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Glacier surges and landforms in a permafrost environment at the tidewater glacier Paulabreen, inner Van Mijenfjorden, Svalbard

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Abstract

This thesis presents a study of the landsystem of the Svalbard tidewater glacier Paulabreen and its late Holocene surge moraines, focusing on the glaciology and the glacial geology. An active surge of Skobreen/Paulabreen was observed and the glacier dynamics and stress regime was studied using satellite images, a time-lapse movie and photographs. A persistent subglacial conduit was found beneath the medial moraine between Paulabreen and Bakaninbreen, and we postulate that this constrained surge propagation by preventing the spread of pressurized water beyond the channel. Fresh glacial submarine landforms in front of the glacier were studied and related to known glacier front positions since 1898. An older moraine system deposited around 600 yrs BP was studied, and we found that the terrestrial and submarine landforms were very similar. A mud apron found in both environments was interpreted as pushed marine sediments deposited in a slurry in front of the surging glacier. Hummocky moraine was formed mainly by squeezing of sediments into basal crevasses. While de-icing is almost complete in some parts of the 600 yrs BP moraine, more than 30 m of buried glacier ice is found in a lateral part of the moraine, Crednermorenen. This great variance in the ice-core preservation we attribute to a higher supraglacial sediment content at the glacier margins. Water temperature measurements and modelling work suggest that permafrost, defined solely by temperature, probably exist in Van Mijenfjorden, but we have no evidence that the seabed is frozen.

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List of papers

I. **Kristensen, L.** & Benn D. I. A surge of Skobreen/Paulabreen, Svalbard documented by a time-lapse movie, aerial and satellite images and photographs. To be submitted to Geosphere.

II. Benn, D. I., **Kristensen, L.** & Gulley, J. D. (2009). Surge propagation constrained by a persistent subglacial conduit, Bakaninbreen-Paulabreen, Svalbard. Annals of Glaciology 50 (52). 81-86.

III. Ottesen, D., Dowdeswell, J. A., Benn, D. I., **Kristensen, L.**, Christiansen, H. H., Christensen, O., Hansen, L. Lebesbye, E., Forwick, M & Vorren, T. O. (2008). Submarine landforms characteristic of glacier surges in two Spitsbergen fjords. Quaternary Science Reviews 27. 1583-1599.

IV. **Kristensen, L.**, Benn, D. I., Hormes, A & Ottesen, D. (2009). Mud aprons in front of Svalbard surge moraines: evidence of subglacial deforming layers or proglacial glaciotectonics? Geomorphology 111. 206-221.

V. **Kristensen, L.**, Christiansen, H. H. & Caline, F. (2008). Temperatures in coastal permafrost in the Svea area, Svalbard. Proceedings of the ninth international conference on permafrost, Fairbanks, Alaska. 1005-1010.

VI. **Kristensen, L**, Juliussen, H. Christiansen, H. H. & Humlum, O. (2009). Structure and composition of a tidewater glacier push moraine, Svalbard, revealed by DC resistivity profiling. Boreas 38. 176-186.

Part 1

1 Introduction

1.1 Outline of thesis

This thesis consists of two parts. Part One provides an introduction, background theory and methods, a summary of the results and a general discussion and conclusions. The results chapter is primarily a summary of the papers but also contains previously unpublished measurements. Part Two comprises the full versions of six papers on which this thesis is based. The order of the papers reflects a temporal sequence of processes, first focusing on the glacier surges responsible for the moraine formation, then on the landform assembly, and lastly on the post-deposition preservation potential. The order of the papers therefore does not correspond to the order of their publication. The papers are referred to by their Roman numbers in the text.

1.2 Motivation

Applying observations of modern processes to analysis of older landforms is the principle of uniformitarianism ("the present is the key to the past") first proposed by James Hutton and Charles Lyell (Encyclopedia Britannica Online). The value of studying processes of modern glaciers as a key to understanding the significance of Pleistocene landforms has long been recognized (Boulton, 1972). In this thesis I attempt to investigate glaciers and permafrost interactions in a landsystem consisting of the surging tidewater glacier Paulabreen, Svalbard, and its late Holocene moraine. The moraine was deposited both on land and at the seabed in Rindersbukta and Van Mijenfjorden. The papers range in subject from glaciology and glacial geology to permafrost. How the papers are related and tied to the surge and landforms theme and the concepts of glacier and permafrost interactions is outlined in this introduction and evaluated further in the discussion.

Surge-type glaciers are internally unstable and switch between fast and slow flow. The fast flow resembles ice streams, and slow flow resembling most non-surge-type glaciers. Understanding the trigger and shut down mechanisms may aid the understanding the dynamics for fast-flowing ice streams. Surge behavior is observed from ice-caps and ice-sheets and there is growing evidence for surge behavior of the midlatitude Pleistocene ice-sheets (Hodgson, 1994; Knight, 2004 ; Kovanen & Slaymaker, 2004; Alley *et al.*, 2006 and

Evans *et al.*, 2008). Ice streams play a major role in ice-sheet drainage and surges may also have a large effect on their dynamics. Identifying the landforms that indicate surges correctly is important for glaciological reconstructions.

From work in Svalbard and Iceland models of both terrestrial (Evans & Rea, 1999 and Christoffersen *et al.*, 2005) and submarine (Solheim & Pfirman, 1985 and Ottesen & Dowdeswell, 2006) landsystems of surging glaciers are emerging. Similarities between the land and submarine landform assembly include the presence of crevasse fill ridges, ice-parallel lineations and large end moraines, while large proglacial debris flows and small annual retreat moraines appear to be a feature of only the submarine assembly.

The Paulabreen tidewater glacier system and its associated late Holocene surge moraine comprise an intriguing landsystem. The moraine consists of a mix of marine clays and terrestrial sediments and is in places ice-cored. Several scientific works have been published since the early 20th century, reflecting a long coal mining history and ease of access, but key issues of the moraine formation and timing remained unsolved. Previous studies of surge moraines have focused either on land (Boulton *et al.*, 1996 and Bennett *et al.*, 1999) or at the seabed landforms (Solheim & Pfirman, 1985 and Ottesen & Dowdeswell, 2006). In this thesis I combine the study of the landforms in both environments. This approach has helped cast light on some peculiar landform features where the earlier explanations were inadequate.

During the initial fieldwork period UNIS was informed that Paulabreen was surging again. I took the opportunity to observe this surge by taking daily photographs of the front which were compiled to a time-lapse movie. Twice I had the opportunity to photograph the entire glacier from the air, and this revealed that the surge was initiated in the tributary glacier Skobreen, which then triggered a surge of the lower parts of Paulabreen. Satellite images allowed a further description on the surge development. The crevasse pattern of a glacier is indicative of the stress and strain the ice has experienced, and this is important for the incorporation of sediment and the resultant landforms. Processes observed at the margin of the actively surging glacier were used as analogues for the older moraine system.

Two intriguing landforms made us consider the possibility of submarine permafrost in the fjord. 1) A submarine hummocky moraine that strongly resembled a terrestrial hummocky moraine, which we believed was ice-cored. The question arose whether the submarine hummocky moraine could still be ice-cored. 2) A mud apron consisting of marine sediments and located on land on the distal side of the hummocky moraine. In all previous studies this mud apron has been explained as the seabed being thrust forward by the glacier as a slab or a solid block (Pewe *et al.*, 1981; Rowan et al., 1982; Gregersen et al., 1983 and Gregersen & Eidsmoen, 1988). Rowan *et al.* (1982) suggested that this competence could have been facilitated by permafrost in the seabed. Therefore I wanted to look into the possibility that the mud on the seabed could contain permafrost/ be frozen and if buried glacier ice could be preserved at the seabed.

The concept of glacier and permafrost interaction has recently received some attention within the cold regions Earth Science communities. Application of the concept have been the significance of permafrost beneath ice sheets (Cutler *et al.*, 2000), on glacio-tectonic processes (Waller & Tuckwell, 2005 and Aber & Ber, 2007), ground ice development, rock glaciers, proglacial, ice-marginal processes and permafrost and related processes (Harris & Murton, 2005). Surges in Svalbard have long been recognized to be related to subpolar or polythermal glaciers, i.e. that the outer part of glacier is frozen to its bed (Schytt, 1969 and Hamilton & Dowdeswell, 1996). This is effectively the permafrost penetrating through the glacier where it is thin and thus this type of surges is another example of permafrost and glacier interaction. Surges in Svalbard are thought to propagate by thermal evolution of the bed, by thawing a thin basal layer above the permafrost (Murray *et al.*, 2000). In this thesis we show how a persistent subglacial conduit may act as a barrier of surge propagation by evacuation pressurized water from the glacier bed at the thermal boundary, which supports this theory. Pro-glacial tectonism often occurs in relation to surging glaciers (Klint & Pedersen, 1995 and Boulton *et al.*, 1999) and may occur both in the presence or absence of permafrost. Permafrost will tend to favor brittle over ductile deformation and both types of deformation are often found even within the same strata (Aber & Ber, 2007). The base of permafrost may in some instances form a décollement surface partly because of the buildup of a high water pressure below the permafrost (Boulton *et al.*, 1999). The papers dealing with the moraine system present evidence of different types of glacio-tectonism which reflects both frozen and unfrozen conditions in the ground. Last we consider the preservation of buried glacier ice in the moraines, which is possible due to the permafrost in the ground.

1.3 Aim and objectives

The aim of the study was to study the glacier and permafrost interactions at the surge type glacier Paulabreen, Van Mijenfjorden, Svalbard and its surrounding late Holocene moraine. For this purpose the following set of objectives were defined:

1.3.1 Objectives

- To study a glacier surge in respect to dynamics, stress and strain and surge propagation pattern
- To investigate the structure and composition of a lateral and a frontal moraine deposited by a surging glacier and evaluate the preservation potential of a buried ice core
- To compare landforms of a glacier surge deposited on land and on the fjord bottom respectively
- To investigate whether subsea permafrost is present in Van Mijenfjorden, and discuss if it is responsible for the formation of some of the observed glacial landforms

2 Theory

2.1 Glaciology

Papers I and II deal with glaciological observations and processes during the 2003-2005 surge of Skobreen/Paulabreen and to a lesser extent the 1985-1995 surge of Bakaninbreen. In the following section I outline a few basic concepts within glaciology that are important for the papers and discussion.

2.1.1 The thermal regime of glaciers

Basal sliding and ice creep are influenced by the ice temperature. The most important distinction is whether the ice is temperate or “warm”, meaning that it is at its pressure melting point, or whether it is cold. The pressure melting point differs from 0°C because the melting point is depressed at depth by 0.072°C per MPa (Benn & Evans, 1998). Ignoring a seasonally warmed and cold surface layer, a glacier is classified as temperate if all the ice is at the pressure melting point. If all the ice is colder than the pressure melting point then the glacier is cold or “polar” while if both cold and temperate ice is found, the glacier is polythermal or “subpolar”. Most of the glaciers in Svalbard are polythermal. The typical situation is that the glacier is temperate beneath most of the accumulation area while the margins are frozen to the bed (Liestol, 1976). A polythermal regime significantly increases a glaciers chance of being of surge type (Hamilton & Dowdeswell, 1996 and Jiskoot *et al.*, 2000).

2.1.2 Glacier flow and balance velocity

Glacier flow can take place as ice creep, as sliding on the bed or by deformation of the bed. The flux through a cross section is essentially controlled by mass balance. If a glacier is to maintain its size and shape it must flow with its balance velocity. The discharge Q through a cross-section at the distance x from the highest point on the glacier is:

$$Q(x) = \Sigma(w_x b_x)$$

where w_x is the width and b_x is the specific net balance. The average velocity v through a cross-section is then given by:

$$v(x) = Q(x) / A(x)$$

where A is the area of the cross section (Benn & Evans, 1998). The concept of balance velocity is useful as it relates an idealized flow everywhere on the glacier to the mass balance. Large differences between measured velocities and calculated balance velocities state that the glacier is out of balance and potentially unstable (Clarke, 1987).

2.1.3 Surge

Glaciers of surge type undergo periodic switches between rapid and slow flow without an external trigger. At periods of slow flow their velocity is considerably slower than their balance velocity, so mass is added to the higher parts of the glacier and lost in the lower regions. During a surge, ice is transferred from the upper (reservoir area) to the lower part (receiving area) of the glacier, which lowers the surface gradient and may sometimes cause an advance of the front (Meier & Post, 1969). Surges last typically 1-10 yrs and are followed by a much longer quiescent period. Ice velocity is typically 10-100 times faster during surges than in the quiescent phase, though there are considerable variations in these figures (Frappe & Clarke, 2007). While surges often occur at uniform intervals (Meier & Post, 1969) changes in mass balance have also been suggested to affect the frequency of glacier surges (Dowdeswell *et al.*, 1995) and may change a surge-type glacier to a non-surge-type glacier (Hansen, 2003).

2.1.3.1 Surge mechanisms

Surges occur because the ice velocity during the quiescent phase is lower than the glacier's balance velocity. The reason why surge-type glaciers flow 'too slow' and the exact mechanisms that trigger a surge are not fully understood. The surge phenomenon has been reviewed in three important papers; Meier & Post (1969); Raymond (1987) and Harrison & Post (2003). Some well studied surges are the 1982-1983 surge of Variegated Glacier, Alaska (Kamb *et al.*, 1985), the c. 1980-2000 slow surge of Trapridge Glacier, Yukon Territory, Canada (Frappe & Clarke, 2007) and the 1985-1995 surge of Bakaninbreen, Svalbard (Murray *et al.*, 1998).

Two main theories of surge behavior have been proposed, which attempt to explain large changes in basal shear strength in response to hydrologic switch and a thermal switch, respectively.

The hydrologic switch mechanism was proposed by Kamb *et al.* (1985) and Kamb (1987) on basis of detailed field observations during a surge of the temperate Variegated Glacier in 1982-1983. Velocities, water pressure and meltwater discharge were measured throughout a surge and water tracer experiments were performed. The surge was initiated by several mini-surges during the summertime while the main surge started in January. Periods of high speed were associated with high basal water pressure; at times borehole water levels were high enough to float the glacier. At the surge termination a pronounced drop in basal water pressure and large discharge floods were observed. Tracer dye experiments indicated a long transition time for meltwater (velocity 0.02 m/s) and wide lateral dispersion during the surge, while after the surge the water velocity was higher (0.7 m/s) and no lateral dispersion was observed. Relatively high discharge volumes of turbid (basal) water in combination with the long transition time during the surge ruled out the possibility of drainage through tunnels.

Kamb *et al*. (1985) proposed that during a surge, inefficient drainage in a linked cavity system led to a high basal water pressure, which reduced basal friction and encouraged rapid sliding. Kamb (1987) showed mathematically that a linked cavity system can be stable in case of rapid basal sliding. The theory does not explain how a linked cavity system develops in the first place, but he suggested that it initiates during winter where the water flux is low and thus does not cause the formation of a normal drainage pattern of low pressure conduits. This is in agreement with the onset of the main surge but not with the initial mini-surges that occurred during summertime. The theory was proposed for a hard bed, but may apply for a till bed, if there are stationary features on the bed that can create lee cavities under the glacier. This disagrees with observations of a readily deformable soft bed below several surge type glaciers; amongst those also Variegated Glacier (Harrison & Post, 2003).

A thermal switch mechanism has been suggested to account for surges in polythermal glaciers. The cold marginal ice in a polythermal glacier may act as a barrier which occasionally may be broken leading to a surge, possibly connected to trapped water upglacier from the cold margin (Schytt, 1969 and Clarke, 1976). Clarke *et al*. (1984) showed that a bulge, later defined as a surge bulge by Frappe & Clarke (2007), on Trapridge glacier was located at the transition between warm basal conditions upglacier and cold basal conditions downglacier. They found evidence of a well developed drainage system at the surge bulge and thus no trapping of water. A similar thermal regime was found on either side of a surge bulge of Bakaninbreen (Murray & Porter, 2001). Here, however, the basal water pressure was often near or above the flotation pressure on either side of the surge bulge indicating a poorly drained bed. Murray & Porter (2001) suggested that in a polythermal glacier, mass builds up in the reservoir area and is lost in the receiving area during the quiescent phase; thus the glacier becomes steeper. This leads to increased basal shear in the upper parts of the glacier, which creates heat and meltwater – further increasing basal sliding as a positive feedback. This is essentially the surge mechanism. The fast flow propagates up and downglacier and the frontal advance depends on the rate that the cold bed can be warmed. Fast flow terminates when to little water is present at the bed to facilitate fast sliding. This could possibly occur by leaking of basal water through holes in the permafrost (Smith *et al.*, 2002).

Björnsson (1998) showed that the drainage of a subglacial lake was much slower during a surge than normally and attributed this to a linked cavity drainage system during the surge in comparison with a normal conduit system. Clarke *et al.* (1984) suggested that a destruction of the normal basal drainage within the basal substrate due to deformation could trigger a surge in a polythermal glacier. Interestingly Murray & Porter (2001) reported a weak correlation between basal water pressure and basal sliding and sediment strength towards the end of the surge of Bakaninbreen. Nevertheless, surges in both temperate and polythermal glaciers appears to be depend on a poorly drained glacier bed.

A deformable till has been found at the sole of several surge type glaciers, namely Trapridge Glacier, Yukon Territory (Clarke *et al.,* 1984), Bakaninbreen in Svalbard, Black Rapids Glacier and probably also Variegated Glacier, Alaska (Harrison & Post, 2003). This is in line with statistical analysis by Hamilton & Dowdeswell (1996) and Jiskoot *et al.* (2000) from Svalbard finding that glaciers on easily erodable sedimentary bedrock is more likely to be of surge-type than glaciers on other lithologies. It seems that surges most often occur on deformable soft beds.

