Strål
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The effect of a glaciation on East Central Sweden: case studies on present glaciers and analyses of landform data

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SSM perspective

Background and objectives

The Swedish Radiation Safety Authority reviews and assesses applications for geological repositories for nuclear waste. When assessing the long term safety of the repositories it is important to consider possible future developments of climate and climate-related processes. Especially the next glaciation is important for the review and assessment for the long term safety of the spent nuclear fuel repository since the presence of a thick ice sheet influences the groundwater composition, rock stresses, groundwater pressure, groundwater flow and the surface denudation. The surface denudation is dependent of the subglacial temperatures and the presence of subglacial meltwater. The Swedish Nuclear Fuel and Waste Management Company (SKB) use the Weichselian glaciation as an analog to future glaciations. Thus, increased understanding of the last ice age will give the Swedish Radiation Safety Authority (SSM) better prerequisites to assess what impact of a future glaciation might have on a repository for spent nuclear fuel.

This report compiles research conducted between 2008 and 2015 with a primary focus on 2012-2014. The primary aim of the research was to improve our understanding of processes acting underneath an ice sheet, specifically during a deglaciation.

Results

Based on the results of studies on present day High Alpine glaciers, the Antarctic and Greenland ice sheets, and glacial geomorphological mapping from land and marine-based data in the Gulf of Bothnia, we have gained an insight into how future glaciations may affect the northeast coast of Uppland, eastern Sweden. Evidence of ice streaming over a large distance in the Gulf of Bothnia, including the presence of numerous glacial lineations of varying scale, contests to a period of fast basal sliding during the Late Weichselian deglaciation. This suggests that there would have been a degree of erosion at the ice sheet bed during this time. These results have been taken into account in SSM:s reviews of SKB:s post-closure safety analysis, SR-Site, for the proposed repository at Forsmark.

The evidences for a Baltic Ice Stream are linked to the long term existence of the Gulf of Bothnia and the Baltic. Unfrozen soft sea bed sediments allowed a fast sliding speed at a stage of glacier advance and the bed remained wet all through the glaciation.

Need for further research

The marine-based data in the Gulf of Bothnia show indications of major landslides in glacial deposits. They might represent one or several earthquakes in connection to the sudden load off during the deglaciation. However, there are no seismic soundings carried out at the site so evidences for this suggested origin are lacking. Further research is thus necessary to increase the understanding of the significance of these structures. Knowledge of any post-glacial earthquakes that have or may possibly occur in the vicinity of a possible final repository for spent nuclear fuel is of high interest for SSM in the continued review of the application for a repository since large earthquakes may induce secondary fault movements in the final disposal volume which could damage deposited canisters.

Project information
Contact person SSM: Lena Sonnerfelt and Carl-Henrik Pettersson
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Contents

Summary

The presence of an ice sheet can exert numerous influences on the ground beneath; the weight of the overlying ice can force isostatic adjustments, and meltwater at an ice sheet bed can affect groundwater flow, ice velocity and the potential for erosion of the substrate. Collecting data on ice temperature, ice velocities, hydrology, ice geometry and meteorological conditions contributes to improving knowledge of physical processes associated with glaciation. The above-mentioned variables can be explored in contemporary ice sheet and glacier settings, but in order to understand past glaciations proxy data from glacial geomorphology must be also be applied. In this report we discuss the analysis of glacial geomorphological data from the northern Stockholm Archipelago and the sea floor of the Gulf of Bothnia, in addition to temperature and geometry changes of contemporary Swedish glaciers.

Studies on a selection of glaciers in High Alpine Sweden reveal that though the temperature distribution adjusts to changes, the speed of change cannot match a quick recession as was recorded during the deglaciation of the Weichselian Ice Sheet. The ice remained cold in analogy with the present Greenlandic Ice Sheet. The colder Antarctic environment is more in analogy with the maximum phase of the Weichselian glaciation. Basal temperature conditions reflect long term changes and are thus more influenced by the maximum phase than by variations in climate during deglaciation. At a late stage of deglaciation, meltwater may have warmed up bed temperatures.

Mapping of bathymetric data provides evidence of significant subglacial water pathways and fast flowing ice along the east coast of Sweden, with further evidence for Baltic Ice Stream onset on the present day Ångermanland-Västerbotten coast during the Late Weichselian. The existence of a Baltic Ice Stream has been shown through modelling experiments to be a necessity to reproduce the Last Glacial Maximum extent in the Baltic sector of Scandinavian Ice Sheet, as inferred by the geomorphological record. However, we believe that an ice shelf or a frozen patch over Åland would be necessary to sufficiently reduce ice flow and thinning in this region, and prevent premature marginal retreat.

The geomorphological landform record, with an impressive array of meltwater channels and eskers of a multitude of spatial scales, suggests that meltwater was plentiful in the Baltic/Bothnian sector during deglaciation. A numerical model was thus used to simulate the Weichselian evolution of the ice sheet and its sensitivity to surface meltwater inputs. The proportion of basal sliding that can be attributed to surface meltwater-enhanced basal sliding varies over space and time, and results of ice sheet modelling highlight the difficulty of implementing the surface meltwater effect through simple parameterizations. This stresses the importance of continued development of physically-based models of surface-to-bed meltwater transfer, needed to better simulate the spatial and temporal variability of

hydrology and dynamics, and their response to changes in climate and meltwater production, within ice sheet models.

The primary aim of this study was to improve our understanding of processes acting underneath an ice sheet, specifically during a deglaciation. Based on the results of studies on present day High Alpine glaciers, the Antarctic and Greenland ice sheets, and glacial geomorphological mapping from land and marine-based data in the Gulf of Bothnia, we have gained an insight into how future glaciations may affect the northeast coast of Uppland, eastern Sweden. Evidence of ice streaming over a large distance in the Gulf of Bothnia, including the presence of numerous glacial lineations of varying scale, contests to a period of fast basal sliding during the Late Weichselian deglaciation. This suggests that there would have been a degree of erosion at the ice sheet bed during this time.

The evidences for a Baltic Ice Stream are linked to the long term existence of the Gulf of Bothnia and the Baltic. Unfrozen soft sea bed sediments allowed a fast sliding speed at a stage of glacier advance and the bed remained wet all through the glaciation. Thus there are good reasons to assume that coming glaciations will act in a similar manner as the Weichselian glaciation did.

Sammanfattning

Ett landskap som är täckt av en inlandsis påverkas på många sätt; isens tyngd orsakar isostatiska rörelser, dess smältvatten påverkar grundvattenrörelser och isen kan erodera underlaget. Genom att studera dessa processer i nutid och spår från dåtid ökar vi kunskapsläget kring en inlandsis inverkan på en miljö. I detta arbete har vi studerat relevanta processer som pågår på svenska glaciärer idag samt de spår som den senaste skandinaviska inlandsisen har lämnat i Stockholms norra skärgård samt i Bottenhavet.

Glaciärstudien har främst handlat om hur temperaturfördelningen i isen förändras i takt med kraftig nettoavsmältning och därmed orsakar en förändring i glaciärens geometri. Det visar sig att temperaturförhållandena förändras i takt med avsmältningen, men inte tillräckligt snabbt för att kunna matcha en mycket snabb avsmältning såsom det antas ha skett vid deglaciationen av Weichselnedisningen. Det innebär att isen förblev kallgradig under deglaciationen, trots att det var förhållandevis varmt klimat. Den nuvarande grönländska inlandsisen är en god analog till deglaciationsläget, medan den Antarktiska inlandsisen är mer lik Weichselisens maxskede. De basala temperaturerna under en inlandsis beskriver förhållanden som verkat under mycket lång tid. Ett maxskede av en glaciation sätter därmed ett större avtryck i den basala temperaturen än vad klimatsvängningar under en deglaciationsfas gör. I ett sent skede av deglaciation kan dock smältvatten ha bidragit till en viss uppvärmning av underlaget.

Sammanställda bottentopografiska data från svenska ostkusten och Bottenhavet visar tydliga spår av snabbt isflöde och stor inverkan av smältvatten under Weichselnedisningen. Det framgår med tydlighet att det har funnits en Baltisk isström som kan följas från Höga kusten ner till Ålands hav. Modellförsök visar också att det måste ha funnits en sådan isström för att isen ska kunna uppnå den utbredning som den hade vid olika tidpunkter under glaciationen. Vi tror att geografiska olikheter i bottentemperaturförhållanden har bidragit till hur deglaciationsförloppet har gått till. Samtidigt som en isström har gått genom Södra Kvarken och Ålands hav har sannolikt isen varit bottenfrusen vid nuvarande Åland. Modellförsöken visar att deglaciationsförloppet annars skulle gå för snabbt och ge en isfrontsgeometri som inte överensstämmer med rådande uppfattning.

De geomorfologiska spåren i främst bottentopografidata visar på en rikedom av smältvattenrännor och åsar av olika åldrar. Spåren vittnar om stora smältvattenmängder under degalaciationsskedet. En numerisk modell användes därför för att simulera Weichselisen och dess känslighet för smältvatteninjektioner i underlaget. Den basala glidningen påverkas mycket konkret av smältvattnet från isytan och processen är svår att modellera eftersom snabba slumpmässiga förändringar kan ge stor effekt under ett deglaciationsskede, vilket har tydliggjorts vid fältstudier på Grönland. De bottentopografidata som har insamlats från Bottenhavet beskriver ett avsmältningsförlopp där vi kan följa tidsmässiga och rumsliga förändringar i vattenflödet som på ett väsentligt sätt ökat vår kunskap kring hur en inlandsis delvis styrs av smältvattenflöden.

Huvudmålet med denna studie har varit att öka vår kunskap kring processer som påverkar underlaget under en glaciation med speciellt fokus på deglaciationsskedet. Med stöd av våra studier av nutida glaciärer och inlandsisar samt geomorfologiska data från land och havsbotten har vi fått en ökad insikt i hur en kommande nedisning kan påverka miljöerna vid norra Upplandskusten. Den isström som har gått från Bottenhavet ner till Ålands hav har lämnat mycket tydliga spår i havsbottnen och rimligen medfört en hög grad av erosion. Väster om den djupränna som isströmmen har gått i är de glaciala spåren också tydliga, men där är erosionen svårare att kvantifiera.

Beläggen för att det har funnits en baltisk isström är kopplade till existensen av Bottenhavet och Östersjön. När en inlandsis avancerar ut över dessa sedimenttäckta havsbottnar så har isrörelsehastigheten ökat väsentligt vilket har givit förutsättningarna för ett effektivt isflöde längs detta nord-sydliga stråk och genom deformationsvärme har området förblivit bottensmältande genom glaciationen. Det finns därmed all anledning att tro att kommande glaciationer kommer att ha ett liknande utvecklingsmönster som Weichselnedisningen hade.

1. Introduction

This report compiles research conducted between 2008 and 2015 with a primary focus on 2012-2014. The aim of this report is to investigate processes occurring at the interface between an ice sheet and its substrate. As part of the research we performed numerical modeling, field studies and a range of data analyses, including the mapping of glacial geomorphological landforms from remotely sensed data. Fieldwork was conducted both on present-day glaciers in Northern Sweden, and in the Northern Stockholm Archipelago. We also planned to extend a radar study of bed topography in East Antarctica, but it was not possible within the time framework of this study. However, data already sampled during the Japanese Swedish Antarctic Expedition (JASE)(Fujita et al. 2011) are used in the discussion to explore the effect of present ice sheets on their bed and landforms.

Our study of glaciers in Northern Sweden focus on how the thermal regime of glaciers respond to major geometry changes forced by climate. These glaciers are all situated in a region typified by the presence of permafrost, where glaciers ae normally “warm” spots. The glaciers in our study area are polythermal, which means that they consist of ice both at pressure melting point and ice below freezing. These glacier studies contribute to our understanding of the interaction between climate and the internal ice temperature in a glacier. Research conducted in the archipelago involved glacial geomorphological mappings on numerous islands, which was later extended by bathymetric data from the Swedish sector of the Gulf of Bothnia. The bathymetric data came to us late in this process and is thus not fully analyzed herein. The Gulf of Bothnia is surprisingly rich in traces from the Weichselian glaciation and landforms predating the glaciation also exist. The Weichselian glaciation is used by Svensk Kärnbränslehantering AB (SKB) as a typical glaciation in their research on possible external stresses on a storage for nuclear waste at Forsmark. Figure 1.1 shows values of solar insolation calculated from past and future variations in accordance with the so called Milankowitch cycles. These cycles show when warming and cooling events occurs but they do not show the actual response of the climate system.

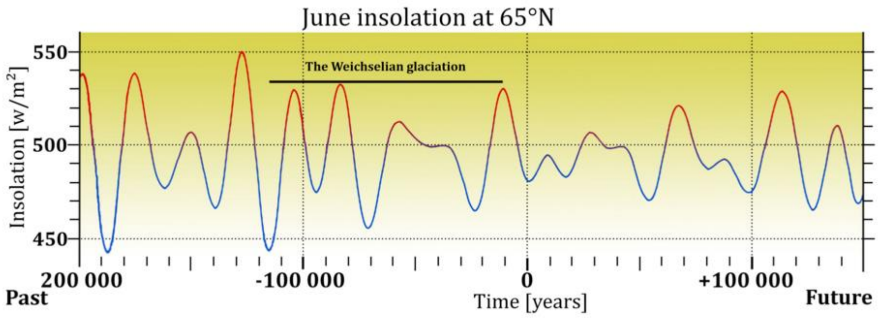


Figure 1.1. Calculated variations in insolation from the sun based on the Milankowitch theory. It includes variations in Earth orbit and angle changes in earth axis. The figure shows past and future variations at 65 degrees north (Berger and Loutre 2002).

The research described herein focusses on glacial processes, with specific focus on meltwater flow, glacial erosion and glacial morphology. The processes investigated here are not well understood and thus we are advancing on existing scientific knowledge. We are using mountain glaciers as laboratories upon which to examine contemporary processes, East Central Sweden as our area of specific interest, and Antarctica and Greenland as analogues for a glaciation in Scandinavia, with numerical modelling to evaluate ice sheet sensitivity to external forcings. By combining results from these scientific and geographic fields our understanding of how an ice sheet affects its substratum during a cycle of glaciation is improved. A glaciation will not only affect the ground by direct glacial processes, but also by periglacial processes like permafrost. The latter is herein only reviewed from other studies.

1.1. Permafrost

Permafrost is widely found in the Sub and High Arctic (King 1984, Harris et al. 2001, Harris et al. 2003). Discontinuous permafrost occurs at annual mean temperatures between -2°C and -5°C, while under colder conditions, permafrost is continuous. Lakes and rivers act as heat sources and constitute warm spots within permafrost areas; so called taliks. Except for the inner part of Greenland, the ground in the High Arctic and Sub Arctic is characterized by summer melt and the formation of an active layer, normally up to a meter thick.

There are several parameters influencing the occurrence of permafrost, such as climate, substrate type, vegetation, exposure, groundwater flow and geothermal heat flux. Climate forcings do not only include meteorological variables such as temperature, precipitation, and wind speed, they also include large-scale temporal and spatial effects such as the seasonal distribution of rain and snow, occurrence of temperature inversions and the influence of topography. But local meteorological conditions can deviate significantly from regional patterns.

Global warming inevitably causes permafrost thawing. Most research on permafrost focusses on changes in its distribution and in the thickness of the active layer. Carbon and methane fluxes from the thawing ground to the atmosphere have the potential to boost future atmospheric warming. A study of past permafrost distribution on the North American continent was carried out by French and Millar (2014). They mapped fossil ice wedges, patterned ground and other landforms associated to permafrost, eluding to a broad permafrost extent of along the margin of the ice sheet during the last maximum of the Laurentide Ice Sheet. Similar mappings have been conducted in Europe, with fossil ice wedges in Skåne dated to the Younger Dryas (Svensson 1964, 1974). Ice wedges are believed to form at annual mean temperatures lower than -6- -8°C (French and Millar 2014).

Physical data on permafrost at drill sites are compiled by Global Terrestrial Network for Permafrost (GTN-p). The majority data points are from shallow holes with depths up to 20 m. Deep drilling has been carried out for the purpose of prospecting for oil or gas, and to a lesser extent for scientific purposes. There are a relatively large number of deep drill holes with depths exceeding the permafrost thickness. Clow (2014) made an analysis of data from 23 boreholes in Alaska for the period 1973-2013. The drill holes were between 227 and 884 m deep and the permafrost depth varied between 210 and 400 m, though most of them had depths around 300 m. The geothermal gradient at the sites were dependent on type of bedrock and climate impact, and varied between 2 and 4°C/100 m (Clow 2014). At some sites it showed variations with depth, which may be an effect on past climatic changes.

Between1998-2002 an EU funded project named Permafrost and Climate in Europe (PACE) was conducted (Harris et al. 2001, 2003). 100 m deep drill holes in rock were equipped with temperature sensors and have been monitored since. There are about 10 sites from Svalbard in the north to the Sierra Nevada in the South, and all indicate a warming trend since installation.

In April 2000, one 100 m and two 15 m holes were drilled into the bedrock close to Tarfala Research Station (Harris et al 2001, Isaksen et al 2001). One 15 m deep hole was drilled by the station at 1130 m above sea level (a.s.l.) at the site of the weather station which has been operating during summer since 1946 and automatically all year round since 1965. At Tarfalaryggen, a saddle area at 1550 m a.s.l., one 100 m and one 15 m deep borehole were drilled and a weather station was erected. The annual mean air temperature is -6°C, the annual precipitation rate is about 500 mm and the permafrost depth is estimated to exceed 300 m. The measured borehole base (100m) temperature was -2.7°C. At Tarfala Research station the annual mean air temperature is -3.9°C, the annual mean precipitation is about 1000 mm and the permafrost is discontinuous. The top 7 m at the site consists of a till cover and the bedrock beneath showed below freezing temperatures when the hole was drilled, but successive surveys have since recorded zero degrees. (Isaksen et al 2001, Isaksen et al 2007).

1.2. Temperature in ice

The temperature within glaciers is more complicated to estimate than the temperature in the ground. Glaciers are moving and snow and temperate ice is permeable to meltwater which releases latent heat. The temperature of ice influences its viscosity, such that an ice body at zero degrees deforms 40 times quicker than a corresponding ice body at -20°C. The temperature also influences the extent to which liquid water is present within or at the base of a glacier, and further influences ice flow by enhancing sliding between the ice and the bed. The ice temperature is thus of fundamental importance for discussions on how a glacier interacts with and affects its substrate.

Ice is a relatively bad conductor of heat. A sinusoidal temperature wave on an ice body needs 5.5 months to reach a depth of 10 m (Cuffey and Paterson 2010). If climate slowly changes, a glacier will adapt its thermal regime to a new situation, but in the case of quick changes the temperature regime will be out of balance with the climate. When the Weichselian Ice Sheet retreated over Uppland the recession rate was on the order of 300 m/year indicating a melt rate (thinning via melting of glacial ice) of more than 10 metres per year. This means that only the very surface of the retreating ice was affected by the warm Holocene climate, while the majority of the ice column remained cold.

There have been some studies on how the temperature regime changes within a glacier with a changing climate. On McCall Glacier in Alaska Delcourt et al (2013) made an analysis of the effect of a gradual change in Equilibrium Line Altitude (ELA) as had been recorded over the last 50 years. The measurements of temperature distribution showed a significant lag in their response. The change in ELA caused a cooling of the glacier as the accumulation area gradually turned into an ablation area, thus no latent heat was released within the ice due to refreezing processes (Delcourt et al 2013). In this report we describe how the thermal regime of selected Swedish glaciers has responded to both changes in climate and geometry.

1.3. Traces of past glaciations in East Central Sweden

Mapping of landforms included field surveys on a number o islands in the North Stockholm Archipelago and analysis of remotely sensed bathymetric data. The bathymetric data, collected by multibeam, are completely new to the scientific community. In Figure 1.2 the areas of bathymetric data used within our reseach are highlighted. The colored area is used specifically for this project while datasets indicated by red lines area areas for which mapping was conducted from multibeam data available at Sveriges Geologiska undersökning (SGU). All data are from Swedish waters and have been delivered by Sjöfartsverket. In addition we have made use of seismic data (not indicated in figure 1.2) provided by, and in cooperation with SGU.

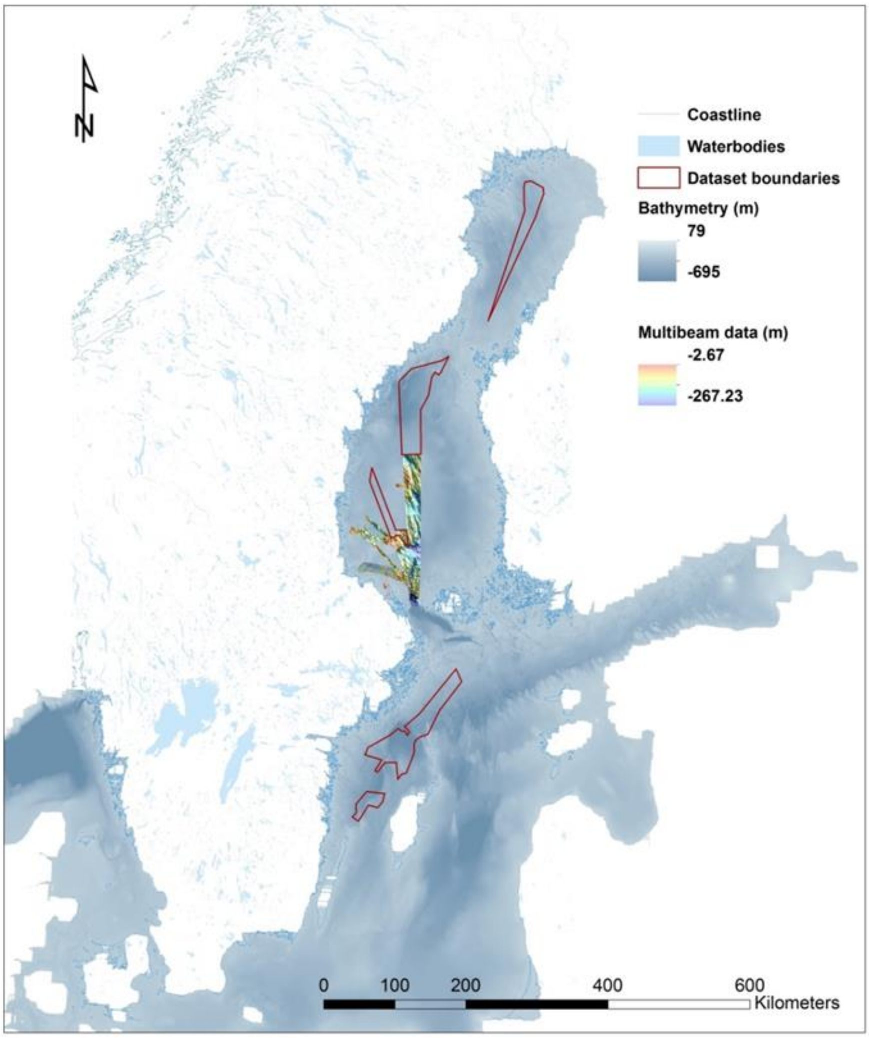


Figure 1.2 The Bathymetric data used for this study. Background bathymetry is from the Baltic Sea Bathymetry Database at 500 m resolution, multibeam data depicted is sourced from Sjöfartverket at 5 m resolution, with areas outlined in red mapped from multibeam data available at SGU. The multibeam data depicted were specifically provided for this study, and are described in more detail in chapters 4 and 5.

