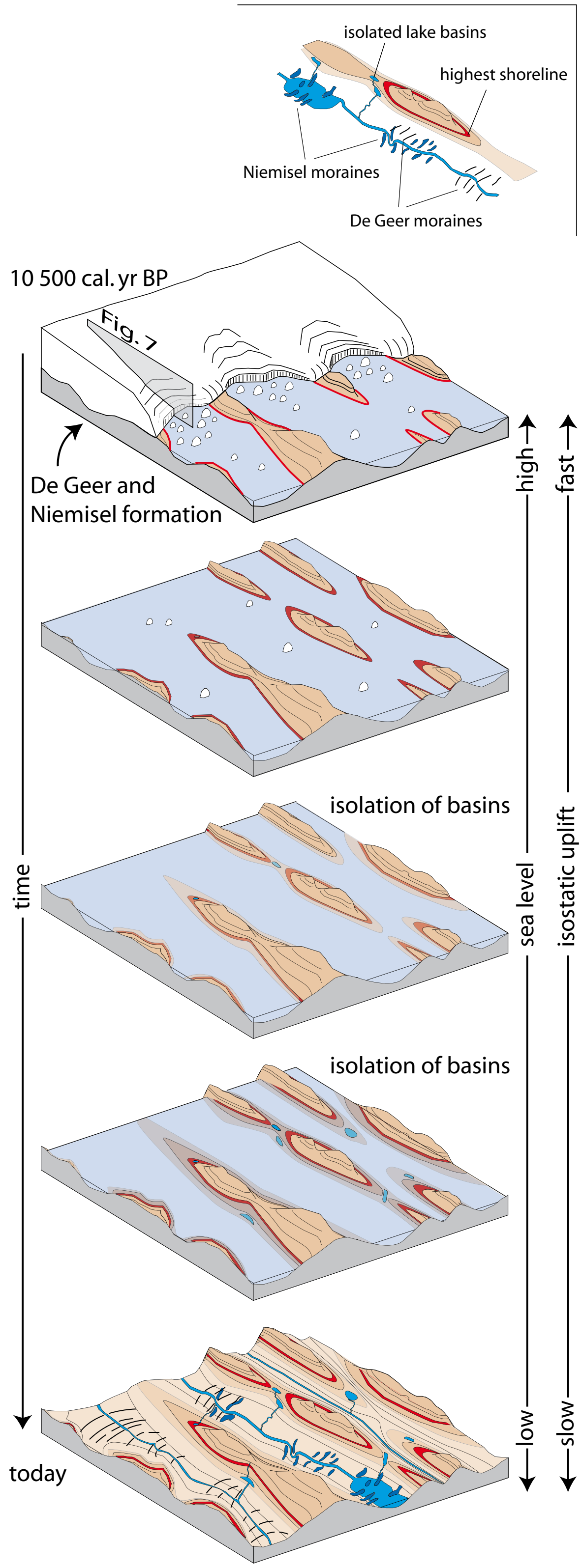


Fig. 4. Evolutionary sketch of the isostatic uplift history in Norrbotten, showing the development of the highest shoreline and the successive isolation of lake basins. The relative sea level and the isostatic uplift are further explored in Fig. 5. Frame marks the position of Fig. 7.

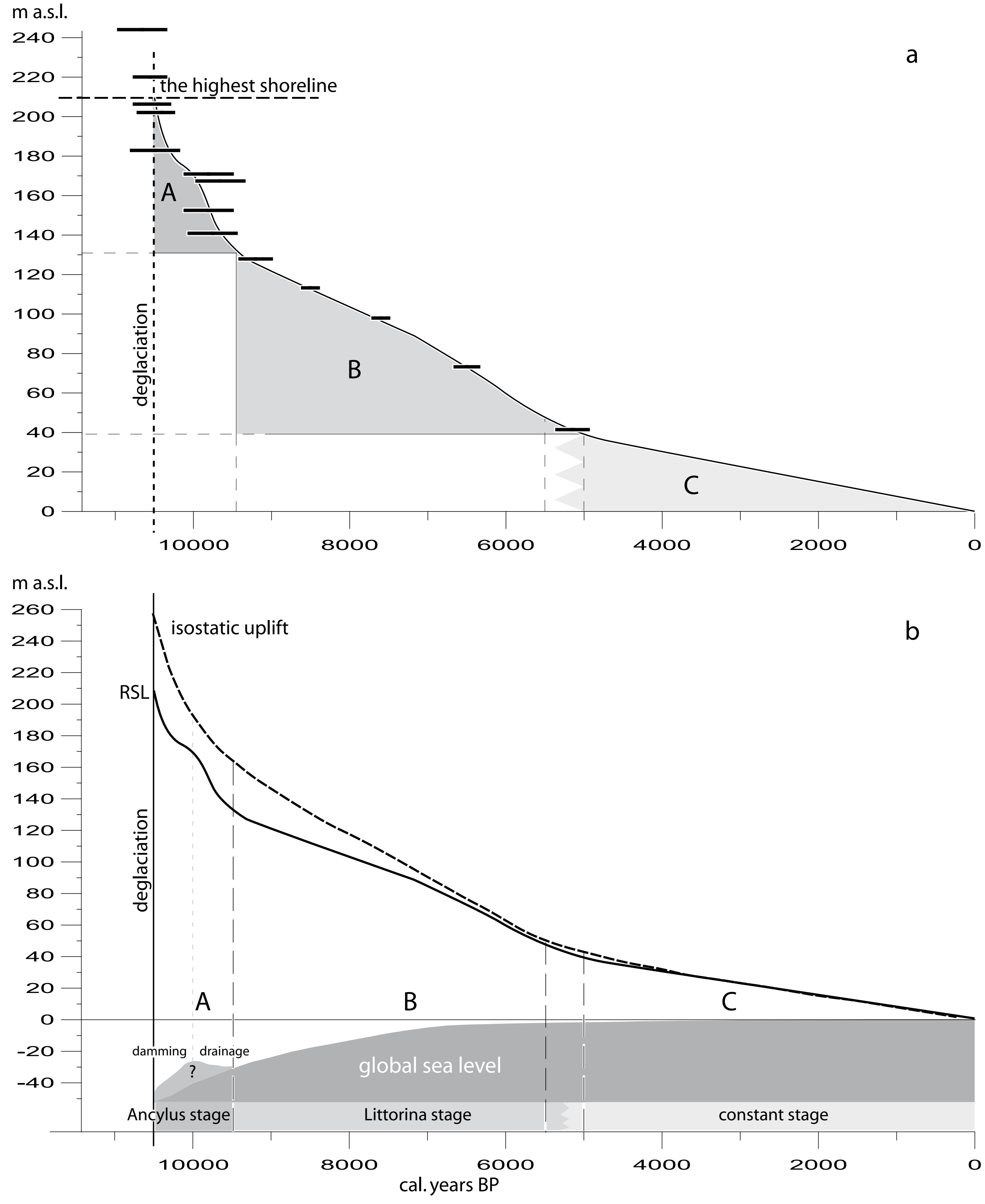


Fig. 5. (a) Relative sea level (RSL) curve for coastal Norrbotten, compiled from dated lake isolations projected to the 210 m a.s.l. highest shoreline isobase (black bars, 2σ age deviation or projected, see Fig. 4 in Appendix III). The different shaded areas A-C correspond to the phases in Fig. 5b: A, the Ancylus Lake stage; B, the Littorina Sea stage; C, phase of constant uplift. (b). Isostatic uplift for eastern Norrbotten (at the 210 m a.s.l. highest shoreline isobase) compiled from; - RSL + global sea level + Lake Ancylus. The global sea level curve (K. Fleming, pers. comm. 2004) is updated from Lambeck and Chappell (2001) and the Ancylus transgression (c. 10 500-9500 cal. yr BP, peaking c. 10 000 cal. yr BP) used is equivalent to a postulated 12 m up-damming of the Baltic. The isostatic uplift curve has been inverted to visualise the relationship to the RSL curve.

shape of the curve. The slower regression rate during 10 300-10 000 cal. yr BP, compared to before and after, could be interpreted as an effect of the Ancylus transgression as it was dammed above sea level. Likewise, the onset of the very rapid regression at 10 000 cal. yr

BP (Fig. 5) might be explained by the sudden drainage of the Ancylus Lake, producing a forced regression in the order of 10-15 m. However, if such an interpretation is correct, it would imply that the Ancylus transgression began later than postulated earlier, lasted for a shorter time and ended slightly later than what has been calculated (Björck *et al.,* 2001). This means that the Ancylus transgression occurred c. 10 500-9500 cal. yr BP, peaking c. 10 000 cal. yr BP. Some time around 9500 cal. yr BP the regression rate in Norrbotten seems to have slowed down.

