Controls on debris-covered glacier fronts in Central and Western Svalbard

A study of arctic glacier termini

Ineke Irene Rookus

Master Thesis submitted for the degree of
Master of Science in Geosciences
60 credits

Institute of Geosciences
Faculty of Mathematics and Natural Sciences

UNIVERSITY OF OSLO

November 2016

Controls on debris-covered glacier fronts in Central and Western Svalbard

A study of arctic glacier termini

Ineke Irene Rookus

Master thesis in geosciences

UNIVERSITY CENTRE IN SVALBARD
UNIVERSITY OF OSLO

November 2016

2016

Controls on debris-covered glacier fronts in Central and Western Svalbard

Ineke Irene Rookus

Cover picture: Longyearbreen in September 2014, by Ineke Irene Rookus

http://www.duo.uio.no/

Print: Reprosentralen, Universitetet i Oslo

Abstract

Debris-covered ice is both common and can greatly change the response of a glacier to climate change. Therefore, realistic predictions of future water availability and global sea-level change need to include debris-covered ice. This study presents an empirical model that estimates the debris-covered ice surface of glaciers of up to 15 km^2, based on the clean ice surface area. Analysis of 91 glacier termini in Nordenskiöld Land, Oscar II Land and Prins Karls Forland, on which the model is based, result in a clear negative correlation, and probable relation, between a glacier's clean surface area and its debris-covered fraction. Differences in geology, topography and climate do not seem to impact the correlation between clean glacier size and debris cover. However, surge-type glaciers are unlikely to form debris-covered termini.

Longyearbreen, a glacier of which the front is covered by on average 0.5 m of debris (n=36), serves as a field example in this study. The debris on this glacier is sourced by avalanches and rockfall events and transported supra- and englacially, occasionally by glacial streams. Temperature sensors have been installed at different depths in the debris layer. The resultant time series are combined with weather data and on-field observations, like slumping and collapses during and straight after mid-freezeback rain showers. Analysis of the data indicates that the debris cover is a better insulator during summer-time than during winter-time. This may explain the great impact that a debris cover can have on a glacier's response to climate change. Overall, climatic changes towards dry conditions are likely to enhance debris-covered ice preservation, but the perfect scenario to preserve debris-covered ice would combine wet autumns, cold, dry and/or windy winters and dry summers.

Acknowledgements

Special thanks go to my supervisor, professor Ole Humlum for his encouraging and enthousiastic manner of suppervision and his help and understanding when progress was slow. I could in all honesty not have wished fo a better supervisor.

Many people helped me out during the fieldwork campaigns and this thesis would not have been here without their help. Great thanks go to Dani Roehnert, for all her help and companionship in the field and in particular for her relentless digging. Untill one attempts to dig a meter deep hole in a glacial debris layer, my gratetute for her help in this cannot be fully understood. My thanks also go to professor Hanne Christiansen for providing the temperature sensons and extracting the data from them. Without her help at these times when professor Ole Humlum was not in Svalbard the temperature measurements couldn not have been performed. I would also like to thank Kjersti Kalhagen and Stefan Schöttl for their long days out doing snowdepth measurements on Longyearbreen with me. Although the results could unfortunately not be used to their full potential, as choises have to be made and time is always limited, these first days of fieldswork formed the best thesis start anyone can have.

And last but definately not least, I ow a big thank you to Emil Kiesbye Larsen and Stijn Hofhuis for helping me dig the temperature sensors up again a year later. I must confess they could have hardly realized what kind of harsh work they signed up for, just a few weeks after moving to Svalbard. And I am much obliged to their enthousiastic attitude and relentles digging efforts in ground that is simply not ment for digging.

Contents

1. Introduction

A large portion of all glaciers in the world is debris-covered (IPCC, 2013; Scherler et al., 2011). Yet debris-covered ice is often ignored or otherwise neglected in climate studies (Raper and Braithwaite, 2006; Rees and Collins, 2006; Immerzeel et al., 2010; Slangen et al., 2014; Mutz et al., 2016). This is perhaps not surprising. As the latest IPCC report stated, modelling short-term glacier response, or the long-term response of more complex glacier types (for example those that are heavily debris-covered) is difficult and requires knowledge that, for the majority of glaciers worldwide, is simply unavailable. Many of the remote-sensing based assessments chose to not discriminate between these glacier types (IPCC, 2013). More mass-balance studies from heavily debris-covered glaciers, inclusion of debris cover in glacier inventories, and adequate models covering large spatial scales that allow for the effect of debris cover are needed, which are currently all missing (Scherler et al., 2011).

One of the great obstacles preventing the study of debris-covered ice is the remote location of many glacial regions. This greatly hampers ground-based monitoring. When mass-balance data are unavailable, scientists can refer to changing glacier front positions (Oerlemans, 2005) and surface areas. Such data can more easily be retrieved via aerial photography and optical satellite imagery and thus does not require in-situ observations. But the resultant data often only comprehends the clean, debris-free ice and overlooks ice surfaces hidden beneath debris.

However, for realistic predictions of future water availability and global sea-level change, debris cover and its influence on glacial-melt rates should be included. The insulating effect of a debris cover can greatly change the response of a glacier to climate change (Mattson et al., 1993). Central Himalaya is home to several debris-covered glaciers with stagnant tongues that extend several kilometers upstream from their termini (Bolch et al., 2008; Quincey et al., 2009; Scherler et al., 2011). More than 65% of the monsoon-influenced glaciers that were observed by Scherler et al. (2011) are retreating, but heavily debris-covered glaciers with stagnant low-gradient terminus regions typically had stable fronts. Although growing meltwater ponds and surface lowering show that the mass balances of these Himalayan glaciers, and others alike in Canada and New Zealand, are negative, their fronts remain remarkably stable (Ogilvie, 1904; Kirkbride, 1993; Bolch et al., 2008). The fact that their fronts do not retreat indicates that even greater negative mass balances are prevented by the insulating debris cover.

By not accounting for debris-covered ice, current climate studies not only underestimate glacier masses, but also misrepresent the response of certain glaciers to climate change. It is clear that this subject cannot be ignored. Scientists must work towards finding ways to account for debris-covered ice and its influence on glacial melt rates.

A first step towards including debris-covered ice in climate scenario studies is to identify debris-covered glacial areas. It is the aim of this study to determine how much of the surface area of glaciers in Central and Western Spitsbergen is debris-covered and to provide a simple model to predict the debris-covered fraction of a glacier. The study additionally aims to form a better understanding of processes that influence the formation and degradation of debris-covered glacial ice. To do so this thesis looks into subjects like debris supply and transport, headwall area, surging and responses to climatic and seasonal differences or changes. Temperature sensors have been installed at different depths in the debris layer of the glacier Longyearbreen (figure 1.1) in order to conduct a case study. This yearlong data series has been combined with data from the nearby Adventdalen metrological station to determine the isolating properties of the debris layer throughout the seasons.

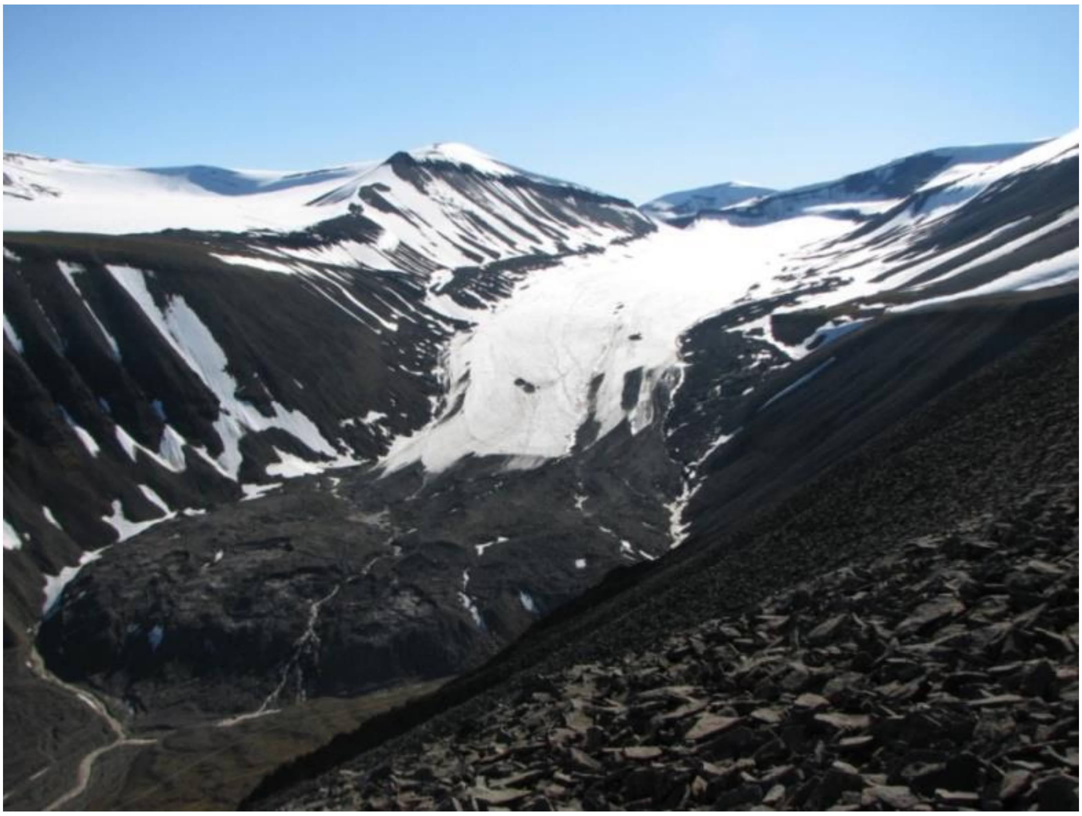

Figure 1.1: Longyearbreen is a relatively small glacier with a large, rounded, debris-covered front and a steep frontal edge. The transition from covered to clean ice is gradual, although it looks sharp on this photo. Picture by Ole Humlum.

2. Description of study areas

2.1. Location of study areas

Svalbard is a Norwegian archipelago that consists of several islands and covers an area of about 61.020 km^2 (Yao et al., 2012). The biggest island is Spitsbergen with 37.673 km^2 (Bringedal, 2004). Nordaustlandet, Barentsøya, Edgeøya, Kong Karls Land, Hopen, Prins Karls Forland, Bjørnøya and several other islands are also part of the archipelago. The archipelago of Svalbard lies in the northern part of the Barents Sea between 74°N to 81°N and 10°E to 35°E. Most of the human activities have occurred on the main island: Spitsbergen. All four permanent settlements in Svalbard (Longyearbyen, Barentsburg, Ny-Ålesund and Sveagruva) are found on this island. The fieldwork for this project was done on Longyearbreen (located by a red arrow in figure 2.1), a glacier near the town of Longyearbyen.

The landscape of Svalbard is varied. The highest mountains are found on the island of Spitsbergen, with Newtontoppen spanning the crown at 1712 m. Not surprisingly, Spitsbergen is also where you find the alpine peaks that gave the island its name. But this is not to say that the whole island has an alpine signature. The island centre is dominated by plateau-shaped mountains, interspersed with wide valleys with extensive braided river systems. The east of Spitsbergen and the smaller (eastern) islands of the archipelago are hilly or flat, rather than mountainous. The east is also the domain of extensive ice-caps, covering up much of the terrain underneath.

To address the differences and/or similarities between glacier-margins on the West Coast of Svalbard and in Central Spitsbergen this study gives special attention to Nordenskiöld Land, Oscar II Land and Prins Karls Forland (encircled in figure 2.1). Including glaciers of such different climatic settings ensures that the results found later in this study account for climatic variability and will thus be able to produce results that are likely to hold for many glaciers in different climatic settings. On the other hand, if the glaciers prove to be strongly influenced by their climatic setting, these influences can be further analysed.

Nordenskiöld Land is located between the two large fjords Isfjorden and Van Mijenfjorden. This area is of particular interest as it covers the central part of Spitsbergen, an area in which many glaciers with large debris-covered fronts are found that form the object of this study. One of these glaciers, Longyearbreen, has been the fieldwork site for this study. Nordenskiöld Land lies between 78.4°N and 77.7°N and covers an area of about 7170 km^2. The highest mountains in the area are above 1000 m and can be found in the inland part of Nordenskiöld Land, whereas the more alpine mountains to the west are around 700 m high. As mentioned before, the inland area of Nordenskiöld Land is a landscape dominated by plateau-like mountains and wide valley- and fjord-systems that are roughly orientated east – west. The major valleys have a distinct U-shape and contain great braiding river systems. The shape of these valleys presumably is the result of erosion by repeated glaciations. However, one would expect the tributary valleys to have a similar shape, since glacial erosion is likely to have affected these valleys as well. The tributary valleys however, often display a characteristic V-shape. Therefor it may be that the spacious U-shaped valleys are in fact V-shaped on the bedrock level. Today's shape could be the result of fill-in by fluvial sediments and slope-angle decrease towards the valley's base as a result of talus fans.

V-shaped tributary valleys can even be found underneath a glacier. An example is Longyeardalen. The glacier Longyearbreen, where the fieldwork for this project has been done, hides the underlying valley, but radio-echo backscatter revealed the V-shaped channel-profile that lies beneath the ice. This indicates that the area is one of low subglacial activity with a high possibility of preglacial form preservation in the presence of cold ice (Etzelmüller et al., 2000).

Oscar II Land and Prins Karls Forland are of interest to this study because of the many coastal glaciers they comprise. This sets a distinctly different climatic setting from the inland glaciers of Spitsbergen. Oscar II Land lies between the largest fjord of Spitsbergen, Isfjorden, to the south and the smaller Kongsfjorden, harbouring Ny-Ålesund, to the north (figure 2.1). The 30 km long grounded tidewater glacier Sveabreen (Chapuis and Tetzlaff, 2014) forms the eastern border of the study area. Oscar II Land spans from 79°N at Kvadehuken to 78.2°N at Isfjorden. The highest mountain in the area is Hofgaardtoppen with 1125 m. It is however not much more than a rocky peak penetrating the ice. Like other high peaks in Oscar II Land, this mou8ntain is in fact a nunatak; its base is hidden from view by glacial ice. Oscar II Land is, unlike Nordenskiöld Land, heavily glaciated. Here one will not find the wide valleys and large river systems that characterise Nordenskiöld Land. Instead valleys are buried under hundreds of meters of glacial ice, leaving mountain peaks buried or protruding in the form of nunataks.

But the fact that Oscar II Land is heavily glaciated is not to mean that the area is uniform. The coastal zone of Oscar II Land is less glaciated than the inland area and displays a much more varied picture. It is home to large glacier tongs fed by inland ice-caps, but also houses small glaciers much like the ones in Nordenskiöld Land and every imaginable glacier size in between. This varied coastal zone, incorporating fjords and regular seasides, is subjected to further study in later chapters in this thesis.

The island Prins Karls Forland is located roughly 15 km west of Oscar II Land and spans from 78.9°N to 78.2°N. The 86 km long, narrow island covers an area of 615 km^2 (measured using the mapping service of SvalbardKartet), effectively sheltering the whole coastline of Oscar II Land. With North Greenland being the nearest Land to the west, waves from the Arctic Ocean break along the western shore of the island. The landscape is varied, hosting alpine summits, grassy plains and barren rock deserts. Historically, Prins Karls Forland had many sea-terminating glaciers, but widespread retreat in recent time (figure 2.4) has made quite a few of these glaciers land-terminating. Calving glaciers however do still exist on the island and today's collection of sea-terminating and former sea-terminating glaciers, mixed with land-terminating small (shorter than 5 km) ones and even full-fledged rock glaciers (Berthling et al., 2000), makes Prins Karls Forland glacially a very diverse island.

Figure 2.1: The study areas of Longyearbreen (red arrow), Nordenskiöld Land (encircled) and Prins Karls Forland together with Oscar II Land (encircled) in Svalbard (basemap from Svalbard Kartet, Norwegian Polar Institute). The inset map has been taken from Jakobsson et al. (2008) and shows the location of Svalbard in the Arctic ocean.

2.2. Recent climatic history and current climate

Svalbard has the longest meteorological series of the High Arctic. Meteorological observations were initiated in 1898 (Nordli et al., 2014). The data of the first 13 years was acquired on scientific and hunting expeditions. The first more or less continues temperature recordings in Svalbard started at Green Harbour, near Barentsburg (figure 2.2), in 1912 (Hanssen-Bauer et al., 1990, Førland et al., 1997). From then on, more weather stations were erected in, among other areas, Longyearbyen, Isfjord Radio and Svea (figure 2.2). These long-term series have long been combined to form the 'Svalbard Airport composite series'. Since 2014 this composite series also includes the early, expedition based, measurements (Nordli et al., 2014). By erecting new automatic weather stations near former expedition cabins, with the specific aim of calculating transfer functions between the old sites and the present Svalbard Airport station, (Nordli et al., 2014) could incorporate these early observations.

Around 1920 a few years of rapid warming changed the mean annual air temperature (MAAT) at sea level from *ca.* -9.5 °C to -5.5 °C (Humlum et al., 2003). This drastic change is usually considered to represent the termination of the Little Ice Age (LIA) in Svalbard (Humlum et al., 2003). From 1957 to 1968 Svalbard experienced a period of cooling of about 4 °C followed by a gradual warming towards the end of the 20th century (Humlum et al., 2007). Today's early 21st century MAAT of about -5 °C (Humlum et al., 2007, Siewert et al., 2012) is similar to the warm temperatures recorded during the 1930s and about 4 – 5 °C higher than the MAAT of the late LIA.

The current MAAT of about -5 °C does not represents the average temperature for the whole of Svalbard, as there are large regional differences. Warm waters of the Gulf Stream reach all the way up to the west coast of Spitsbergen. This flow of warm water from south to north, known as the West Spitsbergen Current, is responsible for a clear temperature division in Spitsbergen. The Islands east coast is cold, the west coast relatively warm.

Precipitation in Svalbard is even more liable to regional variations. While the centre of Spitsbergen is considered to be arid with an average annual sum of about 180mm water equivalent (w.e.) (Humlum et al., 2007), the coastal areas of the island receive much more precipitation. At Isfjord Radio, 50 km south west of Longyearbyen, the annual amount (435 mm w.e.) is more than twice the precipitation recorded at Longyearbyen (Humlum, 2002). Both stations are low altitude weather stations close to the shore.

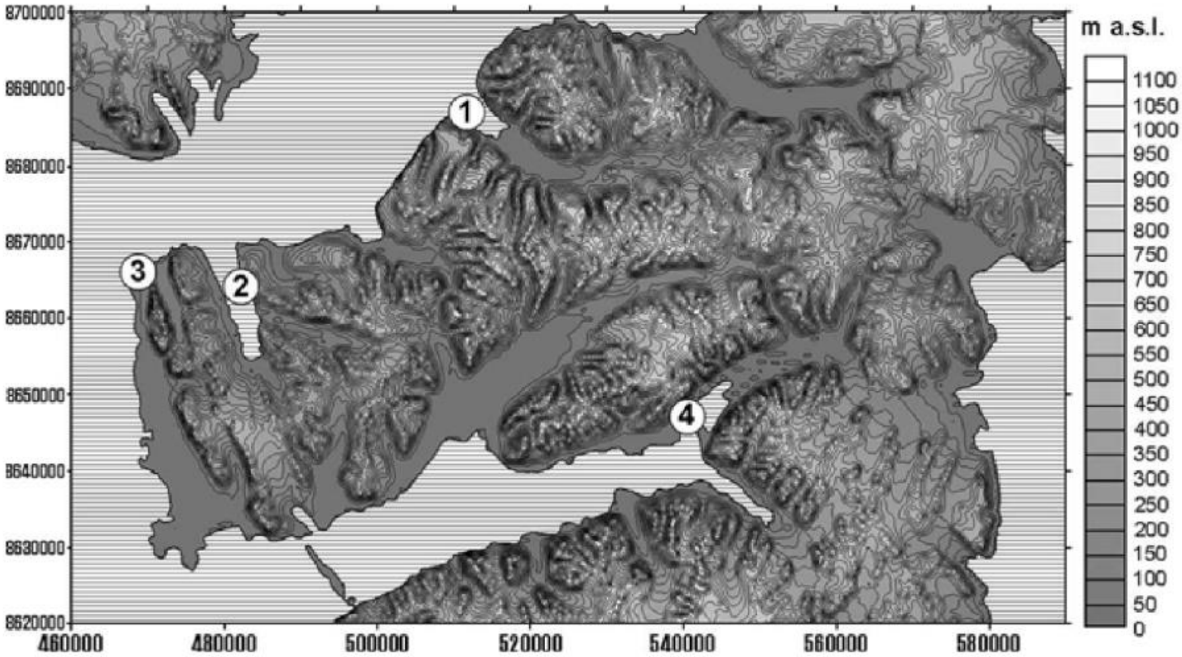

Figure 2.2: Reproduced from Humlum (2002). Topographic map of Nordenskiöld Land. North is towards the top of the diagram. Data source: Global Land One-kilometre Base Elevation, National Geophysical Data Center (NOAA). Altitudinal scale in m a.s.l. The horizontal dimensions of the diagram are 130 and 80 km, respectively. The numbers 1 to 4 indicate the locations of Longyearbyen, Barentsburg, Isfjord Radio and Svea, respectively.

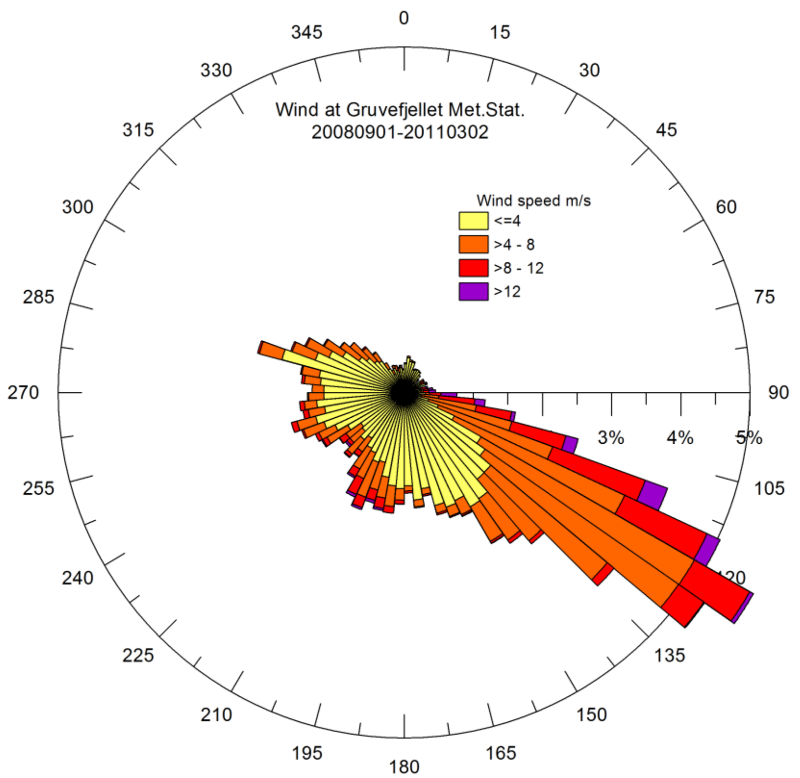

Figure 2.3: The prevailing wind direction at the Gruvefjellet meteorological station, nearby Longyearbreen. This station is located on a mountain-plateau and thus not influenced by valley-funnelling.

Due to the open and spacious character of Svalbard, free of trees and other sheltering structures, the islands are fully exposed to the wind. Especially in the relatively arid centre of Spitsbergen, glaciers are therefore heavily reliant on snow redistribution by wind (Humlum, 2002). The general wind direction within the study area is ESE and is well portrayed by the weather station Gruvefjellet near Longyearbyen (see section 5.2.6 'The influence of wind on debris-covered glacier margins') (figure 2.3).

This dependence of glaciers on wind can even be recognised by simply looking at a topographical map of the Nordenskiöld area and the orientation of the glaciers on the map. Glaciers will often be orientated away from the dominant wind direction, on the leeward side of a headwall. When taking the local wind direction of each glacier into account one will often find that the glacier is orientated perfectly opposite the local wind direction. In chapter 5.2.6 'The influence of wind on debris-covered glacier margins' the relation between wind direction and glacier orientation will be further analysed.

In the more extensively glaciated areas of the archipelago relatively large sums of precipitation have allowed for the formation of extensive glaciers that have effectively buried many peaks and headwalls. Snow distribution by wind presumably pays a role here as well, but it is les recognisable in the form of glacier orientation.

2.3. Glaciation extent and distribution over the archipelago

More than 60% of the land area of Svalbard is covered by glaciers (Hagen et al., 1993, Hagen et al., 2003). On Spitsbergen, the main island of the archipelago, glaciation is especially extensive along the coasts. While both the east and west coast receive high amounts of precipitation, the central part of Spitsbergen is relatively arid and is dominated by small glaciers, that are heavily reliant on snow redistribution by wind (Humlum, 2002).

In Nordenskiöld Land, the central part of Spitsbergen highlighted in this study and the area where all fieldwork has taken place, the glacial landscape is characterised by cirque and valley glaciers of up to 5 km^2 (Etzelmüller et al., 2000), although larger glaciers can be found away from the centre of Nordenskiöld Land. As is seen all over Svalbard, here too the glaciers tend to get larger towards the coast, but are still restricted to the valley and cirque types. The great glacier caps of Svalbard are mostly found on the east coast, with the ice-covered island Nordaustlandet (Fig 1.1) as the ultimate example. Longyearbreen and the other glaciers in Nordenskiöld Land are relatively small and thin, resulting in an abundance of glaciers with a basal temperature below the pressure melting point. Largely or entirely cold-based conditions are common for small (<5km long) land-terminating valley glaciers all over Svalbard. Although this is unlikely to have been the case at their LIA maxima, when the glaciers were substantially thicker and covered more ground (Hodgkins et al., 1999, Stuart et al., 2003, Hambrey et al., 2005, Midgley et al., 2013, Lovell et al., 2015). The surface of today's small glaciers is even and relatively crevasse-free. Drainage takes place supraglacially and englacially. Large englacial meltwater channels reopen and reshape each year. The channels are found in all shapes and sizes. Some form spaces and structures much like an ice-made cathedral. The local tourism industry even offers ice-caving trips into some of the more easily accessible meltwater channels in Longyearbreen and Larsbreen.

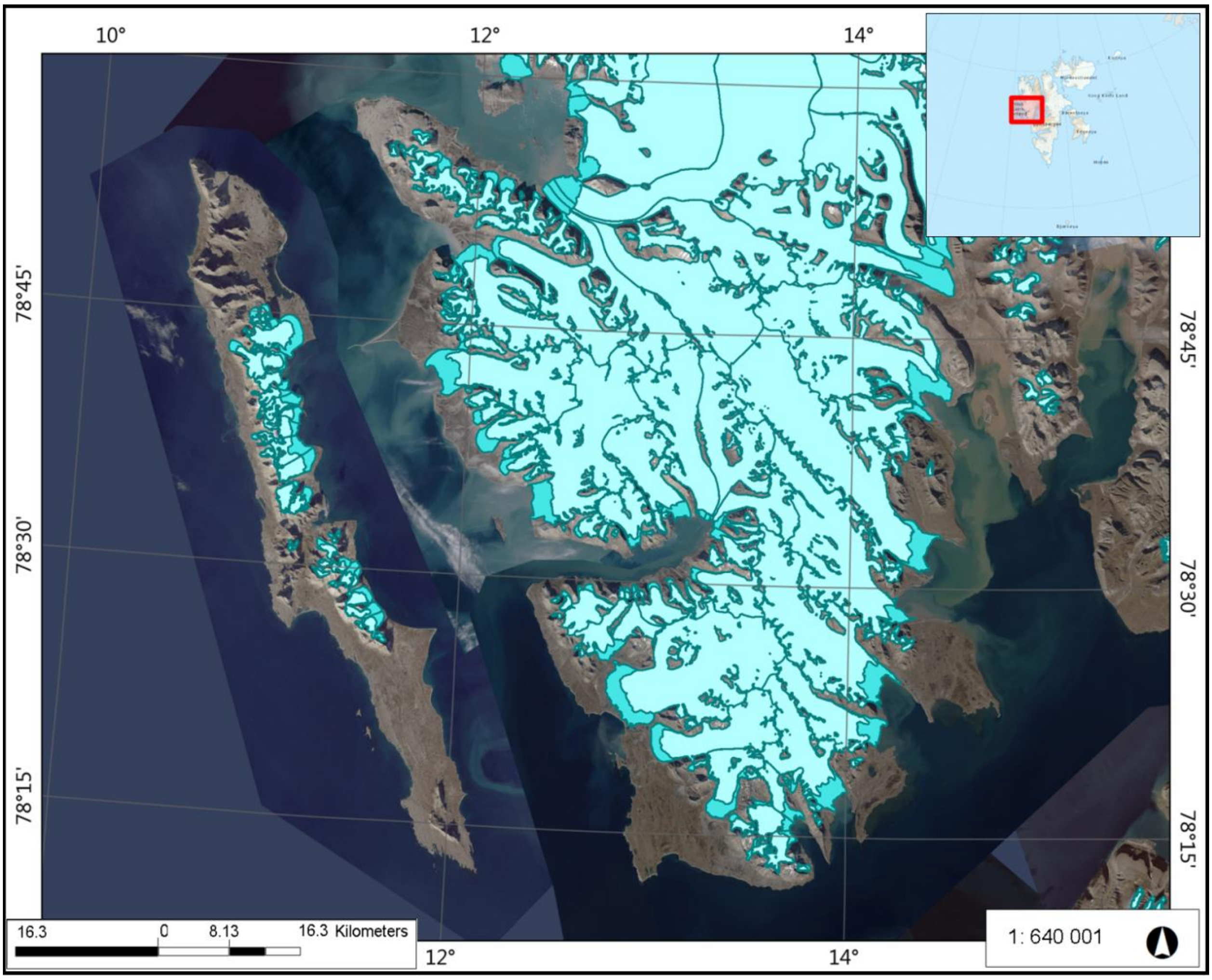

Figure 2.4: This map zooms into Prins Karls Forland and Oscar II Land, as an example of a coastal area of Spitsbergen. Glaciers on the coast of Prins Karls Forland and Oscar II Land are studied in this thesis along with glaciers on the coast and inland area of Nordenskiöld Land. The light blue shows the glaciated area in the period 2001 to 2010. The darker blue displays the glacier extent between 1936 and 1972 (data from Svalbard Kartet, Norwegian Polar Institute).

Many glaciers in Svalbard are known or inferred to be of surge-type, including small valley glaciers, tidewater glaciers, and land-terminating outlet glaciers, with varying thermal regimes from cold-based to polythermal (Dowdeswell et al., 1991, Hagen et al., 1993, Dowdeswell et al., 1995, Jiskoot et al., 2000, Sund et al., 2009, Lovell et al., 2015, Sevestre et al., 2015). The surges are characterized by a long quiescent phase followed by a relatively slow surge compared to other regions (Dowdeswell et al., 1991, Sund et al., 2009). The active phase typically takes several years, while the quiescent phase lasts anywhere from 30 to more than 150 years (Dowdeswell et al., 1991, Hagen et al., 1993, Murray et al., 2003a, Murray et al., 2003b, Sund and Eiken, 2004). At the fieldwork site of this study, Longyearbreen, there is no evidence for surging (Etzelmüller et al., 2000, Humlum et al., 2005) and currently the glacier is entirely cold-based (Sevestre et al., 2015).

It is clear that most glaciers experience a net mass loss since the termination of the LIA around 1920. Today’s glaciers are not in equilibrium with the present climate (Hagen et al., 2003). Fresh trimlines can often be seen 10-50 m above the modern glacier surfaces, indicating a corresponding amount of ice thinning. Most glaciers have shown a pattern of retreat (figure 2.4) and/or thinning. These observations are consistent with what is found elsewhere in the High Arctic (Walsh et al., 2012, Koch et al., 2014).

But using aerial images to determine glacial extent can be deceiving. Frontal glacier thinning is often accompanied by debris meltout and the subsequent formation of a debris cover on top of the glacial ice. Without detailed observation and the field based skills necessary to correctly interpret areal images, glacier thinning can easily be mistaken for glacier retreat. Add to that the fact that debris-covered glacier fronts have a history of being misinterpreted as ice-cored moraines (Ziaja, 2001, Humlum and Ziaja, 2002, Lønne and Lyså, 2005, Lukas et al., 2007) and the retreat-or-thinning confusion is complete. This study will focus on the ice that can be hidden underneath a layer of debris in these marginal glacier zones. It is the loss of glacial ice mass that has our interest when talking about glacier retreat and glacier thinning. When glacier retreat is mentioned in this study it will point to actual retreat: the loss of ice mass resulting in a smaller area being covered by glacier ice.

2.4. Extent and distribution of debris-covered glacier ice

Debris-covered glacier fronts are a common sight for most inhabitants of Spitsbergen. Glaciers near the main town in Svalbard, Longyearbyen, are all debris-covered to some degree. Examples are Longyearbreen and Larsbreen, both of which sit at the head of Longyeardalen, not more than half an hour walk from the outskirts of the town. It is therefore not surprising that these, and other nearby glaciers have been the object of previous studies (Hodgkins et al., 1999, Etzelmüller et al., 2000, Sletten et al., 2001, Lukas et al., 2005, Lukas et al., 2007). What this study may ad is a very valid question, and an easily explained one.

Debris cover on glaciers has been studied in Svalbard before, but previous studies have, although providing very valuable in-depth information on certain glaciers, failed to look at the bigger picture. There was, until this small pilot study, no work that incorporates a large number of glaciers to study the extent and distribution of debris-covered glacier ice on Svalbard. By mapping the debris-covered zones of 164 glaciers, in the area of Nordenskiöld Land and north of Nordenskiöld Land along the coasts of Prins Karls Forland and Oscar II Land, this study makes a start at describing the extent and distribution of (partially) debris-covered glaciers in Svalbard.

3. Theoretic background

3.1. Bigger picture: Holocene history of Svalbard and Longyearbreen

Svalbard can rightfully be called a high Arctic Archipelago. More than 60% of the land area of Svalbard is covered by glaciers (Hagen et al., 1993, Hagen et al., 2003) and the mean annual air temperature (MAAT) lies well below zero. The islands have however not always looked like today. In recent history, during the Holocene, both warmer and colder times have characterised this arctic zone. The Holocene can globally be divided into the three sections early, middle and late. Following the recommendations of Walker et al. (2012) the time span that each section covers are as follows: The Early Holocene starts at 11 700 yr. BP and lasts until 8200 yr. BP, the Mid Holocene lasts from 8200 yr. BP until 4200 yr. BP and the Late Holocene ranges from then up until the present. Usually work published before the publication of Walker et al. (2012) has approximately followed this division when discussing early-mid- and late-Holocene without specifying dates. But one must realize that the range in which the time-boundaries of the Holocene have been placed can be very large. The same issue of Quaternary International dealing with the Middle Holocene Archaeology of South America for example placed the beginning of the Middle Holocene in a range from 8 to 6 ka BP, while the end of the Middle Holocene varied between 5 and 2.5 ka BP (Hoguin and Restifo, 2012, Walker et al., 2012).

Even with these rather vague time-borders, sediment cores or other material, previous research does reveal a clear pattern of Svalbard's climate throughout the Holocene, since research has commonly provided dates of certain climatic events based on the dating of fossils. Based on marine data Salvigsen (2002) showed that the Svalbard region experienced warmer conditions during the early and mid- Holocene, compared with the present-day climate. At that time, most glaciers in Svalbard probably were smaller than at present or even absent. The warm period described by this research is based on the dating of shells that favour a warm climate. Radiocarbon dating of these shells reveals that a warm period took place at Svalbard from 9400 yr. BP to at least 5300 yr. BP. This is in line with the findings of a decade earlier. Back then five species of guide fossils indicative of the warm Holocene where dated from before 9500 yr. BP to about 3500 yr. BP. Based on lacustrine evidence from western Spitsbergen and a marine sediment core in Billefjorden, Svendsen and Mangerud (1997) also demonstrated a warm early- and mid-Holocene climate. They showed that that in Western Spitsbergen at Linnévatnet there were no glaciers in the catchment from about 10000 to 4400 yr. BP. It is only after this time that today's glacier started to form (Svendsen and Mangerud, 1997, Solomina et al., 2015). A climatic optimum seems to have taken place around the shift from early to mid-

Holocene (Salvigsen et al., 1992, Salvigsen, 2002). Overall the glaciated area on Svalbard was greatly reduced during the early- and mid-Holocene and many of the glaciers present today did not exist during that time (Svendsen and Mangerud, 1997).