2. Spatial and temporal changes of size and thermal regime of alpine Glaciers

2.1. Response to climate change

Glaciers are by definition very good climate indicators as changes in their size and form are forced by climatic conditions. The exponential relationship between ice thickness and ice flow results in responses to small climatic perturbations. However, the response time of Swedish subarctic glaciers are typically 50-100 years (Holmlund 1988a, 1993), so observed length changes are results of not only present day climatic changes, but also climate fluctuations over the last century. Most Swedish glaciers have a polythermal temperature regime, meaning that they are in part temperate and in part cold, or below the pressure melting point (Blatter and Hutter 1991, Holmlund and Eriksson 1989, Pettersson et al. 2003). The relative extent of cold and temperate ice differs between glaciers due to local climatic effects. In the simplest case the glaciers are temperate in their accumulation area and more or less cold in the ablation zone. But in permafrost regions the temperature regime is complicated by ice cored moraines and perennial snow fields which are physically linked to the glaciers, and thus influence their response to climatic fluctuations (Holmlund 1998).

In the springs of 2008-14 radar soundings were performed and more than 80 glaciers were surveyed, some glaciers in much detail and others by a single length profile. The number of glaciers studied make up more than a quarter of all glaciers in Sweden, but the selection of glaciers is biased to relatively large glaciers with historical scientific information on size changes (Bolin 2014). In this report data from these studies are presented and analyzed, including examples of extreme physical settings and responses.

The temperature distribution in a glacier is determined by the pattern of snow accumulation, the rate of mass turnover and the air temperature (Pettersson et al. 2003). In the wintertime the surface of the glacier is cooled to a certain degree and a certain depth. In the spring when melting occurs, meltwater penetrates the snow pack and efficiently raises the temperature to the melting point via energy released from refreezing of meltwater. In the accumulation area of a glacier this process will also warm up past years snow (firn). If meltwater meets an ice surface within or below the snow pack the meltwater cannot penetrate beneath this surface so it refreezes as superimposed ice or flows horizontally across the ice surface. Cooling from the winter thus remains in the ice and over time forms a permanent below freezing point layer. In glaciers with low ice mass turn over, the thickness of this layer is large and in glaciers with a high rate of mass turn over this layer is shallower. A typical value for the cold surface layer depth in Swedish glaciers is 100 m in dry areas and 10-20 m in wetter areas. On glaciers originating from multiple accumulation areas there can be significant spatial variation in thermal conditions.

If a polythermal glacier thins its average temperature is believed to decrease (Pettersson et al. 2003), which may result in small glaciers freezing on to their bed, preventing them from sliding. The glacier may become steeper by this process, but at the same time could hinder the transportation of ice mass to the glacier front. This could in turn cause a dramatic change in areal extent. In the event of a shift to a more favourable climate, a small, steep glacier can reach pressure melting point at the base and thus make a dramatic quick advance of its front position (Murray et al 2000). Glaciers in dry environments may become stagnant, yet remain for a relatively long time. The temperature at the base of such ice bodies is low, due to a lack of melting and water penetration through the ice, and due to the relatively low annual temperatures in the Swedish mountains.

Topography is very important both for the mass balance of and the temperature distribution within a glacier. A steep headwall will provide a glacier with snow to ensure net accumulation in the upper reaches of the glacier. This could also result in a sufficiently deep snow/firn layer in order for the ice to be formed at the pressure melting point. If there is no headwall or a gap in the headwall, much less snow will be accumulated and ice temperatures and thus ice velocities will be low.

2.1.1. Data

Eighty glaciers were studied during several field campaigns distributed over seven years. A continuous-wave stepped-frequency (CWSF) radar system was operated from a helicopter platform with its geographical position sampled by a GPS receiver. The radar soundings were performed in late spring (April) when it is still cold but days are bright and long. The radar system consists of a Hewlett Packard Network Analyzer (8753ET) (Hamran and Aarholt, 1993, Hamran et al. 1995) and log-periodic antenna. The system gives full control of the transmitted bandwidth, but for our purposes we chose to use a center frequency of 820 MHz and a bandwidth of 100 MHz. The radar acquired ~2 records per second giving a trace distance of ~10 m, and the collected data was transformed to the time-domain using inverse Fourier transform with no further filtering. This radar system has been successful in imaging the thermal regime of polythermal glaciers (Holmlund et al. 1989, Björnsson et al, 1996, Pettersson et al. 2003), and the mapped boundary between the cold surface layer and temperate ice is within a accuracy of ±1 m (Pettersson et al., 2003).

The results of these radar studies depict large differences in thermal structure between individual glaciers. All glaciers are below freezing point at their terminus, but the up-glacier longitudinal extension of this cold zone varies significantly; from a few tens of metres up to a kilometer. This highlights the differences in their respective response to climatic changes. A glacier which is frozen to its bed for several hundreds of metres cannot easily advance, while a more temperate glacier, aided by water at the bed, can. Therefore very few glacier advances have been reported from Sweden during the last century, and would require a period of significant cooling or increased precipitation in order to build mass and advance.

2.2. A case study of temperature and geometry changes of five Swedish glaciers

In this chapter we describe the changing geometry and thermal regime of a sample of five Swedish glaciers over time. Changes in geometry were observed from the creation of digital terrain models from data obtained through aerial photography, while changes in the thermal regime were surveyed using ice penetrating radar.

2.2.1. How do Swedish glaciers respond to warming?

Annual variations in glacier surface elevation result from winter accumulation and summer melting, but under unchanging climatic conditions result in the same net ice volume from year to year. While surface expression of response to climatic changes can be immediate, ice volume and terminus response is much slower. A time series of ice surface change of a typical Swedish glacier affected by a step-increase in temperature would show an adaption migrating from the upper reaches of the glacier to the terminus. When the terminus has adjusted to a new climatic situation, a new steady state is achieved. This way of adjusting the geometry is also applicable at ice sheet scale, though the response time is much longer; up to thousands of years. Geometry changes thus help to tell us how close a glacier is to a steady state or whether a glacier mass balance is unhealthy.

Table 1. Total volume change (106 m3) during 1990-2009 for selected Swedish glaciers. Volume changes for Kårsaglaciären were calculated from data in Rippin et al., 2011.

Glacier	Volume 1990/92	Volume 2008/09	Volume change	Percentage change
Kårsaglaciären	$40x10^6$ m^3 (1992)	$18x10^6$ m^3 (2009)	$-22x10^6$ m^3	-55%
Mikkaglaciären	$550x10^6$ m^3 (1990)	$494x10^6$ m^3 (2008)	$-56x10^6$ m^3	-10%
Mårmaglaciären	$440x10^6$ m^3 (1991)	$404x10^6$ m^3 (2008)	$-36x10^6$ m^3	-8%
Pårteglaciären	$880x10^6$ m^3 (1992)	$721x10^6$ m^3 (2008)	$-159x10^6$ m^3	-18%
Storglaciären	$310x10^6$ m^3 (1992)	$308x10^6$ m^3 (2008)	$-2x10^6$ m^3	-0.6%

2.2.2. Ice surface elevation surveys

Digital elevation models (DEMs) were used in this study to quantify the long-time volume variations on Pårteglaciären, Mårmaglaciären and Mikkaglaciären. By calculating the difference between two or more DEMs separated in time, it is possible to obtain a measure of the elevation change. This method is appropriate for long-term studies of glacier changes (Klingbjer and Neidhart, 2006). The DEMs allow calculations of the mean net mass balance, although the results do not provide any information on annual variations in accumulation or ablation. The correlation between traditional mass balance measurements and geodetic mass balance calculations have been found to be generally good in several studies (Holmlund, 1987; Krimmel, 1999; Klingbjer and Neidhart, 2006).

Elevation datasets from the National Land Survey of Sweden were used to create the DEMs. The elevation was calculated digitally from aerial photos from around 1960, 1990 and 2008 taken at the end of the melt season. From the DEMs the volume changes were calculated for the time periods available. The original elevation data were imported and converted to raster format in ArcGIS. The elevation rasters were subtracted from each other to obtain the change in elevation between the datasets. Rasters of the surface change were cropped based on the extent of each glacier for each year and the area changes, volume changes and net surface lowering were calculated using the Zonal Statistic Tool.

The bottom topography of the tongue of Mikkaglaciären was mapped using a 8 MHz pulsed Mark II radar with dipole antennae, and the data recorded on photographic film, in May 1983 (Holmlund 1986). Mårmaglaciären was mapped using the same equipment and method in 1984 and 1987. The profiles were matched to maps constructed based on aerial photographs from August 1980 and August 1978 respectively. As surface maps and ice thickness surveys are close in time the bottom maps are believed to be reliable. The bed of Storglaciären was mapped in 1979 (Björnsson 1981) and resurveyed during the 1980s (Eriksson et al 1993). Eriksson et al. (1993) made use of a map based on aerial photos from 1980 to create a bed topography map. In these two studies, the temperature distribution was mapped using a continuous-wave stepped-frequency (CWSF) radar operated from a helicopter platform and positions were sampled by a GPS receiver.

2.2.3. Physical settings and analyses of the glaciers

2.2.3.1. Mikkaglaciären

Mikkaglaciären, 67°25´N, 17°42´E, is a polythermal valley glacier with two accumulation areas. It was documented and mapped for the first time in 1895 (Hamberg 1901). Its current size is about 7 km^2 and its volume is estimated to be 0.5 km^3. It extends vertically from 1000 to 1825 m a.s.l., and is 150-160 m at its maximum ice thickness. The glacier tongue flows through a U-shaped valley with a smooth longitudinal profile. The glacier has been documented since 1895 via surveys of front position, velocity and ablation, and through the use of terrestrial and aerial photography. The balance gradient of Mikkaglaciären, which is the rate of change in net balance with elevation, has been estimated as approximately 1 m water equivalent (w.e.)/100 m (Holmlund 1986). Surveys of the ice surface and front position indicate a dramatic retreat of Mikkaglaciären over the last 100 years. With a steady front recession rate of approximately 15 m/year during the 20th century it has increased to 25 m during the last 15 years. The mass change since 1960 shows a decline in mass loss with time and the last period 1990-2008 shows no increased trend in ablation. The total mass loss 1960-2008 was 0.16 km^3.

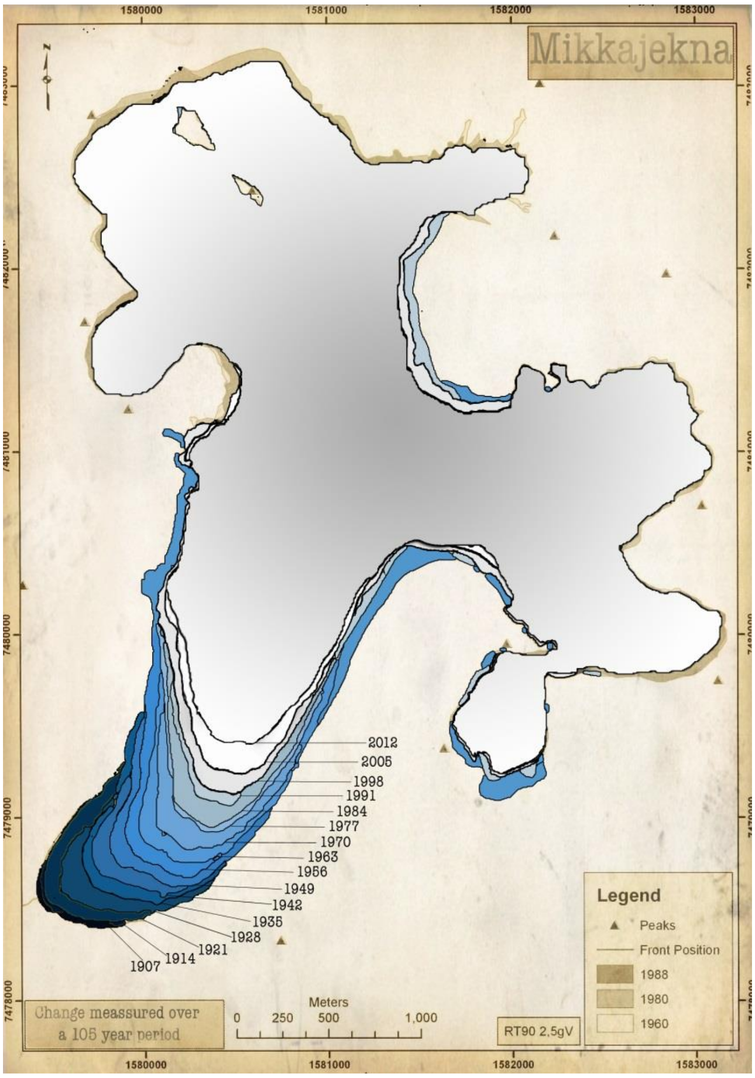
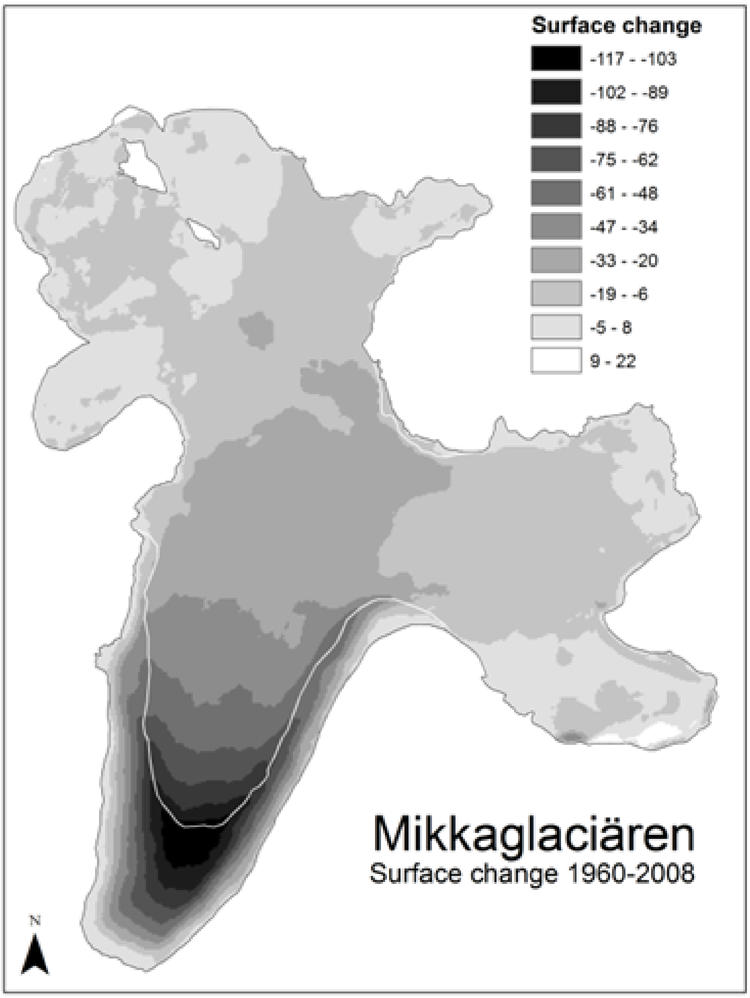


Figure 2.1.The frontal recession of Mikkaglaciären between 1907 and 2012 (Holmlund 2012).



Figure

Figure 2.2. Surface elevation change of Mikkaglaciären between 1960 and 2008, indicating dramatic thinning over the tongue, with more modest surface lowering in the accumulation zone.

Between 1895 and 1901 ice velocity surveys were conducted along transverse profiles on the glacier tongue (Hamberg 1901 and 1910). In 1983 the tongue was radio echo sounded and a surface map constructed based on aerial photographs from 1980. These two datasets allowed for the calculation of balance ice flux, from which it was determine that in 1900 the glacier was close to its maximum, but was far from a balanced state, as the balance flux exceeded the surveyed ice flux by 3-4 times in the upper part of the glacier tongue (Holmlund 1986). The mass balance gradient needed to support this long, slowly moving ice tongue is on the order of 0.1 m w.e. /100 m, which is extremely low (Holmlund 1986). Balance flux calculations also showed that the position of the front was not in a stable state, and since the climate during the 19th century was cold and probably drier than the 20th century, this indicates either that there was a delicate balance between different climatic parameters to stop it at the position it stood at a century ago, or that the glacier underwent a surge and the old velocity surveys were carried out during a quiescent period.

High frequency radar soundings of Mikkaglaciären indicate temperate conditions in the accumulation area and polythermal conditions in the lower part of the tongue, with a cold surface layer approximately 20 m thick. At the confluence of ice flowing from the two accumulation zones, the cold layer is extensive, with a thickness of approximately 70 m.

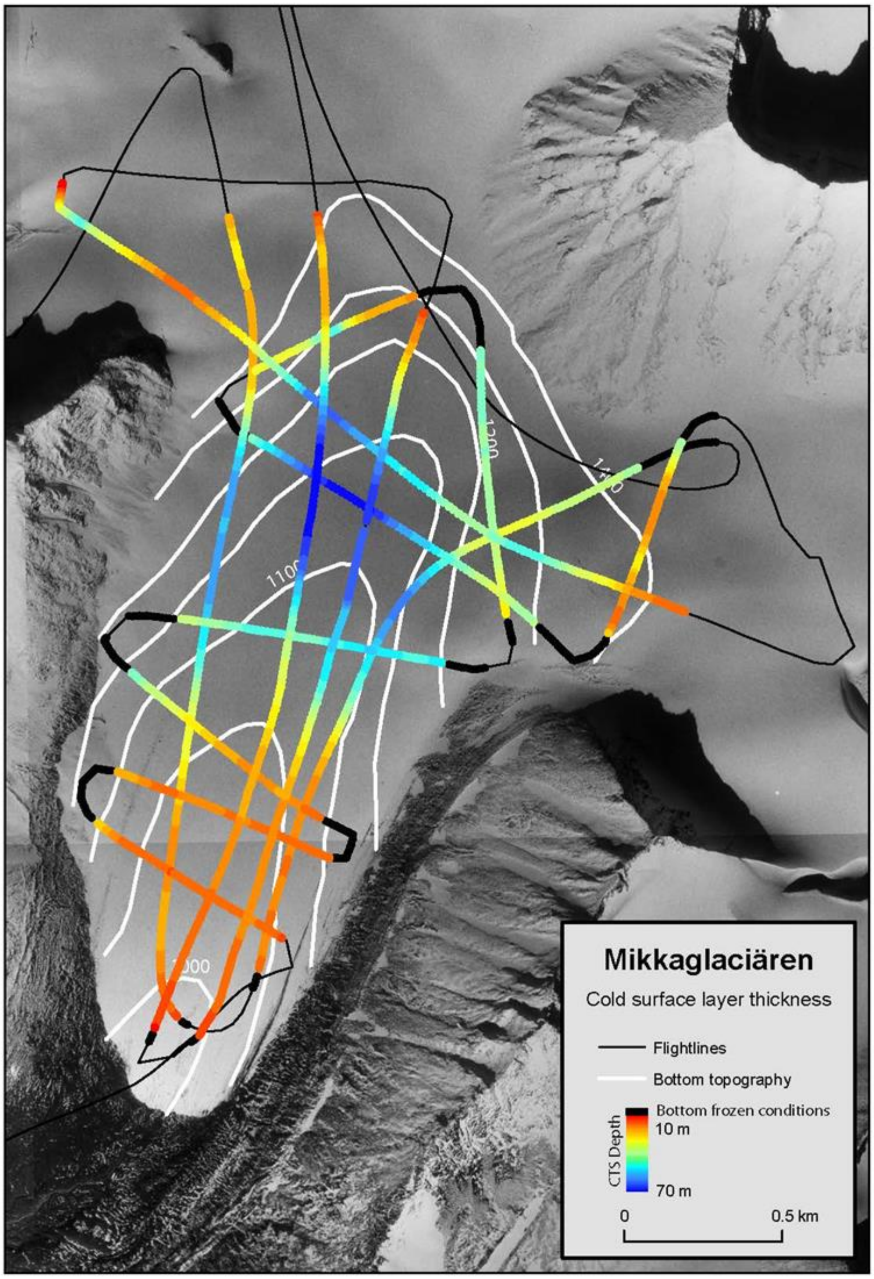


Figure 2.3. Thickness of the cold surface layer as surveyed along flight lines. Figure credit: R.Pettersson.

A hundred years ago the ELA of Mikkaglaciären was situated downstream of the confluence between the two accumulation areas. The ice at this confluence was most probably temperate, with the ice at the bed at the pressure melting point. When temperatures increased during the 20th century the ELA slowly migrated to a position upstream of the confluence, becoming an area containing superimposed ice, and later became an extension of the ablation area. The temperature of the ice at the confluence then cooled, and today it is characterized by cold ice throughout the ice column. This forms an obstacle for ice flow towards the front and acts like a dam to the ice upstream. The lack of ice flux towards the front may then explain the strong response to the recent relatively warm summers. If climate should change such that it promotes glacier growth, the glacier might grow thicker and the ice at the bed of the confluence area might become temperate, creating the prerequisites for a glacier surge (Clark et al. 1984, Murray et al 2000).

2.2.3.2. Storglaciären

Storglaciären is by far the most studied glacier in Sweden, with a mass balance program, the longest in the world, having been initiated in 1945/46 (Schytt 1947). The glacier is located on the east side of the Kebnekaise massif and extends from 1750 to 1140 m a.s.l., covering an area of 3.1 km^2, with a volume of 0.3 km^3 (Bolin 2014, Holmlund 1993, Zempf et al. 2012). Photographic documentation of Storglaciären began in 1886, and the glacier front was surveyed for the first time in 1897 (Svenonius 1910). The thermal structure of Storglaciären was mapped for the first time in 1989 (Holmlund and Eriksson 1989) and a time series of thermal changes was analyzed by Pettersson et al (2007).Terrain models have been constructed based on aerial photos from 1959, 1969, 1980, 1990, 1999 and 2008 (Zempf et al 2012), and radar surveys have been conducted frequently since 1989.