The apparent contrast between the hydrologic switch and the thermal switch mechanisms made Murray *et al.* (2003b) suggest that the surge mechanisms could be altogether different on temperate and polythermal glaciers. This point was supported by Jiskoot & Juhlin (2009) based on the contrast between the slow surge of the polythermal Sermeq Peqippoq and many other East Greenland glaciers compared to the larger temperate Sortebræ glacier, which experienced a more rapid surge. To the contrary Clarke *et al*. (1984) and Frappe & Clarke (2007) pointed out that both temperate and polythermal surge-type glaciers exists in St. Elias Mountains, Canada. They found it unlikely that two entirely different mechanisms should be responsible for the surge-behavior in the same regional cluster of surging glaciers.

2.1.3.2 Patterns of surge propagation

Surges can initiate in the upper part of the glacier and propagate downglacier (Raymond *et al.,* 1987); in the lower part and travel upglacier (Dowdeswell & Benham, 2003; Murray *et al.,* 2003a and Murray *et al.*, 2003b) or initiate in the central part and travel downglacier (Murray *et al.*, 1998) or upglacier (Pritchard *et al.*, 2005). There is some evidence that the surges of tidewater glaciers are usually initiated in the front and propagate upstream while surges in glaciers terminating on land start further up and propagate downstream which could be due to less restriction to flow at the front of tidewater glaciers (Dowdeswell & Benham, 2003 and Murray *et al.*, 2003b). Hagen *et al.* (1993) observed that the surges of tidewater glaciers usually affects the entire glacier system, while often only a single flow unit is affected for glaciers terminating on land. In combination, these observations may imply that when the surge propagates upstream it tends to draw in the tributaries while when the surge propagates downglacier it leaves the tributaries unaffected. In paper II we discuss the mechanisms of surge propagation.

2.1.3.3 Occurrence of surging glaciers in Svalbard

Glaciers of surge-type tend to cluster in certain regions of the world and Svalbard is such a region. The estimate of how many of the glaciers in Svalbard that are of surge-type ranges from 13% (Jiskoot *et al.*, 1998) to 90% (Hagen *et al.*, 1993) while Hamilton & Dowdeswell (1996) in a thorough study limited to of a part of Spitsbergen suggested 36.4%. As both the active surge and the quiescent period are estimated to be somewhat longer lasting for Svalbard glaciers (50-500 yrs) than for surge type glaciers in other regions of the world (Dowdeswell *et al.*, 1991), the number of glaciers in Svalbard that are of surge-type is likely to be underestimated in statistical analysis, where only observed surges are included. Nuth *et al.* (2007) found from analysis of maps made on the basis of aerial photos in 1936/38 and 1990 that most glaciers on Spitsbergen had in general lost mass but many had thickened in their upper parts. This could be attributed to increased precipitation or dynamics (as a build up to a surge). Sund *et al.* (2009) concluded from studies of mass displacement over time on 50 Svalbard glaciers that also "partial surges" are common. The works of both Nuth *et al.* (2007) and Sund *et al.* (2009) suggest that the statistical studies have probably underestimated the percentage of Svalbard glaciers being of surge type.

2.1.4 Crevasses on glaciers – in particular in relation to surges

Crevasses form when a tensile stress exceeds the tensile strength of the ice. They open in the direction of the stress, so the crevasse pattern is indicative of the stress and strain pattern in the glacier. Glaciers undergoing surges become heavily crevassed. Most crevasses that form during a glacier surge, are extensional (transverse) (Herzfeld *et al.*, 2004) because the glacier is stretched when ice is transferred from a glacier's upper region to its lower region. Longitudinal or flow parallel crevasses form due to compression and in surges are related to shortening associated with a downward advancing surge front (Lawson *et al.*, 1994). Murray *et al.* (2003b) however suggested that a downward moving surge front does not necessarily cause compression. If the thermal evolution of the bed advance downstream faster than the ice velocity, then nothing restricts the ice progression and compression would not occur. In paper I we investigate the crevasse pattern that formed on Skobreen/Paulabreen during its latest surge and use this to infer the initiation point of the surge as well as the basal thermal conditions.

2.2 Glacial landforms

Glaciers are powerful tools for landscape modification. The Svalbard landscape is a good example, where repeated Pleistocene glaciations have eroded deep and wide U-shaped valleys and dumped the sediment primarily in trough-mouth fans of the shelf break (Ottesen *et al.*, 2007). At the Last Glacial Maximum (LGM) the Svalbard-Barents Sea ice sheet reached the shelf edge west and north of Spitsbergen (Ottesen *et al.*, 2007). Isfjorden and Van Mijenfjorden were deglaciated to near the present level of glaciation ~ 10.000 yrs BP (Salvigsen & Winsnes, 1989 and Svendsen *et al.*, 1996), but glaciers today still cover about 60% in Svalbard (Hagen *et al.*, 1993).

In Svalbard the depositional landforms of Pleistocene age are mostly found on the shelf or in the fjords, though a few key sections on land give information on the deglaciation (Mangerud *et al.*, 1998). Glaciers in Svalbard grew again after mid Holocene and many probably attained their maximum Holocene position during the Little Ice Age (Humlum *et al.*, 2005 and Mangerud & Landvik, 2007). The young moraines allow for studies of modern glacial deposition in detail, and numerous workers have taken advantage of this since the classic works of Gripp (1929) and Boulton (1972). The fluctuating margins that are the result of surge activity allows for studies at various stages of advance and retreat.

Landscape modification by glaciers is composed of erosion, debris entrainment and transport as well as deposition. I will not write further on erosion as it is not important for the thesis, but will shortly comment on entrainment and transport and then go into the landform assembly. I have used primarily but not exclusively Svalbard examples in the review.

2.2.1 Debris entrainment

The origin of sediment can often be inferred from clast shape and roundness (Evans & Benn, 2004). Rockfall or avalanches onto glaciers supply a significant amount of sediments to the surface of valley glaciers in Svalbard (Hambrey *et al.*, 1999) and as this sediment is transported in primarily a supraglacial or englacial position it tends to be angular. Another sediment source is at the glacier sole. The sediment must be entrained into the glacier for significant amounts to be transported. Regelation around basal obstacles can entrain small amounts of basal debris. Basal freeze-on at the sole may also occur by conductive cooling or hydraulic supercooling (Alley *et al.*, 1997 and Cook *et al.*, 2006). Thrusting and folding are yet other mechanisms for bringing basal sediment into a glacier (Boulton, 1970), which are both favored by a sharp stress gradient and compressive flow (Alley *et al.*, 1997). Such conditions are typical for a downward propagating surge fronts and may also be found in polythermal glaciers in places where temperate ice meets cold ice that is frozen to the bed (Hambrey & Huddart, 1995).

Incorporation of sediments in the glacier by thrusting has been proven for a number of glaciers in Svalbard (Murray *et al.*, 1997 and Hambrey *et al.*, 1999) though there is a controversy on its overall geomorphic significance of the process (Hambrey *et al.*, 1997 and Lukas, 2005). This subject is discussed in greater details in paper VI. Sediment incorporation can also occur by squeezing of sediments into basal crevasses (Woodward *et al.*, 2002). Alley *et al.* (1997) states that while this process may be locally important; the stress state is rarely conducive for the formation of basal crevasses. We present evidence for the presence of basal crevasses within buried ice in paper VI. A requirement for the formation of basal crevasses is a basal water pressure that almost equals the ice overburden pressure (van der Veen, 1998), which are the type of conditions that are likely beneath surging tidewater glaciers in Svalbard (Kamb, 1987 and Engelhardt & Kamb, 1997).

2.2.2 Sediment transport

Sediment-transport by a glacier can be supraglacial, englacial, subglacial or by deformation of a soft bed. Much sediment will also be transported by water. Sediment transported supraglacially or englacially will undergo limited modification while sediments transported at the ice-rock interface are subject to crushing and abrasion.

Sediment in deformable beds is also transported by glaciers, though it is not directly entrained in the ice. Deformable beds are important for both ice dynamics and for the formation of glacial tills (Boulton & Jones, 1979; Hart, 1995; Murray, 1997 and Evans *et al.*, 2006). Attention has lately been drawn to the idea of sticky spots where deformation does not occur. These points are able to support the ice-sheet by impeding ice motion and thus affect the dynamics. Deformation is now thought to occur in a time transgressive rather than pervasive manner (Piotrowski *et al.*, 2004 and Stokes *et al.*, 2007). Where a glacier advances over lacustrine or marine deposits deformation will more likely be pervasive (Alley *et al.*, 1997) which is the type of conditions that meet the surging tidewater glaciers of Svalbard. Subglacial tectonism is a low-grade deformation where the end product – a glaciotectonite - is not homogenized and where inherent structures are preserved while it still bears signs of deformation (Evans *et al.*, 2006). Proglacial tectonism is the term for sediment or rocks in the foreland that is transported (deformed) by the glacier.

In paper IV we describe a section that bears evidence of both proglacial and subglacial tectonism. We reinterpret a proglacial landform as a result of both gravitational and tectonic processes and pushed in front of the glacier in a soft mode rather than being thrust forward as competent slab of the seabed. The origin of this and its subsea counterpart, we suggest, reflects bulldozing in front of the glacier rather than from an extruded deforming layer as previously suggested (Hald *et al.*, 2001).

2.2.3 Landforms of surging glaciers

Surging glaciers themselves may easily be recognized by their crevasses and looped moraines, but following deglaciation the landforms deposited are more subdued. The landforms are not individually exclusively indicative of surges and a land-system approach for identification of surges is useful. Typical features includes crevasse fill ridges, concertina eskers, ice-parallel streamlining and thrust or push moraines (Evans & Rea, 1999; Bennett *et al.*, 1999 and Christoffersen *et al.*, 2005). Push- or thrust moraines have been given particular attention as they can be used as analogues for the Pleistocene moraines found in mid-latitudes (Bennett, 2001) whereas small landforms such as crevasse fill ridges might have disappeared since deglaciation.

There is some confusion around the term “push moraine”. Bennett (2001) defined push-moraines as glacio-tectonic ice-marginal moraines and it is used in that sense by Humlum (1985), Croot (1988), Lønne & Lauritsen (1996), Etzelmuller *et al.* (1996), Hart & Watts (1997), Boulton *et al.* (1999) and Bakker & Meer (2003). Huddart & Hambrey (1996) and Benn & Evans (1998) argues that the term push moraine should be reserved for bulldozing of soft sediment while thrust moraine should be used for tectonic deformation (brittle or ductile) in the foreland, and this terminology is used by Pedersen (1996), Bennett *et al.* (1999), Christoffersen *et al.* (2005) and Evans *et al.* (2008). Unfortunately I have myself been inconsistent. In paper VI I have used “push-moraine” for the combined complex of an ice-cored lateral moraine and its distal pushed apron. In the paper IV I have used the term “moraine” for the combined frontal moraine complex including the partly ice-cored part and the proglacial mud apron. For the remaining part of this introduction I will use “thrust-moraine” where there is evidence of tectonic deformation and “push” for glacial bulldozing in line with Benn & Evans (1998).

In Svalbard thrust moraines indicative of ductile deformation appear to be restricted to glaciers terminating below the Holocene marine limit. The reason is that salt and marine sediments weaken the permafrost sediment strength and allow for ductile deformation below the marine limit (Etzelmuller *et al.*, 1996, Hart & Watts, 1997). Ductile deformation is typical for frozen marine sediments compared to brittle deformation of glacioflucial sediments (Huddart & Hambrey, 1996).

2.2.4 Submarine landforms of surging glaciers

The submarine landform assembly of surging glaciers seems to resemble terrestrial counterparts apart from two additional features. These are large proglacial debris flows and small annual retreat moraines (Liestol, 1977, Ottesen & Dowdeswell, 2006). The submarine landforms of surges in Rindersbukta (and Van Keulenfjorden) are described in details in paper III and in paper IV the landforms are compared with their terrestrial counterparts.

2.2.5 Ice-cored moraines

Ice-cored moraines mark the maximum Holocene extent of most of Svalbard's glaciers. They exist because the margins of Svalbards surging and/or polythermal glaciers contain large amounts of sediments, as outlined above, which melt out on the glacier surface and prevents the ice from melting. The ice-cored moraines are up to 50 m high and their debris cover is between 0.1 – 4 m (Etzelmuller *et al.*, 1996; Lysa & Lonne, 2001 and Lukas *et al.*, 2005). They are also typically found on the proximal side of the proglacial thrust moraines (Hambrey & Huddart, 1995; Hart & Watts, 1997; Bennett *et al.*, 1999 and Boulton *et al.*, 1999). As most of Svalbard's glaciers have an ice-cored morarine, they are not exclusively found in relation to surging glaciers. In paper VI we present the internal structures of the largely ice-cored Crednermorenen. In Damesmorenen (paper IV) we see small remnants of an ice-core, but melting is largely complete here.

2.2.5.1 Mass wasting processes in ice-cored moraines

The end product following complete melting of buried ice debris covered moraines is hummocky moraines. Two mass wasting processes account for the melting, 1) down-wasting, which is defined as melting from the top (or bottom) ice surfaces and 2) back-wasting, which is lateral retreat of nearly vertical ice-walls (Kruger & Kjaer, 2000). Down-wasting rates decrease as a debris cover increases. First a thin layer of debris on a glacier actually increases the melt-rate due to increased absorbtion of short wave radiation, but when the layer reaches a thickness of about 2 cm, the debris cover insulating effect becomes larger than the effect of increased energy absorbtion and the melt-rate is reduced. This was empirically described and measured by Ostrem (1959).

Ice-melting triggers a range of redeposition processes such as fall, slumping and flowing (Kruger & Kjaer, 2000). Debris flows may depend more on sediment water content than of slope angle, and flow can expose pure glacier ice which increases melting (Lukas *et al.*, 2005). In Svalbard and other permafrost environments, the buried glacier-ice is partly preserved by the permafrost. Bottom wasting does not occur and down-wasting ceases when the sediment cover thickness had reached the active layer thickness. As debris is gradually released on the glacier surface, backwasting takes over at the dominant or only melting process of the glacier ice.

2.2.5.2 Mass wasting rates in ice-cored moraines

Mass wasting from an ice-cored moraine in a non-permafrost environment (Iceland) was between 1987-1995 measured to 2.5m/yr, and an average back-wasting rate of 6.9 cm/day (Kruger & Kjaer, 2000). Interestingly the down-wasting on Iceland was dominated of bottom melting over top-melting, which reflects the lack of permafrost in the foreland and probably also the high geothermal heatflux found in Iceland. Investigated over an area, bottom melt and backwasting were of comparable magnitude while top melting was much smaller (Kruger & Kjaer, 2000). Attempts to quantify the rate of surface lowering of ice-cored moraines in the permafrost of Svalbard have also been made. Etzelmuller (2000) subtracted photogrammetrically derived DEMs from four moraines at some time intervals and found an annual lowering rate of 0.1 – 0.2 m/yr (Erikbreen), while down-wasting at Vestre Lovénbreen, Austre Brøggerbreen and most of Finsterwalderbreen was limited. Back-wasting was significant where till was flowing and along fluvial channels. At the margin of the debris covered zone of Finsterwalderbreen (thin debris cover) surface lowering was up to 3m/yr. Schomacker & Kjaer (2008) found a backwasting rate of 9.2 cm/day for the ice-cored part of the moraine of Holmströmsbreen while the overall surface lowering was 0.9m/yr between 1984-2004. From Larsbreen Lukas *et al.* (2005) found backwasting rates of 3.5 – 7.8 cm/day.

Back-wasting rates are thus similar in the Svalbard permafrost environment and on Iceland while down-wasting rates are lower. Etzelmuller (2000) suggest that a rim of ice-cored moraine in Svalbard would disappear within a period of 200 yrs, though surges may have deposited new moraines before they disappeared. The largely ice-cored Crednermorenen, described in paper VI has an age of ~ 600 yrs, and this suggests that Etzelmuller (2000) underestimated the survival time of at least some of the ice-cored moraines in Svalbard. In the results section I present some unpublished measurements of the backwasting rate on Crednermorenen.

2.3 Permafrost

Permafrost is defined as ground (soil or rock and included ice or organic material) that remains at or below 0°C continuously for at least two consecutive yrs (http://ipa.arcticportal.org/). As this definition is based purely on temperature, both temperate and polar glaciers would fall into the category of permafrost. Nevertheless glaciers are normally implicitly excluded from the permafrost definition.

The heat flow in and out of the ground us governed by the surface energy balance. The net radiation Q* is given by

$$Q^* = S^{\downarrow} - S^{\uparrow} + L^{\downarrow} - L^{\uparrow}$$

where S is shortwave (solar) radiation and L is longwave (thermal) radiation. The arrows indicate the direction of the radiation. The net-radiation is usually positive in daytime and negative at nighttime. The equation shows how a high albedo ($S\uparrow/S\downarrow$) gives a smaller net radiation.

