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Combining terrestrial and marine glacial archives: A geomorphological map of the Nordenskiöldbreen forefield, Svalbard

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Geology
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Abstract

Recent research suggests that Svalbard glaciers have retreated and thinned since the Little Ice Age. The analysis of glacial landforms and sediment deposits in the terrestrial and marine environments permits the reconstruction of the temporal and spatial variability of recent deglaciations. This study is founded on the analysis and interpretation of a geomorphological map (1:10000) of the foreland of Nordenskiöldbreen (16°43’35” - 17°11’32” E and 78°37’40” - 78°41’15” N). The map covers an area of ca. 43 km^2 and is constructed from a combination of field observations and remotely sensed imagery from both the terrestrial and marine environments. Landforms are classified into seven genetic categories: (1) subglacial, (2) ice-marginal, (3) supraglacial, (4) proglacial, (5) glaciofluvial, (6) periglacial and (7) coastal. Glacier front positions from 1896 to present are reconstructed using historical data, oblique and vertical aerial imagery and LANDSAT and ASTER imagery.

The results of this investigation suggest Nordenskiöldbreen has been a dynamic glacier with components of all glaciers found in Svalbard: it is a polythermal, tidewater, calving, outlet glacier with both terrestrial and marine termination, with surge imprints in the southern part. The retreat pattern reconstructed in this thesis indicates that water depth and bedrock pinning points (sills) played an important role in the antecedent stability of Nordenskiöldbreen. Additionally, existing landsystem models fail to capture the complexity of glaciers with a combined terrestrial and marine terminus – such as Nordenskiöldbreen. The application of these models should therefore be carefully considered when investigating glaciers with a mixed terminus. This study contributes to an improved understanding of Svalbard glaciers and their response to recent climate fluctuations.

Sammendrag

Isbreene på Svalbard har det siste århundre, siden kulminasjonen av den Lille Istiden, blitt tynnere og trukket seg tilbake. Dette studiet presenterer et detaljert marint-terrestrisk geomorfologisk kart over det nylig eksponerte området foran Nordenskiöldbreen, Svalbard. Eksisterende landsystemmodeller gjelder for isbreer med *enten* terrestrisk eller marin brefront og det komplekse landskapet foran Nordenskiöldbreen kan ikke bli forklart via en slik landsystemmodell. Dette studiet understreker forskjeller og likheter mellom de terrestriske og marine arkiver gjennom et geomorfologisk kart, detaljerte landformbeskrivelser og tolkninger. Studiet presenterer også et omfattende sammendrag av tidligere publisert litteratur og en rekonstruksjon av brefrontens posisjoner fra 1896 til i dag. Det geomorfologiske kartet er basert på analyser av høyoppløselige flyfoto (2009) og akustiske havbunnsdata (multibeam) (2009, 2015) og er utarbeidet i ArcMap 10.3. Feltverifisering ble utført i august 2014 og 2015, hvor de kartlagte landformene ble inspisert, fotodokumentert og sedimentologiske undersøkelser ble utført på viktige lokaliteter. Det geomorfologiske kartet dekker et område på ca. 43km^2 og er i skala 1:10 000. Landformene er inndelt i følgende kategorier: *subglasiale-, randglasiale-, supraglasiale-, proglasiale-, glasifluviale, periglasiale- og kystlandformer*. Brefrontposisjonene har blitt rekonstruert fra historiske data, skrå og vertikale flybilder og LANDSAT/ASTER satellittbilder. Tilbaketrekningsmønsteret indikerer at vanndybde og berggrunnsrelieff spiller en viktig rolle for stabiliteten til Nordenskiöldbreen. Nordenskiöldbreen har vist seg å være en ytterst dynamisk isbre med komponenter fra alle typer isbreer som finnes på Svalbard. Selv om breen er i tilbaketrekningsfasen er transverse morener blitt avsatt på fjordbunnen. Bevaringspotensialet til de ulike landformene varierer fra det terrestriske til det marine arkivet. Mange isbreer på Svalbard har tilsvarende til Nordenskiöldbreen en kombinert terrestrisk og marin rand og dette understreker at landsystemmodeller bør appliseres med forsiktighet når det gjelder isbreer på Svalbard. Studiet bidrar til en bedre forståelse av Svalbards breer og deres respons på klimaendringer, og er en del av forskningsprosjektet ”Holosen historie for Svalbards iskapper og isbreeer” (RIS #7567).

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Longyearbyen, 12.05.2016

Lis Allaart

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1. Introduction

1.1 Motivation

Glaciers are very sensitive to climate change and the landforms found in a glacier forefield can be used to reconstruct past fluctuations in the glacial margins (Benn & Evans 2010). Jakobsson *et al.* (2014, pp. 6) stated that "*ultra-high resolution records are needed to enable us to reconstruct past environmental change on time-scales comparable to current and near-future environmental change*". High-resolution data for the last century from the forefield of the polythermal, tidewater glacier Nordenskiöldbreen are available. The high-resolution aerial imagery and swath bathymetry is background for a geomorphological map of the glacier forefield. The rate of change within a recently deglaciated forefield is expected to be high (Ballantyne 2002), and a geomorphological map is a high-resolution "snap shot" of the landscape at the time when the map is created. In future perspectives such a map can work as baseline for estimating the rate of landscape change in the forefield (Schomacker & Kjær 2008). Detailed geomorphological mapping further enhances the understanding of individual landforms and the link between them, and sediments deposited during advance and retreat (Schomacker *et al.* 2014).

The glaciers in Svalbard have been retreating and thinning the last century (Plassen *et al.* 2004; van der Meer 2004; Kohler *et al.* 2007; Nuth *et al.* 2007; Rachlewicz *et al.* 2007; Błaszczyk *et al.* 2009; König *et al.* 2014). The mass-balance has generally been negative since the end of the Little Ice Age (LIA) and some glaciers have retreated with a rate exceeding 200 m pr. year (Hagen *et al.* 2003). A better understanding of how glaciers responded to LIA cooling and modern the subsequent warming, can improve our abilities to predict the response of Svalbard glaciers in the future.

1.2 Aims, objective and approach

The aim of this project is to produce a combined geomorphological map of the Nordenskiöldbreen forefield based on high-resolution aerial images and swath bathymetry, for highlighting glacial imprints left both in the marine and terrestrial realms. Preservation potential of the different landforms vary between these environments.

Another important focus of this study has been to plan and carry out field campaigns whereby ground verification of remote sensing interpretations of the terrestrial part of the mapped area could be carried out. Both the northern and southern parts of the glacier forefield were investigated and sedimentological surveys conducted at key sites. The easier accessibility to the northern shore led to that more field days were spent there.

The difference between handling the terrestrial and marine data concerns ground verification. The sedimentology of the seafloor has not been sampled with coring or otherwise, and the interpretations are therefore based on high-resolution acoustic sea-floor (multibeam) and subbottom (chirp) data, supported by previously published marine geological work from the area and an unpublished acoustic profile (Plassen *et al.* 2004; Baeten *et al.* 2010).

A literature review of the previous research of Nordenskiöldbreen has been carried out in order to reconstruct past glacier extent, highlight the rapid revolution in the landscape and support the landform interpretation.

The differences and similarities between the marine and terrestrial realm of the glacial archives have been assessed. Therefore, I have chosen to describe the different landforms not by location but by type. An attempt to divide the landforms into different unifying categories has been made; subglacial landforms, ice-marginal landforms, supraglacial landforms, proglacial landforms, glaciofluvial landforms, periglacial landforms and coastal landforms. Each type of landform will be described and subsequently interpreted.

A final aim in this study has been to assess the applicability of *the landsystem model approach* (Evans & Rea 1999; Glasser & Hambrey 2003) in a setting like Nordenskiöldbreen.

1.3 The landsystem model approach

Conceptual models of the geomorphology of glacier forefields, so-called landsystem models, have been developed for land- and marine-terminating glaciers of both surge- and non-surge type (Evans & Rea 1999; Glasser & Hambrey 2001; Plassen *et al.* 2004; Fig. 1). Landsystem models can be used to identify glacier type. The landsystem model incites a holistic approach where the whole suite of landforms in the glacier forefield is assessed and thereafter the glacier type is identified. No single landform is diagnostic for a glacier type, but some landforms are considered *smoking guns* - to be unique for certain glacier types (Cleland 2013). Searching for *a smoking gun* is a way of testing hypotheses and the *smoking gun* will favour one hypothesis over others. The imprints of recent glacial activity constitute a complicated picture, as some diagnostic forms are intrazonal; e.g. old moraines that have been overridden by more recent surges, features related to proglacial outwash fans, topography constrained features such as collapsed lake plains (Evans *et al.* 2009). Many landforms occur within all the different landsystem assemblages.

A recent study on the landforms from documented surge-type outlet glaciers in Iceland (Brynjólfsson *et al.* 2014) showed little agreement with the surging glacier landsystem model of Evans and Rea (1999). However, the absence of surge-landforms does not exclude past surges of the glacier (Ingólfsson *et al.* 2016). Many landforms also show equifinality and morphology alone cannot explain the origin of a landform (Goudie 2004).

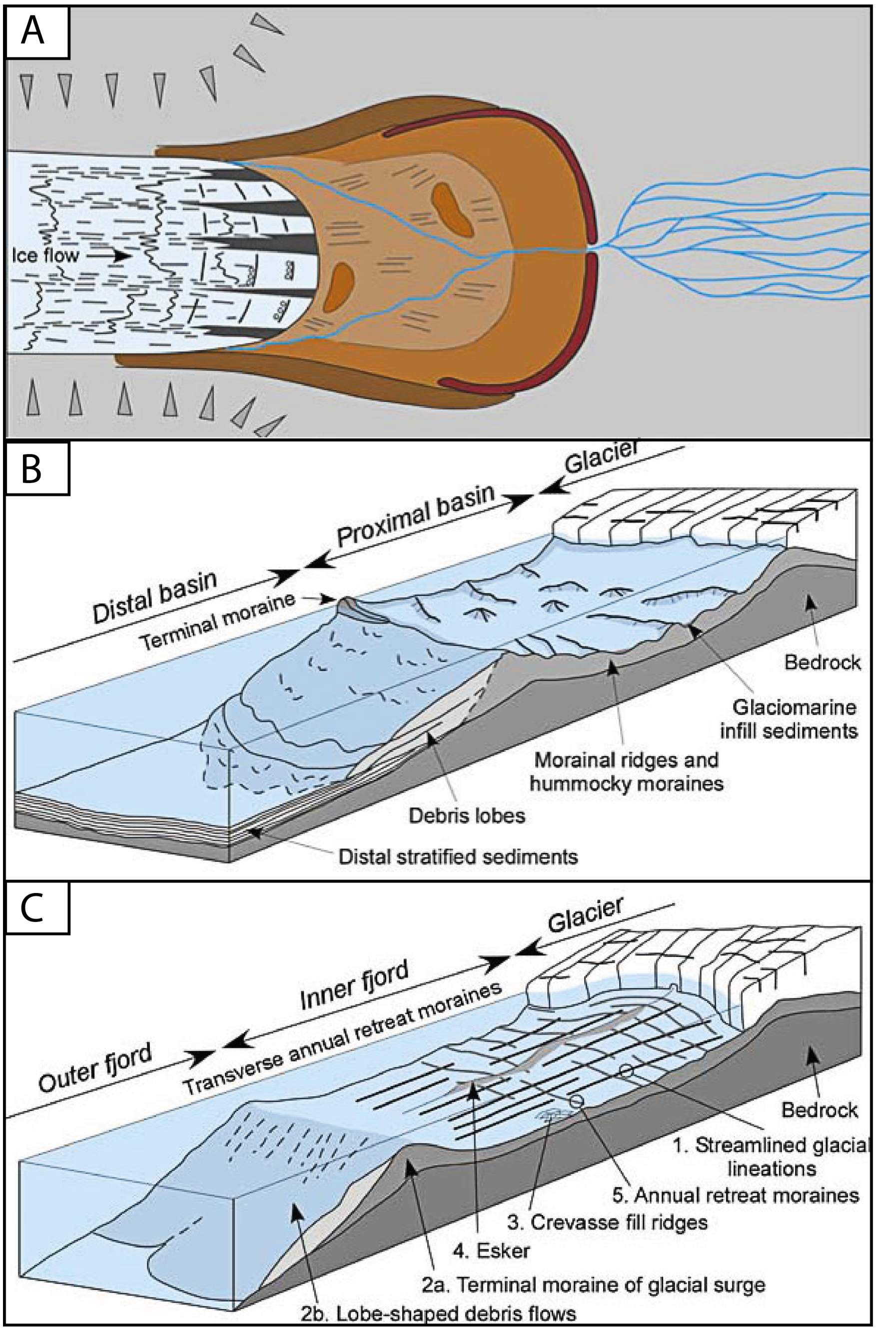


***Figure 1**. Landsystem models. A) Landsystem model from a polythermal, land-terminating glacier. B) Landsystem model for a polythermal tidewater glacier. C) Landsystem model for a surge-type tidewater glacier. Modified from: Glasser and Hambrey (2001); Plassen et al. (2004); Ottesen et al. (2008); Ingólfsson (2011).*

The landsystem models discriminate between glaciers with terrestrial and marine termini (Evans & Rea 1999; Glasser & Hambrey 2001; Plassen *et al.* 2004). Recent retreat has caused that many glaciers in Svalbard have a mixed marine and terrestrial termination (van Pelt *et al.* 2013; Flink *et al.* 2015), and therefore the existing landsystem models do not necessarily apply. The exposed parts of the glacier forefields have increased in size, as the glaciers have retreated and a holistic approach where information from both the terrestrial and marine realm is analysed is necessary in order to improve our understanding of the variety of glaciers in Svalbard.

2. Regional setting

2.1 Billefjorden

Billefjorden is located between 78°27' and 78°45'N and 16°00' and 17°00'E and is a tributary fjord to the Isfjorden fjord system (Fig. 2). Isfjorden cuts into the central part of Spitsbergen and opens up towards the Greenland Sea in the west (Forwick & Vorren 2010). The fjord basin was deglaciated between 12.3 – 11.0 cal. ka BP (Svendsen & Mangerud 1997). The ice front retreated from the outer to the inner part of Billefjorden between 11.3 and 11.2 ka. BP (Baeten *et al.* 2010). The last phase of the deglaciation around 11.2 ka BP was dominated by intense iceberg rafting (Forwick & Vorren 2010). In older literature, the inner part of Billefjorden is called *Klaas Billen Bay* (Wordie 1921; Tyrrell 1922a; Slater 1925).

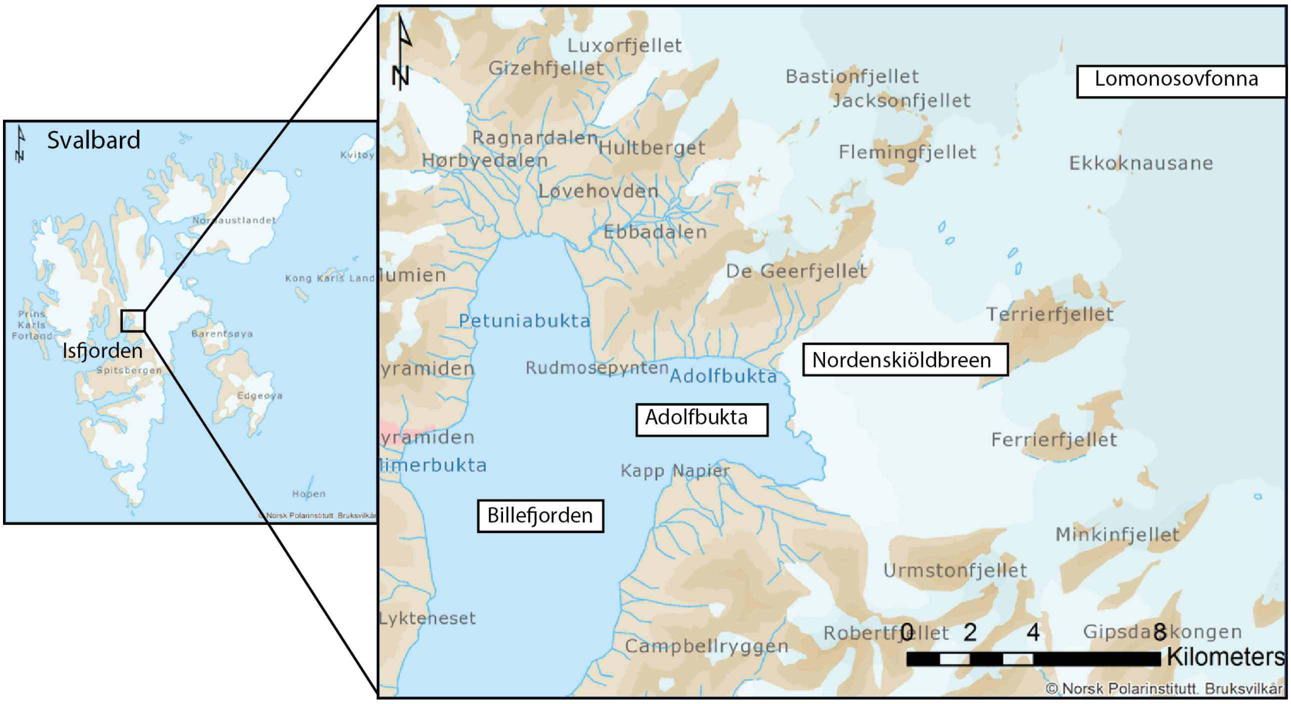


Figure 2. *Location map. A) Svalbard and Isfjorden, B) zoom in of the Nordenskiöldbreen area. Source: Norwegian Polar Institute, 2014.*

Billefjorden is approx. 30 km long, 5 to 8 km wide and up to 226 m deep (Baeten *et al.* 2010). The water exchange with Isfjorden is limited by a shallow sill at the mouth of the fjord (Baeten *et al.* 2010; Strzelecki 2011). The study area is identified in Fig. 2. Billefjorden receives sediment from Nordenskiöldbreen via systems of braided rivers, calving from the ice front and via englacial meltwater channels (Szczuciński *et al.* 2009). Most parts of the fjord floor has a post-glacial sediment cover less than 10 m thick (Elverhøi *et al.* 1995) however the cover increases to 60 m in front of the slope break in front of Nordenskiöldbreen (Plassen *et al.* 2004). The fjord is influenced by semi-diurnal tides with a maximum amplitude of 1.5 m (Szczuciński & Zajączkowski 2012).

2.2 Nordenskiöldbreen

Nordenskiöldbreen is the only tidewater glacier in Billefjorden (Rachlewicz *et al.* 2007), located at the head of the fjord, at 78°40'19"N, 16°54'39"E, terminating in Adolfbukta bay (Fig. 2). Hagen *et al.* (1993) classified Nordenskiöldbreen as a polythermal, tidewater glacier and Błaszczyk *et al.* (2009) described the glacier as fast flowing, with no sign of surge activity. Nordenskiöldbreen is one of the three main drainages of the large Lomonosovfonna Ice Cap, which has a central plateau located ~14 km NE of the present tidewater front of Nordenskiöldbreen.