B. Phase B of the RSL curve, 9500-5500 cal. yr BP, is characterized by an overall gently falling relative sea level, from 130 to 40 m. This coincides with the Tapes transgressions on the Swedish West Coast and with the Littorina Sea transgressions in the southern Baltic Sea; transgressions forced by rising global sea levels in combination with a significantly slowed-down rebound rate in southern Scandinavia. These processes can also be seen in Norrbotten (Fig. 5); the mean isostatic uplift declined from >90 mm/yr between 10 500-9500 cal. yr BP to *c.* 27 mm/yr between 9500-5500 cal. yr BP. With the uplift rate decreased significantly, the eustatic sea level rise is superimposed as a bulge on the Norrbotten RSL curve (Fig. 5) diminishing the regressive trend of the curve. The declining global sea level rise at *c.* 6500 cal. yr BP (Lambeck and Chappell, 2001), marking the final melting of the Laurentide Ice Sheet, can be discerned as an enhanced regression until 5500-5000 cal. yr BP.

C. From *c.* 5500-5000 cal. yr BP the isostasy seems to have become linear with a mean rebound rate of *c.* 8 mm/yr, which is actually the present day uplift value (Ekman, 1996). The outline of the curve is based on present isostatic uplift rate and the assumption that is has not been lower. It is however only based on a single point, although the dates appear consistent (Fig. 4a in Appendix III).

Implications for isostasy and eustasy

The configuration of the shore displacement curve for Norrbotten shows similar trends as nearby curves from northern Finland (Saarnisto, 1981), Ångermanland (Berglund, 2004) and Gästrikland (Berglund, 2005). They all show (i) a slightly visible superimposed effect of the damming and drainage of the Ancylus Lake after the rapid regression following deglaciation, (ii) a substantially slowed-down regression during the Littorina Sea stage and, finally, (iii) a nearly constant uplift for the last 4000-5000 years, close to the present uplift values according to Ekman (1996).

When isolating the isostatic uplift (Fig. 5b, green curve), including hydro-isostasy and geoidal changes, by subtracting the effect of the Ancylus Lake (12 m, see above) and the eustatic contribution (K. Fleming, pers. comm. 2004; see Fig. 5b), the exponentially declining trend of the first phase of the uplift until *c.* 9500 cal. yr BP becomes evident. From *c.* 9500 cal. yr BP (phase B) the isostatic uplift curve seems to be divided into two curve segments: one almost linear segment between 9500-5500 cal. yr BP and the second one from 5500-5000 cal. yr BP until present with a mean uplift rate corresponding to the uplift rate for the last hundred years (Ekman, 1996).

The shape of the isostatic uplift curve during phase A suggests a rapid elastic response in the lithosphere and asthenosphere to the unloading during the deglaciation, on which the Ancylus stages may be superimposed. The subsequent rapidly declining isostatic uplift could possibly be explained in different ways: delayed or disturbed viscous flow response in the upper mantle, temporal changes in mass transfer between the upper and lower mantle or even complex glacial unloading. Other unknown effects are lateral and vertical mantle variations in thickness and viscosity (Lambeck *et al.*, 1998; Wu *et al.,* 1998), which may cause a complex rebound. With, e.g., a low-viscosity layer the early relaxation will be rapid as most of the mantle flow will occur in this layer, but later the flow in the higher viscosity layers will begin to dominate and the exponential nature of the rebound change (K. Lambeck, pers. comm. 2004).

In the reconstructed shoreline displacement curve presented here, some questions arise concerning the resolution, the data set and the interpolation between adjacent isolated lake basins. It is possible to draw the shoreline displacement curve different without violating the isolation ages, and it is also problematic with the resolution in the early Holocene part of the curve. Particularly questionable is the drawn configuration of the damming and drainage of the Ancylus stage, where we in our attempt tried not to overestimate the effect. However, it is possible that both the damming and drainage was considerably faster than presented here. The ^{14}C- plateau around 10 000 cal. yr BP might also have implications for the precision of the curve in this interval. The spatial resolution, especially in the Litorina and the constant stages, is sometimes poor and minor fluctuations might be missing. In the Litorina stage the larger trends are detected, whereas the constant stage is based only on one lake basin isolation and the present isostatic uplift. As the age control in the constant stage is limited the most plausible and in fact the only possible interpolation to make, without violating the present knowledge, is rationally a more or less straight line. This is due to the fact that the

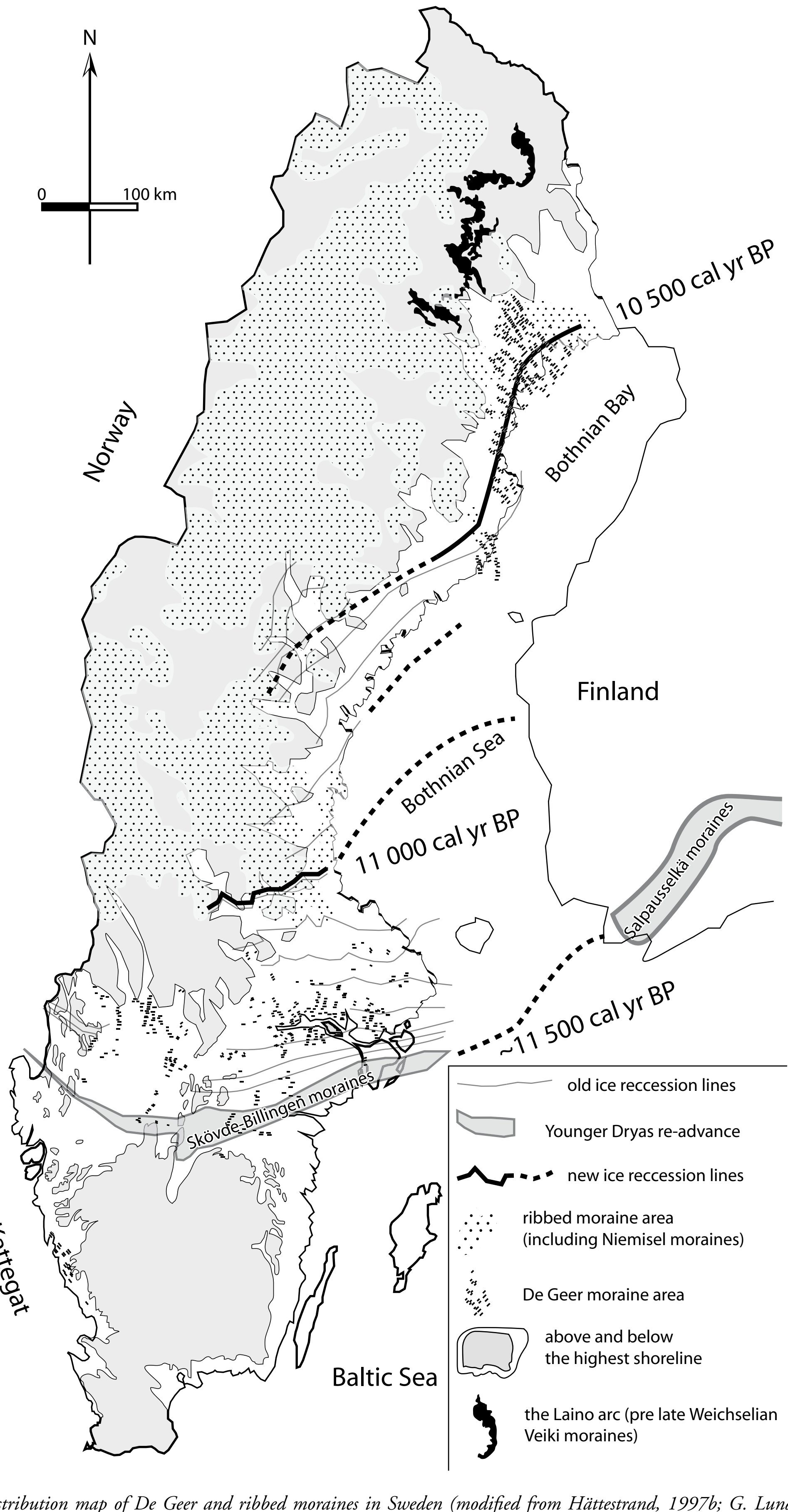


Fig. 6. Aerial distribution map of De Geer and ribbed moraines in Sweden (modified from Hättestrand, 1997b; G. Lundqvist, 1961; J. Lundqvist, 1994). Old (J. Lundqvist, 1994) and new hypothetical ice recession lines for the last deglaciation are also shown.

present isostatic uplift equals the total isostatic uplift for the last c. 5000 years; whether it has been constant or fluctuating does not matter. However, these geophysical speculations and questions can only be unravelled by including our data and the recently published data from Ångermanland (Berglund 2004) and Gästrikland (Berglund, 2005) in new modelling attempts or, alternatively to verify and improve the geophysical models.