The net-radiation is divided into energy fluxes as seen in the energy balance equation; here shown is a simplified form:

$$Q^* = Q_L + Q_H + Q_G$$

where Q_L = is the latent heat, Q_H is the sensible heat and Q_G is the heat conduction to and from the ground. How Q* is divided is highly depended surface conditions, meteorological conditions, etc. see Boike *et al.* (2003). The energy flux in the ground (considering only conduction) is given by:

$$Q_G = -K\frac{dT}{dz}$$

where K is the thermal conductivity, T is temperature and z is depth. Detailed descriptions are given by Oke (1987) or Williams & Smith (1989). What is important is that the presence or absence of permafrost as well as the permafrost temperature is not only given by the mean annual air temperature (MAAT) but is also governed by eg. albedo, surface type, snow-cover etc. The thickness of permafrost is governed by the surface temperature history, the ground thermal conductivity and heat capacity and the geothermal heat flux.

2.3.1 Permafrost in Svalbard

About 25% of the terrestrial part of the earth surface is underlain by permafrost (French, 1996). In Svalbard permafrost is thought to be continuous (possibly with the exception of Bjørnøya); a review of the permafrost in Svalbard is given by Humlum *et al.* (2003). Permafrost thickness is less than 100 m at the coasts and more than 500 m in the mountains. Beneath polythermal glaciers, lakes and fjords taliks (unfrozen ground) are usually present. Taliks and ground water movement in Svalbard is evidenced by pingos and springs that are found in many valleys – in particular near the terminus of polythermal glaciers (Liestol, 1976).

2.3.2 Permafrost in the shore area and subsea permafrost

Permafrost may also be found in the seabed. Globally most submarine permafrost is found on shallow shelves and originated when they were subaerially exposed at periods of low global sea-level – most importantly during the Last Glacial Maximum (LGM) (Osterkamp & Harrison, 1982; Nixon, 1986; Williams & Smith, 1989 and Løvø *et al.*, 1990). The authors use the term 'permafrost' for permanently *frozen* sediments and not the thermal definition. When submerged, salt will infiltrate the seabed, reduce the freezing point and thaw ice present, even while the temperature may remain below 0°C. Rachold *et al.* (2007) found an infiltration rate of 1.2 cm/yr based on modern erosion rates in the shallow Laptev Sea. They define permafrost exclusively thermally and when describing additionally state whether the sediment is frozen or unfrozen; I will follow this usage.

Taliks are usually present beneath lakes and fjords. Unless the ice freezes to the bottom, the water prevents heat from escaping the ground during winter, which strongly affects the local ground heat flow (Williams & Smith, 1989). The effect of fjords (and polythermal glaciers) on the ground temperature in Svalbard was modelled by Werenskiold (1953) who assumed a steady state condition, -8°C temperature boundary for the land surface and a 0° temperature boundary for the water temperature. He found that beneath broad fjords or glaciers permafrost would be absent. (Gregersen & Eidsmoen, 1988) measured the borehole temperature at the shoreline in Longyearbyen and Kapp Amsterdam, near Sveagruva. At Kapp Amsterdam they found subzero temperatures down to more than 100m. Using -6°C and +1°C as land- and sea temperature boundaries respectively Gregersen & Eidsmoen (1988) estimated that permafrost only extends c. 50m horizontally into the fjord.

July temperatures were measured in two basins in Van Mijenfjorden to -1.5°C and -1.3°C (Gulliksen *et al.*, 1985). Sea-water freezes at c. -1.9°C and while sea-ice is present c. December to June the entire water column is isothermal. The unstratified water column during winter and the strong stratification during the summer can be seen in fig. 1. The water temperatures are lower than those used in the models of Werenskiold and Gregersen & Eidsmoen and points to that permafrost may indeed be present in Van Mijenfjorden. In paper V we attempted to model the near-shore permafrost conditions. It was not included in the model whether the permafrost is in a frozen or a thawed state. If the sediment has a salinity comparable to that of seawater, the freezing point of the sediment is even lower than the freezing point of sea-water, because capillarity and adsorption reduces the freezing point further than the salt content alone (Williams & Smith, 1989). Thus a marine deposit may be unfrozen even if the temperature remains below zero during the entire year. A less saline sediment, however, could be frozen in the temperature interval -1.9°C to 0°C. Either it could be a glacial deposit as we see in Van Mijenfjorden, where the sediment is carried from the terrestrial parts of the glacier system or it could have a reduced salinity due to outflow of fresh ground water.

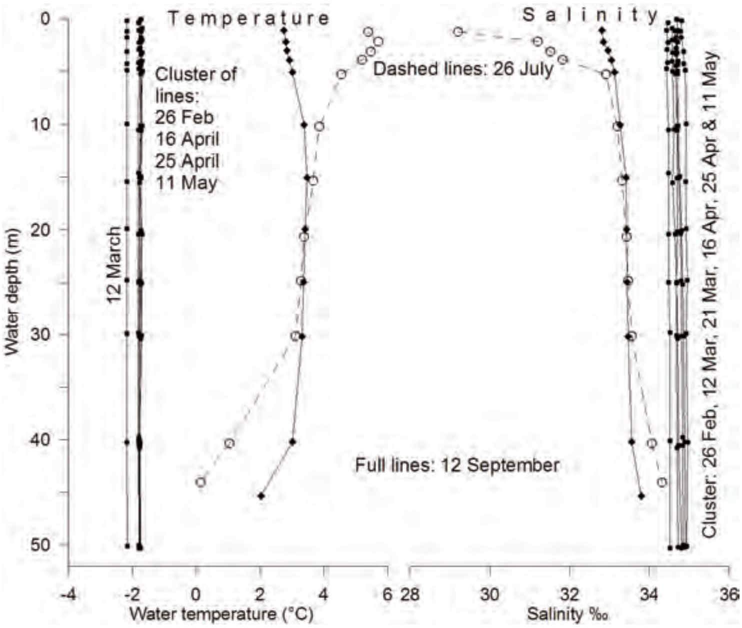

Figure 1 Water temperature and salinity near the mouth of Rindersbukta in 2007. Notice the strong stratification in the summertime, 26 July 2007, and the isothermal and isohaline conditions during the winter. Data from Sanna Markkula, Master thesis at UNIS.

2.3.3 The possibility of buried glacier-ice below the seabed

An irregular and hummocky topography on the seabed led to the idea that glacier-ice might be buried at places in the seabed in Van Mijenfjorden. For glacier ice to remain at the seabed it requires both that 1) the ice does not melt and 2) the ice does not float up to the surface. For the ice not to melt, permafrost has to exist on the seabed. A sediment layer may cover the ice and act as an active layer protecting the underlying ice. For the ice not to float up, the sediment content has to be high. The sediment concentration (C) where the ice would be neutrally boyant can be found by:

$C = (\rho_w - \rho_i)/(1 - \rho_i/\rho_s)$ (Gilbert *et al.*, 2004) where

ρ_w is the density of sea water (1.025g/ cm^3)

ρ_i is the density of glacier ice (here I use 0.870 g / cm^3)

ρ_s is the density of the debris minerals (here I use 2.6 g/ cm^3)

C is 0.233 (233 g/l) corresponding 9 vol % sediment.

Sediment concentrations of this magnitude have been found in basal ice less than 2 m above the bed in Greenland and Norway (Rea *et al.*, 2004 and Knight *et al.*, 2000). Grab samples in front of Swift Glacier, Weddel Sea, Antarctica has contained submerged iceberg fragments (Gilbert *et al.*, 2004) but these were suspected to be the most sediment rich parts of larger icebergs that were broken off and sunken. It may not be realistic for larger parts of glacier ice to have sediment content this high. Sediment covering the ice may act against the floating, but in paper IV we bring forward a different explanation of the hummocky seabed topography.

3 Study site

The study area for this thesis is the Paulabreen Glacier System (PGS) which calves into Rindersbukta, inner Van Mijenfjorden, Svalbard and the moraine that was deposited by a surge of Paulabreen ~ 600 yrs BP. Paper III additionally presents data from the inner part of Van Keulenfjorden, but for the Van Keulen area my contribution was to find old maps and satellite images and draw past glacier-margin positions.

3.1 Geographical outline

Svalbard is an archipelago north of Norway located from 74° to 81° North and 10° to 35° East. North of Svalbard is the Arctic Ocean; west is the Greenland Sea, south is the Norwegian Sea and east is the Barents Sea. Spitsbergen is the largest island in Svalbard, with an area of 39.044 km^2 out of a total of Svalbard of 62.700 km^2. Van Mijenfjorden is a 60 km long fjord on the west coast of Spitsbergen. It is 12 km wide in the outer parts narrowing to 2 km in inner Rindersbukta. Van Mijenfjorden originates in the sound of Bellsund and is separated from Bellsund by the 8 km long and 600 m wide island Akseløya, which leaves only two narrow and shallow straits as passages into the fjord.

Two valleys meet at about 90° angle at the head of Van Mijenfjorden. To the north-east is Braganzavågen, a tidal flat that continues into Kjellstrømsdalen which is a wide U-shaped valley with a braided river typical for Svalbard. To the south-east is Rindersbukta – the innermost branch of Van Mijenfjorden. At the head of Rindersbukta the valley Paulabreen calves into Rindersbukta.

From the settlement Sveagruva, the biggest coal-mine in Svalbard (Svea Nord) is operated by Store Norske Spitsbergen Kulkompani (SNSK). The settlement is partly situated on the moraine of Paulabreen and so is most of its infrastructure, such as its airport and its coal shipping harbor at Kapp Amsterdam. Sveagruva lies approximately 45 km SE of Longyearbyen and houses about 200 workers at all times. Daily flights and a marked snow-mobile route as well as the possibility for accommodation allows for easy access to the area all year. Coal mining in Sveagruva was initiated in 1916, first as a Swedish company and from 1934 by the Norwegian company SNSK. For this reason most of the early literature on the study area is Swedish.

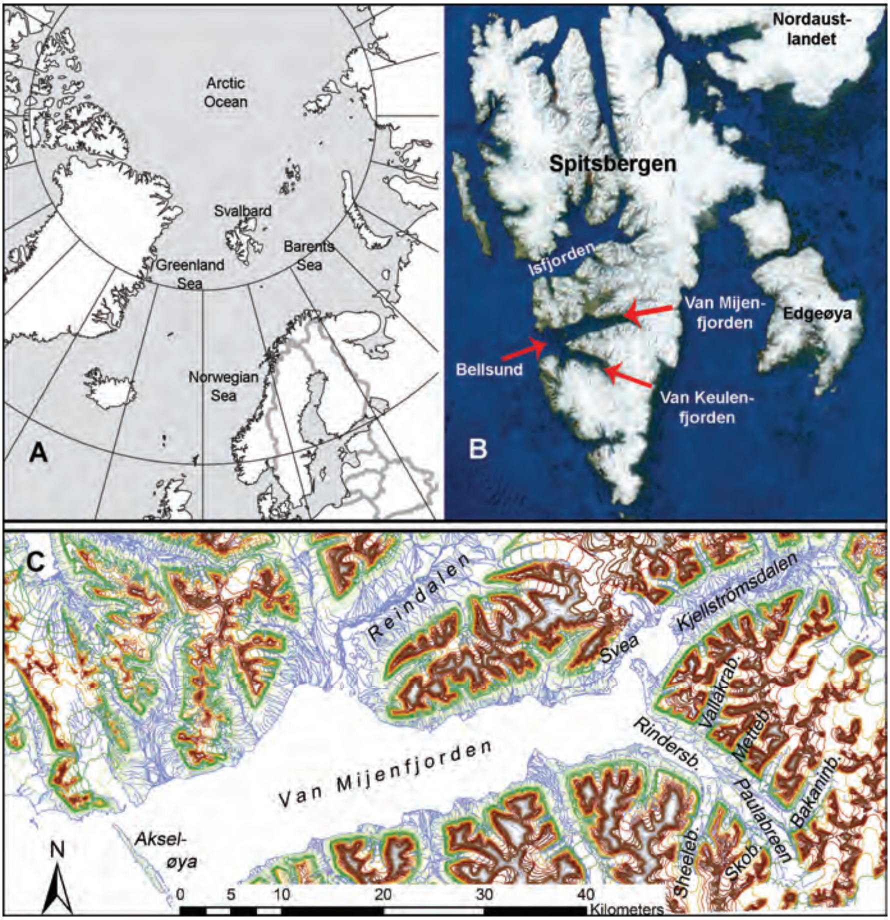

Figure 2 Overview map showing A: Svalbard's location in the north Atlantic; B The location of Van Mijenfjorden and Van Keulenfjorden on Spitsbergen and C: Van Mijenfjorden, Rindersbukta and PGS.

3.2 Meteorology in Svalbard and in Sveagruva

Svalbard has a polar tundra climate and can also be characterized as a polar desert.

The meteorological record of Svalbard is nearly 100 yrs long with the earliest observations being a series from 1911 to 1930 from Finneset, Grønfjorden near the Russian town, Barentsburg. A composite temperature series adjusted to the current Svalbard Airport meteorological station has been compiled from several shorter series from the Isfjorden area (Førland *et al.*, 1997). Work using several daily rather than monthly values as well as a dataseries from different altitudes near Longyearbyen led to a revised composite record that, however, does not change the annual or seasonal trends,

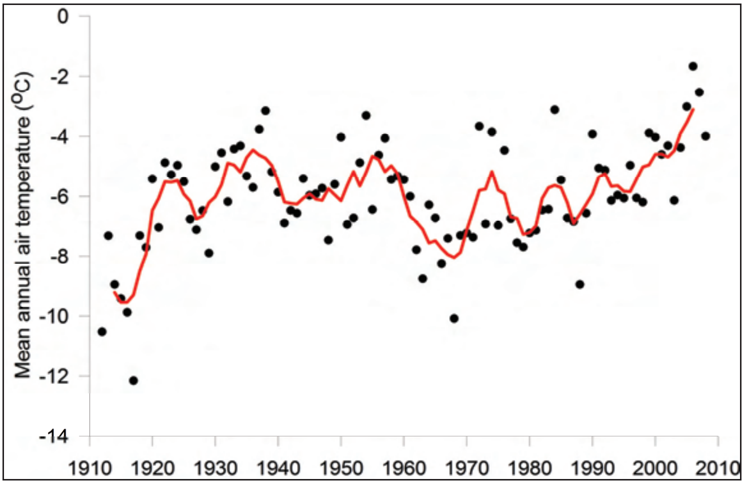

Figure 3 Mean Annual Air Temperatures in Longyearbyen from 1912 to 2008. The red curve is a 5 yrs moving average.

but highlights the importance of temperature inversions during winter (Nordli & Kohler, 2003). Fig. 3 shows the homogenized series of Førland *et al.* (1997). Observable in the meteorological record are the large climatic variations in Svalbard within the twentieth century. The sensitivity is most likely a result of the location as the northernmost tip of the North Atlantic Drift flows on the west coast of Svalbard, and Svalbard is also located on the North Atlantic cyclone track. The sensitivity is enhanced by variations in sea-ice extent, which is coupled to both atmospheric and oceanographic circulation (Humlum *et al.*, 2005). One of the largest changes was a sudden temperature rise ~ 1920 where most of the warming occurred during the winter. About 2/3 of this warming could be explained by an increased cloud cover during winter (Nordli & Kohler, 2003).

A meteorological station has been in operation in Sveagruva since May 1978. It lies 9 m above sea-level (asl.) at the airstrip. Fig. 4 shows a comparison with Longyearbyen Airport (in operation since August 1975, 28 m asl.) of monthly temperatures and precipitation. Temperatures are slightly lower in Sveagruva with a Mean Annual Air Temperature (MAAT) of -7.1°C compared to -6.7°C in Longyearbyen airport. The 30 yrs period is 1961-1990 though none of the stations were operating throughout this time. The winter temperatures are lower in Sveagruva compared to Longyearbyen, while summer temperatures are almost identical. Van Mijenfjorden freezes early every winter compared to Isfjorden, which freezes more rarely and for much shorter time-periods. Perhaps the higher winter temperatures in Longyearbyen can be explained by the release of heat from the nearby Isfjorden. Precipitation (260 mm/yr) is higher in Sveagruva than in Longyearbyen (190 mm/yr). Precipitation is difficult to measure correctly and even harder in the arctic where most precipitation falls as snow, which particularly at high wind speeds tends not to be collected in the gauges (Humlum, 2002). However, observations suggest that snow cover is both thicker and longer lasting in Sveagruva compared to Longyearbyen and the glacier-cover more extensive.

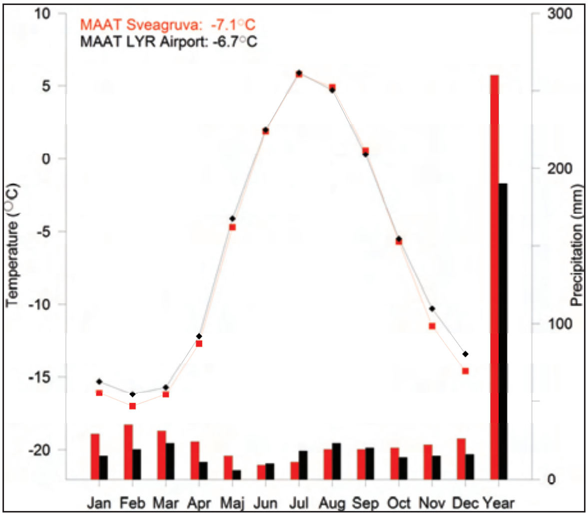

Figure 4 A comparison between the temperature (curve) and precipitation (bars) in Longyearbyen (black) and Sveagruva (red), based on 1960-1990 data (www.met.no).