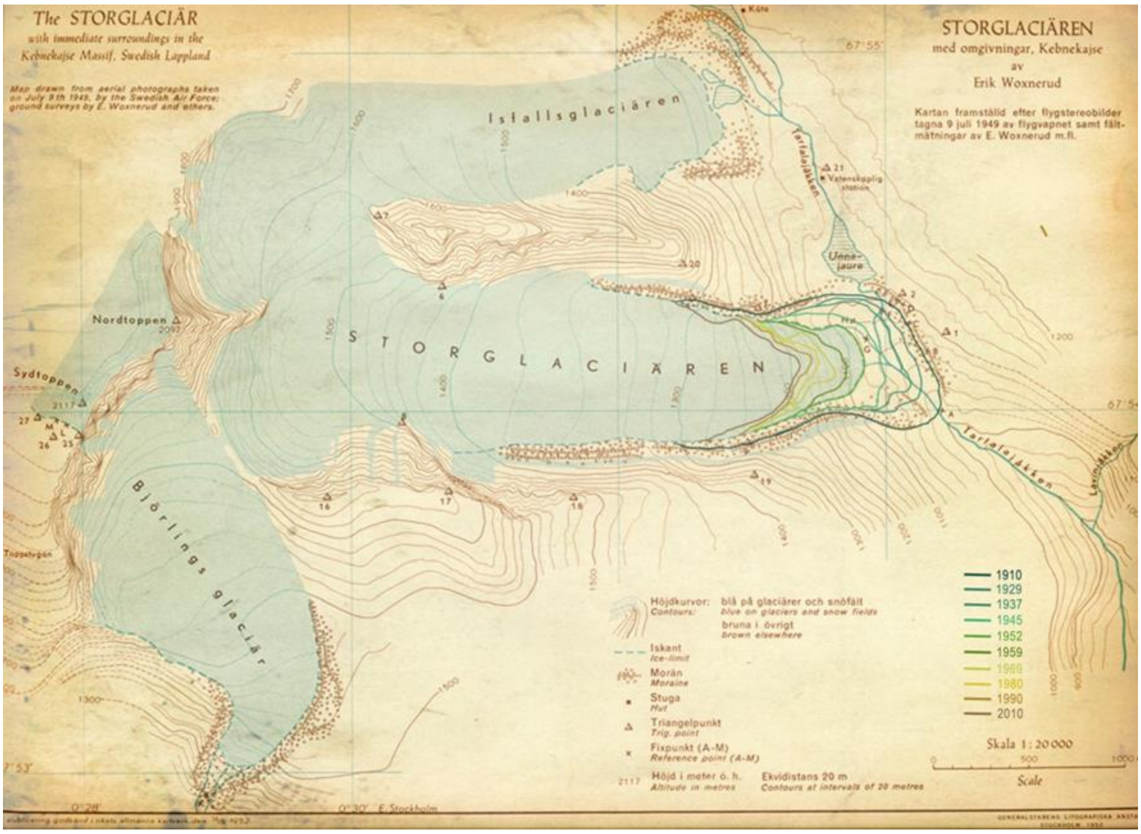


Figure 2.4. The recession of Storglaciären between 1910 and 2012 (Holmlund 2012).

Both mass balance data and front position surveys indicate modest changes during the last three decades. In this sense the glacier is not really representative of the general change in Swedish glaciers. The glacier was close to a steady state in the late 1980s (Holmlund 1988) followed by a period of growth which lasted until c.1996. This has since been followed by a gradual pattern of thinning, albeit somewhat less than that experienced by many other Scandinavian glaciers (WGMS 2012). The resurvey by Pettersson et al. (2007) of the cold surface layer thickness indicated a general thinning of 10 m between 1989 and 2007 (0.55 m/year), with a change in average depth from 30 m to 20 m. Over that period the ice surface experienced negligible thinning, thus the study concluded that climatic warming had forced the change in cold surface layer depth between 1989 and 2007 (Pettersson et al. 2007).

2.2.3.3. Mårmaglaciären

Mårmaglaciären is located in the eastern Kebnekaise area, an area typified by a relatively dry climate. Mårmaglaciären is east facing, ranging between 1350 and 1740 m a.s.l., and is 3.8 km^2 in area, with a volume of 0.4 km^3 (Holmlund 1993, Bolin 2014). The first documentation of Mårmaglaciären are photographs taken in 1918, with more extensive photographic documentation from 1952. The bottom topography was mapped by ground penetrating radar between 1984 and 1987, and the thermal structure in 1995 (Holmlund et al 1996). The glacier was included in the national front surveys in 1968 and field surveys of its annual mass balance have been conducted since 1988. Digital terrain models exist from the years 1959, 1978, 1991 and 2008. Radar mappings were executed in 1995 and 2011-14. Between 1918 and 1959 the total frontal recession amounted to100 m, as the tongue was dammed by ice-cored moraines. When the thinning snout reached the flat bed floor the recession rate increased significantly and is presently about 15 m per year.

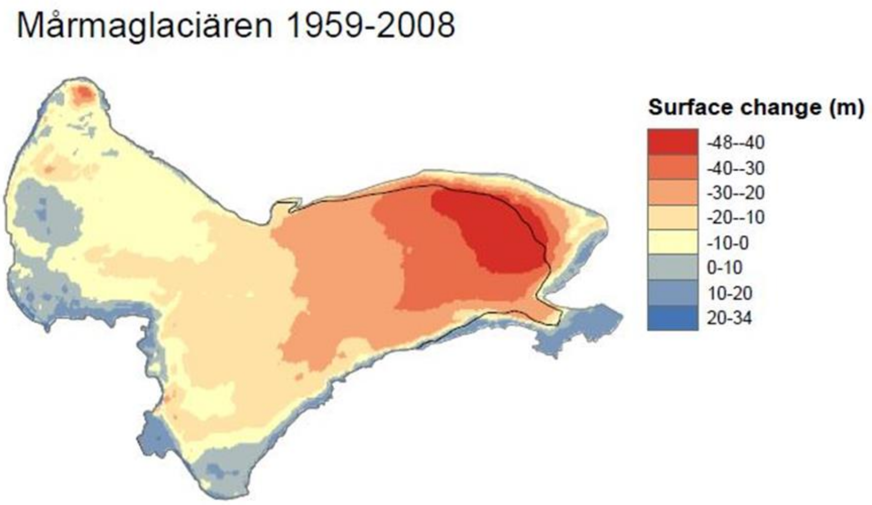


Figure 2.5.Thinning of Mårmaglaciären between 1959 and 2008. The 2014 front position is depicted in black.

The cold surface layer was mapped along a series of radar profiles in 1995 (Holmlund et al. 1996), and between 2010 and 2014. The radar profiles from 1995 are too sparse to make any quantitative comparisons with the present state, but the longitudinal profiles indicate a general thinning trend. In the central part of the tongue the average depth was 74 m in 1995 and the corresponding depth in 2014 was 57 m, which implies a thinning rate of 0.9 m/year. However, as indicated in figure 2.6 there is a strong transverse gradient in the thickness of the cold surface layer on tongue and the 1995 survey was not positioned with GPS so it is difficult to make an in-depth spatial comparison between the present day and 1995.

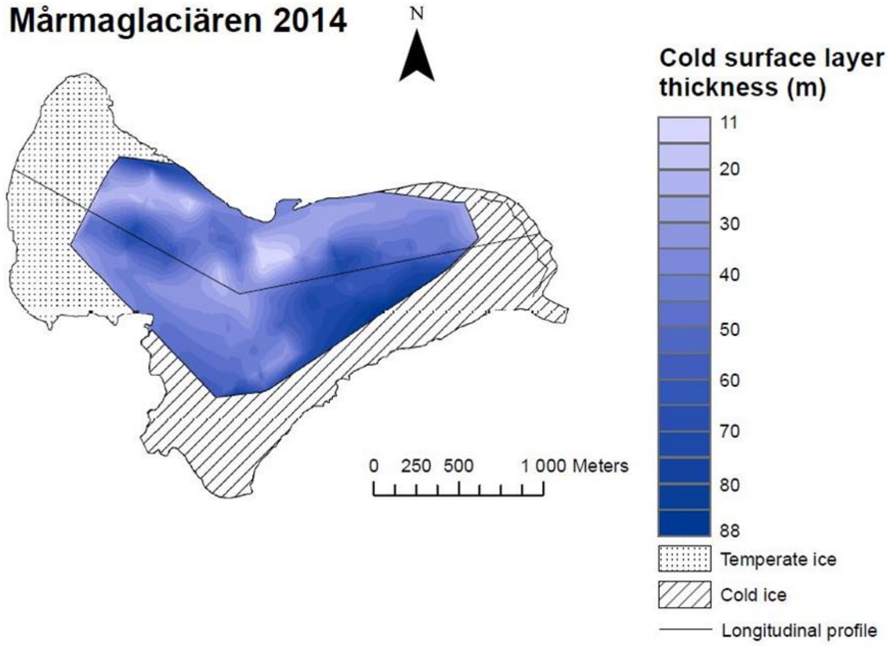


Figure 2.6. The present cold surface layer on Mårmaglaciären as mapped in 2014.

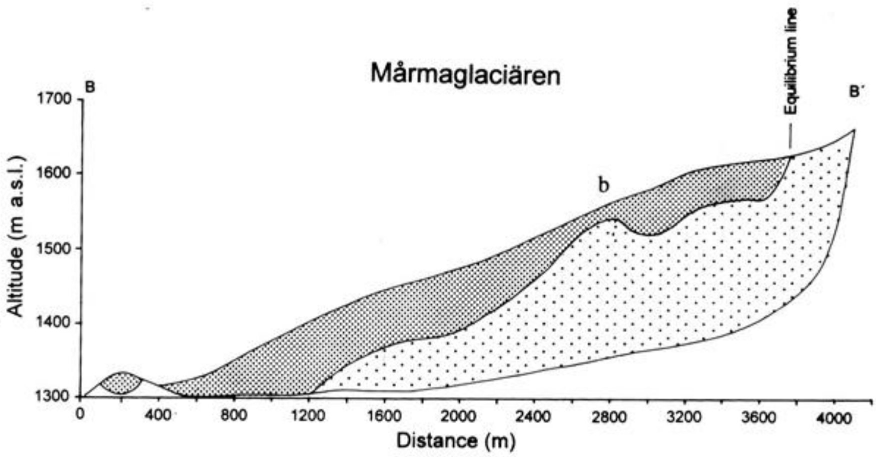


Figure 2.7.The longitudinal profile of Mårmaglaciären from 1995. The grey dotted surface is cold ice and the white dotted area is temperate ice (Holmlund et al 1996).

2.2.3.4. Pårteglaciären

Pårteglaciären is located in the southeastern region of Sarek National Park. It extends from 1860 to 1100 m a.s.l., covering 10km^2 with a volume of approximately 0.8 km^3 (Bolin 2014, Holmlund 1993). The glacier was studied in the 1890s by Hamberg (1901), with mass balance information and photographic documentation existing from this time. In the 1960s it was included in the national glacier front monitoring program, and in 1997/99 a mass balance program was executed by Klingbjer (2001). The annual recession rate 1901-1963 was 4.7 m, and the corresponding rate for 1992-2013 is 15.9 m and 25.1 m for the last five years (Blomdahl 2015). The thermal structure of the glacier was surveyed in 1996 and again between 2010 and 2014. Ice surface digital terrain models exist for the years 1963, 1992 and 2008.

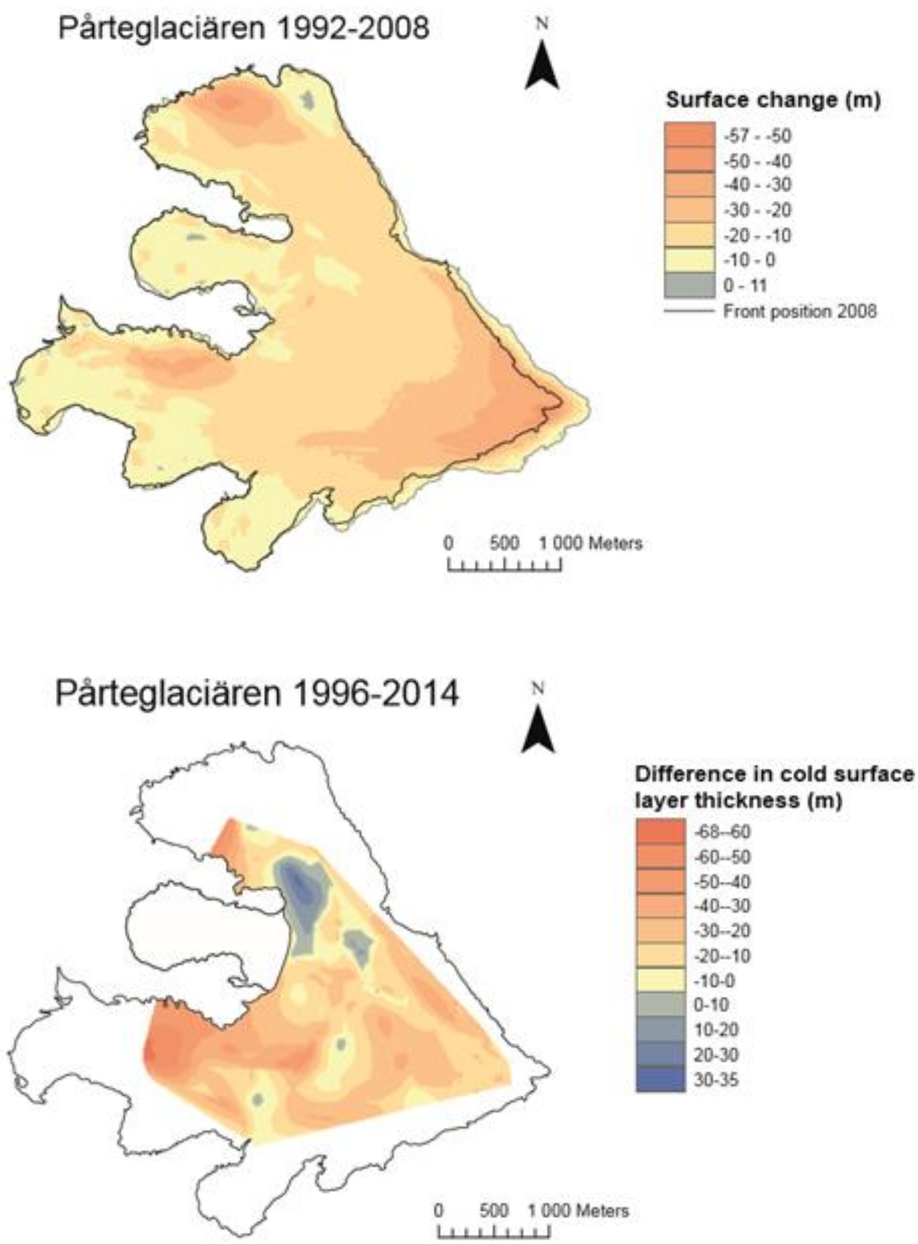


Figure 2.10 The average thinning of the cold surface layer between 1996 and 2014 and ice surface lowering between 1992 and 2008.

2.2.3.5. Kårsaglaciären

Kårsaglaciären (68°21´N, 18°49´E) is situated in the Abisko mountains and has an area of 0.85 km^2 (Bolin 2014). It was first described scientifically by Svenonius in 1884, including surveys of ablation, ice velocities and front position (Svenonius 1910). Mass balance measurements have been carried out occasionally since 1926, when the first detailed map was constructed. At this time the glacier area was 2.63 km^2. The recession since then has been dramatic. In addition to the 1926 map, detailed maps have been constructed for the years 1943, 1959, 1970 and 1992. In 2013 Rippin et al (written communication) constructed a digital elevation model based on laser scanning. They also mapped the thickness of the glacier and the thickness of the cold surface layer (Rippin et al. 2011). The structure of the cold surface layer on Kårsaglaciären is complex as the glacier ice is relatively thin after experiencing major thinning over the last century. The thinning is connected to climate change and explained by an increased winter air temperature since the mid-1980s (Pettersson et al., 2007). Sensitivity analyses indicate that an increase in surface temperature by 1°C would explain most of the observed change in the cold surface layer (Pettersson et al., 2007). An increased air temperature changes the upper boundary condition for the temperature distribution in the ice, which leads to a reduced freezing rate at the base of the cold surface layer.

The cold surface layer on Storglaciären is disappearing in the ablation zone and becoming more temperate, while Kårsaglaciären is losing the zone of temperate ice in the ablation area and consequently becoming colder. Thus there is clearly diversity in how the internal thermal structure of Swedish glaciers will change with a changing climate (Rippin et al., 2011).

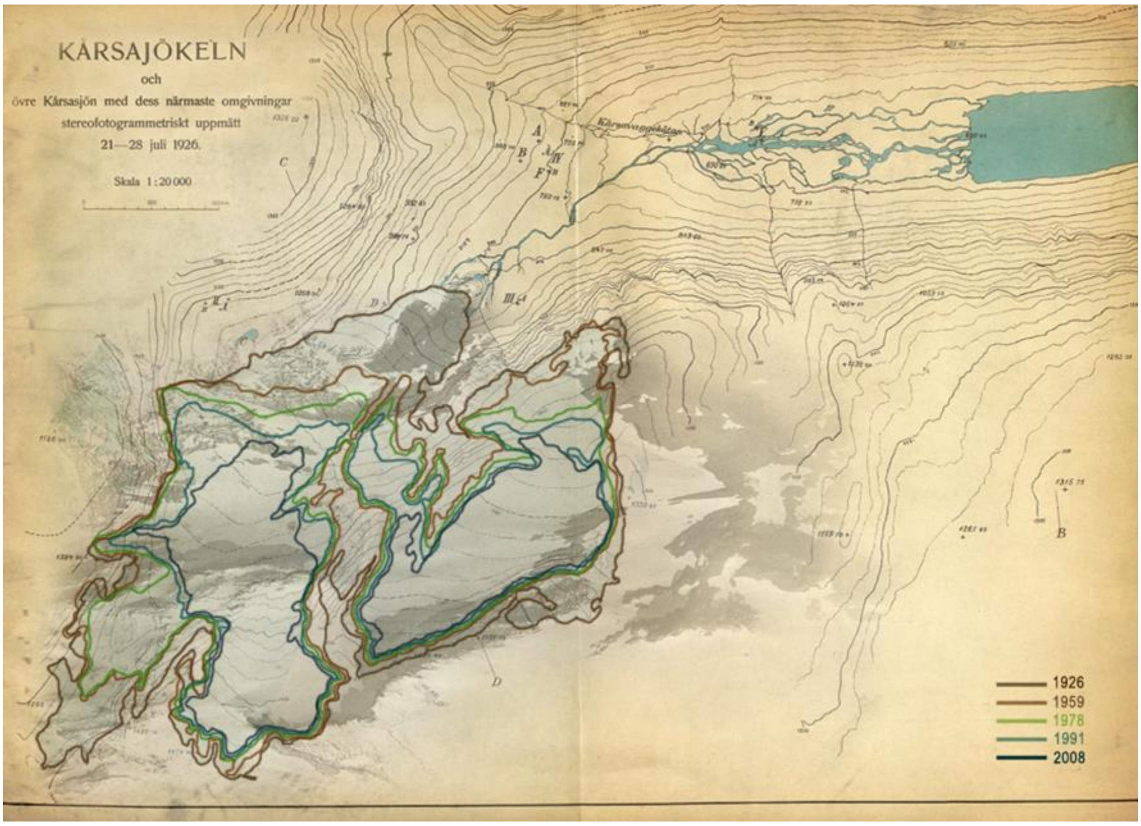


Figure 2.11. The dramatic recession of the Kårsaglaciären between 1926 and 2008. The south-eastern part of the glacier became decoupled from the main part of Kårsaglaciären in the 1950s, and is now a separate glacier named Östra Kårsaglaciären (Holmlund 2012).

2.2.4. Discussion

Rippin et al (2011) suggested that small glaciers are not able to adjust their temperature distribution if they are thinning rapidly. Thinning may thus lead to an average cooling of the ice body, as the cold surface layer remains while the temperate part is vanishing. Kårsaglaciären lost more than 50% of its mass between 1990 and 2008. Pettersson et al. (2007) showed that the change in thickness of the cold surface layer on Storglaciären was caused by climate during the same period, when the glacier lost less than 1% of its mass. These two studies are basically supporting each other, but the question raised herein is what happens during major geometry changes on large glaciers which are capable to match cooling with ice flux? In this study we have added another three glaciers with data on thermal distribution and glacier geometry and their changes over time. The thermal structure of Mårmaglaciären and Pårteglaciären were mapped 1995-96 and again 2013-14. Over this period they have lost a significant amount of ice mass and the cold surface layer has thinned on both glaciers. The spatial pattern of the thinning is evenly distributed over the glaciers, while the geometry changes show a strong correlation with the atmospheric temperature lapse rate causing large ice mass losses at low elevation and modest thinning at high elevation. Our interpretation of the observed temperature regime change is that it is rather caused by climate warming than by thermal/physical effects of the ice mass loss. We may also conclude that the thickness and spatial pattern of the cold surface layer responds sensitively to short term changes in climate. On Mikkaglaciären time series on the temperature distribution are lacking, but are existent on glacier geometry changes. The extent and thickness of the cold surface layer was mapped in 2010-12, but though older data are missing the pattern of the cold surface layer indicate a significant late thickening caused by a gradual increase of the ELA, transforming a gently sloping part of the accumulation area into an ablation area. This area with ice at below freezing point temperatures, act now as an obstacle for ice flux from the accumulation area to reach the snout which has led to a significant increase in recession rate. We have seen changes in thermal structure on all glaciers and the changes in cold surface layer caused by climate is on the order of -1 m a^{-1}, similar in number but spatially different to the ice thinning over the time periods studied.

3. Spatial and temporal changes of ice sheets

3.1. Background

The temperature distribution within a glacier or an ice sheet exerts a control on flow dynamics and hydrology. The ice temperature also affects the glacier hypsometry, interaction with the bed and the interaction between the glacier and groundwater (Cuffey and Paterson 2010).