Deglacial chronology

The two lakes above the highest shoreline provide confident depth/age curves and suggest deglaciation prior to 10 500-10 600 cal. yr BP, whereas the two lakes just below the highest shoreline are more problematic (Fig. 4l-o in Appendix III). Theoretically a minimum age of the deglaciation should be obtained by dating the earliest organic sediments in lake basins at or close to the highest shoreline, as it is plausible that these basins acted as traps for organic remains from the first colonizing plants. However, as organic content is low in deglacial lake sediments, some of our dated bulk sediments are clearly susceptible to contamination from reworked old organic material (Björck and Håkansson, 1982); even low quantities of 'dead' carbon result in too old ages for sediments with only a few percent of organic matter (Björck and Wohlfarth, 2001). We regard this as the most likely source of contamination since the carbonate levels are very low (0.8-1.5 %) and much of this might be crystal water. Both lakes yield quite high ages for their lowermost dated levels, ages older than the oldest dates in Finntjärn and Kallsjön. These dates are therefore considered too old due to contamination from reworked material, and the deglaciation age is suggested to be close to the used isolation ages for these lakes, 10 400-10 450 cal. yr BP, with fairly large error margins. We thus end up in a time span for deglaciation at 10 750-10 400 cal. yr BP, suggesting a deglaciation age of coastal Norrbotten around *c.* 10 500 cal. yr BP, which thus also sets the starting point for the shoreline displacement curve.

The first attempt to date the deglaciation in coastal Norrbotten was done by *De Geer* (1940), who extended his clay varve chronology into this area, resulting in a deglaciation age at the city of Boden around the Swedish varve year +400 (from the zero year) and a deglaciation rate at 350-400 m/year. The deglaciation age has later been revised to approximately correspond to the zero year (Fromm, 1965), suggesting a deglaciation age of *c.* 9238 varve yrs BP according to the revised Swedish varved clay-based chronology (Cato, 1985; Strömberg, 1985). The nearest, more recent varved clay chronology investigation is the study by Bergström (1968) from the county of Västerbotten (Ume River valley), *c.* 200 km to the southwest. An extrapolation of the ice recession lines from Bergström (1968) into coastal Norrbotten resulted according to J. Lundqvist (1994) in a deglaciation of the coastal plain 9300-9100 years BP in the Swedish varved clay-based chronology. As it recently has been shown that the Swedish varve chronology is erroneous due to some 700-800 missing varves (Björck *et al.,* 1996; Andrén *et al.*, 2002), this age interval should correspond to 10 100-9800 cal. yr BP. If our ^{14}C-dated deglaciation age of Norrbotten (Lindén *et al.*, 2006; Appendix III) is correct then this suggests that the deglaciation of coastal Norrbotten was some 500 years earlier than indicated from the extrapolated ice recession lines from Västerbotten, suggesting erroneous correlation of varved clay successions.

Based on the provided deglaciation age for coastal Norrbotten, c. 10 500 cal. yr BP, and recent shore displacement investigations along the coast of the Bothnian Bay (Berglund, 2004, 2005) a revised, though hypotetical, ice recession map is suggested for the Gulf of Bothnia (Fig. 6). In comparison to the ice recession map by J. Lundqvist (1994) a more pronounced embayed configuration of the retreating ice sheet margin is suggested when entering deeper water in the Bothnian Bay, resulting in increased recession rate and break-up of the ice sheet.

A depositional model for the water-terminating ice margin in coastal Norrbotten

The deglaciation of the coastal areas of northern Sweden took place in deep water along a calving ice front. In favourable positions sediments where building transverse moraine areas, *Niemisel moraine*, suggested to be formed beneath the wet-based ice sheet (subglacial landform), and *De Geer moraine*, formed at the receeding ice margin (submarginal landform). *De Geer moraine* ridges superimposed on *Niemisel moraine* ridges show the relative age relationship between the two landforms (Figs. 2c and 3). The relative age relation does, however, not give conclusive evidence about the absolute age relation, and it can not be concluded if *Niemisel moraine* areas were formed during ice sheet advance or retreat. However, there are strong evidences for cold-based conditions both at the inception and during full glacial phases for the northern and central part of the Scandinavian Ice Sheet (MIS 4-2), and a change into wet-based conditions probably did not occur until deglaciation (Kleman *et al.*, 1997; Kleman and Hättestrand, 1999; Kleman *et al.*, 1999). The conditions required for *Niemisel moraine* formation were, thus, only present during the deglaciation phase, and the depositional model is consequently linked to an active ridge-forming zone at some distance behind the finally retreating ice margin. The two ridge-forming depositional models, the *Niemisel moraine* model (Lindén *et al.*, submitted; Appendix I) and the *De Geer moraine* model (Lindén and Möller 2005; Appendix II), can thus be linked into a composite ice-marginal/subglacial sediment-landform system for the water-terminating ice margin in coastal Norrbotten (Fig. 7).

The composite landform system consists of three components, (i) a subglacial drift sheet in flat areas between the ridges, (ii) the subglacial constructional glaciotectonic deformation, thrust stacking, and lee-side cavity formation of *Niemisel moraines*, and (iii) the submarginal sedimentation and constructional glaciotectonic deformation of *De Geer moraines* (Fig. 7).

The subglacial drift sheet

The subglacial drift sheet occurs in flat areas between the moraine ridges and mounds, occasionally not forming any distinct landforms but just an irregular drift sheet. It consists of deformation till and glaciofluvial sediments. However, the drift sheet changes in proximal-distal direction, with the change from a linked cavity system to a braided canal system and from less mature ('Kalixpinnmo'; Beskow, 1935) to mature deformation till. The subglacial drift sheets can thus be related to the subglacial-submarginal formation of *Niemisel* and *De Geer moraine* (see below).

The subglacial ridges – the depositional model for Niemisel moraine formation

The ridge-forming facies and their architectural relationships are summarised below in a five-stage sequential model, graphically expressed in Fig. 8.

1. *Pre-ridge formation stage, T_0.* – The pre-ridge formation stage denotes the subglacial depositional system acting prior to individual