At Linnédalen the long-lasting warm times were followed by a period of renewed glacier growth from 4400 to 4000 yr. BP (Svendsen and Mangerud, 1997). As Svendsen and Mangerud (1997) concluded, no glaciers existed in the catchment area of Linnévatnet throughout much of the Holocene and the present glacier Linnébreen was formed some 4000-5000 years ago. This is relatively consistent with the warm period ending around 3500 yr. BP, as was found by Salvigsen et al. (1992). Svendsen and Mangerud (1997) further showed that the present tide water glaciers started to form 3000-4000 years ago, and, according to a sediment core from Billefjorden, did not reach the seashore until 2000-3000 years ago. Based on Linnévatnet sediments, glacial maxima took place around 2800-2900 yr. BP, 2400-2500 yr. BP, 1500-1600 yr. BP and during the 'Little Ice Age', with the maximum extent of advances occurring in the 19th century (Svendsen and Mangerud, 1997). Subsequent research has usually confirmed this pattern (Solomina et al., 2015). The late Holocene thus marks a switch to colder conditions.

The cold of the late-Holocene most likely came in steps: periods of strong advanced were interchanged with slightly warmer times. Humlum et al. (2005) argued for an ice-free period of at least 800 years at the lower 2 km of Longyearbreen before a subsequent advance at ca 1100 yr. BP. During this less glaciated time at Longyearbreen Linnébreen experienced a retreat. This retreat is dated to about 1600 yr. BP (Reusche et al., 2014). It is important to realize that although the general trend in Svalbard follows a warm early- and mid-Holocene followed by a cold late-Holocene, these overall trends have all been punctuated by both warm and cold spells.

Longyearbreen presumably followed the large-scale temperature pattern in Svalbard, with a relatively warm early- and mid-Holocene and a cold late-Holocene marked by glacier growth. The Holocene history of Longyearbreen has, unlike Linnébreen, not been studied in great detail. However, as part of a master's thesis a basic O-18 isotope study was performed at Longyearbreen and although uncertainty in the dating was considerable it was concluded that Longyearbreen most likely formed about 4000 years ago (Bringedal, 2004). This is in line with the large-scale climate trend that occurred in the whole of Svalbard and places the time of Longyearbreen's formation near the shift from mid- to late-Holocene. The earliest recorded advance of the glacier took place around 1100 yr. BC. This advance was recorded by Humlum et al. (2005) after finding a subglacial exposure of undisturbed palaeosoil and vegetation below cold-based glacier ice of Longyearbreen. This time of glacial advance is in line with the glacial maxima recorded at Linnébreen around 2800-2900 yr. BP. Humlum's find below Longyearbreen showed more than just this glacial advance. Dating demonstrated that the sampling site, which is located about 2 km upstream from the present terminus of Longyearbreen, was ice-free for at least 800 years and possibly much longer prior to the advance. The retreat around 1600 yr. BC at Linnébreen (Reusche et al., 2014) makes it likely that Longyearbreen too retreated around this time. This means that the advance recorded by Humlum et al. (2005) presumably was a re-advance. The isotope record of Longyearbreen supports the notion that the advance at 1100 yr. BC was a re-advance. Bringedal (2004)showed that Longyearbreen's isotope record is characterized by many peaks and falls, indicating multiple climatic highs and lows. The advance of Longyearbreen can be assumed to represents a normal dynamic response to changes in air temperature, precipitation and changes in prevailing wind and the amount of drifting snow, for there is no morphological or structural evidence suggesting past surge behaviour and subglacial in situ vegetation excludes a surge for the last 1000 cal. yr. BP (Humlum et al., 2005). Much like most other glaciers in Svalbard Longyearbreen probably reached its maximal extent in relative recent times, during the Little Ice Age. During this period the coldest conditions of the late-Holocene occurred (Alley et al., 2009, Kaufman et al., 2009). In Svalbard the Little Ice Age had its glacial maximum in the 19th century (Svendsen and Mangerud, 1997).

3.2. Formation of debris-covered glacier margins

Debris-covered glacier margins have often been termed "ice-cored moraines" while in fact consisting of a zone of glacier ice, covered by a layer of supraglacial debris between 0.1 and 4 m thick, which retards the melting of underlying glacier ice (Etzelmüller et al., 2000, Lukas et al., 2005, Lønne and Lyså, 2005, Lukas et al., 2007). These features are not isolated pockets of buried ice that occur within larger bodies of sediment, as is implied by the term ice-cored moraines and misinterpretation can easily lead to a faulty assessment of glacier retreat (Lønne and Lyså, 2005, Lukas et al., 2007).

Debris-covered glacier margins are in essence ice-cored moraines in the making and therefor share the same formation process. A debris-covered glacier margin is a glacier margin hidden from view, buried under a layer of debris. Underneath this layer the glacier is intact and not detached from a coherent and intact body of clean glacier ice up-glacier (Lukas et al., 2007). Further melt of the glacial ice may detach the debris-covered ice form the glacier, thus forming an ice-cored moraine.

Debris covers are created as debris is transported to the glaciers snout. Since cold-based or largely cold-based glaciers hardly erode their bed, the building of a debris cover relies mainly on external debris input; present glacier erosion is less important than periglacial processes (Etzelmüller et al., 2000, Lukas et al., 2005). But even without glacially eroded debris large debris covers can develop. This is possible because polythermal/cold valley and cirque glaciers generally have high supraglacial material input from adjacent slopes (Etzelmüller et al., 2000). Debris from these slopes reaches the glacier in the form of rockfall events, snow-avalanches and occasionally a debris flow. Rockfall is especially common at the backwall of glaciers, while avalanches and debris flows occur along the whole length of the surrounding slopes. Rockfall and debris flows mainly take place during late spring and summer. Snow avalanches are mainly a winter and early spring phenomenon that contribute to debris build-up as a result of debris meltout during the summer (figure 3.1).

Debris that falls onto the ice of a (largely) cold-based glacier can be transported in different manners. First of all, the debris can roll over the ice surface as long as the momentum and surface slope angle permits. After its tumble the debris has three basic options left that can, combined or on their own, transport the material down the glacier.

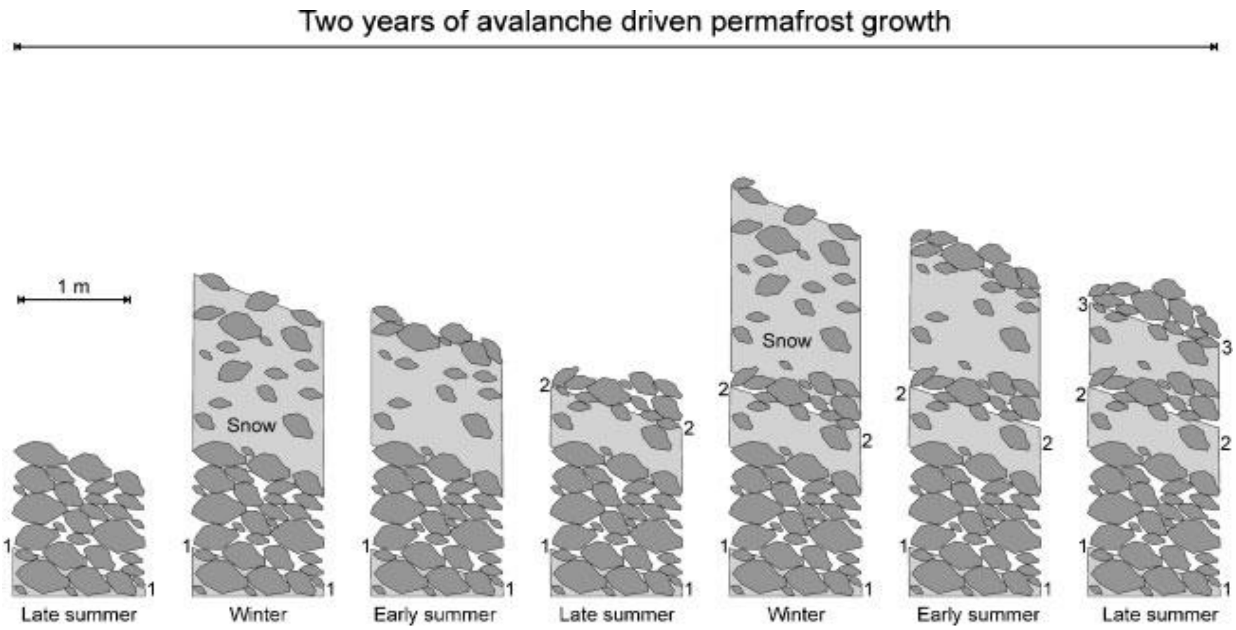

Figure 3.1: Model of avalanche derived build-up of debris- and ice-complex (reproduced from Humlum et al. (2007)). Avalanches contribute to both the accumulation of debris and ice, as the remnant snow will slowly turn to ice. A yearly or otherwise regular occurrence of avalanching can result in the alternating layer of debris and ice.

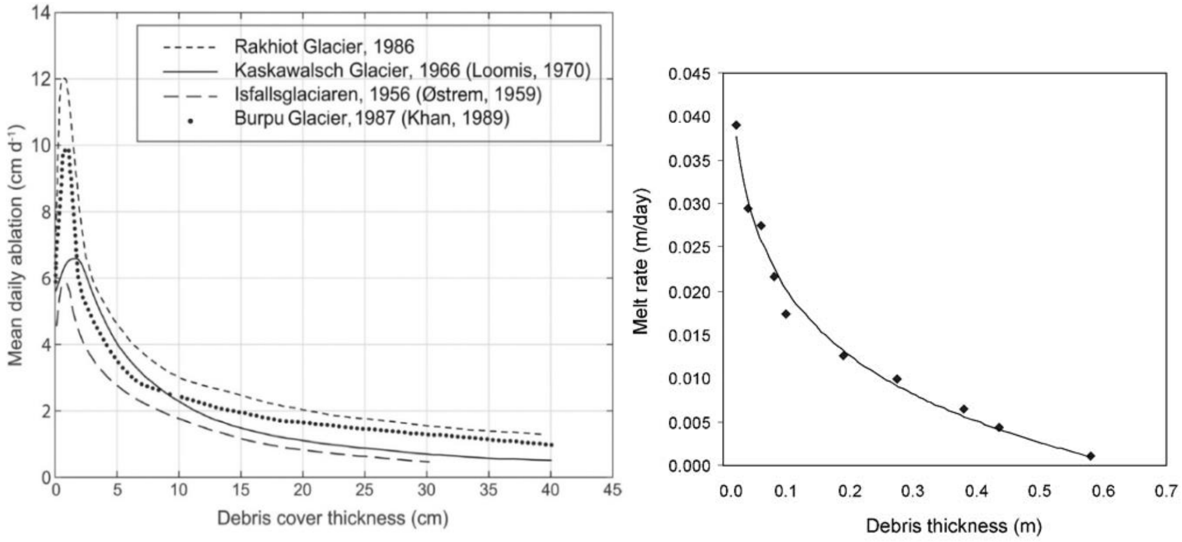

Figure 3.2: Relationship between debris thickness and ablation rate. The figure on the left is taken from Mattson and others (1993). The right-hand plot is reproduced from Lukas et al. (2005) and shows the melt rate underneath glacial debris at Larsbreen (Central Spitsbergen) from July 9 to 20, 2002.

A relatively slow form of transport is for debris to simply remain on top of the ice and be transported sitting on the deforming ice surface. This process can, on its own, transport sediment along the glacier. Since cold-based/polythermal glaciers are largely frozen to their bed, sliding rates will be minimal. Deformation of cold ice is relatively slow, partly due to the lack of liquid water within the ice. If this is the only transport option and great amounts of debris are received from a slope above, the debris will accumulate where the ice-angle no longer permits the rolling and sliding of debris, thus forming an ice-cored deposit. As a result large lateral moraines can form on valley sides with extensive rock walls (Benn et al., 2003).

Alternatively, debris can, if it comes to a stop in the accumulation zone of a glacier, be buried in snow which subsequently transforms to ice. Other ways for material to become incorporated into glacial ice is to melt down into the ice or snowpack or fall into a crevasse. One way or another, the debris finds itself trapped inside the glacial ice and will move with it as a consequence of internal ice deformation. This way of transport too is not the quickest, but given time substantial amounts of debris can be transported down-glacier in this manner.

The last and by far most effective way of transport occurs when relatively small debris finds its way to a meltwater channel. This can happen via one of the two before mentioned routes. A meltwater channel can for example erode/melt the ice containing debris. Off course a rockfall or avalanche can also directly reach a meltwater channel, or a crevasse connected to a meltwater channel. When debris is being transported by water it can quickly make it down all the way to the glacier snout. There it can simply continue with the stream downslope, away from the glacier, but if the debris gets stopped, for example by freezing into the glacial ice, in the frontal part of the glacier it can become part of the debris-covered glacier front. The debris is then added to the debris layer or ice surface directly via shallow supraglacial streams or deposited inside the meltwater channel to form an englacial sediment/debris layer. When the ice on top of this debris melts, the debris layer is exposed and will thicken as more ice melts with time.

Once formed, debris-covered ice takes a long time to disappear. Supraglacial debris covers, with thicknesses exceeding just a few centimetres, lead to a considerable reduction in melt rates (Mattson et al., 1993, Evatt et al., 2015) (figure 3.2). Debris-covered ice is clearly slower to respond to climate warming (Scherler et al., 2011). While the debris-free part of a glacier will lose mass and thus thin, the debris-covered parts will experience little melting. This process results in glaciers that are relatively thin and/or small compared to the size of their debris-covered margins or ice-cored moraines.

3.3. Deformation processes within and underneath debris covers

Most deformation processes on and in a glacial debris cover are, in one way or another, related to the glacial ice underneath. Stresses and strains that act upon the ice naturally affect the material that covers the ice. Like any mass on an inclined surface, glacial ice is subject of a gravitational downwards force. Many small arctic glaciers in Svalbard are largely frozen to their bed (Ingólfsson, n.d.), which means that this force cannot or hardly be transferred to a downward motion. The resultant stresses may cause internal deformation in the form of bending and/or thrusting of the ice.

Additionally, ice has a few properties of its own that interfere with a glacial debris cover. Perhaps unsurprisingly, melt of both snow and ice can have great impact. Englacial air-spaces like meltwater channels can collapse as the ice (partially) melts or degrades. On debris-covered ice this results in not only a collapse of the thinned ice-roof, but also of the debris on top. Due to the differences in permeability of dense glacial ice and its debris cover the border between debris and ice is, when at 0 °C, inherently unstable. When a saturated layer cannot drain water sufficiently rapid pore pressure increases, which can result in slumping and collapses of channels sides.

Not all deformation processes that act upon glacial debris rely on the ice-core underneath the debris. Rivers, fed by meltwater from further up-glacier, affect the ice and debris down-glacier. This happens through physical and thermal erosion. Both erosional processes shape the glacial ice and to a smaller extent the debris layer in and on top of the ice. This process too can cause channel-roof collapses as a meandering stream may cut sideways far enough for the former channel side to become a roof and for that roof to become unsupported.

And last but not least, there is the common transport of debris and sediment by rivers. Like any river, the forceful spring meltwater streams of a glacier can carry sediment and sometimes even small clasts along their stream. Although this process may seem less significant as it does not immediately displace a large part of the debris layer, it can, over time, definitely deform the glacial debris cover.

3.4. Surging and debris-covered glacier termini in Svalbard

A surge-type glacier experiences quasi-cyclic flow that alternates between long periods of slow flow (the quiescent phase) and shorter periods of flow that is typically 10–1000 times faster (the surge phase) (Murray et al., 2003b). The glacier switches between fast and slow flow despite only small changes in driving stress. In other words, these surging events are not directly forced by climatic shifts but by profound changes in processes and conditions beneath the glacier (Murray et al., 2003b). Less than 1% of Earth's glaciers are believed to surge (Jiskoot et al., 2000). Svalbard however contains one of the highest proportions of surge-type glaciers in the world (Sevestre et al., 2015), with the estimates of glaciers classified as surge-type ranging from 13 % to 90 % (Hagen et al., 1993, Jiskoot et al., 2000). Both the Central and Western areas of Svalbard are, like the rest of the archipelago, home to surging glaciers.

Surging has a devastating influence on a glacier's terminal zone and occurs on all types of glaciers, from small inland ones to large calving, tidewater glaciers. The rapid advance of a surge does not only cover the former proglacial area with ice, in doing so the sediments it advances over are deformed and the former glacier front disintegrated (Larsen et al., 2006). It does not take great imagination to see what this would do to a pre-existing supraglacial debris layer.

It is perhaps no surprise that a phenomenon as widespread as this raises a little curiosity. Could surging be responsible for the appearance of certain glacial termini studied in this thesis? Could this perhaps even be a deciding factor in the occurrence of a debris-covered or -free glacier front? Later chapters in this thesis will make a small start at answering these questions.

4. Methods: Glacier margins along the West Coast of Svalbard and in Central Spitsbergen

4.1 Deciding how much of an area is ice-cored

Mapping the debris-covered ice content from an aerial photograph relies heavily on field experience, making it possible to recognise the presence and degradation state of buried ice. Even so it is important to pin down which principles are used in the present study to determine if glacial ice is present underneath the debris. This makes it possible to reproduce the found results and insures that each aerial photograph is analysed in the same way. For the visual analysis, a set of geomorphological indicators was used to recognise the presence and degradation state of debris-covered glacier ice. This list of indicators is purely experience and fieldwork based. The main features indicating the presence of ice hidden below surface debris are the following:

- Deep cut meandering channels, much like supra-glacial channels (figure 4.1).
- Channels do not change course when entering the debris-covered zone.
- Lake edges are often sharp and marked by slumping scars. Lakes are shallow and superimposed on the (buried) glacial ice.
- Slumping evidence is visible in the form of sharp edges and cracks in the debris layer, exposed ice, slump-fans or terrace like steps (figure 4.2).
- Convex debris-covered glacier front, shaped by glacial ice (figure 5.5 of section 5 Results).
- Large scale ripples in the debris layer caused by ice movement or deformation of the debris layer itself (figure 4.3).
- Terminal zone shows few traces of deglaciation (lineations, eskers, lakes). Though crevasse ridges and eskers may exist supraglacially.
- Extension cracks and even crevasses can occur in the debris layer (figure 4.2).

The typical features to look for if the glacier is suspected to be partially debris-covered are:

- Gradual transition from debris-free to debris-covered ice (at lateral moraines and glacier margin). The glacier margin may form a steep edge, as is common for cold-based glaciers, but this marginal zone and thus the edge is debris-covered (figure 5.5 of section 5 Results).
- The suspected debris layer terminates at the glaciers LIA maximum extension. This indicates that the glacial margin is still at its maximum position and little or no retreat has taken place.

Figure 4.1: Meltwater channels in debris-covered glacial ice are much like ordinary supra-glacial meltwater channels; incised deep into the ice and meandering. Slumping and small rockfall events occur on the channel sides resulting in sharp exposed ice edges. Pictures taken at Longyearbreen's main western lateral channel in September 2014, Dani Roehnert and Renée Rookus for scale.

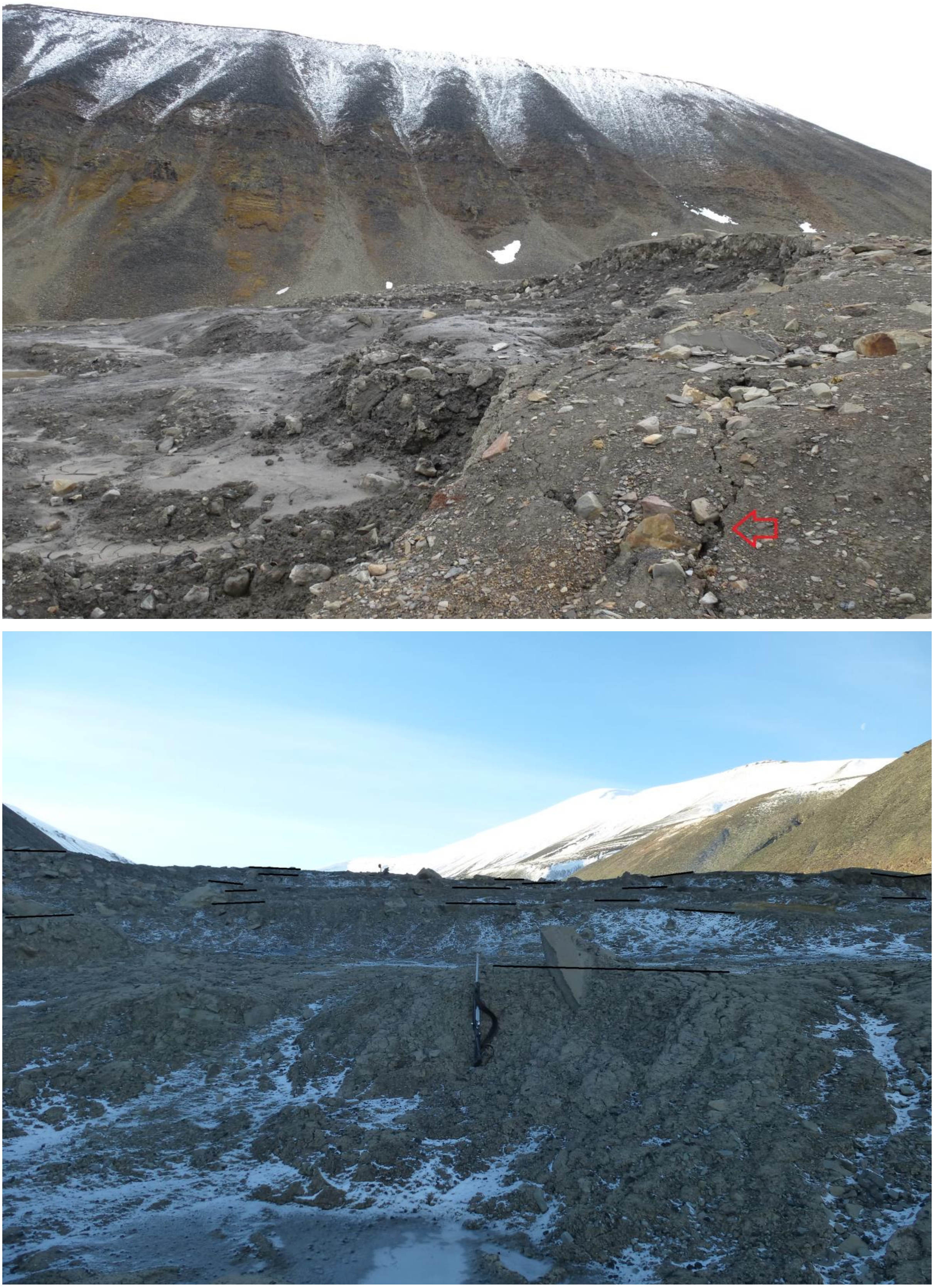

Figure 4.2: Two active slumping zones on Longyearbreen. Top picture shows a fresh slump fan. Mudflows from the previous rainy day are settling, while a new extension crack (highlighted by a red arrow) widens, ready to become the new slump-headwall. Bottom picture shows the characteristic terrace like steps often found in slumping zones. These steps can form either as a mudflow settles, forming a steep front, or as extension cracks widen and release new slumps, forming new headwalls. Pictures taken on Longyearbreen in September 2014.

Off course recognising the geomorphological features that indicate debris-covered ice in the field is different from detecting them on aerial photographs. Luckily the resolution of such photographs today has become incredibly high, making it possible to, with an eye trained by many days in the field, discover buried ice. Examples of how buried ice indicators look in aerial photographs are shown in figure 4.3.

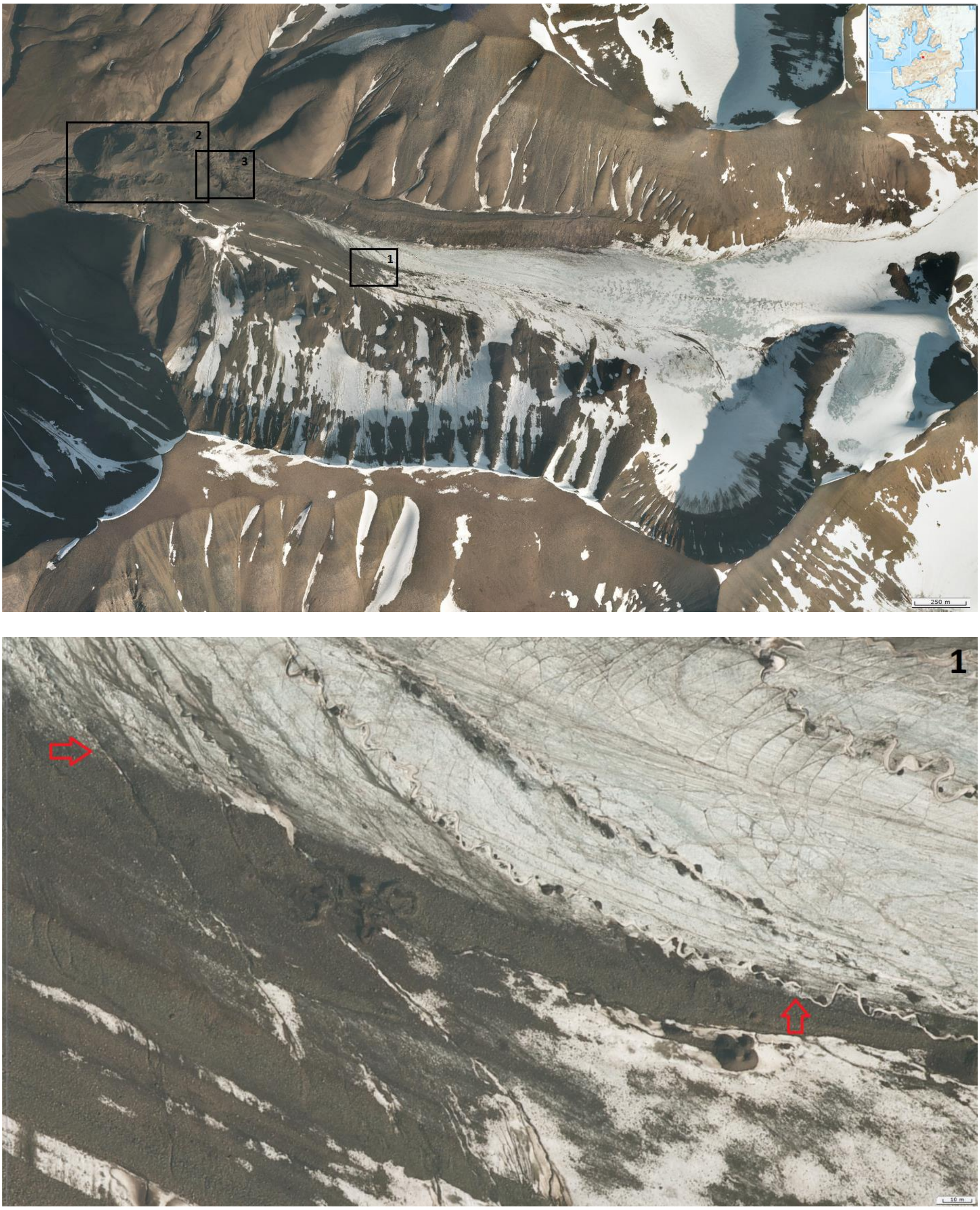

Figure 4.3: Blekumbreen, Nordenskiöld Land. Numbered zoomed in areas show the geomorphological features betraying Blekumbreen as a partially debris-covered glacier. Figure and description resumes on next page.

Figure 4.3 Continued: Zoomed in area 1 shows glacial meltwater channels that do not change at the shift from debris-covered to clean glacier ice. They are incised deep into the ice and meander. Zoomed in area 2 shows a red arrow pointing at an incised meandering stream. Thermal erosion has presumably lowered the stream and set its track as it cut through the ice masses. The black lines in this same picture indicate large scale ripples and bends in the debris layer caused by movement of the underlying ice. They may even be outcropping englacial debris bands (which naturally are also shaped by deformation of the ice in which they are incorporated). In zoomed in area 3 there are 5 arrows pointing towards the headwalls of relatively recent slumps.

Just like there are geomorphological features indicating the presence of debris-covered glacier ice, there are also signs to look for that tell one that no ice is present under the debris. The main features indicating that no ice is present below the debris are:

- Sharps edge (sometimes upstanding) of debris-free ice at the glacier margin.
- Glacier terminus is marked by streams following the rounded shape of the clean ice margin.
- Streams tend to be wide, shallow and braided.
- The clean ice has a convex form, but the debris is found in flat plains, often interrupted by braiding meltwater streams.
- Clear front moraine, disconnected from the clean glacier ice up-valley.
- Lake edges show no signs of slumping.

However, hardly any glacier terminal area in Svalbard is completely free of buried ice. Therefore, one should also look for signs of disconnected buried ice to distinguish between the ice-cored and ice-fee areas. The typical features to look for if the glacier is suspected to be free of debris cover are:

- The LIA moraine, disconnected from the glacier, marks the end of the deglaciated area. The moraine may show tension cracks and slumping due to dead ice meltout.
- The terminal zone of the glacier is dominated by a low and flat area between the glacier front and, if present, the LIA moraine. This (largely) deglaciated area may contain lineations, crevasse fill ridges, eskers and lakes. The deglaciated area itself is free of collapse features, but the area may contain remnant ice-cored mounts and/or moraines that are prone to slumping.

With the use of the above principles, detailed aerial photographs could be studied for ice-cored areas. This visual analyses relies heavily on fieldwork experience, making it possible to recognise the presence and degradation state of buried ice. More than three years of living in Svalbard, filled with field-work based courses and private trips combined with two months (spread out over more than a year) of field work on Longyearbreen have enabled the author to perform the aerial image analysis with some confidence. Without a field-based background, enabling one to read glacial landscapes, aerial image analysis as performed in this study would be impossible.

Based on the aerial photographs, the debris-covered parts of glaciers and their terminal zones have been divided into four classes. Ranging from intact ice core, meaning that the glacial ice is still present below the debris cover without any ice-free zones, to ice-free. With the use of colour coding the presence and degradation state of debris-covered glacial ice has been made visible on TopoSvalbard's topographic maps. This was done for 164 glaciers in the area of Nordenskiöld Land and north of Nordenskiöld Land along the coasts of Prins Karls Forland and Oscar II Land (figure 2.1 and 2.4.).

The first class, intact ice core (coloured light green on the glacier maps), only includes debris-covered ice that is in fact still part of the glacier. This means that the debris-covered ice is still connected to the intact body of clean glacier ice up-glacier and does not contain any ice-free zones. These light green areas are used to calculate the debris-covered glacier surface.

4.2 Length, area and orientation measurements

The online distance and area measurement tools on Norwegian Polar Institute's website TopoSvalbard were used to measure the projected length, width and area of intact ice-cored debris-covered zones, clean glacier ice and possible debris supplying areas like headwalls and side-slopes surrounding a glacier. Where slope angle was of great influence (headwalls and side-slopes), the surface area was calculated to provide a more accurate estimate of the actual area. Orientation was measured by projecting a line on the feature of interest followed by simply reading of the angle with a protractor. Measurements were made based on the most recent aerial images on TopoSvalbard (the same aerial images that were used to map the presence and degradation state of glacial ice).

4.2.1 Modelling debris-covered glacier surface against clean glacier size

The debris-covered percentage of a glacier is calculated as follows:

$$Debris\ covered\ glacier\ surface\ (\%) = \frac{Debris\ covered\ glacier\ surface}{Clean\ glacier\ surface + Debris\ covered\ glacier\ surface} * 100$$

Linear regression data analysis was used to calculate P and R values. However, a visual analysis showed that the obtained data is best represented by a non-linear function (figure 5.3 of result section). This meant that the data representing the x-value in the best-fitting model needed to be modified to straighten the logarithmic profile and allow for linear regression. This is done by taking the logarithm of 'Clean glacier surface' (the x-value) before the data is subjected to linear regression analysis. It is important to note that only the x-value needs to be modified to straighten the logarithmic profile presented by the data. Taking the logarithm of both x (Clean glacier surface (km^2)) and y (Debris-covered glacier surface (%)) would give a double, and erroneous, correction.

The automated regression analysis was checked for accuracy by performing a part of the analysis manually. This output was than compared to that of the automated regression analysis to check for any differences. In practise this meant that the residuals were calculated manually by subtracting the modelled debris-covered glacier surface (%) of each glacier from the measured debris-covered glacier surface (%) of the same glacier. The resultant model errors, in other words the residuals, proved to be identical to the output of the automated regression analysis and thus validates its results.

4.2.2 Calculating headwall and side-slope surface areas

Being able to see which slopes are likely to supply debris to a glacier by simply looking at an aerial photograph is a little more straightforward than mapping the hidden ice-content. Even so, this skill too relies on field experience. For the visual analysis, once again a set of geomorphological indicators is used to recognise the presence, runout-zone and boundaries of debris supplying slopes. The main features indicating the presence of slopes supplying debris to the glacier are:

- Avalanche trails, fans and headwalls above the glacier.
- Rockfall trails and fresh rock exposures indicating release zones above the glacier.
- Debris flow trails and fans above the glacier
- (Lines of) debris on the glacier surface, connected to probable avalanche/rockfall-prone site

Obvious candidates of debris transporting slopes are glacial headwalls and steep side-slopes, but nunataks can also provide debris. The occurrence of slope processes alone does not necessarily imply debris transport to the glacier. The debris transporting slopes must have a free flow path to the glacier ice. Obstructions like moraines, lateral streams or a V-shaped trap between glacier and slope can stop debris from reaching the glacier ice. Therefore, steep avalanche prone slopes may not be included in the debris supplying area if obstructed, unless there is clear evidence of debris supply making it across the moraine like wide and often occurring streams cutting straight through the moraine, opening the way for debris to flow to the glacier ice.

Since slopes prone to rockfall events are generally steep, the projected area that can be easily measured with TopoSvalbard online services differs greatly from the actual surface area. To correct for this the slope angle needs to be taken into account. This was done in six steps (measurements are illustrated in figure 4.4):

1. The projected area covered by debris supplying slopes was measured in TopoSvalbard.

2. The distance from the foot of the debris-supplying slope (at the glacier edge) to the top of the slope was measured in TopoSvalbard. Generally, this is done by three projected length lines within each projected slope area. But depending on how much the slope angle changes within the projected area less or more length lines can be used. The average of these projected lines is used for further calculation of average slope angle of headwall and side slopes.

3. The difference in height covered by each projected length line was read of from the map's contour lines the average height difference covered by the projected lines is used for further calculation of average slope angle of headwall and side slopes.

4. The average slope angle of each projected area is calculated by:

$$Slope\ angle = \tan^{-1} \frac{Height\ covered\ by\ projected\ line}{Length\ of\ projected\ line}$$

5. The surface area of each projected area is calculated by:

$$Surface\ area = \frac{Projected\ area}{\cos\ (Slope\ angle)}$$

6. The total debris supplying surface area is calculated by summing up all the independent debris-transporting slopes delivering debris to a glacier. This total surface area includes headwalls, side slopes and nunataks.

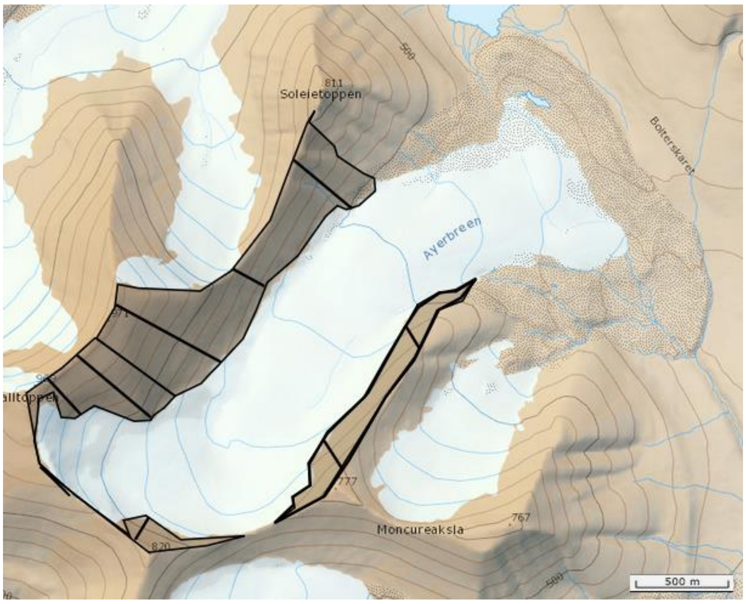

Figure 4.4: Example of area measurement of headwalls and side slopes in TopoSvalbard (North is upwards). Aerial photographs were used to decide which slopes are likely to supply debris to a glacier. The most current glacier outline was also recovered from the areal pictures and if necessary the glacier outline on the map was corrected.