After recent retreat, the ice front now partly rests on land and the significance of mass loss in form of frontal calving has declined (Plassen *et al.* 2004; van Pelt *et al.* 2013). The freshwater flux into Adolfbukta during the melt-season has been estimated to 49 $m^3 s^{-1}$ (Szczuciński & Zajączkowski 2012).

2.3 Climate

The climate in Svalbard is mainly affected by variations in extension of the air masses and sea ice. Mild air masses from the south collide with cold air masses from the north, and the temperature differences result in large weather variations particularly in the winter season (Hanssen-Bauer *et al.* 1990; Humlum *et al.* 2007). Along the west coast of Svalbard, the West Spitsbergen Current, a branch of the Gulf Stream, brings up warm waters, and the climate is modified thereby (Nilsen *et al.* 2008; Rachlewicz & Szczuciński 2008).

Meteorological measurements have been carried out in Longyearbyen since 1912 (Førland *et al.* 2010). There, the average annual precipitation is 180 mm and the mean annual air temperature is about -5°C (Humlum *et al.* 2007). The climate in inner Billefjorden is a little different from the climate in Longyearbyen (Kostrzewski *et al.* 1989; Rachlewicz & Szczuciński 2008). Meterological data have been collected in Petuniabukta, the neighbouring bay to Adolfbukta, during summer campaigns of various length in the 1980's and in the early 2000's (Kostrzewski *et al.* 1989; Rachlewicz & Szczuciński 2008). The highest summer temperatures were recorded in August (monthly average between 6.5 °C and 8.5 °C) and the highest amounts of summer precipitation were recorded in July (Rachlewicz & Szczuciński 2008). On the background of the available data the climate in inner Billefjorden was stated to be of quasi-continental, inner-fjord subtype (Rachlewicz & Szczuciński 2008). The variability in length of the data sets though questions the reliability of this classification.

2.4 Bedrock

The bedrock in the Nordenskiöldbreen area is shown in Fig. 3. Along the outer moraine on the northern shore of Adolfbukta, an area with uneven topography of dark weathered basement of Palaeo- and Mesoproterozoic origin occurs (Dallmann *et al.* 2004). The basement consists of garnet-mica schist, quartzite, amphibolite, few distinct layers of marble and is traditionally termed Hecla Hoek (Maher & Braathen 2011). The basement occurring within the LIA moraine is ice moulded with several sets of glacial striae. Distal to the moraine the Hecla Hoek bedrock is deeply weathered. On the southern shore, the basement sequence is only found inboard of the moraine, proximal in the forefield. The bedrock makes up the newly exposed Retrettøya, located halfway between the northern and the southern shore.

An angular unconformity separates the Hecla Hoek basement from overlying sedimentary rocks of Carboniferous age (Dallmann *et al.* 2004). The Carboniferous deposits vary from shale and sandstone to limestone, dolomite and gypsum/anhydrite and were mainly deposited in half-grabens in the Billefjorden Trough (Maher & Braathen 2011). Billefjorden trough has a clastic, carbonate and evaporate infill with complex disputed stratigraphic relationships (Maher & Braathen 2011). The Carboniferous outcrops occur on both sides of the bay.

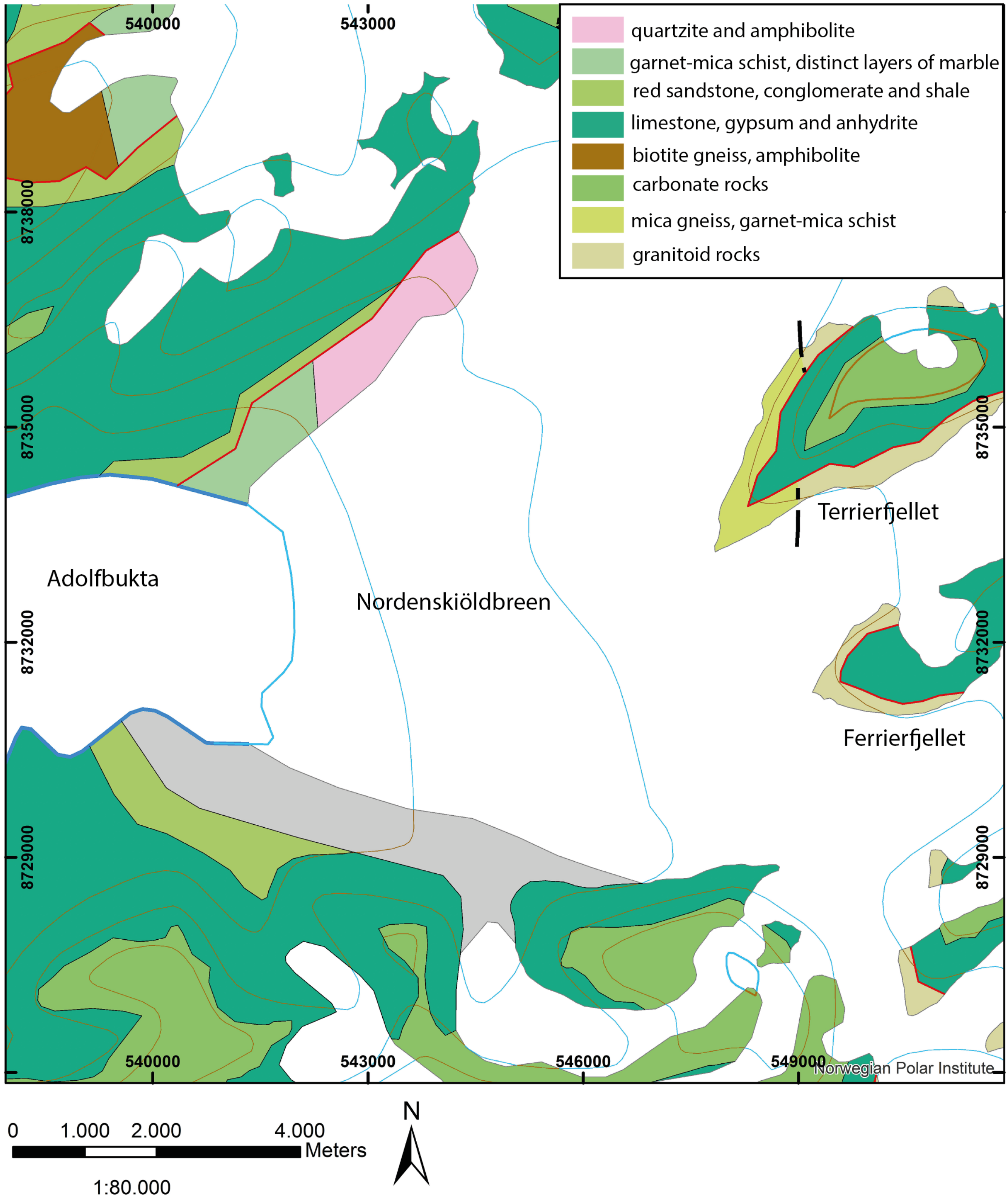


Figure 3. *Bedrock map of the Nordenskiöldbreen area. The lithologies vary from basement rocks of Paleo- and Mesoproterozoic origin exposed at the ice front to sedimentary rocks of Carboniferous age. The latter is seperated from the basement by an angular unconformity. The nunataks Terrier- and Ferrierfjellet located ~6 km from the ice front expose granitoid rocks at their base. Source: Norwegian Polar Institute, 2016.*

3. Late Quaternary history of the Nordenskiöldbreen area

3.1 Holocene Glacial History

Isfjorden was deglaciated between 12.3 – 11.0 cal. ka BP (Svendsen & Mangerud 1997). The ice front retreated to the inner part of Billefjorden by 11.2 ka BP. The last phase of the deglaciation of Billefjorden was dominated by intense iceberg rafting (Forwick & Vorren 2009). In the early Holocene, the glacier margins retreated to their modern limits or even further inland in some cases (Ingólfsson 2011). During the Holocene, the Svalbard archipelago is suggested not to have deglaciated completely (Ingólfsson 2011; Hormes *et al.* 2013).

The Holocene Thermal Maximum (HTM) is recorded in Svalbard by a change from glacier-proximal to marine deposits found in marine sediment cores and exposed sections in coastal cliffs, abundance of thermophilous molluscs, a low number of ice rafted debris (IRD), raised beaches and absence of permafrost at sea level (Salvigsen *et al.* 1992; Humlum 2005; Jessen *et al.* 2010; Ingólfsson 2011). The timing of the HTM varied throughout Svalbard; however, sediment core data suggests that the HTM in the Isfjorden area occurred around 10.8 ka BP (Svendsen and Mangerud 1997).

A neoglacial period is identified on Svalbard. During this period, ca. 5.5 ka BP, the activity of calving glaciers in Billefjorden and Tempelfjorden increased (Baeten *et al.* 2010). Linnébreen glacier, situated near the mouth of Isfjorden, started forming around 5.0 – 4.0 ka BP (Svendsen & Mangerud 1997). Concurrent with this, thermophilous molluscs declined in numbers in the waters around Svalbard (Salvigsen *et al.* 1992). Permafrost in the lowlands of Svalbard valleys started forming at around 3.0 ka BP (Humlum 2005).

In Svalbard, the term the Little Ice Age (LIA) refers to a cold period ending ~1920 (Salvigsen & Høgvard 2006). Areas between the maximum extent of glaciers during Neoglacial and their modern extent are marked by series of ice-recessional moraines for many Svalbard glaciers (Svendsen & Mangerud 1997; Sletten *et al.* 2001). However, it should be noted that the chronological control on the majority of these moraines is poor (Werner 1993). The postulated LIA moraines are found 1–2 km in front of the present day glacier front (Werner 1993). They are assumed to be ice cored and stand out as significant features in the landscape (Sletten *et al.* 2001).

Imprints from Anthropogenic activity are seen in the landscape and lake record of Svalbard, supporting the onset of a new geological epoch: The Anthropocene (Holmgren *et al.* 2010; D'Andrea *et al.* 2012; Wolfe *et al.* 2013). As of yet, the Anthropocene has not been formally recognized as a geological epoch (Waters *et al.* 2016).

3.2 Research history of Nordenskiöldbreen

Nordenskiöldbreen has been subject of glacial/paleoglacial research since the late 1800s. The approach and perspective of the research has changed with time and the development of new methods. New methods have given us aerial imagery, satellite imagery and bathymetrical data, but old photographs and sketches of the glacier front also reveals information on the state of the glacier. The area of Adolfbukta and Nordenskiöldbreen has been mapped repeatedly since the days of the early research (Conway 1898; de Geer 1910; Mathieson *et al.* 1919; Zinger 1965; Karczewski *et al.* 1990). Each map documents the state of the glacier around the time of production. The following literature review has been carried out in order to highlight the landscape changes over the past 120 years.

3.2.1 Early work (1896 – 1976)

In 1897, a pioneer sledge-expedition was carried out by E.J. Garwood and W. Martin Conway to investigate if inland ice of *"Greenland character"* was present in Svalbard (Conway 1898, pp. 138; Frazer 1922). They observed steep moraines in front of Nordenskiöldbreen and the glacier to have a heavily crevassed front. A *"curious long barrel-vault"* of ice and a burst glacier lake were observed close to Terrierfjellet (Conway 1898, pp. 140). Garwood (1899, pp. 684) described the medial moraines of Nordenskiöldbreen as "*interesting"* features being defined by rows of large almost perfectly rounded blocks of hard Archean rocks. The medial moraines were traced up to the nunataks found several kilometres upglacier and large erratics were observed on the glacier surface. Garwood argued that the roundness of the boulders indicated in- or subglacial transport, and he suggested that the rocks originated from an earlier glaciation, prior to the LIA advance. A map showing their ski-route on Nordenskiöldbreen, nunataks, the burst glacier lake and medial moraines together with the report was published (Conway 1898). The conclusion of their work was that true "*inland ice"* could be argued only to exist in North-Eastern Spitsbergen and on Nordaustlandet (Conway 1898, pp. 142).

The Swedish baron Gerard de Geer visited Nordenskiöldbreen on several expeditions between 1882 and 1910, and the results of his work were published in an excursion guide (de Geer 1910). The excursion guide included a detailed map of Nordenskiöldbreen with a description of the work behind the map and detailed observations of the area. The map and hypsometry of the glacier was constructed from detailed photogrammetry; the entire glacier was mapped in scale 1:20 000 in 1896 and 1908 by De Geer and his assistant, N.C. Ringertz. The southernmost part of the glacier *"seemed to have been stationary between 1882 and 1896"* (de Geer 1910, pp. 15), but an advance of a small side glacier had pushed the adjacent medial moraines aside. Soundings showed a depth of ~150 m in the bay.

The excursion guide was made for an eight-day-long excursion to Spitsbergen in connection with the International Geological Congress in 1910. Leading European geologists participated, and Nordenskiöldbreen was one of the excursion sites. The British geologist, G. A. J. Cole published a paper with observations from the excursion (Cole 1911). He described the visit to Nordenskiöldbreen as one of the most successful days of the excursion. Cole observed that the land terminating part of Nordenskiöldbreen was dark with a thin cover of mud and stones. This led him to assume that melting had been prohibited by the surface sediment cover. Cole also observed several large erratics on the glacier surface which is in accordance with Garwood (1899). A photo of the ice-front documented that gravitational fall-sorting of newly exposed intraglacial material took place there.

The Scottish Spitsbergen Syndicate had two expeditions to Spitsbergen in 1919 and 1920 of respectively seven weeks and four months (Brown 1920; Tyrrell 1921; Wordie 1921). The expeditions were led respectively by Dr. W.S. Bruce, after whom Brucebyen is named, and Dr. Rudmose Brown (Brown 1920). Motivated by the signing of the Svalbard Treaty in 1920 the objective of the expeditions was "*to prospect and develop"* the estates of Scotland/the UK in the Svalbard archipelago (Brown 1920, pp. 133; Wordie 1921). Another aim was to update the coastline map, and thereby several glaciers were visited. Due to findings of coal, the area around Nordenskiöldbreen was of interest (Brown 1920). British and Scottish scientists participated in the expeditions and a number of publications including geological and glaciological observations from the Adolfbukta was the outcome (Mathieson *et al.* 1919; Brown 1920; Tyrrell 1921; Wordie 1921; Tyrrell 1922a, b).

Wordie (1921, pp. 37) described the glaciers of Svalbard as retreating and as *"shrunken remains"* of a much thicker ice-cover. Nordenskiöldbreen was described as a large, valley-glacier of *alpine type,* draining highland ice. Monitoring of the glacier was carried out in 1919 by John Mathieson, which resulted in publication of a detailed map of *Klass Billen Bay* (modern day Billefjorden), including Nordenskiöldbreen (Mathieson *et al.* 1919).

G. W. Tyrell published a number of papers from the expeditions (Tyrrell 1921, 1922a, b). His work comprised thorough descriptions of glaciers all over Spitsbergen and a segment with highly detailed observations and studies of the moraines of Nordenskiöldbreen. Tyrrell analysed the lithological composition of the moraines. The terminal moraines were observed to be dominated by blocks/rocks of metamorphic basement, which he interpreted as that the floor underneath the glacier was made up of Pre-Devonian basement. Observations on the medial moraines revealed a different predominant lithology - Carboniferous sedimentary rocks which likely originated from the nunataks of Ferrier- and Terrierfjellet. Tyrrell also observed that the debris cover on the medial moraines limited melt of the underlying ice. A difference in surface elevation of 3 - 6 metres between the medial moraines and the exposed glacier ice was observed - in accordance with Cole (1911).

The peculiar appearance of the moraines in the south-western (de Geer 1910) part was also noted by Tyrrell. He argued that the southern medial moraines originally joined the lateral moraine near the termination of the glacier, but that a recent, energetic advance of the tributary glacier (Gerritbreen) had caused the first five medial moraines to curve (Tyrrell 1922a). A number of glacial phenomena were photo documented (Tyrrell 1922a; b; Fig. 4). 'Boulder clay' (equivalent to *till*) with clasts of various sizes and *shell-fragments* incorporated were also described. On a sketch the terminal moraine was marked with two crests (Tyrrell 1922a, pp. 36).

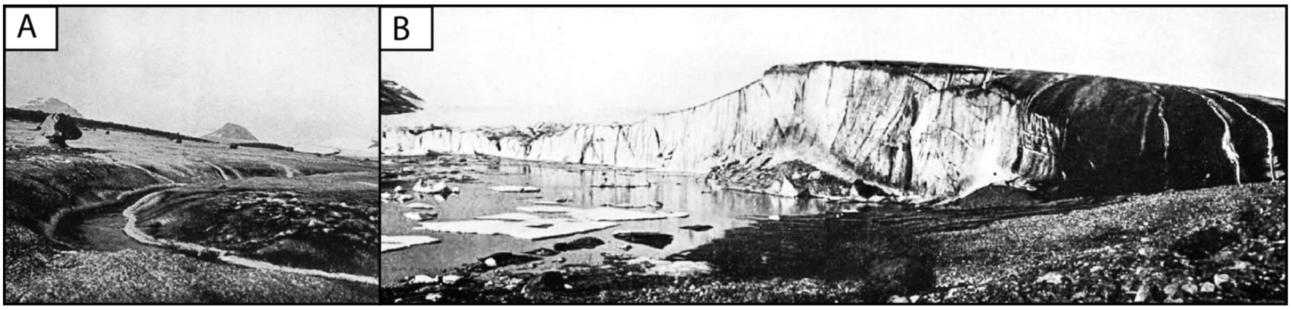


Figure 4. *Historical black and white photos of Nordenskiöldbreen. A) Differential melting caused by large boulders resting on the glacier surface adjacent to a ~ 5 m wide supraglacial channel, B) The south-western part of the debris covered icefront. Modified from Tyrrell (1922a).*

The Oxford University Expedition to Spitsbergen in 1921 focused on topographical work and a number of publications with observations of the Nordenskiöldbreen area were published (Frazer 1922; Odell 1922; Slater 1925; Odell 1927). The 1921 expedition continued the topographical work initiated by de Geer (1910) and Mathieson *et al.* (1919). They parted in a sledge party that went into the interior (Frazer 1922) and a base party that stayed in Brucebyen (Slater 1925).

The sledge party focused on topographical work and was led by N.E. Odell and R.A. Frazer and R.F Stobart were surveyors (Frazer 1922). Frazer (1922) described the lithological differences between the medial and terminal moraine in accordance with the previous expedition parties (Garwood 1899; Tyrrell 1922a). Frazer argued that Garwood's interpretation of the rounded blocks as remnants from a previous glaciation was less likely, and that the rounded boulders rather originated from a small glacier descending from the summit of Ferrierfjellet. The glacial lake observed by Garwood (1899) was also observed by Frazer (1922), but the lake had shrunk significantly.