In a maritime environment, such as the western part of Norway and the south coast of Iceland, the precipitation and ablation rates are high, giving high rates of mass turn over (WGMS 2008). The mass surplus of snow in the accumulation area is heated by refreezing melt-water every summer, forming temperate ice. In the ablation area the ice surface canbe cooled by low winter temperatures but this frozen top layer melts off each year in the summer, allowing the glacier to remain temperate. A temperate glacier is believed to be permeable to melt water. Water drains in a dendritic englacial system, with small veins close to the ice surface and successively larger conduits with depth as a consequence of an increased water-flow and thus increased melt rate (Shreve 1972). Estimations of the content of liquid water in a temperate glacier are often about 1-2%. The direction and inclination of the englacial water flow is governed by the ice surface slope. At the base of a glacier water flow is directed by the force of gravity plus the weight of the overlying ice (Cuffey and Paterson 2010). This simple model for water flow fits well with observations but it implies a system in equilibrium, with a hydrostatic pressure at the bottom equal to the overburden pressure. In practice it would mean that the influence of seasonal changes in meltwater flow is negligible. This is valid for the winter season but probably not during the melt season when the subglacial water pressure drops and has significant variability (Holmlund 1988b). However, studies of polythermal glaciers indicate that the seasonal changes has a minor effect on the general drainage pattern of the glacier (Fountain et al 2005, Holmlund 1988b)

The presence of water at the bed of a temperate glacier allows it to slide over its substratum, with the by-product of erosion. Diurnal and seasonal variations in subglacial water pressure influence the geometry of the drainage system and the sliding speed significantly (e.g. Iken and Bindschadler, 1986). Direct measurement of glacial erosion is difficult, and the most common estimations are based on silt load measurements in proglacial streams. Assuming a uniform erosion rate beneath a glacier and a steady state, silt loads may be transformed into erosion rates but with large degrees of freedom. However, such measurements include sediments derived from parts of the basin which are ice free. In particular, the area immediately in front of the glacier, from which ice may have retreated recently, will contribute with sediment from surface runoff in proglacial areas. This becomes a source of error in estimates of subglacial erosion rates, which is difficult to quantify (Holmlund et al. 1996). The erosion rate for Storglaciären in Northern Sweden is estimated to be approximately 1 mm per year (Schneider and Bronge 1996). Thus, it is reasonable to assume an erosion rate of about 1 mm/year for glaciers in the Scandinavian mountain range.

In drier environments, precipitation and ablation rates are lower giving lower rates of mass turnover (Holmlund and Schneider 1997, WGMS 2008). It is common for the winter cooling depth to exceed the depth of summer melting causing a permanent sub-frozen top layer in the ablation area of the glacier (Holmlund and Eriksson 1989, Björnsson et al. 1996). Depending on winter temperature and summer melt rates, this cold layer can be more or less thick on such polythermal glaciers. In a dry polar climate the entire ice body may be at a temperature well below the freezing point, though melting occurs at the ice surface during the summer. These glaciers are referred to as subpolar. However, as melting occurs during the summer, it is common that temperate ice is produced at the higher reaches of the glacier while the entire ablation area is below the freezing point and frozen to its bed (Björnsson et al. 1996). A glacier which is entirely below freezing point temperature is herein assumed not to slide over its substratum with no liquid water present at the interface between bedrock and ice. Striae formation may still occur via mechanical erosion if basal ice is debris rich (Shreve 1984).

As ice below the freezing point is impermeable to meltwater there is no englacial drainage in cold glaciers except for surface drainage channels which can, if they carry large quantities of water, melt down into the ice and form conduits called cut-and-closure channels (Gulley et al., 2009). In the coldest environments, such as on top of the Greenland and Antarctic ice sheets, no melting occurs and thus the ice surface temperature remains equal to the local annual mean temperature. This is the high polar environment from which deep drilling projects have yielded comprehensive environmental and climatic data. This is also the environment which best corresponds to glacial maxima of the Scandinavian Ice Sheet during the Weichselian. The temperature regime within such an ice sheet is a function of the air temperature at the surface, the vertical movement, advection of ice, heat conduction, internal friction and the geothermal gradient. If there was no ice movement we could simply extrapolate the geothermal gradient from the bed beneath through the ice (Cuffey and Paterson 2010), but this situation is unrealistic for most part of a glacier or an ice sheet. It is unrealistic because the mass balance is positive on the centre of an ice sheet (to maintain the shape of the glacier), thus producing a vertical movement.

The temperature regime governs not only the erosion capacity of a glacier but also the viscosity of the ice, which influences ice flow through deformation. Ice at the pressure melting point deforms ten times more quickly than ice at –20°C (Cuffey and Paterson 2010). Physical factors which influence basal temperatures are: precipitation rates, ice thickness, geothermal gradient, ice movement, and ablation rates. Dissipative heat production via internal friction becomes increasingly important with depth since deformation gets higher with depth due to increasing shear stress and frictional heating at the bed. On the Greenland ice sheet, ice is formed at the surface at temperatures 20-30 degrees below the freezing point. Where the ice is sufficiently thick it reaches the pressure melting point at its base due to the pressure exerted by the overlying ice. Towards the margins, ice will freeze onto the bed as a result of thinner ice. In the case of ice streaming, the high rate of deformation and enhanced basal sliding via meltwater lubrication produced fast velocities, warming the base far into the interior of the ice sheet, along the flow line of the ice stream.

3.2. Basal conditions in East Antarctica

In our search to understand glacial processes three different approaches are typically taken: mapping landforms from past glaciations; studying glacial processes on contemporary glaciers; and conducting modelling experiments. In Antarctica we have the opportunity to take a fourth approach; using geophysical techniques to study the subglacial landscape during an ongoing glaciation. The East Antarctic Ice Sheet can be viewed as an analogue for glacial maxima of the Scandinavian Ice Sheet. A range of basal conditions exist under Antarctica. Where ice is cold based, fragile landforms such as eskers may exist, preserved under the ice sheet, formed under a climate warmer than present conditions. In other areas of thick ice, basal conditions at pressure melting point allow for the formation of subglacial lakes, which recent studies have shown may have connectivity with other lakes (e.g. Wingham et al., 2006; Wolovick et al., 2013).

In western Dronning Maud Land there is considerable diversity in the glacial environment. Close to the coast there are glacially-eroded mountain ranges with large blue ice areas and supraglacial lakes (Holmlund and Näslund 1994). Some of these lakes have been observed to drain through the ice (Holmlund 1993, Jonsson 1988) in a manner similar to the frequent lakes drainages on the Greenland ice sheet. In between the mountain ranges there are deep tectonic troughs extending to 1000 m below sea level which contain ice streams that are currently operating (Fretwell et al 2013, Fujita et al 2009). Results of numerical ice sheet modelling suggest that the coastal area is very sensitive to sea level changes in a manner similar to West Antarctica (Näslund et. al 2000). A major increase in sea level may cause a collapse of the ice stream(s), with the mountain ranges becoming islands, which could have a dramatic effect on the pattern of ice flow and erosion. Further inland the majority of the land surface is above sea level and the ice sheet thus becomes robust (Fretwell et al 2013).

The concept of an Antarctic landscape formed under a past warm climate was first(described by Hans W:son Ahlmann (1944) when he took part of the photographic information sampled by the German Antarctic expedition in 1938/39 (Ritscher 1942). The relief of the landscape showed that the climate must have been different in the past and this fact linked to the question whether or not the present ice sheet was in a state of change as glaciers in the northern hemisphere did by shrinking as a consequence of climate warming (Ahlmann 1953). This idea became one of the arguments for the Norwegian-British-Swedish expedition in 1949-52 though the analysis of the landscape never was developed further.

In the pre site surveys of the European Deep Drilling Project EPICA large areas of Dronning Maud Land was mapped using radioecho technique for bottom topography and satellite positioning systems for surface mapping, and numerical ice sheet models became available. Mappings of the mountain ranges showed that the mountains no doubt were glacially eroded and modeling experiments suggested that this erosion may be very old, probably originating from the formation of the robust East Antarctic Ice Sheet some 17-35 millions of years ago a fact that opens new frontiers of scientific ideas to be developed (Holmlund and Näslund 1994, Näslund et al 2000).

In the early 1990s an increasing number of evidence confirmed the existence of a subglacial lake near the Vostok base. Old seismic data, new radio-echo data and analyses of the lowest lying ice in the deep ice core clearly indicated the existence of a lake which later also was seen on satellite laser altimetry measurements. The existence of a subglacial lake beneath the Antarctic Ice Sheet was an eye opener and soon the number of potential lakes grew to over a hundred and the present figure is more than 400 lakes. However these lakes are of two fundamentally different kinds. The first one is large lakes such as the Vostok lake which has never been bottom frozen and includes organic life. There are only a few of this kind. The other type of lake is formed secondary when the ice sheet has grown thick enough to reach the pressure melting point at the base. These lakes may not have organic life, but play an important role in the subglacial hydrology and for the formation of landforms.

Within the international project BEDMAP data has been sampled by scientists worldwide to compile a bed topography map of Antarctica (Fretwell et al. 2013). This compilation includes radar data from a variety of missions from pure mapping projects to detailed studies of small sites. A large number of sites have been sited where the bottom topography is flat and the reflection of radio-waves indicates liquid water at the base. These sites may not necessarily be proper lakes, but at least a wet surface. However, the interpretations are often based on only little data. As a general rule of thumb there need to be at least 2500 metres of ice on top and the accumulation rate must be low to make the prerequisites for a subglacial lake to persist.

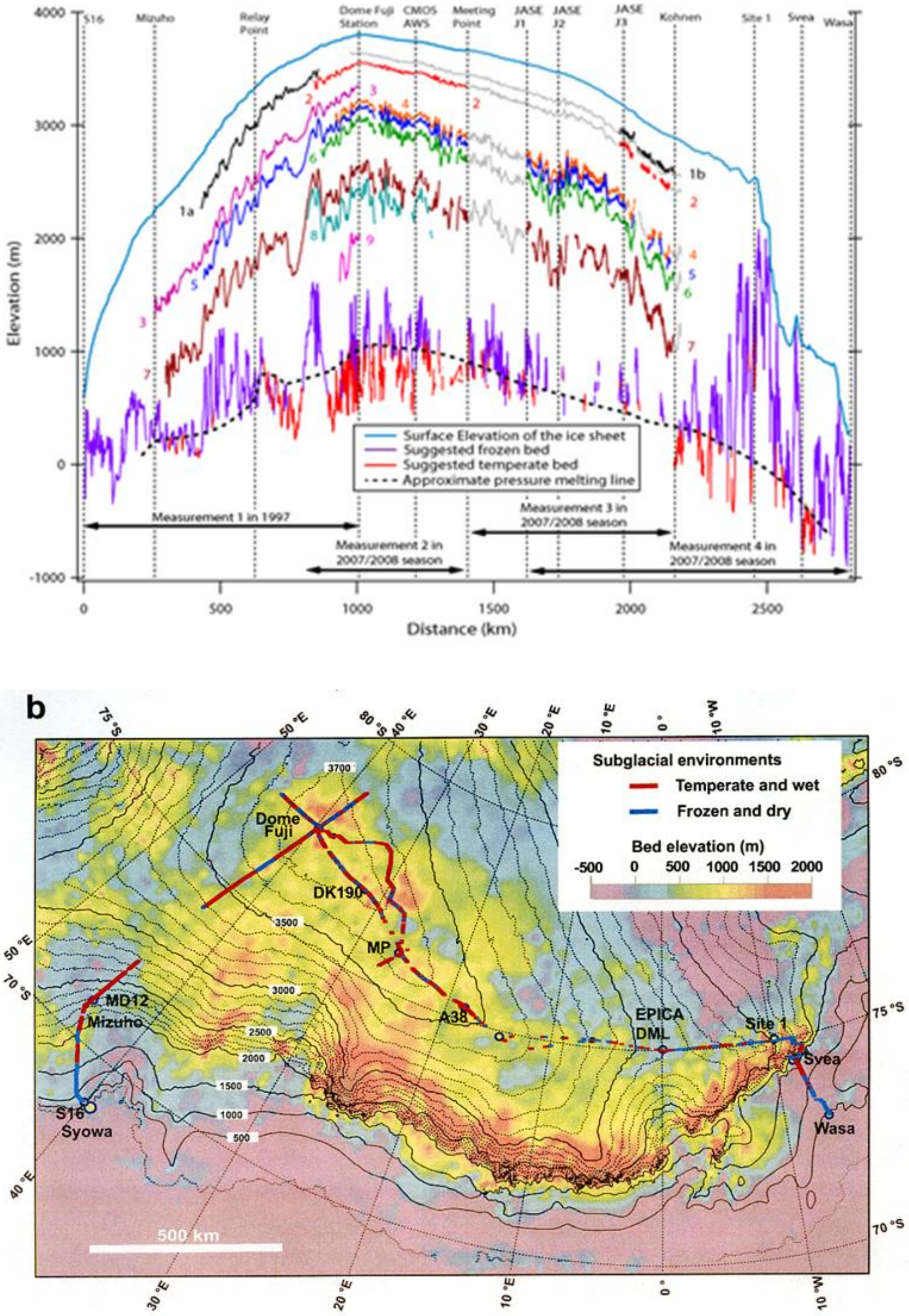


Figure 3.1. Basal conditions along the travelling route of the Japanese-Swedish Antarctic Expedition 2007/08 (Fujita et al. 2012). During this expedition three sites of potential subglacial lakes were investigated and supported earlier indications that they most probably are real lakes. The Amundsenisen site is marked Site 1 in both figures. The dashed line in the upper figure indicates at which depth the ice is believed to reach the pressure melting point. Data shows that along the profile different basal temperature conditions may exist. In the deeper parts these calculations are supported by radar signal processing. The colored surfaces within the ice body are isochrones from distinctive layers interpreted from radar signals.

At some sites such as in the western Dronning Maud Land there are flat basal surfaces seen on radar registrations on sites where basal melting is not expected. On a site at 75°S, 10°W a flat bottom was described by Näslund (1997). Calculations on ice temperatures indicate frozen bed conditions (Holmlund and Näslund 1994) and ice sheet models suggest a surface lowering during ice age conditions (Näslund et al. 2000). The hypothesis is that it is a sediment filled valley and not a lake.

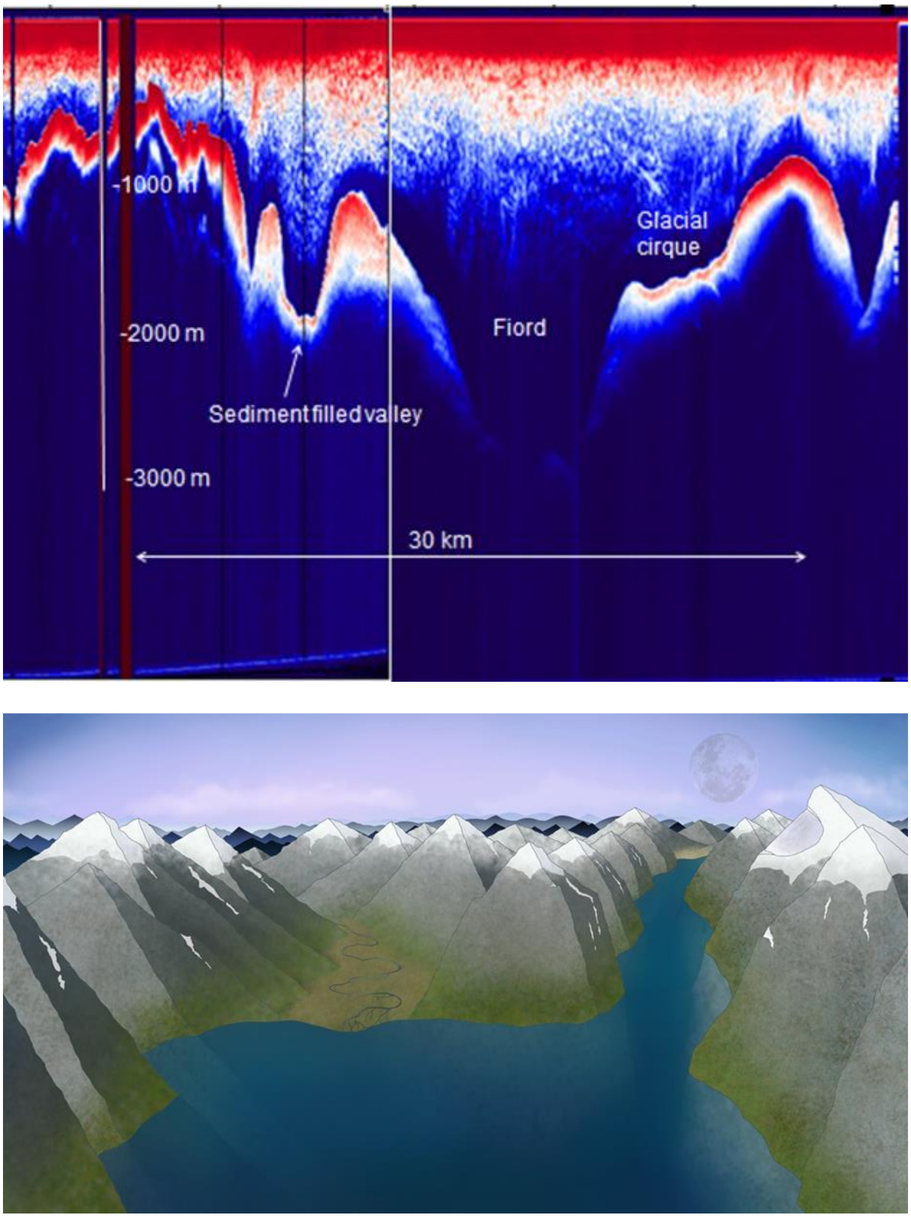


Figure 3.2. The upper figure shows a radargram of a passage of site 1. A monopulse radar at 150 MHz was used from tracked vehicles and it was sampled January 17-19, 2008. The cartoon at the bottom shows what the landscape once may have looked like some 30 million years ago. The most interesting feature is the sediment filled valley shown to the left.

In western Dronning Maud Land mountain ranges act as efficient ice dams and the flow from the inland is concentrated into ice streams. The mountain ranges such as the Heimefront range reaches 2400 m a.s.l. and the ice surface elevation is approximately 2500 m on the proximal side and 1300 m on the distal side. Upstream this range there are other subglacial mountain massifs and ranges that are not visible on the surface. Radar soundings were performed during two days in January 1998 and January 2008 and a surface area of approximately 10x20 km^2 was mapped. The ice surface elevation is 2500 m a.s.l. and the subglacial mountains reach up to 1600-1800 m. On an elevation around 500 m there is a 3 km broad valley with a flat bottom (Fig 3.2). Next to this valley there is a U-shaped fiord with a bottom at approximately -500 m a.s.l. On the summits there are subglacial landforms very much like glacial cirques. If the flat bottom represents a sediment filled valley this landscape has remained more or less undisturbed since the ice sheet once was formed (Fig 3.2).

On the distal side of the Heimefrontfjella mountain range there is a large number of glacier cirques and subglacial indications of a former warmer climate. The Scharffenbergbottnen valley next to the Swedish station Svea was mapped during Swedarp 1987/88 (Holmlund and Herzfeld 1990) and the results highlighted two issues of importance. The valley is approximately 7 km long, 2 km wide and the ice thickness varies from about 300 m in the valley bottom to about 1000 m at the mouth. The valley is glacially over deepened (Figure 3.3). The ice surface is a blue ice area and the surface of the valley bottom is lying 150 m lower than the valley mouth. The present ice flux is directed into the valley as a result of high evaporation rates within the valley. The average annual temperature is -24 degrees. Though the ablation may cause a relative warming of the base there is no reason to believe that the over deepened basin relate to present erosion conditions but rather to something significantly older.

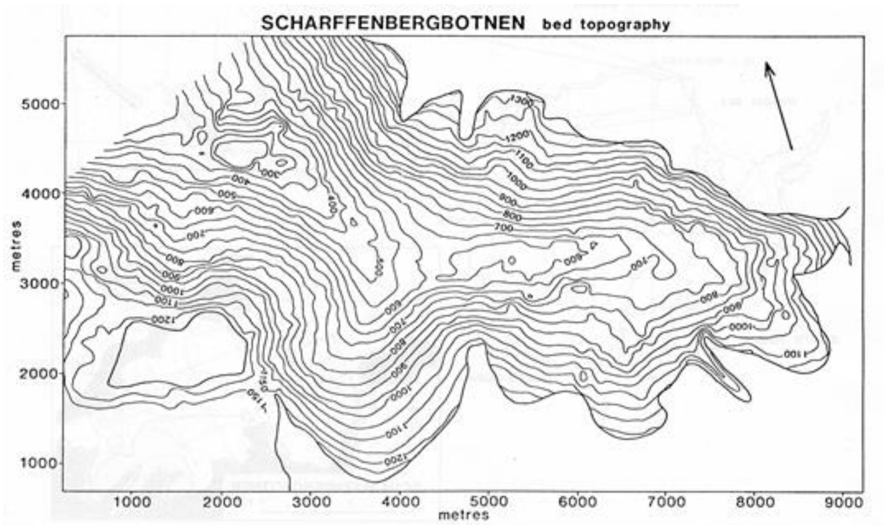


Figure 3.3. Bed topography of the Scharffenbergbottnen in the Heimefrontfjella range, East Antarctica (Herzfeld and Holmlund 1990).

4. Glacial geomorphology in the Gulf of Bothnia, Åland Sea and corresponding shores

4.1. Ice age scenario

Studying the morphology of the deglaciated landscape of Sweden gives an indication of the temperature regime of the former ice sheet. The interior of northern Sweden contains few signs of the effects of glacial erosion, whilst the coast of the Gulf of Bothnia and the Baltic has clear signs (Holmlund and Fastook 1993). In the Västerbotten area, on the north west side of the Gulf of Bothnia, there is a sharp line where drumlins change direction from south east in the inland parts to a more southern direction near the coast, indicating high ice velocities in the Gulf of Bothnia (Eklund 1988). In Tornedalen (north easternmost part of Sweden) the survived traces of fragile landforms indicate cold based conditions (Lagerbäck 1988). On the other hand, the closely located roche moutonées along the Swedish east coast are beautiful examples of both large and small scale glacial erosion. Thus from a glacial morphological viewpoint we may conclude that inland Sweden and, probably, also inland Finland were covered by cold based ice during the maximum phase of the Weichselian. During the phase of recession some areas may have experienced a change in thermal conditions from frozen base to thawed base as the frontal zone migrated backwards (Fastook and Holmlund 1994).

The thermal conditions outlined here compare well with the present day ice sheet of Greenland, for which we can describe the boundary conditions fairly satisfactorily. Thus, the question is; what was the climate like during the last glaciation? Greenland ice cores indicate a climate during the later phase of the Weichselian at least 10-15 degrees colder than present (Johnsson et al. 1992). Assuming these figures are correct, and that the period lasted long enough for an ice sheet to form and to expand from the mountains in north-western Scandinavia towards the south, we will get a high polar ice sheet with a certain ice thickness and extent. Using an adiabatic lapse rate of 0.7-1.0 degrees Celsius per 100 m, we can assume annual temperatures around -40°C at the top center of the ice dome situated over the Gulf of Bothnia. If this estimation is reasonably correct we can conclude that the ice sheet over Scandinavia during the Weichselian was of a high polar type, similar to the present ice sheets over Greenland and Antarctica.

Based on temperature records interpreted from Greenland ice cores, the temperature during the first phase of the Weichselian glaciation may have been 4-5 degrees lower than at present (Johnsen et al. 1992). This likely led to significant increases in the extent of glaciers in the Scandinavian mountain range, with the entire mountain range, except for high mountain peaks, likely ice-covered during the Early Weichselian (Fredin and Hättestrand 2001). At present the glaciers in northern Sweden are polythermal (Holmlund et al 1996). In the event of air temperature cooling of a few degrees, the subpolar character of the glaciers would increase, but the polythermal regime would persist, as the glaciers also would become thicker. Likewise, during the initiation of the Weichselian glaciation, it is probable that there was still temperate ice formation due to summer snow melt, despite the reduction in air temperature, especially in the high mountain areas where localized net accumulation rates may have been large.