In the papers I have given MAAT and precipitation data for different shorter time-periods. By giving an outline here of the downloaded data-series for Sveagruva, I try to explain why. From May 1978 to 31 Jan 2003 a manual station was operated in Sveagruva. The data-series includes temperature, wind, precipitation, humidity and various other parameters (cloud-parameters etc.) and snow thickness was also measured. In general the data appear to be of good quality. Until 1 February 2002 the data were recorded daily at 07.00, 13.00 and 19.00. When I have calculated temperature averages from this period I have used only the 07.00 and 19.00 values as the missing night value would cause a too high average temperature.

An automatic weather-station has been in operation from December 2001. A number of parameters were not recorded by the automatic weather-station – most importantly neither precipitation nor snow-depth were recorded. There are several gaps and errors in the data-series and strangely there are large differences in the downloaded data depending on which settings are selected in Eklima; different settings have caused different gaps in the time-series. I have only recently been aware of this fact. When calculating I have tried to avoid the periods of gaps and errors, which has led to averaging over rather short time periods.

3.3 Oceanography in Van Mijenfjorden

The oceanography in Van Mijenfjorden is strongly affected by the island at the fjord mouth, Akseløya (Fig. 2). Together with the narrow and shallow straits on either side it forms a sill that nearly blocks the water exchange between the fjord and the warmer Atlantic water outside (Nilsen, 2002). The water column is thus dominated by cold local water. Shore-fast ice is usually present from December to June. Bottom water salinity is ~ 34‰ (Hald *et al.*, 2001), and July temperatures of -1.5°C and and -1.3°C have been measured in two basins in Van Mijenfjorden at 112 m and 74 m depth (Gulliksen *et al.*, 1985). As Fig. 1 indicates, Sanna Markkula measured a bottom temperature of 0.1°C in a shallower basin (44 m) on 26 July 2007. Van Mijenfjorden is probably more stratified and colder than any other fjords on the west coast of Spitsbergen.

3.4 Geology

The Paulabreen glacier system and its moraine are located near the eastern limit of the Central Tertiary Basin on Spitsbergen (Fig. 5). This is a foreland basin filled after a fold and thrust belt was formed in western Spitsbergen in the Paleocene, and it stretches from Isfjorden in the north and southward, east of the orogenic belt. At Paulabreen the strata dip gently towards west-southwest as the eastern limb of the basin syncline. From Paulabreen and eastward, shelf deposits of Cretaceous age outcrop while westward almost exclusively the tertiary basin infill is found. Both the Cretaceous and Tertiary strata consist of shales, siltstones and sandstones and several coal-bearing units are located in the lower Tertiary strata (Salvigsen & Winsnes, 1989).

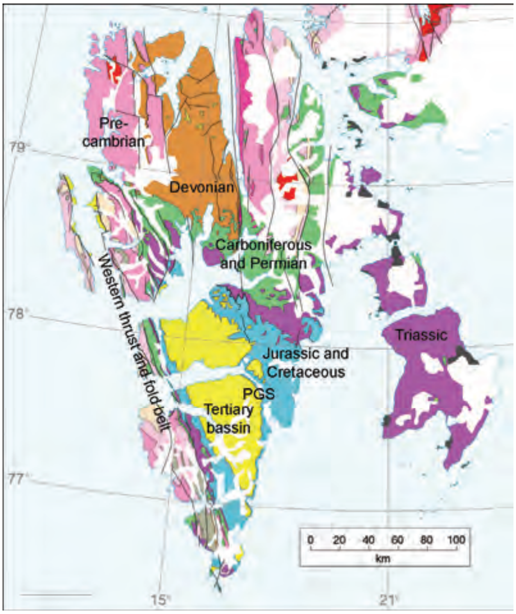

Figure 5 Overview map of the geology of Spitsbergen.

The bedrock weathers and erodes easily and the mountain slopes have large talus aprons covering their lower parts. The geology has been shown to be important for surge behavior of Svalbard's glaciers. Hamilton & Dowdeswell (1996) and Jiskoot *et al*. (2000) show that a sedimentary lithology, in particular shale/mudstone, as found in this area, increases the chance of a glacier to be of surge-type significantly.

3.5 The Paulabreen glacier system

There are many valley glaciers and/or small ice-caps in the area of Van Mijenfjorden, but Paulabreen is together with Fridtjovbreen north of Akseløya, one of the only two glaciers calving into the fjord.

Paulabreen has historically been used as the name for the entire glacier-system that calves into Rindersbukta and thus including Skobreen, Bakaninbreen and all other tributary glaciers. More recently it was used rather for a particular flow-unit, where a distinction from the surging Bakaninbreen was necessary (Murray *et al.*, 1997; Murray *et al.*, 2000 and Fowler *et al.*, 2001). I have adopted both practices; in paper III , IV, V and VI Paulabreen is used for the entire glacier system, while in paper I and II the term is used for a particular flow unit. For the reminder of this overview I use Paulabreen for the flow unit, while Paulabreen Glacier System (PGS) includes all the tributaries.

PGS has an area of 141.8km^2 (measured on an ASTER image from 2005) of which Paulabreen is 64.6km^2 (Hagen *et al.*, 1993). The glaciers reach a maximum altitude of 800m asl. and the equilibrium line altitude (ELA) is between 290 and 380 m a.s.l. (Hagen *et al.*, 1993). In 1898 PGS terminated near the mouth of Rindersbukta and included Scheelebreen and Vallåkrabreen (Kjellström, 1901), which today both terminate on land. From that position and until 2003 the glacier retreated c. 10 km, see front positions on Fig. 3 in paper II. Despite the overall retreat, Scheelebreen, Vallåkrabreen and Mettebreen surged and advanced independently ~ 1919-1925 (De Geer, 1919 and Cöster, 1925). Bakaninbreen, the second largest tributary at 60.8km^2, surged from 1985 to 1995, but the surge terminated before reaching the front, and caused no advance. Skobreen (18.2km^2) surged in 2003-2005 and triggered a surge in the lower parts of Paulabreen as well. These recent surges are described in further details in paper I and II.

3.6 The maximum Holocene moraine of Paulabreen Glacier System

The Holocene maximum moraine of PGS forms a broad belt around Rindersbukta and inner Van Mijenfjorden, which dates to approximately 600 yrs BP. The moraine can be divided into four parts, Crednermorenen, Torellmorenen, Damesmorenen and Geikiemorenen. The moraine consist of two distinct morphological units: (1) a hummocky and partly ice-cored moraine; and (2) a mud apron previously referred to as the Svea marine clay (Rowan *et al.*, 1982 and Gregersen *et al.*, 1983). The mud apron is located on the ice-distal side of the hummocky moraine and can be found on Damesmorenen, Geikiemorenen, Crednermorenen and on the westernmost part of Torellmorenen at Conwentzodden (Fig. 6).

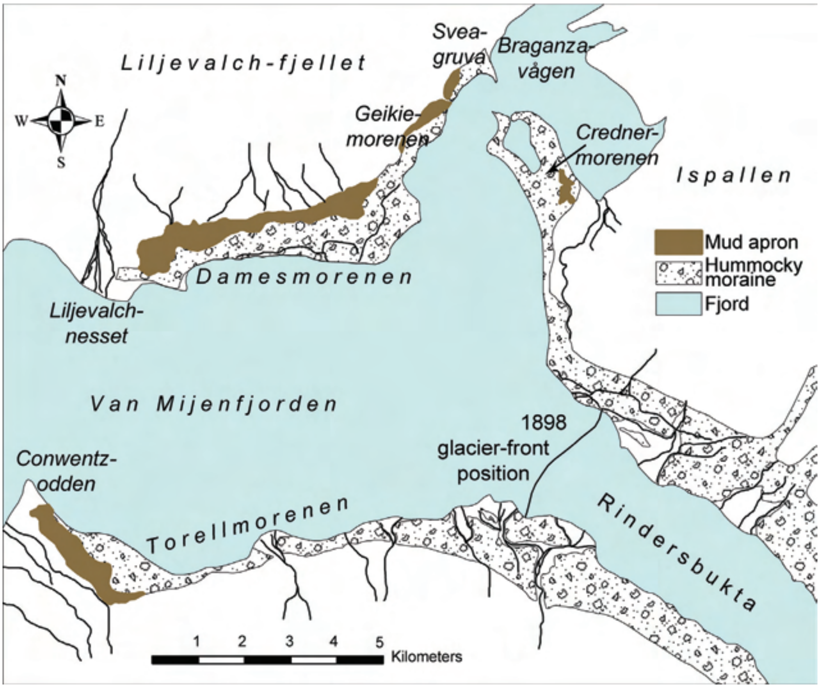

Figure 6 An overview of the distinctly different morphological units of the Holocene maximum moraine of PGS. Also shown is the 1898 glacier front position of PGS.

Much scientific work focusing on its genesis and age has been carried out on the moraine; in particular in the early parts of the 20th century and in the 1980s. A detailed review is given in paper IV including a discussion of the age of the moraine. There is a broad consensus that the maximum moraine of PGS is a result of a glacier surge (De Geer, 1919) occurring ~ 600-700 yrs BP (Punning *et al.*, 1976; Rowan *et al.*, 1982 and Hald *et al.*, 2001), though Punning *et al.* (1976) and Rowan *et al.* (1982) were of the opinion that at least two surges were responsible for the moraine formation.

4 Methods

4.1 Time-lapse photography

Time-lapse photography or cinematography is a motion picture technique that allows slow processes to be viewed at greatly accelerated rate. Classic applications for time-lapse photography include the blossoming of a flower or cloud pattern development (Encyclopedia Britannica Online), but it has also been used to document geological processes, basal sliding of a glacier and volcano dome growth at Mount St. Helens can be seen on http://geology.rockbandit.net. It was used to create a movie of the frontal advance of Paulabreen during its surge in 2005 (paper I). Daily photographs were compiled using Windows Movie Maker with the picture duration set to 0.75 sec. With 86400 seconds a day the motion is thus captured $1.152*10^5$ times the true velocity. Details are given in paper I.

4.2 Use of ASTER images

ASTER images were used in paper I to study geometric changes in Skobreen/Paulabreen during the 2003-2005 surge. ASTER (Advanced Spaceborne Thermal Emission and Reflection Radiometer) is an imaging instrument on the satellite Terra, which was launched in December 1999 (http://asterweb.jpl.nasa.gov). The images were obtained as GeoTIFFs and were the 1B product, where the original image is corrected with radiometric and geometric coefficients. The bands used are in the visible and near infrared (VNIR) as seen in table 1 below. Within these bands the spatial resolution is 15 m. The bands were combined to false-colour images in ArcMap using the tool “Composite Bands”; where the three bands were assigned to colors as can be seen in table 1. Images from late in the ablation season were preferred, as a minimum of snow obscures the surface features of interest. Plenty of images exist of Paulabreen for the period of interest, but most had to be rejected due to cloud cover. Three ASTER images from different stages of the surge (24 July 2003, 23 July 2005, 04 August 2008) were eventually used.

ASTER bandname	Wavelength in ASTER	Color in composite
B1 (VNIR_Band1)	520 – 600 nm	Blue (Real blue wavelength: 450 – 490 nm)
B2 (VNIR_Band2)	630 – 690 nm	Green (Real green wavelength : 490 – 560 nm)
B3 (VNIR_Band3N	780 – 860 nm	Red (Real red wavelength: 630 – 700 nm)

Table 1 How the different bands in ASTER images were assigned to colors in composites.

4.3 Aerial photographs

Aerial photographs were used in paper IV for geomorphological mapping, in paper I for mapping of crevasse pattern and for illustrations of site or as background for the bathymetry data in the other papers. Some of the photographs used as in paper I and IV are from 2006 and were kindly provided by Store Norske Spitsbergen Kulkompani, while all others were obtained from Norsk Polarinstitutt.

4.4 Bathymetry surveying

The recent glacial landscape on the fjord bottom in Rindersbukta was the focus of paper II and the submarine landforms in Van Mijenfjorden were an important part of paper IV. The bathymetry-data originate from two sources: 1) In Rindersbukta NGU and UNIS made a joint effort to survey the fjord using a shallow water swath bathymetry system (GeoSwath) and 2) in Van Mijenfjorden we obtained the data previously surveyed with a multibeam echo-sounder system by Sjøkartverket. The principle behind the two data-collection methods is different as the multibeam system uses a “beam-forming” technique while the GeoSwath uses an interferometric sonar, see eg. Jones (1988). The methods are compared by Gostnell *et al.* (2006). Processing and gridding is explained in the respective papers.

4.5 2D resistivity surveying

Electrical resistivity, measured in Ωm, is the reciprocal of electrical conductivity. As different materials have different electrical resistivities, measurements of the resistivity in the ground can provide information on the ground properties. The distance between the electrodes determines the depth that the measured resistivity is integrated over. The basics of the method is explained by Sharma (1997) who also advises on suitable arrays for different types of surveys.

1 dimensional (1D) resistivity surveying assumes horizontal layering of the subsurface to be investigated. When non-horizontal boundaries are expected, the 2 dimensional (2D) resistivity profiling method is often used – ideally with the profile placed normal to the strike of the structures to be mapped (Sharma, 1997). 2D profiling is often employed as a suitable compromise between very time consuming but more accurate 3D surveying and 1D soundings (Loke & Dahlin, 2002).

2D resistivity profiling was used to image the composition of the moraine of PGS. It was the prime focus of paper VI, and in that paper the application of the method in cold regions is reviewed. It was also used in paper IV. In both cases automatic profiling equipment from ABEM Instrument AB was used in a Wenner array. The software RES2DINV was used to create a geological model from the datapoints that will result in the measured resistivity values (explained further in paper VI).

4.6 Boreholes on Crednermorenen – sediment properties

4 – 8 March 2005, four 9 m deep boreholes were drilled into the frozen sediments of Crednermorenen with a mobile air-pressure driven rig. Core-sampling was attempted but failed. Instead we sampled the blowout sediments during the drilling from three of the holes in 0.5 and 1 m intervals. This method creates two sources of sample contamination: 1) Incomplete blowup and thus mixing of sediments from various depths at the bottom of the borehole and 2) material fall down from the unprotected borehole walls while drilling. Transitions in sediment type were recorded during the drilling (accuracy c. 20 cm), and corresponding sharp boundaries of the sediment properties of the samples suggest that contamination due to mixing was not a big problem. The drill bit crushed the rocks in the sediment, so instead of a grain-size analysis only a qualitative sample description was performed.

In the laboratory we measured salt and water content in each sample. The water content was found using the formula:

Soil water content (%) = 100 x (wet weight – dry weight) / (dry weight – container)

Borehole 2 was drilled in almost pure ice (buried beneath 2.7 m sediment) and strong winds blew the light ice-crystals away during drilling, leaving the heavier sparse debris to be sampled. For these samples the water content could not be measured accurately and thus I only measured the salt content.

The salt content was found by expelling water from the samples with pressurized air and placing a drop of the expelled water in a handheld refractometer. If the sample was dry, sometimes quite some air flowed through the sample before enough water was pressed out, and this process dried the sample further. I believe that for this reason the salt concentration in some samples was overestimated. In other samples the sediment was to dry for water to be pressed out altogether.

The sediment properties are schematically presented in paper VI and details on the lab-measurements can be seen in appendix A.

4.7 Permafrost temperature recordings on Crednermorenen

Permafrost and active layer temperatures were recorded in the four boreholes mentioned above. Each hole was equipped with an eight m long thermistor-strings (from EBA – Consulting Engineers and Scientists) with each 16 thermistors placed in: 0.05, 0.1, 0.25, 0.50, 0.75, 1, 1.25, 1.50, 2, 2.5, 3, 4, 5, 6, 7 and 8m depth. The thermistor-strings were connected to a Lakewood datalogger and temperatures were automatically logged at 2 hours intervals. The manufacturer guarantees that that the thermistor-accuracy is better than 0.2°C, while the precision is probably better than 0.01°C. The inaccuracy may be caused by cable length, the connection between the thermistor and the wire in the cable and battery power of the datalogger (Valeriote, EBA, pers. com.). Some interruptions in the data-series occurred due to logger failure and physical breakage of the cable at two of the boreholes, but those periods were not used in the data presented in papers V and VI.

4.8 Measuring of backwasting rates

Most direct measurements on backwasting rates in Svalbard are measured over short time periods during the summer (Lukas *et al.*, 2005 and Schomacker & Kjaer, 2008). It was the intention that I would investigate the melting processes in the ice-cored Crednermorenen using a combination of direct measurements and photogrammetric techniques. The work with the photogrammetry (both terrestrial and using aerial photos) did not bring convincing results so the work remains unpublished. Two methods were used in the direct measurements: 1) Measuring with steel tape and 2) measurements with differential GPS (DGPS). This data is presented in the result section.

The tape measurements were made in 2006. 21 small steel rods were placed at 10 m distance to three backwasting edges of different exposure. The right angle direction was marked with painted rocks placed towards the edge. Two rods had to be moved during the measuring period as they were near to falling into the backwasting niche. The distances were measured four times throughout the melting season starting from 1 May 2006 and ending 3 September 2006. In 2007 the back-wasting rates were measured using DGPS. Seven niches were surveyed this year. The measurements started on 2 July 2007 and ended on 26 September 2007; the niches were measured four times.

4.9 2D Permafrost modelling using TempW

Permafrost is a thermal phenomenon and ground temperature is controlled by energy fluxes in the air and in the ground. Modelling of ground temperatures can be done in 1D, 2D or 3D depending on the purpose. Both analytical and numerical models have been proposed, though numerical solutions are restricted due to non-liniar heat transfer at phase changes (Mottaghy & Rath, 2006; Boike *et al.*, 2008 and Riseborough *et al.*, 2008).