4.2.3 Measuring wind direction by snow-drifts

Wind direction in Oscar II Land was measured by mapping multi-year snowdrifts on glaciers in the area. As wind-blown snow settles on the lee side of nunataks and mountain tops/ridges, a natural snow-blown 'wind-arrow' forms. The direction of 37 of these features has been measured and portrayed on aerial photographs such as the one shown in figure 5.12. These measured wind features were taken within four different area's in Oscar II Land, selected to insure spatially evenly distributed samples, that together are representative of the whole area. Only snow-drifts located on the higher parts of Oscar II Land's glaciers were measured, since wind features lower down on the glaciers would not be free of local topographical influences like valley-channelized winds.

To make sure that the general wind direction was measured only large snow drift forms have been included in this study. The natural snow drifts found on the ice-caps of Oscar II Land are up to a kilometre in length (figure 5.12). Including only large scale wind-features somewhat limits the number of studied features, but insures that the measured wind-direction is a wind direction that prevailed over multiple years.

5. Results: Glacier margins along the West Coast of Svalbard and in Central Spitsbergen

Deciding when a glacier has a debris-covered front and when to call it an ice-cored terminal zone can be somewhat arbitrary as nature will show all the different stages of ice-core meltout. To be classified as a glacier with a debris-covered terminus in this thesis, the terminal area must have a clearly developed ice core, often tongue shaped. Meltout can take place, but has (except from the occasional incised channel) not led to ice-core free areas. In other words, a solid body of glacier ice is still present below its debris cover and not detached from a coherent and intact body of clean glacier ice up-glacier.

To get an overview of all glaciers within the studied areas, this study does not only focus on the glaciers with a debris-covered front but instead tries to include all different types of glaciers and terminal zones. This is done to include all debris-covered parts of the glacier, without imposing limits for its location on the glacier surface. These debris-covered areas on a glacier are further referred to as areas with an intact ice core (coloured light green on the glacier maps), meaning that the glacial ice is still present below the debris cover and there are no ice-free zones. If debris-covered glacial ice has become patchy by meltout, and is therefore no longer connected to the intact body of clean glacier ice up-glacier, the ice core is no longer considered to be a part of the glacier. Instead it may be referred to as dead ice or a degraded ice core.

5.1 Small-scale study of glaciers within 20 km of Longyearbreen

The first step towards a better understanding of debris-covered glacier ice on Svalbard has been taken in the form of a small-scale study in Central Spitsbergen. This study provides basic data about glaciers and glacier margins in the form of orientation, length and width of each glacier and the debris-covered areas and/or moraines. The area was picked because of the great amount of debris-covered glacier fronts, known to the author by living in the area for three years. Remarkable within this area, in the heart of Central Spitsbergen, is the distinct glacial signature that is different from most of the island. Here, in the archipelago’s centre the glaciers are many, but small.

All 40 glaciers within a radius of 20 km from Longyearbreen have been studied, using the online mapping service of the Norwegian Polar Institute (NPI). The average debris-free (clean) glacier length within this small, central area of Spitsbergen is just 3 km (obtained from NPI's most recent maps). All of the 40 measured glaciers have either a debris-covered terminus or a heavily ice-cored terminal zone.

Twenty-seven of the 40 glaciers within 20 km of Longyearbreen have been defined as glaciers with a debris-covered glacier snout (table 1), meaning that the entire terminal area of the glacier is underlain by an intact ice core, not detached from a coherent and intact body of clean glacier ice up-glacier. This debris-covered terminus is generally substantial and adds on average 0.6 km to the glacier's debris-free length, but can contribute as much as 1.6 km. Which, considering that the mean total glacier length of these 27 glaciers is but 3.4 km, is significant.

The remaining 13 glaciers do have high ice core percentages in the terminal area, and there are even examples of glaciers that could be argued to have a small debris-covered front. The difference with the above mentioned 27 glaciers with a debris-covered front however is that the terminal zones of these 13 glaciers include areas, other than the occasional cut in channel, with less than 100% ice core within the outermost moraines. On average 19 % of the total glacier length is debris-covered within the 20 km radius around Longyearbreen.

Length measurements by themselves are of limited use, since it does not inform about the glacier's actual size. It is however a useful measure when checking if there is a basis for a more in depth study. The data recovered from this small-scale study is promising and paved the way for a more detailed study on a larger scale, covering not only the glaciers in Nordenskiöld Land but also the central west coast of Spitsbergen. Instead of length and width measurements this study relies on areal mapping to obtain more accurate results.

Table 1: A small scale study of glaciers within 20 km of Longyearbreen shows the widespread occurrence and length of debris-covered glacier fronts in the area.

Glaciers within 20km from Longyearbreen	Number of glaciers	Average Length (km)
Clean ice surface; all observed glaciers	40	3
Total glacier surface (debris-covered and clean); observed glaciers with a debris-covered front	27	3.4
Debris-covered terminus; observed glaciers with a debris-covered front	27	0.6

5.2 Large-scale study of glaciers along the West Coast of Svalbard and in Central Spitsbergen

Debris-covered ice is not limited to the frontal part of the glaciers. More often than not the sides of a glaciers hold great debris-covered areas or ice-cored lateral moraines and lines or patches of debris cover elsewhere on the glacier are not uncommon. To account for all the glacial ice covered by debris a broader study has been performed, mapping the extent of the ice core underneath debris. This large-scale study focuses on the glaciers in Nordenskiöld Land and north of Nordenskiöld Land along the coasts of Prins Karls Forland and Oscar II Land (figure 2.1 and 2.4. in the section 'Description of study areas') and shows some interesting trends as to the occurrence of debris-covered glacier snouts and the amount of buried ice in terminal areas that will be further described in the following sections.

The large-scale study can be divided into three steps:

1. Mapping the ice core of the terminal zones with the use of recent aerial photographs. The terminal zones of 164 glaciers in the area of Nordenskiöld Land and north of Nordenskiöld Land along the coasts of Prins Karls Forland and Oscar II Land (figure 2.1 and 2.4. in the section 'Description of study areas') have been mapped (the appendix holds all maps).
2. Measuring the surface areas of the debris-covered and clean intact glacier ice of 91 glaciers (table 1 and 2 of appendix), representative of all 164 glaciers in the study area.
3. Analysing the surface measurements of debris-covered and clean intact glacier ice and fitting a model to the data.

5.2.1 The influence of glacier size

The most distinct find of this study is the clear link between glacier size and the amount of ice-core in the terminal area of a glacier. Debris-covered terminal areas are mostly found on small (< 5 km in length) glaciers, rather than on their larger siblings (figure 5.1 a,b). The general trend that appeared from visual analysis of the 164 studied glaciers is as follows: the smaller the glacier the greater the amount of ice core in its terminal area relative to glacier size. To further analyse this pattern, 91 of the 164 previously mapped glaciers, representative of all 164 glaciers in the study area, where chosen for further analysis. Using the aerial photographs and mapping services provided by the Norwegian Polar Institute, the projected areas of the debris-covered intact glacial ice and clean intact glacier ice, were measured for each of the 91 glaciers.

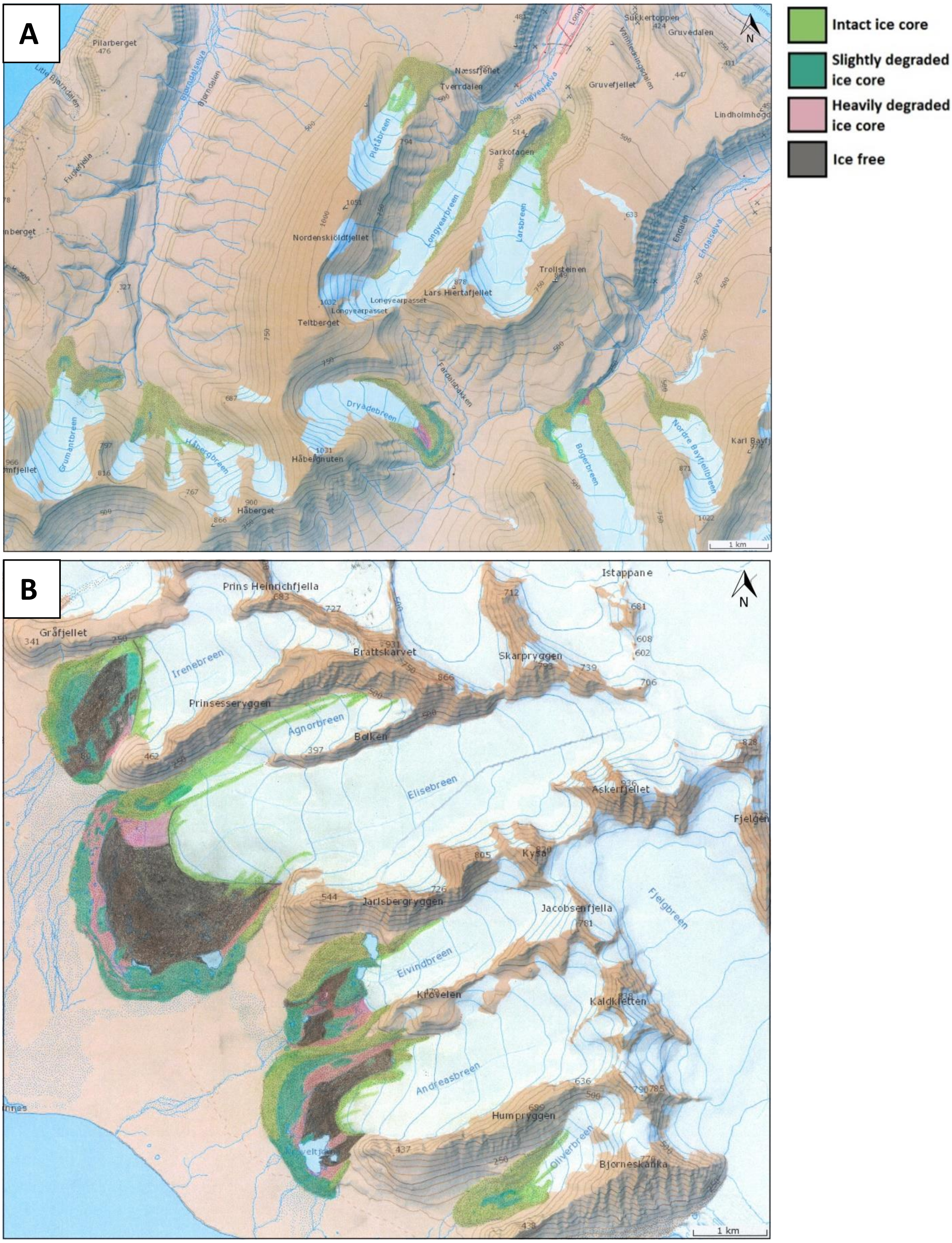

Figure 5.1: Large glaciers have relatively little ice core in the terminal zone, while the terminal area of smaller glaciers tends to be greatly ice-cored. A: Ice-core map of Longyearbreen area, central Spitsbergen, Nordenskiöld Land (for location see figure 2.1 and 2.4). B: Ice-core map of Elisebreen area, west coast of Spitsbergen, Oscar II Land (for location see figure 2.1 and 2.4). Intact ice core means the glacial ice is still present below the debris cover, the area is free from any ice-free areas; Slightly degraded ice core means the area is mostly ice-cored, but few ice-free areas do exist; Heavily degraded ice core means the area is mostly ice-free but contains patches of ice-cored debris. Colours can appear slightly darker or lighter due to transparency with the terrain of the underlying base map.

5.2.1.1 All 91 glaciers: Model fits and statistical analysis

Three of the 91 glaciers are considerably larger than the other 88, presenting a data gap in the observed population (figure 5.2). These are Drønbreen, Charlesbreen and Eidembreen measured at respectively 22, 27 and 136 km^2 of clean glacier ice, while the largest glaciers of the bulk of the measurements is about 14 km^2. Multiple models have been fitted to the total of 91 data points. The two models best resembling the sampled data are described in the present section. But as will become clear, models based on all 91 data points, including the outliers Drønbreen, Charlesbreen and Eidembreen, result in unrealistic scenarios because of the above-mentioned data gap.

The best model fit, including all data points, is based on a logarithmic function and gives a good representation of all glaciers but the three largest ones (figure 5.2). Regression analysis of the models gives an R-square of 0.78 and a P-value of $1.3*10^{-29}$. In other words, roughly 78 % of the data's variation can be explained by this model and the probability that the models equation does not explain the variation in debris-covered glacier surface percentage, i.e. that the fit is purely by chance, is as low as $1.3*10^{-27}$. Furthermore, the residuals average (i.e. the average error) of $-2.5*10^{-9}$ has the model's errors lined up nicely along the zero line. Statistically this is a well-fitting model. But when one looks towards the practical side of things it becomes apparent that something is off; the model predicts negative percentages of debris-covered glacier surface for all glaciers larger than 15.5 km^2. This model clearly needs many more data points of large glaciers in order to reliably predict the percentage of debris-covered glacier surface of such large glaciers.

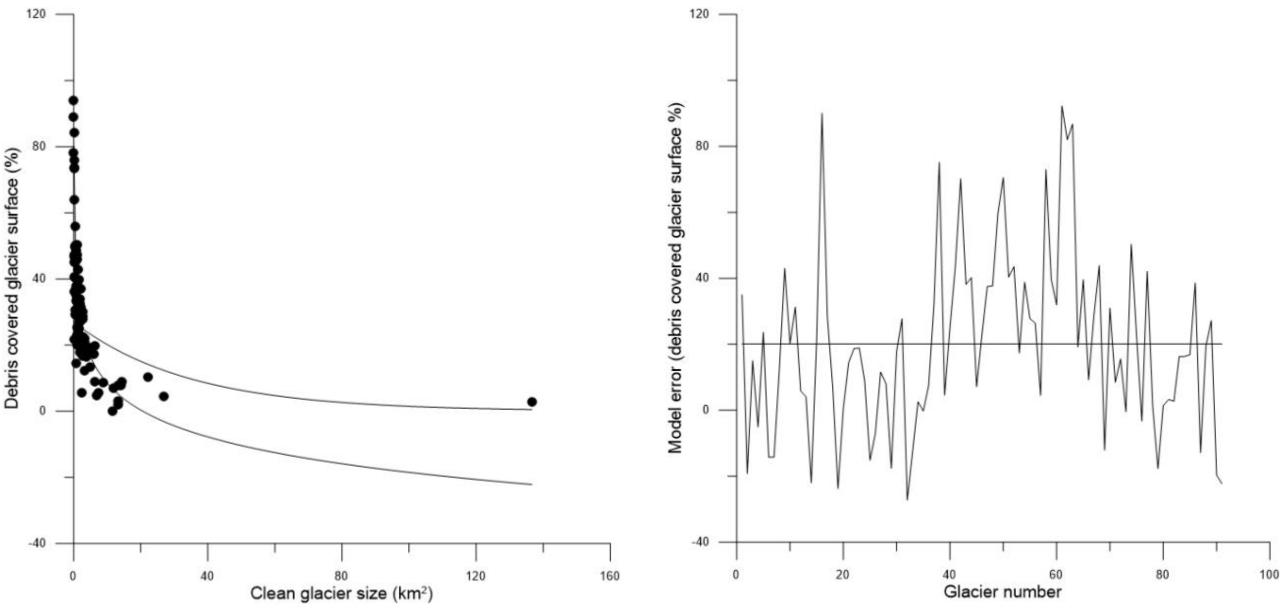

Figure 5.2: Right-hand graph: Plot of all data points and the corresponding exponential (upper) and logarithmic (lower) model fit. Left-hand graph: Plot of the residuals (errors) of the exponential model. The residuals are not centred at the 0 line, showing that the exponential fit is not representative.

The second-best model fit, based on an exponential function (figure 5.2), does not have the problem of predicting negative percentages of debris-covered surfaces. While there is not much wrong with the low P-value of $2.5*^{10-5}$, the R-square of 0.19 is far from promising. And analysis of the model's residuals showed an average error in the prediction of debris-covered glacier surface of 20 % (figure 5.2 right). Even by simply looking at the model's trend line and the data points it is supposed to fit one can see that the model is strongly biased towards the largest data points (the largest glaciers), resulting in a poor fit for the remaining data.

The data gap clearly makes it impossible for either model to accurately predict the percentage of debris-covered glacier surface for the entire range of glacier sizes it incorporates. It is therefore decided to exclude the three outliers from further data analyses.

5.2.1.2 Without the three outliers: Model fits and statistical analysis

The remaining 88 glaciers have been plotted and tested with multiple model fits. The two best fitting models are once again based on a logarithmic and an exponential function. These models where tested and subjected to error analysis, the result of which can be seen in figure 5.4.

The 95 % confidence interval is overall slightly narrower for the logarithmic model, although it is narrow for both models. Confidence intervals provide an estimated range of values calculated from a given set of data that is likely to include an unknown population parameter. Instead of estimating the parameter by a single value, an interval likely to include the parameter is given. The confidence level of 95 % means that the chance that the interval includes the correct model is 95 %. The narrow confidence intervals of the exponential and logarithmic model suggest that sufficient data has been collected to make conclusions based on the model.

While the model based on an exponential function, with an R-square of 0.61, a P-value of $1.3*10^{-19}$ and the models errors lined up close to zero, does an all right job of representing the data, the logarithmic function clearly is a better fit. Not only does the logarithmic function give an R-square as high as 0.79, it also greatly improves the P-value to the remarkably low number of $3.7*10^{-30}$. Meaning that the high correlation between clean glacier size and the glaciers debris-covered surface percentage is not likely to be due to chance. Purely looking at these statistics it is highly likely that the two variables are not only well correlated but also related; there seems to be a strong negative relationship between a glaciers debris-covered surface % and clean glacier size. This logarithmic model's reliability is further supported by the model's error graph (figure 5.4, bottom right), which shows the residuals to be centred perfectly around the zero line, as would be expected. Here too, the logarithmic model is an improvement over the exponential one, with both the error margins being smaller and a better centring at zero. Understandably the logarithmic model has been chosen as the most reliable and accurate fit. In which the debris-covered percentage of a glaciers surface (Y) is given by:

$$Y = -12.72 \ln x + 35.166$$

with x being the clean glacier surface of the glacier in km^2. However, the model is only accurate for glacier of up to 15 km^2 of clean ice surface and should not be used to predict the debris-covered glacier fraction of larger glaciers. This because the model as is will predict negative debris-covered percentages for glaciers larger than 15 km^2. For prediction purposes of large glaciers, the alternative (exponential) model could be useful. In this model the debris-covered percentage of a glaciers surface (Y) is given by:

$$Y = 43.95e^{-0.2106x}$$

with x being the clean glacier surface of the glacier in km^2. Although not as great a fit as the logarithmic model, the exponential model, with an R-square of 0.61, a P-value of $1.3*10^{-19}$ and the models errors lined up close to zero, does an all right job of representing the data. The great advantage of this model is its extended usability for large glaciers. It predicts likely, small but not negative, percentages of debris cover for glaciers larger than 15 km^2. Although it must be noted that this model has not been tested for larger glaciers, as the study area does not contain a sufficient amount of them to test the model.

The 88 measured glaciers in Nordenskiöld Land and along the coasts of Prins Karls Forland and Oscar II Land show a clear correlation, and probable relation, between the fraction of the glacier that is debris-covered and a glacier's clean size. This supports the before mentioned trend for small glaciers to produce large debris-covered glacier fronts, while large glaciers create relatively little ice-core in the terminal zone. The R-square (R^2=0.79) value of the best fitting model is remarkably high, considering the number of observations (n=88). This is underlined by the fact that this is a pilot study without any compensation for other factors impacting the many studied glaciers, and other factors, like climate, do differ. The large study area that covers a distance of 150 km (SE-NW) contains both relatively mild and wet coastal climates and colder and drier inland areas. Geology and topography also differ within the study area, but clearly not to an extent to greatly impact the correlation between clean glacier size and debris cover.

The strong negative trend between a glaciers clean size and its debris-covered fraction is not the only interesting discovery that can be retrieved from the data. The right-hand graphs of figure 5.4 also shows a trend in the distribution of the data points. In small glaciers, there is a large spread of debris-covered percentage, but with increasing glacier size the data points seem to get closer to each other. This may partially be due to the smaller amount of measured large glaciers in the study area. The more data points the more the maximum spread between the data points will be. However, the trend identified is not based on just a few data points at the end of spectrum, but is more substantial than that. While it is clear that there is a strong negative relation between clean glacier size and debris cover, the data also shows that the range of the debris-covered ice fractions is greatest for small (< 3 km^2) glaciers. For glaciers of roughly the same clean size (maximum size difference is 400 m^2), larger glaciers (> 3 km^2) differ in debris-covered fractions by a maximum of 10 %, while for smaller glaciers the range is up to 35 % (figure 5.3).

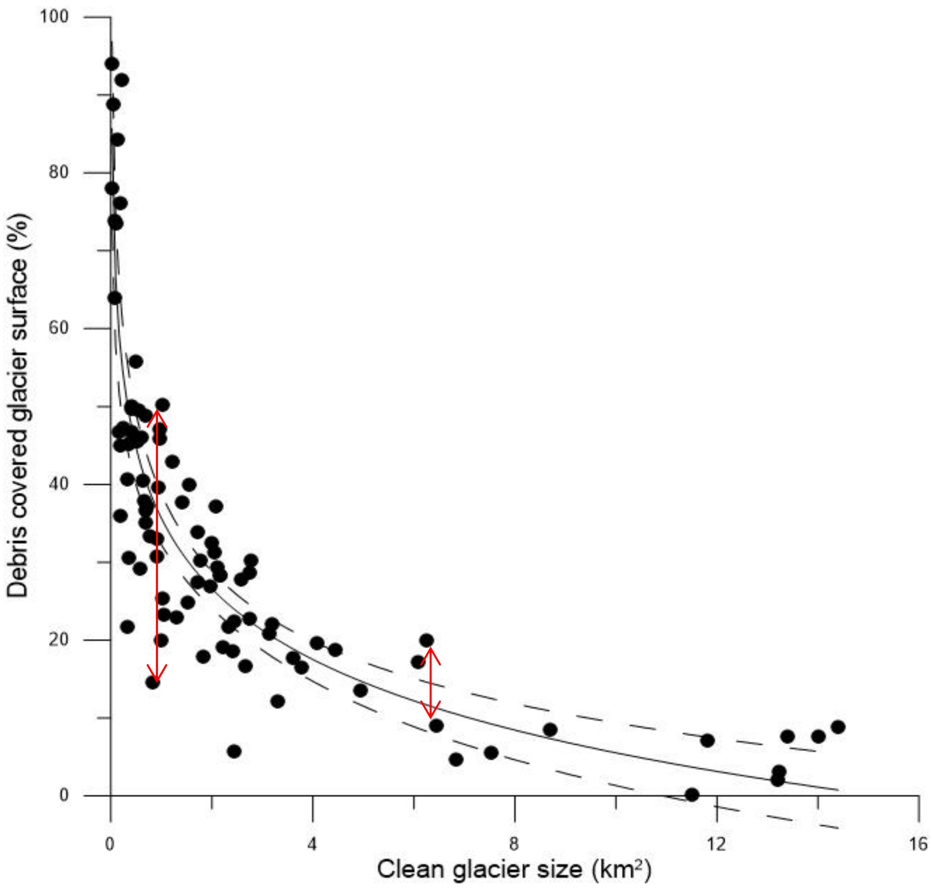

Figure 5.3: The red arrows show the range in debris-covered percentages for glaciers of roughly the same size. The range of debris-covered ice percentages is greatest for small glaciers. Larger glaciers (>3km^2) show a maximum range of 10 % debris cover, while for smaller glaciers the range is up to 35 %.

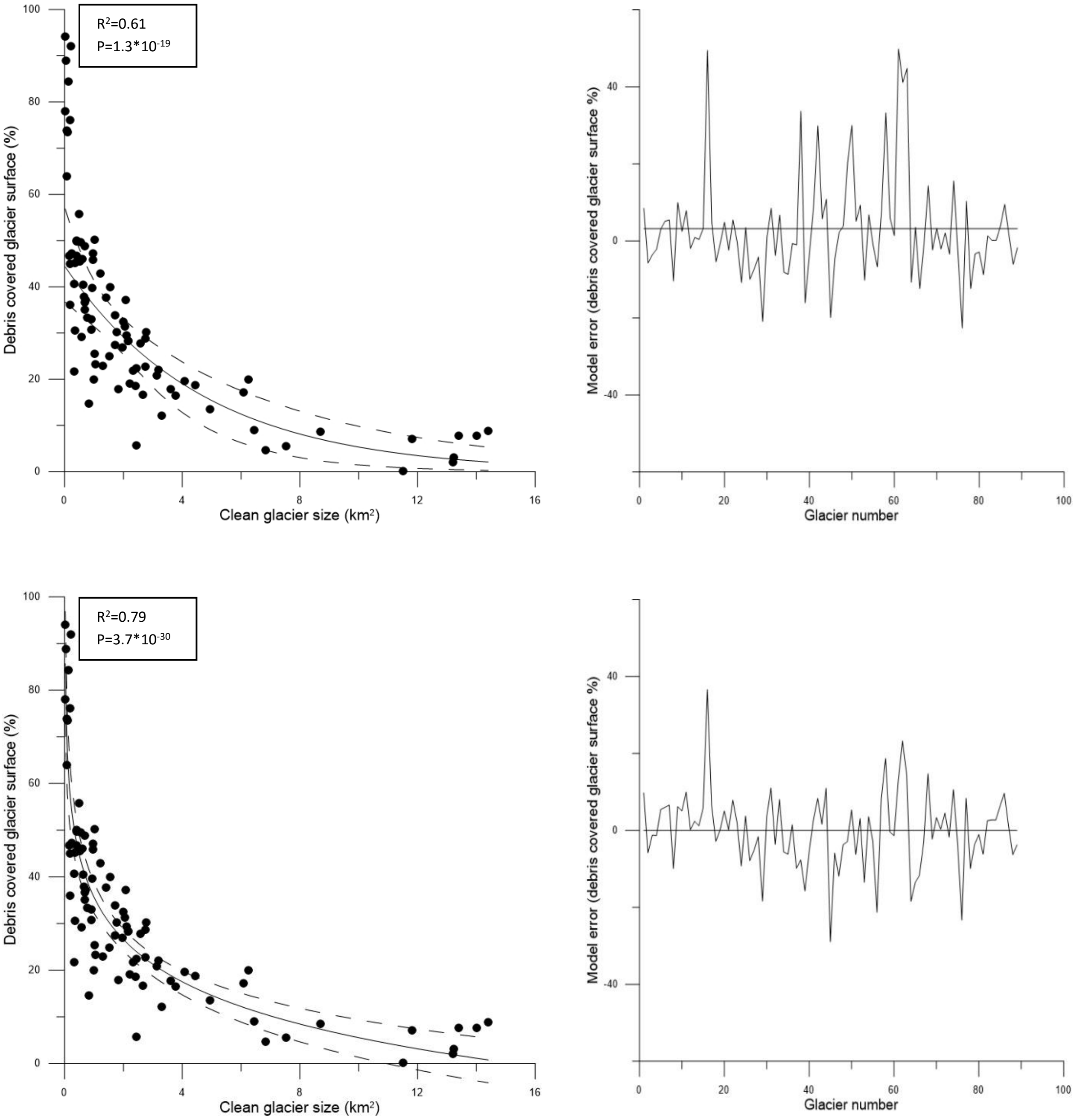

Figure 5.4: To the left the 88 glaciers are plotted and fitted with an exponential (upper graphs) and a logarithmic function (lower graphs). The dotted lines indicate 95 % the confidence interval. Meaning that there is a 95 % chance that the correct model lies between the dotted lines. Each model's deviation from the true data is shown in the error (residual) plots to the right. The straight lines through the error graphs shows the average error of the model.

5.2.2 Terminal shapes and glaciation differences

No wonder we find such a high percentage of glaciers with a debris-covered terminus in central Spitsbergen, an area dominated by small glaciers. The immediate area of Longyearbreen clearly shows the great occurrence of debris-covered fronts and high terminal ice cores amount that seems to be typical for central Spitsbergen (figure 5.1a). Analysis of aerial photographs showed that the debris-covered margins of these glaciers tend to be rounded in shape, whereas the fronts of glaciers in western Spitsbergen, if debris-covered, tend to produce a flatter debris-covered margin with a wide and low central area. A good example of a glacier in central Spitsbergen with such a rounded debris-covered margin is Longyearbreen (figure 5.5). Extensive mapping showed that this type of debris-covered glacier front is common in Central Spitsbergen, especially in the part of Nordenskiöld Land north of Adventfjorden and Adventdalen.

Towards the west coast of Spitsbergen the wetter climate (Humlum, 2002) results in a greater amount of large glaciers (> 3 km^2). These bigger glaciers often have large retreat areas with minimal ice core (figure 5.1b) compared to the terminal areas of their smaller siblings. The front of a typical large glacier may have a narrow very thin band of freshly melted out debris, but this does not seem to indicate that the glacier is starting a debris cover as aerial pictures show retreat, rather than debris build up, even within the last ~20 years. The terminal areas of these big glaciers are often, but not always, rimmed by ice-cored terminal moraines, either intact or degraded into patches,

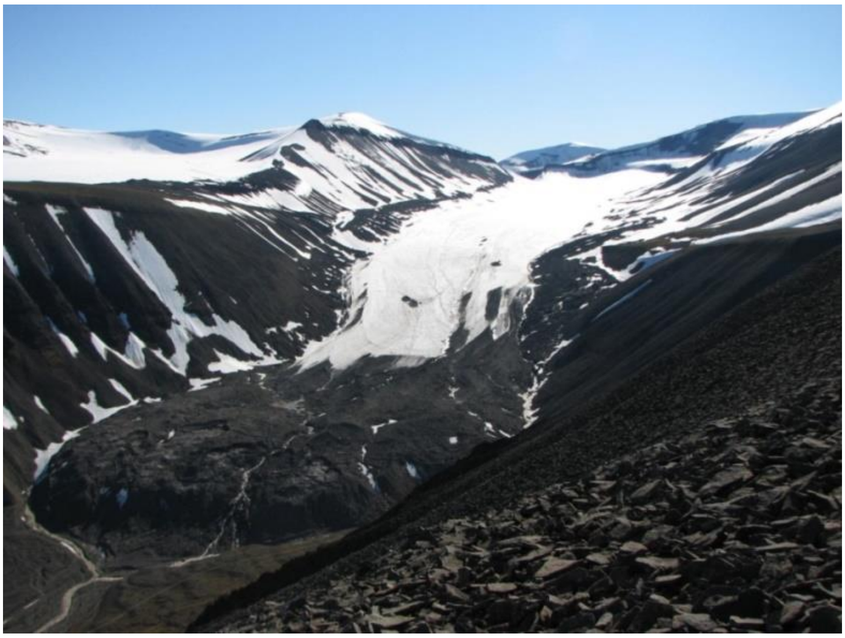

Figure 5.5: Longyearbreen is a relatively small glacier with a large, rounded, debris-covered front. The transition from covered to clean ice is gradual (note the wave like greyish ice at the transition and the lack of a height difference between clean and covered ice). The glacier debris-covered front has steep edge, as is common for cold-based glaciers. Note the extensive slumping due to meltout at the unstable rim of the debris-covered margin. Photo by Ole Humlum.

showing the LIA max advance. Buried ice can be found in the terminal areas of such glaciers, but the ice-core is patchy and limited to moraines and disconnected mounds within the ice-core-free area.

Larger glaciers show a distinctly different response to the recent climatic warming than smaller glaciers in the study area. Whereas negative mass balances have caused large glaciers to retreat, thinning of small glaciers has led to the formation of extensive terminal debris covers. This has enabled small glaciers to remain at, or close to, their previous maximum extent.

But is this difference between small and large glaciers truly rooted in their sizes? Most small glaciers are after all found in Central Spitsbergen, while West Spitsbergen is the home of their greater siblings. Is this a difference between West and Central Spitsbergen or between large and small glaciers? The answer lies in the similarities and differences between terminal zones of small and large glaciers within Central Spitsbergen and within Western Spitsbergen.

Luckily Central Spitsbergen does not only contain small glaciers, just as the west coast does not solely host vast ice masses. The two glaciers, Agnorbreen (figure 5.6) and Oliverbreen, that can be seen on figure 5.1b, form a good example of smaller glaciers near Spitsbergen's west coast. These glaciers are located right next to large ones such as Elisebreen (figure 5.6) and Andreasbreen and can therefore be expected to experience the same climatic conditions. The difference in their glacial margin is nonetheless distinct. Unlike their larger neighbours, both small glaciers, Agnorbreen and Oliverbreen, have a well-developed debris-covered terminus. In fact, the whole of their marginal zones still seems to be glaciated and forms a continuous ice mass with the debris-free glacier ice higher up. As this example shows, the strong negative trend between a glaciers clean size and its debris-covered fraction holds, independent of location. And this example is not on its own. Whether it is in the West or in Central Spitsbergen, larger glaciers tend have relatively little debris cover, while small glaciers often develop an extensive debris-covered terminus.

This is not to say that there are no differences between the margins of equally sized glaciers in Western and Central Spitsbergen. Both Agnorbreen and Oliverbreen are glaciers with a debris-covered glacier terminus, but there is a difference between these debris-covered fronts and for example Longyearbreen's and that is the shape of the covered margin. While small glaciers at the west coast of Spitsbergen clearly do form debris-covered margins, and are well capable of forming the distinct tongue like shape, the centre of the tongue is typically lower (visible in figure 5.1b as dark green patches within the otherwise light green tongue) than the edges. This seems to be less common in central Spitsbergen, where slightly more elevated debris-covered terminal zones can be found.

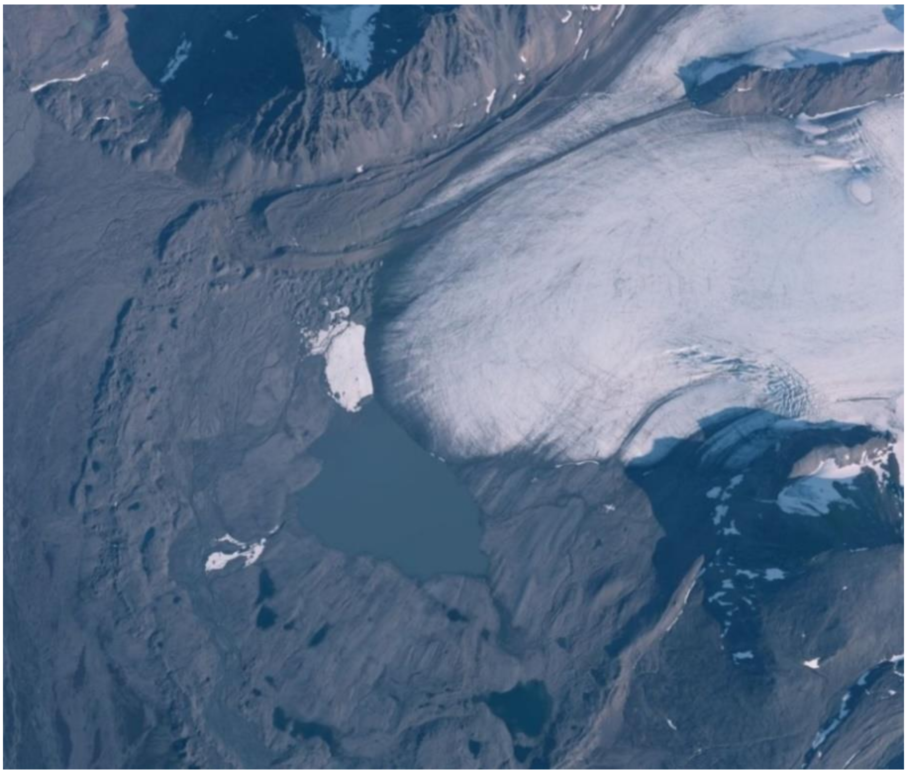

Figure 5.6: Aerial photograph of two glaciers at the west coast of Spitsbergen, Oscar II Land; Elisebreen and the smaller Agnorbreen. Elisebreen has a terminal zone dominated by glacial retreat, exposing fluted or drumlinized areas and ground moraine, while meltout at Agnorbreen has resulted in a debris-covered terminus. Agnorbreen's debris-covered front is tongue shaped and has a lower middle.