Transferring our knowledge of individual glaciers to the entire mountain range during a period of predominantly westerly winds, we can infer that there would be more temperate glacial conditions on the west side of the range and more subpolar conditions on the east side (Holmlund and Schneider 1997). We may then expect that an early Weichselian Ice Sheet would have had more impact on the landscape through erosion on the western side of the mountain range and less impact on the eastern side. Superimposed on this east-westerly trend, there was also a vertical zonation, with areas characterized by basal melting in the deep valleys, and areas with ice below the freezing point at high altitudes. These scenarios are in agreement with observations of preserved old weathering surfaces next to deep glacially excavated valleys.

On the eastern rim of the Scandinavian mountain range there are remnants of preserved ancient landscapes, indicating local frozen bed conditions (Kleman 1992, 1994, Kleman and Borgström 1994). Thus, though glacial erosion was active in the central and western parts of the mountains, inland Sweden may have been characterized by permafrost and frozen interface between ice and bed.

The next phase of the last glaciation was when the ice sheet developed in size but was still smaller than during the glacial maximum. According to the Greenland ice cores, temperatures were only slightly warmer during this phase than during the last maximum phase (Dansgaard et al. 1993). Thus there is no reason to expect anything but a high polar ice sheet to form with frozen conditions under most of the ice sheet. Wet based conditions may have ocurred along the Norwegian west coast, in the Gulf of Bothnia, where large lakes are situated and perhaps some parts along the southern perimeter of the ice sheet.

The third and maximum phase is also characterized by a high polar ice sheet with well below freezing point temperatures throughout the ice body. Wet based conditions may have occurred in the deepest areas, which according to different data sets was centred in the Gulf of Bothnia. The length of the flow line through the Baltic turning west towards Denmark, indicates high flow rates, and the existence of an ice stream (Boulton et al. 1985, Holmlund and Fastook 1994, Torell 1878). This assumption is supported by the beautiful modified landscape which is slowly ascending by land uplift and by bathymetric data from the Gulf of Bothnia and the Baltic.

The temperature regime of the ice remnants during deglaciation was probably quite different than during the growing phase. Ablation rates were high, probably at same order of magnitude as thermal conduction speed in ice. During such conditions an ice body will preserve its thermal regime. A cold ice will remain cold during deglaciation.

In order to study past basal ice conditions in the region of northern Uppland, landforms were analyzed. Bathymetric data was received from Sjöfartsverket, seismic and bathymetry data from SGU and on field studies were performed along the shore from Norrtälje to Örskär. The bathymetric data from Sjöfartsverket cover 7645 km^2 and field controls were carried out along the main land coast and on approximately 50 islands during 2012-2014. These data sets are herein described separately as 4.2 and 4.3. The seismic data provided information of sediment depth.

4.2. Bathymetry of the Swedish sector of the Gulf of Bothnia

4.2.1. Meltwater beneath the Scandinavian Ice Sheet

Glacial landforms documented on both land and the seafloor in Sweden may provide evidence for the transfer of meltwater to the bed of the deglaciating Scandinavian Ice Sheet inland of the ice sheet margin. This meltwater may have originated from the drainage of supraglacial lakes, as have been observed up to and beyond 2000 m a.s.l. in regions of the present-day Greenland Ice Sheet, with the area over which lakes are observed having increased significantly inland in line with warming and rise in the equilibrium line altitude over the past decade (Howat et al., 2013). Strömberg (2010) documented the presence of rare features, referred to as polished flutes, on numerous sites in the Åland Sea archipelago. These flutes, some of which were documented again during fieldwork in 2012 (Figure 5.5), are typically 2-10 mm wide, 5-20 cm long and 2-5 mm deep (Strömberg, 2010). They are smooth in comparison to the surrounding striated bedrock, and deviate in direction from the glacial striae they are found beside and in some cases cross-cut the striae, suggesting that the flutes are the younger feature. We hypothesise in line with Strömberg (2010) than these flutes were formed by a large supply of meltwater delivered to the local subglacial environment within a short period of time, enabling erosion of this magnitude. The drainage of supraglacial lakes to the ice sheet bed or moulins penetrating the full ice thickness (i.e. Das et al., 2008; Catania and Neumann, 2010; Doyle et al., 2013); could provide meltwater volumes capable of forming these meltwater-eroded features. Given that flutes are the youngest features on the land surfaces documented by Strömberg (2010), and were not subsequently eroded by abrasion, their formation was likely close to the ice margin during the late Weichselian deglaciation, during a period rich in surface meltwater production.

Figure 4.1. Examples of polished flutes in the northern Stockholm archipelago (Photos: C. Clason, 2012).

Mapping of glacial landforms was conducted from high resolution bathymetric data collected in the Baltic and Bothnian Basins by a Reson 7125 SV multibeam and a Kongsberg EM2040 multibeam for the Swedish Maritime Administration (Sjöfartsverket). The raw data, delivered in x,y,z (easting, northing, elevation) format, were resampled from variable spatial resolutions using the nearest neighbour method and gridded at 5 m resolution. The grids were stored within a mosaic dataset to allow for consistent visualisation. The data were projected in the ETRS 1989 coordinate system within UTM Zone 33N. To optimize data visualisation for mapping purposes a combination of slope maps and relief shading, derived from the raw bathymetric data, were employed. These maps, depicted in greyscale, were overlain by a semi-transparent layer illustrating bathymetric surface, depicted using a broad colour ramp to best visualize the range in topographic values within the data. Mapping was largely conducted by digitizing the crestline of landforms (or topographic lows in the case of meltwater channels), and stored as polyline features, with the exception of areas of ribbed moraine which were mapped as polygons. Standards employed for the production of geomorphic maps in Sweden are outlined in Peterson and Smith (2013), and have been adapted for mapping of offshore landforms.

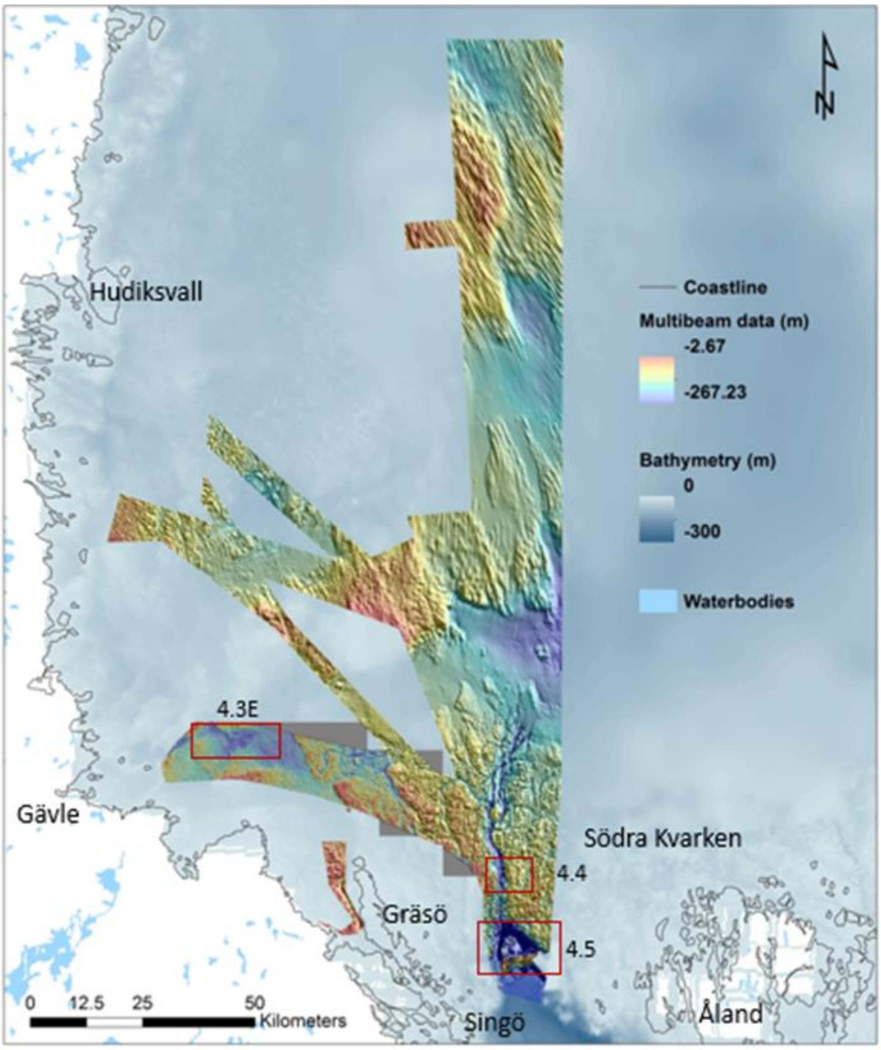


Figure 4.2.Detailed multibeam bathymetric data provided by Sjöfartsverket and used in this study. Background bathymetry is from the Baltic Sea Bathymetry Database.

Mapping of glacial landforms conducted from multibeam data in the Gulf of Bothnia as part of this project reveals a rich meltwater landform record, providing evidence for abundant meltwater in the subglacial environment beneath the Baltic Ice Stream (Figure 4.3). Numerous eskers and meltwater channels are documented, providing evidence for subglacial channelization in at least the marginal areas of this sector of the ice sheet. Mapped eskers are up to c. 17 km in length, but the size of individual data swaths mean that some of these eskers may form part of the same system, stretching for many tens of kilometres. Meltwater channels range greatly in size, with the largest extending up to 4 km in width, some of which appear to exploit ancient fluvial systems extending from onshore Sweden. The dominance of this meltwater record, preserved surprisingly well in comparison to other marine locations, suggest meltwater-driven retreat in the Gulf of Bothnia.

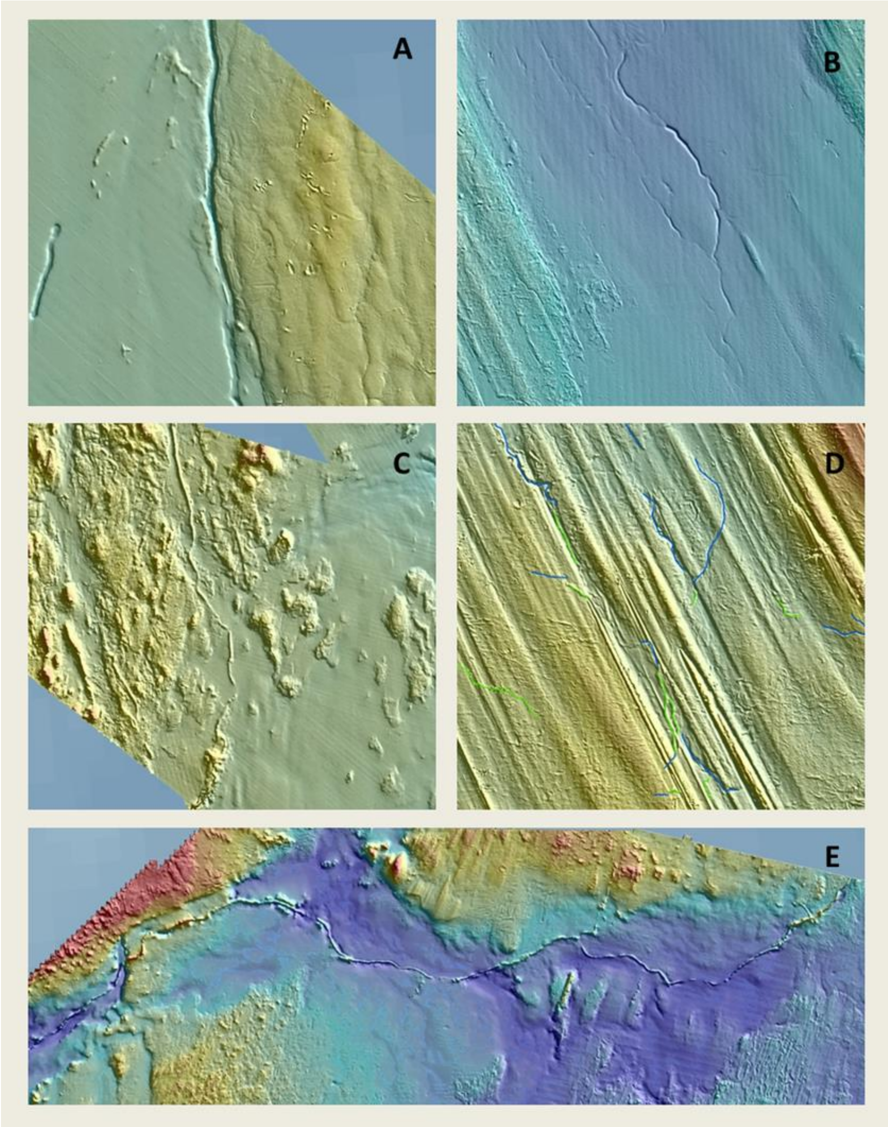


Figure 4.3. Meltwater landforms observed in multibeam data in the Gulf of Bothnia. A) large meltwater channels c. 90 m in width, B) smaller meltwater channels c. 40 m in width, C) an esker c. 12 km in length, D) connectings netoworks of channels (blue) and eskers (green), and E) an esker c.17 km in length.

The pattern of lineation indicates the main ice flow directions. On shore towards the coast in Västerbotten drumlins change in direction from a north west flow to a northern flow. In the Gulf of Bothnia the main ice flow direction has been along the deepest part which means a south eastward direction in the north and south westward in the south. In the Åland Sea the direction is from north to south.

The Uppsala Esker can be followed far north and several smaller eskers are identified in the data set. The esker system between Öregrund and Västra banken (Figure. 4.3E) indicates a dramatic shift in ice flow direction from north to east or west when the ice front had passed Örskär. This is indicated by an east-west oriented esker, but not seen in the lineations. De Geer (1940) suggested a surge like advance from west to east in this area. There is very little support for this idea, but it would explain the appearance of the esker.

Melt water spillways and channels are commonly existing features. An interesting detail is that the river channels of the on shore nowadays existing rivers can be followed far out in the Gulf of Bothnia. In addition, a river channel connects the deep south-eastern part with the Åland Sea.

The large scale scouring is primarily seen in the south where the ice has excavated the through between Åland and Sweden (Figure 4.4). The seismic data indicate that the bedrock is at approximately 300 m depth but the through is now filled with 100 m glacial clay. It is in a tectonic fault line but clearly influenced by glacier ice. Further north, in the proper Gulf of Bothnia, signs of dramatic ice flows are seen in connection to the former river channel draining an ancient Bothnia lake.

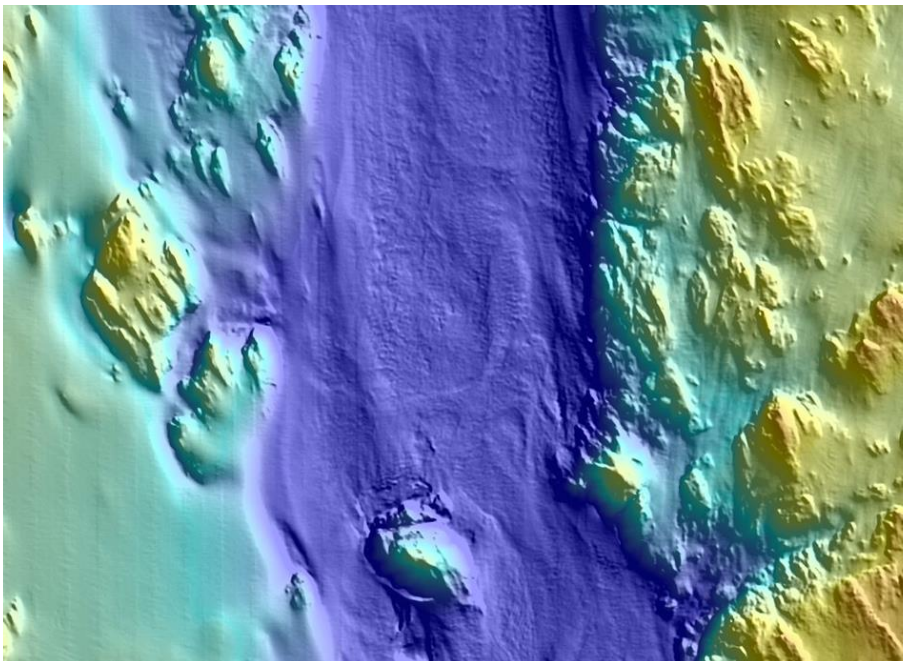


Figure 4.4. Traces of ice stream flow in the ancient drainage channel of the Gulf of Bothnia. The position of this site is indicated in Figure 4.3.

In the northernmost Åland Sea and along the island Väddö there are indications of major landslides in glacial deposits (Fig. 4.5). There are no datings on these events but they might represent one or several earth quakes in connection to the sudden load off during the deglaciation. The larges slide has a 1 km^2 scar and a sediment tongue of similar size. This landslide is on a hill, which together with a neighboring one, have the appearance of rotated rock blocks. However, there are no seismic soundings carried out at the site so evidences for this suggested origin are lacking.

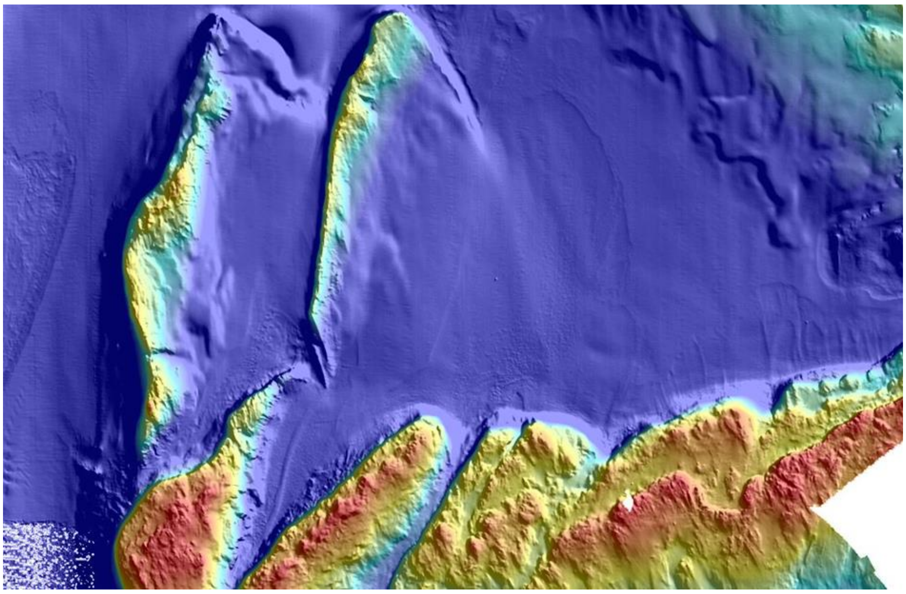


Figure 4.5.Large landslides in Södra Kvarken. Scale is approximately 1:50 000. The blue colour indicate sea floor at a depth of approximately 200 m and the rock out crops in redish brown are just a few tens of metres below the sea surface. The position of this site is indicated in Figure 4.2.

Traces of ice berg calving and ice berg scouring is frequently found, especially on sea floor out of glacial clay. When data on the entire Gulf of Bothnia exists, frequency analyses can be carried out giving input to variability in calving rate but the present data is not sufficiently covering the sea floor.

Moraines of De Geer type is found in the northern Gulf of Bothnia, but are rarely seen in the central and south part. The reason for the low number of moraines is somewhat unclear but is related to the recession speed and the water depth during deglaciation.

4.3. Glacial geomorphology of the Swedish coast

This mapping has been carried out by boating along the shores and getting ashore on islands and on the coast if they looked interesting or if they were located at a site where special features were expected. In total about 50 islands were visited and it was in the archipelagos of Norrtälje, Singö and Gräsö and in Öregrundsgrepen, including the islands outside Forsmark.

The isostatic uplift rate in the area is on the order 50 cm per 100 years and the vegetation free area is roughly speaking the lowermost three meters above sea level. Detailed glacial morphological mappings thus only cover what has been exposed during the last 500-700 years. Rocks exposed earlier are now covered by vegetation and the vegetation on these outer skirt islands not only hide the rocks it is also so dense that it is very difficult to cross by foot. So morphological mapping along coastlines only shows what has been exposed during a limited timeslot. This is a significant limitation.

The deglaciation from the Mälaren to Norra Kvarken has been described by several authors (DeGeer 1940; Hoppe 1961; Lundqvist 1980; Strömberg 1981, 1989, 1990, 2005; and others). The reconstructions are based on clay varves and striations. De Geer (1940) suggested a surgelike advance from west to east between Gävle and Hudiksvall. This suggestion was based on very few data points and it lacks support from other investigations. The data on land shows that a calving bay was formed in south western Gulf of Bothnia with a very fast recession of the front while the recession on the eastern side of the Gulf of Bothnia was slower (Strömberg 1989, 2005).

The Uppsala esker was seen on bathymetric charts up to a position at a level of Hudiksvall (Hoppe 1961; Strömberg 1989).

Lagerbäck et al (2005) made a survey of traces of late- or postglacial faulting in the Forsmark region. They made stratigraphic studies in pits and trenches along several eskers and they studied the bedrock fracturing and quarrying in the area. The overall conclusion was that there was no evidence of any large earth quake, but at several locations the glacial stratigraphy showed distorted sequences of glacial clay interpreted to be caused by sliding (Lagerbäck et al 2005).

4.3.1. Norrtälje archipelago

This excursion was carried out in June 2012. Bo Strömberg who has been working in this area since the 1960s, showed us sites with “polished flutes”. These are depressions in rocks where the bed is striated in a direction deviating significantly from the glacial striaes and rather points towards the glacial trough in the Åland Sea. They are interpreted as water injections from the ice surface to the bottom (Strömberg 2010). This type of erosion marks have been found here, on Åland, and in the northern part of Stockholm archipelago, but never elsewhere. Field surveys were done on Håkansskär, Västerskär and Vattunga (Figure 4.1).

4.3.2. The Singö archipelago

Visits in this archipelago were made in 2012 and 2013. Based on aerial photos a selection of about ten islands were chosen. The island Halsaren was the most fabulous island. It is a cigar formed island in crystalline rock next to the deep trench between Sweden and Åland. It is about 200 m long and 10 m high and the water depth around the island is about 60 m. It is extremely glacially eroded.

The island Storskäret had a moved rock slab, 2 m thick and some hundreds of m^2 indicating movement at a late stage of the glaciation. Understen had a lot of glacial forms, but this island has been heavily modified by the military. The same history applies to the island Måssten. On aerial photographs it looks like a large low lying glacier polished island perfect for glacial geology. But unfortunately it has been a military fortress so most of the island is blasted apart and very little of the former rocks are preserved.