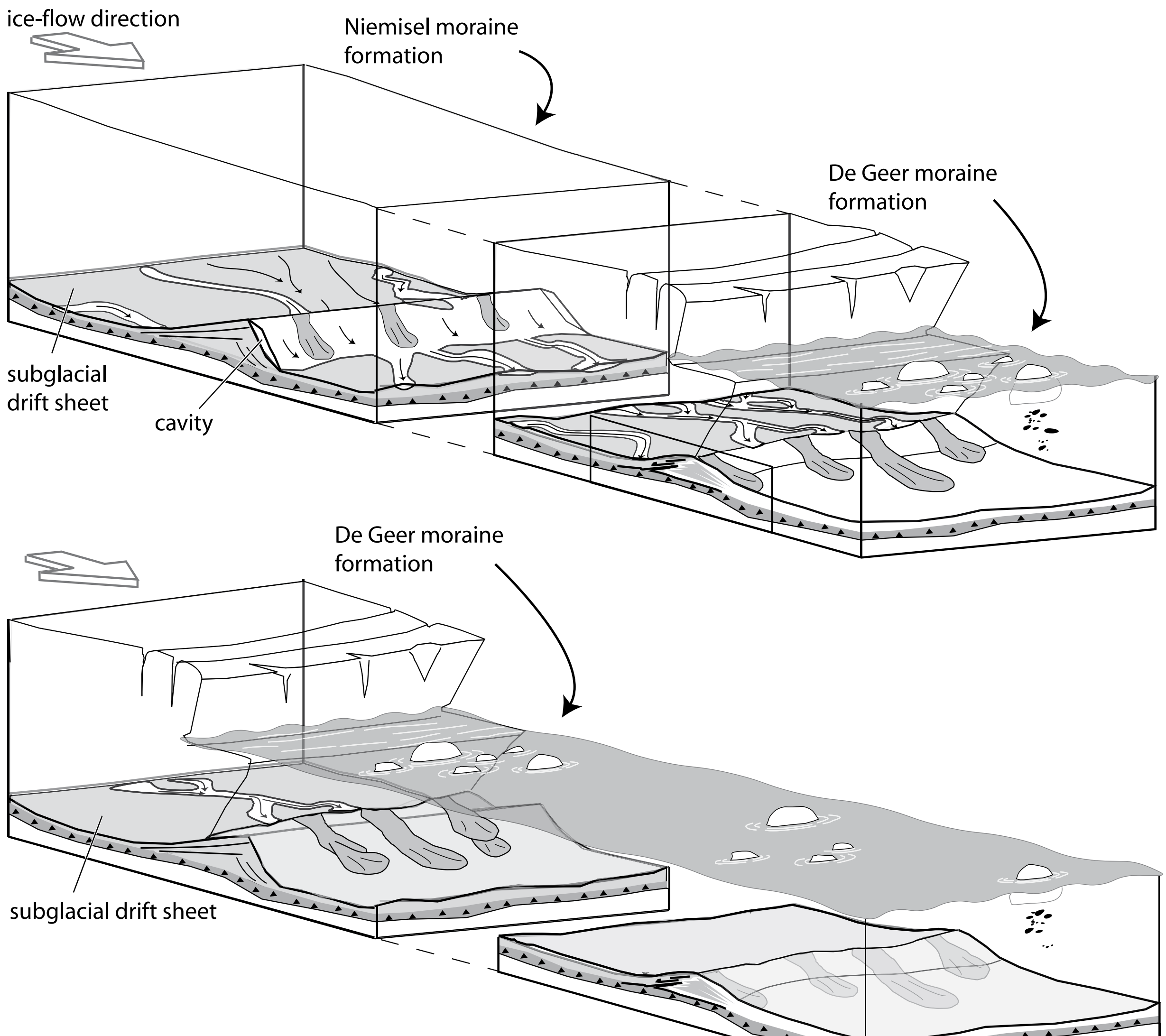


Fig. 7. Proximal-distal depositional model for Niemisel and De Geer moraine formation during subsequent recession of the ice margin.

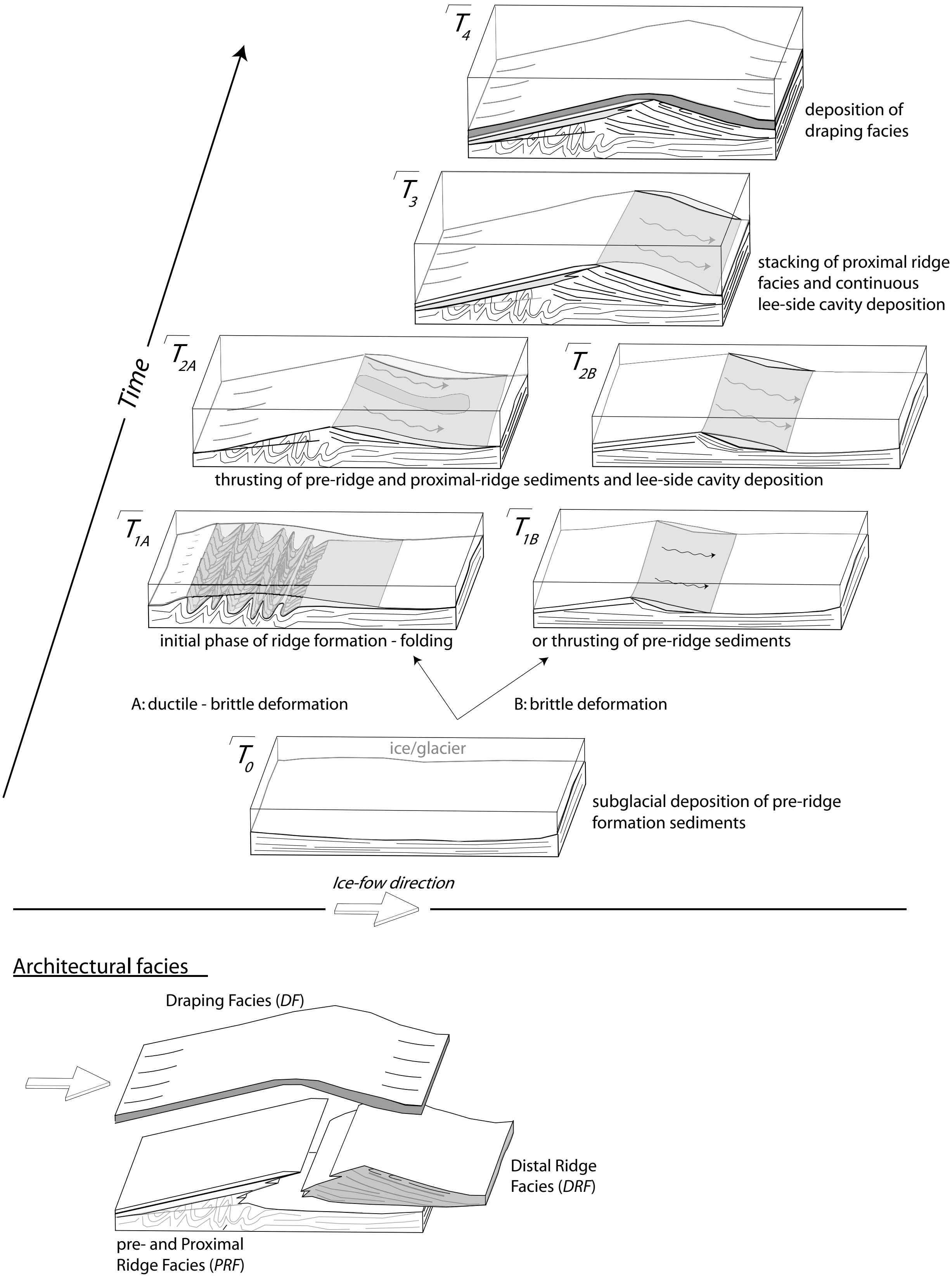


Fig. 8. Depositional model for Niemisel moraine formation. For explanations, see text (see also Appendix I).

ridge formation, as inferred from the sediment facies revealed in sediment beds, post-depositionally stacked into the proximal parts of the *Niemisel* moraine ridges. According to the genetical interpretations this was a subglacial environment predominated by bed deformation resulting in vertical accretion of both massive and tectonically laminated diamicts and glacitectonites. A distributed subglacial drainage systems (cf. Walder and Fowler, 1994) also formed, with deposition of low- to high-energy sorted sediments, which post-depositionally were more or less deformed.

2. *The initiation stage, T_1.* – The initiation of ridge formation is marked by a localized change from bed extension (deforming bed) to bed compression. Previously deposited sediments are either folded (T_{1A}) or thrust (T_{1B}) into lineaments transverse to ice flow, forming a transverse thickening of the bed, a *proto*-ridge. The proto-ridge forms an obstruction to glacier flow and a lee-side cavity develops with dimensions controlled by effective stress variations over a flow obstruction, by height to width ratio of the obstruction, and by ice-flow velocity and melt-water discharge/pressure (cf. Boulton, 1982).
3. *3-4: Ridge formation stages, T_2-T_3.* – Initial folding is followed by repetitive erosion events and thrust (nappes) stacking of sediment (T_{2A}). At places the proximal ridge formation is a continuous thrust stacking of sediment (T_{2B}). The thrusting produce drag structures or fully developed drag folds within the footwall sediment, and sediment is injected into tension veins, i.e. clastic dykes. The proximal part of the *Niemisel moraine* ridges is thus built up from sediments received from an up-glacier direction, 'constructive glacitectonism'.
4. As the proximal part of the ridge grows in vertical and proximal direction, so does the lee-side cavity. Distal cavity infilling in stages T_2-T_3 is characterised by repetitive deposition of glaciofluvial sediments interbedded with stratified and massive diamicts deposited from sediment gravity flows, introduced into the lee-side cavity by the deforming bed conveyor belt. All cavity sediments contain out-sized clasts delivered from the melting cavity roof (drop-stones). The sediment accretion in the lee-side cavity is subjected to both extensional and compressional deformation due to coupling/decoupling at the ice-bed interface.
5. *5: The draping stage, T_4.* – The draping stage is a late-stage smoothing of the transverse ridge by erosion of both proximal ridge and distal ridge sediments. The erosion is demarcated by a major angular unconformity at the base of the draping facies sediments but also by folding of underlying sediments. A massive to shear-laminated deformation till is finally draped over the ridge.

The ice-marginal ridges – the depositional model for De Geer moraine formation

Ridge-forming facies and their architectural relationships, and also their relations to ridge-distal facies, all indicate that *De Geer moraines* were formed along grounding lines associated with the glacier retreat in a subaqueous environment. They form ramp-like moraine ridges that are members of the morainal bank concept as proposed by, e.g., Powell and Domack (1995). A four-stage sequential formation model is summarised below and presented in Fig. 9.