5.2.3 Exceptions to the rule: small glaciers with little ice-core in the marginal zones

The general trend in the studied areas is for small glaciers to produce large debris-covered glacier fronts, while large glaciers create relatively little ice-core in the terminal zone (figure 5.4). This trend holds for the entire study area (central Nordenskiöld Land, coast of Nordenskiöld Land, coast of Oscar II land and the island Prins Karls Forland), but as with any trend there are exceptions to the rule.

Contrary to the what is found elsewhere, two very small (just above 1 km in length) glaciers at St. Jonsfjorden, on the west coast of Oscar II Land, do have relatively large ice-free retreat areas (figure 5.8a). Only a few small glaciers in this area are such, but they clearly do exist. Another exception is Linnébreen, a small (just under 2.5 km in length) glacier near the coast of Nordenskiöld Land (figure 5.7). The terminal area of Linnébreen is marked by a great ice-cored terminal moraine. Between which and the current glacier front lies a large ice-free area dominated by braided meltwater streams and small shallow lakes.

In central Spitsbergen too, there are small glaciers without debris-covered margins. Tellbreen (figure 5.8b) and Scott Turnerbreen for example have a marginal zone characterized by a sequence of ice-cored moraines in between which largely ice-free areas exists.

These small glaciers without a debris-covered margin are relatively uncommon, but nonetheless present. The other possible exception to the rule, a large glacier with a debris-covered margin, has not been discovered within the 164 studied glaciers in Nordenskiöld land, Prins Karls Forland and the coastal areas of Oscar II Land.

The interoperation section in this thesis will come back to the interesting examples of small glaciers without a debris-covered front and look into possible explanations, including the effects surging may have on the formation or destruction of debris-covered glacier margins.

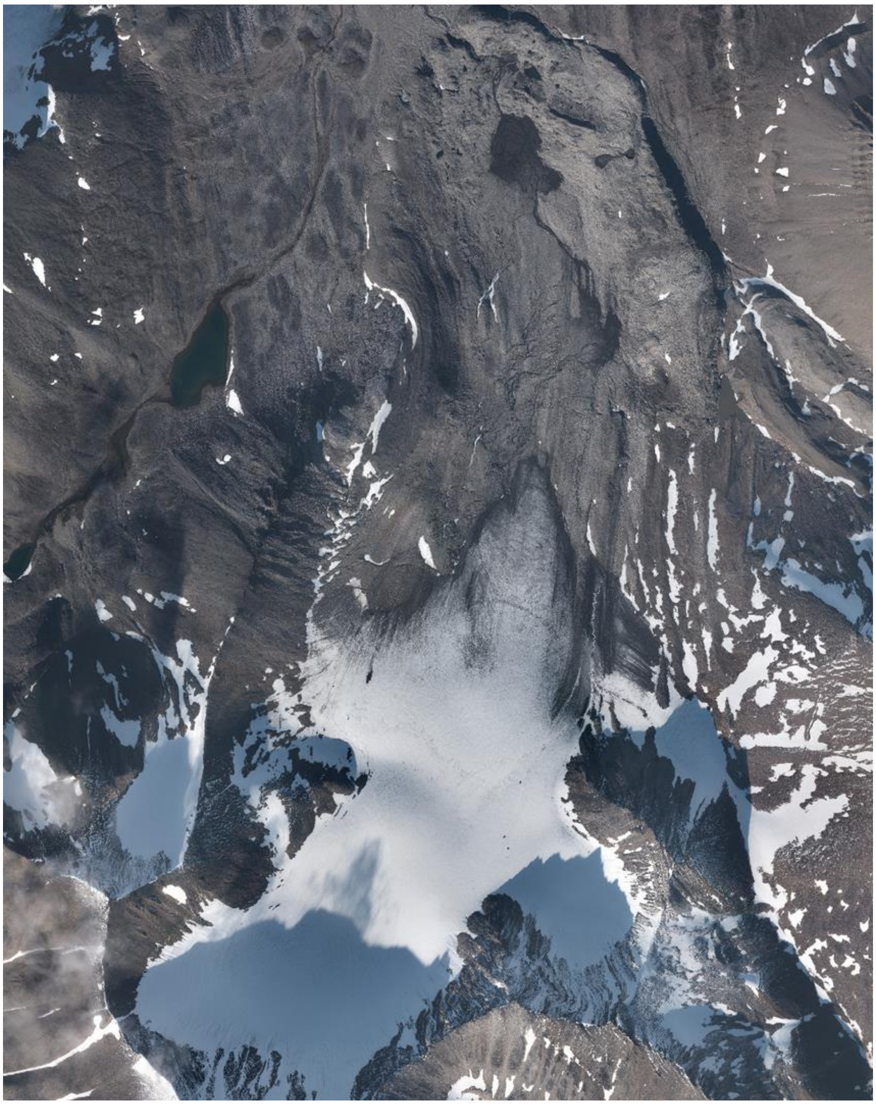

Figure 5.7: Aerial photograph of the glacier Linnébreen at the west coast of Nordenskiöld Land, Spitsbergen. Even though Linnébreen is a small glacier, no debris-covered front has formed.

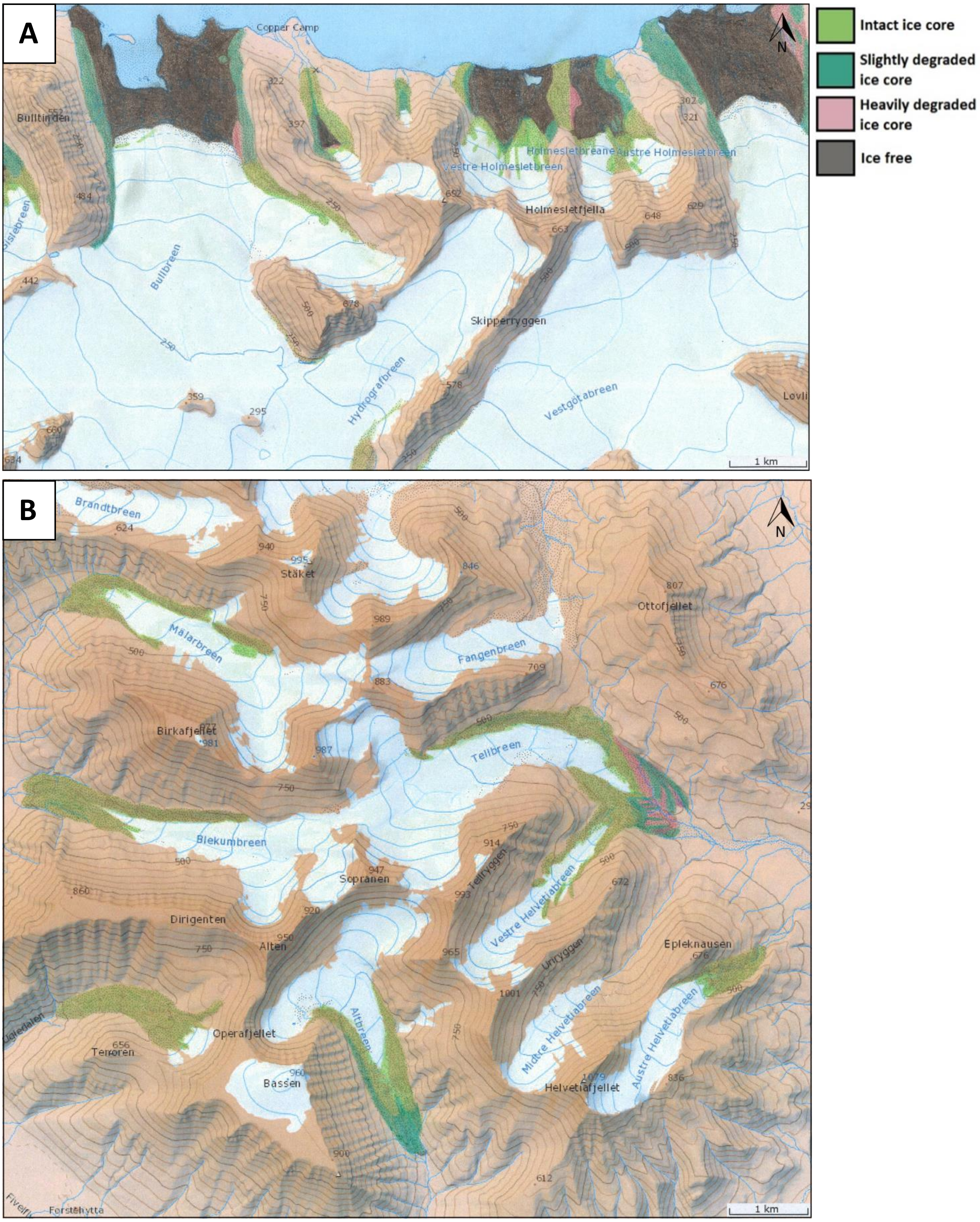

Figure 5.8: A: Ice-core map of Holmesletfjella area, west coast of Spitsbergen, Oscar II Land (for location see figure 2.1 and 2.4). The small glaciers at Holmesletfjella have clear retreat areas which is unusual for glaciers of this size. B: Ice-core map of Tellbreen area, central Spitsbergen, Nordenskiöld Land (for location see figure 2.1 and 2.4). Intact ice core means the glacial ice is still present below the debris cover, the area is free from any ice-free areas; Slightly degraded ice core means the area is mostly ice-cored, but few ice-free areas do exist; Heavily degraded ice core means the area is mostly ice-free but contains patches of ice-cored debris. Colours can appear slightly darker or lighter due to transparency with the terrain of the underlying base map.

5.2.4 The influence of headwall area on debris-covered ice

Theoretically a large debris supplying surface area would be expected to lead to a large debris-covered glacier area. Especially for the many cold- and largely cold-based glaciers of Svalbard, supraglacial debris supply via avalanches and rockfall is thought to be of great importance (Etzelmüller et al., 2000). To test the relevance of supraglacial debris sourcing on glaciers in the study area (figure 2.1 and 2.4. section 'Description of study areas') a possible relation between debris-sourcing slopes and debris-covered ice is investigated. Small glaciers inherently have a large headwall and side-slope area relative to glacier size. A possible relation between the surface areas of theses slopes and the debris-covered surface areas of glaciers could thus explain why small glaciers tend to have a large debris-covered percentage and vice versa.

The surface areas of slopes likely to supply supraglacial debris to a glacier have been calculated for 22 Glaciers, ranging in size from 15 to 0.9 km^2 and distributed over Nordenskiöld Land and the coasts of Prins Karls Forland and Oscar II Land (figure 2.1 and 2.4. section 'Description of study areas'). These surface areas can include both side slopes and headwalls and even nunataks (which is counted as headwall area since they usually sit high on the glacier). Figure 5.9 relates the surface area of debris-covered ice to the surface area of debris supplying slopes and headwalls.

The result is not exactly encouraging: the 22 glaciers form a cloud of data points without any direction. R-square values of a fitted trend lines are extremely low at 0.01 for a linear function and 0.07 for an exponential function. Also, when only side-slopes or only headwall areas are plotted against the debris-covered glacier surface there is no correlation. For the glaciers tested, the headwall and side-slopes, do not seem to be deciding factors on whether a glacier has a large or a small debris-covered area.

Does this mean that headwall and side-slope surface areas are wholly unrelated to the debris-covered surface area of glaciers? That is rather unlikely, since avalanches and rockfall events will inevitably to supply debris to the glaciers below. This has even been observed on Longyearbreen (chapter 9.1 'Formation of Longyearbreen's debris-covered terminus'). It is more probable that the effects of supraglacial debris supply are either misrepresented due to the low amount of data points, inadequate methods of debris-supplying area calculation or because the relation is overshadowed by another, more dominant process. The interpretation chapter of this thesis will further analyse this.

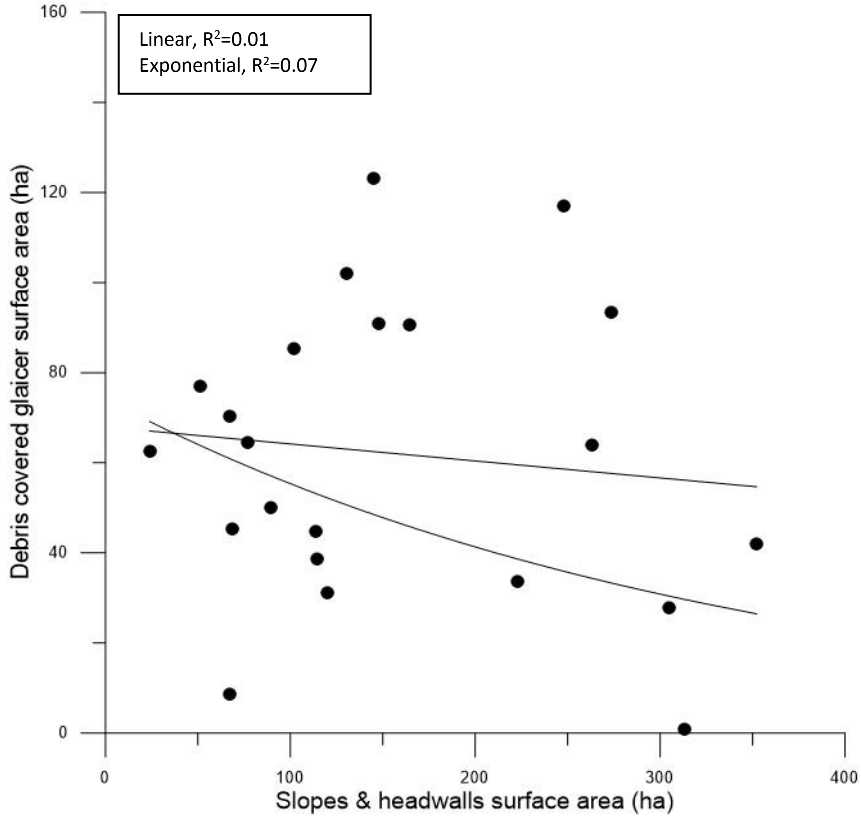

Figure 5.9: The surface area of debris-covered zones on glaciers seems to be unrelated to the surface area of the adjoining debris-supplying slopes.

5.2.5 Surging and debris-covered ice

A process that is certain to strongly influence the terminal and proglacial zone is surging. It is therefore important to look into the implications surging has for the glaciers in this study. The hypotheses for glaciers in Central and Western Spitsbergen is: Surge-type glaciers are unlikely to form debris-covered termini. This hypothesis is based on the assumption that surge events hinder the building of a debris-covered glacier front in the two ways. First of all, an existing debris-covered front and any other debris-covered ice in the terminal zone would be disintegrated. Either by the breaking up off glacial ice on which the debris lays, or by ice advancing over ice-cored structures in the terminal zone. Either way, the effects on the debris-covered ice are destructive. Secondly a recurrent surging cycle may prevent the building of a consistent debris cover all together if there is not enough time to build a post-surge debris cover before the coming of a subsequent surge. The time needed to build a post-surge debris cover would depend on the amount of englacial debris and the rate at which the debris melts out of the ice.

Table 2: Inside and near (within 15 km of) the study area the following glaciers have been observed to surge or hold strong evidence to support a former surge-event. The study area is composed of Nordenskiöld Land and the coasts of Prins Karls Forland and Oscar II Land (figure 2.1 and 2.4. section 'Description of study areas'). Italic writing indicates a calving glacier.

In the study area: Surge-type glaciers	**Nearby the study area: Surge-type glaciers**
Aavatsmarkbreen (Sobota et al., 2016)	Austre Brøggerbreen (Hagen et al., 1993)
Drønbreen (Hagen et al., 1993)	*Fridtjovbreen (Lønne, 2006)*
Elisebreen (Larsen et al., 2006)	Hyllingebreen (Hagen et al., 1993)
Lunckebreen (Hagen et al., 1993)	*Kongsvegen (Melvold and Hagen, 1998)*
Marthabreen (Hagen et al., 1993)	*Kronebreen (Hagen et al., 1993)*
Møysalbreen (Hagen et al., 1993)	Midre Lovénbreen (Hagen et al., 1993)
Scott Turnerbreen (Hagen et al., 1993)	*Nansenbreen (Hagen et al., 1993)*
Tellbreen (Sevestre et al., 2015, Lovell et al., 2015)	*Osbornebreen (Dowdeswell et al., 1991)*
	Skutbreen (Hagen et al., 1993)
	Glacier under Slottsmøya (Hagen et al., 1993)
	Glacier under Storknausen (Hagen et al., 1993)
	Vendombreen (Hagen et al., 1993)

A topic like this undoubtedly could fill an entire thesis by itself. It follows that this small sub-chapter cannot be expected to provide a definite answer, but even so, a start can be made. For research on surging glaciers is not new to Svalbard. On 8 glaciers within the study area surging has either been observed or strong evidence has been found to support a former surge-event (table 2). Twelve more glaciers nearby the study area (within 15 km) have been identified as surge type (table 2). Presumably many more glaciers in and near the study area have a history of surging. But not every glacier studied in this thesis has been the subject of previous surge-research and due to the remote location of the archipelago observations may have been limited or non-existent. As a result, several surges may have occurred which have not been recorded.

To discover if there is evidence for a negative relationship between surging and debris-covered ice content in a glacier terminus one can look at the aerial photographs and ice-content maps of the glaciers. Since the aim here is to discover debris-covered ice in the terminal zone, a land-based terminal zone must exist in order to do so. This means that the calving glaciers Aavatsmarkbreen, Fridtjovbreen, Kongsvegen, Kronebreen, Nansenbreen and Osbornebreen are excluded from the following analysis.

Starting by looking at the mapped ice content and aerial photographs of the surge-type glaciers in the study area, a pattern appears. None of the seven land-terminating surge-type glaciers have a well-developed debris-covered front. The glacier Lunckebreen however does have a partially degraded debris-covered terminus (5.10). All other six land-based surge-type glaciers in the study area have a strongly degraded and/or largely ice-free retreat area. Marthabreen (figure 5.10) shows the typical terminus of a surge-type glacier in the study area: the glacier has clearly retreated, leaving degraded ice-cored structures in the forms of moraines and bumps amidst apparently ice-free terrain, while the current glacier front itself is largely debris-free. Glaciers with a surge history in the study area thus seem to follow the predicted pattern and are unlikely to form debris-covered fronts. Off course one cannot draw a firm conclusion from such a small sample of glaciers. For this reason, the study sample has been enlarged. All glaciers within 15 km of one of the 164 previously studied glaciers (chapter 5.2 'Large-scale study of glaciers in Nordenskiöld Land and the west coast of Spitsbergen') for which a surge has been recorded are investigated as well.

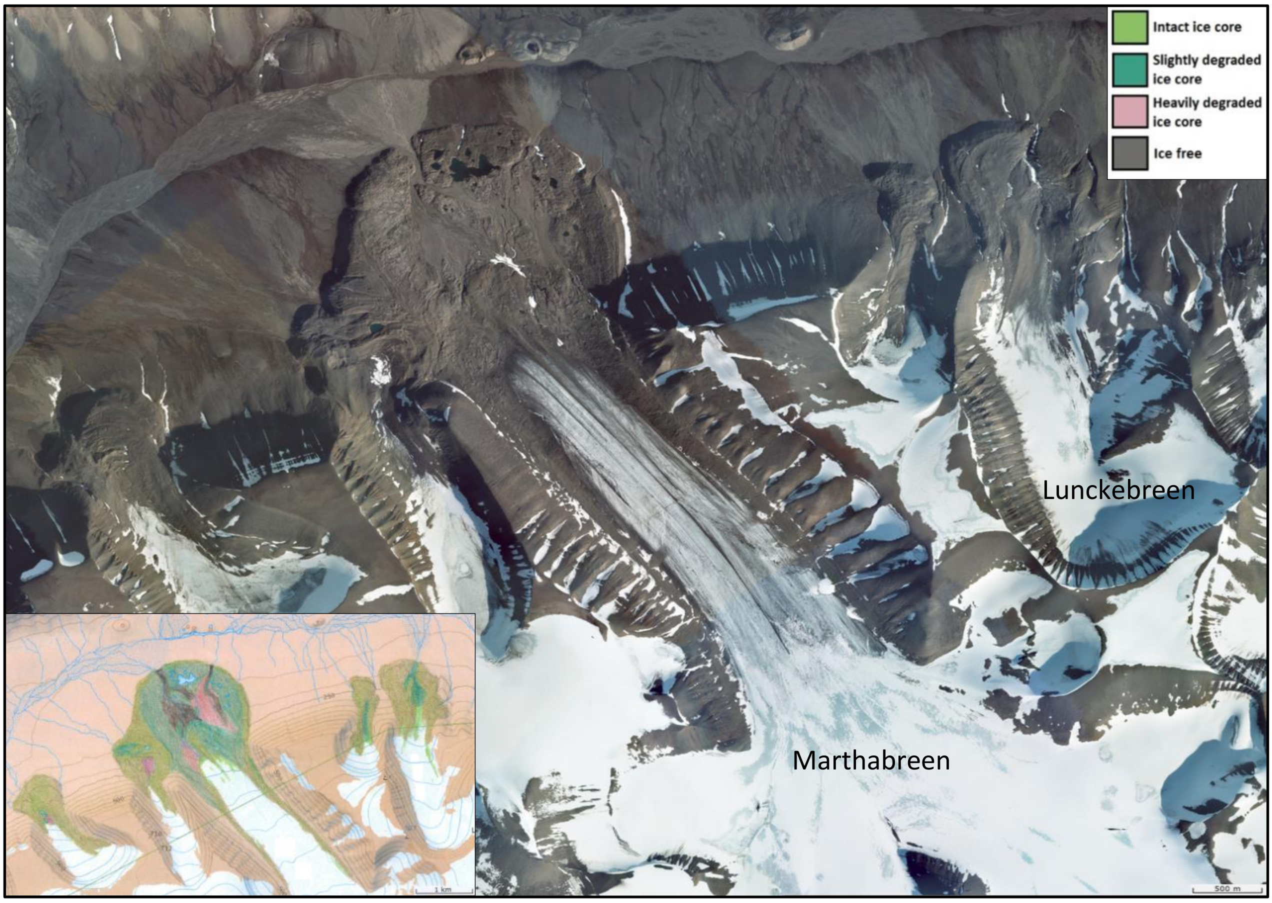

Figure 5.10: Aerial photograph of Marthabreen and Lunckebreen. Both glaciers experienced a surge event in c. 1925 and c. 1930 respectively (Hagen et al., 1993). Inserted is an ice core map of the same area. Marthabreen has little debris-covered ice in its terminal zone, just like all other surge-type glaciers in the study area, but Lunckebreen. Lunckebreen on the other hand has formed a somewhat degraded debris-covered glacier front and is the only known surge-type glacier within the study area of 144 glaciers to have done so. A legend explains the degradation state of the terminal ice. Intact ice core means the glacial ice is still present below the debris cover, the area is free from any ice-free areas; Slightly degraded ice core means the area is mostly ice-cored, but few ice-free areas do exist; Heavily degraded ice core means the area is mostly ice-free but contains patches of ice-cored debris. Colours can appear slightly darker or lighter due to transparency with the terrain of the underlying base map.

Enlarging the former study area by 15 km adds 12 surge-type glaciers to the study (table 2), 7 of which are land-terminating and thus of use to this study. Once again none of the glaciers have a large well-developed debris-covered front. But apart from that the results are not as clear as for the glaciers within the former study area. Although the majority of the glaciers (4 out of 7) show the so-far typical surge-type terminus, with little debris-covered ice in the terminal zone and a clear retreat area, three glaciers do not quite fit this pattern. Figure 5.11 shows the large Skutbreen (surged in 1930) and two smaller unnamed glaciers underneath the mountains Slottsmøya and Storknausen (both surged in 1960). Skutbreen's terminal zone fits the majority of surge-type glaciers in the area. The retreat area is largely ice-free and little debris covers the glacier front, just as is found on the three glaciers Austre Brøggerbreen (surged c. 1890), Midre Lovénbreen (surged c. 1890) and Vendombreen (surged c. 1934). The small unnamed glaciers however are a little different and, to a lesser and greater degree, do have debris covering the glacial ice (figure 5.11).

The 3rd a-typical glacier, Hyllingebreen, seems to somehow have managed to both have a small debris-covered front and a clear post-surge retreat area (figure 6.2). It goes to show that there likely is a middle ground to the proposed hypothesis. After a surge-event a glacier indeed does retreat and has little debris cover, but depending on the amount of englacial debris, the melt-rate and the surge-free time-span, a new debris cover can be built fairly quickly. Hyllingebreen surged in 1970 to 1980 (Hagen et al., 1993) and has since managed to build a small debris-covered glacier front.

Surging in and nearby the study area seems to be indiscriminate, affecting all types of glaciers, from small inland ones to large calving, tidewater glaciers. This is in line with larger scale studies (Hagen et al., 1993). The land terminating surge-type glaciers in and near the study area range from 2.1 km^2 to 13.2 km^2 of clean glacier ice and 3.3 km^2 to 13.5 km^2 of total glacier size. This large range covers the major part of the glaciers studied in this thesis. The calving surge-type glaciers are even larger.

Although the sample size of 14 glaciers with a surge history is admittedly too small to draw any definite conclusions, the proposed hypothesis does seem to hold to a certain extent. The majority of the investigated surge-type glaciers have little supra-glacial debris and none of the 14 glaciers have a large well developed debris-covered front. However, a small relatively well developed debris-covered front is found on Hyllingebreen and a somewhat degraded debris-covered glacier front on Lunckebreen. Additionally, debris cover is present to differing degrees on the two small unnamed glaciers underneath the mountains Slottsmøya and Storknausen. All in all, 13 out of 14 glaciers do not have an intact debris-covered front. And 10 out of 14 glaciers show the same, by now typical, picture: A post-surge glacier with little supra-glacial debris and a large retreat area with some degrading ice-cored structures.

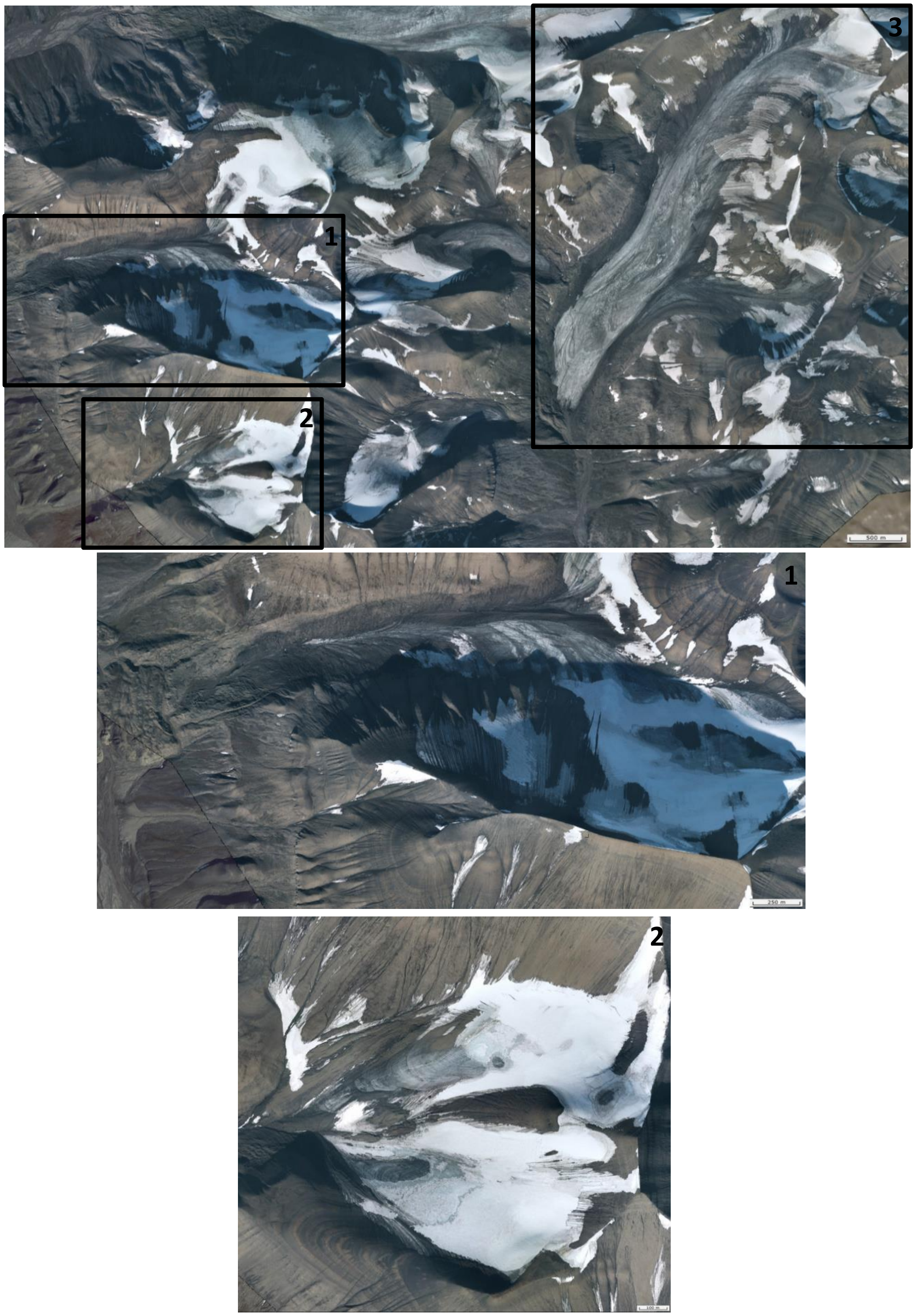

Figure 5.11: Glaciers up to 15 km outside the study area with a surge history. The top picture shows 3 surge-type glaciers, boxed and numbered. Glacier 3, Skutbreen, has little debris-covered ice in its terminal zone. Skutbreen is one of 3 surge-type glaciers nearby the study area that shows this typical surge-type terminus. The unnamed glaciers 1 (under mountain Slottsmøya) and 2 (under mountain Storknausen) are slightly different. They have no well-developed debris-covered front, but clearly are partially debris-covered: glacier 1 to a small extent and glacier 2 to a large extent. Pictures from 2009.

5.2.6 The influence of wind on glacier orientation

Wind-blown snow can be of great importance to glacier creation, growth and orientation. Therefore, it would logically also effect the debris-covered part of a glacier. This section studies the relation between wind and the orientation of glaciers and their debris-covered margins.

The prevailing wind direction in Nordenskiöld land is ESE (figure 5.13c, measured at Gruvefjellet meteorological station). This weather station is located on top of a mountain plateau and thereby exposed to the general wind direction, relatively uninfluenced by local topography. Other stations in the area exist, but these are often located in valleys, where air flow is channelized, and are therefore less well situated to measure the general wind direction.

In Oscar II Land and Prins Karls Forland there is no weather station well placed to measure the general wind direction. There is however an easy way to discover the regional prevailing wind-direction in the area. As wind-blown snow settles on the lee side of nunataks and mountain tops/ridges, a natural snow-blown 'wind-arrow' forms. The direction of 37 of these multi-year snowdrifts in Oscar II land has been measured and portrayed on aerial photographs such as the one shown in figure 5.12. The natural 'wind-arrows' found on the ice-caps of Oscar II Land are up to a kilometre in length (figure 5.12). Prins Karls Forland does not contain large enough glaciers to be free of local topographical influences. It therefore proved not suited to this way of wind direction measurements.

Figure 5.12: Wind features on the ice caps of Oscar II Land, notice the large scale of these features (1 km scale bar is located in the bottom right).

The wind-direction revealed by the 37 snow and ice features measured in Oscar II Land is shown in rose-diagram D of figure 5.13. It is almost identical to the wind chart of the Gruvefjellet weather station in Nordenskiöld Land (figure 5.13c). The average wind direction on Oscar II Land is ESE, with a range from South to East and one outlier with a direction of WSW. This makes the general wind direction for study area ESE (The general wind-direction in Prins Karls Forland is however unknown).

Wind dependence of glaciers is most visible in the arid part of the study area. Since the windward facing valleys here do not receive enough snow to allow for glacier build up, these valleys are ice-free. The arid inland area of Nordenskiöld Land forms a perfect area to study the influence of wind on glacier orientation in Central Spitsbergen. Forty glaciers within a radius of 20 km from Longyearbyen were studied to find a possible relation between wind direction and the orientation of a glacier or its debris-covered margin. Diagram A and C of figure 5.13 clearly show that glaciers in inner Nordenskiöld Land are orientated away from the prevailing wind direction. The debris-covered glacier margins too are orientated downwind in relation to the general wind direction, largely reflecting the orientation of the glaciers that formed them. The glaciers in inner Nordenskiöld Land (central Spitsbergen) are hardly ever orientated towards the prevailing wind direction. Instead glaciers are orientated downwind, on the leeward side of a headwall. Local variations in wind direction due to valley-funnelling mean that glacier-orientation is not simply the opposite of the prevailing wind direction, as would be expected if local topography was of no influence. But even with the influence of local topography figure 5.13 clearly shows that glaciers in inner Nordenskiöld Land are orientated away from the prevailing wind direction. When taking the local wind direction of each glacier into account one will often find that the glacier is orientated perfectly opposite the local wind direction.

It is however noteworthy that it does not seem to be the main wind direction that is the driver of glacier and debris-covered glacier margin orientation, otherwise the main glacier orientation would be WNW rather than the now found NNE. The absent wind directions seem to be of bigger impact. In other words, the direction from where the wind never comes is where we find the glaciers. This 'wind-gap' in towards which the glaciers are orientated spreads from WNW to E (figure 5.13 a,b,c).

The zone of the study area where wind dependence of glaciers is most visible, Central Spitsbergen, is interestingly also the zone where most glaciers with a debris-covered terminus are found. One should however be careful with drawing any conclusions from this. For wind is unlikely to be a great direct contributor to debris covers. There is after all a limit to the sizes of particles suspendable by wind. And although dust may contribute to debris build-up, the relative contribution would be minor. If a strong relation between wind and debris cover exists it is likely to be indirect.

In the more extensively glaciated areas of the archipelago, such as Oscar II Land, relatively great amounts of precipitation have allowed for the formation of large glaciers that have effectively buried many headwalls and their leeward sides and filled even the windward facing valleys. Snow distribution by wind certainly pays a role here as well. The wind-blown features of up to 1 kilometre in length are widespread and often occurring in this part of the study area. But a simple orientation analyses, as performed in inner Nordenskiöld Land cannot be performed here due to the large size of the glaciers, burying most valleys in glacial ice.