Figure 4.6.The eastern part of the island Halsaren. Photo P.Holmlund July 2013.

4.3.3. Gräsö archipelago

This is a very large and wide spread archipelago which herein is defined as the area within Singö-Örskär. This archipelago is exposed to the Gulf of Bothnia, its bathymetry is poorly mapped and it includes a number of restrictions concerning getting ashore due to wildlife and military, so it is not easy to access. The southern and northernmost parts are easier to reach and were visited in 2011-2012. The central part was visited three times during 2014. More than 30 islands were examined in this part of the archipelago. In general the stoss side of the islands is modulated by the ice and the southern lee side is covered by rocks, so they are classic roche moutonnée . On the outermost islands such as Västerbådan, Klykholmen and Fluttuskär the glacial influence was most prominent with different type of erosion markers. At Västerbådan scars were found, similar but not identic, to what was found in Norrtälje. On at least five islands there were rock slabs that had been moved by the ice. The most beautiful examples were found on Rödskäret (Figs. 4.7 and 4.8). The largest rock slab on this island was about 5 m thick and 50x100 m wide. All islands showed major glacial erosion features. The island Örskär showed major meltwater features.

Figure 4.7.Displaced rockslab on Rödskäret. Photo P.Holmlund 2012

Figure 4.8. Rock slab on Rödskäret. Photo P. Holmlund august 2012

4.3.4. Öregrundsgrepen

This is here defined as the area in between Öregrund-North of Forsmark-Örskär. About 10 islands were visited over the years 2011-2014. This area includes an esker passing Kallrigafjärden with a submarine continuation north-west of Forsmark. The islands in the path between the visible and the marine esker is rock (till) covered. The islands next to the esker are heavily eroded, among them Mellanskäret with a high rate of erosion and interesting pools which may be results of thrusting or large scale ice plucking. One of the islands on which the cooling water outlet from the nuclear power station has been built also show rocks modulated by both ice and meltwater.

Figure 4.9.Polished rocks at Mellanskäret close to Forsmark. Photo P.Holmlund, September 2014.

4.4. Discussion

The channels of rivers entering the coast of the Gulf of Bothnia continue along the sea floor to present sea depths exceeding 50 m. These submarine river channels have not been possible to excavate since the early Weichselian when they twice were covered by first ice and later seawater. If these channels are ancient, and predate the period of large glaciations, negligible glacial erosion has acted on them. However, the relative water level in the Gulf of Bothnia is highly variable over time and the richness of glacial erosion features around the gullies rather point in another direction. The warm Eemian period had an extension of about 10,000 years, and was followed by a gradual cooling. It took another 40,000 years until the ice sheet reached the Gulf of Bothnia. At this point the Gulf of Bothnia was a fresh water lake drained southward into the Åland Sea. The present isostatic uplift around the Gulf of Bothnia is 0.5-1.0 cm a year. A large portion of the present seafloor of the Gulf of Bothnia was dry land during the initiation of the Weichselian, and when the glacier approached the present shore line large quantities of melt water were drained by these gullies. It also implies that the time for permafrost to develop on land is significant.

The bathymetric data show clear signs of an ice stream from the position of the ice divide close to the coast of Västerbotten all the way down to Åland Sea. It is not a straight pathway but rather a sinuous shaped path starting south, then southeast, and a turn to southwest and finally south into the Åland Sea. The morphology along this path varies with distance. In the north the landscape is highly modulated by the ice and in the south where the bedrock is crystalline the bedrock obstacles are all eroded free from sediments while the valleys are filled with glacial clay. Outside the pathway of the ice stream the glacial lineaments are more general from north to south and large well developed eskers occur. At some sites the subglacial water flow has made use of the ancient river channels shown by the position of eskers.

The most dramatic geomorphological features are shown in the Åland Sea east of the islands Gräsö and Väddö. The channel trough which was used by the ice stream is approximately 300 m below the present sea level. Large landslides are seen in the bathymetric data, indicating earthquakes possibly when the ice front was close, as no such features exist north of this area. To our knowledge there are no evidences of tsunamis hitting the shore of the Baltic this early though Lagerbäck et al. (2005) describe erosional unconformities in stratigraphic sequences in northern Uppland and Mörner (2003) describes possible traces of earthquakes. On Bornholm Andrén and Andrén (2001) found evidences for a tsunami like flood wave dated to 8 000 ca years before present (BP). This is somewhat late for having a connection to the observations in the Åland Sea. Andrén et al (2001) linked this event to a major slide at Storegga in southern Norway.

5. Investigating the hydrology and dynamics of the Scandinavian Ice Sheet

5.1. Meltwater transfer to the bed of the Greenland Ice Sheet: a contemporary analogue for a deglaciating Scandinavian Ice Sheet

Present day Greenland may be viewed as an analogy for a deglaciating Scandinavian Ice Sheet, with plentiful surface-generated meltwater. An ice sheet hydrological system dominated by the transfer of meltwater through the supra-, en- and subglacial systems may resemble the conceptual ice sheet depicted in Figure 5.1. The transfer of meltwater to the subglacial drainage system drives evolution from an inefficient system of fast flow and high water pressures (distributed), to one of efficient, fast flow with lower water pressures (channelized). The fluctuation of effective pressure (ice overburden pressure minus water pressure) in the subglacial drainage system has long been theorized to directly influence ice dynamics (Iken and Bindschadler, 1986; Jansson, 1995; Sugiyama and Gudmundsson, 2004). Observed velocity response to increased surface melting, the ‘surface meltwater effect’, provides evidence of surface-induced ice dynamic forcing at ice sheet scale in Greenland (i.e. Zwally et al., 2002; Bartholomew et al., 2011; Hoffman et al., 2011).

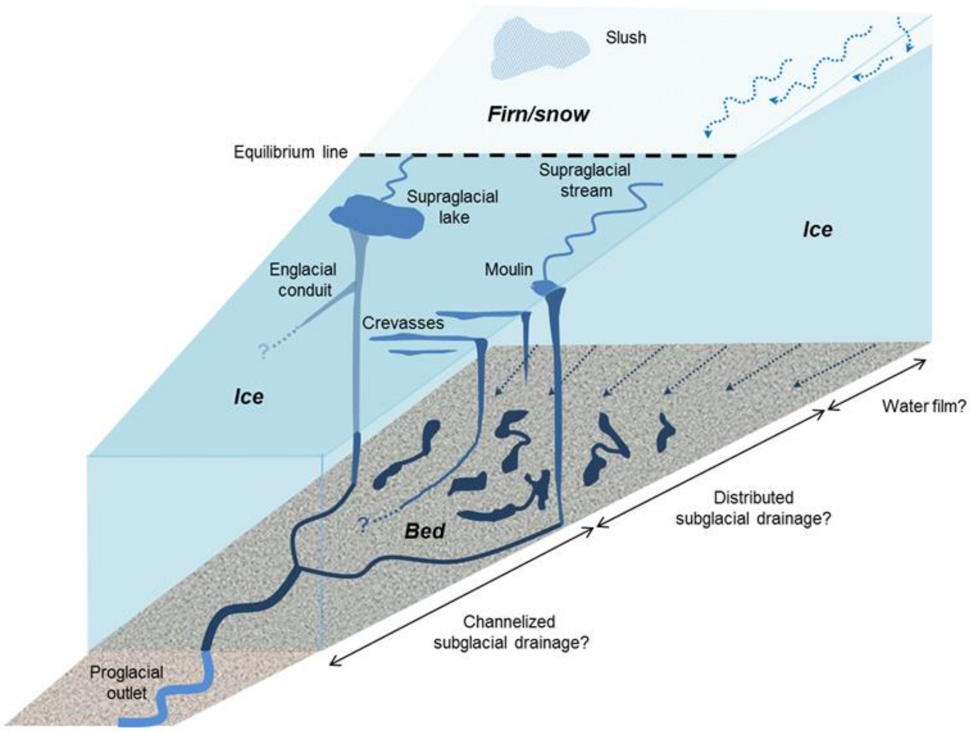


Figure 5.1. Conceptual model of an ice sheet hydrological system (adapted from Figure 1 of Margold et al., in review).

Despite such documented response of ice surface velocities to increased meltwater inputs on the Greenland Ice Sheet, recent studies now suggest that annually-averaged velocities do not generally correlate positively with melt rates (Sole et al., 2013; van de Wal et al., 2015), particularly in the lower ablation zone. Increased melting might in fact result in decreased winter velocities (Sole et al., 2013), counteracting any meltwater-forced summer speed-up. However, in the upper ablation zone observed velocity response to melting (van de Wal et al., 2015) indicates that increased melting and access of meltwater to the bed of the ice sheet at high elevations may become progressively more important under climatic warming and deglaciation, as the equilibrium line altitude increases, and supraglacial lakes form further inland of the margin (Leeson et al., 2015). This may in part be attributed to the inability of an efficient, channelized subglacial drainage system to form more than a few tens of kilometers from the ice sheet margin, due to large ice thicknesses and low ice surface slopes (Chandler et al., 2013; Meierbachtol et al., 2013). It is thus imperative to focus efforts on investigating the areal extent of meltwater penetrating to the bed (i.e. Clason et al., 2012; Clason et al., 2015) if the relative importance of meltwater-enhanced basal sliding on velocities and retreat rates of both contemporary and palaeo ice sheets is to be better understood.

Clason et al. (2015) applied a predictive model for the formation of ice surface-to-bed connections in the form of moulins and drainage of supraglacial lakes (Figure 5.2) to quantify meltwater transfer to the bed of the Leverett glacier catchment on the southwest Greenland Ice Sheet. The model was run for the 2009 and warmer 2010 melt seasons, producing timing and locations of meltwater delivery to the bed that match well against observed temporal and spatial patterns of ice surface speed-up.

Figure 5.2. Left: Moulin on the Leverett glacier, west Greenland (Photo: C. Clason, 2009). Right: Supraglacial lake in east Greenland (Photo: N. Selmes, 2008).

The necessity of modelling the process of meltwater transfer through fracture propagation, including supraglacial meltwater routing, was demonstrated when comparing model results against a control simulation (Figure 5.3). All surface-generated meltwater was permitted to reach the bed locally and instantaneously in the control simulation, neglecting meltwater routing across the ice surface, ice fracture propagation, and storage of meltwater at the surface and within crevasses. Above 750 m elevation (Figure 5.3d, e, f) the control simulation resulted in a mismatch between meltwater delivery and measured ice velocity (Bartholomew et al., 2011), which becomes increasingly worse with elevation. Modelled meltwater transfer, however, matches much more favourably with observed spatial and temporal patterns of acceleration in ice surface dynamics. The drainage of supraglacial lakes play an important role above 1000 m a.s.l., while moulins provide the main pathways for meltwater transfer in the lower catchment. The spatial pattern and timing of modelled lake drainages also compare favourably against those observed on repeat satellite imagery.

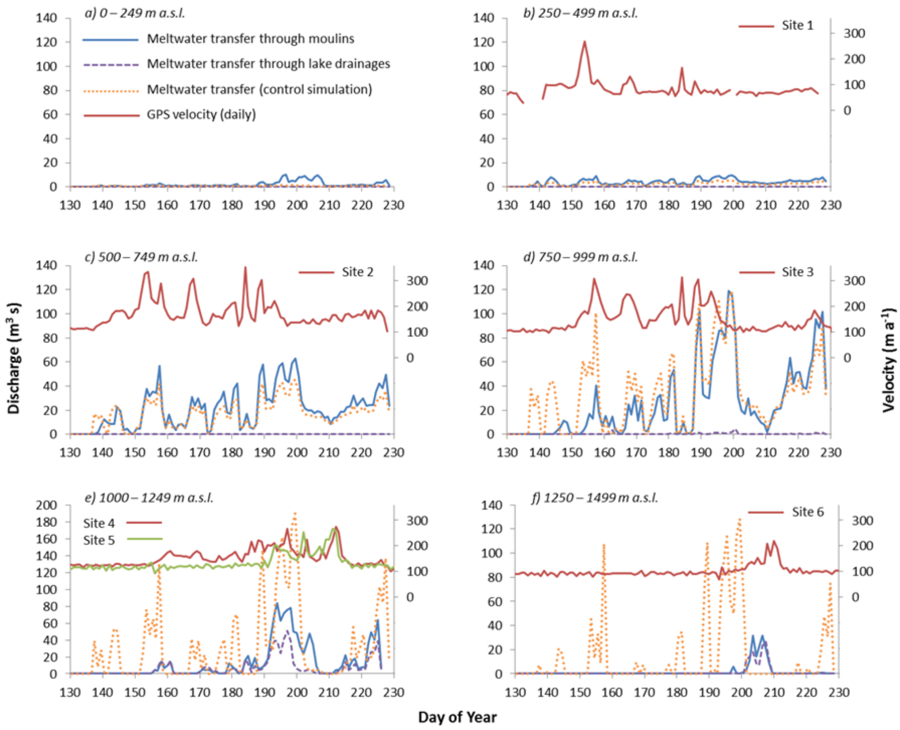


Figure 5.3. Supraglacial meltwater delivered to the bed each day through modelled lake drainages, moulins, and for the control simulation within ice surface elevation bands of 250m during 2009. Ice surface velocities from GPS sites 1–6 are plotted within their respective elevation bands (after Bartholomew et al., 2011). Note the extended y-axis in (e). (Figure 8 in Clason et al., 2015).

To evaluate the formation of moulins, drainage of lakes and the transfer of meltwater to the bed under a future warming climate, the model was run for the IPCC A1B June, July and August scenarios for the Arctic. Figure 5.4 illustrates the temporal and elevation changes in moulins and lakes drainages for the A1B mean scenario (+2.1°C). There is a temporal shift in moulin formation for the A1B scenario, with formation at high elevation beginning much earlier in the melt season. Importantly, applying a warming scenario results in increased drainage of lakes at high elevation, producing more widespread surface-to-bed meltwater transfer. This may have important consequences for ice dynamics in regions of the ice sheet where large ice thicknesses and low surface slopes hinder the formation of channelized, efficient subglacial drainage (Meierbachtol et al., 2013; van de Wal et al., 2015). In addition to an absolute increase in moulins and lake drainages, applying the A1B mean scenario also produced a 14% increase in the proportion of supraglacial meltwater reaching the ice sheet bed, in comparison to modelling of the 2009 melt season.

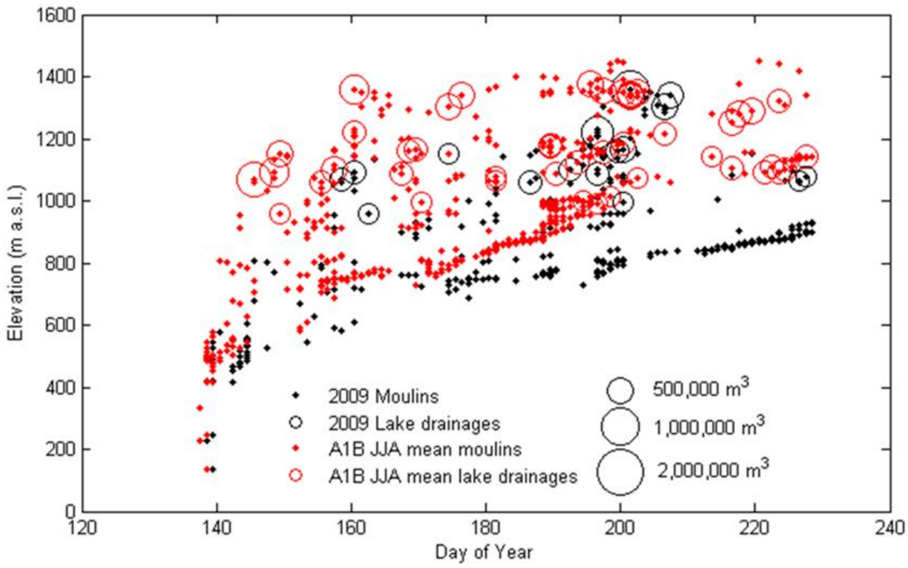


Figure 5.4. Spatial distribution of moulins and lake drainages for the 2009 melt season and the A1B mean June, July and August Arctic scenario (IPCC Fourth Assessment Report) applied to 2009 meteorological data. (Figure 6 in Clason et al., 2015).

5.2. Flow dynamics of the Baltic Ice Stream

In addition to a rich meltwater record, mapping of glacial lineations (streamlined features in the direction of ice flow) from submarine multibeam data reveals fast flowing ice in both the Bothnian and Baltic basins. These lineations include mega scale glacial lineations, drumlins, and rat-tails (Figure 5.5), the length of which can be used to infer relative ice flow velocity (Figure 5.6). Multiple flow events are recorded in the mapped lineations, evidenced clearly by crosscutting lineations (Figure 5.5D), and extensive patches of ribbed moraines. Very few marginal features, such as moraines, are observed, suggesting that there were no prolonged standstills during retreat of the ice sheet margin in this sector of the Scandinavian Ice Sheet in the Late Weichselian.

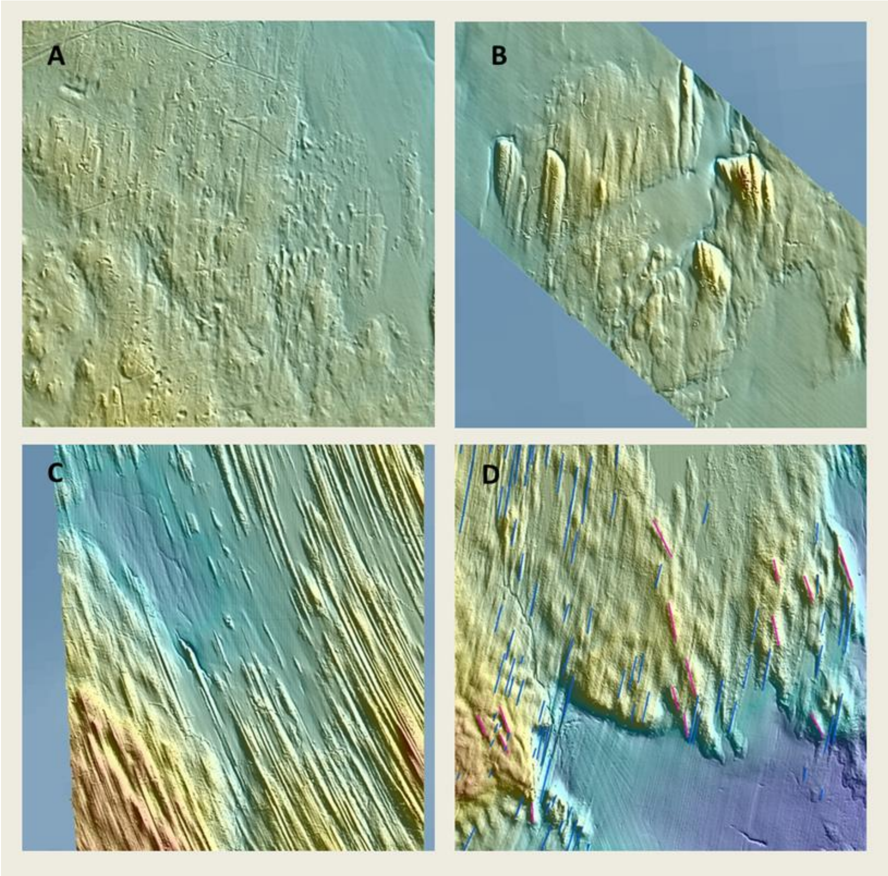


Figure 5.5. Glacial lineations observed in multibeam data in the Gulf of Bothnia. A) rat tails c. 200 m in lenght, B) drumlins c. 1-2 km in length, C) mega scale glacial lineations c. 10-20 km in length, and D) crosscutting lineations depicting changing ice flow directions.

In Greenwood et al. (in review), we identify a system of lineations crossing the terrestrial-marine boundary, based on integrated mapping of glacial landforms from land-based LiDAR data and marine-based multibeam and sidescan sonar data (Figure 5.6). These lineations indicate late stage ice streaming in a relatively narrow corridor of ice flow, originating on the Västerbotten and Ångermanland coast. Flow onset is in a south/south-easterly direction on land, turning sharply south/south-westward on the seafloor. The continuous nature of these mapped lineations across multiple data sets confirms that they are part of one flowline, and suggests that flow from this section of the present-day Swedish coast did not terminate in southeast Finland as part of an ice sheet lobe. Crevasse-squeeze ridges are also present in the multibeam data, overprinting the marginal end of lineations, and may indicate the end of a late stage surge or advance.

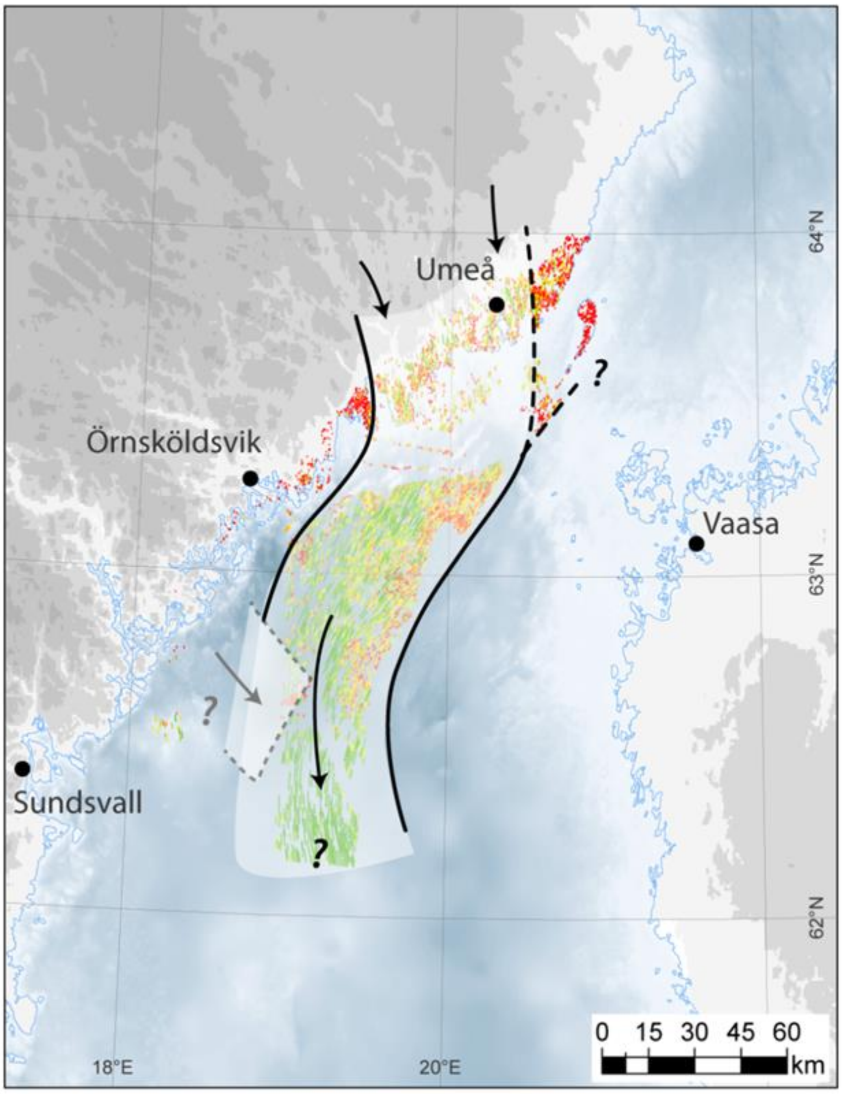


Figure 5.6. Pathway of ice stream flow into the Bothnian Sea from onset on the Västerbotten and Ångermanland coast. Note that short, red lineations infer slower flow, while longer, green lineations are indicative of fast flow or ice streaming (Figure 5 in Greenwood et al., in review).