1. A subglacial deforming bed system, coupled to a superimposed migrating braided canal network (Fig. 9a) exists some distance behind the retreating grounding line. Deformation till with strongly developed parallel clast orientation is deposited as a basement for the later deposited *De Geer moraines*.
2. An up-glacier relocation of the grounding line occur due to calving of the ice margin. The subglacial system becomes submarginal and the deforming bed transport system changes to a predominantly depositional system (Fig. 9b). Constructional glaciotectonic deformation (Hart and Boulton, 1991) generates a local thickening of the subglacial bed, i.e. the initiation of *De Geer moraine* formation. Outside the grounding-line is a glaciolacustrine environment with suspension settling, sediment gravity flows and melt-out of debris from the ice-cliff and from icebergs, but also with direct delivery from the deforming bed conveyor-belt and the debouching braided canal networks. Redeposition due to sediment failure also occur.
3. With a temporary stable grounding line, subglacially derived sediments are brought continuously to the submarginal-marginal zone, where the sediments are successively stacked, either as deformation till units on the proximal slope or as suspension settling and sediment gravity flow units on the distal slope (Fig. 9c). This results in a continuous vertical, but also proximal and distal build-up of a ramp-like moraine with an interfingering architecture. Syn- and postdepositional deformation, by folding of beds or bedsets and ploughing boulders, occur predominantly in the proximal part of the evolving ridge.
4. After a new calving event and further up-glacier relocation of the grounding line, the former submarginal-marginal environment change

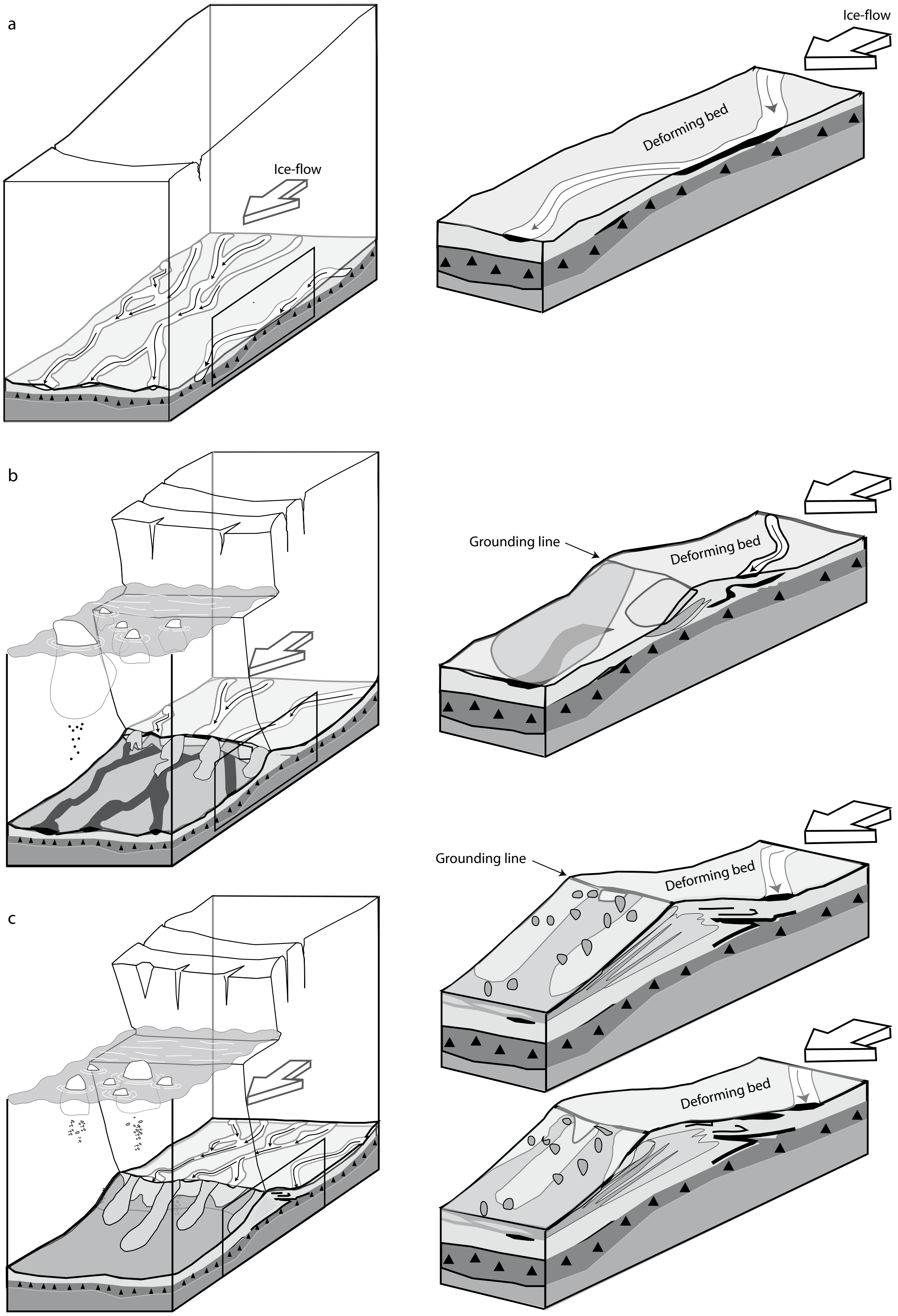


Fig. 9. Depositional model of De Geer moraine formation. Stages a-c in the figure are related to stages 1-3 in the depositional-model text, respectively, whereas stage 4 in the text correspond to the on-lapping glaciolacustrine sedimentation occurring distally of stages a-c in the figure. See text for explanation (see also Appendix II).

into an entirely glaciolacustrine environment, dominated by suspension settling of fine-grained sediment from density underflows and ice-rafted debris. Glaciolacustrine sediments form an on-lapping fining-upward succession, indicating gradually increasing distance to the ice margin due to further grounding-line retreat.

The most important link between the *subglacial drift sheet*, the *Niemisel moraine* ridges and the *De Geer moraines* is the subglacial hydrology with a distributed drainage system and a deforming bed. This suggests, that during the successive deglaciation of coastal Norrbotten, c. 10 500 cal. yr BP (Lindén *et al.*, 2006, Appendix III), *Niemisel* and *De Geer moraines* were deposited in a coherent, successively retreating subglacial-submarginal depositional system associated with deforming bed conditions (Fig. 7). The existence of a wide infra-marginal zone with deforming bed conditions is also supported by Kleman *et al.* (1997) who, based on aerial photographic interpretation of glacial landforms, identified a wet-based deglacial fan over the investigation area.

The proposed model for *Niemisel moraine* build-up suggests that glaciotectonic processes were the initiating cause for the ridge formation, and that the deposition of distal ridge sediments was a secondary effect. The initial thrusting and folding is suggested to be a result of near-marginal differences in bed rheology during early or late melt season, generating a transverse to ice-flow compressive zone due to a down-glacier change into less mobile bed (Appendix I). The process was highly dependent on lateral differences in basal water pressure and normal stress (Boulton, 1987; Boulton and Hindmarsh, 1987). The lee-side cavities are considered a secondary result of ice-bed separation distal to the created obstruction, and the cavities formed an integrated part of a distributed subglacial drainage network, regulated by the degree of interconnection, fluctuations in basal water pressure and by differences in basal ice-flow velocity (Boulton, 1982; Kamb, 1987; Iken and Truffer, 1997). Existing subglacial drift sheets of deformation till were disrupted during the initiation stage of the *Niemisel moraines* formation, and the draping sediment facies on top of the ridges is deformation till. Seasonal variations are suggested as an important factor in obtaining a compressive zone. The possible geochronological value, whether more than one *Niemisel moraine* was developed each year, as the compressive zone migrate up- or down-glacier, is unknown. A comparison between the ‘end of melt-season scenario’ and superimposing of ‘ribbed moraines’ on mega scale lineation in the Dubawnt Lake area in Canada (Stokes and Clarke, 2003), associated with the shut-down of the Dubawnt Lake ice-stream and with a stick-and-slip movement over cold-based patches can possible be made.