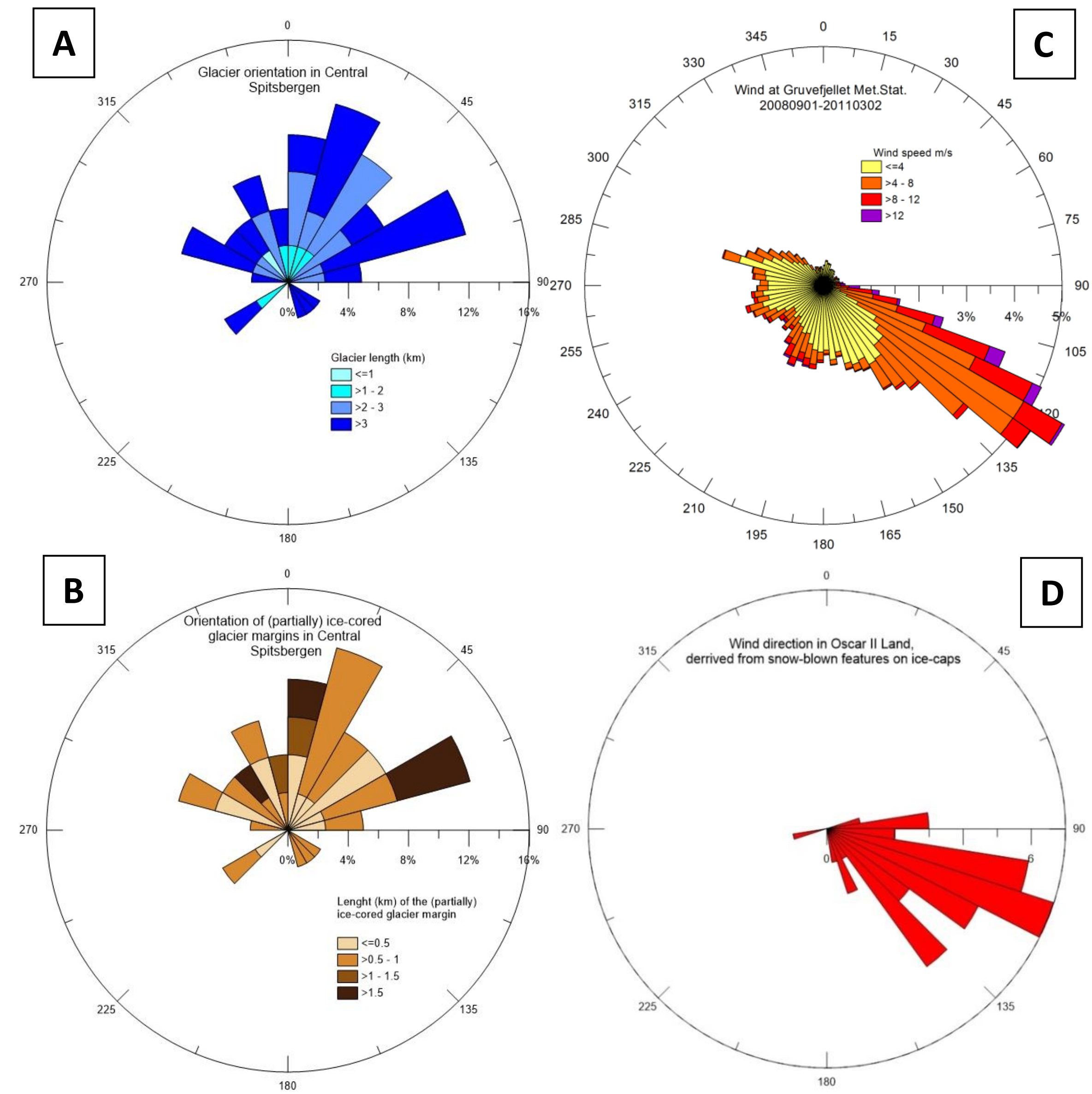

Figure 5.13: Glacier length and orientation is displayed in diagrams A and B (dataset of 40 glaciers in inner Nordenskiöld land within a 20 km radius from Longyearbreen). Diagrams C and D show the prevailing wind direction in Nordenskiöld Land (at Gruvefjellet meteorological station, nearby Longyearbyen) and Oscar II Land (derived from 37 wind-blown features on glaciers in the area). Gruvefjellet meteorological station is located on a mountain-plateau and thus not influenced by local topography.

6. Interpretation: Glacier margins along the West Coast of Svalbard and in Central Spitsbergen

6.1 Small glaciers, large debris covers

The 88 measured glaciers in Nordenskiöld Land and along the coasts of Prins Karls Forland and Oscar II Land show a clear negative correlation, and probable relation, between the fraction of the glacier that is debris-covered and the glacier's clean (free from debris cover) surface area. Small glaciers tend to produce large debris-covered glacier fronts, while large glaciers create relatively little ice-core in the terminal zone (figure 6.1). For glaciers of up to 15 km^2 the study provides a useful model that predicts the degree to which a glacier is debris-covered. In which the relation between the debris-covered percentage of a glaciers surface (Y) and the clean glacier surface of the glacier in km^2 (x) is described as:

$$Y = -12.72 \ln x + 35.166$$

This simple formula makes it possible to predict the debris-covered fraction of a glacier within the study area. The fact that this model is valid for both the relatively wet and warm West and the arid, colder Centre of Spitsbergen is encouraging; it is possible that the model will hold for a large range of climatic settings. It must however be noted that this is only a preliminary study. Large scale testing is necessary to validate the findings of this study and the model's usability outside the study area.

For the model, it has been chosen to work with the clean (free of debris cover) glacier area rather than the total glacier area. It is important to realize the reasoning and consequences of this decision, because it does have its drawbacks. Inherently there is always going to be some form of a negative relation between debris-covered ice and clean ice of a glacier. Because if a large portion of a glacier is debris-covered the non-covered portion is going to be correspondingly small. This misleading effect is not that strong here since we use the absolute clean glacier size, not the relative one. In the study area, a large absolute clean glacier size always belongs to a large glacier and a small absolute clean glacier size always belongs to a small glacier, without exception. Large glaciers with (very) small absolute clean glacier sizes do not exist in the study area, so there is no misleading effect here. There are however small glaciers with very small absolute clean glacier sizes and great debris-covered portions, so some bias may affect the data here.

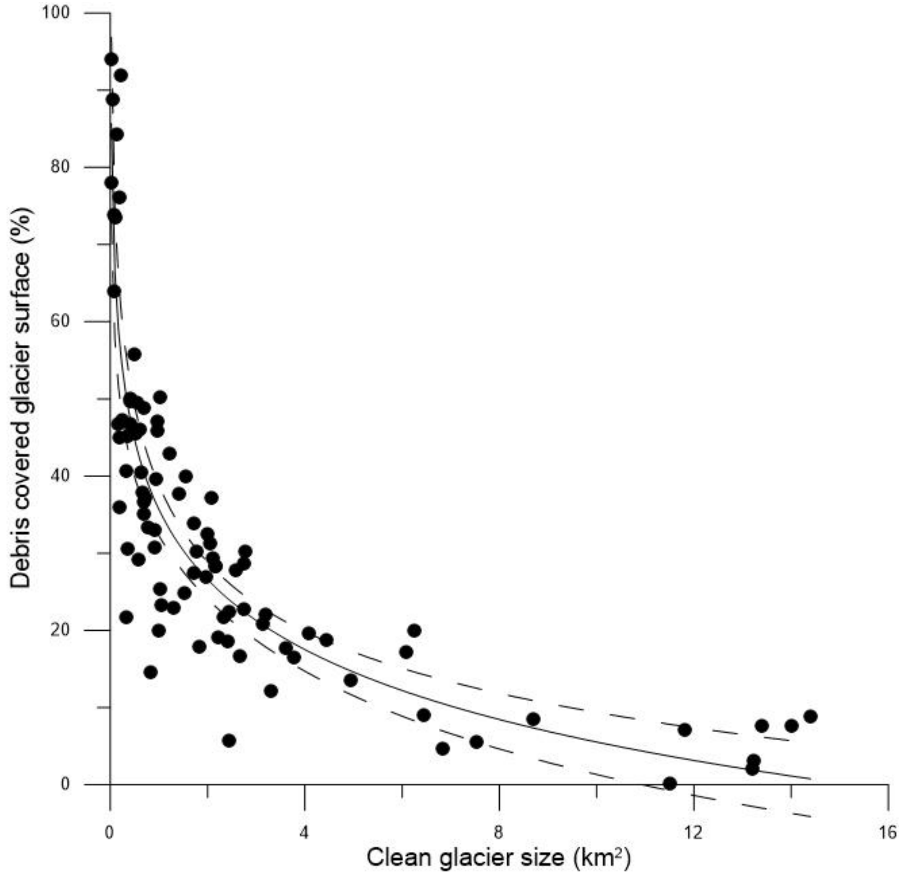

Figure 6.1: 88 glaciers are plotted and fitted with a logarithmic function. The dotted lines indicate 95 % the confidence interval. Meaning that there is a 95 % chance that the correct model lies between the dotted lines.

To avoid the possible bias, the total glacier size could be used (including both debris-covered and debris-free). This too gives a strong negative relation, although not as strong as the relation with clean glacier size. This indicates that, yes, the correlation in figure 6.1 is inherently slightly biased but the major part of the displayed correlation is real. Then why do we use clean glacier size and not total glacier size? This is done for practical purposes. Clean glacier size is what we see and perceive as a glacier. This is the area that in basically all other research will be called the glacier size, as if it encompasses the total ice area. To allow this research to be of use to-, and comparable to data from current and previous papers the choice was made to relate to clean glacier size.

All in all, the results of this study show that there is a clear negative correlation, and probable relation, between the fraction of the glacier that is debris-covered and a glacier's size. Off course these findings spring the immediate question of 'why?'. What is it that causes this negative trend between a glacier's clean size and its debris-covered fraction? A small start is made towards answering this question. To do so the study looks into the effects of supraglacial debris supply from headwalls and side-slopes. Providing a comprehensive answer to this question however lies outside the scope of this study. The focus of this research has been to first of all map glacier margins over a widespread area. Only when debris-covered fronts indeed did show to be widespread over a large area, the focus could change to discovering trends and ultimately developing a model to predict a glacier's debris-covered fraction. But as stated, this study does include a first step towards understanding the mechanisms that cause a small glacier to form a relatively large debris cover. A theory relating a glaciers headwall size to its debris-covered area has been tested (see chapter 5.2.4 'The influence of headwall area on debris-covered ice') but has yet failed to bring forward any conclusive results. Possible research following up on the findings of this study is highly recommended to further analyse the processes causing the negative relation between a glaciers clean size and the debris-covered fraction.

The following sections will present theories aimed at understanding the results brought forward in earlier chapters. This includes the effect climate may have on the different types of debris-covered glacier fronts found in the study area. Next to this a theory based on glacial surging is proposed as an explanation for the odd few small glaciers without a debris-covered front.

6.2 Climatic differences and different types of debris-covered glacier fronts

Towards the west coast of Spitsbergen the relatively high amount of precipitation (Humlum, 2002) results in a greater number of large glaciers (> 5 km in length) than in the island's dry centre. As was shown in earlier sections, large glaciers are unlikely to develop (large) debris-covered fronts. This is true for both the inner and western parts of Spitsbergen. The relationship does not depend on location or climatic differences. Small glaciers in the centre of Spitsbergen tend to form debris-covered fronts as do small glaciers in the west, examples of which are Agnorbreen and Oliverbreen, shown in figures 5.1b and 5.6 of the result section. These glaciers are located right next to large ones such as Elisebreen (figure 5.6) and Andreasbreen and can be expected to be exposed to the same climatic conditions. But, as shown in the result section of this thesis, the glacial margins of the neighbouring large and small glaciers are far from similar. Both small glaciers (Agnorbreen and Oliverbreen) have a debris-covered glacier front. But the terminal areas of the large neighbouring glaciers (Irenebreen, Elisebreen, Eivindbreen and Andreasbreen) are, apart from their ice-cored moraines, largely ice-free. A very different picture from the glacier fronts of Agnorbreen and Oliverbreen.

The occurrence of debris-covered glacier fronts appears to be independent of climatic differences within the studied area. However, as noted in section 5.2.2 'Terminal shapes and glaciation differences' there is a difference between the debris-covered fronts of small glaciers near the west coast and in central Spitsbergen. In Western Spitsbergen, the centre of the debris-covered glacier tongues is typically more degraded and thus lower (shown in figure 5.1b as dark green patches within the otherwise light green tongue) than the edges. In Central Spitsbergen, a rounded debris-covered margin, like the one on Longyearbreen (figure 5.5, result section), is more common.

The shape of a debris-covered glacier front merely mimics the shape of the underlying ice. Therefore, climatic differences influencing the formation or reduction of glacial ice can be held responsible for shape differences of debris-covered fronts in different climatic settings. The rounded fronts in the centre of Spitsbergen versus the flattened or hollow fronts in the west could well be related to climate.

Western Spitsbergen is wetter and warmer than the inner parts of the island (Humlum, 2002), not ideal conditions for preserving ice. In such conditions one could expect faster degradation of debris-covered ice as more melt-water and rain will percolate through the debris cover, warming the ice below. This can occur via ice-melt directly induced by water percolation through the debris layer, especially prominent in ice covered by a high-permeability thin supraglacial debris cover (Reznichenko et al., 2010), but may also take place in the form of warming by refreezing of melt- or rain water (Polashenski et al., 2014). However, refreezing can actually help ice-preservation. The more ice-rich the debris is, the higher its conductive properties become. This helps to cool the glacial ice in winter.

There are however more ways in which the wetter and warmer climate of Western Spitsbergen influences debris-covered ice. A wet climate also produces more snow and a thick snow cover can effectively diminish conductive cooling of the glacial ice during winter. A thick snow cover will additionally lead to more meltwater in the glacial streams and rivers. This leads to strong thermal river erosion which can result in degradation of the ice within and underneath the debris layer.

In Central Spitsbergen, the effects of melt water, rain and snow cover will be less, as there simple is less snow or rain to fall. This not only reduces the warming effect seen in the wetter climate of Western Spitsbergen, it even gets to the point where the summer soil (and debris cover) is prone to drying up. This can often be seen towards the end of summer in the area around Longyearbyen. Just like winter winds carry snow, summer winds carry dust. As pores become air filled the conductivity of the soil or debris layer greatly decreases. This dried ground, partially filled with water and partially filled with insulating air-filled pores, forms a very effective insulating layer, protecting the ice below. The process of evaporation also helps to cool the debris layer. As pore water evaporates heat escapes the soil, the net effect of which is cooling of the ground. Summer-time soil drying thus has a double positive effect on ice core preservation.

Overall wet conditions are unlikely to enhance ice preservation. Wet summers prevent the debris from drying out and thereby hinder its insulating effect when it is most needed. Additionally, a thick snow cover insulates the ground when air temperatures are low. This lessens conductive cooling. Resulting enhanced ice degradation could theoretically lead to a lowering of glacier fronts, resulting in the typical flattened or hollowed out debris-covered glacier-tongues that are common in western Spitsbergen. This theory however needs testing before it can be held accountable for any finds in this study.

A climate with relatively dry summers and winters seems, based on the above-mentioned factors, to be ideal for preserving debris-covered ice. The climate-forced differences may (partially) explain the different types of debris-covered glacier fronts (hollowed versus rounded) that are found in Western and Central Spitsbergen. Since the slightly degraded hollowed types are more common in the wetter and warmer West, while the Islands arid and colder Centre has preserved many of the original rounded, debris-covered fronts. The role played by temperature and moisture from rain and snowmelt is further analysed in the second part of this thesis which focuses on one glacier in Central Spitsbergen: Longyearbreen. Year-long temperature measurements at three different depths have been taken in the debris layer of Longyearbreen to gain a better understanding of the insulating properties of supra-glacial debris.

6.3 Causes for the negative trend between glacier size and the debris-covered area

Analysis of 22 glaciers did not show any correlation between the surface area of debris supplying slopes and the surface area of debris-covered zones of a glacier. So, does headwall and side-slope surface area have an impact on a glaciers debris cover at all? Avalanches and rockfall events definitely do bring debris from slopes to the glacier below and are thought to be of great importance to the debris supply of many cold- and largely cold-based glaciers of Svalbard (Etzelmüller et al., 2000). Furthermore, the angular debris observed on glaciers near Longyearbyen are indicative of a supraglacial source, at least at those locations. Not finding a clear correlation does not mean that there is no relation between headwall/side-slope debris supply and the debris-covered portion of a glacier. It just means that with the approach used a possible relation could not be discovered. A larger sample size and a more precise calculation of debris supplying areas could bring different results.

The currently used method for calculating headwall and side-slope areas could well be insufficient. Two procedures that may have introduced errors are:

- In order to compute the surface area of a slope it is of great importance to adjust for the slope angle. To calculate the slope angle height differences were measured with the use of mapped contour lines. These lines occur at an interval of 50 m, which off course creates the need to estimate elevation levels in between the contour lines.
- In order to calculate the surface area of slopes in the manner described in chapter 4.2.2 'Calculating headwall and side-slope surface areas' strong simplifications have had to be made, which resulted in the slope area being reduced to three to four smooth and even surfaces, all on an angle averaged for each area. This makes it impossible to account for small scale topography and terrain roughness.

On the other hand, a correlation between headwall area and the occurrence of a debris-covered front may be overshadowed by a more prominent process. Assuming that the sample size and methods of the current study is sufficient to discover a small correlation, if a correlation exists at all, the differences in the size of the debris supplying area do not explain the different percentages of debris cover on large and small glaciers. In this case one is left with the conclusion that glacier size is the deciding factor. But how could glacier size by itself decide or even influence the extent to which a glacier is or becomes debris-covered? There are inherent differences between large and small glaciers that make the two respond differently to exactly the same climatic changes. One example could be the position of the equilibrium line altitude, also called the ELA.

A change in climate towards increasing temperatures and/or decreasing precipitation could lead to a rise of the ELA. Now take in mind a small glacier, starting at 600 m altitude and coming down to 400 m. A larger glacier, starting at 800 m altitude and coming down to 200 m is located next to it (and thus experiences the same climatic changes). Say that 100 years ago, the ELA lay at 500 m. Now imagine a rise of the ELA of 100 m. The climatic changes impacting both glaciers are the same but its effect is much more extreme for the smaller glacier. The ELA is now located at the headwall of the small glacier, turning the whole of the glacier to ablation area. The larger neighbouring glacier on the other hand has a relatively large accumulation zone left, spanning an altitude of 200 m. In other words, the same climatic changes led the small glacier to increase its ablation zone by 100% while the large glacier increased its ablation zone by 33 %.

Since debris covers only form due to the process of debris meltout they only form below the ELA, in the ablation zone. This process of debris meltout now takes place on the entire surface of the small glacier. Thus, after a rise of the ELA the small glacier is presumably more prone to debris meltout resulting in the formation of debris covers. Consequently, the main factor deciding which glacier gets debris-covered could well be the position of the ELA.

In this example, it is crucial that for the larger glacier a greater percentage of the surface belongs to the accumulation zone than for the small glacier. This can be by a higher maximum altitude, which is likely to be the case for larger glaciers, since they tend to cover their headwalls more extensively. Additionally, the accumulating fraction of the glacier could be larger due to a wide accumulation zone and a narrow tongue. This too is more easily realized by large glaciers since they are less restricted by the topography.

Whether or not this idea holds must of course be tested by additional research before any conclusions can be made. A succeeding study might focus on the average maximum height of glaciers of different sizes as well as the relative width of the accumulation areas of large and small glaciers.

6.4 Surging and debris-covered ice

The rapid glacial advance called a surge-event greatly effects a glacier’s terminal zone. The former proglacial area is quickly covered by the advancing ice. The terrain over which the glacier advances deforms and the former glacier front disintegrates (Larsen et al., 2006). Off course anything covering the disintegrating glacial ice is also impacted. If a debris cover existed prior to the surge, is likely to break up and thin as debris slides into opening crevasses. Material in the terminal zone, including any ice-cored moraines, would be overrun or pushed forward by the advancing ice.

An advance as forceful as is seen during a surge-event must indeed have a destructive and deforming influence on any form of debris-covered ice in the terminal zone. It is however not only the possible destruction of a glacial debris cover that may lead to surge-type glaciers displaying termini with relatively little debris-covered ice. It may be that a recurrent surging cycle prevents the building of a consistent debris cover all together. The formation of a debris cover relays on relatively slow and steady processes, as will be discussed in 10.1 of this thesis. Glacial flow is a slow process, particularly in the many small cold-based glaciers of Spitsbergen. When one realises that the formation of a debris cover is heavily reliant on englacial debris transport (chapter 10.1), debris cover building does not seem to be a likely match to surge cycles with time spans of decades or even centuries. Glacier flow velocities do however not need to compete with surge cycle time spans as long as a constant high amount of debris has been transported englacially. Meltout will reveal this incorporated debris after a surge. Whether a glacier can sustain a debris cover therefor is dependent on the relative amount of englacial debris, melt rates, and the length of time in which debris cover building by meltout is not interrupted by a surge-event. It goes without saying that, based on these considerations, a glacier that is not disturbed by surge events presumably has a higher probability of creating a debris cover than a glacier hindered by the destructive forces of surging.

The hypotheses for glaciers in Central and Western Spitsbergen is consequently stated: Surge-type glaciers are unlikely to form debris-covered termini. It is however important to note that surge-type glaciers with debris-covered tongues do exist. Studies have shown that the Karakoram region of the Himalayas for example is home to many surge-type glaciers with extensive debris mantles on the lower tongues (Barrand and Murray, 2006, Quincey et al., 2011). These glaciers are largely or wholly avalanche nourished (Quincey et al., 2011), and clearly have been able to build and sustain or quickly rebuild a frontal debris cover despite surging events. It seems that debris cover building mechanisms of these Himalayan glaciers steadily carry enough material, have sufficiently high ablation rates near the terminus and operate fast enough to outweigh the surge-cycles.

However, the glacial characteristics of Svalbard's high arctic are vastly different form the high alpine glaciers of the Himalayas. Due to the sheer scale differences and associated flow rates it is hard to compare glaciers form such topographically and climatically different areas. What is true for the Himalayan glaciers does not necessarily hold for glaciers in Svalbard. Equally, what has been found in this study is unlikely to hold for glaciers in a vastly different setting, but may apply to glaciers similar in size, thickness, flow rates and with equal forms of debris supply.

As can be concluded from the result section (5.2.5) the proposed hypothesis does seem to hold. The majority of the investigated surge-type glaciers have little supra-glacial debris and debris-covered fronts are highly uncommon. However, a small relatively well developed debris-covered front is found on Hyllingebreen and a somewhat degraded debris-covered glacier front on Lunckebreen. Additionally, debris cover is present to differing degrees on the two small unnamed glaciers underneath the mountains Slottsmøya and Storknausen. The few atypical termini of surge-type glaciers may be explained by differing amounts of englacial debris. Which in turn could be related to either debris entrainment by the surge itself or the extent in which debris is supplied by supraglacial slopes. Climatic differences may also impact the glacier termini as will the time that has passed since the last surge, although the surge-free time-span seems to have been of little influence for the glaciers investigated.

Hyllingebreen, one of the atypical glaciers, serves as a good explanation for the differing degrees of debris-coverage that is found on the surge-type glacier fronts in and nearby the study area (figure 6.2). The glacier surged in 1970 to 1980 (Hagen et al., 1993) and has since build a small debris-covered front. Yet a clear post-surge retreat area can be found beyond the current covered glacier front. This glacier holds evidence of the destructive forces that act on a (covered or debris-free) glacier margin during a surge. Still the example also shows that a new debris cover can be built fairly quickly, as long as the amount of englacial debris, the melt-rate and the surge-free time-span suffice.

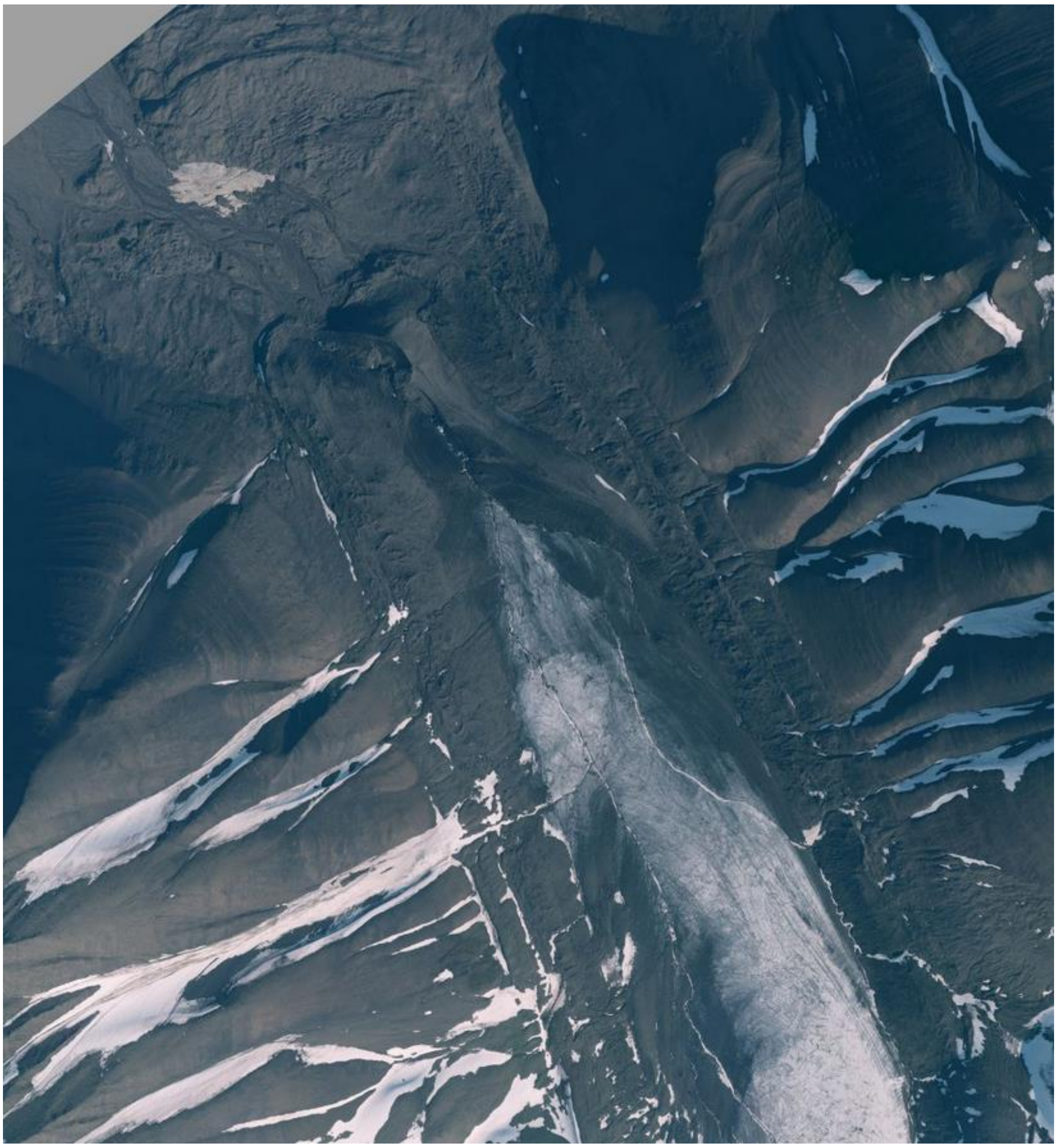

Figure 6.2: Hyllingebreen has a clear post-surge retreat area, marked by degrading ice-cored topographic highs among ice-free terrain. Yet, the glacier has since managed to form a small debris-covered front.

The proposed hypothesis does seem to hold to a certain extent. The majority of the investigated surge-type glaciers have little supra-glacial debris and debris-covered fronts are uncommon; only 1 out of 14 surge-type glaciers has an intact debris-covered front. These results must however be considered very carefully. A much larger sample size is needed to thoroughly test the hypothesis. Next to this it is difficult to compare the results to surge-free glaciers in the area, since most glaciers have not been the subject of surge-studies, meaning that their history remains unknown.

Surging occurs on all glacier types, from small inland ones to large calving, tidewater glaciers. Since surging is indiscriminate it could explain the existence of some of the small glaciers without a debris-covered margin, that although uncommon, are nonetheless present in the study area (Nordenskiöld land, Prins Karls Forland and the coastal areas of Oscar II Land). Chapter 5.2.3 'Exceptions to the rule: small glaciers with little ice-core in the marginal zones' introduced a few remarkable glaciers that did not fit the negative trend between a glaciers size and its debris-covered fraction. The glaciers that were investigated are: Holmesletbreane (Vestre & Austre Holmesletbreen), Linnébreen, Scott Turnerbreen and Tellbreen. Considering the findings of the small surge study in this thesis, it is indeed interesting that two of these four glaciers have a surge history, namely Scott Turnerbreen (Hagen et al., 1993) and Tellbreen (Lovell et al., 2015, Sevestre et al., 2015).

Since surge-studies are limited in Svalbard, due to the remote location of the archipelago, the fact that two of the clear exceptions to the trend, could be the result of surging does seem to be telling. And the other two exceptional glaciers/glacier-zones, could they have surged as well? Holmesletbreane has not been the subject of any study looking into a possible surge history. It is therefore unknown if a surge event has occurred. The final glacier, Linnébreen does appear in the literature in relation to surging. Svendsen and Mangerud (1997) found no evidence of surging at Linnébreen. It must however be noted this study focussed on dating the glacier's advances and retreats during the Holocene. Although extensive fieldwork was performed in the proglacial zone, determining whether or not Linnébreen had surged never was the objective to the study. More recent studies have repeated the statement of Svendsen and Mangerud (1997), but its validity has not been tested by any study ever since. This is worth some consideration, since changing conclusions on a glacier's possible surge history is not new to Svalbard's literature. Tellbreen for example has, by a study investigating its current thermal structure and drainage system, been defined as a non-surge-type glacier without comprehensive geomorphological analysis of the proglacial zone. A recent study that did take this geomorphological, field-based approach revealed new evidence suggesting that Tellbreen indeed did experience a surge-event (Lovell et al., 2015, Sevestre et al., 2015).

7. Introduction to an in-depth study of a Central Spitsbergen glacier: Longyearbreen

The 88 measured glaciers in Nordenskiöld Land and along the coasts of Prins Karls Forland and Oscar II Land show a clear negative correlation, and probable relation, between the fraction of the glacier that is debris-covered and a glacier's clean (free from debris cover) size. Small glaciers produce large debris-covered glacier fronts, while large glaciers create relatively little ice-core in the terminal zone. But what do glaciers with a debris-covered front actually look like? How do the seasons impact them? How are they formed and deformed and in what manner does the debris cover impact the ice below? To answer questions like these the large-scale study now zooms in on the area where we find most debris-covered glacier ice: Central Spitsbergen. The glacier Longyearbreen, which has been found to be representative of most glaciers in the area, is put through an in-depth and field-based study.

Longyearbreen is located the arid inner zone of Spitsbergen. The average annual sum in the area is about 180 mm water equivalent (Humlum et al., 2007). The highest mountain of Nordenskiöld land, Nordenskiöldfjellet, sits to the west of the glacier. Longyearbreen covers a large elevation range. It hangs of the eastern side of Teltberget, that, with a height of 1032 m, is only about 20 m lower than the highest mountain in the area. The debris-covered snout of the glacier reaches down to slightly below 200 m. Next to mountain peaks like these, the nearby area is dominated by extensive mountain plateaus, reaching a height of around 500 m. Longyearbreen itself sits in a V-shaped tributary valley. Radio-echo backscatter revealed the V-shaped channel-profile that lies beneath the ice. This indicates that the area is one of low subglacial activity with a high possibility of preglacial form preservation in the presence of cold ice (Etzelmüller et al., 2000). The glacier is characterized by a rounded, well-formed debris-covered front (figure 5.5 of the previous result section).

8. Methods specific to Longyearbreen

8.1. Geomorphological observations, mapping and debris thickness measurements

The first step of field campaign of autumn 2014 on Longyearbreen has been to precisely determine the outline of the debris-covered area. This was done in a simple yet effective way: walk along the outline and track ones' steps with a hand-held gps system (accuracy 3m) using an interval of just 1 second. The next step was to record the geomorphology of the glacier front, using the same approach. A total of 105 geomorphologically interesting sites on Longyearbreen's debris-covered front have been identified. The location of these sights was recorded/tracked with the use of a hand-held gps unit and the feature was then photographed, if needed with scale. Any noteworthy observations were recorded, including clast/sediment size, distribution and sorting, stability of the matrix and clast/pebble rounding. In this manner ridges, slumping zones, mudflows, streams, ice caves, etc. were listed and subsequently uploaded into ArcMap to create a geomorphological map of the glacier margin.

Notes and pictures were further analysed in order to understand the manner in which the structures had been formed and to discover if (and in what manner) deformation was currently taking place. For this purpose, it was of great importance that the fieldwork was executed over a relatively long period of two months, as this allowed one to think certain observations over and test preliminary hypotheses.

The thickness of the glacial debris cover was measured by accessing incised meltwater channels. This provides a simple way to measure the debris layer thickness, since both the debris layer and the underlying ice were naturally cut straight through. Before each measurement, the vertical profile containing both debris and ice was cleaned until the boundary between clean ice and debris could be clearly defined (figure 8.1). A total of 36 debris thickness measurements have been taken. The measurements were distributed as evenly over the glacier front as the topography of Longyearbreen would allow. Due to the unstable, slump-covered nature of Longyearbreen's steep frontal edge, the debris thickness could not be recorded on this slope. This area is simply to hazardous to conduct fieldwork safely. Luckily this zone is very narrow as a result of its high slope angle, meaning that only a minor fraction of the glacier front was excluded from measurements.

Methods specific to Longyearbreen

While the field work on Longyearbreen has provided a great amount of valuable information on its own, the indirect value of the work has been even greater. While the author (and field-worker) of this thesis naturally was experienced in making glacial-geomorphological observations as a result of long term field-based studies at UNIS (Svalbard), the fieldwork on Longyearbreen has certainly reinforced this knowledge and experience. By going to the glacier front day in, day out, valuable field knowledge and skills were gained, which enabled detailed interpretation of aerial images as performed in this thesis.

Figure 8.1: Example of a debris thickness measurement on Longyearbreen, the measurement stick is extended to 1 m. The picture illustrates the great need of cleaning the profile before measurements are conducted. The ice in the bottom-centre and right has been cleaned until the border with the debris layer could be determined. The ice in the bottom-left has not been cleaned and can easily be mistaken for a part of the debris cover. As a standard measure the ice and debris at the debris-ice interface has been slightly hacked away where the measurement is taken, which is done as a final check to make sure that this is indeed where ice and debris meet.

8.2. Temperature recording

Four temperature sensors were installed in a 1 m depth pit in the debris-covered front of Longyearbreen (figure 8.2). Installing loggers deep in a debris layer poses difficulties due to the stone-filled nature of the ground. Digging is extremely time-consuming. Drilling could be a less labour intensive option. Although this too is problematic, as debris-covered glacier-fronts are often actively moving and deforming, the result of which is rough terrain that can only be accessed on foot. Drills able to drill a wide core through such an unforgiving matrix of boulders, clay and gravel come dear and are hard to transport. This makes them less than ideal machinery for this study. The temperature pit in Longyearbreen was dug in the effective, but labour intensive way; a sturdy shovel, a hammer and willing hands.

A hammer and screwdriver proved to be simple yet effective tools, enabling one to precisely form the slits in which the delicate temperature sensors were inserted. The electrical wires were given plenty of slack before closing the hole. This is of great importance as the debris layer deforms quite a bit over a year long period. A wire that was loose in autumn can be pulled tight in spring solely by soil movement. A little extra buried length allows the wires to move with the soil, preventing snapped wires and resultant data loss. When closing the temperature pit, the earth was stamped down as much as possible and a little mount was left over the hole to allow for settling of the soil. Over time this mount flatten out, leaving the surface of the filled pit at the same height as the surrounding ground.

Figure 8.2: Installing the temperature loggers in Longyearbreen's frontal debris layer in September 2014. Measurement stick of 110 cm for scale. Sensors have been installed. The loggers in the yellow boxed are still uncovered.

Methods specific to Longyearbreen

In winter Longyearbreen services as a snow-mobile 'highway' between the two populated cities in Spitsbergen; Longyearbyen and Barentsburg. To avoid damage to the temperature loggers by possible snow-scooter traffic the pit was dug away from the gullies that become snow-scooter pathways in winter. Still, there is little stopping snow-scooters from driving over the temperature pit, which leaves the loggers themselves in need of protection. Since there are plenty of sturdy stones to be found on a debris-covered glacier front, it is not surprising that a few of them were used to protect the loggers and wires. A stone cairn was built next to the temperature pit to house the loggers and a small stone trail paced on top of the wires where they run from pit to cairn. As no damage was seen on the wires when they were retrieved in 2015 this manner of protection seems to have been effective.

Measurements were conducted from September 2014 to September 2015. The temperature pit was located towards the front of the debris-covered area, centred at approximate equal distance of the eastern and western mountainsides (figure 8.3). The ground temperature was measured hourly at depths of 6 cm, 16 cm, 38 cm and 90 cm. Unfortunately, only the uppermost three sensors survived the yearlong burial in the actively moving and deforming debris layer. However, the recordings of the three surviving temperature sensors do provide good insight into the process of heat transfer within Longyearbreen's debris layer.