Slater, the leading glaciologist of the base party of the expedition, described Nordenskiöldbreen as "*exhibiting a striking spectacle"* and with a heavily crevassed surface close to the front (Slater

1925, pp. 411). He linked the appearance of transverse crevasses on the surface to a ridge of metamorphosed rocks underneath the ice. Detailed descriptions of two types of moraines; surface- and marginal, occurring in the forefield were carried out (Slater 1925). The term *surface moraine* is equivalent to what in present day terminology is called *medial moraine.* The ice front was described to have a horse-shoe shape, and it was inferred to be a result of more rapid mass-loss by calving in the marine terminating part of the glacier. The marginal moraines were described as extensive features with the most prominent one found in the south-western moraine, being ~ 80 feet (~ 25 m) high.

A detailed sketch of the south-western, terrestrial part of the forefield showed that the glacier just recently had started to retreat and the ice front now rested a few metres behind the terminal moraine. The ice retreat was monitored by stake measurements at four stations (I – IV) during periods of 23 – 27 days (Table 1). Slater also pointed out a difference in lithological composition of the medial- and terminal moraine.

Table 1. *Results from stake measurements of the retreat of the ice cliff from Slater (1925).*

Station	Measuring period	Retreat
I	24 days	3 feet 11.5 inches
II	23 days	3 feet 6 inches
III	27 days	3 feet 1.5 inches
IV	24 days	1 foot 1.5 inches

Stott (1936) investigated the *Marine foods of birds in an inland fjord region in west Spitsbergen*, focusing on upwelling of nutritious waters in front of Nordenskiöldbreen in Adolfbukta Bay. The glacier was described to have a ~ 3.2 km wide ice front. Brown detritus-laden meltwater streaming from two tunnels in the ice front into the fjord was observed, and a correlation between high air temperatures, meltwater production and how far the meltwater plume was able to extend from the ice front was made.

A Soviet led expedition in 1965 did glaciological and meteorological measurements on both Lomonosovfonna and Nordenskiöldbreen (Zinger 1965). A detailed map with glacier hypsometry and ten ice front positions (from 1898 to 1965) was included in the publication. General observations of the moraine were also completed. The retreat of Nordenskiöldbreen was noted to be slower than Paulabreen (in Van Mijenfjorden), which also had been investigated by Zinger.

Boulton (1970b, 1976) used the Nordenskiöldbreen forefield as a case study on respectively subglacial melt-out tills, and the formation of flutes. The aim was to find support for existing theories of the genesis of both melt-out tills and flutes. Occurrences of stagnant debris-rich ice where tills had completely or partially melted out were observed (Boulton 1970b, 1976). Boulton described recently exposed striated metamorphic bedrock in the immediate vicinity of the glacier front. Based on the pattern of striations Boulton assumed that the undulating bedrock topography controlled the glacier movement. This was in support of Slater's interpretation of the connection between crevasse formation and obstructing bedrock underneath the ice (Slater 1925). The bedrock was described to be partially submerged by tills and fluvial deposits (Boulton 1970b, 1976). The latter was suggested to originate from ice-marginal streams migrading east-wards according to glacier retreat. A series of sections parallel to glacier flow direction, starting ~ 150 m from the glacier margin were studied. In the western-most sections, bedrock, overlain by unfrozen till, was observed. The same bedrock sequence was also identified more glacier proximal, there with glacier ice resting on top. Relatively well-preserved shell-fragments were found in the lowermost 60 cm of the till and fragmented shells were found in the uppermost part of the till (Boulton 1970b). The glacier ice was described as *resting* on the bedrock summits while subglacial material was accumulating in the hollows.

In Boulton's 1976 paper *on flutes,* a natural cavity at the northern frontal margin of Nordenskiöldbreen was described. There, three flutes all initiated by a low hummock on the bedrock surface were observed. The features were ~ 0.3 m high and had furrows on each side. The roof of the cavity had deep grooves starting from the point where the glacier sole and the hummocky bedrock was still connected and the location of the grooves matched the flute location (Boulton 1976). Further distal in the forefield, the study of the internal composition and structure of other flutes was enabled by erosion caused by a proglacial stream. Curving/bending flutes were also observed and Boulton argued that the ice flow there had been disturbed by an obstacle and subsequently flowed around it. Conclusively Boulton (1976) argued that the term *flute* should be used as a genetic term only: to describe *long parallel-sided ridges* indicating the former direction of ice movement.

3.2.2 Recent studies (1989 to present)

Until 1989, the surveys of Nordenskiöldbreen had focused on the terrestrial archive with an emphasis on the glaciology and morphology of the glacier, except a few soundings from the bay published by de Geer (1910). From 1989 a significant increase in resolution of aerial imagery and the proliferation of sea floor geographers increased the level of detail in the studies of the area.

Klysz *et al.* (1989) described the landforms present in the Nordenskiöldbreen forefield, identifying roches moutonnées formed in the Hecla Hoek (bedrock), lateral ice-cored moraine (h ≤ 20 m), fluvial outwash plains, traces of earlier outwash plains and a fluted moraine plain. Grain size analysis on samples of subglacial till collected close to the glacier margin was conducted. A narrow strip of marine terraces along the Adolfbukta coast was also described. There, accumulations of sub-fossil mollusc shells in sediments underlying till deposits were discovered and the material was suggested to of marine origin. The moraine of Nordenskiöldbreen was sketched with two ridges and till deposited in between.

Next year, the same working group published a geomorphological map in scale 1:40 000 of Petuniabukta and the northern part of Adolfbukta (Karczewski *et al.* 1990). The map was based on aerial images taken in 1961 by the Norwegian Polar Institute. The northern shore of Nordenskiöldbreen was included in the map and ice-cored lateral/frontal moraines, fluted moraine, outwash plains, appearance of single erratics and roche moutonée were mapped there.

Plassen *et al.* (2004) opened up the window to "new" data in the marine archive. A depositional model for surging and non-surging tidewater glaciers in Svalbard was presented. The model was based on acoustic soundings (Fig. 5) and sediment core data from investigations of glacigenic deposits in Spitsbergen fjords - and one of the case studies was on Adolfbukta bay. From acoustic profiles, ice-marginal deposits extending about 3 km beyond the present glacier front were observed. Their observations of ice-marginal deposits correlated well with the 1896 and 1908 AD glacier frontal position on the De Geer's map (de Geer 1910; Plassen *et al.* 2004). The 1896 position was assumed to be the Neoglacial maximum position. This location also corresponded with the bedrock topography and a seismic line revealed a clear slope break at this point (Plassen *et al.* 2004, Fig. 5). Distal thereof, a package of acoustically stratified sediments was underlying a debris lobe deposit with several slip surfaces in the upper part. Bathymetrical data of low resolution was presented and the area between the Neoglacial maximum and the present ice front was described as hummocky terrain (Plassen *et al.* 2004). The reflection patterns from the proximal part of the basin were chaotic with some internal incoherent reflections. The pattern was argued frequently to be interpreted as till (Plassen *et al.* 2004).

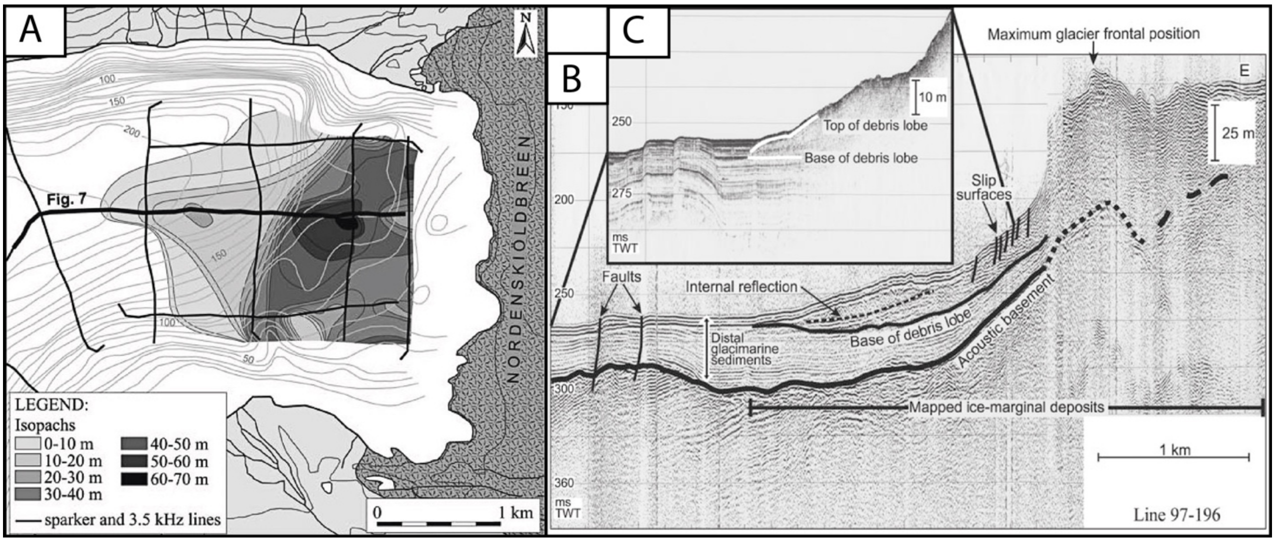


Figure 5*. Acoustic data from Adolfbukta 2004. A) Top view of Adolfbukta with isopacs and location of acoustic lines are marked with black lines. B) Sparker profile of the ice-marginal deposits. C) 3.5 kHz profile of the debris lobe and the distal glacimarine sediments. Modified from Plassen et al. (2004).*

Rachlewicz *et al.* (2007) surveyed post LIA retreat rates of the glaciers around Billefjorden. Remote sensing and archive data were the basis for the survey and it was supported by field measurements carried out from 2000 to 2005. All glaciers in the surveyed area showed retreat with Nordenskiöldbreen showing the most rapid retreat. The mean linear retreat rate of Nordenskiöldbreen was calculated to 35 m a^{-1}. The rates were assumed to have increased due to climate warming. Secondary factors suggested were: water depth at the grounding line, surge history, altitude, shape and aspect of glacier margin, and bedrock relief. Glacier retreat was also suggested to be influenced by the degree of erosion, and to affect the preservation potential of the glacially shaped landscape.

More work on Nordenskiöldbreen was published in 2010. From March 2006 to May 2009 den Ouden *et al.* (2010) used Nordenskiöldbreen as a case study on precise measurements of ice-flow velocities with stand-alone single-frequency GPS. Nine GPS receivers were installed on the glacier. The study revealed annual average ice velocities of 40 to 55 m yr^{-1} at the central flow line. Peak velocities were observed in the beginning of July, during the period with highest melt-rates. Maximum velocities varied between 60 and 90 m yr^{-1}. Fluctuations with a period of 15 to 20 days were observed in winter and tidal effects were suggested as explanation. Nordenskiöldbreen was concluded not to be a *very* fast flowing glacier.

Baeten *et al.* (2010) published a full bathymetrical dataset of Billefjorden, where bathymetrical data from 2005 and 2006 and results from analyses of one sediment core (retrieved in 1997), were presented. From the sediment core, it was inferred that Nordenskiöldbreen was comparatively small during the early Holocene and that its activity increased around 5470 cal. BP (Baeten *et al.* 2010). The glacier reached its LIA maximum around AD 1900, where it generated glacial lineations and a terminal moraine in the inner fjord. Transverse ridges inboard the terminal moraine observed in the bathymetry were linked to post-LIA annual retreat.. Iceberg plough marks were observed outboard of the LIA moraine, and due to their fresh appearance they were interpreted to have formed in the Late Holocene. Baeten *et al.* (2010) argued that Nordenskiöldbreen had been a tidewater glacier throughout the entire Holocene.

Nordenskiöldbreen was used as base for two modelling studies by van Pelt *et al.* (2012) and van Pelt *et al.* (2013). The first focused on simulating melt, runoff and refreezing. They were able to simulate mass balance, radiative fluxes and subsurface profiles that were in good agreement with observations. A problem in this study was that the main part of the meteorological data used in the model was from Svalbard Airport, which is approx. 60 km from the glacier front, and the climate in inner Billefjorden is described to differ from the climate in Longyearbyen (Rachlewicz & Szczuciński 2008).

van Pelt *et al.* (2013) modelled the basal topography of Nordenskiöldbreen. Included in their study was ground penetrating radar data achieved in 2010, which revealed bed heights below sea level upglacier of the present glacier front (Fig. 6). This might support the conclusion by Baeten *et al.* (2010) that Nordenskiöldbreen had existed as a tidewater glacier throughout the entire Holocene, but suggestively further behind the present glacier front. The GPR data were used to validate the modelled bed, but the coverage of the GPR lines was sparse (Fig. 6).

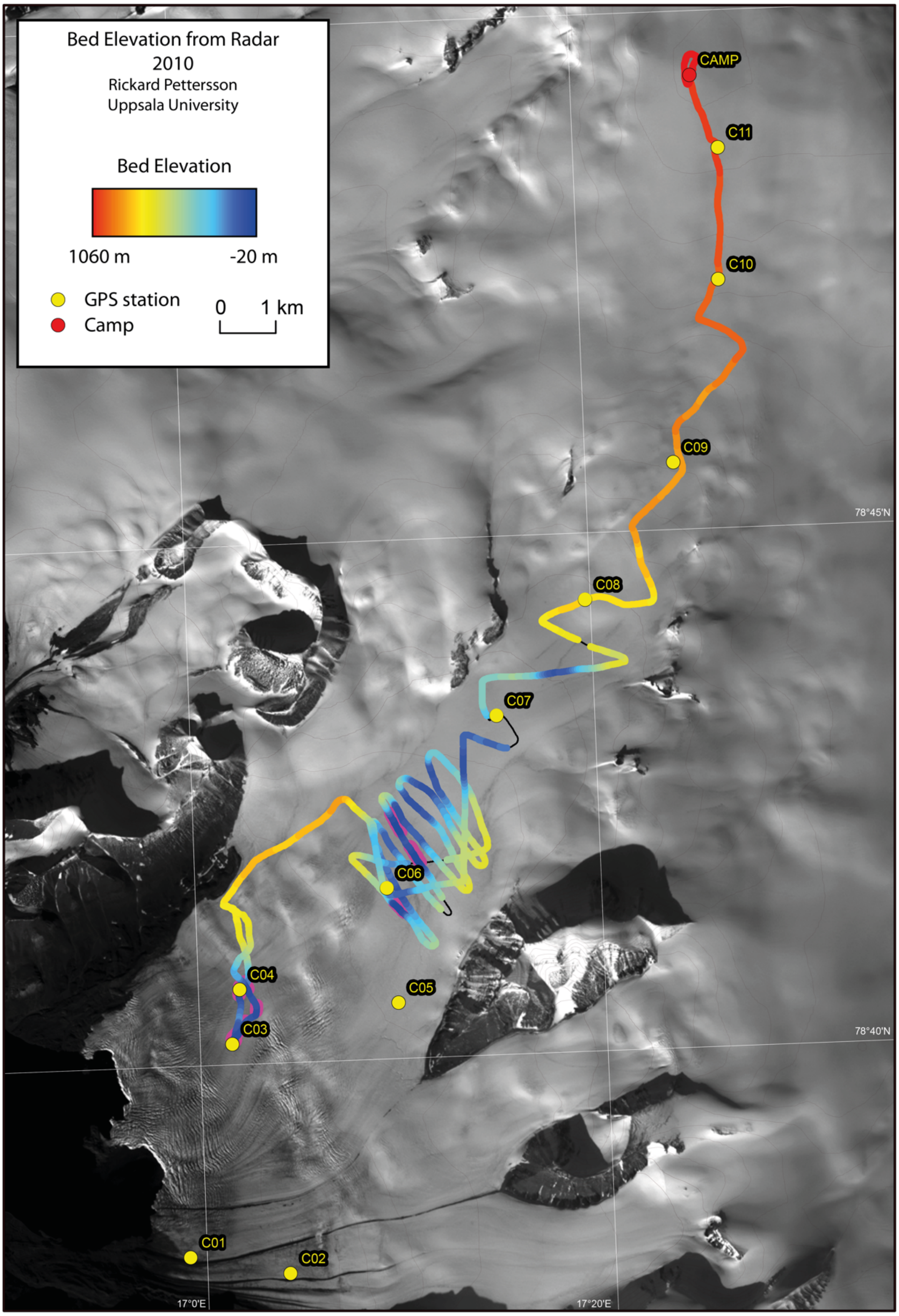


Figure 6. *GPR track of the upper part of Nordenskiöldbreen and Lomonosovfonna reveiling bed heights below sea level. (R. Petterson 2010, unpublished data)*

4. Material and methods

4.1 Material

4.1.1 Aerial images

Aerial images of Adolfbukta and Nordenskiöldbreen were taken in 1936, 1948, 1956, 1961, 1990 and 2009 (Norwegian Polar Institute; Table 2). The images were collected during flight campaigns carried out by the Norwegian Polar Institute. The aerial images from 1936, (Fig. 7) are oblique, while the rest are vertically oriented.

4.1.2 Satellite images

Twelve satellite images were obtained from USGS Earth Explorer. The satellite images used are multispectral scanner (MSS) images from LANDSAT 2, LANDSAT 4 and LANDSAT 5 and ASTER (Advanced Spaceborne Thermal Emission and Reflection Radiometer) L1T imagery (Table 2). Only images with little or no cloud cover above Nordenskiöldbreen were used.

4.1.3 Historical maps

Historical maps of Adolfbukta and Nordenskiöldbreen were created by Backlund (1908), de Geer (1910), Mathieson *et al.* (1919), Slater (1925) and Zinger (1965) (Table 2). Original versions of the maps by Backlund (1908), de Geer (1910) and Mathieson *et al.* (1919) were acquired, while Slater (1925) and Zinger (1965) were acquired in copied versions.

4.1.4 Acoustic data

High-resolution swath bathymetry datasets (2009, 2015; Fig. 7) with a grid size of 2 x 2 meters of the inner part of Billefjorden were used in this project, and the depth data are reproduced according to the permission No 13/G706 by the Norwegian Hydrographic Service.

A 6.35 KHz sub bottom profile (SBP) was achieved together with the bathymetrical dataset in 2015. The material is not processed and the material is therefore raw and is only used as support for the interpretation of the sediment cover of the seafloor.