The build-up of *De Geer moraines* suggest advection of deforming bed sediments towards the ice margin, where it was deposited as a gradually thickening till wedge (Lindén and Möller; 2005; Appendix II). Deposition occurred close to the ice/bed separation point at the grounding line. The ice was close to floatation giving a decrease in effective stress, and the glacially induced shear stress dropped to zero. The depositional processes are suggested to have operated mainly during melt-season and the spatial distribution of *De Geer moraine* ridges indicates step-wise retreat of the grounding line, with deposition during temporary halts. On average four *De Geer moraines* were formed each melt season with a retreat rate of 600 m/yr, calculated from recently provided deglaciation ages (Lindén *et al.*, 2006; Appendix III). The retreat of the ice margin is suggested to be due to slab and flake calving of the ice cliff above the waterline, forming a gradually widening subaqueous ice ledge which eventually broke off to a new grounding line (Hughes, 2002; Lindén and Möller, 2005). Every calving event was followed by regained sediment delivery and a new ridge build-up at the new grounding-line. Fining-upward successions within on-lapping glaciolacustrine sediments indicate a gradually increasing distance to the ice margin due to further grounding-line retreat.

The spatial distribution data (Fig.3b) indicate lateral differences and restrictions in the depositional environment. *Niemisel moraines* occur in the lowest parts of the valleys, where also some of the largest and most continuous *De Geer moraines* are found. *De Geer moraines* are, however, also found on higher ground, on valley sides and interfluves, but then often discontinuous with a less pronounced morphologic expression. On altitudes above 150 m a.s.l., corresponding to water depths of c. 70 metres at deglaciation, only a few *De Geer moraines* have been found, indicating insignificant sediment flux and thus low ice-flow velocities (Fig. 3 in Appendix II). It confirms the assumption that, with a relatively smooth ice surface the highest porewater pressure occurred in the central deeper parts of the valleys, where the substrate was most prone to deform. Consequently, higher terrain experienced the opposite situation, a subglacial environment unsuitable for *Niemisel moraine* formation and with weak conditions for marginal *De Geer moraine* formation (too high effective stress). Higher ice velocities and sediment flux along the valleys also promoted a high calving rate and ice-margin embayment due to glacier thinning (van der Veen, 1996). This is demonstrated in the *De Geer moraine* configuration; a quite linear, slightly embayed ice margin in the flat coastal area, and a curved embayment further inland where the relief increases. However, the degree of embayment does not increase with successive recession according to the calving rate/water depth equation proposed by, e.g., Funk and Röthlisberger (1989) and Warren (1992). Instead the embayment is suggested to be related to and regulated by high calving rate and the thinning of ice caused by rapid ice-flow, which promoted the curvature (van der Veen, 1996), but counterbalanced by the ice flow towards the margin. A balance between calving rate and ice flux in the valleys appear to have been maintained during the recession as no adding-up effect is observed in the embayment (Fig. 3 in Appendix II).

The glaciodynamic situation in Norrbotten with a rapidly retreating ice margin indicates a strong negative mass-balance of the Scandinavian Ice Sheet, and a mass loss driven by the climate. Calving, enhanced by high ice-flow velocities, generated a thinning of the marginal zone. However, the effect of the cold ice in the interior zone was probably a high surface ablation rate and a supraglacial run-off during the melt season. The large catchment area in the interior zone generated high influx of supraglacial meltwater to the near-marginal warm-based zone where it could percolate and lower the effective stress at the ice-bed interface. The interior of the ice sheet can thus be regarded as fairly stable during parts of the deglaciation, although major displacement of the ice-divide zone, changing the ice-flow stream-lines, may have occurred during the period of rapid ice recession after the Younger Dryas re-advance (600 km between 11 500-10 500 cal. yr BP).

Areas with similar landforms and deglacial conditions

In Sweden *De Geer moraines* are mainly found in the central Swedish lowlands and in the coastal zone of northern Sweden, regions with significant water depths during the deglaciation (Fig. 6). *De Geer moraines* have not been observed in the coastal area between those two regions. This is an area with a different topography characterised by a high relief valley landscape and a steep coast along the Bothnian Sea. The general picture is that *De Geer moraines* preferentially occur in areas with smoother topography and limited areas of high ground. The only divergence in this general picture is found in the inland of Norrbotten where *De Geer moraines* occur in an area with glacially streamlined bedrock of higher relief, which coincides with the only area where *Niemisel moraine* is found.

The main distribution of *De Geer moraines* in Scandinavia is generally restricted to areas inside the Younger Dryas moraines, i.e. the Skövde-Billingen moraines in Sweden (Fig. 6) and the Salpausselkä moraines in Finland. However, *De Geer moraine* also occur outside these re-advance moraines, on the west-coast in south Sweden (G. Lundqvist, 1961) and in southern Finland (e.g. Zilliacus, 1987, 1989; Aartolahti *et al.*, 1995). The evident consequence of the Younger Dryas cold-event was, without explaining the causes, that the recession of the Scandinavian Ice Sheet was interrupted. The most plausible cause for this change to positive mass balance and 'sudden' re-advance was probably a decrease in meltwater production and an increase in winter precipitation, but probably also an increase in sea ice production preventing calving in subaqueous positions. In western Norway (Larsen *et al.*, 1991) *De Geer moraines* have been found to occur outside the Younger Dryas ice-marginal position, but appear to be more common inside the Younger Dryas ice-marginal zone, e.g. in northwestern (Blake, 2000) and northeastern Norway (Sollid and Carlson, 1984). This implies that rapid ice recession with extensive calving and *De Geer moraine* formation mainly occurred after the Younger Dryas re-advance (Fig. 6). This distribution pattern is also seen in North America where *De Geer moraine* distribution preferentially is restricted to areas subjected to rapid Holocene ice recession (e.g. 8300 BP in central Quebec; Hardy, 1976), in areas with former ice-dammed lakes and other areas of significant water depths at deglaciation.

According to the recently published deglaciation ages in Norrbotten (Lindén *et al.*, 2006; Appendix III), Ångermanland (Berglund, 2004) and Gästrikland (Berglund, 2005) the majority of the *De Geer moraine* ridges in Sweden were formed within a period of c. 1000 years (c. 11 500-10 500 cal. yr BP). Prior to this period the ice divide migrated from the Gulf of Bothnia to the Scandinavian mountain range (e.g. J. Lundqvist, 1994; Kleman *et al.*, 1997), setting up the conditions for ice recession along a flow-line from the Salpausselkä moraines in Finland to coastal Norrbotten. Along this flow-line *De Geer moraines* occur frequently inside the Salpausselkä – Younger Dryas – moraines in Finland (Zilliacus, 1987, 1989; Aartolahti *et al.*, 1995), but are absent on the seafloor of the Gulf of Bothnia (Andrén, 1990). The absence implies unfavourable conditions for *De Geer moraine* formation here, possibly due to the great water depths. A low-gradient ice surface profile and increasing water depths, exceeding c. 230 m outside Norrbotten, could have led to very rapid calving and a collapse of the ice sheet in the Gulf of Bothnia (Berglund, 2004, 2005; Lindén *et al.*, 2006). This scenario is possibly the most plausible explanation for the absence of *De Geer moraines* along the Swedish coast of the Bothnian Sea, where the deglacial water depth infer gradually increasing water depth during ice retreat. Temperate glaciers cannot survive in deeper water than at a critical threshold, and will calve back until grounded in shallow water (Powell, 1984), as probably was the case along the present coast where *De Geer moraine* reoccur. This implies that the threshold for *De Geer moraine* formation was locally regulated by the relation between ice sheet thickness/profile and water depth. Local maximum water depth for *De Geer moraine* formation also has been reported from the Chapais area, Canada (180 m; Hardy, 1976). The presence of *De Geer moraine* is thus suggested to coincide with a subaqueous ice terminus, having a more or less regularly retreating grounding-line, but also associated with a rapid ice recession driven by calving. These glaciodynamic conditions should be applicable to *De Geer moraine* formation everywhere and have a global significance as these are associated with rapid ice sheet recession and significant water depths in both Scandinavia and in North America.

The knowledge of the spatial distribution of *Niemisel moraine* is more problematic, as there is only one area (Norrbotten) where these have been identified and investigated in detail (Lindén *et al.*, submitted; Appendix I). This area, with transverse moraines genetically identified as *Niemisel moraine* has, however, been included into the more general concept of ribbed moraine (Hättestrand and Kleman, 1999), and interpreted by them to be formed due to processes not compatible to sediment facies and ridge architecture, as shown by Lindén *et al.* (submitted; Appendix I). The problem is thus to differentiate *Niemisel moraine* from other types of geomorphologically similar transverse moraine ridges (e.g. Rogen moraines; Möller, 2006). Transverse moraines with some similarity to *Niemisel moraine* in architectural build-up have, however, been described from one area in west-central Sweden (in Härjedalen; Minell, 1977). Otherwise ribbed moraines must be regarded as a polygenetic group of moraines with in varying degree similar morphology, but of different origins. Such a ribbed moraine type is Rogen moraine which, however, is related to quite different processes and depositional environments (Möller, 2006) than *Niemisel moraine* and do not fit into a context of water-terminating ice margins.