Figure 8.3: Temperature pit at Longyearbreen marginal zone, debris layer thickness at the pit > 1m. Dismantling of the stone cairn to recover the data loggers, before opening the pit to retrieve the measurement instruments in September 2015. The clean part of the glacier, seen in the back, is blanketed by the first snow of the season.

9. Results specific to Longyearbreen

Central Spitsbergen is an area where debris-covered glacier fronts are especially common. When one wishes to further analyses the systems that act upon the debris-covered parts of a glacier, it is needless to say that a glacier from Central Spitsbergen makes a good candidate. Due to its proximity to the town of Longyearbyen, Longyearbreen is easily accessible. Logistically this makes it a good candidate for multiple months of fieldwork. But is the glacier also a good choice for this study? Is Longyearbreen representative of Central Spitsbergen glaciers?

The answer to this question can be found in the previously conducted studies, starting with the small-scale study of glaciers located within 20 kilometers of Longyearbreen (chapter 5.1). All 40 glacier termini in this area contain debris-covered ice (either in the form of ice-cored moraines, patches of debris on the glacier or as a debris-covered glacier front). 68 % of the 40 glaciers have a debris-covered snout. Longyearbreen, being one of those, fits this common glacier type of the area. The same small scale study shows that Longyearbreen is slightly larger than most glaciers within the 20 km radius (table 3), but lies well within the range. Overall Longyearbreen is about 1.3 km longer than the average of the group. This extra length comes purely from clean ice, free of glacial debris. The length of the glaciers debris-covered front is right on the spot, exactly the calculated average of all 40 glaciers. The fact that Longyearbreen is relatively long can partially be explained by its shape and partially by a slightly greater glacier size. Longyearbreen is, due to the shape of its confining valley, a relatively narrow, long glacier. All in all, the glacier lies well within the range of glacier-sizes/-shapes found in its area and the size and shape of Longyearbreen's debris-covered snout is especially representative.

Table 3: A small scale study of all 40 glaciers (including Longyearbreen) within 20 km of Longyearbreen (all of which had debris-covered ice in their terminus and 27 of which had an actual debris-covered glacier front) and how they compare to Longyearbreen. The range (min – max) is given in brackets behind each average.

Small scale study: Length	*Glaciers within 20 km from Longyearbreen Average (km)*	*Longyearbreen (km)*
Clean ice (n=40)	3 (0.80 – 5.25)	4.3
Debris-covered terminus of the glaciers (n=27)	0.6 (0.17 – 1.92)	0.6
Small scale study: Width		
Clean ice (n=40)	0.65 (0.27 – 3.04)	0.58
Debris-covered terminus of the glaciers (n=27)	0.69 (0.14 – 2.08)	0.50

To get an even broader picture, Longyearbreen is also compared to other 87 measured glaciers of the large-scale glacier study (chapter 5.2). This study does not concern length and width measurements and can therefore not give any indication of glacier-shape, but mapping of the projected area does give a much more accurate estimate of glacier-size than what can be recovered from length and width measurements. When compared to this extensive study Longyearbreen is very near to the average size of both clean glacier ice and debris-covered parts of a glacier (table 4).

Table 4: A large scale study of the 88 measured glaciers (including Longyearbreen) in the area of Nordenskiöld Land and north of Nordenskiöld Land along the coasts of Prins Karls Forland and Oscar II Land (figure 2.1 and 2.4. in the section 'Description of study areas'). The range (min – max) is given in brackets behind each average.

Large scale study (n=88): Area	*Glaciers in and north of Nordenskiöld Land, and along the coast of Prins Karls Forland and Oscar II land (km^2)*	*Longyearbreen (km^2)*
Clean ice	2.72 (0.02 – 14.40)	2.11
Debris-covered parts of glacier (includes all debris-covered ice that is not detached from the clean glacier ice)	0.60 (0.01 – 1.55)	0.88

Length and area analysis show that Longyearbreen, although not being the spot-on average, well represents the glaciers in both its immediate and its broader surroundings. But of course, there are more aspects in which glaciers can differ. Wind-dependence and ice-mass orientation for example can define certain glaciers and be less important to others. The earlier performed study on this subject (chapter 5.2.6) showed that in general glaciers in central and western Spitsbergen are wind-dependent. It is therefore highly relevant to look into the wind-dependency of Longyearbreen. Figure 9.1 portrays the glacier orientation in Central Spitsbergen, based on the small-scale study of 40 glaciers within 20 km from Longyearbreen. The red arrow that represents Longyearbreen is located right at the mean glacier orientation and could not be directed any better towards the so-called wind-gap of the area. With respect to glacier orientation and wind-dependence Longyearbreen is an ideal candidate to represent the other glaciers studied in this thesis. Of course, it must be noted that all variables tested here show large ranges. The representativeness of Longyearbreen thus certainly does not mean that a small valley glacier is equal to a large coastal glacier elsewhere on Svalbard. It is the average, the mean signature of all studied glaciers that is of interest. And to conclude, it can be answered that yes, based on size, shape, orientation and wind-dependence, Longyearbreen can indeed be considered representative of the larger spectrum of glaciers studied in this thesis.

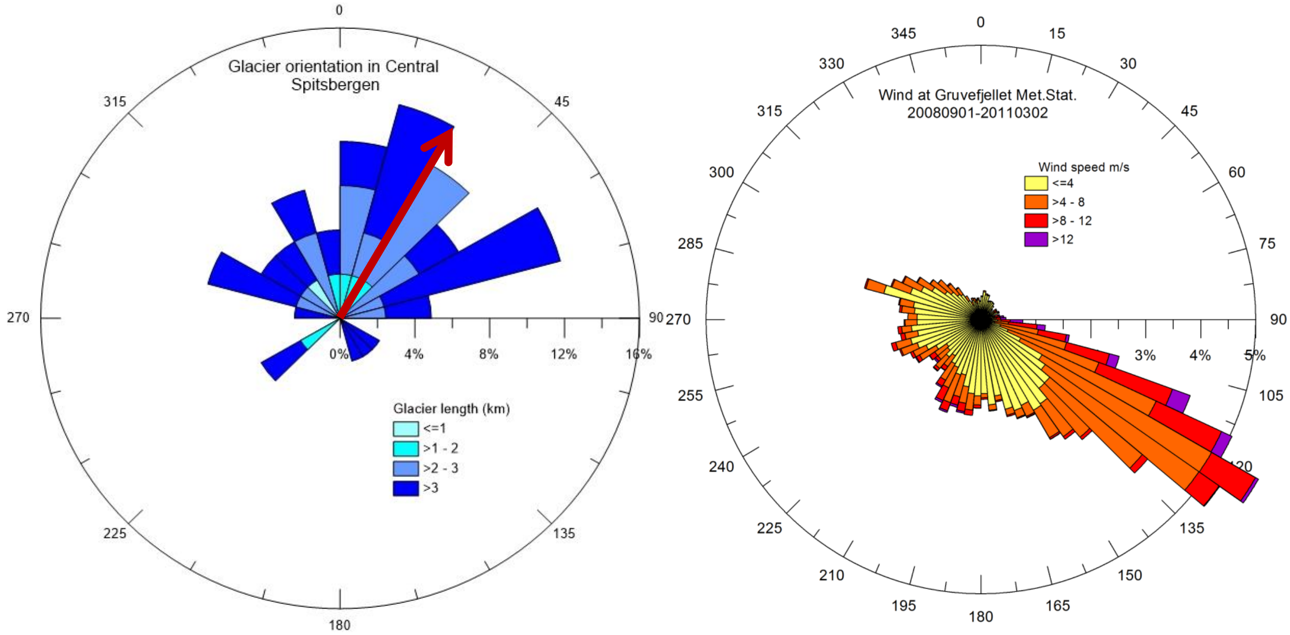

Figure 9.1: Left: Orientation of Longyearbreen (red arrow) compared to the orientation of all other glaciers within a 20 km radius of Longyearbreen. Right: Wind direction and speed measured at Gruvefjellet, nearby Longyearbreen. Gruvefjellet meteorological station is located on a mountain-plateau and thus not influenced by local topography.

9.1. Formation of Longyearbreen’s debris-covered terminus

Previous research at Longyearbreen has brought forward valuable information with regard to the formation of the glaciers debris-covered front. These results revealed much about Longyearbreen’s subglacial erosive properties, which can be a great source of debris that may later-on help form a glacial debris cover. Information on Longyearbreen’s subglacial erosional properties are therefore highly relevant to this study.

Etzelmüller et al. (2000) performed radio echo soundings to reveal Longyearbreen’s subglacial profile. Interestingly the subglacial valley profile beneath lower Longyearbreen turned out to be V-shaped. This indicates that Longyearbreen has hardly eroded it’s bed, which would have caused a more U-shaped valley profile (Etzelmüller et al., 2000). It does look like basal sliding at Longyearbreen is either absent or minimal. Indeed Etzelmüller et al. (2000) showed that there is slight basal sliding during summer, but the flow velocities remain low; in the lowermost ablation area, flow velocities decrease to below 1 ma^{-1}. Higher up on the glacier frozen in situ soil and vegetation has even been found below the cold-based glacier ice. Dating of the relict vegetation indicates that, at least at this location, the glacier has not eroded its bed for at least 1100 years (Humlum et al., 2005). These findings are strongly in favour of a primarily supraglacial sourced debris cover at Longyearbreen. Although it must be noted that a small portion of less angular clasts, a few of which were striated, have been found at Longyearbreen, which indicates that to a small extent subglacial transport and/or glacial-river-reworking has taken place in the past (Lukas et al., 2005). All in all, the debris-covered ice-margin appears to dominantly have been formed by supraglacial processes, while subglacial erosion may have played a minor role.

Since Longyearbreen hardly erodes its bed, moraine building relies strongly on external debris input from the slopes above the glacier. These slopes are highly unstable and form an easily available debris source if plucked by slope processes like rockfall, avalanches and debris flows. There is little data on rockfall and debris flows at Longyearbreen, but avalanches, some of which terminated on the glacier, have been monitored in the area. Avalanching is held responsible for moraine formation at Larsbreen, located within one kilometre from the study site (Humlum et al., 2007). Eckerstorfer and Christiansen (2011) showed that avalanches are common on the slopes around Longyearbreen (figure 9.2).

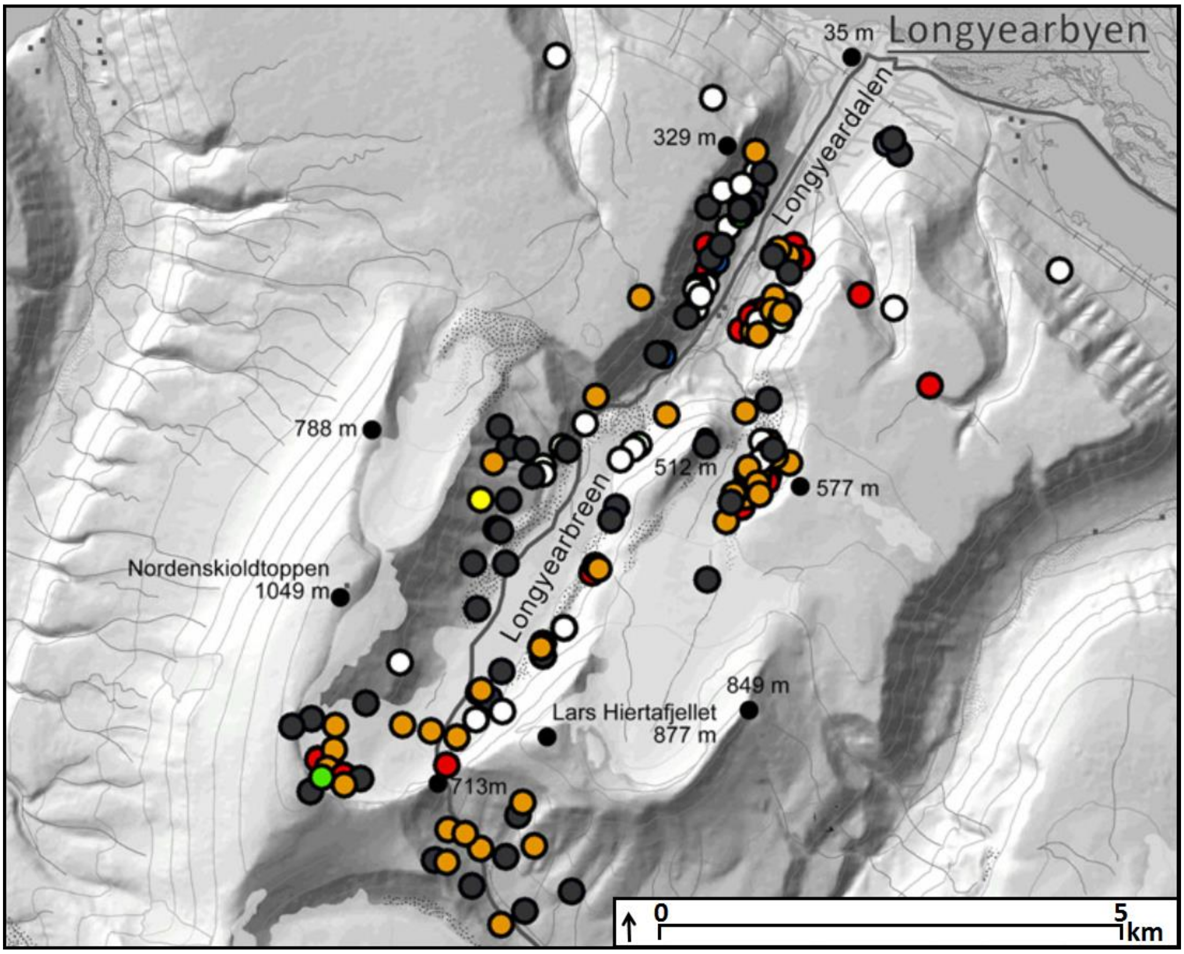

Figure 9.2: Avalanches with a volume of 100m^3 or larger, spotted during the 2006–2009 observation period. Colour coding of the avalnches: Grey = slab avalanches, Red = cornice fall avalanches, Orange = cornice fall / slab / loose snow avalnches, Green = loose snow avalanches, Yellow = sluch flows. Modified from Eckerstorfer and Christiansen (2011).

Eckerstorfer et al. (2009) monitored the avalanches in the area between 12 October 2008 and 25 May 2009 and observed a total of 36 avalanches in the period from 11 March to 14 May, released along the snow mobile track from Longyearbyen over Longyearbreen and further through Fardalen, with most avalanches observed on Longyearbreen (47 %). Depending on the runout zone, avalanches could be a debris source at Longyearbreen. This is in line with the conclusions earlier drawn by Lukas et al. (2005), who stated that the debris-covered ice-margins were dominantly formed by the incorporation of rockfall and avalanche material sourced from free faces overlooking the glacier in their source areas.

During the fieldwork on Longyearbreen, recording avalanches has not been the head objective of this study, but when one is in the field and spots an avalanche it is off course recorded. Most of these did not flow onto the glacier itself, but avalanches near the headwall of the glacier do usually spread out over the glacial ice. Often one or more avalanches could be spotted coming down the headwall of the glacier towards the start of summer (season 2015 and 2014). In addition to this each year's melt season is accompanied by debris input from the headwall in another way, that is perhaps best described as minor rock fall events. Figure 9.3 shows the situation at Longyearbreen's headwall at the 30th of May 2015. Two processes that may release debris in/onto the glacier can be seen in this photograph. First of all, a fresh avalanche has just released from the headwall (this is hard to see against the low arctic sun, but snow blocks rolled down all the way onto the flattened part of the glacier). The appearance left by the second process is best described as brownish fan-shaped strips, high up on the steep face of the glacier. Off course the colour alone already indicates that more than just snow was being transported here, even if it could not be called a traditional avalanche or a rockfall.

Although a full avalanche or rockfall study was not incorporated in the fieldwork of this thesis, the findings of these studies are in line with the results of previous studies. During the field seasons of 2014 and 2015 both avalanches and (minor) rockfall events indeed showed to be common on Longyearbreen. These events happened on the side slopes, where they usually have little impact as the glacier is often not reached, but also commonly took place at the glaciers headwall. These headwall releases did reach the glacier itself and paved the way for subsequent glacial debris transport to the glacier front.

Like Lukas et al. (2005), this study's fieldwork in 2014 and 2015 too revealed that not all debris on Longyearbreen's front is angular. Clasts are definitely far from rounded, and, if not compared to the

Figure 9.3: Supraglacial debris transport at Longyearbreen's headwall during the seasonal melting of 2015. Encircled is a freshly released avalanche. During the following month, the avalanche turned brownish, revealing the sediment it had carried. The arrow points towards one of the many minor 'rockfall' events that can be seen at the glacier's headwall each year.

mountainside debris on the slopes surrounding the glacier, can easily be called angular. But this is exactly the point: compared to the mountainside debris some of the clasts on the glacier are more rounded. Clast analysis performed by Lukas et al. (2005) has classified the debris on Longyearbreen's margin as dominantly angular, with the entire debris mixture covering a range from very angular to sub-rounded. This is not to say that extensive subglacial transport must have taken place, most debris still is too angular to suggest that such transport has been important. And although previous research has discovered striations on some of the more rounded clasts, no striated clasts have been found during the fieldwork of this study. Some reworking however has taken place. Interpretation section 10.1, 'Glacial debris transport: What formed the debris cover?', aims to answer this question with the use of a simple theory based on sights photographed on Longyearbreen as part of the fieldwork in the autumn of 2014.

9.2. Current state of Longyearbreen's debris-covered terminus

Whether it is by rockfall, avalanches or the occasional debris flow, debris clearly makes its way to the terminus area of Longyearbreen. During the field campaign of 2014, 36 debris thickness measurements have been taken. This showed that the snout of Longyearbreen is covered with a debris layer of on average 46 centimeters (n=36), although the layer is far from even. At one instance the debris cover thickness went from 38 cm to 2 m within a horizontal distance of 3 m. The only feature separating the two observation points was an incised glacial channel. The 36 thickness measurements ranged from 2 to 200 cm. The debris layer thickens towards the glacier front and thins to just a few cm further up-glacier; at the border between debris-covered ice and bare glacier ice.

During the field campaign of autumn 2014 a total of 105 geomorphologically interesting sites on Longyearbreen's debris-covered front have been identified. The location of these sights was recorded, the feature mapped and photographed, if needed with scale. Longyearbreen's debris-covered front is highly irregular as a result of deformation. Common geomorphological features include ridges, mudflows, slump headwalls and fans and incised streams. With the use of field based gps measurements a basic geomorphological map of the glacier front has been created (figure 9.4). The following chapter 'Deformation of Longyearbreen's debris-covered terminus' will further analyse these features and the deformation processes that shape them.

As mentioned before most of Longyearbreen's debris-covered glacier front shows a distinctly uneven topography. Debris is found as a layer covering the ice, but also as single clasts stuck in the otherwise clean ice or as debris layers frozen into the ice. Clasts are predominantly angular, with some slightly more rounded finds. The ground is bumpy and full of ridges. However, these topographic highs generally do not indicate a thickening debris layer but merely reflect the topography of the underlying ice. This ice core also lies at the base of the many slumps, and tension cracks that characterize the debris-covered glacier front. A widespread channel network of sharply incised streams transports meltwater through the glaciers front. Streams may disappear and reappear into and from the debris-covered ice (figure 9.5). This describes the general picture of Longyearbreen's margin, but one place on the glacier front shows a very different picture, even though no change could be discovered in the glacial ice underneath the debris. The two light-blue framed areas in figure 9.4 consist of relatively flat even ground and are shown in figure 9.6. The sediment is relatively well sorted and shaped sub-angular to sub-rounded. This presumably is the result of frequent shallow flooding. During and after heavy rain showers or melt events these flat areas would often be flooded. These events were characterized by slow but continues water movement. In the flooding-zones one can find fluvial layering, ripple-marks, and a (for Longyearbreen) well sorted state of debris (figure 9.6). This indicates that these relatively gentle flooding events are not new to 2014, when they were recorded. These floodplains are exceptional. The only other areas in Longyearbreen were such indicators of high fluvial forcing can be seen is not on top of- but inside the glacier. Englacial debris layers will often to a certain degree display sorting, rounding, layering and a common debris orientation.

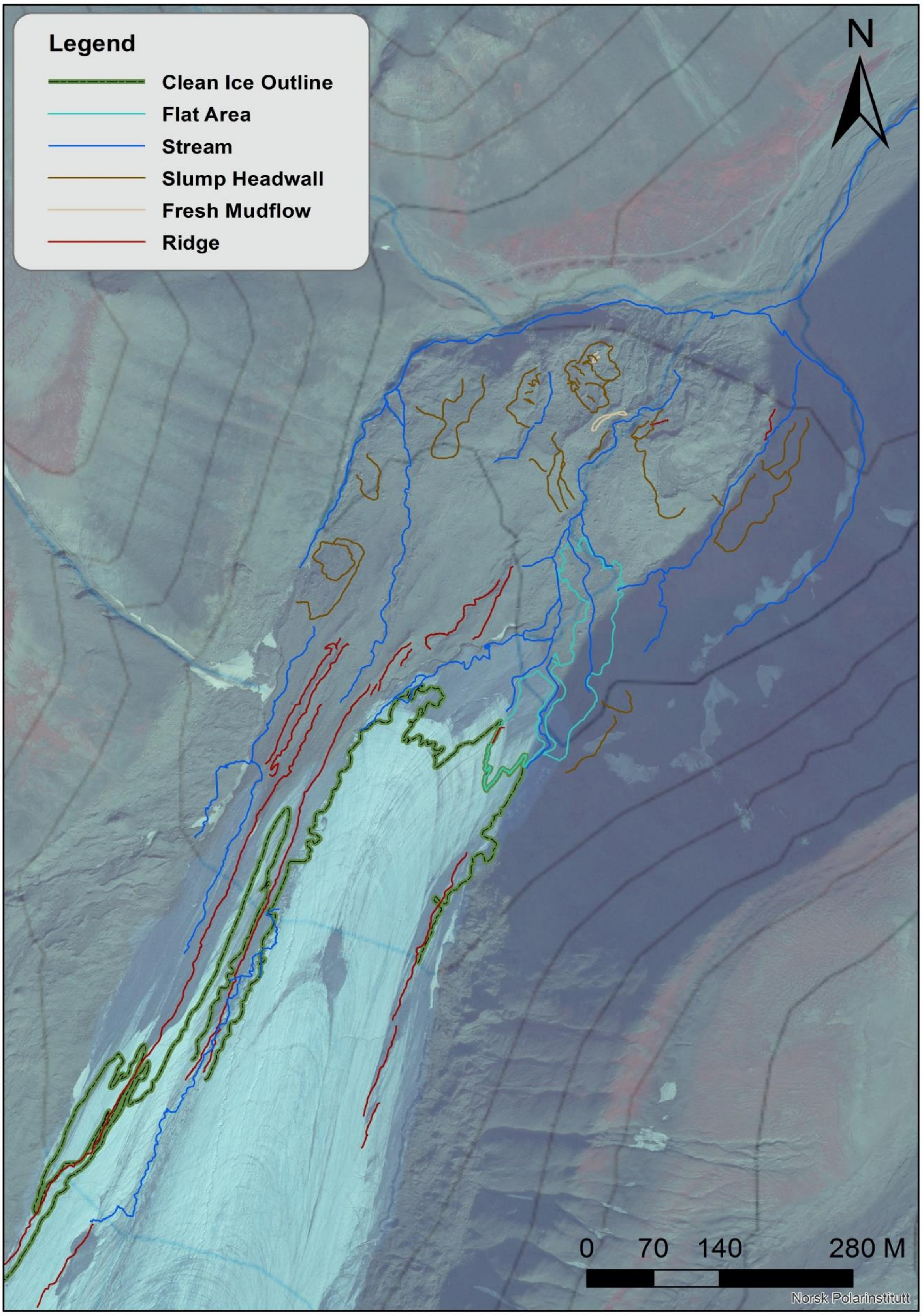

Figure 9.4: Thematic geomorphological map of Longyearbreen's terminus. Measurements were taken during the autumn of 2014. Gps-tracked features are projected on an aerial image of the glacier from 2009 overlaid with a topographical map. In 2014 the debris cover has slightly spread up-glacier. Slumps are rampant and occur in both new and former slumping zones. At slumping zones that are active in both 2009 and 2014 the headwall has retreated. When mapped rivers disappear at the frontal edge, it was too hazardous to track the stream. Naturally the stream continues downslope.

Figure 9.5: The western lateral channel is shown from above and inside (D. Roehnert for scale) Longyearbreen. Like all meltwater channels on this glacier, it is sharply incised. This lateral channel is the deepest cut supraglacial stream on Longyearbreen's front. Notice the debris on the channel floor and the dirty ice layers: debris is not just reserved to the glacier surface. The top picture shows the channel splitting up. The right channel-branch will continue supraglacially, but the left branch disappears into the debris-covered ice to continue its way englacially. A few hundred meters further up-glacier the channel surfaces again for a short distance until it once again dips under the surface.

Figure 9.6: The left top photograph gives an overview of one of the so called ‘flat areas’ on the debris-covered margin of Longyearbreen. The plain is covered by an ice layer that formed when temperatures dropped after a flood. To the right of this picture is a vertical profile of the same floodplain (measurements tick extended to 60 cm) that contains layering and ripple-marks in fine sediment. What also stands out is the common orientation of most pebbles, aligned parallel to the surface of the flood plain. The lower most picture shows the flood plain’s surface. The debris, that consists largely of pebbles, is extremely well sorted compared to what is generally found on Longyearbreen. Boulders and clasts greater than 0.1 m are uncommon. Furthermore, the pebbles are generally sub-angular to sub-rounded, with a few more rounded examples. All of these findings are indicative of long term fluvial activity in the form of flooding.

9.3. Deformation of the debris cover on Longyearbreen's terminus

In figure 9.4 the geomorphological features like streams, ridges, slumps, etc. are projected on top of a georeferenced aerial photograph of the glacier front from 2009. The data of 214 has purposely been projected on a five year older areal image as this enables one to quickly see to what extent the debris-covered front has changed during the time span of a few years.

The overall shape and form of the glacier margin, has, as one might expect, not visibly changed over the five-year period. However, looking at the geomorphological map, almost all before mentioned geomorphological features seem to be actively changing. A few slump headwalls that existed in 2009 are clearly located higher up-glacier in 2014, allowing for a larger slump fan. To the western side of the glacier front multiple new slumping zones have formed. Although some headwalls retreat faster, and thus more visibly than others, all seem to actively move up-glacier. From August to October 2014 headwall retreat was visually observed on all slumping zones. Figure 9.8 shows one of the many headwalls on Longyearbreen retreated during the autumn of 2014. The process of headwall retreat generally starts with a narrow tension crack that forms behind the headwall (figure 9.7).

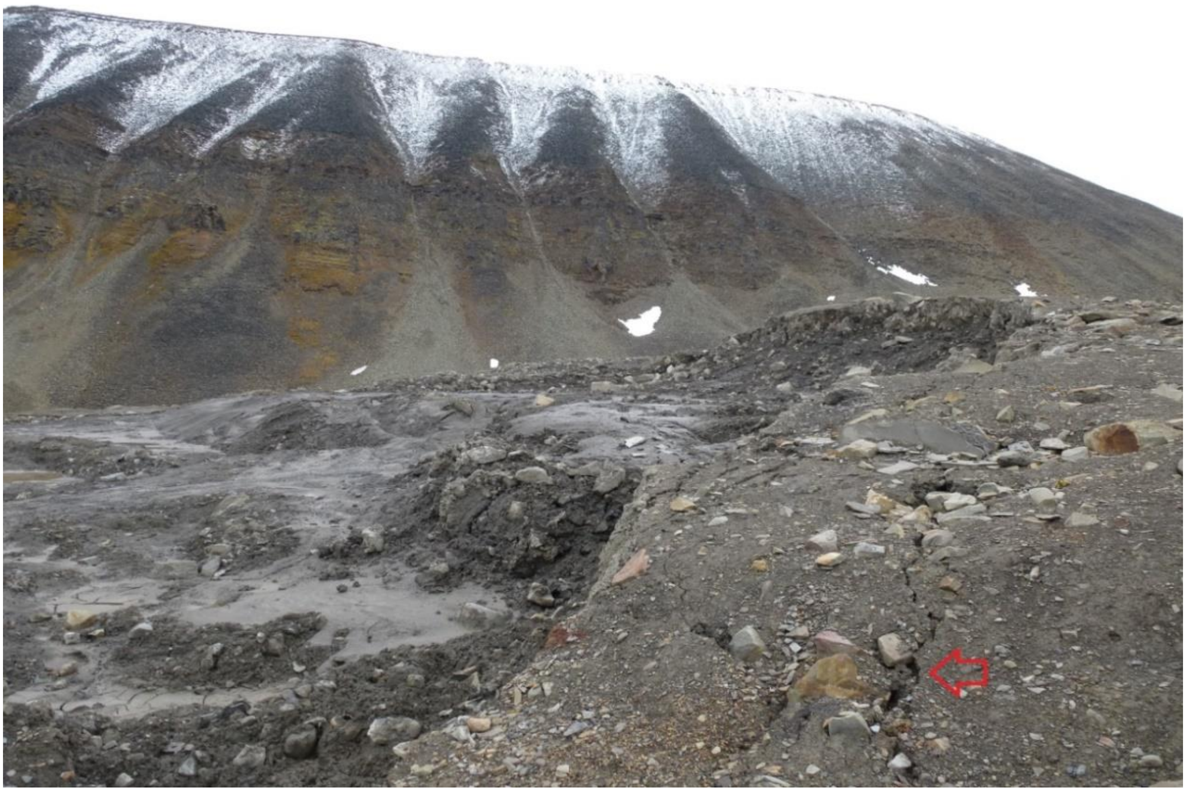

Figure 9.7: An active slumping zone on Longyearbreen in September 2014. Muddy soil, released after the previous rainy day, is settling. Meanwhile a new extension crack (highlighted by a red arrow) widens, ready to become the new slump-headwall.

This crack slowly widens and deepens, until, usually during or shortly after a large rain event, the soil collapses and the backwall of the former tension crack becomes the new headwall. It is important to realize the scale on which this happens. Active slumping zones cover more than half of the steep edge of Longyearbreen's front. The headwalls of these slumps are up to 150 meters in length (this measurement only includes the backwall, when both side walls are taken into account the headwall length would be even greater). During the freezeback period days of sudden above-freezing temperatures and rain where common which caused the slumping zones to be especially active. The activity and sizes of these headwalls make slumping, on a small and narrow glacier like Longyearbreen, a major deformation process.

As mentioned before, slumping zones are not the only actively deforming features on Longyearbreen. The many ridges (figure 9.4) for example are loose and unstable unless they are frozen. They are also subject to deformation due to melt of the underlying ice core, for the ridges on Longyearbreen are not a result of a thicker debris cover but simply reflect the topography of the ice. It is therefore perhaps not surprising that some ridges that are very pronounced in 2009 are not easily retraceable in 2014. And likewise, the fieldwork of 2014 revealed two ridges that cannot be detected on the areal image of 2009. The larger scale ridges however seem to be more stable and follow the same track in both 2009 and 2014. This is consistent with field observations.

Along meltwater channels deformation of the adjoining debris cover takes place on a major scale. Roof and side collapses have been observed along all mapped channels. As was found earlier for the slumping, areas these events too were especially common during and straight after mid-freezeback rain showers, which resulted in massive, cobble carrying, first-day-of-spring like floods. At the centre of the covered glacier front, at the edges of or in the 'flat areas' (figure 9.4 and 9.6) ice melt and the resultant collapse of underground channels was especially common. In this manner, previously undetected underground channels would expose itself and sometimes become a part of the visible, supra-glacial, channel system, thereby widening and deepening the supraglacial channels. After a major rain event one channel in the central 'flat area' of the glacier margin even had to be re-mapped since it had changed its route dramatically as a result of ice core collapse. Figure 9.8 shows a less extreme and more common partial channel-roof collapse that occurred towards the end of September 2014 at the western lateral channel, relatively high up the glacier front.

In meltwater channels that cut down through the glacial ice one can often find debris layers in-between otherwise clean ice. The englacial debris bands generally show signs of sorting, rounding, layering and a common debris orientation (right-hand picture in figure 9.8). Larger rocks are also found incorporated the ice. These rocks may be aligned or grouped together, but larger clasts (greater than

20 cm) usually show no apparent pattern. Behind these observations of debris inside glacial ice lies another process of deformation. Water movement shapes and sorts the debris. Smaller clasts, pebbles and sediment may be carried by the stream until the water level and/or velocity drops. At this point it can settle and form the debris bands that can be found in the ice today. Larger clasts are unlikely to be transported but are nonetheless shaped and rounded by the collision with smaller particles in fluvial transport. This transporting and shaping of debris influences the external (by debris dropping or melting into a channel) and internal (inside former and current englacial channels) characteristics of the glacier front.

The deeply incised meltwater channels of Longyearbreen's debris-covered terminus enable inspection of a vertical profile of the glacial ice below. This does not only reveal information about fluvial forced debris transport and abrasion, it also discloses the structure of the ice (left picture in figure 9.9).

Figure 9.8: Partial channel roof collapse at the western lateral channel on Longyearbreen's front. The collapse incorporates both glacial ice (a red arrow highlights a block of glacial ice, the bending ice layers that are present in the intact glacial ice can still be recognised on this detached block) and supraglacial debris (blocky boulder highlighted by orange arrow). Backpack and rifle for scale to the bottom left of the collapsed material (encircled).

Debris layers and dirty ice bands presumably have been deposited in a rather surface in the uppermost part of the glacier, either by river transport or the accumulation of snowfall, avalanche material, rockfall remnants or debris flows. When debris layers or sediment rich ice layers therefor do not follow a relatively straight horizontal line ice-deformation likely is to be the cause. Realizing this, it is rather impressive that the fieldwork of 2014 showed that practically all ice-revealing channel sides on Longyearbreen are made up of inclined and/or curved ice and debris layers (figure 9.9). Ice deformation in Longyearbreen is so common that finding a not-deformed ice section is great job indeed. This rampant ice deformation definitely contributes to the highly uneven and irregular topography of Longyearbreen's terminus.

Although a variety of geomorphological information of Longyearbreen's glacier front has been gained by this fieldwork, the main result of the fieldwork is perhaps not the great acquaintance with Longyearbreen, but rather a better understanding of and familiarization with the deformation processes that take place on debris-covered ice. The fieldwork at Longyearbreen lies at the base of the extended investigation of 164 glaciers and their terminal zones in Nordenskiöld Land, Prins Karls Forland and Oscar II Land.

Figure 9.9 Left: Bended sediment rich ice layers in the western lateral channel of Longyearbreen's front show that the ice has undergone deformation (rifle and walking pole for scale). It is however important to note that the vertical dark bend does not contain much sediment, but consists of clear and air-free ice. Presumably this the infill of an old crevasse. Right: Sorting and layering of fine material in an englacial debris layer. Measurement-stick of 40 cm for scale.

9.4. Temperature recording: Heat transfer in the debris layer

9.4.1. Background

From September 2014 to September 2015 temperature sensors measured the ground temperature with an hourly resolution at 6, 16 and 38cm depth (the sensor at 90cm depth did not survive a yearlong burial) in the debris covering lower Longyearbreen. These measurements were conducted in the frontal zone of Longyearbreen on what is commonly and mistakenly called the 'moraine', centred at approximate equal distance of the eastern and western mountainsides (figure 8.3 of the method section). The debris layer at the location of measurement is slightly thicker than 1 m. The aim behind the gathering of ground temperature data is to gain a better understanding of the insulating properties of supra-glacial debris.

Measuring ground temperature at different depths enables one to visualise the process of heat transfer throughout the debris. Heat transfer in the debris layer is of great interest since it is this layer that insulates, and thereby protects the glacial ice below. Temperatures of supra-glacial debris are of course related to air temperatures. The response of a debris layer to air temperature changes is what decides how well a layer can protect the ice below. By analysing the ground temperatures and plotting them against hourly air-temperature data from the nearby Adventdalen meteorological station the response time of each depth is visualised. The delayed heat transfer within Longyearbreen's debris layer is further analysed by comparing the temperature changes and gradients at different depths.