A 2D finite element model, TempW from GEO-SLOPE International Ltd., was used in paper V to estimate the permafrost distribution in the near shore area of Svea, in a profile from land and into the sea. Limited details are given on the model in the paper due to constraints to the paper length, so here I include a bit more model information as well as choice of inputs. For all details on the program, I refer to the product manual (GEO-SLOPE, 2001). TempW is an ‘of the shelf’ software program designed to model thermal changes in the ground due to environmental changes or manmade construction. It is well suited for permafrost studies as it includes the effect of latent heat at phase changes (freezing and thawing). Heat flux by conduction (q) in a material is a function of the temperature gradient and the thermal conductivity (k) as already stated:

$$q = -k\frac{\partial T}{\partial x}$$

where

T = Temperature

x = distance

The governing equation used in TempW in 2 dimensions (x and y) is:

$$\lambda \frac{\partial T}{\partial t} = \frac{\partial}{\partial x}(k_x \frac{\partial T}{\partial x}) + \frac{\partial}{\partial y}(k_y \frac{\partial T}{\partial y}) + Q$$

where

kx and ky = thermal conductivity in the x and y directions respectively

Q = applied boundary flux

λ = capacity for heat storage

t = time

The equation states that a change in stored energy equals the flux to and from a volume of soil at some point in time.

λ is composed of both the volumetric heat capacity (material property) and the unfrozen water content (temperature dependent). It thus changes dramatically around 0°C. Thermal conductivity in ice is 4 times higher than in water (Oke, 1987). Thus it is also a function of unfrozen water content (and thus temperature). Both unfrozen water content and thermal conductivity are to be defined as functions of temperature in the model. I used the values by Gregersen *et al.* (1983) for unfrozen water measured for sediment from Damesmorenen 4 km from Crednermorenen. This sediment is similar to that in the lower part of borehole 1 (paper VI and appendix A). The thermal conductivity function was estimated by averaging a table value of clay with the appropriate water and ice content for a range of temperatures. Volumetric heat capacity in frozen and unfrozen state was estimated using the same crude assumption. The total water content (in thawed condition) was taken as an average of the lower part of borehole 1 (appendix A). The model allows for different thermal properties and functions to be supplied to different layers, but I assumed a homogeneous subsurface due to a lack of information of the true subsurface composition.

The modelled profile stretches from land and into the seabed. Separate temperature boundary conditions were applied to the submarine and subaerial parts of the profile. First the model was run in a steady state mode to obtain realistic starting values using average annual air and water temperatures and a geothermal heat flux of 35mW/m^2. Then the model was run in a transient mode in the period 1 January 2006 to 1 October 2007. Here a meteorological file with daily values was constructed to force the model as a boundary of land surface. The file contained maximum and minimum temperature, relative humidity, wind speed and precipitation. The meteorological data from Sveagruva was used to create this file, apart from the precipitation, where data from Longyearbyen was used. For the water temperatures, a temperature-function as 14 days average was created on basis of water temperature measurements in Sveasundet (presented in paper V) and this was set as a time-dependent boundary functions for the part of the profile that was below the water surface. The sides of the profile were set as zero flux boundaries. Two scales were used in the model simulations. The model-blocks and boundary settings for the large scale model can be seen in Fig. 7.

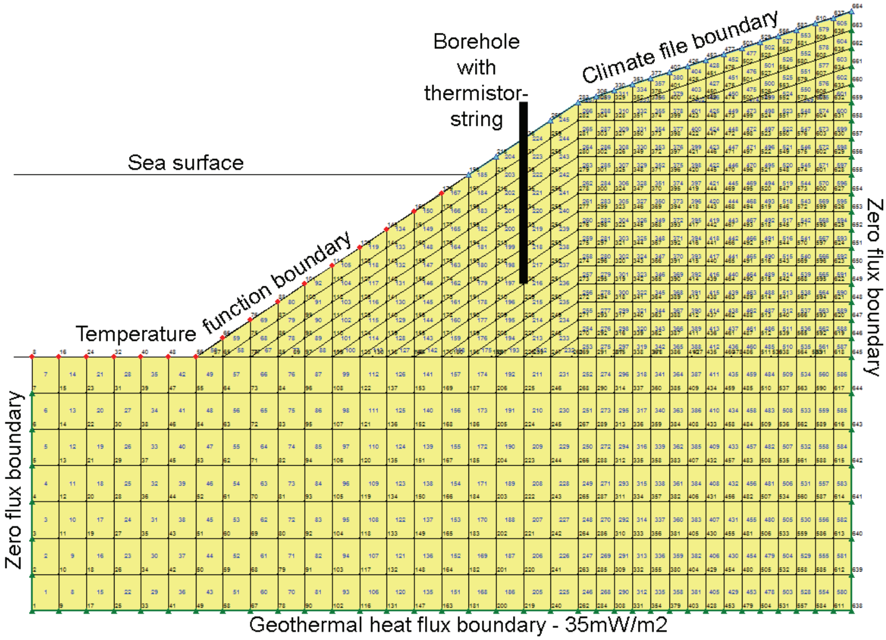

Figure 7 The model blocks of the large scale model and the flux boundaries supplied. Drawn is also borehole 1, which was used as model validation.

The model results would improve with better input parameters, both in terms of subsurface layers and of the thermal properties. Better information on sea-temperature including stratification, as well as snow-cover would also improve the reliability.

4.10 2D section logging

A section drawing was presented in paper IV to illustrate the glaciotectonic deformation structures in Damesmorenen. While photographs today are often used instead, drawings often show detailed information more clearly (Evans & Benn, 2004) while also providing an interpretation of the section.

4.11 Radiocarbon dating

We used radiocarbon dating on molluscs and driftwood in paper IV and the method is carefully described in that paper. The previously published dates (Punning *et al.*, 1976 and Rowan *et al.*, 1982) caused considerable confusion around both the moraine formation and its age, and for that reason we wished to obtain a better age control.

5 Results

5.1 Backwasting rates on Crednermorenen (previously unpublished)

The backwasting rates measured with measuring tape on three niches on Crednermorenen in 2006 can be seen in table 2 below.

Rod nr/ melting interval	12 May-30 Jul (m)	12 May-30 Jul (cm/day)	30 Jul–20 Aug m	30 Jul–20 Aug (cm/day)	20 Aug–3 Sep (m)	20 Aug–3 Sep (cm/day)	Annual melt (m)	Annual melt (cm/day)
1 East	4.40	5.7	2.10	10.0	0.99	7.1	7.49	6.6
2 East	3.77	4.9	2.93	14.0	1.89	13.5	8.59	7.5
3 East	3.73	4.8	2.67	12.7	1.56	11.1	7.96	7.0
4 East	3.43	4.4	2.67	12.7	1.29	9.2	7.39	6.5
5 East	3.45	4.5	1.45	6.9	0.35	2.5	5.25	4.6
6 East	3.06	4.0	1.59	7.5	0.03	0.2	4.68	4.1
7 East	-0.14	0	0.84	4.0	0.25	1.8	0.95	0.8
8 South	7.2	9.4	2.31	11.0	1.50	10.7	11.01	9.7
9 South	5.63	7.3	2.50	11.9	1.03	7.4	9.16	8.0
10 South	5.03	6.5	3.25	15.5	1.15	8.2	9.43	8.3
11 South	0.65	0.8	0.15	0.7	-0.01	-0.1	0.79	0.7
12 South	-0.05	0.0	0.05	0.2	-0.02	-0.1	-0.02	0.0
13 West	0.93	1.2	-0.10	-0.5	0.01	0.1	0.84	0.7
14 West	0.83	1.1	-0.01	0.0	0.00	0.0	0.82	0.7
15 West	1.1	1.4	0.56	2.7	-0.31	-2.2	1.35	1.2
16 West	2.83	3.6	0.28	1.3	0.22	1.6	3.33	2.9
17 West	3.12	4.1	0.32	1.5	0.05	0.4	3.49	3.1
18 West	2.62	3.4	1.21	5.8	0.00	0.0	3.83	3.4
19 West	4.5	5.8	1.4	6.7	0.05	0.4	5.95	5.2
20 West	5.33	6.9	1.93	9.2	0.67	4.8	7.93	7.0
Ave. for interval	**3.07 m**	**4.0 cm/day**	**1.41 m**	**6.7 cm/day**	**0.54 m**	**3.8 cm/day**	**5.01 m**	**4.4 cm/day**

Table 2 Backwall retreat and backwasting rates on three different niches, measured in 2006.

Some negative values are seen, which mainly reflects measurement uncertainty. On the large negative value, 3 September, rod 15, a wide crevasse or crack was observed in the ground in front of the backwasting edge. This crack probably explains the increased distance to the edge. The largest edge retreat is seen in the first period 12 May to 30 July, but this is also the longest interval. The maximum edge retreat rate in this period was 9.4cm/day and the average is 4cm/day. In the period 30 July to 20 August the melt rate was faster with a maximum of 15.5cm/day and an average of 6.7 cm/day. The last period, 20 August to 3 September, had a maximum backwasting rate of 9.7cm/day and an average rate of 4.4 cm/day. For the entire summer the maximum retreat of the slope edge was 11.01 m (9.7 cm/day) and the average was 5.01 m (4.4cm/day).

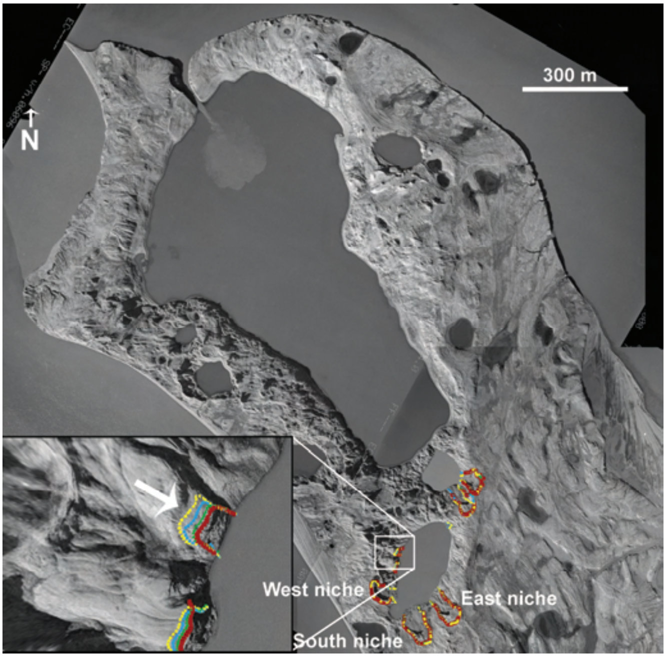

Figure 8 An overview on the surveyed niches in 2007 with an enlargement of the most active niche. Red dots: 02 July 2007, green dots: 16 July 2007, blue dots: 7 August 2007, yellow dots: 26 September 2007. The labeled niches were surveyed using a tape measure in 2006.

Fig. 8 shows an overview of the DGPS measurements in 2007 and a closeup shows the retreat of the most active niche. Between 02 July and 16 July the retreat was max 1.2 m (8.6cm/day), between 16 July and 7 August the retreat was max 2.5 m (11cm/day) and between 7 August and 26 September the retreat was 3.5m (6.9cm/day). However, in the three very active niches in 2006, the total backwasting was less than 2 m retreat in 2007. The measurements started late in 2007, which may explain some of this variation; however it appears that there is large interannual variability of the individual niches.

With an average temperature of -1.7°C, 2006 is the warmest year on record between 1912 and 2008. 2006 has by far the warmest winter and spring temperatures, while the summer temperatures are closer to average, though the second warmest on record. The warm winter temperatures in 2006 may have initiated an early melt of the active layer and set the setting for exceptional backwasting rates and debris flow activity this summer. The impression from several visits to the site is that there was much more debris flow activity in 2006 compared to 2007.

5.2 Summary of papers

I: Lene Kristensen, Douglas I. Benn (To be submitted). A surge of Skobreen/Paulabreen, Svalbard documented by a time-lapse movie, aerial and satellite images and photographs. To be submitted to *Geosphere*

In 2003-2005 Skobreen and the lower parts of Paulabreen experienced a major surge. This paper presents a time-lapse movie showing the advance of the front of Paulabreen. Several satellite images and aerial photographs from different stages during the surge, as well as digital photographs from land and the air document the surge in detail. The movie shows that the glacier advanced as a coherent block (plug-like flow). This shows that the motion is almost entirely dominated of basal sliding or basal till deformation rather than ice creep. The observation supports the idea that surges are facilitated by trapped, pressurized water that reduces the basal drag, while clearly indicating thawed conditions at the glacier sole. Pro-glacial thrusting at the margin was observed and indicates a transfer of compressive stresses into the foreland. The movie shows that the surge terminated during the polar night in the winter 2005 – 2006. An ASTER image shows that by 24 July 2003 most of Skobreen was surging and the surge bulge was about 700m upstream from Skobreen's confluence to Paulabreen. From land it was observed that on 2 April 2005 the SW side of the front of Paulabreen was heavily crevassed and advancing while the NE side was undisturbed; in August 2005 the entire calving front was advancing. Between 24 July 2003 and August 2008 the maximum measured downstream ice displacement was 2600 m, while the frontal advance was 1800 m. This significantly underestimates the front's true advance, because calving occurred between the surge termination and August 2008. Assuming a surge onset 24 July 2003 and a termination in December 2005, the average ice velocity of the main trunk of Paulabreen was 3.2 m/day. The surge resulted in a down-draw of the Skobreen surface of c. 50 m as evidenced by glacier ice stranded on the mountain slopes. The presence of longitudinal crevasses signifying compression points to an initiation point of the surge almost halfway between the backwall and the confluence to Paulabreen c. 4 km from the confluence. Wide irregular crevasses indicate shearing in the margin between the surging ice of Paulabreen and the non-surging ice upstream. Shearing and rotation was also seen in the ice that has 'turned the corner' from Skobreen and into Paulabreen. The central area of Paulabreen was largely uncrevassed throughout the surge, suggesting that no surge-front had passed downstream. We tentatively interpret this as signifying a lack of permafrost below the lower parts of Paulabreen. In the front and at the margins brecciated ice and chaotic crevasses and debris bands indicate intense compression and englacial thrusting, probably because the marginal zone was cold-based. In the ice-cored lateral foreland several crevasses developed during the surge. These indicate compressive and tensile stresses transferred into the foreland by the surging glacier. At the side of Bakaninbreen we observed intensely folded debris rich glacier ice lying on top of much less deformed glacier ice. A bubbly ice-layer interpreted as compressed snow separated the two units. This was interpreted as the surging ice having advanced over pre-existing almost stagnant ice without incorporating or activating it. If most of Paulabreen was warm-based, as argued, then this may have occurred at the transition between warm and cold basal conditions at the margin.

II. Douglas I. Benn, Lene Kristensen and Jason D. Gulley. *In Press*. Surge propagation constrained by a persistent subglacial conduit, Bakaninbreen/ Paulabreen, Svalbard. In press, *Annals of Glaciology*.

This paper deals in more detail with the mechanisms and patterns of surge propagation. We argue surge propagation is constrained by a persistent subglacial conduit that runs below the medial moraine between Bakaninbreen and Paulabreen. Bakaninbreen surged in 1985-1995 and Skobreen/Paulabreen in 2003-2005. The surge propagation rate for Skobreen/Paulabreen was 7.8 m/day which is much faster than the ice displacement rates that was reported in paper I, and twice as fast as the surge propagation of the Bakaninbreen surge. In both instances the surges did not transgress the medial moraine between Bakaninbreen and Paulabreen. This was most obvious for the Bakaninbreen surge, which propagated 7 km downstream alongside Paulabreen without transgressing the medial moraine, while during the surge of Skobreen the medial moraine was displaced somewhat towards north in the lower parts. On the bathymetry dataset of Rindersbukta we see an esker running along the fjord. Glacier front positions from old maps and aerial photographs reveal deep embayments at the location of the esker, and Cöster (1925) observed upwelling at the location of an embayment. The embayments and the submarine esker were in every instance associated with the medial moraine on the glacier. A moulin is found at the confluence of Bakaninbreen and Paulabreen fed by supraglacial lateral meltwater channels from both Bakaninbreen and Paulabreen. The moulin dropped and led to a Nye-channel that extended downglacier so it appears that the subglacial conduit follows the medial moraine at the bed towards the front and that this pattern has persisted for at least 100 yrs. The presence of pressurized water at the sole is believed to be important for the surge propagation and maintenance of fast flow. We argue that the low-pressure subglacial conduit acted as a “firewall” between the two glaciers, preventing the pressurized water and surge front to propagate into the other glacier by draining the pressurized water. This mechanism may explain why sometimes a single flow unit may surge while not affecting neighboring flow units.

III. Dag Ottesen, Julian A. Dowdeswell, Douglas I. Benn, Lene Kristensen, Hanne H. Christiansen, Ole Christensen, Louise Hansen, Erland Lebesbye, Matthias Forwick, Tore O. Vorren. (2008). Submarine landforms characteristic of glacier surges in two Spitsbergen fjords. *Quaternary Science Reviews* 27 (15-16). 1583-1599.