5.3. Modelling the influence of meltwater on the evolution of the Weichselian Eurasian Ice Sheets

With ample geomorphological evidence for a meltwater-rich deglaciation of the Baltic Ice Stream sector of the Scandinavian Ice Sheet, Clason et al. (2014) investigated the role of meltwater-enhanced sliding on the Late Weichselian Eurasian Ice Sheets through application of the ice sheet model SICOPOLIS. Forced with an ice core-based scaling between present day and modelled Last Glacial Maximum (LGM) temperature and precipitation, the extent, evolution and dynamics of the ice sheets were simulated, with a focus on the LGM and subsequent deglaciation. By first running the model neglecting any basal sliding enhancement from surface meltwater the model produces a double-peaked LGM and an ice sheet with two domes, forming distinctive but joined Scandinavian and Barents-Kara ice sheets. Total ice volume peaks at 25 ka BP, with the Scandinavian Ice Sheet connecting with the British Ice Sheet across the North Sea (Figure 5.7, top panel), and by 20 ka BP this connection has been broken, while the depression between the Scandinavian and Barents-Kara ice sheets has largely disappeared (Figure 5.7, bottom panel). The modelled LGM extent compares favourably with the asynchronous extent drawn by Svendsen et al. (2004) derived from regional glacial records of geomorphology, stratigraphies, palaeo proxies and dating.

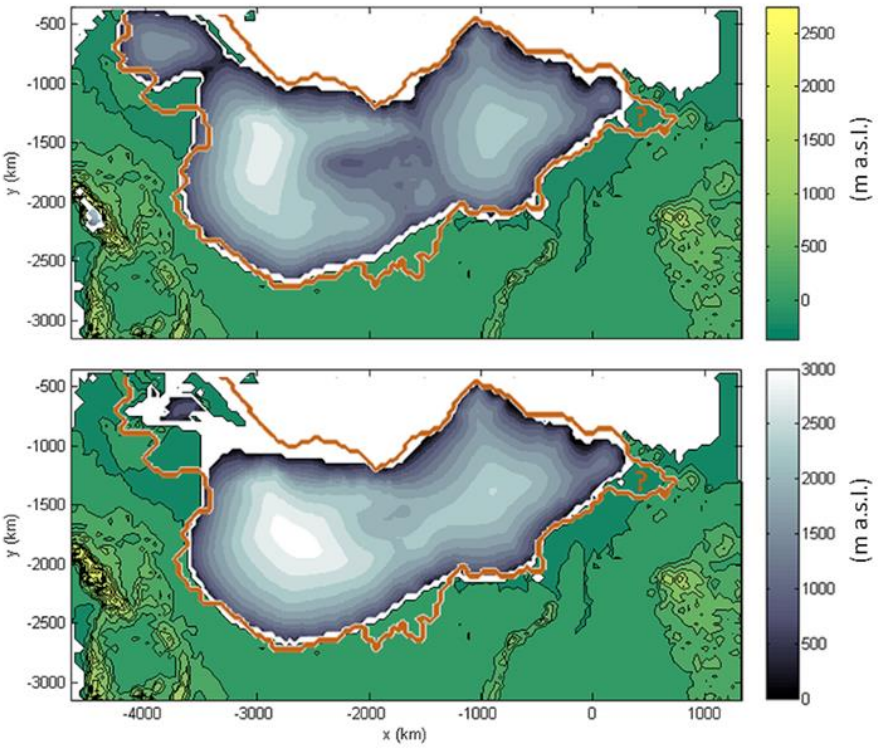


Figure 5.7. LGM maxima at 25 ka BP (top) and 20 ka BP (bottom). Blue shading represents ice surface elevation, and green represents land surface elevation. The LGM extent from Svendsen et al. (2004) is depicted in orange for comparison. (Figure 2 in Clason et al., 2014).

To investigate the influence of supraglacial meltwater reaching the bed of the Eurasian Ice Sheets a simple parameterization of the surface meltwater effect was implemented in the model, as applied previously within the Greenland domain of SICOPOLIS by Greve and Sugiyama (2009). This parameterization relates surface meltwater to basal sliding using a surface meltwater coefficient which accounts for the acceleration of basal sliding by supraglacially-derived meltwater at the bed. The surface meltwater effect is influenced either by the local surface melt rate alone, or by the local surface melt rate and local ice thickness, accounting for the less efficient surface-to-bed meltwater transfer through increasingly thick ice. Figure 5.8 illustrates the effect of this parameterization on modelled ice volume over time, where run 1 is a control simulation not including the surface meltwater effect, runs 2 to 5 relate enhanced sliding only the local surface melt rate (Figure 5.8, top panel), and runs 6 to 9 (Figure 5.8, bottom panel) relate enhanced sliding to both the local surface melt rate and ice thickness. The value of the surface meltwater coefficient is increased over runs 2 to 5 and 6 to 9 at the same rate to test the effect of increased melt-enhanced acceleration of basal sliding, and the relative influence of local melt rate and ice thickness. Furthermore, for areas of the ice sheet underlain by deformable sediments the sliding coefficient is tripled in comparison to that over bedrock.

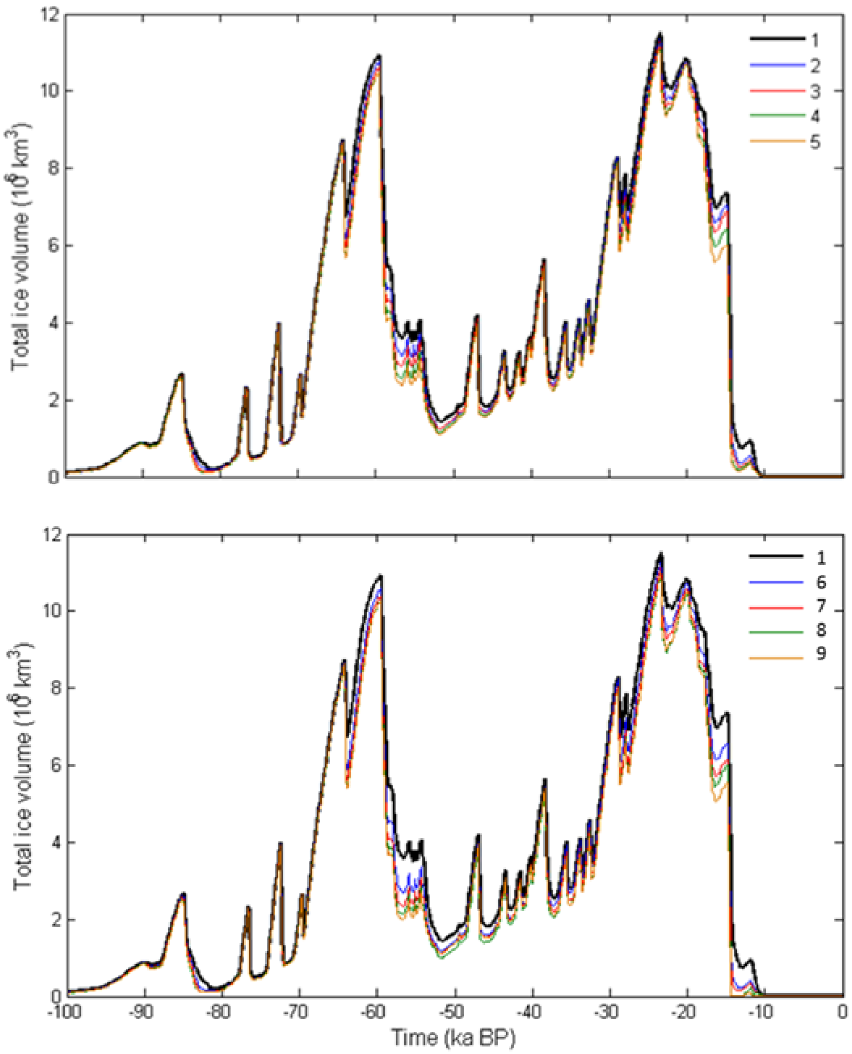


Figure 5.8. Total ice volume over time for model runs relating basal sliding to the local surafce melt rate (top) and to the local surafce melt rate and ice thickness (bottom), (adapted from Figure 4 and Figure 5 in Clason et al., 2014).

The result of including the surface meltwater effect within modelling is a decrease in total ice volume, which is most pronounced during periods of deglaciation. The percentage reduction is up to 19 % for run 5 and 25 % for run 9 in comparison to the control simulation (run 1) at 17 ka BP (Figure 5.10), approaching a complete early glacial termination at 15 ka BP for the most extreme surface meltwater effect parameterizations (runs 8 and 9). Inclusion of the surface meltwater effect produced spatial changes in ice sheet extent most notably in the southwest sector of the ice sheets. This forces a premature retreat of the ice margin into the Gulf of Bothnia, shutting off fast flow in the region of the Baltic Ice Stream. Three key conclusions can be drawn from these results: firstly, that a single value of the surface meltwater coefficient cannot be applied uniformly across the ice sheet, as the proportion of basal sliding that can be attributed to the surface meltwater effect varies spatially and temporally, highlighting the importance of developing physically-based models of surface-to-bed meltwater transfer such as Clason et al. (2012; 2015); secondly, that the Baltic/Bothnian sector of the Scandinavian Ice Sheet was very sensitive to meltwater inputs, which the geomorphological meltwater landform record suggests were plentiful; and thirdly, that the Baltic Ice Stream may have been buttressed by an ice shelf. Holmlund and Fastook (1993) concluded from their modelling study of the Scandinavian Ice Sheet that simulation of a Baltic Ice Stream was necessary to reproduce the LGM ice sheet extent, but required prescription of a frozen patch on Åland to slow flow and thinning of this sector during deglaciation. However, the presence of an ice shelf would also buttress ice flow, with any subsequent ice shelf disintegration leading to accelerated flow of glaciers and ice streams as observed on the present day West Antarctic Ice Sheet (e.g. Rignot et al., 2004; Dupont and Alley, 2005; Berthier et al., 2012).

6. Conclusions

- The river channels along the east coast stretches far out in the present Gulf of Bothnia, indicating that the Gulf has been significantly smaller during long periods. A wide and deep drainage channel in the south indicate that the Bay has been a lake. The last time this lake existed was probably during the initial phase of the Weichselian before the ice sheet had grown large enough to enter this area.
- Thrusting in rocks and movement of rock slabs on islands along the Gräsö archipelago indicate earth quakes short before glaciation. And, major landslides in the Åland Sea, between the light houses Understen and Märket and along the Väddö coast could be due to post glacial faulting after deglaciation.
- The northern Åland Sea is characterized by bare crystalline rocks and deep trenches. The deep north south facing spillway between Understen and Märket has a bed floor presently at approximately 200 m depth and it is infilled with 100 m clay. It is believed to have been the center of ice flow during deglaciation.
- The deglaciation development as described by Strömberg (1989) is well supported by bathymetric data, especially the opening of a calving bay next to the present east coast of Sweden.
- An east west oriented esker in southern Gulf of Bothnia cannot be explained by present recession models.
- The Uppsala Esker can be followed at least to the level of Söderhamn. Data lacks north of this level.
- The investigated area of Gulf of Bothnia show evidences of significant glacial impact on the ground. There is a richness in landforms, there is thrusting and melt water traces, both accumulation and erosion.
- Subglacial data from Antarctica show the potential for fragile landforms to survive a glaciation though neighboring sites are affected.
- The temperature regime of alpine glaciers respond quickly to climate changes and to geometric changes of the ice mass. However, during catastrophic changes, such as the deglaciation of South and central Sweden, the temperature distribution will never respond to current climate. The mass loss by melt is quicker than the heat conduction within the ice.
- The temperature at the base of, and within a glacier governs its deformation rate, its capability to slide and thus to provide ice flux to match ablation at the front. Changes in the temperature may cause variations in both direction of ice flow and on ice flux rates.
- Meltwater can penetrate a temperate ice mass and the drainage system may form a dendritic pattern governed by ice flow, gravity and ice surface geometry. In an ice mass at below freezing point temperatures meltwater at the ice surface is forced to run off supraglacially. However, in access of surplus of meltwater on the ice surface and horizontal strain in the ice water can penetrate thick ice masses. The latter process is exemplified by sudden drainage of

supraglacial lakes on the Greenland Ice Sheet causing sudden and severe water injections into the substratum.

- Despite glacial hydrology and its influence on ice dynamics playing a central role in current glaciological research, the topology of and processes occurring within the glacial drainage system remain poorly understood for both past and contemporary ice sheet domains. In order to forward our understanding of hydrology for past and contemporary ice sheets, a concerted effort must be made to forge better collaboration between the theoretical, contemporary observational and palaeo-glaciological research communities. In doing so we can contribute to better predictions of ice sheet response under a warming climate, using the past to inform the future.

7. References

Ahlmann, H., W:son., 1944: Nutidens Antarktis och istidens Skandinavien. Några jämförelser. Geologiska Föreningens i Stockholm Förhandlingar, 66(3): 635-652.

Ahlmann, H., W:son., 1953: Glacier Variations and Climate Fluctuations. Bowmans memorial lectures, series three. The American Geographical Society, New York, 51 p.

Andrén, T., and Andrén, E., 2001. Did the second Storegga slide affect the Baltic Sea? Baltica, 14:115–121

Bartholomew, I.D., Nienow, P., Sole, A., Mair, D., Cowton, T., King, M.A. and Palmer, S., (2011), Seasonal variations in Greenland Ice Sheet motion: Inland extent and behaviour at higher elevations, Earth and Planetary Science Letters, 307, 271-278

Berger, A., Loutre, M.F., 2002: An Exceptionally Long Interglacial Ahead? SCIENCE, 297: 1287-1288.

Berthier, E., Scambos, T.A. and Shuman, C.A., (2012), Mass loss of Larsen B tributary glaciers (Antarctic Peninsula) unabated since 2002, Geophysical Research Letters, 39 (13), L13501

Björnsson, H., 1981: Radio-echo sounding maps of Storglaciären, Isfallsglaciären and Rabots glaciär. Northern Sweden. Geografiska Annaler 63A (3-4): 225-231.

Björnsson, H., Gjessing, Y., Hamran, S.-E., Hagen, J. O., Liestøl, O., Pálsson, F. and Erlingsson, B.1996: The thermal regime of sub-polar glaciers mapped by multi-frequency radio-echo sounding. Journal of Glaciology, 42(140): 23–32.

Blatter, H. (1987). On the thermal regime of an arctic valley glacier: A study of White glacier, Axel Heiberg Island, N.W.T., Canada. Journal of Glaciology, 33, pp. 200-211

Blatter, H., and K. Hutter (1991).Polythermal conditions in arctic glaciers. Journal of Glaciology, 37, pp. 261-269

Bolin 2014: http://bolin.su.se/data/svenskaglaciarer/

Boulton, G.S., Smith, G.D., Jones, A.S. & Newsome, J. 1985: Glacial geology and glaciology of the last mid-latitude ice sheets. Journal of Geological Society, London 142, 447-474.

Catania, G.A., and Neumann, T.A., (2010), Persistent englacial drainage features in the Greenland Ice Sheet, Geophysical Research Letters, 37, L02501

Chandler, D.M., Wadham, J.L., Lis, G.P., Cowton, T., Sole, A., Bartholomew, I., Telling, J., Nienow, P., Bagshaw, E.B., Mair, D., Vinen, S. and Hubbard, A., (2013), Evolution of the subglacial drainage system beneath the Greenland Ice Sheet revealed by tracers, Nature Geoscience, 6, doi: 10.1038/NGEO1737

Clarke, G., S. Collins, D. Thompson (1984). Flow, thermal structure and subglacial conditions of a surge-type glacier.Canadian Journal of Earth Sciences, 21, pp. 232-240

Clason, C.C., Mair, D.W.F., Burgess, D.O. and Nienow, P.W., (2012), Modelling the delivery of supraglacial meltwater to the ice/bed interface: application to southwest Devon Ice Cap, Nunavut, Canada, Journal of Glaciology, 58 (208), 361-374

Clason, C.C., Applegate, P. & Holmlund, P., (2014), Modelling Late Weichselian evolution of the Eurasian ice sheets forced by surface meltwater-enhanced basal sliding, Journal of Glaciology, 60 (219), 29-40

Clason, C.C., Mair, D.W.F., Nienow, P.W., Bartholomew, I.D., Sole, A., Palmer, S. and Schwanghart, W., (2015), Modelling the transfer of supraglacial meltwater to the bed of Leverett Glacier, Southwest Greenland, The Cryosphere, 9, 123-138

Clow, G.D., 2014: temperature data acquired from the DOI/GTN-P Deep Borehole Array on the Arctic Slope of Alaska, 1973-2013. Earth System Science data 6: 201-218.

Cuffey, K.M., Conway, H., Hallet, B., Gades, A.M. and Raymond, C.F., 1999: Interfacial water in polar glaciers and glacier sliding at –17degrees C. Geophysical Research Letters, 26(6): 751-754.

Cuffey, K.M., Conway, H., Gades, A.M., Hallet, B., Lorraine, R., Severinghaus, J.P., Steig,

E.J., Vaughn, B., White, J.W.C., 2000: Entrainment at cold glacier beds. Geology 28 (4): 351-354.

Dansgaard, W., Johnsen, S.J., Clausen, H.B., Dahl-jensen, D., Gundestrup, N.S., Hammer,

C.U., Hvidberg, C.S., Steffensen, J.P., Sveinbjörnsdóttir, A.E., Jouzel, J. And Bond, G., 1993: Evidence for general instability of past climate from 250-kyr ice-core record. Nature 364: 218-220.

Cuffey, K., and Paterson, W.S.B., 2010:The physics of Glaciers. 4th ed. Oxford: Elsevier

Das, S.B., Joughin, I., Behn, M.D., Howat, I.M., King, M.A., Lizarralde, D. and Bhatia, M.P., (2008), Fracture Propagation to the Base of the Greenland Ice Sheet During Supraglacial Lake Drainage, Science, 320, 778–781

Delcourt, C., Van Liefferinge, B., Nolan, M., Pattyn, F., 2013: The climate memory of an Arctic polythermal glacier. Journal of Glaciology 59 (218): 1084-1092.

De Geer, G., 1940: Geochronologia Sueccica Principles. Kungliga Svenska Vetenskapsakademiens Handlingar, Tredje serien, Bd18:6, 367 pp.

Doyle, S.H., Hubbard, A.L., Dow, C.F., Jones, G.A., Fitzpatrick, A., Gusmeroli, A., Kulessa, B., Lindback, K., Pettersson, R., and Box, J.E., (2013), Ice tectonic deformation during the rapid in situ drainage of a supraglacial lake on the Greenland Ice Sheet, The Cryosphere, 7, 129–140

Dupont, T.K. and Alley, R.B., (2005), Assessment of the importance of ice-shelf buttressing to ice-sheet flow, Geophysical Research Letters, 32 (4), L04503

Eriksson, M.G., Björnsson, H., Herzfeld, U.C., Holmlund, P., 1993: The Bottom topography of Storglaciären. A new map based on old and new ice depth measurements, analysed with geostatistical methods. Department of Physical geography, Stockholm University (ISSN 0346-7406), Forskningsrapport 95, 48 pp.

Fredin, O., Hättestrand C., 2002: Relict lateral moraines in northern Sweden – evidence for an early mountain centred ice sheet. Sedimentary Geology 149: 145-156

Eklund, A., 1988: 20K Umeå/20L Holmön, 19J Husum/20J Vännäs and 20I Björna/21IFredrika. Kvartärgeologiska kartan. Sveriges geologiska undersökning, SGU, Ak, 5:2, 6:2 and 7:2; Isrörelser, stratigrafi och högsta kustlinjen. Maps at scale 1: 100 000.Etzelmüller, B., and J.-O.Hagen (2005).Glacier-permafrost interaction in Arctic and alpine mountain environments with examples from southern Norway and Svalbard. Geological Society, London (Special Publications, 242)

Fastook, J. and Holmlund, P., 1994: A glaciological model of the Younger Dryas event in Scandinavia. Journal of Glaciology 40 (134): 125-131.

Fountain, A.G., Jacobel, R.W., Schlichting, R., Jansson, P., 2005: Fractures as the main pathways of water flow in temperate glaciers. Nature, 433 (7026): 618-621.

French, H.M., Millar, S.W.S., 2014: Permafrost at the time of the last Glacial Maximum (LGM) in North America. Boreas 43: 667-677. (10.1111/bor.12036. ISSN 0300-9483).

Fretwell, P., (+59 co-authors, incl. Holmlund, P.) 2013: Bedmap2: Improved ice bed, surface and thickness datases for Antarctica. The Cryosphere (doi: 10.5194/tc-7-375-2013) 7: 375-393.

Fujita, S., Holmlund, P., Andersson, I., Brown, I., Enomoto, H., Fujii, Y., Fujita, K., Fukui, K., Furukawa, T., Hansson, M., Hara K., Hoshina, Y., Igarashi, M., Iizuka, Y., Imura, S., Ingvander, S., Karlin, T., Motoyama, H.., Nakazawa, F., Oerter, H., Sjöberg, L. , Sugiyama, S., Surdyk, S., Ström, J. Uemura, R. and Wilhelms, F., 2011: Spatial and temporal variability of accumulation rate on the East Antarctic ice divide between Dome Fuji and EPICA DML. The Cryosphere (doi: 10.5194/tc-5-1-2011) 5: 1-24.Greenwood, S.L., Clason, C.C., Mikko, H., Nyberg, J., Peterson, G. & Smith, C.A., Integrated use of LiDAR and multibeam bathymetry reveals onset of ice streaming in the northern Bothnian Sea, in review in GFF

Fujita, S., Holmlund, P., Matsuoka, K., Enomoto, H., Fukui, K., Nakazawa, F., Sugiyama, S., Surdyk, S., 2012: Radar diagnosis of the subglacial conditions in Dronning Maud Land, East Antarctica. The Cryosphere (doi: 10.5194/tc-6-1-2012) 6: 1-17.