Since the data series at Longyearbreen runs from September 2014 to September 2015 the impact of water phase changes can be analysed. The insulating properties of a debris layer will change as its properties change. In other words, a matrix with ice-filled pores acts differently from a matrix with water filled pores. And when this same debris layer (partially) dries out, creating a matrix with both water filled and air filled pores, heat transport within the layer will once again alter. The ground temperature data series at Longyearbreen is analysed with the impact of seasonal changes in mind.

9.4.2. Data analysis

Figure 9.10 shows the temperature in the debris layer at 6, 16 and 38 cm depth. Ground temperatures in Longyearbreen's debris layer clearly mimic the air temperature record. Even short melting periods impact the debris layer. Changes in air temperature are reflected in the ground

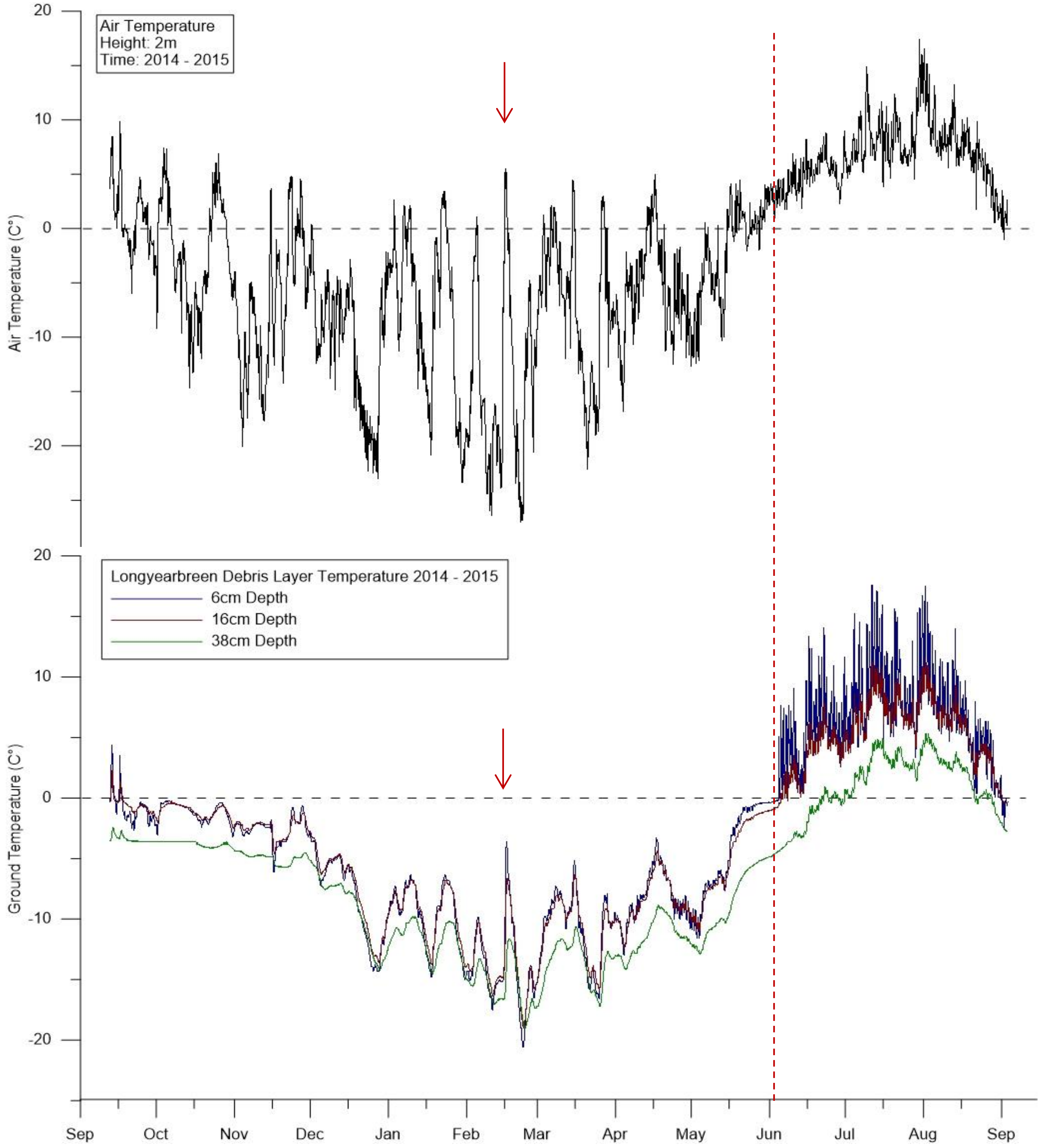

Figure 9.10 Air temperature (top graph, measured at Adventdalen meteorological station) and the temperature recordings of the three surviving ground sensors in Longyearbreen's debris layer. Ground temperature variability is most pronounced in the uppermost layer and gets dampened further down. Right after the onset of summer (red dotted line) and thus snowmelt a sudden shift takes place; all sensors, but particularly the uppermost ones, show a much more variable signal. The ground sensors clearly mimic the air temperature record and even short periods of positive degrees have impact on the debris layer, although the signal is dampened deeper in the debris (the red arrow shows an example). This dampening is more pronounced in summer than in winter.

temperature record. The strongest effect can, as one would expect, be seen in the upper most layer, while the alteration in ground temperature is present but dampened further down. An example is shown by the red arrows in figure 9.10 which highlight an abrupt temperature change in both the ground and the air temperature. This highlighted event is however nothing special. Air temperatures, especially during winter, vary often and strongly in Svalbard (Nordli et al., 2014). Although the ground temperatures at Longyearbreen's debris layer follow the air temperature record, both differ in range and seasonal variability. The ground temperatures, within the measured layer of 6 to 38 cm depth, cover a range from +18 °C to -21 °C. The air temperatures range from +17 °C to -29 °C. Seasonally changes, both internal and on the surface (snow cover) are likely (partially) responsible for this. This will be further discussed in chapter 10 'Interpretation and discussion specific to Longyearbreen'. The presence of a snow layer may, by acting as a buffer, also cause the minimum ground temperature to differ from the minimum air temperature.

The snow cover on Longyearbreen's front likely has great impact on the temperature of the debris and ice below. In early May 2014 (when the snow pack has reached its maximum thickness) a few days of fieldwork have been spent to obtain a general picture of this snow layer. This was done by probing for the snow-depth and analysing snow-pits using a standard snow density gauge. The average thickness of the cover was 0.8 m (n=647), but the snow pack was highly irregular. Topographic highs would sometimes be completely free of snow while miniature valleys were filled with up to 3.2 m of snow. The snow pack was rather dense with a water fraction of 36 % (n=4). This value is in line with the wind packed state of the snow.

The onset of thawing brings a time of change. One interesting change deals with how the near surface ground mimics the air temperature. In a thawed state the ground temperatures in the topmost layer tend to vary much more than the air temperatures, while for below-freezing ground temperatures the opposite seems to be true; the air temperature is more variable than the ground temperatures. This means that, at least for the summer situation, the ground temperature cannot be linked to air temperature alone. Another effector, likely related to air temperature since the ground temperature does mimic the air temperature, must play a roll. Solar radiation is a strong candidate for such an effector. Although radiation data is sadly unavailable for this study, the hold solar radiation has on the ground temperature has been well documented in previous studies (Nikita-Martzopoulou, 1981, Lacelle et al., 2016). Longyearbreen's debris-covered summer surface is snow-free and thus directly exposed to solar radiation. This snow-free debris surface also has both a high albedo and constant excess to solar radiation due to the midnight sun and is therefore prone to warming up.

The shift from freezing to thawing temperatures brings a scene of more abrupt changes. To further see what happens within the debris layer at this time of change one may look at a variable that displays temperature change as a function of depth: the vertical temperature gradient. The temperature gradient utilises a simple number to show how the temperature changes with depth. A negative number indicates that the ground temperature decreases with depth. If the temperature gradient is positive the ground temperature increases with depth. As the soil measured on Longyearbreen is underlain by ice and surrounded by permafrost, in an overall warming climate, one would expect the mean temperature gradient in the deeper layers to be negative. The lower graph of figure 9.11 show that, at least for the season 2014 -2015, this is the case at Longyearbreen. The shallower layer (upper graph in figure 9.11) is very responsive to changes in air temperature. Positive temperature gradients are not uncommon here. They may occur after a warm spell, just as the air temperature plummets again. Due to delayed heat transfer the deeper parts of the measured layer are still warm while the top already responses to the present cold weather, resulting in a positive temperature profile. Although such behaviour is usually reserved for the shallow ground, even the deeper layers can, in short and rare spells, warm up to such a great degree that the temperature gradient inverses (figure 9.11).

The pattern drawn in figure 9.11 shows a sudden change at the onset of summer, not unlike the change in ground temperature variability seen in figure 9.10. It can perhaps be expected that a seasonally induced increase in temperature variability reflects on the temperature gradient. What is of interest is the distinct difference in temperature between 6 – 16 cm depth and 16 – 38 cm depth. This dampening effect can off course be studied best when the soil is actually exposed to great variations in temperature: in summer. And indeed, as summer comes, the top layer (16-6cm) shows great short term variability, while the lower layer, though impacted by the arrival of summer, paints a less extreme picture (figure 9.11). This difference between the seasons is striking. In winter the pattern drawn by the temperature gradient in the lower ground is very similar to that of the shallow layer. In summer, the signal of both layers changes more rapidly than in winter, however, the pattern of the temperature gradient between upper and lower layer differs greatly. For the upper layer the average range of the varying temperature gradients is much larger in summer than in winter. For the lower soil this is not the case.

It is important to realize that this dissimilarity is caused by a minor depth difference. All measurements were recorded within a soil pack of 32 cm. Even within this small depth change the variability of the temperature gradient is greatly dampened in summer. This dampening effect could already be seen in winter, but as summer comes seasonal changes in soil characteristics amplify the

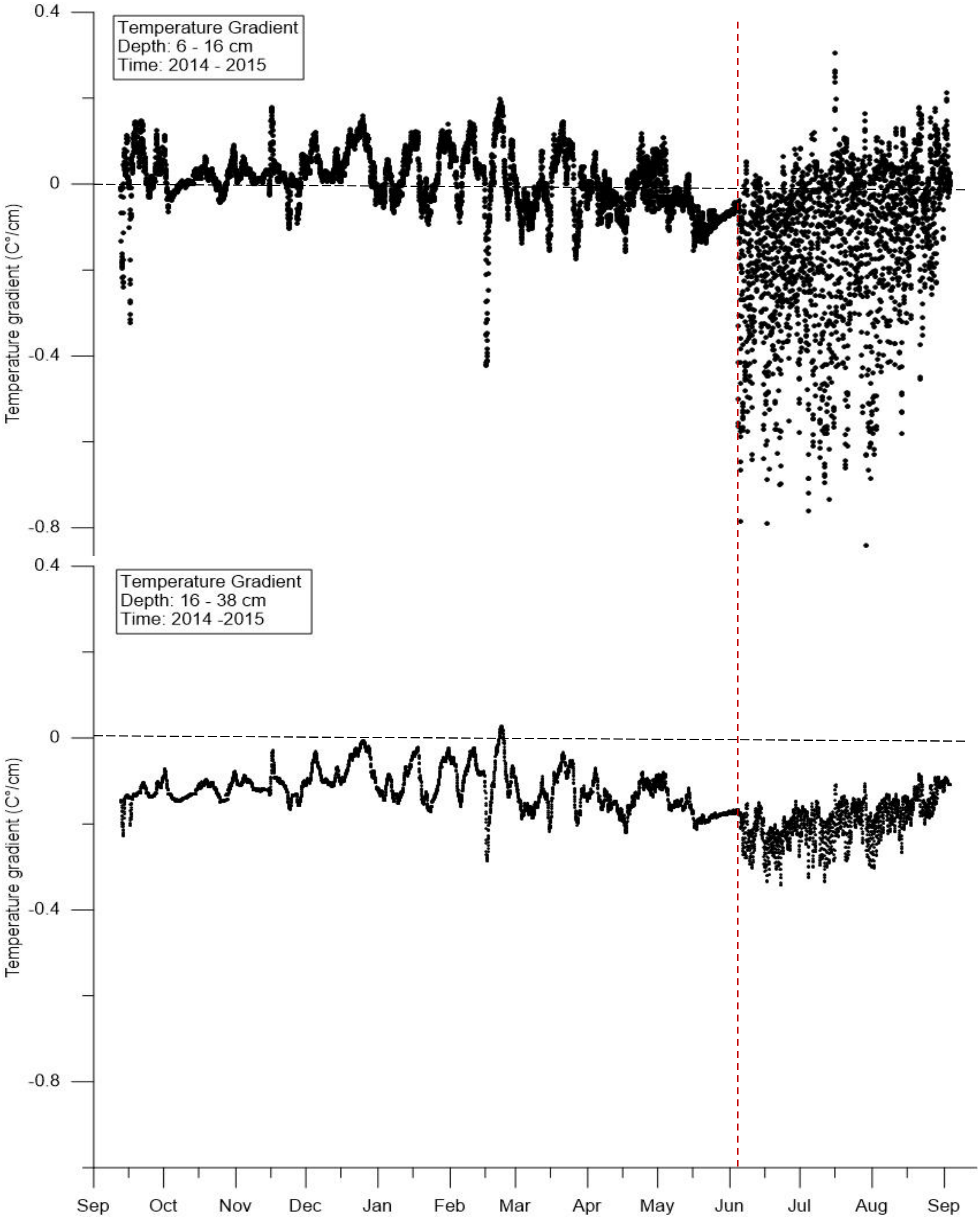

Figure 9.11: When a temperature gradient reads 0 °C, there is no heat transport by conduction. Any temperature gradient other than 0 indicates that the heat transfer within the soil takes time and deeper layers can only respond after shallower layers have adjusted to the current air temperatures. If the temperature gradient is negative the ground temperature within the measured layer decreases with depth. For positive temperature gradients, the ground temperature increases with depth. Positive temperature gradients may occur after a warm spell when the air temperature plummets again. Due to the delayed heat transfer within the soil deeper ground is still warm while the top layer already responses to the present cold weather. At the onset of summer a sudden shift takes place (red dotted line). Especially the top layer (16-6cm) shows much more short term variability. The temperature gradient is able to change much more rapidly in summer than in winter. This is likely linked to the presence/absence of an exposed snow layer, more information on which can be found in chapter 10 'Interpretation and discussion specific to Longyearbreen'. The pattern drawn by the temperature gradient in upper and lower soil layer is quite similar during winter. But during summer the variation shown by the lower layer is much less extreme than that of the upper layer. It seems that heat transport through the debris covering Longyearbreen is less effective in summer than in winter.

difference between the shallow and deeper layer. It seems that heat transport through the debris cover on Longyearbreen is less effective in summer than in winter.

Figure 9.12 is a composite plot of all 8540 temperature observations (hourly recording) in the uppermost part (6 and 16 cm depth) of the active layer. The plot visualises the general control of the near-surface (6 cm depth) on the shallow depth debris (16 cm depth). A very pronounced change can be seen at the shift from frozen to thawed ground (highlighted by the dotted lines and zoomed in box in figure 9.12). In winter the data points are aligned at roughly 45°, but in summer there is a cloud of data points rather than a line. Since both axes of figure 9.12 have the same variable (ground temperature) and are measured at the same place at the same time, the sole factor that can cause discrepancy between the x and y values is soil depth. The near-surface ground temperature is controlled by the

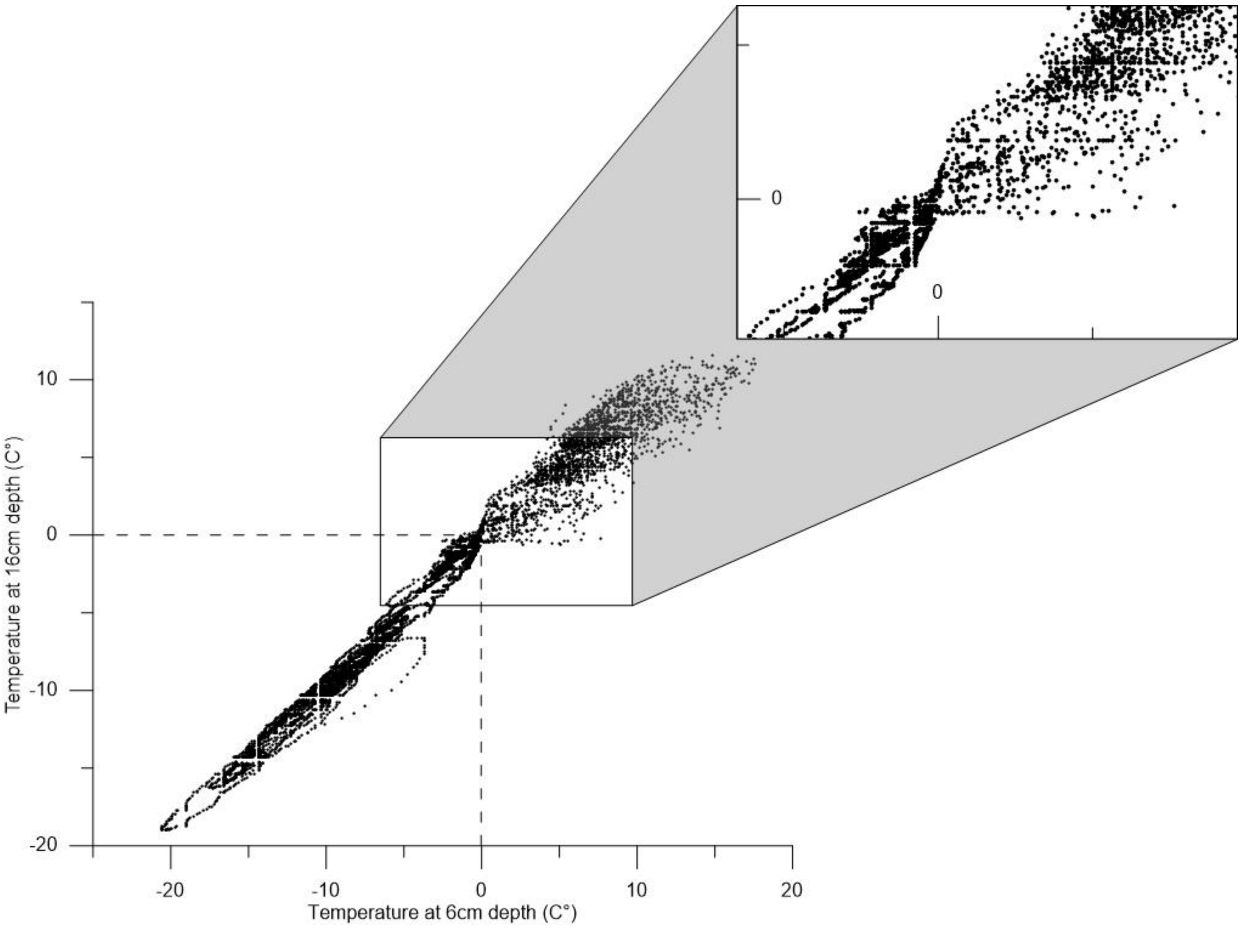

Figure 9.12: At the phase change from ice to water highlighted by the dotted lines and zoomed in box) there is a clear shift in heat transport within the 16 – 6 cm layer. If the data points are aligned perfectly on the 45° angle, heat transport between 6cm and 16cm depth takes place instantaneously (the temperature gradient is 0). The further the points differ from the 45° angel, the more time is involved transferring heat from one layer to the other. Thus, the plot shows that the insulating capacities of the soil within the 10 cm thick shallow layer (from 6 to 16 cm depth) are greater in summer than in winter. The white cross like shapes that may be noticed in the lower left of the plot have no physical basis. They are the result of the temperature sensors being slightly biased (by up to 0.5 °C) to certain values. While this does not have severe impact, it does produce slightly odd looking crosses in the plot for certain temperatures.

air temperature and solar radiation. The ground temperature further down is controlled by the temperature of the ground above, and so on. What can be read from figure 9.12 is thus that at freezing ground temperatures the soil at 6cm depth effectively controls the temperature further down at 16cm depth. After the seasonal melt has set in, the data points are much more scattered. Clearly the 6cm depth layer is no longer very effective in controlling the temperature 10 cm further down. In other words, the debris layer is a far more effective insulator in summertime than in wintertime. And for this shallow layer it even seems to change from insulator in summer to conductor in winter. Such behaviour of the glacial debris cover is good news for the glacial ice below. Better conductive properties of the soil in winter allow the glacial ice to cool when the air temperatures are negative. While the great insulating properties of the summertime debris layer protect the underlying ice from seasonal warmth.

Like the previous plot, figure 9.13 shows how one soil layer controls the temperature of a deeper soil layer. This time the plot displays the correlation between the temperature at 16 cm and at 38 cm depth. The data points don't follow as narrow a line as the measurements from the shallow layer in

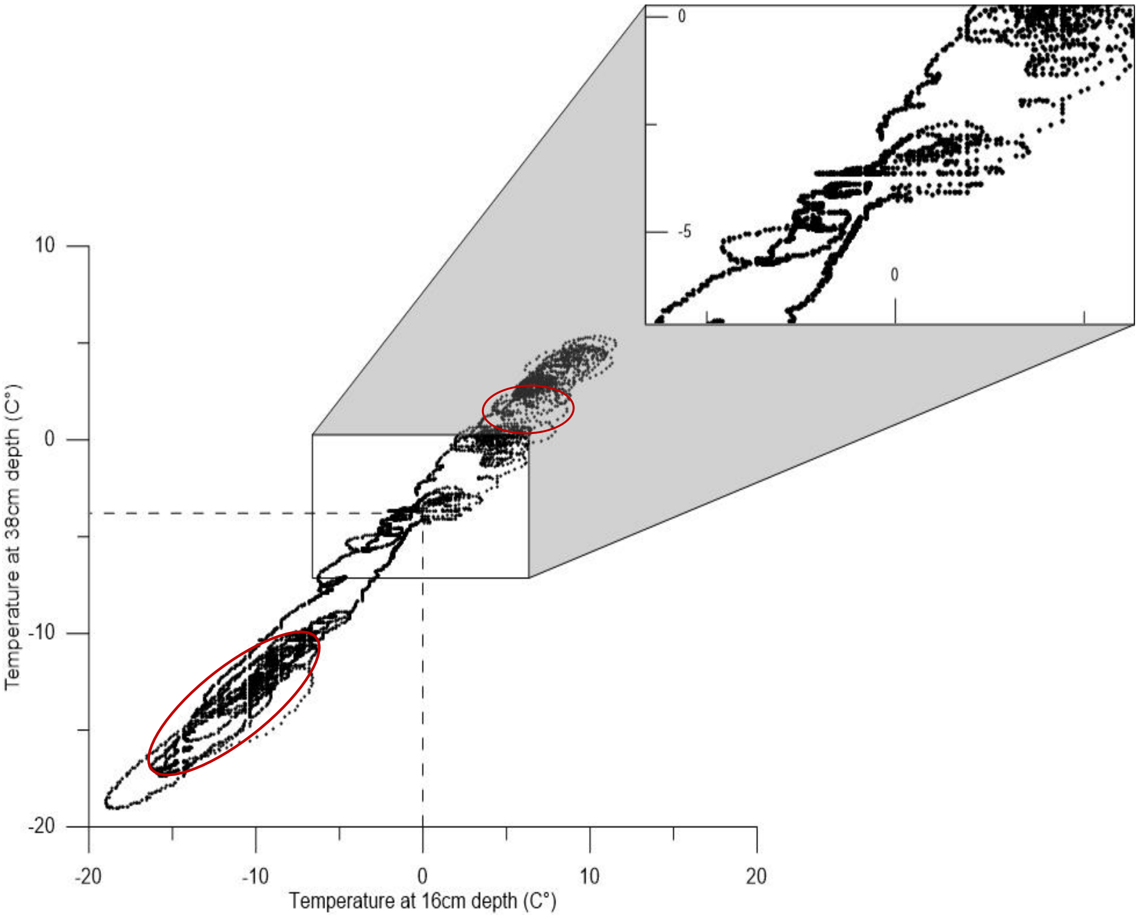

Figure 9.13: Although not as clear as closer to the surface the change from ice to water is a visible in the 16 cm to 38 cm layer as well (see zoomed in area). Below the melting point the delay in heat flow follows an ellipses at an angle near 45°. After entering the liquid water state the angle of these ellipses seems to flatten out.

figure 9.12, but instead follow a characteristic ellipse-like pattern (highlighted by the two red ellipses in figure 9.13). Since every data point is taken one hour after the previous one the ellipses formed can be tracked trough time. Doing so reveals that each ellipse has an anticlockwise movement. All that needs to be done is follow the track of the ellipse and one can read how the temperature changes with depth. This works as follows: Say the ground temperature at 16 cm depth increases. First the temperature at 16 cm depth increases more than at 38 cm depth (bottom of ellipse, moving anti-clockwise). Then the lower layer catches up and increases in temperature just like the less deep layer. Until, at the top of the ellipse, the deeper layer is still warming up while the layer closer to the surface no longer changes temperature. Inevitably a warm spell will be followed by a change to colder conditions. When this cold spell comes the temperature at 16 cm depth will first drop without much of an effect on the deeper layer (top of ellipse moving anti-clockwise). But soon the deeper layer starts to decrease in temperature as well, although not as much as the 16 cm depth level, until both layers show the same level of temperature decrease (angle is 45°). After this the roles reverse and the deeper layer is cooling down more than the upper layer, which is already closer to the current air-temperature. Towards the bottom of the ellipse the deeper layer is still cooling down while the upper layer is already stable. And having reached the bottom of the ellipse, we are back at square one, ready for another warm spell.

The angles of the ellipses in figure 9.13 hold the key. They tell how effectively a warm or cold spell is transmitted through the debris layer. When in frozen state the ellipses are lined up at an angle slightly below about 45°, indicating that the temperature at 16 cm efficiently, but with a slight delay, controls the temperature at 38 cm depth. If figure 9.13 would only show a line angled at 45° the temperature at 16 cm depth would be identical to the temperature at 38 cm depth. In reality however we find that it takes time for the ground to acclimatise, resulting in the ellipses found in the figure 9.13. In summer the angle of the ellipses in figure 9.13 tends to flatten out (it follows an angle of about 15°). The process told by the ellipses is the same for thawed ground as for frozen ground. But the lower angle means that the temperature transfer, in other words the adjusting of one layer to the temperature of the other layer, is slower in summer than in winter. In summer, it is common to find that the upper layer is warming or cooling, while the lower layer is not or hardly changing temperature. The delay may even cause the lower layer to still be cooling (as a response to a previous drop in air temperature), while the higher layer is already warming up, as a result of the current air-temperature. Such strong delays, causing prolonged opposite behaviours within 22 cm, can be seen at the lower half of the left-hand rounding and upper half of the right hand rounding of each summer ellipse. The lower angle of the summer ellipses is yet one more sign that the debris layer is a far more effective insulator in summer-time than in winter-time.

10. Interpretation and discussion specific to Longyearbreen

10.1. Glacial debris transport: What formed the debris cover?

Avalanches and rockfall events that originate at or near the glaciers headwall presumably are the dominant source of debris (see result section). But how did this debris become part of the layer covering Longyearbreen's terminus? The secret here is kept by individual clasts in the debris cover. Clasts on Longyearbreen are dominantly angular, which indicates that supra- or englacial transport was most important when the debris-covered glacier front was formed (Lukas et al., 2005). Some more rounded clasts can however be found. This has previously been interpreted as subglacial reworking, supported by the documentation of striated clasts at nearby Platåbreen (Lukas et al., 2005). However, during the fieldwork for this thesis no striated clasts could be found at Longyearbreen. Off course this does not prove that such clasts don't exists. Striated clasts have previously been recovered at Longyearbreen's front (O. Humlum 2014, personal communication, 6 November) and it is not the intention of this study to discredit those valuable finds. This study merely intends to propose an alternative explanation for the many non-angular clasts without any sign of glacial striations either on the clasts or on any clast or boulder nearby.

What kind of reworking is it that shapes the clasts on their way to the glacier's front? Sights photographed on Longyearbreen as part of the fieldwork in the autumn of 2014 may help answer this question (figure 10.1). In reality a glacial debris layer often is not just a layer of debris on the glacial ice. The border between debris layer and clean glacier ice is not that clear cut. One can for example find multiple layers of debris in glacial channels (figure 10.1, left). As their routes (slightly) change each year, multiple ice floors are created, each of which may or may not have its own debris cover. Additionally, clasts can be observed inside the otherwise debris-free glacier ice. Channels are also not reserved to just the glacial ice (figure 10.1, right). They have been discovered in the ice, the debris layer or cutting through both. A clast may be deposited onto the glacier by an avalanche or rockfall and find itself incorporated into the ice by subsequent snowfall. Ice deformation and slow movement may bring it towards the glaciers front. But on the way, it could easily get caught in a glacial stream, not to be transported away from the glacier, as most clasts are simply too big to be carried by water, but to be moved somewhat, slightly rounded and finally frozen back into the ice. Once near the glacier margin, melt will free it of its icy prison, making it appear on the surface. A process like this may lay at the start of many of the less angular clasts of Longyearbreen's debris cover.

Clasts are predominantly angular, with some slightly more rounded finds (Lukas et al., 2005). Previous work has shown that Longyearbreen is hardly moving at its base and in some places, has been frozen to its bed for over a thousand years (Etzelmüller et al., 2000, Humlum et al., 2005, Lukas et al., 2005, Sevestre et al., 2015). On the other hand, there is evidence of widespread channel network in and on the glacier. Debris layer thickness measurements performed for this study at Longyearbreen have also shown that the boundary between clean ice and debris can be diffuse: multiple layers, complete with layering, sorting and some rounding, and single clasts are often found frozen into clean ice. These findings can be explained by a combination of supra- and englacial transport followed by meltout; with occasional glacial river transport. If Longyearbreen was sufficiently thick in earlier times, a little subglacial erosion may also have taken place along warm-based parts of the glacier.

Figure 10.1: The left-hand photograph shows the western lateral channel cut inside the debris-covered front of Longyearbreen. One debris layer covers the current channel floor (bottom red arrow). The second debris layer depicts a former channel floor (top red arrow). A person leans against a large clast that sticks out of otherwise clean glacier ice (due to sandy dust this ice looks rather dark, but scraping of the ice revealed that it was clean). The right-hand picture shows a large supra-glacial channel that has cut through the debris layer (without reaching the underlying ice). Another stream has formed a tunnel in the debris layer and streams out into the main channel (highlighted by red circle). Collapses of such tunnels happen often and are usually preceded by a rain storm.

10.2. Deformation on and inside the debris-covered glacier margin

It became very apparent during this study that deformation processes like mudslides, rockslides, slumping and channel collapses do not gently change the glacier fronts appearance throughout the season. The debris-covered glacier front will in fact appear quite stable on most days. Deformation generally happens suddenly and on a massive scale during short events. This is the case during, for example, the first few days of spring-melt. Afterwards the system seems to enter a quasi-stable mode and summer, with its gentle light rain, usually does not bring many large-scale deformation events. During the freezeback period there are once again a few great deformation events. All large-scale deformation events that were recorded during the fieldwork on Longyearbreen happed during and shortly (a few hours) after sudden days of positive air temperatures and rain. These rainy warm days occurred amidst the otherwise cold and often freezing Autumn. With each event the soil not only received rainwater but also had to deal with meltwater of all ice and snow from the previous days.

Figure 10.2: The picture shows a channel at Longyearbreen on the 5th of August 2014. The blue long arrow depicts the position and direction of the meltwater stream. The side of the channel has recently collapsed (crack is highlighted by the red arrow), dumping a part of the former channel side into the riverbed. Note the colour difference between the top layer of the sediment and the lower part of the matrix. The top layer has dried a little which gives it a light colour. The rest of the debris layer is dark, since sediment is saturated with water. Oversaturation presumably led to the collapse.

Large amounts of liquid water clearly can destabilize the glacial debris cover. This is interpreted as the consequence of strongly differing degrees of permeability. Glacial ice does not even need to melt to destabilize the debris cover. The dense glacial ice is near-impermeable (Fountain and Walder, 1998), while the debris above is permeable. Most water that percolates down the debris must therefore leave the trough the debris layer itself, either via evaporation or fluid movement. For small quantities of water this is not a problem, but when large amounts of melt or rain water drain through the debris the bottom of the matrix may become over-saturated and detach from the ice below. Combine this with a slight slope angle and it is not hard to see why glacial debris may be released and slide downwards. This process can result in slumping and frequent collapses of channel sides (figure 10.2 and, in the result section, figure 9.7). This interpretation of water-saturation induced deformation can explain why deformation of a glacial debris cover is most persistent during spring melt and during or right after heavy rain showers in the freezeback period.

10.3. Prospects

Already at the beginning of this millennium Etzelmüller et al. (2000) noted that debris from the total backwall of Longyearbreen often does not enter the bergschrund but instead accumulates below the backwall. During the fieldwork performed in 2014 and 2015 for this thesis rockfall events and avalanches released from the glacier's backwall have been found to reach down onto the glacial ice and are therefore likely to be incorporated in the slow transport down-glacier. Debris accumulation into the bergschrund has not been noted but neither have efforts been made to observe it. Following Etzelmüller's findings however, englacial debris transport is unlikely to start right at place of deposition via incorporation into the bergschrund. What remains is the possibility of debris snowing in and becoming trapped in the glacial ice trough subsequent years of ice build-up. While this is a natural phenomenon in the accumulation zone of the glacier, and therefore likely to have played a substantial role in the glacial debris transport in the past, it is not as likely to happen today and even less so in the future. The accumulation zone on Longyearbreen is decreasing fast. As this section is being written, in September 2016, the glacier is in its entirety snow-free. No snow, not even right at the bergschrund, has survived the summer melt. All accumulation of last winter has already melted before the start of winter, in other words: currently the glaciers accumulation zone is non-existent. In 2014 and 2016 the picture was not as bleak as today, but the accumulation zone was, although existent, extremely narrow.

Of course, these are short term observations that may simply represent a spike in the general trend. But even so, the future of Longyearbreen, and with it the possibility to incorporate debris inside it's ice, does not look promising. As long as this situation holds, debris will not be incorporated into the ice. Since the glacier is already near stationary, this means that the current warming trend (Isaksen et al., 2007) and resultant retreat of the accumulation zone, may in time result in a debris-covered glacier area right below the backwall. Alternatively, the ice may melt to quickly for the debris to accumulate in time. This would be a death sentence to all but the current debris-covered margin of Longyearbreen, as without a protecting cover and with the current warming trend the glacier will, given time, simply melt away.

10.4. Temperature recording: Heat transfer in the debris layer

10.4.1. Seasonal Changes

Seasonal changes seem to greatly influence how heat is transferred throughout a glacial debris cover. This is true for both the ground temperature itself as well as its gradient. The alteration of the soil characteristics caused by the arrival of summer and the coming of winter likely has a major impact on the debris layer's insulating abilities. But what changes do the seasons actually bring? The accumulation and melt of a snow layer for one, certainly has impact on the soil below. Snow can form an effective insulting layer covering the supra-glacial debris. Next to this the presence of a snow layer changes the surface albedo. A snowpack functions as a reflector due to its high surface albedo (typically 0.8 to 0.9) , but the bare summer soil has a rather low surface albedo (typically around 0.2) (Humlum, 1998). In other words, the exposed summer surface is relatively prone to changes in air temperature and quick to absorb short-wave radiation. Additionally, phase changes of soil water may impact the ground temperature very directly by the movement of cold or warm water throughout the debris layer. Moreover, even when without physical water movement the heat capacity of the active layer changes with the shift from pore ice to liquid water. As ice is a better thermal conductor than water (Ratcliffe, 1962, Ramires et al., 1995, Akyurt et al., 2002), conductive heat transport through the soil is more effective in winter than in summer. What may also have effect is the relatively low amount of precipitation in the interior of Central Spitsbergen (Humlum, 2002). Summer dust storms are in fact a common sight in nearby Adventdalen during summer and autumn. Once the spring melt comes to an end and pore water starts to slowly evaporate, the soil is left with partly water, partly air-filled pores. Since air has an extremely low conductivity air filled pores increases the insulating properties of the soil.