Table 2. *List of data types and sources used for geomorphological mapping and ice front reconstruction. Year, date, type of data, source, ground resolution/scale and the original file-numbers are listed where the meta data were available.*

Year	Date	Data type	Source	Ground resolution/scale	Original numbers
1896	-	Historical map	De Geer, 1910	1 : 50,000 m	-
1896	-	Historical map	Backlund, 1908	1 : 168,000 verst	-
1896	-	Historical map	Conway, 1898	English mile	-
1908	-	Historical map	De Geer, 1910	1 : 50,000 m	-
1919	-	Historical map	J. Mathieson, 1919	1 : 100,000 miles	-
1919	-	Historical map	Zinger, 1966	-	-
1921	-	Historical map	Slater, 1925	1inch : 1mile	-
1924	-	Historical map	Zinger, 1966	-	-
1936	-	Oblique aerial	NPI	-	S36_1110
1936	-	Historical map	Zinger, 1966	-	-
1938	-	Historical map	Zinger, 1966	-	-
1948	21.08	Vertical aerial	NPI	-	104, 105, 106, 107
1949	-	Historical map	Zinger, 1966	-	-
1956	05.08	Vertical Aerial	NPI	1 : 35,000 m	6074 - 6080
1960	07.07 – 09.07	Vertical aerial		1 : 45,000 m	7004 – 7006 or
1961	23.08 – 25.08	Vertical aerial	NPI	1 : 40,000 m	2960 - 2962
1963		Historical map	Zinger, 1966	-	-
1965		Historical map	Zinger, 1966	-	-
1976	09.07	MSS Satellite Landsat 2,	USGS Earth Explorer	Pixel size 57 x 79 m	-
1982	19.10	MSS Satellite Landsat 4,	USGS	Pixel size 57 x 79 m	-
1987	19.09	MSS Satellite Landsat 5	USGS	Pixel size 57 x 79 m	-
1990	24.08	Vertical aerial	NPI	1 : 50,000 m	1902 – 1904 1855 - 1859
2002	13.07	Satellite ASTER L1T	USGS	Pixel size 15 x 15 m	-
2003	12.07	Satellite ASTER L1T	USGS	Pixel size 15 x 15 m	-
2004	13.07	Satellite ASTER L1T	USGS	Pixel size 15 x 15 m	-
2005	13.06	Satellite ASTER L1T	USGS	Pixel size 15 x 15 m	-
2007	31.07	Satellite ASTER L1T	USGS	Pixel size 15 x 15 m	-
2009	11.08	Vertical aerial	NPI	1 : 50,000 m 0.4 m pixel size	0476 - 0479
2009	-	Multibeam	Norwegian hydrographic survey	2 x 2 m grid	-
2010	03.08	Satellite ASTER L1T	USGS	Pixel size 15 x 15 m	-
2012	08.07	Satellite ASTER L1T	USGS	Pixel size 15 x 15 m	-
2014	14.08	Satellite ASTER L1T	USGS	Pixel size 15 x 15 m	-
2015	22.08	Satellite ASTER L1T	USGS	Pixel size 15 x 15 m	-
2015	August	Multibeam	UNIS	2 x 2 m grid	-

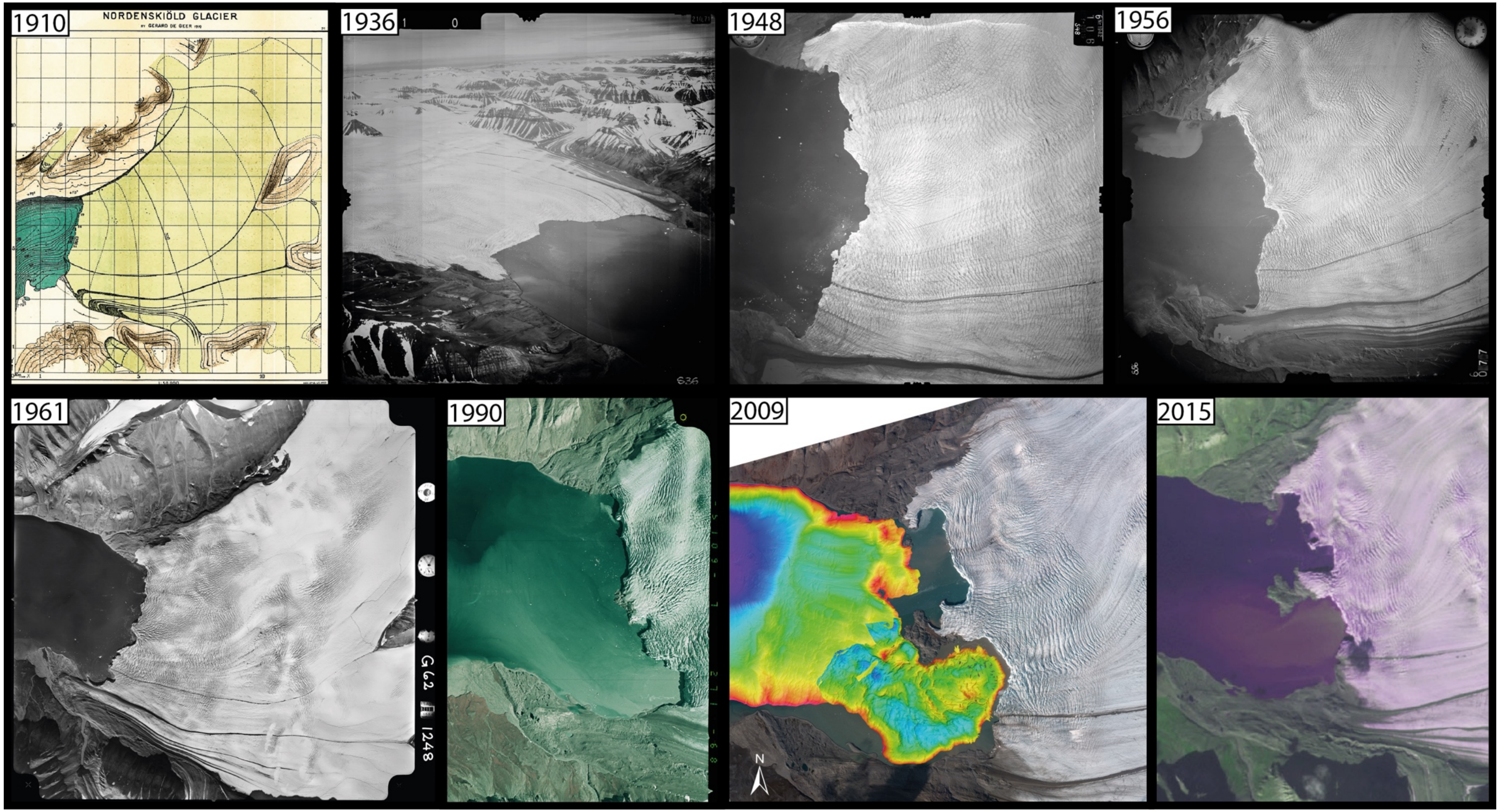


Figure 7. *Development in type of data sources from 1910 to 2015. From early work to recent studie.. De Geer's map from 1910 is highly detailed and has been easy to georeference in ArcMap. The oblique aerial image from 1938 was not possible to project and georeference in a satisfying way, however, it is still considered a valuable data source for observation of the ice front position. From 1948 all aerial images are vertical, and can be georeferenced. From 1990 the level of detail increases and the images are now in colour. From the early 2000s batymetrical data have been collected from the fjord, above the bathymetrical datasets from 2009 and 2015 are shown. Satellite imagery is available from 1976, here an example from 2015. (de Geer 1910; NPI, 1936, 1948, 1956, 1961, 1990, 2009; Norwegian Hydrographic Survey, 2009, UNIS 2015; USGS, 2015)***.**

4.2 Methods

4.2.1 Geomorphological mapping

A geomorphological map is compiled based on analyses of aerial images and bathymetrical data sets from Adolfbukta bay. Mapping will always be an interpretation of the landscape by the mapper and will therefore not be completely objective.

Aerial image pairs (2009) were mapped in stereo using the ERDAS IMAGINE 2015 with the SAFA extension for ArcMap10.3 (Fig. 8). The pixel size of the aerial images is 0.4 meters. A digital elevation model (DEM) with a 0.4 m grid was constructed from the stereo pairs from 2009. This was completed using the enhanced Automatic Terrain Extraction (eATE) tool within ERDAS IMAGINE. The contour lines are based on this DEM. To limit the detail level of the map a zoom level of 1:1500 was chosen as the limit for mapping of features in the terrestrial part. Features that could not clearly be identified at this level were not mapped.

Figure 8. *Stereoscopic mapping in the polar night. Photo: Jordan Mertes, 2015.*

The bathymetric data sets were viewed in Fledermaus software and mapped in ArcMap10.3. The landforms appearing in the bathymetry were mapped based on their shape and surface expression.

The whole basin is draped by postglacial sediment, that settled from suspension (marked with small F-letters on the map). A single unprocessed acoustic subbottom (chirp) profile was used to examine the sediment structure using Discover software and aid in the interpretation of the sediment cover.

The final geomorphological map was compiled in ArcMap10.3. The map is in scale 1:10 000, when printed on A0 (appendix). An overview map in scale 1:25 000 m and five map sections of respectively the northern shore, the southern shore and the fjord were created.

4.2.2 Georeferencing and projection

All data were imported to ArcMap and georeferenced using the UTM 33N, a Transverse Mercator projection in WGS84 datum. All data, except the already orthorectified images from 2009 and the bathymetrical data sets, were georeferenced. A number of points were identified as persistent features on the orthophotos (2009) and were used as ground control points (GCP) for georeferencing: Rudmosepynten on the northern shore, Kapp Napier and the summit of Cadellfjellet on the southern shore and the two nunataks, Terrier- and Ferrierfjellet situated ~ 6.5 km from the present glacier front (Fig. 2). On the datasets where the glacier has retreated from Retrettøya the island is used in addition. All points are not present in all the data sets.

The exact location of the spit of Kapp Napier may be changing due to wave action and redeposition, but the extensive set of raised beach ridges making up Kapp Napier is considered to be persistent enough for the time span. Rudmosepynten is made up of sedimentary rocks of Carboniferous age and is considered to be a persistent feature in the landscape within the timeframe of the reconstruction. The summit of the nunataks and Cadellfjellet are also considered to be persistent features. Retrettøya is made of Precambrian basement and is since its exposure also considered a persistent feature.

4.2.3 Symbols and legend

Polygons represent different types of surface cover and point symbols are used to indicate details in the surface cover (Fig. 9). Polylines represent linear shaped landforms such as channels, flutes and drumlinoids. Similar legend and symbols have been applied to the terrestrial *and* the subaqueous part of the map. This is done to highlight the parallels and differences between the two realms. The difference in resolution between the two data set has been a challenge, but symbol size has been chosen to fit both archives.

4.2.4 Terminology

It is important to distinguish between descriptive and genetic terms when working with geomorphology. Two terms of importance for this thesis have been subject to extraordinary many discussions between geomorphologists, terrestrial- and marine geologists.

The first problematic term is *drumlin*. In Benn and Evans (2010), the term *drumlin* is describing an oval shaped hill with a blunt proximal side and a gently sloping distal side, respectively called stoss and lee-side. This description fails to apprehend the wide range of 'drumlin' morphologies and internal compositions hereof that are found in deglaciated areas (Stokes *et al.* 2011; Dowling 2016). In a doctoral dissertation, entitled 'The drumlin problem', Dowling (2016) sheds light on the various theories concerning drumlin formation (Menzies 1979; Rose 1987; Patterson & Hooke 1995; Stokes *et al.* 2011). He argues that the less coloured term *drumlinoid* should be used to avoid problems with the different, competing theories of drumlin formation (Menzies 1979; Shaw 1983; Krüger & Thomsen 1984; Rose 1987; Shaw *et al.* 1989; Patterson & Hooke 1995; Hart 1997; Smith *et al.* 2007; Clark *et al.* 2009; Benn & Evans 2010; Johnson *et al.* 2010; Dowling 2016). Therefore, the term *drumlinoid* onwards from here is adapted in this thesis.

The other problematic term is De Geer Moraines (DGM). The term *De Geer Moraines* is according to the first description of them as annually shaped transverse moraines [*Swedish*: Årsmoräner] (de Geer 1889) an interpretational term. However, transverse ridges found in bathymetric data are with a descriptive term called *transverse ridges* as long as no clear idea of genesis is present (Ottesen & Dowdeswell 2006). This term is adapted here.

4.2.5 Field work

The terrestrial part of the map was ground verified during two field campaigns in August 2014 and August 2015, with 21 field days in all. Six days were spent on the southern shore and the rest on the northern shore. The different landforms were photo documented and sedimentological investigations were carried out at key sites. The sedimentary units were classified according to Krüger and Kjær (1999). Since sedimentology is a minor focus in this thesis, references will be made to Friis (2015), where the sedimentology of the northern, terrestrial part of the forefield was in focus.

A difference between marine and a terrestrial based Quaternary geology exists. It is logistically easier to survey the internal composition of terrestrial landforms, where in marine geology either acoustic profiles or marine sediment cores are needed to enable a final interpretation. The spectre of the snap shot you get from the two different records is thereby different. In the terrestrial part of a geomorphological map, the interpolation of the sediment cover is based on more data points, where in marine geology the interpolation is based on a limited amount of data points.

4.2.6 Retreat reconstruction

Nordenskiöldbreen has been in overall retreat since 1896 (de Geer 1910; Slater 1925; Rachlewicz *et al.* 2007). The ice front positions have been reconstructed from historical maps, historical aerial images, ortorectified aerial images and satellite images of Adolfbukta bay. Six glacier front positions between 1948 and 2009 were reconstructed from aerial images, 10 glacier front positions between 1896 and 1965 were reconstructed from historical maps and 12 glacier front positions between 1976 and 2015 were reconstructed from satellite images (Fig. 7, Table 2). The imagery was imported to ArcMap and georeferenced. The background image on the retreat line figure is the most recent ASTER image, from 2015, used in the survey (Fig. 7).

5. Results

The map covers the inner part of Adolfbukta; the marine and terrestrial forefield of Nordenskiöldbreen, making up an area of ~ 43.4 km^2 in total. Bathymetrical data constitutes ~ 14.1 km^2, the dataset from 2009 and 2015 covering ~10.9 km^2 and 3.1 km^2, respectively, with a small overlapping area. For the overlapping area, the most recent dataset is used as basis for the analyses. Area with no bathymetrical data makes up ~ 2.6 km^2. In the legend, this is labelled 'fjord, no bathymetry data'. The two small islands in the fjord together are ~0.2 km^2. The northern and southern part of the terrestrial area constitutes ~ 1.9 km^2 and ~ 4.5 km^2, respectively, and glacier and medial moraines together constitute and area of ~ 20.1 km^2.

5.1 Subglacial landforms

5.1.1 Scoured bedrock

<u>Description</u>

The total area of this type of landscape is ~ 2.5 km^2 (Figs 10 – 15): ~ 0.4 km^2 on the northern shore, ~ 0.15 km^2 on the southern shore, ~ 0.2 km^2 on Retrettøya in the fjord and ~ 1.7 km^2 in the subaqueous record. The rocks rise up to ~ 80 m.a.s.l. on the northern shore.

On the northern shore, it dominates the proximal part of the forefield and the area along the coast (Figs 11, 13). On the southern shore it is only found very proximal to the glacier (Fig. 13). Distal to the terrestrially exposed rocks, glaciofluvial deposits are found (Figs 11, 13). In the fjord basin, it dominates the most proximal part of the bathymetrical dataset and it can be traced on to Retrettøya in the fjord (Fig. 10). The area appears irregular on the bathymetrical dataset, and it occurs mainly at shallow water depths (≤ 50 m).

The subaerially exposed rocks generally have smooth surfaces and some surfaces are observed to be carved. Small erosional furrows (0,5 – 1 cm wide) occur on all the exposed rock surfaces, with orientations varying from 44° - 70° (azimuth angle). A thin layer of glacial drift occurs spotwise on the surface. Several small channels have cut into the rocks and small amounts of sorted sediment have accumulated in the depressions. The distal sides are often plucked, and the mean strike of the plucked surfaces is ~160°.

<u>Interpretation</u>

The area is interpreted as glacially scoured bedrock, and most of the bedrock knobs can be classified as *roche moutonées* based on their smooth proximal faces and plucked distal sides (Sugden *et al.*

1992; Benn & Evans 2010). The carved surfaces are interpreted as plastically moulded forms (p-forms), indicating presence of high pressures at the base of the ice during advance (Dahl 1965; Benn & Evans 2010). Since the till plain is located in the downflow direction of the p-forms, they are assumed to have formed due to carving by saturated till flowing between the ice and bedrock (Benn & Evans 2010).

The small furrows are interpreted as glacial striae, representing former ice flow direction (Kleman 1990). The plucking on the distal side of the rock obstacles differs from the mean orientation of streamlining in the landscape and is interpreted to correspond with planes of weakness in the bedrock (Sugden *et al.* 1992; Benn & Evans 2010). However, it could suggestively also represent pre-LIA ice flow directions.

The exposed bedrock is interpreted to have worked as pinning points for the glacier during retreat (Warren 1992; Wiles *et al.* 1995; Ottesen *et al.* 2008). On the aerial image from 1956 (Fig. 7) it is seen that the ice covers the bedrock in the eastern part of the forefield. The image from 2009 (which the map is based on) shows that the ice has just recently retreated across the bedrock. Since the ice has rested on the bedrock for a while, it is likely to assume that a continuous supply of englacial meltwater has limited the accumulation of subglacial material. This could suggestively also explain the low amount of glacial drift observed in the bedrock area.