Greve, R. and Sugiyama, S., (2009), Decay of the Greenland Ice Sheet due to surface-meltwater-induced acceleration of basal sliding, arXiv:0905.2027

Gulley, J.D., Benn, D.I., Muller, D. and Luckman, A., 2009a.A cut-and-closure origin forenglacialconduits in uncrevassed regions of polythermal glaciers. Journal ofGlaciology, 55(189), 66-80

Gusmeroli, A., P. Jansson, R. Pettersson, T. Murray (2012). Twenty years of cold surface layer thinning at Storglaciären, sub-Arctic Sweden, 1989-2009. Journal of Glaciology, 58, pp. 3-10

Hamberg, A., 1901: Sarjekfjällen, en geografisk undersökning. Ymer 1901 (2): 145-204.

Hamberg, A., 1910: Die gletscher des Sarekgebirges und ihre untersuchung. Eine Kurze übersicht.In: The Gletscher Schwedens im Jahre 1908. Sveriges Geologiska Undersökning.Ca 5 (3), 26 pp.

Hamran, S.-E.andAarholt, E.1993: Glacier study using wavenumber domain synthetic aperture radar. Radio Science, 28(4): 559–570.

Hamran, S.-E., Gjessing, D. T., Hjelmstad, J. and Aarholt, E.1995: Ground penetrating synthetic pulse radar: dynamic range and modes of operation. Journal of Applied Geophysics, 33: 7–14.

Harris, C, Aaeberli, W., Mühll, D.Vonder. and King, L., 2001: Permafrost Monitoring in the High Mountains of Europe: The PACE project in ist Global Context. Permafrost and periglacial Processes 12 (1): 3-12.

Harris, C., Mühll, D.V., Isaksen, K., Haeberli, W., Sollid, J.L., King, L., Holmlund, P., Dramis, F., Guglielmin, M. and Palacios, D., 2003: Warming permafrost in European mountains. Global and Planetary Change 39: 215-225.

Harris, C., and J. Murton (2005).Cryospheric Systems: Glaciers and Permafrost, Geological Society, London (Special Publications, 242)

Herzfeld, U. C. and Holmlund, P., 1990: Geostatistics in glaciology: Implications of a study of Scharffenbergbottnen, Dronning Maud Land, East Antarctica. Annals of Glaciology 14: 107-110.

Herzfeld, U. C. and Holmlund, P., 1990: Geostatistics in glaciology: Implications of a study of Scharffenbergbottnen, Dronning Maud Land, East Antarctica. Annals of Glaciology 14: 107-110.Holmlund, P.,

Hoffman M.J., Catania, G.A., Neumann, T.A., Andrews, L.C. and Smith, J.A., (2011), Links between acceleration, melting and supraglacial lake drainage of the western Greenland Ice Sheet, Journal of Geophysical Research – Earth Surface, 116, F04035, doi:10.1029/2010JF001934

Holmlund, P., 1986: Mikkaglaciären: Bed topography and response to 20th century climate change. Geografiska Annaler. 68 A (4): 291-302.

Holmlund, P. (1987). Mass balance of Storglaciären during the 20th century.Geografiska Annaler, 69, pp. 439-447

Holmlund, P. 1988a. Is the longitudinal profile of Storglaciären, northern Sweden, in balance with the present climate?.Journal of Glaciology, 34, pp. 269-273

Holmlund, P., 1988b: An application of two theoretical melt water drainage models on Storglaciären and Mikkaglaciären, northern Sweden. Geografiska Annaler 70 A (1-2): 1-7.

Holmlund, P., 1993: Interpretation of basal ice conditions from radio-echo soundings in the Eastern Heimefrontfjella and the southern Vestfjella mountain ranges, East Antarctica. Annals of Glaciology 17: 312-316.

Holmlund, P., 1993: Surveys of Post Little Ice Age glacier fluctuations in Northern Sweden. Zeitschrift für Gletscherkunde und Glazialgeologie 29 (1): 1-13.

Holmlund, P.,1998: Glaciers as detailed climatic records in subpolar environments, Ambio 1998(4): 266-269.

Holmlund, P., 2012: Glaciärer- Gnistrande smycken som ännu pryder våra fjäll. Votum & Gullers förlag (ISBN 978-91-85815-83-8), 160 s.

Holmlund, P. and Eriksson, M.1989: The cold surface layer on Storglaciären. Geografiska Annaler,71A(3–4): 241–244.

Holmlund, P. and Fastook, J., (1993), Numerical modelling provides evidence of a Baltic ice stream during the Younger Dryas, Boreas, 22 (2), 77–86

Holmlund, P. and Fastook, J., 1995: A time dependent glaciological model of the Weichselian

Ice sheet.Quaternary International, 27: 53-58.

Holmlund, P., J.-O. Näslund, C. Richardsson (1996). Radar surveys on Scandinavian glaciers in search of useful climate archives. Geografiska Annaler, 78A, pp. 147-154

Holmlund, P. and Näslund, J-O., 1994: The glacially sculptured landscape in Dronning Maud Land, Antarctica, formed by wet-based mountain glaciation and not by the present ice sheet. Boreas, 23: 139-148.Howat, I.M., de la Pena, S., van Angelen, J.H., Lenaerts, J.T.M. and van den Broeke, M.R., (2013), Expansion of meltwater lakes on the Greenland Ice Sheet, The Cryosphere, 7, 201-204

Holmlund,P. and Schneider, T.,1997: The effect of continentality on glacier response and mass balance. Annals of Glaciology 24: 272-276

Hoppe, G., 1961: The continuation of the Uppsala esker in the Bothnian Sea and ice recession in the Gävle area. Geografiska Annaler 43 (3): 329-335.

Iken, A. and Bindschadler, R.A., (1986), Combined measurements of subglacial water pressure and surface velocity of Findelengletscher, Switzerland: Conclusions about drainage system and sliding mechanism, Journal of Glaciology, 32 (110), 101-119

IPCC: Climate Change 2007: The Physical Science Basis, in: Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, edited by: Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K. B., Tignor, M., and Miller, H. L., Cambridge University Press, Cambridge, UK, 2007

Isaksen, K., Holmlund, P., Sollid, J.L., Harris, C., 2001: Three deep alpine-permafrost boreholes in Svalbard and Scandinavia. Permafrost and Periglacial Processes 12 (1): 13-25.

Isaksen, K., Sollid, J.L., Holmlund, P. and Harris, C., 2007: Recent warming of mountain permafrost in Svalbard and Scandinavia. Journal of Geophysical Research, 112, F02S04, doi: 10.1029/2006JF000522, 2007.

Jansson, P., (1995), Water pressure and basal sliding on Storglaciären, northern Sweden, Journal of Glaciology, 41 (138), 232-240

Johnsen, S.J., Clausen, H.B., Dansgaard, W., Fuhrer, K., Gundestrup, N., Hammer, C.U., Iversen, P., Jouzel, J., Stauffer, B. and Steffensen, J.P., 1992: Irregular glacial interstadials recorded in a new Greenland ice core. Nature 359: 311-313.

Jonsson, S., 1988: Observations on the physical geography and glacial history of the Vestfjella nunataks in western Dronning Maud Land, Antarctica. , Stockholms universitet, Naturgeografiska institutionen, Forskningsrapport 87.King, L., 1984: Permafrost in Skandinavien, Untersuchungesergebnisse aus Lappland, Jotunheimenund Dovre/Rondane. Heidelberger Geographisches Arbeiten 76.

Kleman, J., 1992: The Palimsest glacial landscape in north-western Sweden – Late

Weichselian deglaciation landforms and traces of older west-centered ice sheets. Geografiska Annaler 74 A (4): 305-325.

Kleman, J., 1994: Preservation under ice sheets and ice caps. Geomorphology 9:19-32.

Kleman, J. & Borgström, I., 1994: Glacial land forms indicative of a partly frozen bed. Journal of Glaciology 40(135): 255-264.

Klingbjer, P., and F. Neidhart (2006).The thinning and retreat of Pårteglaciären, Northern Sweden, during the Twentieth Century and its relation to climate.Arctic,Antarctic and Alpine Research, 38, pp. 104-112

Krimmel, R. (1999). Analysis of difference between direct and geodetic mass balance measurements at South Cascade Glacier, Washington.Geografiska Annaler, 81 A, pp. 653-658

Lagerbäck, R. 1988: Periglacial phenomenon in the wooded areas of northern Sweden - relicts from the Tärendö interstadial. Boreas 17, 487-500.

Lagerbäck, R., Sundh, M., Svedlund, J-O.And Johansson, H., 2005: Forsmark site investigation. Searching for evidence of late- or postglacial faulting in the Forsmark region.Results from 2002-2004. Svensk Kärnbränslehantering AB, ISSN 1402-3091. SKB Rapport R-05-51, 50 p.

Leeson, A.A., Shepherd, A., Briggs, K., Howat, I., Fettweis, X., Morlighem, M. and Rignot, E., (2015), Supraglacial lakes on the Greenland ice sheet advance inland under warming climate, Nature Climate Change, 5, 51-55

Lundqvist, J., The deglaciation of Sweden after 10,000 B.P. Boreas 9: 229-238.

Margold, M., Clason, C.C., Greenwood, S.L. & Helanow, C., Processes and products: theoretical, contemporary observational and palaeo perspectives on ice sheet hydrology, in review in Earth-Science Reviews

Meierbachtol, T., Harper, J. and Humphrey, N., (2013), Basal Drainage System Response to Increasing Surface Melt on the Greenland Ice Sheet, Science, 341 (777), DOI: 10.1126/science.1235905

Murray, T., Stuart, G., Miller, J., Woodward, J., Smith, A., Porter, P., Jiskoot, H., 2000: Glacier surge propagation by thermal evolution at the bed. Journal of Geophysical Research, 105: 491-513.

Mörner, N-A., 2003: Paleosesmicity of Sweden. A novel paradigm.A contribution to INQUA from its Sub-commission on Plaeoseismology (ISBN-91-631-4072-1).JOFO Grafiska AB, Stockholm, 320 pp.

Näslund, J-O., 1997: Subglacial preservation of valley morphology at Amundsenisen, West ern Dronning Maud Land, Antarctica. Earth Surface Processes and Landforms. 22: 441-455.

Näslund, J.O., 2001: Landscape development in central Dronning Maud Land, East Antarctica. Antarctic Science, 13(3): 302-311.

Näslund, J.O., Fastook,J.L. and Holmlund, P., 2000: Numerical modelling of the ice sheet in western Dronning Maud Land, East Antarctica: impacts of present, past and future climates. Journal of Glaciology, 46(152): 54-66.Pettersson, R., Jansson, P. and Holmlund, P. 2003: Thinning of the cold surface layer on Storglaciären, observed by repeated ground penetrating radar surveys. Journal of Geophysical Research, 108(F1, 6004): doi:10.1029/2003JF000024.Pettersson, R., P. Jansson, P. Holmlund (2003). Cold surface layer thinning on Storglaciären, Sweden, observed by repeated ground penetrating radar surveys. Journal of Geophysical Research, 108, pp. 5.1-5.9

Pettersson, R., P. Jansson, H. Huwald, H. Blatter (2007). Spatial pattern and stability of the cold surface layer on Storglaciären, Sweden. Journal of Glaciology, 53, pp. 99-109

Porter, S., 1989: Some geological implications of average Quaternary glacial conditions. Quaternary Research 32: 245-261.

Rignot, E., Casassa, G., Gogineni, P., Krabill, W., Rivera, A. and Thomas, R., (2004), Accelerated ice discharge from the Antarctic Peninsula following the collapse of Larsen B ice shelf, Geophysical Research Letters, 31 (18), L18401

Rippin, D., J. Carrivick, C. Williams (2011). Evidence towards a thermal lag in the response of Kårsaglaciären, northern Sweden, to climate change.Journal of Glaciology, 57, pp. 895-903

Ritscher, A., 1942: Deutsche Antarktische Expedition 1938/39. Erster band.Koehler&Amelang, Leipzig. 304 printed pages+photographs and maps.

Schytt, V., 1947: Glaciologiska arbeten i Kebnekajse. Ymer (1): 18-42.

Schytt, V., 1974: Inland ice sheets - recent and Pleistocene. Geologiska Föreningens i Stockholm Förhandlingar 96, 299-309.

Schneider, T. and Bronge, C., 1996: Suspended sediment transport in the Storglaciären drainage basin. Geografiska Annaler 78 A (2-3): 155-161.

Shreve, R.L.,1972: Movement of water in glaciers. Journal of Glaciology, 11(62): 205-214.

Sole, A., Niewnow, P., Bartholomew, I., Mair, D., Cowton, T., Tedstone, A. and King, M., (2013), Winter motion mediates dynamic response of the Greenland Ice Sheet to warmer summers, Geophysical Research Letters, 40, 3940-3944

Strömberg, B., 1981: Calving bays, striae and moraines at Gysinge-Hedesunda, Central Sweden. Geografiska Annaler 63 A (3-4): 149-154.

Strömberg, B., 1989: Late Weichselian deglaciation and varve chronology in East-Central Sweden. Sveriges geologiska undersökning, Ca 73. 70 pp.

Strömberg, B., (1990), A connection between the clay varve chronologies in Sweden and Finland. Annales Academiae Scientiarum Fennicae Geologica-Geographica, 154, 32pp

Strömberg, B., (2005), Clay varve chronology and deglaciation in SW Finland. Annales Academiae Scientiarum Fennicae Geologica-Geographica, 167, 49pp

Strömberg, B., (2010), Rare forms of meltwater erosion on bedrock: polished flutes in the Åland Sea area, Sweden – Finland, Annales Academiae Scientiarum Fennicae Geologica-Geographica, 169, 39pp

Sugiyama, S. and Gudmundsson, G.H, (2004), Short-term variations in glacier flow controlled by subglacial water pressure at Lauteraargletscher, Bernese Alps, Switzerland, Journal of Glaciology, 50 (170), 353-362

Summerfield, M. (1991).Global geomorphology. Harlow: Pearson Education Limited

Svendsen, J.I. and 30 others, (2004), Late Quaternary ice sheet history of northern Eurasia. Quaternary Science Reviews, 23 (11–13), 1229–1271

Svensson, H., 1964: Fossil Tundramark på Laholmsslätten. Sveriges geologiska undersökning, C598, 29 p.

Svensson, H., 1974: Distribution and chronology of relict polygon patterns on the Laholm plain, the Swedish west coast. Geografiska Annaler 54 A(3-4): 159-175.

Svenonius, F., 1910: Studien über den Kårso- und Kebnegletscher nebst Notizen über andere Gletscher im Jukkasjärvigebirge. In: The Gletscher Schwedens im Jahre 1908. Sveriges Geologiska Undersökning.Ca 5 (1), 54 p.

Torell, O., 1873: Undersökningar öfver istiden. II. Skandinaviska inlandsisens utsträckning under isperioden. Öfversigt af Kungliga Vetenskaps-akademiens förhandlingar 30 (1), 47-64.

van Heteren, S., D. FitzGerald, P. McKinlay, I. Buynevich (1998). Radar facies of paraglacial barrier systems: coastal New England, USA. Sedimentology, 45, pp. 181-200

van de Wal, R.S.W., Smeets, C.J.P.P., Boot, W., Stoffelen, M., van Kampen, R., Doyle, S.H., Wilhelms, F., van den Broeke, M. R., Reijmer, C.H., Oerlemans, J. and Hubbard, A., (2015), Self-regulation of ice flow varies across the ablation area in south-west Greenland, The Cryosphere, 9, 603-611

Wahlenberg, G., 1808: Berättelse om mätningar och observationer för att bestämma lappska fjällens höjd och temperatur vid 67 graders polhöjd, förrättade 1807. Swedish Academy of Sciences, KVA, Stockholm 1808, 58p.

WGMS, 2008: Global Glacier Changes: facts and figures.eds. Michael Zemp and Jaap van Woerden. (ISBN978-92-807-2898-9) World Glacier Monitoring Service, 88 p.

WGMS, 2012: Fluctuations of Glaciers 2005-2010. World Glacier Monitoring Service (ISSN 1997-910X, 336 pp.

Wingham, D.J., Siegert, M.J., Shepherd, A. and Muir, A.S., 2006. Rapid discharge connects Antarctic subglacial lakes. Nature, 440(7087), 1033-6.

Wolovick, M.J., Bell, R.E., Creyts, T.T. and Frearson, N., 2013. Identification and control of subglacial water networks under Dome A, Antarctica. Journal of Geophysical Research: Earth Surface, 118(1), 140-154.

Yang, D., D. Kane, L. Hinzman, X. Zhang, T. Zhang, H. Ye (2002).Siberian Lena river hydrologic regime and recent change.Journal of Geophysical Research, 107, D234694

Zwally, J.H., Abdalati, W., Herring, T., Larson, K., Saba, J., and Steffen, K., (2002), Surface Melt-Induced Acceleration of Greenland Ice Sheet Flow, Science, 297, 218–222

Ångström, A., 1974: Sveriges klimat. Third edition. Stockholm, generalstabens litografiska anstalt

Östrem, G., Haakensen, N., Melander, O., 1973; Atlas over breer i Nord-Skandinavia. Norges vassdrags og elektrisitetsvesen. Meddelse nr 22 fra hydrologisk avdelning, 315 pp.

Appendix

Publications and presentations

A1 Peer-reviewed articles

A1.1 Published:

Greenwood, S.L., Clason, C.C., Mikko, H., Nyberg, J., Peterson, G. & Smith, C.A., 2015: Integrated use of LiDAR and multibeam bathymetry reveals onset of ice streaming in the northern Bothnian Sea, GFF, DOI:10.1080/11035897.2015.1055513

Clason, C.C., Mair, D.W.F., Nienow, P.W., Bartholomew, I.D., Sole, A., Palmer, S. & Schwanghart, W., 2015: Modelling the transfer of supraglacial meltwater to the bed of Leverett Glacier, Southwest Greenland, The Cryosphere, 9, 123-138

Clason, C.C., Applegate, P. & Holmlund, P., 2014: Modelling Late Weichselian evolution of the Eurasian ice sheets forced by surface meltwater-enhanced basal sliding, Journal of Glaciology, 60 (219), 29-40

A1.2 In review:

Greenwood, S.L. Clason, C.C., Helanow, C. & Margold, M., Theoretical, contemporary observational and palaeo perspectives on ice sheet hydrology: processes and products, Earth-Science Reviews

A1.3 In preparation:

Clason, C.C., Greenwood, S.L., Selmes, N., Lea, J.M., Jamieson, S.S.R., Nick, F.M. & Holmlund, P., Testing controls on the post-Younger Dryas retreat of the Bothnian Sea Ice Stream through numerical flowline modelling

A2 Master Thesis

Blomdahl, K 2015: Changes in the cold surface layer on a polythermal glacier during substantial ice mass loss, Degree Project, Uppsala University, Uppsala. (ISSN 1650-6553) Supervisor: Per Holmlund. Department of Earth Sciences, Uppsala University, Villavägen 16, SE-752 36 Uppsala.

A3 Conference presentations

Greenwood, S.L., Clason, C.C., Holmlund, P., Jakobsson, M. & Nyberg, J., 2015: Ice streams, surges and meltwater-driven ice sheet decay in the Gulf of Bothnia revealed by a rich submarine glacial landform record, International Union for Quaternary Science Congress

Greenwood, S., Clason, C., Nyberg, J., Hell, B., Öiås, H., Holmlund, P. & Jakobsson, M., 2014: Signatures of ice flow, retreat and meltwater delivery in the Gulf of Bothnia, European Geosciences Union General Assembly

Clason, C., Holmlund, P., Applegate, P. & Strömberg, B., 2012: Exploring the role of surface-to-bed meltwater transfer on the evolution of the Scandinavian Ice Sheet during the Late Weichselian, American Geophysical Union Fall Meeting

Clason, C., Holmlund, P., Applegate, P. & Strömberg, B., 2012: Exploring the role of surface-to-bed meltwater transfer events on the evolution of the Scandinavian Ice Sheet during the Weichselian, International Glaciological Society British Branch Meeting & International Glaciological Society Nordic Branch Meeting.

Greenwood, S., Clason, C., Lea, J., Selmes, N., Nyberg, J., Jakobsson, M., Holmlund, P., 2016: Dynamics of and controls on post – Younger Dryas retreat of the Bothnian Sea ice stream. Nordic geological winter meeting 2016.

Holmlund, P., Clason, C., 2013: The influence of thermal structure of glaciers on hydrology and their response to climate change. IASC workshop on the dynamics and mass budget of Arctic glaciers, February 2013 in Obergurgl, Austria.

Holmlund, P., Clason, C., Blomdahl, K., 2014: Present changes in extent and thermal regime of Swedish glaciers. International Glaciological Society, Nordic Branch meeting in Iceland.

Holmlund, P., Clason, C., Blomdahl, K., 2015: Present changes in extent and thermal regime of Swedish glaciers. IUGG-IACS in Prague

Holmlund, P., Clason, C., Greenwood, S., Jakobsson, M., 2015: An Investigation of ice flow and retreat of the Baltic Ice Stream from a new rich submarine glacial landform record. IUGG-meeting in Prague.

A4 Planned future publications

Description of mapped glacial geomorphology from multibeam data in the Baltic and Bothnian basins (with Sarah Greenwood, Per Holmlund, Martin Jakobsson and Johan Nyberg).

A combined modelling and geomorphological study of the dynamics and hydrology of the Bothnian Sea Ice Stream (with Sarah Greenwood).

A study focussed on the subglacial hydrology of the Baltic and Bothnian sea sector, including the possible role of subglacial lakes (with Greenwood and Per Holmlund).

Isostatic rebound and tectonic movements in the Gulf of Bothnia (with Per Holmlund, Martin Jakobsson and Sarah Greenwood).

2016:21

The Swedish Radiation Safety Authority has a comprehensive responsibility to ensure that society is safe from the effects of radiation. The Authority works to achieve radiation safety in a number of areas: nuclear power, medical care as well as commercial products and services. The Authority also works to achieve protection from natural radiation and to increase the level of radiation safety internationally.

The Swedish Radiation Safety Authority works proactively and preventively to protect people and the environment from the harmful effects of radiation, now and in the future. The Authority issues regulations and supervises compliance, while also supporting research, providing training and information, and issuing advice. Often, activities involving radiation require licences issued by the Authority. The Swedish Radiation Safety Authority maintains emergency preparedness around the clock with the aim of limiting the aftermath of radiation accidents and the unintentional spreading of radioactive substances. The Authority participates in international co-operation in order to promote radiation safety and finances projects aiming to raise the level of radiation safety in certain Eastern European countries.

The Authority reports to the Ministry of the Environment and has around 300 employees with competencies in the fields of engineering, natural and behavioural sciences, law, economics and communications. We have received quality, environmental and working environment certification.

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