In this paper we describe, compare and interpret the submarine landforms of two similar fjords, Rindersbukta and Van Keulenfjorden using bathymetric data. Both fjords were recently affected by glacier surges. Historical maps and aerial photographs provide details of the glacial history and glacier front positions within the surveyed area aiding the interpretation of the landforms. PGS surged, filling most of Rindersbukta shortly before 1898 while Nathorstbreen surged into Van Keulenfjorden in the 1870s or 1880s. In both fjords streamlined lineations were found parallel to the fjord axis interpreted as megascale glacial lineations. In Rindersbukta three broad sedimentary transverse ridges were found of which the most proximal of these is located close to the 1898 position of PGS. In Van Keulenfjorden a similar single broad transverse ridge was found, which was so high that only a narrow corridor of this ridge was mapped due to shallow water. These large transverse ridges were interpreted at terminal moraines. Distal to the three terminal moraines in Rindersbukta and the single terminal moraine in Van Keulenfjorden are semicircular lobes dipping evenly away from the ridges. These were interpreted as glacigenic debris flows either originating from failure of the frontal slopes of the terminal moraines or as being extruded from beneath the glacier at the surge termination. Sinuous ridges running in general parallel to the fjord axis were found in both fjords. There were interpreted as eskers formed by infill in subglacial meltwater channels. Numerous small subparallel ridges were found in both fjords forming in some places a quite complex crosscutting pattern. We divided the ridges in a concordant set aligned parallel to the former known front position and a discordant set aligned oblique to the former margin positions. The concordant we interpret as annual push moraines whereas the discordant were interpreted as crevasse-fill ridges. From this we present a landform assemblage model for Svalbard tidewater glaciers, which differs from its terrestrial counterpart by including annual push moraines and being better preserved.

IV. Lene Kristensen, Douglas I Benn, Anne Hormes, Dag Ottesen. (*Accepted*). Mud aprons in front of Svalbard surge moraines: evidence of subglacial deforming layers or proglacial glaciotectonics? Accepted for *Geomorphology*

In this paper we compare the submarine and terrestrial landforms of a 600 yrs BP surge of PGS. For the geomorphological mapping we use bathymetry data from Van Mijenfjorden collected by Sjøkartværket and aerial photographs and a digital elevation model on land. DC Resistivity profiling and section studies were used to investigate the internal structure of the landforms. The focus is two landforms that were found in both environments; a hummocky moraine and a distal proglacial mud apron. The hummocky moraine on land contained small amounts of buried glacier ice, but less than originally expected, showing that deicing is almost complete. We suggest that thrusting and in particular the squeezing of sediment into basal crevasses is responsible for the hummocky topography. This is evidenced on several places where ridges consist of almost pure marine muds while the sediment in most of the landform is a mix of terrestrial and marine origin. The hummocky topography on the seabed was formed by a similar mechanism and does probably not contain an ice-core, which we had considered. The submarine mud apron in Van Mijenfjorden resembles the debris flows of Rindersbukta (paper III) as well as in front of moraines of other tidewater glacier surges in Svalbard. Flow structures covering most of its surface support a debris flow interpretation. The landbased mud apron consists of marine muds and was previously interpreted as being a competent slab of the seabed that was thrust in front of the glacier – possibly facilitated by permafrost in the seabed. Comparing the land-based and the submarine mud apron we realized they are very similar, both dipping evenly away from the former glacier front position and being wedge-shaped (as evidenced in geophysical profiles). The landbased mud apron is covered by a dendritic gully network signifying intense fluvial erosion not seen in the submarine counterpart and also some weakly defined transverse ridges are only found on the terrestrial mud apron. The transverse ridges we interpreted as compressional ridges formed by ice push. The slope angle of the terrestrial mud apron is greater than the submarine mud apron. We suggest that the terrestrial mud apron was deposited similar to the submarine; as a debris flow. The low amplitude compressional ridges and the steeper angle of the terrestrial counterpart indicate a higher viscosity, as water could drain during deposition. The hypotheses of the origin of the mud aprons being a deformable bed extruded from beneath the glacier, was excluded on basis of the resistivity profile. The terrestrial debris flow consisted of a saline low resistivity sediment body that only reached the ice-contact slope. Proximal to this more mixed sediments were found. We suggest instead that the marine sediment was bulldozed in front of the glacier in a continuously failing soft mass and let to rest, as the surge terminated. These mud-aprons can be seen as an endmembers of proglacial landforms that have thrust-block moraines as the opposite endmember.

Two clusters of dates have previously led to the interpretation that Damesmorenen was deposited by two separate surge events, at 8-9000 yrs BP and at c. 600 yrs BP. Here we suggest that the 8-9000 cluster was caused by a local extinction of the extremely abundant *Chlamys islandica* at that time, possibly in relation to the 8.2 ka cooling event; which is supported by new evidence from a sediment core from Van Mijenfjorden (Hald & Korsun, 2008). Sampling other species we obtain dates of various Holocene ages as would be expected of a sediment that has been eroded of the seabed by a surging glacier.

V. Lene Kristensen, Hanne H. Christiansen and Fabrice Caline. (2008). Temperatures in coastal permafrost in the Svea area, Svalbard. *Ninth International Conference on Permafrost*, University of Alaska Fairbanks, 1005-1010.

Early in the project I considered whether subsea permafrost could be responsible for some of the landforms seen on the seabed in Van Mijenfjorden and at Damesmorenen. For this reason we wished to investigate the possible permafrost distribution in the fjord and in the coastal zone. In this paper we present borehole temperatures that suggest that permafrost temperatures in the low-lying peninsula Crednermorenen are strongly influenced by the proximity to the fjord. Furthermore we present water temperature measurements showing that the average annual water temperature is below 0°C, suggesting that permafrost would be present in the fjord. Previous studies and modelling attempts of coastal and subsea permafrost have assumed mean annual water temperatures on 0°C and +1°C respectively. Using the new data we ran the 2D finite element program (TempW) for a profile from land and into the fjord. The program was first run on a steady state mode, using annual average temperatures for air and water and a realistic geothermal heat flux as forcing. It shows that thin permafrost is probably present in the fjord and that the warming effect of the fjord is felt on the ground temperatures about 100 m inland and vise versa. Then the model was run in a transient mode using actual air- and water temperatures as forcing. The model results were compared to the measured borehole temperatures in the nearshore area, and were in reasonable correspondence to the measured temperatures. The modelling indicated subsea permafrost, and it probably underestimated its thickness because the water temperatures used in the model was measured near the surface of the fjord, which is strongly stratified during the summer. Whether the permafrozen seabed is actually *frozen* or *thawed* will depend on primarily its salinity. A marine sediment would be unfrozen at temperatures down to at least -1.9°C while a less saline sediment may be frozen.

VI. Lene Kristensen, Håvard Juliussen, Hanne H. Christiansen and Ole Humlum. (2009). Structure and composition of a tidewater glacier push moraine, Svalbard, revealed by DC resistivity profiling. *Boreas 38 (1)*. 176-186.

Here we studied a lateral part of the 600 yrs BP moraine of PGS, Crednermorenen. Six resistivity profiles were used to image subsurface structure and composition of the moraine. Three of the profiles passed a borehole, where the sediment composition was known, and that provided confidence to the geological interpretation. The western proximal side of the moraine was hummocky and buried glacier ice was exposed in niches on several places, covered by a debris layer of 1-2 m. The profiles in the hummocky areas showed a high resistivity layer, interpreted as glacier ice, covered by a thin lower resistivity layer interpreted at the active layer. In some of the profiles the glacier ice was thicker than the maximum penetration depth (30 m) so while the maximum thickness of the buried glacier ice is uncertain; we know it is at least 30 m on several places. The distal mud apron described for Damesmorenen (Paper IV) is also found on Crednermorenen; on many places covered by beach sediments (as in borehole 4) because of a glacier dammed lake when PGS was at its maximum position. Three profiles on the mud apron showed very low resistivities here. Borehole 4 showed that the sediment is a saline mud; the salinity being similar to sea-water. We interpreted the deposit as proglacially pushed or subglacially extruded marine muds, though later (paper IV) we came to favor the first of these hypotheses. Areas of intermediate resistivity we interpreted as frozen glacial till. In one profile zones of high resistivity was penetrated from below by a wedge-shaped feature of low resistivity. This we interpreted as the infill of marine muds in a basal crevasse. Thrusting may transport basal sediments into the glacier, but the size of this feature (35 m wide) is unrealistic for a thrusted sediment band, while it is comparable with crevasse fill features observed in Borebukta (Ottesen & Dowdeswell, 2006). Furthermore thrusts normally dip upglacier, while this feature was symmetrical around a vertical axis. Compared to other geophysical techniques, the DC resistivity profiling provides easy to interpret information of the subsurface composition, while smaller structures such as thrust faults may be better imaged using ground penetrating radar.

6 Discussion and perspectives

In this thesis I have studied both terrestrial and marine parts of the PGS landsystem by a combination of glaciology, glacial geomorphology and permafrost approaches. Direct observations of the actively surging Skobreen/Paulabreen were applied to evaluate and interpret the Holocene maximum moraines of PGS. The surge mechanisms observed during recent surges of Skobreen/Paulabreen and Bakaninbreen are most likely similar to those responsible for the formation of the 600 yrs BP moraines of PGS. Combining a study of both glaciology and the terrestrial and submarine landforms has improved our understanding of the processes, landforms and glacier and permafrost interactions in those environments. In this discussion I have tried not to repeat the discussion already found in each of the papers, but rather link the findings in the papers

The measurements of the 2003-2005 surge of Skobreen/Paulabreen presented in Papers I & II are of three distinct, but related, aspects of the glacier motion. First, the maximum total *ice displacement* was about was about 2700 m giving an average velocity of 3.2 m/day. Second, the *surge front* movement was faster than the average ice velocity, with a rate of 7.8 m/day, which is about twice the velocity of the surge front on Bakaninbreen (Murray *et al.*, 1998). Before the surge front reaches the terminus, the ice and surge-front velocities are usually not identical, as the surge front propagates by thermal evolution of the bed which also can take place upglacier or sideways. Third, the glacier-front velocity reflects a combination of ice velocity and ice-front meting and calving. When the surge-front reaches a terrestrial glacier terminus, the ice-, glacier-front, and surge-front velocities are closely similar, because ice melt rates are small compared with velocities during a surge. At tidewater margins, however, the glacier-front velocity may be much less than the ice velocity, due to rapid calving losses from deeply fractured ice.

A crude estimate of a glacier's dynamics can be obtained by studying crevasses and other features on ASTER images, as attempted here. It would be interesting to study the dynamics in much greater temporal and spatial detail, as was accomplished during the famous 1982 - 1983 surge of Variegated Glacier (Kamb *et al.*, 1985).

During the 2003-2005 surge of Skobreen/Paulabreen, brecciated ice, thrusting and crevassing of the foreland at the margins indicated compression and shearing. Sedimentological evidence of tectonic deformation shows that similar compressional stresses were transferred into the foreland during the 600 yrs BP surge (paper IV). In 2008 we found crevasse-fill ridges melting out at the margin of the 2003-2005 surge. This is in line with the evidence for crevasse fill ridges in the 600 yrs BP moraine of PGS, presented in papers IV and VI.

At the side of Bakaninbreen, surging glacier ice had overridden much less deformed stagnant ice. This shows that, at some point, the surging ice was thrust along a shear-plane up onto the stagnant non-surging ice. This probably occurred near the margin where the ice is thinner and thus is able to exert a lower pressure on the bed. Englacial thrusting is often seen in polythermal glaciers at the transition between warm basal ice and a cold-based margin (Hambrey & Huddart, 1995; Alley *et al.*, 1997 and Murray *et al.*, 1997) and the thrust fault may have occurred at such a basal thermal boundary. To support the inferred thermal properties it would be very useful to combine such observations with direct observations by drilling boreholes to the bed, and using geophysical methods (geo radar) to provide evidence for the thermal conditions in the ice and at the glacier sole. In combination with a more detailed study of the dynamics and the basal water pressure it may be possible to find if there is a different surge mechanism for polythermal and temperate glaciers, for which there is not a concensus (Kamb, 1987; Murray *et al.*, 2003b and Frappe & Clarke, 2007).

The only glacier in PGS which is classified as a surge-type glacier in the Glacier Atlas of Svalbard and Jan Mayen (Hagen *et al.*, 1993) is Bakaninbreen. Jiskoot *et al.* (2000) also categorized Scheelebreen as surge-type glacier. Skobreen, Vallåkrabreen, or Mettebreen were not previously classified as surge-type glacier though there is sound historical evidence that the latter two and Sheelebreen surged ~ 1920 (De Geer, 1919 and Cöster, 1925). These surges (including the surge of Bakaninbreen) appear to be limited to the individual flow units. In paper II we propose that the reason for this may be low pressure drainage conduits, located below the medial moraines. Such conduits may act as 'firewalls' preventing injection of high pressure water at the sole which melt the permafrost. It is possible that this mechanism has worked during the early 20th century surges of different flow units in the PGS. For supporting the conduit/firewall hypothesis it would be useful to study the discharge and turbidity of meltwater release of such a conduit which borders a surging flow-unit throughout a surge.

With observations of five different flow units of the PGS surging during the last 100 years, I find it likely that all of the tributary glaciers in PGS are of surge-type, though this is not proven. Recent observations of patterns of glacier elevation change in Svalbard

(Nuth *et al.*, 2007 and Sund *et al.*, 2009) suggest that the majority of glaciers in the archipelago are of surge-type, contrary to the results of the statistical analysis (Jiskoot *et al.*, 1998 and Hamilton & Dowdeswell, 1996) where only observed surges were included. None of the above mentioned glaciers have been observed to surge more than once since 1898, which suggest that they have a quiescent period of more than 100 yrs. Only four glaciers in Svalbard have been observed to surge twice, with intervals of 54, 70, 107, and 130 yrs respectively. Tunabreen has been observed to surge three times with periods of 30 to 40 yrs. These are probably among the glaciers in Svalbard that have the shortest surge intervals (while perhaps also being located where they were easily observed in earlier times). Surge intervals in western North America are typically 20-30 yrs (Meier & Post, 1969) but Dowdeswell *et al.* (1991) found that the quiescent period in Svalbard was relatively long (50-500 yrs). The observations from PGS support that the quiescent period of the glaciers in Svalbard is long.

The ca. 600 yrs BP surge of PGS was one of the greatest surges recorded in Svalbard with moraines 25 km beyond the 2003 position. Even more striking, between this moraine and the moraine at the head of Rindersbukta there are no evidence of any earlier or later surges, such as the overridden end moraines of Borebukta (Ottesen & Dowdeswell, 2006). The 600 yrs BP surge apparently reached 12 km further into the fjord than any earlier or later surges. Tentatively I suggest that all the major glaciers or flow units in PGS coincidently surged at the same time ~ 600 yrs BP, and their combined effort caused this spectacular advance. De Geer (1919) may have implied something of this nature with his statement: 'Paulabreen's oscillation is much larger than seen at all other known Spitsbergen glaciers so that without a doubt a unique coincidence of circumstances must have occurred to facilitate such an exceptional event' though he wrote this before glacier surges were an established scientific phenomenon. The hypothesis of a coincidental simultaneous surge of several flow units may also explain why the age of the maximum Holocene moraine of PGS is slightly out of sync most other Svalbard glaciers, which were at their maximum Holocene position ~ 1920 (Dowdeswell *et al.*, 1995 and Mangerud & Landvik, 2007).

One cause of why flow units may either surge separately or combined may be found in previously reported pattern of surge propagation. As stated in section 2.1.3.2, surges of tidewater glaciers often initiate at the glacier front and propagate upstream, due to less restriction to flow at the front of these glaciers; while surges in glaciers terminating on land initiate higher and propagate downstream (Dowdeswell & Benham, 2003 and Murray *et* *al.*, 2003b). Hagen *et al.* (1993) found that surges of tidewater glaciers usually affects the entire glacier system, while often only a single flow unit is affected for glaciers terminating on land. In combination, these observations may imply that surges that propagate upstream tend to draw in the tributaries while surges propagating downglacier leave the tributaries unaffected. This fits well with the observations from Skobreen/Paulabreen. Above the initiation point (identified in paper I) ice from all niches were drawn into the surge, while below the initiation point only the particular flow unit was included. It may be that the 600 yrs BP surge of PGS was initiated at the front, and for this reason all flow units were drawn in. This may set a boundary to the effectiveness of a subglacial conduit as a surge propagation firewall; possibly the mechanism works only for a downstream or sideways propagation of a surge front. While being a plausible cause for the major 600 yrs BP advance and the single flow unit surges the last 100 years, it is not adequately proven. I anticipate more studies of patterns of surge propagation patterns.

Paper III provides a detailed description of the landforms deposited by glacier surges in Rindersbukta and Van Keulenfjorden and relates them to the historical record of front position of the last ~100 yrs. This allowed us for the first time to validate the landform interpretation with information of the front position and glacier behaviour. The landforms in the two fjords differ from those in front of Borebreen and Wahlenbergbreen by including prominent submarine esker systems and less extensive crevasse fill ridges. As suggested in paper II the esker systems probably are a result of subglacial conduits forming at the confluence of large tributaries, namely Bakaninbreen and Paulabreen in Rindersbuta and Liestølbreen and Nathorstbreen in Van Keulenfjorden. Borebreen and Wahlenbergbreen have both only minor tributaries. The historical evidence of a medial moraine on PGS and a submarine esker in Rindersbukta shows that the 1898 front position was comprised of two flow units; most likely Bakaninbreen and Paulabreen.

We state in paper III that hummocky moraine is not seen in the marine record and suggest that when hummocky moraine is found in the terrestrial record they may be subareally degraded crevasse fill ridges. In paper IV we show that a submarine hummocky moraine may indeed be a part of the submarine surge landform assembly. We discuss the hummocky topography both on land and on the seabed later in the discussion.