10.4.2. The influence of a snow cover

The sudden change to strongly variable ground temperatures and temperature gradients that can be seen in the figure 9.10 and 9.11 of the result section (highlighted by a red dotted line) occurs simultaneously with the onset of snow melt. This leads one to think of the insulation and change in surface albedo caused by snow. At below freezing air temperatures , the ground surface is typically covered by a snow pack with high surface albedo (0.8 to 0.9) (Humlum, 1998). This snow pack insulates the ground below and protects the ground surface form direct radiation, reducing ground temperature variations (even as near to the surface as 6 cm depth (figure 9.10 & figure 9.11). But in summer positive air temperatures result in snow melt, leaving the surface with a low albedo (typically around 0.2) (Humlum, 1998). As a result, incoming short-wave radiation heats the unprotected summer surface significantly more than is indicated by the air temperature alone. Snow-free ground responses rapidly to changes in air temperature and is quick to absorb the abundant short-wave radiation often coinciding with high temperatures.

It is only logical that the shift from a relatively smooth to a highly variable ground temperature (gradient) signal happens suddenly. Seasonal snowmelt in Svalbard happens quickly. And although the snow pack will already start to thin in May (figure 9.10), causing the ground temperature to rise, it is the simple presence or absence of an exposed snow cover that determines the surface albedo. Therefor the change in surface albedo will usually happen rapidly. Combine this with the quick nature of the spring snow melt in Svalbard and one must conclude that anything influenced by a snow layer must show a sudden change as the snow disappears.

If the presence of snow influences the temperature (gradient) variations it must also impact the range of the ground temperatures within the soil. A delayed response of the soil (enhanced by the presence of an insulating snow layer) to changes in air temperature may result in a relatively small ground-temperature range. The ground temperatures, within the measured layer of 6 to 38 cm depth, cover a range from +18 °C to -21 °C. The air temperatures range from +17 °C to -29 °C. It is likely that the presence of a snow layer, by acting as a buffer, causes the minimum ground temperature to strongly differ from the minimum air temperature. The effect of this snow-buffer is impressive. Especially when realizing that the debris layer is a better conductor in winter than in summer, as had to be concluded from the data presented in figure 9.12 and 9.13 (result section).

It is however important to recall that the snow cover on Longyearbreen is highly uneven (ranging from 0 to 3.2 m.) and mainly controlled by the topography of the debris layer. Since the ground temperature measurements were done on a relatively even and flat area it is unlikely that the snow cover thickness differed spatially in this zone, but on other locations on the glacier front the impact of snow could be much more, less or even inexistent.

All in all, it is highly likely that the presence of a snow layer is responsible for the little short term variation of both ground temperature and ground temperature gradient that was found in the winter season of 2014-2015. And likewise, the absence of this snow layer in summer results in a greatly increased short term variability of the ground temperature (gradient). The sudden shift from a smooth to highly variable signal at the onset of summer is also in favour of the great influence of the snow layer.

10.4.3. Introducing a conceptual heat transfer model

Phase changes of soil water impact the ground temperature, as the heat capacity of the active layer changes with the shift from pore water to ice. It is probable that changes like these, influencing the conductivity of- and transport within the soil, cause the soil heat transport to differ so greatly from summer to winter. This difference in heat transport efficiency may be explained by the different conductive properties of ice and water and the degree to which both are mobile. In the form of a conceptual model, figure 10.3 aims to visualize how heat transport through a glacial debris layer changes with the change of seasons. Each season is depicted in a simplified manner, showing the processes that in all likelihood greatly influence heat transport during a certain season. This includes conductive as well as non-conductive processes like the effect of rain or snow events, melting, freezing, water percolation through the debris, (wind enhanced) evaporation and solar radiation in relation to surface albedo and the angle of the sun.

In wintertime, the pores of the debris layer will be filled with ice, allowing for efficient conductive heat transport (shown in figure 10.3 as a blue coloured debris layer). As thaw sets in, the lower conductivity of liquid water results in far less effective heat transport within the debris layer: the debris layer becomes a better insulator. As long as the pore water is relatively stationary the net effect of the phase change to water is protecting the ice core beneath by increasing the insulating properties of the debris layer.

At the onset of summer though, the phase change from ice/snow to water has a strong warming effect on the debris and underlying ice. As the figure 9.10 and 9.11 show, the shift from freezing to thawing ground temperatures, and thus the shift from snow-covered to snow-free surface conditions, happens quickly. Meltwater percolates into the debris layer via pores, cracks and other

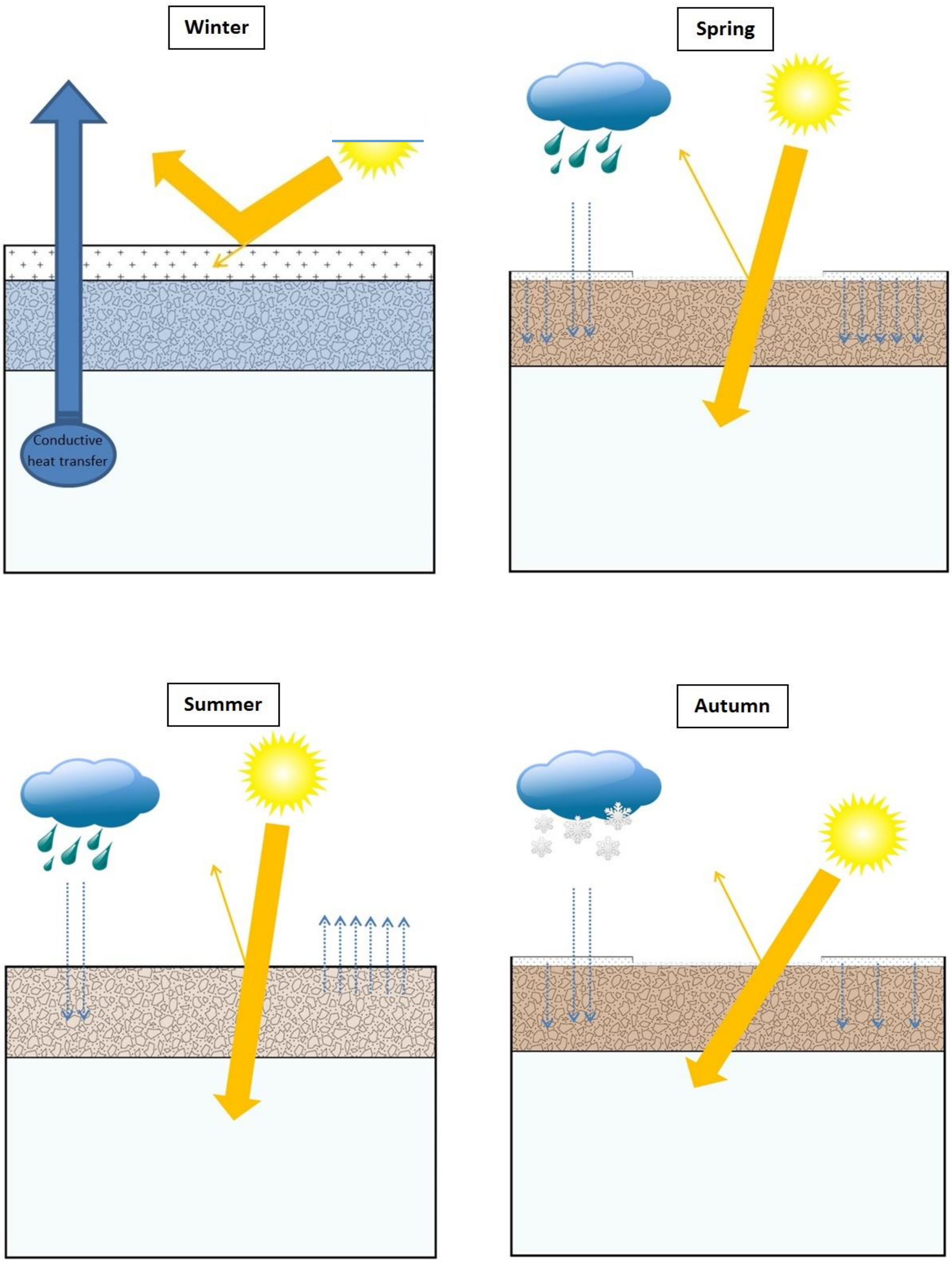

Figure 10.3: Conceptual model of seasonally impacted heat transfer within and underneath a glacial debris cover. During winter a glacial debris layer will be frozen (hence the blue coloured debris). This allows for effective conductive cooling which may to a certain degree be hindered by the presence of a sufficiently thick snow cover. During early- to mid-winter solar radiation is of little impact due to the high albedo of the surface and the low angle or even absence of the sun (polar night). When spring melt arrives, percolating water will quickly heat and thaw the debris layer. The disappearing snow cover leads to a lowering of the surface albedo. Solar radiation is free to heat the debris and possibly the underlying ice. After the melting has calmed down, liquid pore water will become less mobile: water only percolates downwards during and after showers. Meanwhile, evaporation consumes heat and thereby cools the debris layer. The result of both the pore water becoming less mobile and the drying by evaporation is a well-insulating summertime debris layer, right at the time when solar heating is most powerful. This is shown in the figure as a lighter brown (dried out) debris layer in summer. When Autumn comes evaporation from the surface is hindered by a lowering of the sun's angle, shorter days and the coming of an unstable snow cover. As early snowfall repeatedly settles and melts, percolating melt water cools the relatively warm debris. Until finally, the debris layer freezes up, a snow layer starts to settle and winter has come.

discontinuities (shown in figure 10.3 as dotted blue arrows that start at a patchy snowpack and point into the debris layer). This nonconductive heat transfer process results in a quick temperature rise within the debris layer (Humlum, 1998); so quick that temperature control of upper layers on lower layers seems to be almost immediate. Figure 9.12 and 9.13 illustrate this rapid phase change by the extremely narrow range of data points at the phase change from ice to water.

This highly effective warming by meltwater percolation only last a short while. For the data points become highly scattered and/or the angle of the ellipses falls dramatically when above-freezing ground-temperatures are reached (figure 9.12 and 9.13). In other words, heat transport through the debris layer in summer is slow. Physical heat transport by pore-water movement certainly does not hinder heat exchange from one soil layer to another. The meltwater must therefore drain or become stationary (partially by refreezing) relatively quickly. As soon as melting has settled down the now less-conductive properties of the non-frozen debris layer hinder further warming of the underlying ice.

The effect of a low conductivity becomes even greater as summer progresses. The low amounts of precipitation in Central Spitsbergen mean that the ground is prone to drying up. Pores become air filled and extremely low conductive. This dried ground, partially filled with water and partially filled with air-filled pores, forms a very effective insulating layer, protecting the ice below. In figure 10.3 this partially dried up debris layer is visualized by a slightly lighter shade of brown than in the previous season of spring melt.

The process of evaporation also helps to cool the debris layer. As pore water evaporates heat is consumed, leading to a cooling of the ground (visualized in figure 10.3 by dotted blue arrows pointing upwards from the debris layer). Evaporation may be assisted by wind movement. Naturally this makes topographic highs, like for example a steep debris-covered glacier front, especially susceptible to evaporation. Add to this that Longyearbreen is, like the rest of Svalbard, a place exposed to frequent and relatively strong winds and one must conclude that wind assisted evaporation is likely to be of importance.

In a debris layer consisting of courser material than found at the measurement site the influence of air movement can play an even bigger role. Air filed pores that are connected to one another allow winds to move through the soil. Naturally the temperature of the air impacts the temperature of the soil it travels through. Previous research has demonstrated the how exposure to high wind speeds can result in forced ventilation in the active layer of rock glaciers (Humlum, 1997). In summer this leads to efficient active layer warming. As autumn comes and air temperature drops the same process can effectively cool the debris layer. However, wind pumping and other forms of forced ventilation can only take effect if the debris surface layer is sufficiently course grained. But at the measurement site the debris layer consisted of clay, silt, fine sand, gravel, pebbles and rocks. This matrix supported by fine sediments does not give much room for air movement compared to, for example, the much coarser debris layers found on rock glaciers. Forced ventilation and the resultant cooling or warming is therefore not likely to play a substantial role at Longyearbreen.

At the onset of winter (melt) water percolation may once again strongly influence the debris layer's thermal regime. As early snow showers repeatedly settle and melt they wet the debris layer. This once again can be seen in figure 10.3 as dotted blue arrows starting at the patchy snow layer and ending in the debris below. At this time of year however the meltwater will be colder than the soil it travels through. A wet autumn can therefore speed up the cooling of glacial debris layers. The effect of a wet active layer during this time goes further. Even though it takes extra energy to freeze this pore-water, a process that initially releases heat to the debris, if the debris layer is soaked when it freezes at the start of winter the end result is a highly conductive ice-debris mixture. This allows for effective conductive cooling of the glacial ice core during winter. Especially if in the early winter months snowfall is limited or unable to settle due to high winds. Even when spring melt arrives the ice core still benefits from the previous wet start of the winter since great amounts of energy must be used the melt the ice within the debris layer before any warmth can reach the glacial ice underneath. The energy needed to melt the pore-ice must of course be equal to the energy (warmth) released to the debris layer when freezing took place. These thawing and freezing processes cancel out against each other, bringing the net result to zero. BUT, in the meantime, during the long arctic winter, the debris layer has been ice-filled and allowed for highly efficient conductive cooling. A dry summer, creating insulating are filled pores in the debris-layer, would be the end to a perfect ice-preservation scenario.

A dry onset of winter followed by a period of snowfall and little wind will, based on the same theory, hinder the cooling of any debris-covered ice core. When summer arrives, thawing a debris layer with minimal pore-ice content would take little energy. This is scenario both prevents effective cooling and speeds up the melting of debris-covered ice. Add to this a wet summer, preventing the debris-layer from drying out and reaching its maximum insulating properties and you have a true nightmare for the glacial ice core.

11. Conclusions

Analysis of 164 glacier termini in Nordenskiöld Land, Oscar II Land and Prins Karls Forland showed that debris-covered glacier termini are very common in Svalbard and occur both along the West Coast as well as in Central Spitsbergen. A large-scale study of 88 glacier termini, representative of the 164 previously mapped termini, show a clear negative correlation, and probable relation, between the fraction of the glacier that is debris-covered and a glacier's clean surface area. Small glaciers tend to produce large debris-covered glacier fronts, while large glaciers create relatively little ice-core in the terminal zone. Differences in geology, topography and climate do not seem to impact the correlation between the clean glacier surface and the debris-covered area.

In Central Spitsbergen, small glaciers (less than 5 km^2) are very common resulting in many glaciers with a debris-covered front. Small glaciers along the west coast of Svalbard form debris covers as well, but since glaciers here are generally larger, there are not as many with debris-covered fronts. Debris-covered glacier termini in Central Spitsbergen also tend to be slightly more convex than in the west. This may well be related to climatic forced differences in debris layer conductivity, since coastal Svalbard receives more precipitation than inner Spitsbergen.

Analysis of 88 glaciers and their terminal zones indicates that it is possible to predict the debris-covered fraction of a glacier with the use of a simple empirical model, with clean glacier surface area as its only input variable. This may well be an important step towards including debris-covered ice in future climate studies, without the need of expensive and cumbersome in-situ observations in remote areas. Although additional research is needed to firmly cement the accuracy of the model, the practical implications are very encouraging. For glaciers of up to 15 km^2 the relation between the debris-covered percentage of a glaciers surface (Y) and the clean ice surface in km^2 (x) is described as:

$$Y = -12.72 \ln x + 35.166$$

A process that is likely to disturb the relation between glacier size and its debris-covered fraction is surging, which may be responsible for the odd few small glaciers without a substantial debris-covered zone. Analysis of 14 land-terminating glaciers (in and nearby the study area) that are known or believed to have surged showed that surge-type glaciers are unlikely to form debris-covered termini.

Presumably a surge-event would destroy a debris-covered front, or, if surging is cyclic, even prevent it from forming at all.

Debris-covered ice acts differently in response to climatic changes than clean ice, meaning that it cannot be added to climate models without accounting for these differences. This study does not aim to singlehandedly solve these issues. However, big progress comes in small steps. Field data have been gathered from the glacier Longyearbreen to work towards reaching a better understanding of the processes that influence the formation and degradation of debris-covered glacial ice. It also enables a close look at the current state of a glacier with a debris-covered front. Longyearbreen's front is covered by on average 0.5 m of debris (n=36). The glacier, like many others in Central Spitsbergen, seems to rely on supra-glacial debris sources like avalanches and rock fall events. The debris is then transported supra- and englacially and occasionally by glacial streams. Near the glacier front material is released by meltout. But frequent collapses and slumping events keep the debris active. These forms of deformation are especially common during and straight after mid-freezeback rain showers.

Ground temperature measurements at different depths in Longyearbreen's debris layer show that, while the total insulating effect during winter is considerate due to the snow cover, the properties of the debris cover seem to change seasonally, thus enabling cooling of the ice during winter. The debris itself presumably is a better insulator during summer-time, when it is needed most, than during winter-time, when the low air temperatures cool the covered ice. This may explain the great impact that a debris cover can have on a glacier's response to climate change.

Although climatic differences within Spitsbergen do not seem to decide if a debris covered front forms or not, they do have an effect. Climatic changes towards overall dry conditions are likely to enhance debris-covered ice preservation. The perfect scenario to preserve debris-covered ice would combine wet autumns, cold, dry and/or windy winters and dry summers.

12. Recommendations for future research

It is important to recall that this study is a preliminary one. Further research is needed to validate its findings. It would be of great use to perform additional field work on a few of the glaciers that are included in this study in order to validate the ice-core maps, which have formed the base of all large-scale analysis.

Next to validating findings, it might be of interest to further investigate a few subjects introduced in this thesis. In other words, picking up where this study ends. A start has been made at interpreting the differences between debris-covered fronts. Additionally, the clear trend between a glacier's size and its debris-covered fraction has been investigated. Further research may continue on such subjects that could not been fully investigated. For example:

- The differences between debris-covered glacier fronts in different locations may well have to do with climate. It is known that the Centre of Spitsbergen tends to be arid, while the precipitation increases towards the (west) coast. But following the conceptual heat transfer model, introduced in chapter 10.4.3, the timing of precipitation is of major influence. Dry summers and winters generally enhance debris-covered ice preservation, but this does not seem to be the case for the autumn. It deserves recommendation to look into the seasonal climate properties of Central and West Spitsbergen in order to rightly interpret the differences between debris-covered glacier fronts on the archipelago.

- Small glaciers tend to produce large debris-covered glacier fronts, while large glaciers create relatively little ice-core in the terminal zone. An explanation for this may lie in the manner in which large and small glaciers respond to a rise of the ELA, leading large glaciers to presumably experience a smaller relative increase in ablation area because of a higher maximum altitude or a wider spread accumulation zone relative to the ablation zone. Both maximum height and width likely are greater for large glaciers, since they tend to cover their headwalls more extensively. Whether or not this idea holds must of course be tested by additional research before any conclusions can be made. A succeeding study might focus on the average maximum height of glaciers of different sizes as well as the relative width of the accumulation areas of large and small glaciers.

13. References

AKYURT, M., ZAKI, G. & HABEEBULLAH, B. 2002. Freezing phenomena in ice–water systems. *Energy Conversion and Management,* 43**,** 1773-1789.

ALLEY, R. B., BRIGHAM-GRETTE, J., MILLER, G. H. & POLYAK, L. 2009. Past climate variability and change in the Arctic and at high latitudes. *Synthesis and Assessment Product.* Washington, D.C.

BARRAND, N. E. & MURRAY, T. 2006. Multivariate Controls on the Incidence of Glacier Surging in the Karakoram Himalaya. *Arctic, Antarctic, and Alpine Research,* 38**,** 489-498.

BENN, D. I., KIRKBRIDE, M. P., OWEN, L. A. & BRAZIER, V. 2003. Glaciated valley landsystems. *Glacial landsystems***,** 372-406.

BERTHLING, I., ETZELMÜLLER, B., ISAKSEN, K. & SOLLID, J. L. 2000. Rock glaciers on Prins Karls Forland. II: GPR soundings and the development of internal structures. *Permafrost and Periglacial Processes,* 11**,** 357-369.

BRINGEDAL, A. 2004. *Oxygen isotopes from the surface of Longyearbreen, central Spitsbergen: palaeoclimatic implications.* Master Thesis, University of Bergen.

CHAPUIS, A. & TETZLAFF, T. 2014. The variability of tidewater-glacier calving: origin of event-size and interval distributions. *Journal of Glaciology,* 60**,** 622-634.

DOWDESWELL, J., HAMILTON, G. & HAGEN, J. O. 1991. The duration of the active phase on surge-type glaciers: contrasts between Svalbard and other regions. *Journal of Glaciology,* 37**,** 388-400.

DOWDESWELL, J., HODGKINS, R., NUTTALL, A., HAGEN, J. & HAMILTON, G. 1995. Mass balance change as a control on the frequency and occurrence of glacier surges in Svalbard, Norwegian High Arctic. *Geophysical Research Letters,* 22**,** 2909-2912.

ECKERSTORFER, M. & CHRISTIANSEN, H. H. 2011. Topographical and meteorological control on snow avalanching in the Longyearbyen area, central Svalbard 2006–2009. *Geomorphology,* 134**,** 186-196.

ECKERSTORFER, M., CHRISTIANSEN, H. H. & NEUMANN, U. Avalanches and snow mobile traffic around Longyearbyen, Svalbard. International Snow Science Workshop, at Davos, 2009.

ETZELMÜLLER, B., ØDEGÅRD, R. S., VATNE, G., MYSTERUD, R. S., TONNING, T. & SOLLID, J. L. 2000. Glacier characteristics and sediment transfer system of Longyearbreen and Larsbreen, western Spitsbergen. *Norsk Geografisk Tidsskrift - Norwegian Journal of Geography,* 54**,** 157-168.

EVATT, G. W., ABRAHAMS, I. D., HEIL, M., MAYER, C., KINGSLAKE, J., MITCHELL, S. L., FOWLER, A. C. & CLARK, C. D. 2015. Glacial melt under a porous debris layer. *Journal of Glaciology,* 61**,** 825-836.

FØRLAND, E., HANSSEN-BAUER, I. & NORDLI, P. 1997. Climate statistics and longterm series of temperature and precipitation at Svalbard and Jan Mayen. *DNMI report,* 21**,** 43.

FOUNTAIN, A. G. & WALDER, J. S. 1998. Water flow through temperate glaciers. *Reviews of Geophysics,* 36**,** 299-328.

HAGEN, J. O., LIESTØL, O., ROLAND, E. & JØRGENSEN, T. 1993. Glacier atlas of Svalbard and Jan Mayen. *Meddelelser,* 129**,** 167.

HAGEN, J. O., MELVOLD, K., PINGLOT, F. & DOWDESWELL, J. A. 2003. On the Net Mass Balance of the Glaciers and Ice Caps in Svalbard, Norwegian Arctic. *Arctic, Antarctic, and Alpine Research,* 35**,** 264-270.

HAMBREY, M. J., MURRAY, T., GLASSER, N. F., HUBBARD, A., HUBBARD, B., STUART, G., HANSEN, S. & KOHLER, J. 2005. Structure and changing dynamics of a polythermal valley glacier on a centennial timescale: Midre Lovénbreen, Svalbard. *Journal of Geophysical Research: Earth Surface,* 110.

HANSSEN-BAUER, I., SOLÅS, M. K. & STEFFENSEN, E. 1990. *Climate of Spitsbergen*, DNMI.
HODGKINS, R., HAGEN, J. O. & HAMRAN, S.-E. 1999. 20th century mass balance and thermal regime change at Scott Turnerbreen, Svalbard. *Annals of Glaciology,* 28**,** 216-220.
HOGUIN, R. & RESTIFO, F. 2012. Middle Holocene archaeology: Dynamics of environmental and socio-cultural change in South America. *Quaternary International,* 256**,** 1.
HUMLUM, O. 1997. Active layer thermal regime at three rock glaciers in Greenland. *Permafrost and Periglacial Processes,* 8**,** 383-408.
HUMLUM, O. 1998. Active Layer Thermal Regime 1991-1996 at Qeqertarsuaq, Disko Island, Central West Greenland. *Arctic and Alpine Research,* 30**,** 295-305.
HUMLUM, O. 2002. Modelling late 20th-century precipitation in Nordenskiold Land, Svalbard, by geomorphic means. *Norsk Geografisk Tidsskrift - Norwegian Journal of Geography,* 56**,** 96-103.
HUMLUM, O., CHRISTIANSEN, H. H. & JULIUSSEN, H. 2007. Avalanche-derived rock glaciers in Svalbard. *Permafrost and Periglacial Processes,* 18**,** 75-88.
HUMLUM, O., ELBERLING, B., HORMES, A., FJORDHEIM, K., HANSEN, O. H. & HEINEMEIER, J. 2005. Late-Holocene glacier growth in Svalbard, documented by subglacial relict vegetation and living soil microbes. *The Holocene,* 15**,** 396-407.
HUMLUM, O., INSTANES, A. & SOLLID, J. L. 2003. Permafrost in Svalbard: a review of research history, climatic background and engineering challenges. *Polar Research,* 22**,** 191-215.
HUMLUM, O. & ZIAJA, W. 2002. Discussion of "Glacial Recession in Sarkappland and Central Nordenskioldland, Spitsbergen, Svalbard, during the 20th Century" by Wieslaw Ziaja. *Arctic, Antarctic, and Alpine Research,* 34**,** 226.
INGÓLFSSON, Ó. n.d. *Outline of the physical geography and geology of svalbard* [Online]. Available: https://notendur.hi.is/oi/svalbard_geology.htm [Accessed October 2 2016].
ISAKSEN, K., SOLLID, J. L., HOLMLUND, P. & HARRIS, C. 2007. Recent warming of mountain permafrost in Svalbard and Scandinavia. *Journal of Geophysical Research: Earth Surface,* 112**,** F02S04.
JAKOBSSON, M., MACNAB, R., MAYER, L., ANDERSON, R., EDWARDS, M., HATZKY, J., SCHENKE, H. W. & JOHNSON, P. 2008. An improved bathymetric portrayal of the Arctic Ocean: Implications for ocean modeling and geological, geophysical and oceanographic analyses. *Geophysical Research Letters,* 35.
JISKOOT, H., MURRAY, T. & BOYLE, P. 2000. Controls on the distribution of surge-type glaciers in Svalbard. *Journal of Glaciology,* 46**,** 412-422.
KAUFMAN, D. S., SCHNEIDER, D. P., MCKAY, N. P., AMMANN, C. M., BRADLEY, R. S., BRIFFA, K. R., MILLER, G. H., OTTO-BLIESNER, B. L., OVERPECK, J. T. & VINTHER, B. M. 2009. Recent warming reverses long-term Arctic cooling. *Science,* 325**,** 1236-1239.
KOCH, J., CLAGUE, J. J. & OSBORN, G. 2014. Alpine glaciers and permanent ice and snow patches in western Canada approach their smallest sizes since the mid-Holocene, consistent with global trends. *The Holocene,* 24**,** 1639-1648.
LACELLE, D., LAPALME, C., DAVILA, A. F., POLLARD, W., MARINOVA, M., HELDMANN, J. & MCKAY, C. P. 2016. Solar Radiation and Air and Ground Temperature Relations in the Cold and Hyper-Arid Quartermain Mountains, McMurdo Dry Valleys of Antarctica. *Permafrost and Periglacial Processes,* 27**,** 163-176.
LARSEN, N. K., PIOTROWSKI, J. A., CHRISTOFFERSEN, P. & MENZIES, J. 2006. Formation and deformation of basal till during a glacier surge; Elisebreen, Svalbard. *Geomorphology,* 81**,** 217-234.
LØNNE, I. 2006. Low-velocity glacial surges—Processes unlocked by modern surge on Svalbard. *Geology,* 34**,** 553-556.
LØNNE, I. & LYSÅ, A. 2005. Deglaciation dynamics following the Little Ice Age on Svalbard: Implications for shaping of landscapes at high latitudes. *Geomorphology,* 72**,** 300-319.

LOVELL, H. 2014. *On the ice-sediment-landform associations of surging glaciers on Svalbard.* PHD, Queen Mary University of London.

LOVELL, H., FLEMING, E. J., BENN, D. I., HUBBARD, B., LUKAS, S. & NAEGELI, K. 2015. Former dynamic behaviour of a cold-based valley glacier on Svalbard revealed by basal ice and structural glaciology investigations. *Journal of Glaciology,* 61**,** 309-328.

LUKAS, S., NICHOLSON, L. I. & HUMLUM, O. 2007. Comment on Lønne and Lyså (2005):"Deglaciation dynamics following the Little Ice Age on Svalbard: Implications for shaping of landscapes at high latitudes", Geomorphology 72, 300–319. *Geomorphology,* 84**,** 145-149.

LUKAS, S., NICHOLSON, L. I., ROSS, F. H. & HUMLUM, O. 2005. Formation, Meltout Processes and Landscape Alteration of High-Arctic Ice-Cored Moraines—Examples From Nordenskiold Land, Central Spitsbergen. *Polar Geography,* 29**,** 157-187.

MATTSON, L. E., GARDNER, J. & YOUNG, G. 1993. Ablation on debris covered glaciers: An example from the Rakhiot Glacier, Panjab, Himalaya. *IAHS publication,* 218**,** 289-296.

MELVOLD, K. & HAGEN, J. O. 1998. Evolution of a Surge-Type Glacier in its Quiescent Phase: Kongsvegen, Spitsbergen, 1964-95. *Journal of Glaciology,* 44**,** 394-404.

MIDGLEY, N. G., COOK, S. J., GRAHAM, D. J. & TONKIN, T. N. 2013. Origin, evolution and dynamic context of a Neoglacial lateral–frontal moraine at Austre Lovénbreen, Svalbard. *Geomorphology,* 198**,** 96-106.

MURRAY, T., LUCKMAN, A., STROZZI, T. & NUTTALL, A.-M. 2003a. The initiation of glacier surging at Fridtjovbreen, Svalbard. *Annals of Glaciology,* 36**,** 110-116.

MURRAY, T., STROZZI, T., LUCKMAN, A., JISKOOT, H. & CHRISTAKOS, P. 2003b. Is there a single surge mechanism? Contrasts in dynamics between glacier surges in Svalbard and other regions. *Journal of Geophysical Research: Solid Earth,* 108.

NIKITA-MARTZOPOULOU, C. 1981. Effect of solar radiation on soil temperature in pots housed in cold frames. *Agricultural Meteorology,* 24**,** 263-274.

NORDLI, Ø., PRZYBYLAK, R., OGILVIE, A. E. J. & ISAKSEN, K. 2014. Long-term temperature trends and variability on Spitsbergen: the extended Svalbard Airport temperature series, 1898-2012. *Polar Research,* 33.

POLASHENSKI, C., COURVILLE, Z., BENSON, C., WAGNER, A., CHEN, J., WONG, G., HAWLEY, R. & HALL, D. 2014. Observations of pronounced Greenland ice sheet firn warming and implications for runoff production. *Geophysical Research Letters,* 41**,** 4238-4246.

QUINCEY, D. J., BRAUN, M., GLASSER, N. F., BISHOP, M. P., HEWITT, K. & LUCKMAN, A. 2011. Karakoram glacier surge dynamics. *Geophysical Research Letters,* 38.

RAMIRES, M. L. V., NIETO DE CASTRO, C. A., NAGASAKA, Y., NAGASHIMA, A., ASSAEL, M. J. & WAKEHAM, W. A. 1995. Standard Reference Data for the Thermal Conductivity of Water. *Journal of Physical and Chemical Reference Data,* 24**,** 1377-1381.

RATCLIFFE, E. H. 1962. The thermal conductivity of ice new data on the temperature coefficient. *Philosophical Magazine,* 7**,** 1197-1203.

REUSCHE, M., WINSOR, K., CARLSON, A. E., MARCOTT, S. A., ROOD, D. H., NOVAK, A., ROOF, S., RETELLE, M., WERNER, A., CAFFEE, M. & CLARK, P. U. 2014. 10Be surface exposure ages on the late-Pleistocene and Holocene history of Linnébreen on Svalbard. *Quaternary Science Reviews,* 89**,** 5-12.

REZNICHENKO, N., DAVIES, T., SHULMEISTER, J. & MCSAVENEY, M. 2010. Effects of debris on ice-surface melting rates: an experimental study. *Journal of Glaciology,* 56**,** 384-394.

SALVIGSEN, O. 2002. Radiocarbon-dated Mytilus edulis and Modiolus modiolus from northern Svalbard: Climatic implications. *Norsk Geografisk Tidsskrift - Norwegian Journal of Geography,* 56**,** 56-61.

SALVIGSEN, O., FORMAN, S. L. & MILLER, G. H. 1992. Thermophilous molluscs on Svalbard during the Holocene and their paleoclimatic implications. *Polar Research,* 11**,** 1-10.

SCHERLER, D., BOOKHAGEN, B. & STRECKER, M. R. 2011. Spatially variable response of Himalayan glaciers to climate change affected by debris cover. *Nature Geoscience,* 4**,** 156-159.

SEVESTRE, H., BENN, D. I., HULTON, N. R. J. & BÆLUM, K. 2015. Thermal structure of Svalbard glaciers and implications for thermal switch models of glacier surging. *Journal of Geophysical Research: Earth Surface,* 120**,** 2220-2236.

SIEWERT, M. B., KRAUTBLATTER, M., CHRISTIANSEN, H. H. & ECKERSTORFER, M. 2012. Arctic rockwall retreat rates estimated using laboratory-calibrated ERT measurements of talus cones in Longyeardalen, Svalbard. *Earth Surface Processes and Landforms,* 37**,** 1542-1555.

SLETTEN, K., LYSÅ, A. & LØNNE, I. D. A. 2001. Formation and disintegration of a high-arctic ice-cored moraine complex, Scott Turnerbreen, Svalbard. *Boreas,* 30**,** 272-284.

SOBOTA, I., WECKWERTH, P. & NOWAK, M. 2016. Surge dynamics of Aavatsmarkbreen, Svalbard, inferred from the geomorphological record. *Boreas,* 45**,** 360-376.

SOLOMINA, O. N., BRADLEY, R. S., HODGSON, D. A., IVY-OCHS, S., JOMELLI, V., MACKINTOSH, A. N., NESJE, A., OWEN, L. A., WANNER, H., WILES, G. C. & YOUNG, N. E. 2015. Holocene glacier fluctuations. *Quaternary Science Reviews,* 111**,** 9-34.

STUART, G., MURRAY, T., GAMBLE, N., HAYES, K. & HODSON, A. 2003. Characterization of englacial channels by ground-penetrating radar: An example from austre Brøggerbreen, Svalbard. *Journal of Geophysical Research: Solid Earth,* 108.

SUND, M. & EIKEN, T. 2004. Quiescent-phase dynamics and surge history of a polythermal glacier: Hessbreen, Svalbard. *Journal of Glaciology,* 50**,** 547-555.

SUND, M., EIKEN, T., HAGEN, J. O. & KÄÄB, A. 2009. Svalbard surge dynamics derived from geometric changes. *Annals of Glaciology,* 50**,** 50-60.

SVENDSEN, J. I. & MANGERUD, J. 1997. Holocene glacial and climatic variations on Spitsbergen, Svalbard. *The Holocene,* 7**,** 45-57.

WALKER, M., BERKELHAMMER, M., BJÖRCK, S., CWYNAR, L., FISHER, D., LONG, A., LOWE, J., NEWNHAM, R., RASMUSSEN, S. O. & WEISS, H. 2012. Formal subdivision of the Holocene Series/Epoch: a Discussion Paper by a Working Group of INTIMATE (Integration of ice-core, marine and terrestrial records) and the Subcommission on Quaternary Stratigraphy (International Commission on Stratigraphy). *Journal of Quaternary Science,* 27**,** 649-659.

WALSH, K. M., HOWAT, I. M., AHN, Y. & ENDERLIN, E. M. 2012. Changes in the marine-terminating glaciers of central east Greenland, 2000–2010. *The Cryosphere,* 6**,** 211-220.

YAO, Y., ZHAO, Q., BERA, S., LI, X. & LI, C. 2012. Pollen morphology of selected tundra plants from the high Arctic of Ny-Ålesund, Svalbard. *Advances in Polar Science,* 23**,** 103-115.

ZIAJA, W. 2001. Glacial Recession in Sorkappland and Central Nordenskioldland, Spitsbergen, Svalbard, during the 20th Century. *Arctic, Antarctic, and Alpine Research,* 33**,** 36-41.