Nevertheless, it is possible that the glaciodynamic conditions required for *Niemisel moraine* formation were special but existed during the deglaciation of coastal Norrbotten, the only area where *De Geer moraine* is known to be associated with *Niemisel moraine*.

A comparison between De Geer moraine areas in northern and central Sweden

The most pronounced *De Geer moraine* areas in Sweden are found in the lowland of south-central Sweden and along the Bothnian Bay coast in northern Sweden (Fig. 6). However, the combination of the two landform elements, *De Geer moraines* and *Niemisel moraines,* are only present in the northern area, and a universal applicability of the above presented landform system and depositional model from Norrbotten may be limited. The model is here tested on the type-area of classical *De Geer moraines* in east central Sweden (De Geer, 1932; H. Möller, 1962; Strömberg, 1965). In the comparison following parameters must be considered: water depth during deglaciation; ice thickness; ice-retreat rate; landscape morphology/topography; bedrock geologi/till composition; glacier temperature/basal temperature; melt-water production/ablation rate; subglacial drainage system.

The physical setting of the two *De Geer moraine* areas is approximately the same. There are no significant regional differences in deglacial water depth (140-190 m), landscape morphology and bedrock geology. Ice-retreat rate, meltwater production and thermal conditions were probably also relatively equal, and the supraglacial ablation must have been high during this part of the Holocene climatic amelioration. One important difference is, however, obvious; *De Geer moraine* ridges in east central Sweden are associated with large, continuous eskers and the moraine ridge configuration often indicates a pronounced ice margin embayment, i.e. calving bays around the eskers (cf. Strömberg, 1989).

The presence of large eskers indicates Röthlisberger channels connected to a well developed discrete subglacial drainage system (Rötlisberger, 1972; Benn and Evans, 1996). Such a drainage system differs from a distributed drainage system, as suggested for *De Geer* and *Niemisel moraine* formation in Norrbotten, where also very few eskers are associated with areas of *De Geer* and *Niemisel moraine*. A deforming bed is generally related to distributed drainage systems (Benn and Evans, 1996), and the subglacial transport of material to the ice margin diminishes substantially with a discrete subglacial drainage system of Röthlisberger channels, which increases the effective stress and thus the shear strength of the substrate. Higher basal shear strength and concordantly a lower ice-flow velocity implies a thicker ice sheet and a steeper ice surface profile (Paterson, 1994).

Obviously, the divergent subglacial drainage system also indicates different processes in the *De Geer moraine* formation in central Sweden. Lindén and Möller (2005) suggested that winter ice-push could be an additional processes involved in *De Geer moraine* formation. Subaqueous push moraines have also been reported from modern glacial settings (e.g. Boulton, 1986: from Aavatsmarkbreen, Spitsbergen) where several subaerial push moraines could be followed into the frontal subaquatic environment. Unfortunately, earlier investigations on mid-Swedish *De Geer moraine*, were focused only on geomorphology and spatial distribution of the ridges, and detailed sedimentological observations are missing (e.g. H. Möller, 1962; Strömberg, 1965). However, H. Möller (1962) observed *De Geer moraine* ridges with a core of a coarse, weakly sorted till with compressional structures and boulders indicating debris release from the ice sole or ice cliff.

The glaciodynamic conditions during the formation of the mid-Swedish *De Geer moraine* areas rather imply hard-bed conditions, basal transport of debris within the ice and possibly ice-push at the ice margin instead of sediment advection by a deforming bed towards the ice margin. Consequently, the depositional model for Norrbotten does possibly not apply to *De Geer moraine* areas associated with a discrete subglacial drainage system. It also appears as the large Röthlisberger channels in central Sweden remained operating during the winter, preventing any deforming bed from developing. Contrary, it appears as the drainage system in Norrbotten was shut-down during winter, and started from scratch every melt season.

The occurrence of *Niemisel moraine* and *De Geer moraine* associated with deforming bed suggests an ice marginal/subglacial landform association in many ways differing from the landforms in central Sweden. Of interest here are areas in central Québec from where transverse moraines, interpreted as Rogen moraine and *De Geer moraine* with transitional features and *De Geer moraine* grading into Rogen moraine have been described (Beaudry and Prichonnet, 1995). However, neither the interpretations of Rogen moraine, the transitional features nor the grading of *De Geer moraine* into Rogen moraine are supported by sedimentological data.

Implications and ideas for the future

Sedimentological data from the ice-marginal and subglacial landforms of coastal Norrbotten imply a depositional system with sediment transport typically associated with a deforming bed. A subglacial drainage pattern with linked cavities and braided canals suggest a distributed subglacial drainage system. A comparison to areas in central Sweden, with similar environmental conditions during deglaciation, shows conclusive evidences that *De Geer moraine* areas also were formed at the margin of an ice sheet drained by large Röthlisberger channels, and subsequently with a discrete subglacial drainage system. Significant differences in effective normal stress and basal sediment transport must have occurred beneath the ice sheet margin during the deglaciation of the two areas, and the models for *De Geer moraine* formation in Norrbotten can not uncritically be used as a universe model for *De Geer moraine* formation. The model for *Niemisel moraine* is also limited to special subglacial conditions and is not expected to serve as a general model for ribbed moraines. The two models may, however, be applicable in areas with a combination of the two landforms where large eskers are absent. The combination of landforms is decisive.

The polygenetic origin of ribbed moraine is intriguing, especially as *De Geer moraine* and 'ribbed moraine' in some regions are found together, occasionally also as transitional forms. The term ribbed moraine (including Rogen moraine) is problematic and the need for genetically derived terms instead of morphologically such are obvious. *Niemisel moraine* is considered a specific form of moraine ridges formed subglacially beneath the ice sheet with a distributed drainage system, but probably close to the ice margin.

With these perspectives in focus the issue for future investigations of *Niemisel moraine* and other ribbed moraine and *De Geer moraine* is obvious. How were the *De Geer moraine* areas in central Sweden formed? How do the moraine ridges and the till areas in between differ from the drift sheet and moraine ridges in Norrbotten? Is *Niemisel moraine* specific for Norrbotten, or do they exist in other areas? Why did the subglacial drainage system in Norrbotten differ from the drainage of the retreating ice sheet in central Sweden? Why did not large Röthlisberger channels develop during the deglaciation of Norrbotten, and what effects did the rapid ice-flow velocity and deglaciation rate have on the subglacial drainage?

Conclusions

Norrbotten is the only area where the *Niemisel/De Geer moraine* landform association has been documented. From this ice marginal/subglacial landform association it can be concluded that:

- The primary components in the early stage of *Niemisel moraine* build-up, forming complex sequences of folded, thrust and stacked packages of both sorted and diamict sediment, were primarily deposited prior to ridge formation during deforming-bed conditions. Continuing proximal and vertical accretion/stacking of sediment grade/interfinger in distal direction with lee-side cavity-deposits consisting mainly of gravity flow and fluvial deposits, i.e. suggesting a linked cavity system.
- *Niemisel moraine* formation is suggested to have been initiated in a transitional zone between deforming bed and rigid bed conditions, the

latter propagating up-glacier as a result of falling basal meltwater pressure at the end of the meltwater peak season or, alternatively, propagating down-glacier at the beginning of the melt season when basal drainage system is recharged, gradually increasing basal meltwater pressure.

- The *Niemisel moraine* ridges are seen to be erosively cut and/or draped by a massive to shear-laminated diamict, a deforming-bed till associated with *De Geer moraine* formation and the approaching (retreating) ice margin.
- *De Geer moraine* ridges were preferentially formed during the summer season at temporary halts in the grounding-line retreat as a result of subglacial sediment advection to the ice margin. During ice-margin retreat *De Geer moraine* ridges were occasionally superimposed on previously formed *Niemisel moraine* ridges, demonstrating the proximal-distal and temporal relations between these ridge types. The proximal part of *De Geer moraines* is built up in submarginal position as stacked sequences of deforming bed diamictons intercalated with glaciofluvial canal-infill sediments. The distal part of *De Geer moraines* is built up from the grounding-line by prograding sediment gravity flow deposits, distally interfingering with fine-grained sediments deposited in the frontal glaciolacustrine environment.
- *De Geer moraine* ridges probably also form during winter re-advances of the ice margin as push moraines, though this is not supported from our field studies.