In paper III as well as in earlier works on both submarine and terrestrial surging glacier landsystems, the existence of streamlined landforms as indicators of surges or fast flow have been emphasized (Evans & Rea, 1999; Christoffersen *et al.*, 2005 and Evans *et al.*, 2008). While we find streamlined landforms in Rindersbukta and in Van Keulenfjorden they seem to be absent in the bathymetry data in Van Mijenfjorden from the 600 yrs BP surge of PGS (paper IV). The surveyed strip does not cover the entire width of the fjord, so it can not be ruled out that streamlined landforms are found beyond this strip. Nevertheless it may be fair to say that an absence of streamlined landforms does not rule out the possibility of a glacier surge having occurred.

Several workers have found an age of the major surge of Paulabreen to 600-700 yrs BP (Punning *et al.*, 1976; Rowan *et al.*, 1982 and Hald *et al.*, 2001) from dating of driftwood and an IRD peak in a marine sediment core. Punning *et al.* (1976) and Rowan *et al.* (1982) however both proposed that Damesmorenen was deposited by (at least) two separate surges because of an age cluster (between 8555-7850 yrs BP) of seven marine molluscs. Careful geomorphological investigations, however, provided no evidence for the 'two surge' hypothesis. Furthermore, both the 'young' driftwood and the 'old' molluscs were both collected in various locations in both the hummocky moraine and in the pro-glacial mud-apron and Hald *et al.* (2008) report an IRD minimum on a core ~ 8.2 ka yrs BP. The two age-clusters were not properly accounted for by invoking two separate surges and we have long been mystified by the implications of these dates. Anne Hormes (Associate Professor, Quaternary geology, UNIS) dated molluscs from the Kapp Amsterdam section of Damesmorenen and from the moraine surface. She found various Holocene ages for the mollusc shells, apart from Chlamys islandica, where all specimens belonged to the cluster of 'old' ages. We find now the most likely explanation for the cluster is a local extinction of the species at that time. We speculate that the cause of the extinction might be the 8.2 ka event, as Hald *et al.* (2008) present evidence of this event from a sediment core from Van Mijenfjorden.

The presence or absence of subsea permafrost was found to be less important for the geomorphology than originally assumed. First we found that the hummocky terrain on Damesmorenen contained only small amounts of buried ice (paper IV) showing that deicing was largely complete. Instead there is evidence that infill of basal crevasses is important for the hummocky topography, and this may account for the hummocky topography on the seabed as well. For some reason, neither on land nor at the seabed do they form as regular a rhombohedral network as that found in the fjord Borebukta, Svalbard (Ottesen & Dowdeswell, 2006). In paper VI we present geophysical evidence of a crevasse fill ridges on Crednermorenen, still surrounded by buried glacier ice.

Secondly we abandoned the assumption that the distal mud-apron in front of the moraine was deposited as a solid slab, which Rowan *et al.* (1982) suggested might have been facilitated by permafrost in the seabed. This realization came from comparing this landform to the large submarine debris flow on the distal side of the Holocene maximum moraine of PGS and in other Svalbard fjords. The close similarity between the terrestrial mud apron and the submarine debris flow caused us to realize, that the landform more likely was deposited as slurry in advance of the surging glacier (paper IV). The origin of the submarine slurry has been suggested to be a subglacial deforming layer (Hald *et al.*, 2001) but a geophysical profile through the proglacial terrestrial mud apron showed that the composition is markedly different from the subglacial sediments of Damesmorenen. The high salinity and homogeneous silty-clayey composition points to an almost pure marine origin, and we suggest that the debris flow originated as marine sediments pushed or bulldozed in front of the advancing glacier.

In the lateral Crednermorenen > 30 m of buried ice was found. This contrasts to the Damesmorenen where deicing is almost complete. We assume that the two were deposited simultaneously by PGS ~ 600 yrs BP. The reason for the difference may be due to a greater sediment supply to the lateral areas of the glacier by avalanches and rock fall. This supply does not reach the central regions and so the frontal zone is cleaner than the lateral zone. The lateral sediment cover insulated and preserved the glacier ice better on Crednermorenen than on Damesmorenen. This hypothesis is supported by comparing the sediment on Crednermorenen and Damesmorenen. Crednermorenen is in its hummocky and ice-cored regions dominated by angular terrestrial sediments, whereas the sediment on Damesmorenen has a greater marine component.

The buried ice in Crednermorenen points to a much longer preservation potential of the ice-cored moraines in Svalbard than the c. 200 yrs estimated by Etzelmuller (2000). However, the longevity of ice-cored moraines in Svalbard appear significantly smaller than in northeastern Greenland (75°N), where an ice-cored moraine of Pleistocene age has been found (Houmark-Nielsen *et al.*, 1994). Measured back-wasting rates of Crednermorenen were higher than measurements from other places in Svalbard and on Iceland in the record warm year of 2006. However the back-wasting rates were smaller in 2007 and large interannual variations of the debris flow activity in front of the individual niches was observed. While the back-wasting rates are perhaps quite normal on Crednermorenen; the area affected by back-wasting is limited, as the topography has become quite subdued over time. Thus the ice-core here is not likely to disappear any time soon.

Permafrost and glacier interactions are important for several aspects of this thesis. Of particular interest is the question of whether the thermal evolution of polythermal glaciers in Svalbard is fundamental to their surge behaviors (Murray *et al.*, 2000). Our observations of the surge of Skobreen/Paulabreen support this notion, though possibly the central part of Paulabreen was warm-based before the surge. Permafrost is an important factor for proglacial tectonics. We see this evidenced in the tectonic landforms of the 600 yrs BP surge of PGS. These vary between push of a soft marine slurry, which acts as a debris flow, and the thrusted and folded beach gravels that is found at the Kapp Amsterdam section. Lastly, the preservation of ice in ice-cored moraines is influenced by the presence of permafrost. Bottom-wasting will not occur in a permafrost environment and top wasting becomes inhibited by a debris cover that is as thick as an active layer, leaving only backwasting as a process of melting the buried ice.

7 Conclusions

Skobreen/Paulabreen surged in 2003-2005 and a range of processes were observed or inferred from observations of this surge relating both to dynamics and stress pattern. This surge and the 1985-1995 surge of Bakaninbreen were confined to either side of the Paulabreen/Bakaninbreen medial moraine, which is associated with a subglacial drainage channel. Drainage through the conduit prevented could have constrained thermal evolution of the glacier bed by injection of pressurized water beyond the conduit.

Skobreen was not previously classified as a surge-type glacier and neither were Vallåkrabreen or Mettebreen which surged early in the 20th century. These observations suggest that quite likely all the glaciers in the catchment are of surge-type, and that the surge interval is more than 100 yrs. Due to the long quiescent phase in Svalbard more and more new glaciers appear to fall in the “surge-type” category, and points to a greater surge probability than suggested by earlier statistical analysis.

A major surge ~ 600 yrs BP left deposits on land and on the seabed and the landsystem elements are largely similar in both environments. This moraine is 25 km further into the fjord than the glacier today and 12 km ahead of the second largest moraine. This points to an extraordinary event and I suggest that all major glaciers in the PGS coincidently surged simultaneously then, while normally surging only as a single flow unit, as a result of the proposed firewall mechanism. The lateral part of the moraine (Crednermorenen) contains > 30 m buried glacier ice while at the frontal part (Damesmorenen) de-icing is almost complete. This reflects a much greater supraglacial debris content at the lateral margin, which has prevented the ice from melting.

Permafrost as thermally defined is probably present in Van Mijenfjorden as it has subzero temperatures in the basins even during summertime due to its strong stratification. Whether the submarine sediments are *frozen* will depend primarily on their salinity, particle capillarity and absorption as those factors reduce the freezing point in sediments. Results from a 2D temperature simulation suggest that heating from the fjord is felt c. 100 m inland and cooling from land is felt c. 100 m into the fjord.

Despite the probable submarine permafrost in the fjord we have abandoned the earlier published assumption that the terrestrial mud apron of Damesmorenen was a slab of seabed thust onto land by the surging glacier. Instead we suggest that it was pushed ahead of the glacier as a slurry, which, however, was slightly folded by push of the glacier. We have also rejected the hypothesis that the hummocky topography in Van Mijenfjorden was due to buried glacier ice in the fjord and suggest that it is rather a result of thrusting and infill of sediments in basal crevasses.

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Appendix

Sediment description from the boreholes at Crednermorenen

Hole Nr	Depth (m)	Description	water content (%)	salt content (%)
Hole 1	0 - 1	Diamicton	8.84	0.1
Hole 1	1 - 2	Diamicton	7.40	0.2
Hole 1	2 - 3	Diamicton	10.79	0.4
Hole 1	3 - 4	Diamicton	9.79	0.8
Hole 1	4 - 5	Diamicton	20.83	2.2
Hole 1	*5.4 m*	*Sharp transition*	-	-
Hole 1	5 - 6	Diamicton/silty clay	23.87	4.6
Hole 1	6 - 7	Silty clay - shell found	27.67	4.4
Hole 1	7 - 8	Silty clay	21.20	5.2
Hole 1	8 - 9	Silty clay	20.99	6
Hole 2	0 - 1	Diamicton	7.99	Too dry
Hole 2	1 - 2.	Diamicton	8.91	Too dry
Hole 2	*2.7 m*	*Sharp transition*	-	-
Hole 2	2 - 3	Diamicton/ice	22.66	0.3
Hole 2	3 - 4.5	Ice	Not available	0.1
Hole 2	4.5 - 6	Ice	Not available	0.1
Hole 2	6 - 7	Ice	Not available	Too dry
Hole 2	7 - 8	Ice	Not available	0.2
Hole 2	8 - 9	Ice	Not available	0.2
Hole 4	0 - 0.5	Mixed gravels	4.71	Too dry
Hole 4	0.5 - 1	Mixed gravels	5.74	Too dry
Hole 4	1 - 1.5	Mixed gravels	23.03	0.7
Hole 4	1.5 - 2	Mixed gravels	21.46	1.8
Hole 4	2 - 2.5	Mixed gravels	16.06	1.5
Hole 4	2.5 - 3	Mixed gravels	19.41	2.6
Hole 4	*2.9 m*	*Sharp transition*	-	-
Hole 4	3 - 3.5	Silty clay - few pebbles	21.67	4
Hole 4	3.5 - 4	Silty clay - few pebbles	21.37	4
Hole 4	4 - 4.5	Silty clay - few pebbles	23.18	3.6
Hole 4	4.5 - 5	Silty clay - few pebbles	33.73	4
Hole 4	5 - 5.5	Silty clay - few pebbles	31.58	2.6
Hole 4	5.5 - 6	Silty clay - few pebbles	25.82	4.6
Hole 4	6 - 7	Silty clay - few pebbles	26.42	4
Hole 4	7 - 8	Silty clay - few pebbles	34.67	3.4
Hole 4	8 - 9	Silty clay - few pebbles	34.03	3.5

Part 2

V

Temperatures in Coastal Permafrost in the Svea Area, Svalbard

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Abstract

Temperature data from three boreholes located on an ice-cored moraine near sea level are analyzed. One of these boreholes was drilled 6 m from the shore and shows significantly higher temperatures than the holes about 150 m from the shore. Using meteorological data and measurements of water temperatures, we model the permafrost distribution into the fjord as well as the influence of the sea on permafrost temperatures near the shore. The model results suggest that permafrost, as defined solely on temperature, is present beneath Van Mijenfjorden.

Keywords: borehole temperatures; coastal; geothermal modeling; permafrost; subsea permafrost; Svalbard.

Introduction

Permafrost on Svalbard is classified as continuous. It is more than 500 m thick in the highlands and less than 100 m near the coasts (Humlum et al. 2003). While considerable knowledge of permafrost conditions in the mountains exists from the extensive coal mining, little has been published on the permafrost in the shore areas of Svalbard. Exceptions are Gregersen & Eidsmoen (1988) that compares deep borehole temperatures at the shore with inland boreholes in Longyearbyen and Svea, and Harada & Yoshikawa (1996) that uses DC resistivity soundings to estimate the permafrost thickness of marine terraces at the shore of Adventfjorden near Longyearbyen.

The aim of this paper is to describe the permafrost conditions in the ice-cored moraine, Crednermorenen,a peninsula in Van Mijenfjorden in central Spitsbergen (Fig. 1). Since April 2005 temperatures have been logged every two hrs in three boreholes on the moraine each with 16 thermistors down to eight m. One of the boreholes is located only six m from the shoreline. Permafrost conditions in the shore area are important since the permafrost here is "warm" and thin. When constructing in such areas, particular attention must therefore be given to the permafrost conditions. In most other parts of Svalbard, permafrost is thicker, colder and more stable.

Due to the few previous studies of near-shore permafrost on Svalbard, an attempt was made to model both the effect of the sea on the onshore permafrost as well as the possibility of subsea permafrost. We use a transient 2D finite element geothermal model (TEMP/W from Geoslope International, Calgary, Canada; Krahn 2004) that is forced with meteorological data and measured water temperatures as boundary conditions.

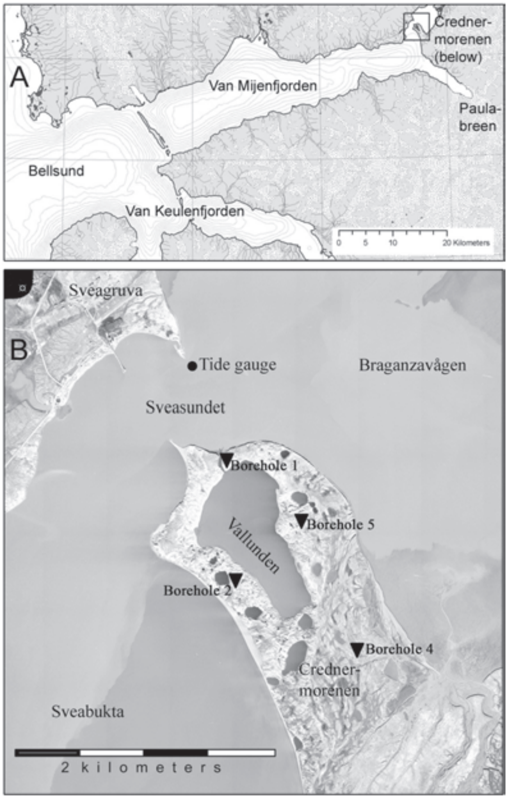

Figure 1. Field site. A) Van Mijenfjorden with Crednermorenen (in upper right corner). Map from: http://www.iopan.gda.pl. B) Crednermorenen with the location of the boreholes and the tide gauge. Air photo: Norsk Polarinstitutt, 1977.

Field Site

Crednermorenen is a lateral moraine deposited by a surge of the tidewater glacier Paulabreen (Fig. 1) around 1300 A.D. (Hald et al. 2001). The moraine forms a peninsula (1x3 km),

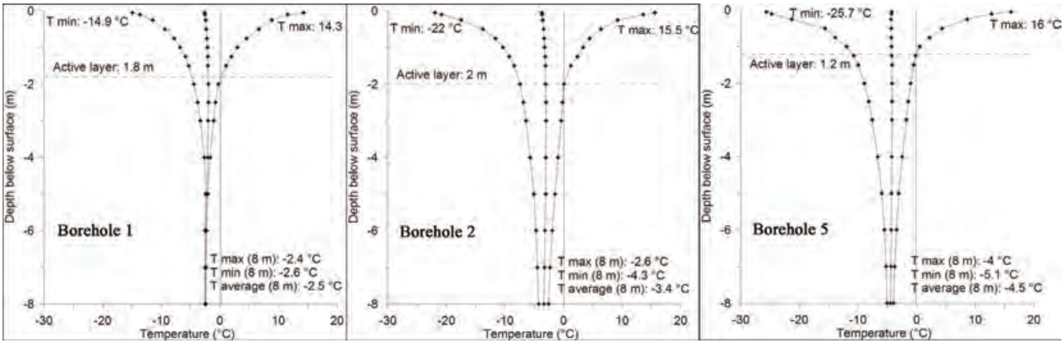

Figure 2. Maximum, minimum, and average temperatures in all depths from 10 September 2006 to 9 September 2007 for Boreholes 1, 2, and 5.

Table 1. Average annual temperature at the ground surface, at the top of the permafrost (TTOP).

	(Air)	Hole 1	Hole2	Hole 5
T surface (°C)	-4.6	-2.6	-3.7	-4.2
TTOP (°C)		-2.0	-3.1	-4.3

partly ice-cored and partly consisting of proglacially pushed marine clays (Kristensen et al., in press). It is surrounded by water on three sides and in its northern part lies a 1 km long lagoon, —Vallunden—that, is connected to the sea by a 15m-wide channel near Borehole 1. The water in Vallunden is salty as the tide flows in and out through the channel.

The oceanography in Van Mijenfjorden is strongly affected by an island at the fjord mouth, Akseløya (Fig. 1A), which nearly blocks the water exchange between the fjord and the warmer Atlantic water outside (Nilsen 2002). The water column is thus dominated by cold local water. Shore-fast ice is usually present from December to June. Bottom water salinity is around 34‰ (Hald et al. 2001), and July temperatures of -1.53°C and and -1.27°C have been measured in two basins in Van Mijenfjorden at 112 m and 74 m depth (Gulliksen et al. 1985). The climate in Sveagruva is slightly colder and more humid than in Longyearbyen 45 km to the NNE, but the meteorological record is shorter and more irregular. Mean annual air temperature close to sea level was -5.4°C in the period 1997–2006, and precipitation in the period 1995–2002 was on average 244 mm/y (www.met.no).