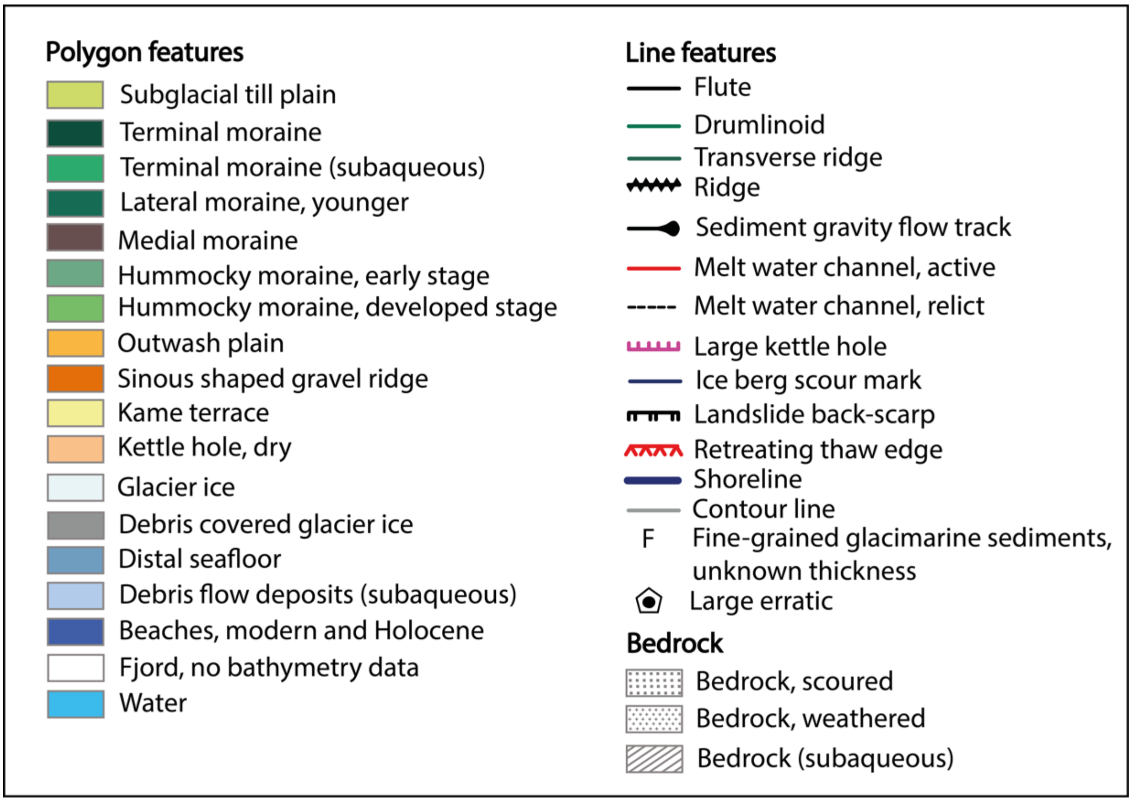


Figure 9. *Legend for the geomorphological map.*

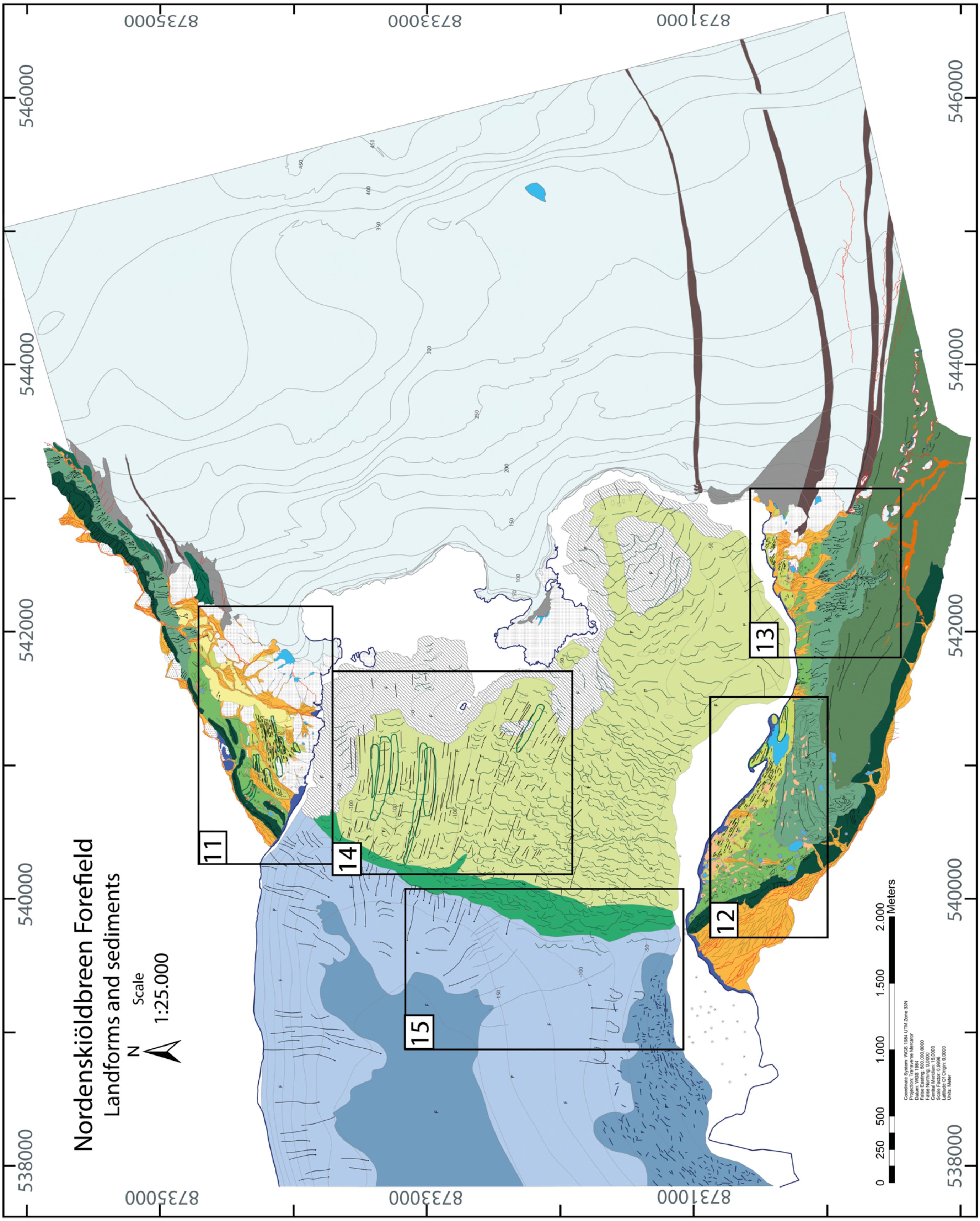


Figure 10. *Geomorphological map of the Nordenskiöldbreen forefield. See online supplementary data for full version of map. Black boxes indicate map sections.*

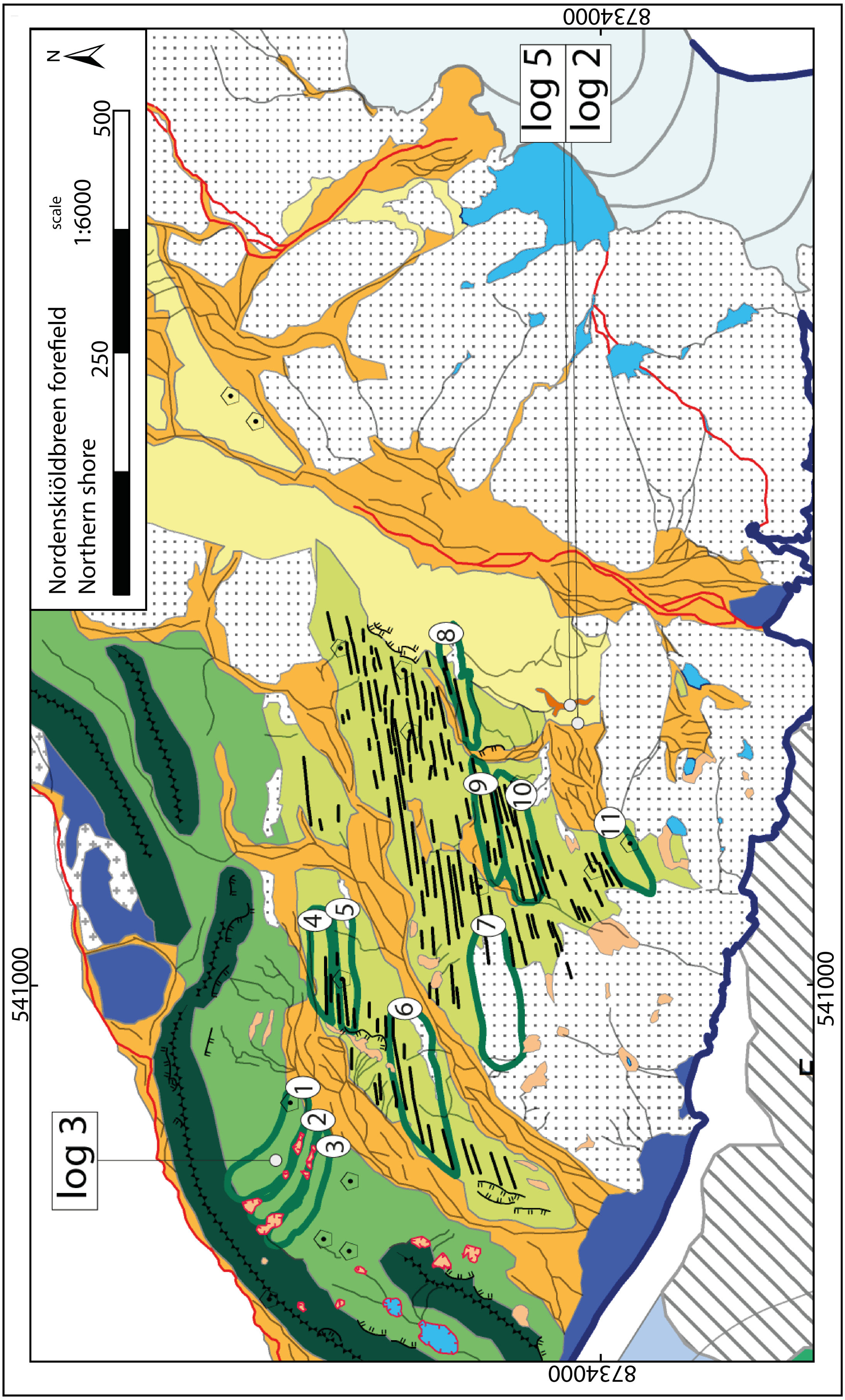


Figure 11. *Map section of the northern, terrestrial forefield of Nordenskiöldbreen. Drumlinoids are labelled no. 1 – 11. Location of sections no. 2, 3 and 5 are marked (log2, 3 and 5).*

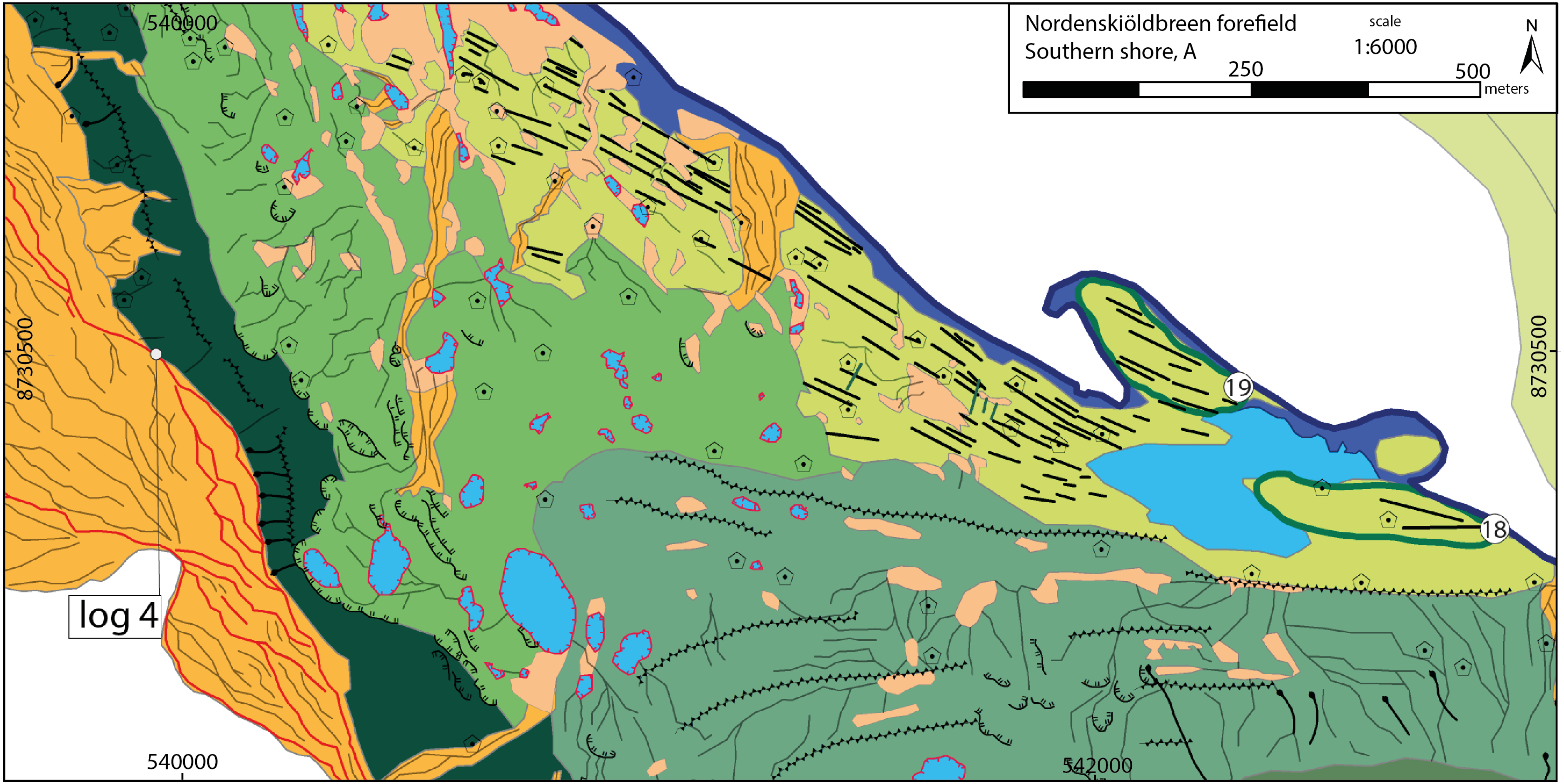


Figure 12. *Map section of the western part of the southern terrestrial shore of Nordenskiöldbreen forefield. Drumlinoids no. 18 and 19 are marked, as well as location of section 4 (log4).*

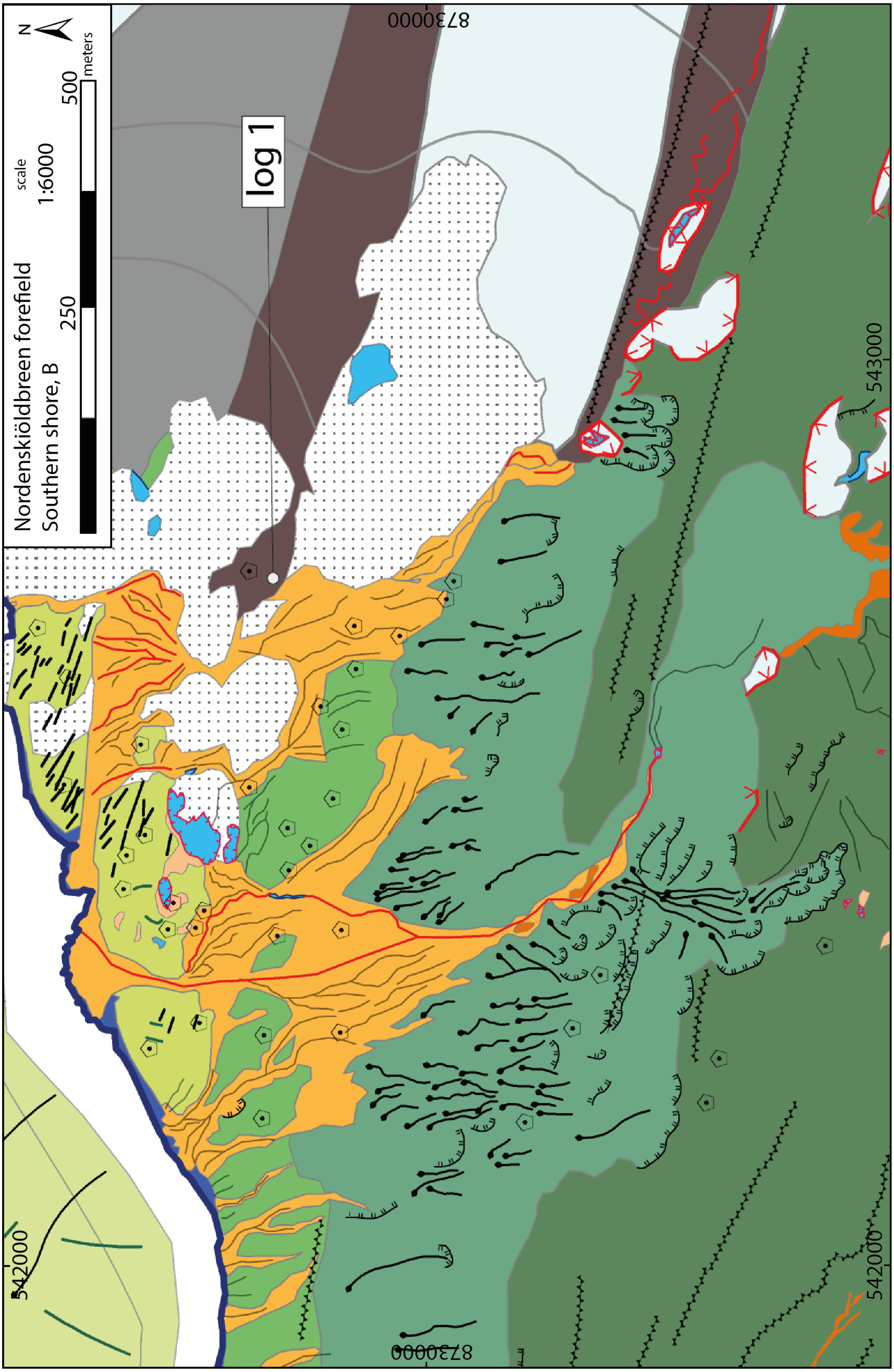


Figure 13. *Map section of the eastern part of the southern, terrestrial shore of Nordenskiöldbreen forefield. Location of section 1 is marked (log1).*

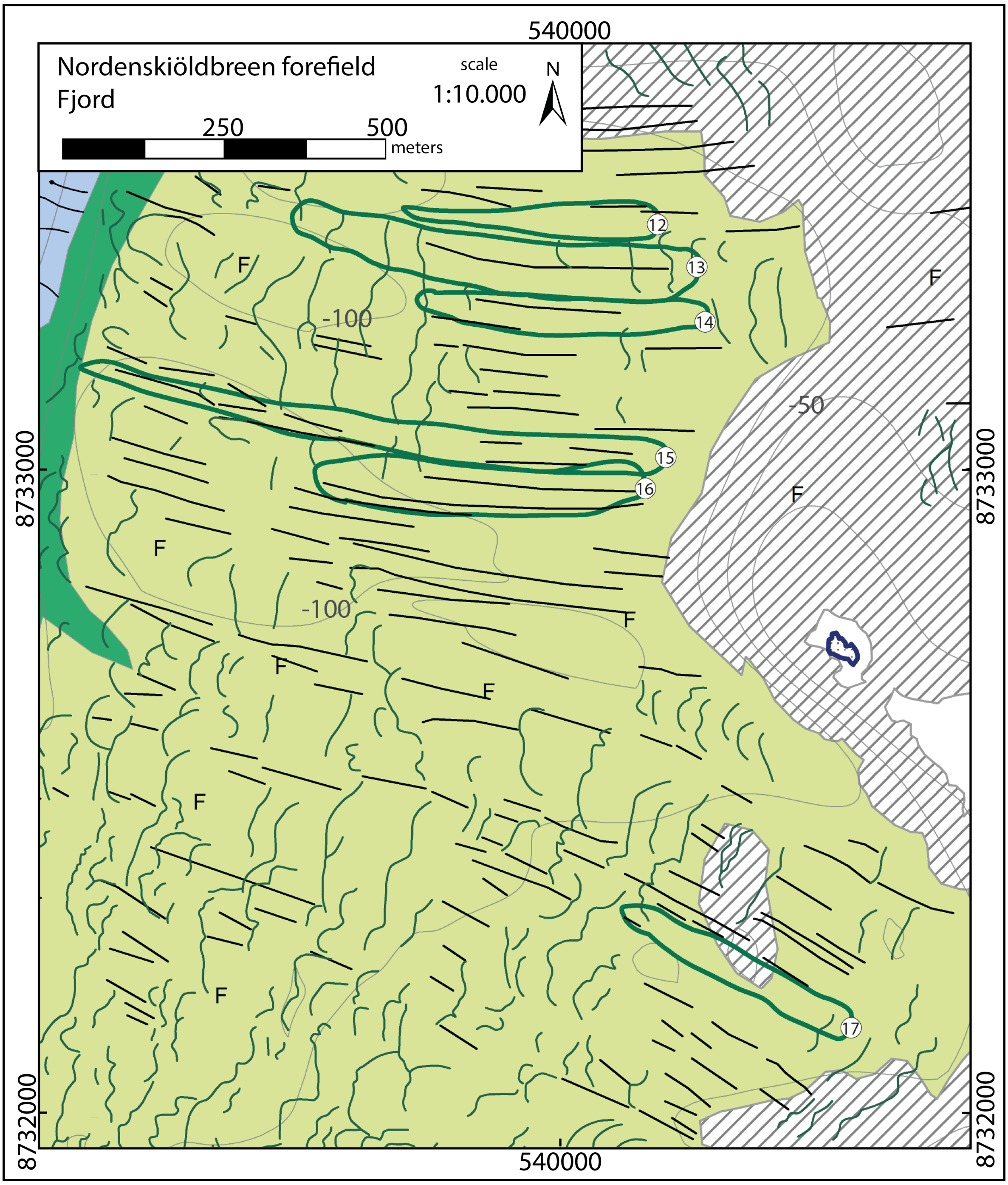


Figure 14. *Map section of the proximal part of the basin of Adolfbukta. Drumlinoids no. 12 – 17 are labelled. Characteristic for this map section is a streamlined substratum with transverse ridges superimposed thereon. Scoured subaqueous bedrock is seen in the eastern part of the map section. Maximum water depth there is ~100m.*