From the obtained deglacial chronology and the reconstructed shoreline displacement and isostatic uplift in Norrbotten, as a response to the unloading during the deglaciation, it can be concluded that:

- The rapid marginal recession rate (c. 600 m/yr) was driven by rapid calving, in turn enhanced by fast ice flow and marginal thinning of the ice due to deforming bed conditions.
- The step-wise retreat of grounding lines, as indicated from the spatial distribution of *De Geer moraine* ridges, suggests subaqueous ice ledge calving to new grounding-line positions, followed by regained sediment delivery and ridge build-up. The spatial distribution of *De Geer moraine* ridges in Norrbotten suggests an average formation of *c.* 4 'summer moraine' ridges every year.
- Coastal Norrbotten was deglaciated *c.* 10 500 cal. yr BP, 500 years earlier than previously thought. This implies rapid break-up of the ice sheet in the Gulf of Bothnia.
- The damming and drainage of Lake Ancylus is evident as an overprint on the oldest part of the constructed RSL curve, suggesting that the Ancylus transgression in the southern Baltic began later, lasted for a shorter time and ended slightly later than previously thought, *c.* 10 500-9500 cal. yr BP. The resulting isostatic uplift curve is an exponential function for the first c 1000-1200 years of uplift history then it slows-down to a semi-constant isostatic uplift until *c.* 5000 cal. yr BP, where slows-down to an almost constant isostatic uplift until today.

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Svensk sammanfattning

Under kvartärtiden (de senaste 2,6 miljoner åren) har istider (glacialer) och mellanistider (interglacialer) avlös varandra. Mellanistider är perioder (ca 10 000 år) mellan istider (ca 100 000 år), vilket innebär att istid varit det normala under kvartärtiden och att Holocene, den istid vi nu lever i, endast representerar ett kortare intervall. Under istiderna täckte inlandsisar stora delas av norra Europa (bla Skandinavien) och Arktis för att vid övergången till mellanistid snabbt smälta bort.

Syftet med det här doktorandprojektet har varit att med stöd av sedimentologiska och strukturella undersökningar av två typer av moränryggar, *Niemisel* och *De Geer moräner*, rekonstruera både landformsbildande processer och inlandsisens dynamik vid dess avsmältning (deglaciation) i Norrbotten. För att få en uppfattning om när moränerna bildades startades ett fristående projekt med avseende att fastställa tidpunkten för deglaciationen av Norrbotten samt jordens respons på den omfördelning av tyngd som skedde vid deglaciationen, då is blev till vatten i haven, som återspeglas i landhöjningen. Således behandlar avhandlingen bildningen av *Niemisel moräner* (Appendix I) och *De Geer moräner* (Appendix II), samt tidpunkten för inlandsisens avsmältning och landhöjningen (Appendix III).

Undersökningarna av fyra *De Geer moränerna* och fem *Niemisel moränerna* visar på en retirerande subakvatisk inlandsis, bottensmältande i områdena nära iskanten och en hög ishastighet över ett deformerande underlag. Den höga ishastigheten ledde till uttunning av frontområdena och ökad kalvning, vilket i sin tur ledde till högre reträtt hastighet.

Undersökningarna av *De Geer moränerna* visar vidare att de bildades under iskanten, där deformations morän transporterades fram och avsattes, som ett resultat av en låg effektivspänning nära iskanten och därför ingen inducerad skjuvspänning. Detta som ett resultat av att iskanten var nära flytpunkten vilket ledde till att transporten (deformationen) av underlaget upphörde med deposition dom följd. Iskantens närvaro kan ses i de distala sedimenten som visar på en glaciolakustrin miljö (utanför iskanten) medan de proximala sedimenten visar på en subglacial miljö (under isen).

Undersökningarna av *Niemisel moränerna* visar att de bildades under isen (subglacialt) genom deformation av redan avsatta sediment samt ifyllnad av glaciofluviala (vattenavsatta) sediment i läsides kaviteter, bildade genom flödesseparation nedströms om de upp-staplade – deformerade – proximala sedimenten. De proximala sedimenten har initialt veckats eller skjuvats upp längs proximalt lutande skjuvplan och återfinns som staplade packar av subglacialt avsatta sediment. Detta visar på, framförallt den återkommande uppskjuvningen, en vertikal och distal uppbyggnad av moränryggar orienterade transversellt mot den dominerande isrörelseriktningen.

Det finns inga bevis för att *Niemisel* och *De Geer moräner* skulle vara bildade vid olika istider, men inte heller något som talar för att de är bildade vid samma deglaciation. Den enklaste slutsatsen, baserad på förutsättningarna för en deformerande bed och bevarings potentialen, favoriserar däremot att bägge morän typerna bildas mer eller mindre samtidigt i ett, distalt-proximalt varierande, landforms system.

Undersökningar av borrkärnor från 15 sjöar belägna på olika nivå över havet visar att inlandsisen försvann från Norrbottens kustzon för 10 500 år sedan och att landytan höjts med 200-220 m i förhållande till dåvarande Bottenvikens nivå sedan dess. Vilket innebär att den högsta kustlinjen utbildades som bildades för 10 500 år sedan och återfinns 200-220 m ö h i området. Sedan dess har den relativa havsytan successivt sjunkit allt eftersom landet återhämtat sig från inlandsisens nedtryckning vilket även medfört att kusten flyttats 30-100 km längre ut i Bottenviken. Landhöjningen kan ses som utbildandet av kalottberg, olika typer av strandterrasser och ryggar samt som svallad morän i området. Högsta kustlinjen är som högst vid dagens kust och minskar inåt landet. Variationen i högsta kustlinjens nivå har två orsaker;

- Då inlandsisens var som störst (ca 22 000 år sedan) var isen som mäktigast i Bottenhavet, varför återhämtningen är störst längs kusten och sedan successivt minskar in i landet.
- Vid deglaciationen, då landhöjningen redan startat och var som snabbast, flyttades inlandsisens centrum till fjällkedjan. Detta medförde att inlandsisens kant retirerade inåt landet och att landet redan hunnit höjas en del från det att kustzonen blev isfri och att det samma skedde i inlandet.

Undersökningarna av borrkärnorna visar att två av sjöarna, de högst belägna, befann sig över havsytan även vid deglaciation medan de övriga 13 sjöarna utgjorde bassänger vid samma tidpunkt. Övergången från bassäng till sjö, isoleringen, kan ses i borrkärnorna som en övergång från grå sand, silt eller lera till brun lergyttja och sedan gyttja med successivt ökande organisk halt. Miljöförändringen vid isoleringen, alternativt när isen smälte bort från de två sjöarna belägna ovanför högsta kustlinjen, medförde att den organiska produktionen startade och denna har daterats med ^{14}C-metoden. Eftersom sjöarna ligger på olika nivåer har både deglaciationen kunnat dateras och en strandförskjutningskurva reproduceras utifrån när olika bassänger isolerades. Genom att subtrahera effekten av Ancylussjöns dämning och dränering samt ”väldshavets” stigande nivå, pga vatten från inlandsisarna, kan den absoluta landhöjningen – isostasin – rekonstrueras. Den visar att landhöjningen var som störst de första 1000 åren efter deglaciationen (~9 m/100 år) för att sedan minska till en relativt konstant nivå fram till för ca 5000 år sedan då landhöjningen åter igen minskade till en konstant nivå motsvarande dagens landhöjning (~0,8 m/100 år). Den första exponentiellt avklingande fasen beror på den elastiska responsen i litosfären och astenosfären medan Litorina fasens relativt konstanta landhöjning kan skyllas responsen i mantelns mer viskösa lager samt responsen på en senglacial förskjutning av iscentrum mot fjällkedjan, dvs en förskjutning av massan. Den avslutande konstanta fasen med en landhöjning motsvarande dagens är mer häpnadsväckande och mer detaljerade undersökningar är att föredra innan utförligare tolkningar görs. Det är dock inte enbart i Norrbotten denna uppträder utan fenomenet är även känt från Gästrikland, Ångermanland och finska Österbotten.

Inom studien gjordes även en jämförelse mot OSL-daterade strandterrasser. Resultatet visade att tillförlitligheten hos metoden inte var jämförbar med ^{14}C-dateringarna av isoleringar.

Tillsamman pekar de tre artiklarna på en snabb deglaciation av Norrbotten. Den tidiga dateringen av deglaciationen stöds av de subglaciala indikationerna, högt porvattentryck och ett deformerande underlag, samt höga ishastigheter. Detta leder till en uttunning av inlandsisens marginella delar vilket i sig tur ger förutsättningar för omfattande kalvning och därför en snabb isreträtt.

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