Temperature Measurements in Boreholes

Four 10 m deep boreholes (Fig. 1B) were drilled on Crednermorenen in March 2005 using an air pressure driven drilling rig. Coring was not possible, but the pulverized blown up sediment was collected from three of the holes, described and analyzed for water content and salinity. Borehole 1 was located two m above sea level and six m from the channel that connects Vallunden with Sveasundet. Borehole 2 was drilled into the ice-cored part of the moraine 17 m a.s.l. and 150 m from Sveabukta. Borehole 5 was established on top of the moraine ridge 145 m from Vallunden and 20 m a.s.l. Hole four was located on the marine clay part of Crednermorenen, but was destroyed by a bear in October 2006, and therefore no data are presented from that hole. In each borehole an eight m thermistor string (EBA Engineering, Edmonton, Canada) with 16 thermistors at decreasing spacing towards the surface was inserted. The uppermost sensor in each hole was placed at roughly three cm depth, and was measuring the surface temperature in this paper. Temperatures have been logged every two hour using Lakewood dataloggers; the accuracy of the thermistors is around 0.1°C. The annual temperature envelopes recorded in the three boreholes are shown in Figure 2. The maximum surface temperature for holes one and five occurred on 16 July 2007, and for hole two on 18 June 2007. The maximum temperature was very similar for the three holes, probably indicating that the barren surface provided similar summer conditions. The minimum surface temperature for all three holes occurred on 23 January 2007, which was contemporary with the minimum air temperature (-32.9°C) being recorded. The minimum surface temperature was much lower in hole two and five than in hole one. The latter had usually a snow cover of around 20 cm whereas both holes two and five were usually snow-free in winter due to wind redistribution in these more exposed sites. In Borehole 1 the seasonal temperature fluctuation became insignificant (0.25°C) below six meters depth, whereas the difference between annual maximum and minimum temperature at eight m in Boreholes 2 and 5 were 1.7°C and 1.1°C respectively.

Table 1 shows that all ground surface temperatures in the investigated period were higher than the mean air temperature. Hole one was warmest, reflecting the thickest snow cover during winter. The snow insulated the surface against cold winter temperatures creating a positive surface offset as demonstrated by, for example, Smith & Riseborough (1996). Hole 5 had the smallest surface offset as this site is usually never snow covered.

Smith & Riseborough (1996) also demonstrated that, due to higher thermal conductivity in frozen ground than in unfrozen, temperatures will tend to decrease from the ground surface to the top of the permafrost table (TTOP). Table 1 shows

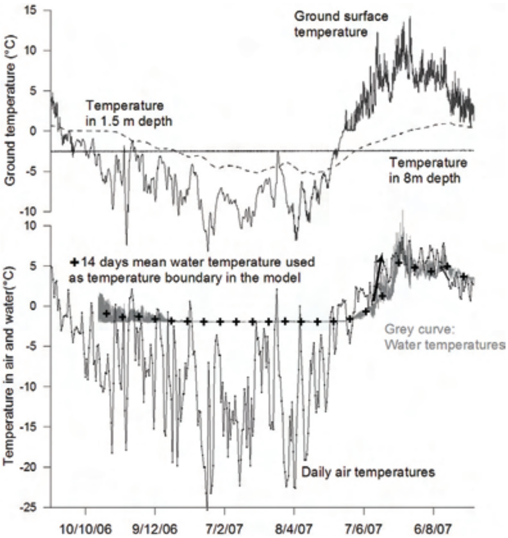

Figure 3. Upper part: Temperatures in Borehole 1 at the surface, in 1.5 m depth (dotted line) and in 8 m depth (no variation). Lower part: Air temperatures and water temperatures.

that in Boreholes 1 and 2 TTOP was higher than the surface temperature whereas in hole five it was practically the same. The active layer offset therefore seemed not to be important for the ground temperatures on the moraine, whereas snow depth in winter certainly was.

Tide and Water Temperature Measurements

Water temperatures in the narrow and shallow strait Sveasundet were measured and recorded every 20 min from 10 October 2006 to 9 September 2007 by a tidal gauge placed at two m water depth. The data can be seen in Figure 3 together with temperature measurements from Borehole 1 from three selected depths and air temperature measurements over the same time interval.

The freshwater from Kjellstrømsdalen passes the Sveasundet strait, and strong tidal currents flowing in and out of Braganzavågen ensure mixing of salt and fresh water here. For this reason the summer water temperatures were high compared to what has been measured in the deep basins in the fjord during summer. Winter temperatures however appeared to be constant around -1.93°C in most of the fjord.

The measured water temperatures can be divided in three distinct periods:

1) An autumn period when the temperature fluctuated in relation to the tide between -1.9 and +1.7°C lasting from 21 October to 26 December. Temperatures rose when the tide was moving through Sveasundet into the tidal flat Braganzavågen and fell when the tide flowed out again. This is consistent with observations that sea ice started forming in Braganzavågen before other places in the fjord. An example

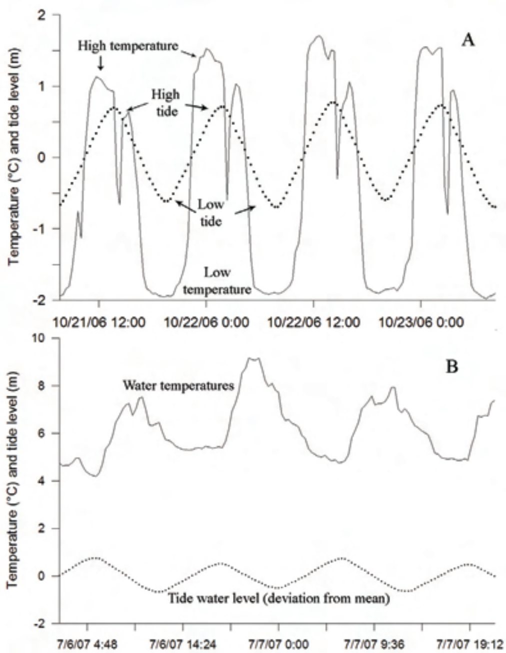

Figure 4. Examples of the relationship between water temperatures and the tide. A: Autumn conditions when the first sea ice forms, B: summer conditions.

of the coupling between tide and temperatures in the autumn can be seen in Figure 4A**.** November 28 was the last time the temperature rose above zero.

2) A winter period from 26 December to 22 May 2007. The temperature was nearly constant around -1.93, and no temperature fluctuations with tide were observed. The temperature corresponded to the freezing point of sea water and the fjord was covered by ice throughout this time. This winter period with constant water temperatures is easily identifiable in Figure 3.

3) A summer period from 22 May 2007 to the end of the measurements. Here the water temperatures increased and gradually approached the air temperature. The water temperature rose above zero for the first time on 13 June 2007. The temperature fluctuations were again controlled by tidal currents and were opposite in phase to those of the autumn. Now rising tide was associated with lower temperatures and falling tide with increasing temperatures (Fig. 4B). The reason is that water was now heated in the tidal flat Braganzavågen, where it was cooled during the autumn.

Modeling the Effect of the Sea on the Permafrost Temperatures

Studying the measured borehole temperatures, one can see that Borehole 1 deviated significantly from Boreholes

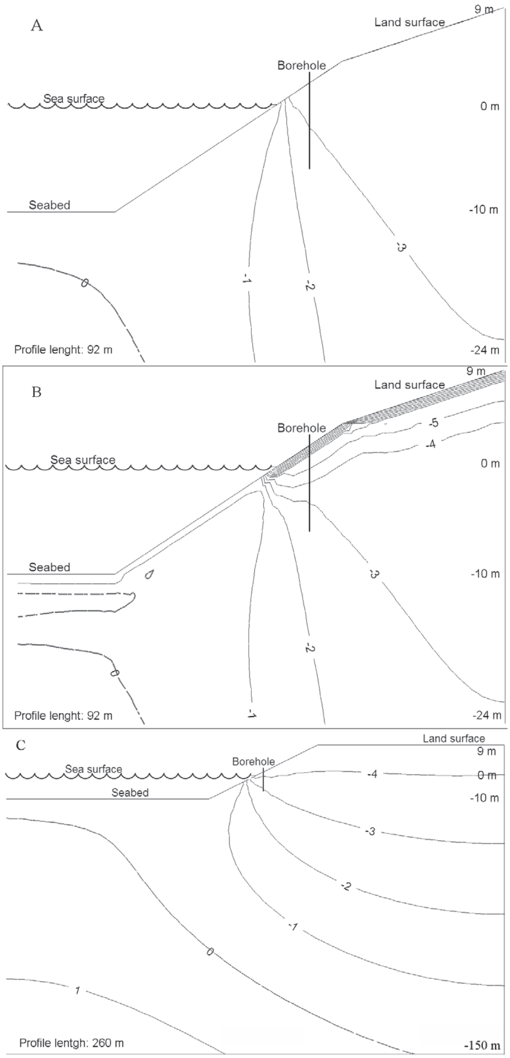

Figure 5. Three model output results. A: Steady state simulation of the ground temperatures in the immediate shore area. B: The same section but run in a transient mode forced with measured climatic data and sea water temperatures. Snapshot from 7 April 2007. C: Steady state simulation of a wider and deeper section with the boundary conditions the same as in A.

2 and 5 in respect both to the thermal regime and the depth of zero annual amplitude. While some of this deviation can be explained by a thicker snow cover during winter, most likely the proximity (6 m) to the sea affected the permafrost temperatures here. Located two m a.s.l. and being eight m deep, most of the borehole also lay below sea level. At eight m depth in Borehole 1, the temperature was -2.5°C. This indicates that permafrost probably extends into the seabed

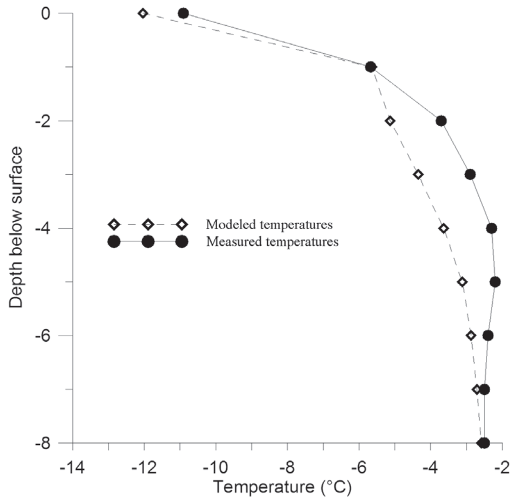

Figure 6. Comparison of the temperatures measured in Borehole 1 on 7 April 2007 and the modeled temperatures on the same day.

from the shore. An attempt was made to model both the effect of the sea on the onshore permafrost temperatures as well as the possibility of subsea permafrost existence. Gregersen & Eidsmoen (1988) previously tried to model the possible subsea permafrost in the area, but they had no information on the water temperatures in the fjord.

Model description and input

A 2D finite element program (TEMP/W) was used to model the permafrost thickness in Crednermorenen and the extent below the fjord bottom. The model is described in detail by Krahn (2004). Two temperature-dependent input functions (unfrozen water content and thermal conductivity) and overall water content are laboratory data from a nearby moraine, Damesmorenen, four km from hole 1, published by Gregersen et al. (1983). Volumetric heat capacity was set to 2000 and 3000 kJ/(m^3 x K) for frozen and unfrozen states respectively. Only one set of thermal properties was supplied to the model. It is obviously incorrect to assume homogeneous subsurface conditions, but we have no other thermal properties data available nor information on the subsea stratigraphy. A geothermal gradient of 35 mW/m^2 was set as a flux boundary condition at the bottom of the profiles. The vertical profile sides were set as zero flux boundaries.

To obtain a first estimate of the subsurface temperatures, the model was initially run in a steady state mode (Figs. 5A, 5C) using an estimated annual surface temperature and the average water temperatures as upper boundary conditions. A ground surface temperature of -4°C was used; this is slightly warmer than the mean annual air temperature due to the surface offset demonstrated in Table 1. A temperature of -0.1°C was used as the seabed boundary condition; this is the mean annual measured water temperature with interpolated temperatures for the missing 1.5 months of data.

The model was also run in a transient mode (snapshot in Fig. 5B) to compare the model results with the temperatures measured in Borehole 1. In the transient mode, eight simulations were run per day, and each node result was input to the next model run. Here the model was forced with meteorological data from 1 Jan 2006 to 10 September 2007. The meteorological inputs were maximum and minimum daily temperature, maximum and minimum daily humidity, and average wind speed. Latitude and longitude were supplied and the TEMP/W program used an energy balance approach to model the surface energy balance. To simulate the seabed temperature, a time dependent te mperature function was supplied as boundary condition, consisting of the average measured water temperature on 14 day basis. These are seen as crosses in the lower part of Figure 3.

No attempts were made to simulate the tidal fluctuation and its affect on the ground temperatures.

The model was run on two profile sections of different lengths and depths to both obtain detailed information on the near surface conditions, and impressions of the larger-scale ground temperatures in the coastal zone. The profiles were 92 m and 260 m long respectively and simplify a profile across the moraine and into Vallunden crossing Borehole 1.

Model results

Figure 5A shows a steady-state simulation for the immediate shore area. A high horizontal thermal gradient is seen in a narrow zone just below the shoreline. Since the mean annual water temperature is slightly below zero (-0.1°C), permafrost is modeled to be present in a thin layer below the seabed.

Figure 5B shows a snapshot plot from the transient model run from 7 April 2007. The sharp decrease of near surface temperatures reflects the winter freezing on land. A -1 °C isotherm has formed close below the seabed reflecting that the water temperatures approach -2°C during winter.

Figure 5C is a model run of the larger and deeper section but with the boundary conditions as those of Figure 5A. The pattern is similar as the one in Figure 5A but suggests that at depth, the presence of the sea will affect the ground thermal conditions more than 100 m from the shore, and similarly, that the cold temperatures from land will affect the sub-seabed temperatures at a similar distance offshore.

Figure 6 compares the measured and modeled temperatures in Borehole 1 for 7 April 2007. The discrepancy of model temperatures near the surface and towards the bottom is quite small, whereas the modeled temperature is up to 1.4°C wrong in the middle of the borehole. This and other snapshots throughout the year show that, while the general pattern is simulated reasonably well, there are discrepancies. These are often larger than those shown in Figure 6. However, the reasonable agreement of the modeled to the measured temperatures gives us some confidence in the general modeling results.

Discussion

The pronounced sill, Akseløya, restricts warm coastal water from entering Van Mijenfjorden and probably makes this fjord colder than other western Spitsbergen fjords. Sea ice cover is longer-lasting and more stable here. Therefore, this fjord is a primary candidate for possible subsea permafrost in western Spitsbergen fjords.

The modeling results of the subsea permafrost extend presented here should be seen as a minimum scenario. This is because the water temperatures were measured in a shallow, high-current strait, where the fjord water is strongly mixed with warmer fresh water during the summer. July temperature measurements from two deep basins in Van Mijenfjorden (Gulliksen et al. 1985) of -1.53°C and -1.27°C, respectively, indicate that water temperature in the deeper parts of the fjord remains below 0°C all year.

Permafrost, as defined solely on the basis of temperature, may not necessarily indicate cryotic subsea conditions. Sea water freezes at temperatures slightly above -2°C but capillarity and adsorption—in particular in fine-grained sediments—can further reduce the freezing point (Williams & Smith 1989). Thus, depending on the sediment properties, the seabed may well have permafrost by definition but still remain unfrozen. If the seabed consists of saline marine deposits, they will not be cryotic, even if thermally defined permafrost exists.

The 1300 A.D. surge of Paulabreen deposited lateral moraines in a rim around the inner parts of the fjord. A new detailed bathymetric survey indicates that glacial deposits also occupy the seabed here (Ottesen et al., in prep.). A seabed consisting of terrestrial sediments of glacial origin and with fresh rather than saline porewater could actually be frozen, but this hypothesis has not yet been tested.

The Crednermorenen moraine contains large amounts of buried glacier ice. It is possible that the unusual cold water conditions in Van Mijenfjorden are influencing the preservation potential of the ice-core in this moraine.

Conclusions

The permafrost temperatures measured in three boreholes in the ice-cored Crednermorenen moraine were studied for a period of a year. The surface temperatures in all holes were higher than the corresponding air temperature. The highest surface temperature was measured in Borehole 1 that normally has a snow cover of 20 cm while the two other boreholes are nearly snow free during winter. Most likely the warmer surface temperature in Borehole 1 is due to a surface offset (Smith & Riseborough 1996) caused by the insulating effect of snow.

Increasing temperatures were observed from the surface down through the active layer to the top of the permafrost in two of the boreholes. This is opposite to what would be expected if higher thermal conductivity of frozen ground compared to unfrozen ground causes an active layer (or thermal) offset. So this offset appears not to be important here; probably the ground is too dry.

Borehole 1 is located six m from the shore and is significantly warmer than two other boreholes both about 150 m away from the shoreline. To investigate the effect of the proximity to the sea, the finite element program TEMP/W was used to model the ground temperatures at the shore and below the seabed in both a steady-state and transient mode. Meteorological data and water temperature measurements were used to force the model.

The program manages reasonably well to simulate the ground temperatures in the near-shore borehole. The simulations indicate that permafrost, as defined solely on temperature, is present in a thin layer beneath the seabed of Van Mijenfjorden. Whether it is frozen or unfrozen will depend on the material properties.

At depth, the warming effect of the sea on the ground temperatures is modeled to penetrate more than 100 m inland and the cooling effect of land is affecting the seabed at an equal distance. The temperatures closer to the surface, however, are primarily locally controlled.

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