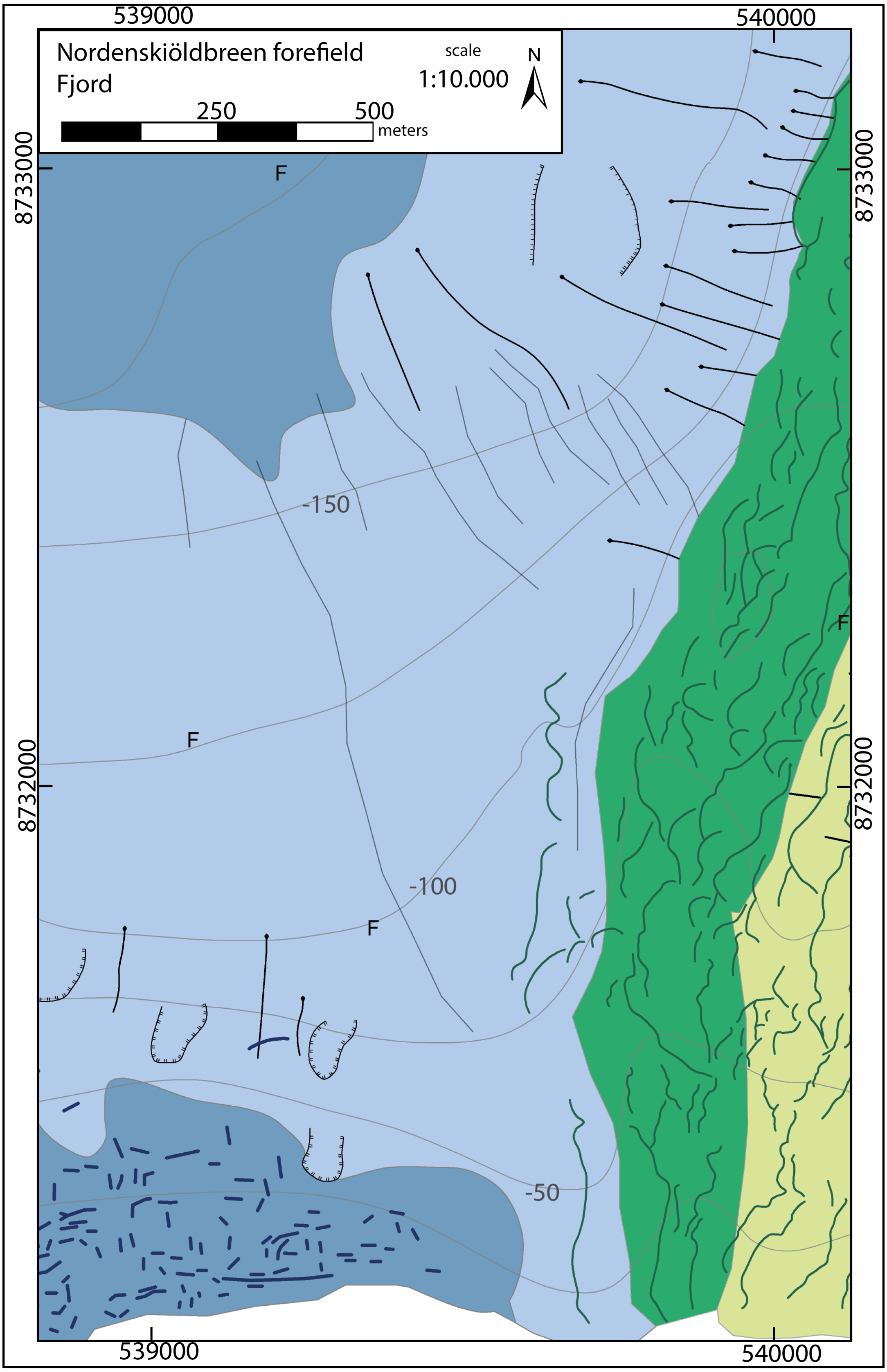


Figure 15. Map section of the distal part of the Adolfbukta basin. Water depth increases distal to the terminal moraine, ice berg plough marks occur in the shallowest part of the basin (<50m depth). Transverse ridges overprints the terminal moraine. Debris flow lobes occur outboard of the terminal moraine. Sediment gravity flow tracks and a few land-slide back-scarps occur on the lobe surfaces.

5.1.2 Subglacial till plain

Description

The area with this label covers ~ 5.4 km^2, ~ 0.1km^2 on the northern shore (Fig. 11), ~ 0.2 km^2 on the southern shore (Figs 12, 13). It covers the central part of the northern shore, in between the exposed bedrock and terminal moraine. On the southern shore, the same material occurs in two 100 - 200 m wide bands parallel to the beach. The two bands are separated by a ~ 500 m long area (mapped with a different legend) with extensive slumping activity.

On the southern shore, it is only elevated a few metres above the sea, but on the northern shore it is ~ 25 m.a.s.l. The plain consists of a clast-rich, silty to sandy diamict. It shows clear fissility when excavated. On the surface, the diamict is coarse with a large number of scattered erratics ($d \leq 2m$). Large boulders that are lodged in the diamict have several sets of superficial striae. Very large erratics occur on the southern shore. Shell fragments occur both on the surface and within the diamict.

From a natural section cut by the active meltwater channel right in front of the glacier on the southern shore, a succession of sediments could be observed (Fig. 16). At the base, dark basement rocks of Hecla Hoek are exposed. A ~ 90 cm thick unit of diamict overlies the basement. There is a colour change throughout the unit from almost similar colour as the bedrock at the base to brighter colours in the upper part. It contains clasts of various sizes and roundness, but the main lithology is Hecla Hoek. The unit is given the lithofacies code DmF(m_3)3, according to Krüger and Kjær (1999). The unit is interbedded with the next unit, a sandier, extremely firm diamict, given the lithofacies code DmF(m_2)4. The boundary to the following unit is gradational. This unit is a coarse-grained diamict, it is easy to excavate compared to the two underlying units, and it is given the lithofacies code DmC(m_2)2. A single lense of sandy material is observed within this unit. On the surface are boulders with little or no fine-grained material in between observed. Fissility is observed in the two lower units.

Interpretation

The clast-rich, silty to sandy diamict found in the northern shore is, on the basis of its composition, fissility, the appearance of striated and lodged boulders both at the surface and within the unit, interpreted to be subglacial till (Boulton 1970a; Ruszczyńska-Szenajch 2001; Evans *et al.* 2006). The abundance of shell-fragments indicate that the till at least is partly composed of re-worked marine deposits (Friis 2015). It is interpreted to have been reworked and deposited during the most recent advance of Nordenskiöldbreen, the LIA advance (Slater 1925; Rachlewicz *et al.* 2007).

From the section the following interpretations were made. The diamict unit immediately above the bedrock is interpreted to represent local bedrock that has been grinded and incorporated into the subglacial material. Both this lower unit and the unit above it, are interpreted to be subglacial till, formed by a mix of grinding of local bedrock and reworking of previously present material. These two units appear similar to the sediment described from the subglacial till plain on the northern shore by Friis (2015). The unit at the top is interpreted as a meltout till from en- and supraglacial material and is interpreted to originate from the medial moraine. The material from the medial moraine has settled down onto the subglacial till.

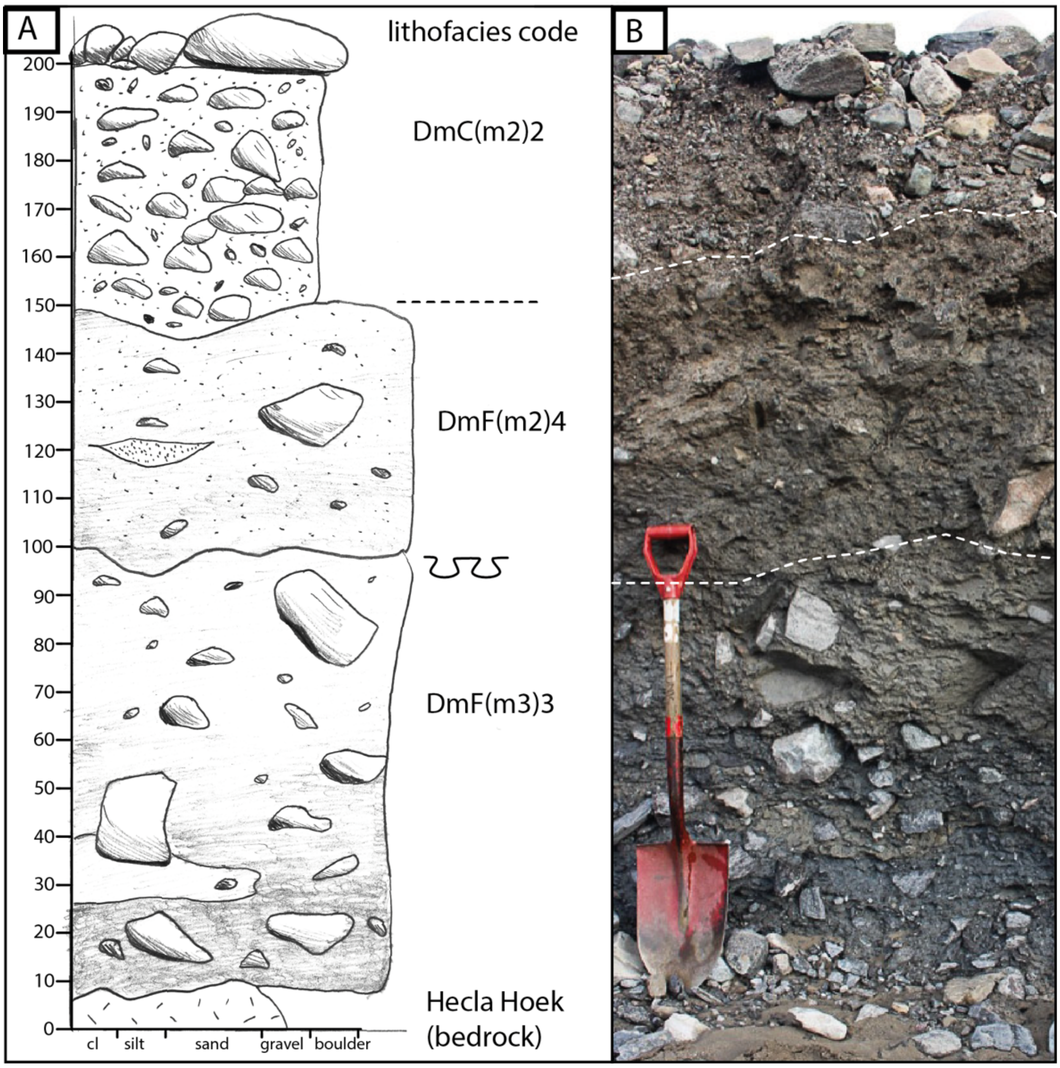


Figure 16. *Log of section 1, for location see Fig. 13. A) The lowermost part of the section is interpreted as subglacial till with a large component of incorporated local bedrock. The colour gets brighter towards the top. The uppermost unit is interpreted as a melt-out till. B) Photo of the section.*

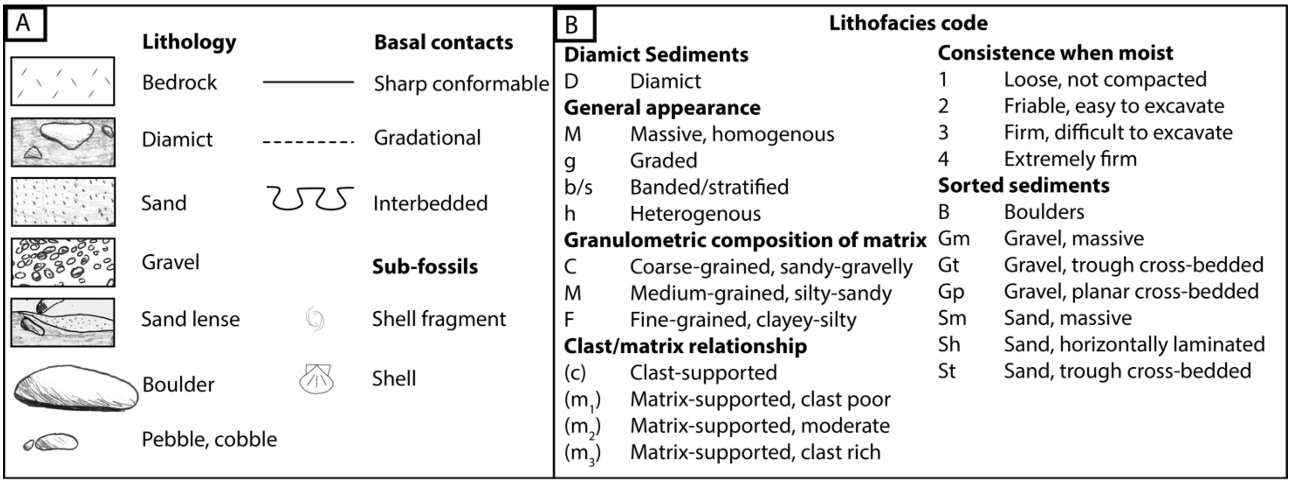


Figure 17 *Explanation of lithofacies codes and symbols used in the five logs/sketches.*

5.1.3 Glacigenic material, sculptured

Description

Sculptured material making up an area of ~ 5.0 km^2 occurs in the subaqueous realm (Figs 10, 14, 15). It covers the area in between the terminal moraine and exposed bedrock - in the eastern part of the bathymetrical dataset. The surface expression is dominated by elongated ridges parallel to ice flow and ridges transverse to ice flow.

On the SBP (Fig. 18b) repeated hyperbolas with up to 5 m thick packages of acoustically stratified material in between are observed. The acoustic basement is ~ 10 m below the sea floor surface.

Interpretation

Due to the presence of streamlined landforms, the clear acoustic basement and transverse ridges, the area is interpreted to consist of diamict material, that has been deposited by the glacier (Plassen *et al.* 2004; Flink *et al.* 2015). The transverse ridges only form when diamict material is available (Flink *et al.* 2015), and the acoustic basement is interpreted to be till. Penetration depth is not deep enough to reach the bedrock floor. Each hyperbola on the SBP is interpreted to represent a transverse ridge and the stratified material in between is interpreted to be laminated glacimarine sediment. It is assumed that a layer of laminated glacimarine sediment, of unknown, various thickness drapes the whole area (Plassen *et al.* 2004).

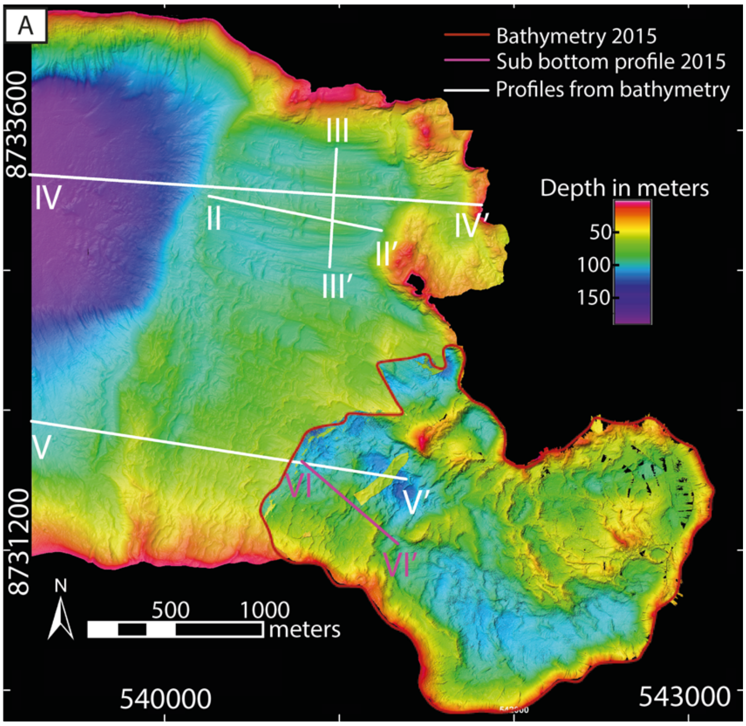
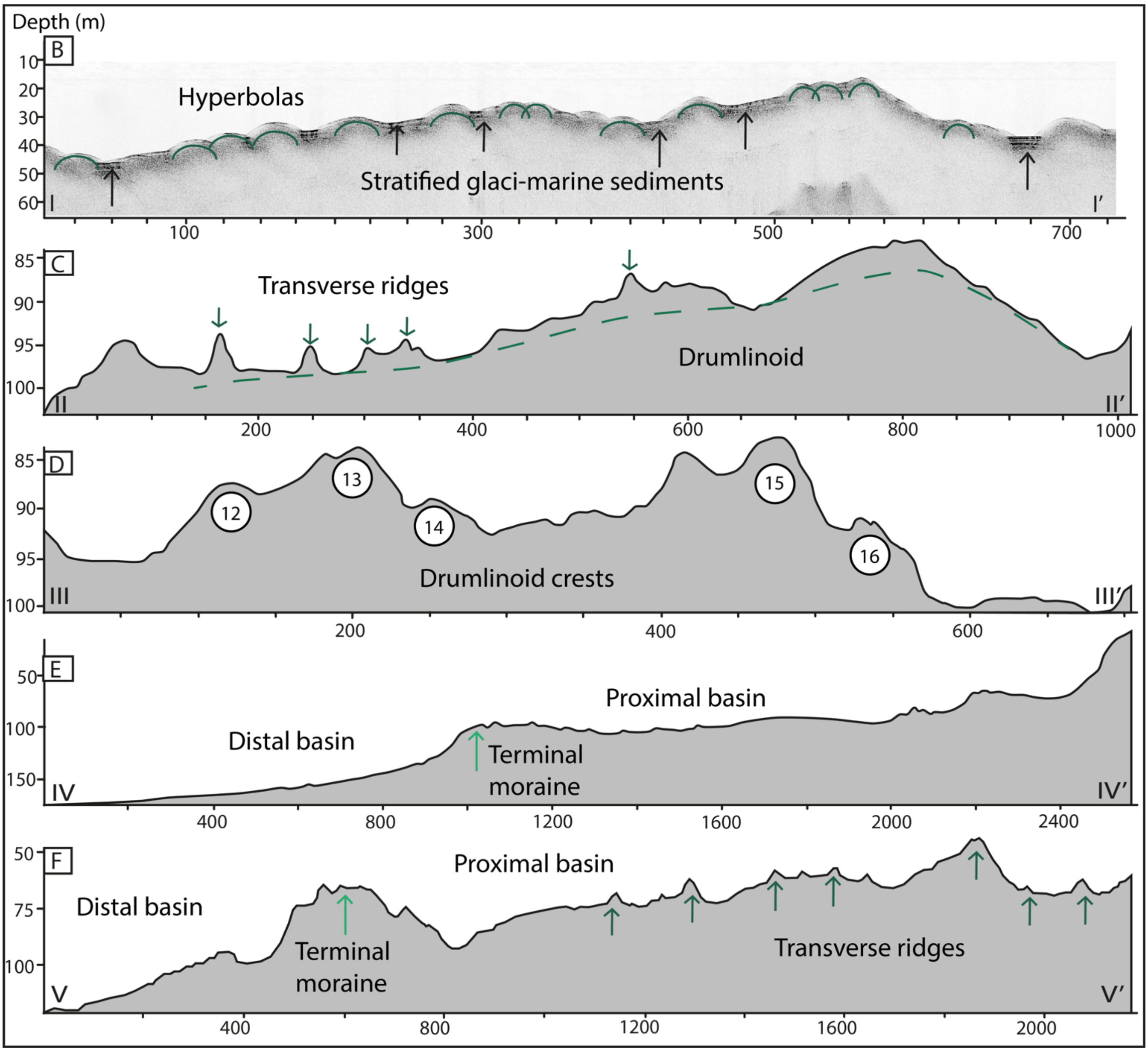


Figure 18. Acoustic data. A) Overview of the bathymetrical datasets from Adolfbukta. The northern basin is dominated by streamlined landscape and the southern basin by transverse ridges. The irregular surfaces in the north-eastern part are interpreted as scoured bedrock. B) An unprocessed SBP (2015) showing acoustically stratified glacimarine sediments in between hyperbolas. The hyperbolas are interpreted as reflections from transverse ridges. C) Bathymetry profile of drumlinoid no. 15 with superimposed transverse ridges. D) Bathymetry profile of drumlinoid 12 – 16. E) Bathymetry profile of the northern basin showing the slope break and the low amplitude terminal moraine. F) Bathymetry profile of the southern basin showing a terminal moraine of higher amplitude and transverse ridges.



5.1.4 Glacial lineations (flutes)

Description

A total number of 499 glacial lineations have been mapped (Fig. 10). 172 were identified on the northern shore (Figs 11, 19a), 151 on the southern shore (Figs 12, 13, 19b) and 176 in the subaqueous realm (Figs 14, 18a). They appear in swarms, closely spaced. The landforms vary in length from ~ 3.9 to ~ 438.9 m, heights are up to 3 m and the widths range from ~ 1 to ~ 20 m. Orientation of the landforms (azimuth angles) range from 56° to 141°. The landform orientation on the northern shore varies from 59° to 141°, the subaqueous from 80° to 141° and at the southern shore from 89° to 135°.

On land, the lineations are prominent, low relief ($h \leq 0.5$ m) features. The lengths of the terrestrial lineations range from ~ 3.9 m to ~ 145.1 m and widths from ~ 1.0 to ~ 7.6 m. Some ridges curve around obstacles or other lineations. The surface contains coarse subrounded to subangular clasts of various lithologies and sizes and fines are absent. Large boulders are found in the proximal end of several of the landforms (Fig. 19). The surface expression varies, and some ridges have a more pronounced relief than others. The landforms with the least surface expression consist almost only of larger rocks.

The subaqueous examples range from ~ 24.7 to ~ 438.9 m in length and ~ 5.0 to ~ 20.0 m in width. Distance between the crests of the ridges vary form ~ 10 to ~ 100 m and the landforms rise ~ 0.5 to ~ 3 m above the surrounding seafloor. Due to the resolution (2 x 2 m grid) of the bathymetry, it is not possible to identify if the lineations have initiating boulders in subaqueous record. There is a larger concentration of lineations in the northern part of the basin.

Interpretation

The appearance of the landforms in sub-parallel swarms, the diamictic composition and inclusion of erratic clasts in the material leads to the interpretation of the landforms as glacial lineations/flutes (Hoppe & Schytt 1953; Boulton 1976; Benn 1994; Friis 2015). Their orientations are interpreted to reflect the ice-flow direction during the LIA advance. They are interpreted to have been produced at the ice-bed interface by soft-sediment deformation (Benn & Evans 2010; Flink *et al.* 2015). This is in accordance with observations by Boulton (1976). Fabric analyses presented by Friis (2015) also support this. The absence of fine material in the surface of the terrestrial examples could be explained by nivation and delation processes, caused by spring snow melt and katabatic winds (Christiansen 1998; Frenot *et al.* 1998; Ballantyne 2002). The preservation potential of the landforms is therefore limited by the efficiency of these processes.

The flutes observed in the marine record are significantly longer than the ones in the terrestrial record. This could be explained by the coarser resolution of the bathymetrical dataset (2 x 2 m grid), whereby smaller landforms are not captured.

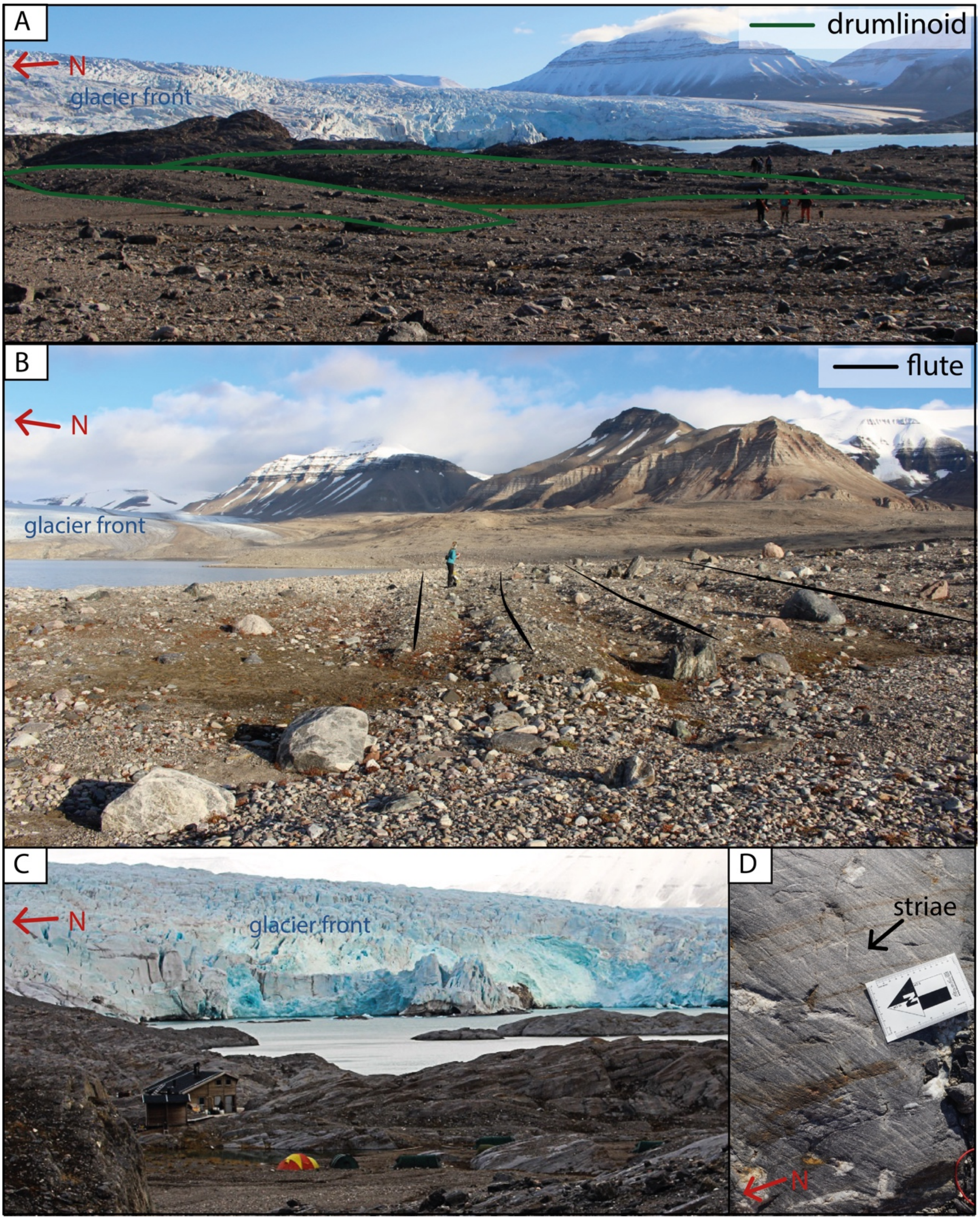


Figure 19. *Subglacial landforms. A) Fluted till plain on the northern shore, drumlinoids no. 9 and 10 are seen central in the photo. B) Fluted till plain on the southern shore. C) Scoured bedrock in the glacier proximal area, northern shore. D) Scoured bedrock with several sets of striae. Photo13D: Nina Friis 2015.*

5.1.5 Drumlinoids

Description

Half egg shaped hills with blunt stoss sides and tapering lee sides occur on the subglacial till plain (Fig. 10). Nineteen landforms, labelled no. 1 – 19 on the map (Figs 11 - 15), have been identified as this type, six in the subaqueous record (Figs 14, 18c, d), eleven on the northern shore (Fig. 11) and two on the southern shore (Fig.12). The lengths range from ~ 86 to ~ 921m, heights are up to ~ 15 m and widths are from ~ 20 to ~ 89 m. The largest examples of these bedforms occur in the subaqueous record, with lengths up to twice the length of their terrestrial counterparts. The heights above the surrounding seafloor range from ~ 10 to ~ 15 m. The highest terrestrial example is ~ 5 m. The elongation ratios range from 14.6:1 to 2.7:1 (Table 3).

Some are easy identified from aerial imagery, others were identified in the field and others again only from the stereoscopic mapping. The landforms create a gentle relief in the ground moraine, and they are overprinted by small glacial lineations in both archives (Figs 18, 19).

Table 3. *Drumlinoid elongation ratio, for location of each drumlinoid see Figs 12 – 14. Drumlinoids no. 1-11 (marked with green in the table) are located on the northern shore, drumlinoids no. 12 – 17 (light blue) are located in the fjord and drumlinoids no. 18 – 19 (darker green) are located on the southern shore.*

Drumlinoid no.	L (length, m)	W (width, m)	Elongation ratio (L/W)
1	128	33	3.88
2	123	20	6.15
3	134	25	5.36
4	106	27	3.93
5	131	30	4.37
6	161	52	3.10
7	154	46	3.35
8	129	24	5.38
9	124	20	6.2
10	139	36	3.86
11	86	32	2.69
12	398	62	6.42
13	643	89	7.22
14	449	86	5.22
15	921	63	14.62
16	513	86	5.97
17	415	64	6.48
18	269	55	4.89
19	219	62	3.53

Interpretation

Based on the morphology and elongation ratio, the hills are interpreted as drumlinoid bedforms (Menzies 1979; Ottesen *et al.* 2008; Johnson *et al.* 2010). Drumlinoid formation and composition have been discussed through decades (Smalley & Unwin 1968; Shaw & Kvill 1984; Rose 1987; Hart 1997; Briner 2007; Dowling 2016) and the term drumlinoid is used here to prevent pre-judgement of the genesis of the landforms. To determine the genesis, sedimentological investigations of the internal composition are needed.

Drumlinoids are described to have long axes >100 m long and elongation ratios up to 7:1 (Rose 1987). All the landforms in this study, except two, categorized as drumlinoids accomplish these drumlinoid criteria. Drumlinoid no. 11 has a long axis of ~ 86 m and no. 15 has an elongation ratio of 14.6:1. However, since the landforms in this work are called drumlinoids and not drumlins, the two landforms that do not fit with the 'drumlin criteria' by Rose (1987) are kept in the drumlinoid category.

Drumlinoids no. 4 to 10 were easily identified as drumlinoids during the first period of fieldwork in 2014. No. 8, 9 and 10 have been investigated by Friis (2015) and can therefore be interpreted as 'true' drumlinoids. Drumlinoid no. 1 - 3 were first categorized as curving, streamlined ridges, but their elongation ratio and shape fit with the drumlinoid classification. However, due to their location in immediate extension of the broad flutes it can be argued that drumlinoid 1-3 represent a transitional stage between flutes and drumlinoids.

5.2 Ice-marginal landforms

5.2.1 Debris flow lobes (subaqueous)

Description

Approximately 7.1 km^2 are mapped with this category in the forefield. Distal to the broad ridge in the middle of Adolfbukta, lobes are observed in the bathymetry (Figs 10, 15). The lobes that occur parallel to ice flow, can be traced across the fjord and they have lengths up to ~ 860 m. Their profiles are steep the proximal part and flatten out towards the fjord bottom, and the slope descends from ~ 100 m.b.s.l to ~ 160 m.b.s.l. Similar lobes, but perpendicular to ice flow, are observed in the outer part of Adolfbukta descending from the coastlines. They reach lengths up to ~ 1200 m. The lobes perpendicular to ice flow have steeper profiles in the northern part of the basin, which is reflected by the close spacing between the contour lines on the map. Erosive furrows, scars perpendicular to ice flow and gravity flow tracks occur on the surface of the lobes. The individual lobes are not mapped, as they overlap.

Interpretation

Due to their shape and occurrence of debris flow tracks on the surface the lobes are interpreted as debris flow lobes (Broster & Hicock 1985; Dowdeswell *et al.* 1998; Ottesen & Dowdeswell 2006). They are interpreted to have built out while the glacier was at its LIA maximum position. Large quantities of material were transported to the ice front and debris flows initiated when the slope got unstable. Baeten *et al.* (2010) interpreted the material as mass-transport-deposits. From an acoustic profile only one layer of debris flow lobes outboard of the LIA moraine was identified (Plassen *et al.* 2004; Fig. 5). Similar flow lobes have been observed in front of surge-type tide water glaciers in Svalbard, however there they are stacked (Ottesen *et al.* 2008; Flink *et al.* 2015).

5.2.2 Large lobate ridge (terminal moraine)

Description

The total area mapped with this polygon makes up ~ 1.1 km^2: an area of ~ 0.2 km^2 is identified on the northern shore (Figs 11, 20), ~ 0.3 km^2 on the southern shore (Figs 12, 13, 20) and ~ 0.55 km^2 in the subaqueous record (Figs 10, 14).

The outer margin of the forefield on both northern and southern shore and in the subaqueous record is marked by a semi-continuous ridge. The ridge on the northern shore is up to ~30 m high compared to the surrounding terrain, and the width varies from ~ 20 to ~ 80 m. In the very proximal part of the southern shore the ridge is up to ~ 100 m high. The width varies from ~ 30 to ~160 m. A broad ridge, ~ 70 to ~ 350 m wide, transverse to iceflow, can be traced across the bay in the bathymetrical dataset. It rests on a slope break and splits the basin in two parts. The seafloor drops to ~ 150 m.b.s.l. distal to the break.

On land, the slopes of the ridges are steep, and the distal slope tends to be steepest (Fig. 20). The surface material consists of subrounded to angular debris, with some very large boulders in between. The crest of the ridge on the northern shore is roughly oriented NE-SW and it almost merges with the ice front in north eastern corner. The landform rises above the remaining terrain. On the northern shore a face with a lot of slumping activity and a ridge with a large ice content have been exposed (Fig. 20c). Where the ridge reaches the sea it is eroded by waves and the surface material slumps (Fig. 20d). On both the northern and southern shores the distal parts of the ridges are cut by fluvial channels. The ridge some places appear double crested with an elongated depression in between (Fig.

20a). Particularly, the southern shore ridge has been subject to fluvial outflows where fine-grained material has been washed away. The ridges have a higher relief closer to the glacier front.

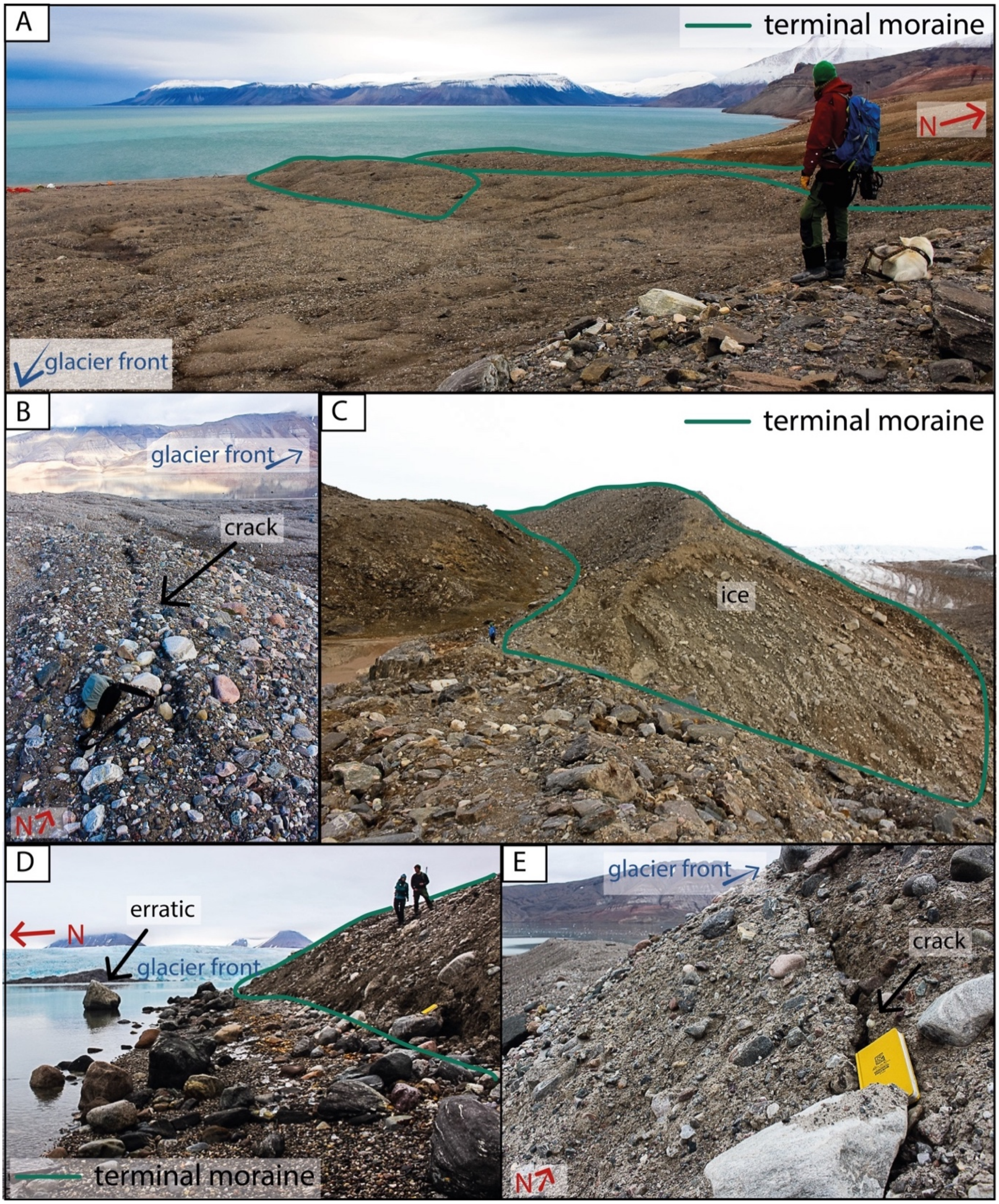


Figure 20. *Terminal moraine. A) Lower relief, terminal moraine in the distale part of the forefield, northern shore, B) crack in terminal moriane, southern shore. C) Interior ice in terminal moraine, northern shore. D) Wave erosion of terminal moraine, southern shore - note the large amount of erratics. E) Crack in terminal moraine southern shore.*

In the basin the ridge is observed to have a higher relief in the southern part of the bay. An offset between the northern terrestrial ridge and the northern part of the subaqueous ridge occurs (Fig. 10). Streamlined landscape, areas of heavily slumping activity and scoured bedrock dominates the area inboard of the broad ridge. Outboard the landscape is deeply affected by periglacial weathering processes; the bedrock has a very irregular surface and examples of well developed sorted circles are found immediately north of the broad ridge on the northern shore. The semi-continuous ridge frames the forefield and the landscape outboard of the ridge differs significantly from the landscape inboard of the ridge.

Interpretation

The height, width and location of the semi-continuous ridge as an outer frame of the glacier forefield leads to the interpretation of the landform as a terminal moraine (Bennett 2001; Benn & Evans 2010). The interpretation is further supported by the clear difference in landscape expression in- and outboard of the ridge. The terminal moraine has previously been interpreted to indicate the LIA maximum extent (Plassen *et al.* 2004; Baeten *et al.* 2010; Benn & Evans 2010). Due to the similarity in material and its semi-continuity, the lobe-shaped ridge is described as one landform. On the background of the abundant cracks in the surface, gravitational sorting, slumping on the slopes and the case where backwasting was observed (Fig. 20c) it is assumed that the terrestrial part of the ridge contains an ice-core (Etzelmüller 2000; Schomacker & Kjær 2008).

5.2.3 Small-scale transverse ridges (recessional moraines)

Description

A total number of 641 transverse ridges have been mapped (Fig. 10): 635 from the bathymetrical datasets (Figs 14, 18) and 6 from the terrestrial, southern part (Fig. 13). The landforms in the subaqueous part are ~ 15 to ~ 505 m long, ~ 0.5 to ~ 3 m high and ~ 10 to ~ 20 m wide. The distance between the crests of the ridges varies from ~ 40 to ~ 150 m. The six examples on land have low relief ($h \leq 0.5$ m), are ~ 12 to ~ 37 m long and range from ~ 2.85 to ~ 3.85 m in width.

The overall orientation (azimuth) of both terrestrial and marine ridges range from 0° to 180°, generally transverse to the fjord axis. They appear as pointy, symmetric features in the cross profile of the terrace in front of the glacier (Fig. 18a, b, f). The ridges dominate the morphology inboard the LIA in the fjord basin, and the concentration is highest in the southern part of the basin (Fig. 18a). The ridges are superimposed on the glacial lineations both in the northern part of the basin and on the southern shore (Figs 10, 18). The terrestrial examples have a very low relief, appear very close to the sea and are *only* found on the southern shore.

Interpretation

The rhythmicity, close spacing and overprinting of the glacial lineations lead to the interpretation of the landforms to be recessional moraines (Lindén & Möller 2005; Ottesen & Dowdeswell 2006; Flink *et al.* 2015). Similar landforms are found in front of other tidewater glaciers in Svalbard (Ottesen & Dowdeswell 2006; Ottesen *et al.* 2008; Flink *et al.* 2015) and their exact genesis is widely discussed (Bouvier *et al.* 2015; Flink *et al.* 2015). They have been documented to form in the subaqueous environment (Bouvier *et al.* 2015). The proximity to the sea makes it reasonable to interpret the terrestrial examples as similar landforms as the subaqueous transverse ridges. They appear in the transition zone, where the glacier ice has been affected by the dynamic behaviour of the marine part of the glacier.

5.2.4 Kame terraces

Description

About 0.13 km^2 are mapped with this label, and it is only found on the northern shore (Figs 11, 21, 22). The outline of the area is shaped as an amphitheatre, with terrace steps descending towards the dry river bed (Fig. 22). The orientation of the terrace steps mimics the glacier front.

A 1.30 m section was cleared at the edge of the western-most terrace, where ten units were identified (Fig. 21). The lower-most unit constitutes massive gravel (Gm) and the next four units alternates between trough cross bedded sand and gravel (St and Gt). There above a unit of massive sand (Sm) occurs, which is followed by two units of trough cross bedded sand and gravel (St and Gt). The two upper-most units constitutes massive sand (Sm) and boulders (B) occur at the surface.

Interpretation

The westernmost terrace step matches the location of the ice front on the 1956 aerial image (Fig. 22a). Together with the shape and the sorted material in the section this leads to an interpretation of the area as kame terraces, deposited in ice-dammed lakes (Donnelly & Harris 1989; Goudie 2004). The trough cross stratified units indicate presence of bottom currents and the ice-walled lake is interpreted to have been shallow.

A modern analogue is seen on the aerial image from 2009. There small ice dammed lakes occur in front of the glacier. The lakes do not exist today and seen from the satellite images from 2015 it is clear that the ice has retreated, and the ice supported lake therefore has drained since 2009.

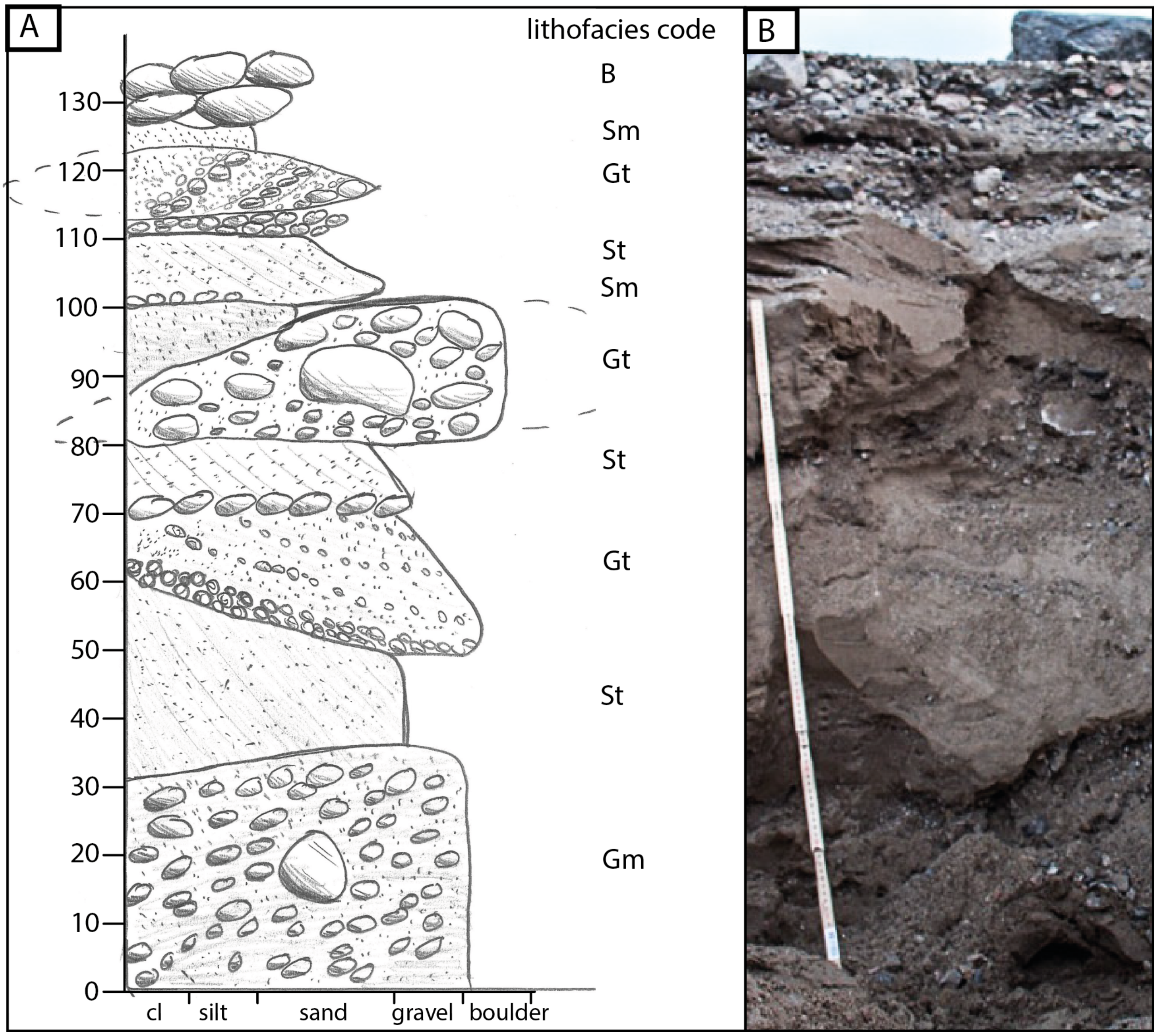


Figure 21. *Log of section 2, see Fig. 11 for location. A) Trough cross stratified gravel and sand. B) Photo of the section. The material got unstable when excavated and some of the units have already collapsed on the photo.*

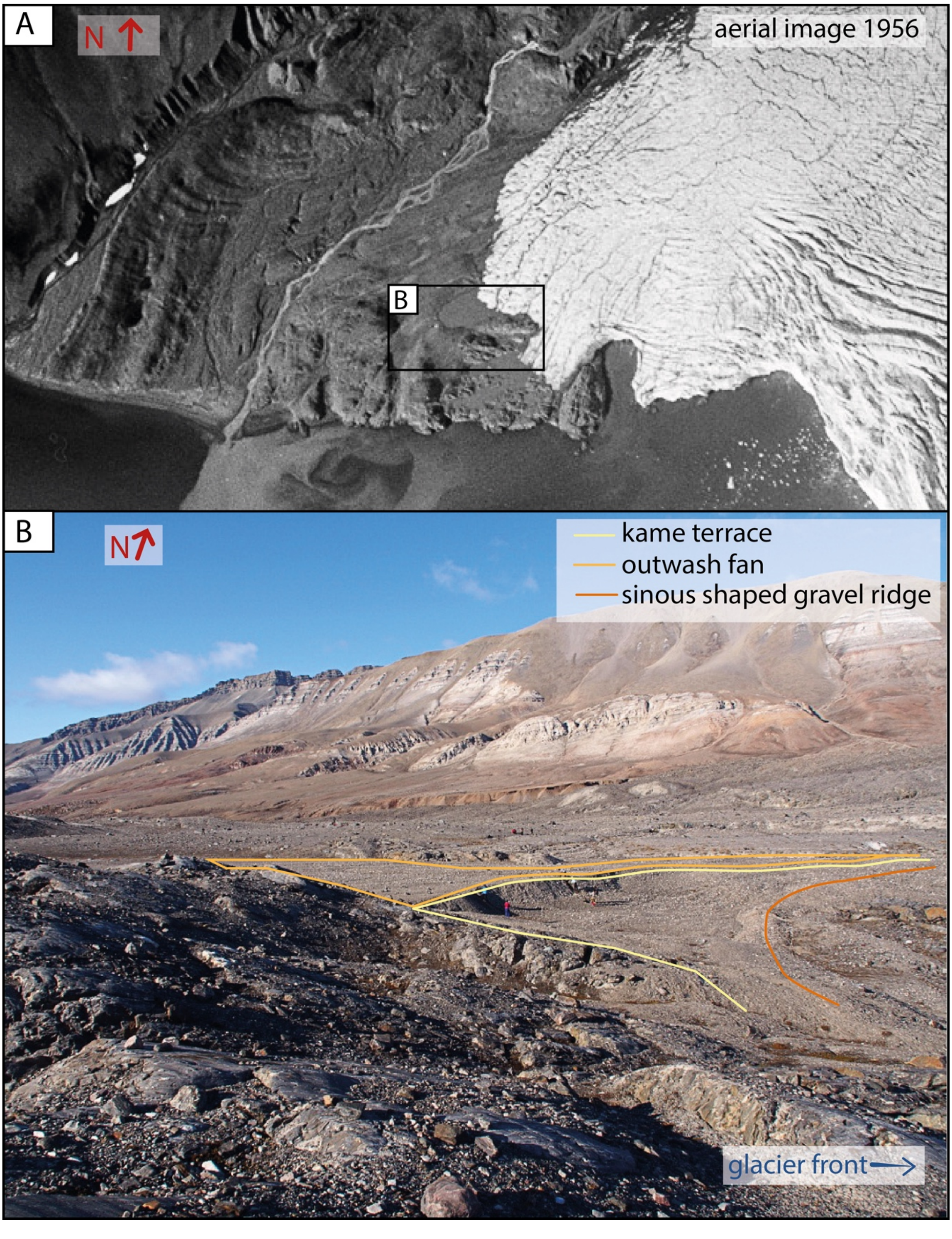


Figure 22. *Kame terraces. A) Aerial image from 1956 of the northern shore. The ice front is located further west than today. An active glaciofluvial channel is cutting through the central part of the forefield and an ice dammed lake is observed at the ice margint. B) Modern photo of the area (marked in the black box on 21A). The ice dammed lake has drained since 1956. Photos: 21A, Norwegian Polar Institute 1956, 21B, Ólafur Ingólfsson, 2014.*

5.3 Supraglacial landforms

5.3.1 Sediment gravity flow tracks

Description

380 pronounced, downslope tracks have been mapped (Fig. 10): 157 on the northern shore (Fig. 11), 133 on the southern shore (Figs 12, 13) and 90 in the subaqueous record (Figs 14, 15). The lengths vary from ~ 16 to ~ 253 m, the terrestrial examples from ~ 16 to ~ 152 m and the subaqueous from ~ 101 to ~ 253 m.

A high concentration of tracks occurs in the north-eastern corner of the terminal moraine on the northern shore, descending from the terminal moraine. On the southern shore the highest concentration occurs in the ice proximal area ~ 500 m south of the shore line. In the subaqueous record, the highest concentration appears in the northern part of the basin outboard the LIA terminal moraine. Most tracks there initiate either at the base of the terminal moraine (Fig. 10, appendix) or at the shoreline and some reach all the way to the fjord bottom.

Many of the terrestrial tracks appear as erosional furrows with levees and they spread and flatten in the downslope direction (Fig. 23a, c, d).

Interpretation

The appearance of the landforms as erosional furrows with levees, the concentration of the tracks in particular areas of the forefield and the downslope appearance lead to the interpretation of the landforms as sediment gravity flow tracks (Wright & Anderson 1982; Schomacker 2007). Sediment flows are essential parts of the degradation processes of areas with high dead-ice content (Etzelmüller 2000; Schomacker & Kjær 2008). They are interpreted to form due to liquefaction of the sediment and to indicate recent slope activity (Lawson 1979; Etzelmüller 2000; Schomacker 2007; Benn & Evans 2010). The most recent sediment flow tracks occur in the proximal part of the forefield. The tracks initiating at the base of the LIA moraine in the subaqueous record are interpreted to be relict features (Plassen *et al.* 2004). They are interpreted to have formed synchronously or subsequent to the formation of the LIA moraine (Plassen *et al.* 2004). The tracks that initiate where the outwash fans meet the fjord (Fig. 10, appendix), are suggested to have been active more recently.

The total number of debris flow tracks in the forefield has not been established and mapped. The outline sometimes is blurred which makes it hard to identify them on the aerial image. The appearance of debris flow tracks is later in this chapter used to classify stages of hummocky moraine development.

Figure 23. *Supraglacial landforms***.** *A) Gravity flow tracks, northern shore. B) Gravity sorting, southern shore. C) Debris flow tracks, southern shore. D) Debris flow tracks, northern shore. E) Landslide back-scarp, southern shore. F) High-relief moraine, southern shore.*

5.3.2 Kettle hole

Description

Two hundred and fifty-seven small depressions with a various but often circular outline are observed spread over the forefield (Fig. 10). 139 of the depressions are dry and small amounts of fine-grained sediment are found deposited in the depressions. The *dry* depressions cover an area of ~ 0.072 km^2 and 42 individual depressions occur on the northern shore (Fig. 11) and 97 on the southern shore (Figs. 12, 13). 119 of the landforms have water present in the bottom of the depression; 34 on the northern shore and 87 on the southern shore. The depressions with water cover an area of ~ 0.101 km^2. Small mounds with positive relief consisting of material similar to that observed in the depressions are also observed in the area, however they are not designated a specific symbol in the legend. The landforms are concentrated in the distal part of the forefield. Their dimensions range from ~ 6 x ~ 6 m to ~ 60 x ~ 20 m (Fig. 24c).

Interpretation

The depressions are interpreted to have formed due to dead-ice degradation (Clayton 1964; Schomacker 2008; Benn & Evans 2010). The irregular surface of the terrain reflects the topography of the dead-ice and water is present in the depressions in the early stage of degradation (Benn & Evans 2010). The fine-grained material is interpreted as waterlain sediments (Balescu & Lamothe 1994). With time, the activity is expected to shift throughout the forefield and some depressions will dry out and fine-grained material will be left (Benn & Evans 2010). The features with positive relief are interpreted to represent inversion landforms (Kjær & Krüger 2001; Benn & Evans 2010), where the previously surrounding ice has melted away and the small amount of lacustrine sediment is left as a little hill in the landscape. All three types of landforms are characteristic for dead-ice landscape (Benn & Evans 2010).

5.3.3 Landslide back-scarp

Description

A total number of 202 scars are observed in the forefield: 54 on the northern shore (Fig. 11), 138 on the southern shore (Figs 12, 13) and 10 in the subaqueous record (Figs 14, 15). They vary in length from ~ 9 to ~ 253 m.

The features appear as scars in the landscape, where an amount of material has slid out and deposited in immediate vicinity (Fig. 20e). The appearance is often connected to slopes where debris flows tracks have occurred. Most scars occur close to other scars.

Interpretation

The scars are interpreted to be landslide back-scarps and to be a part of the ongoing mass moving processes in the dead-ice terrain (Etzelmüller 2000; Ballantyne 2002; Benn & Evans 2010).

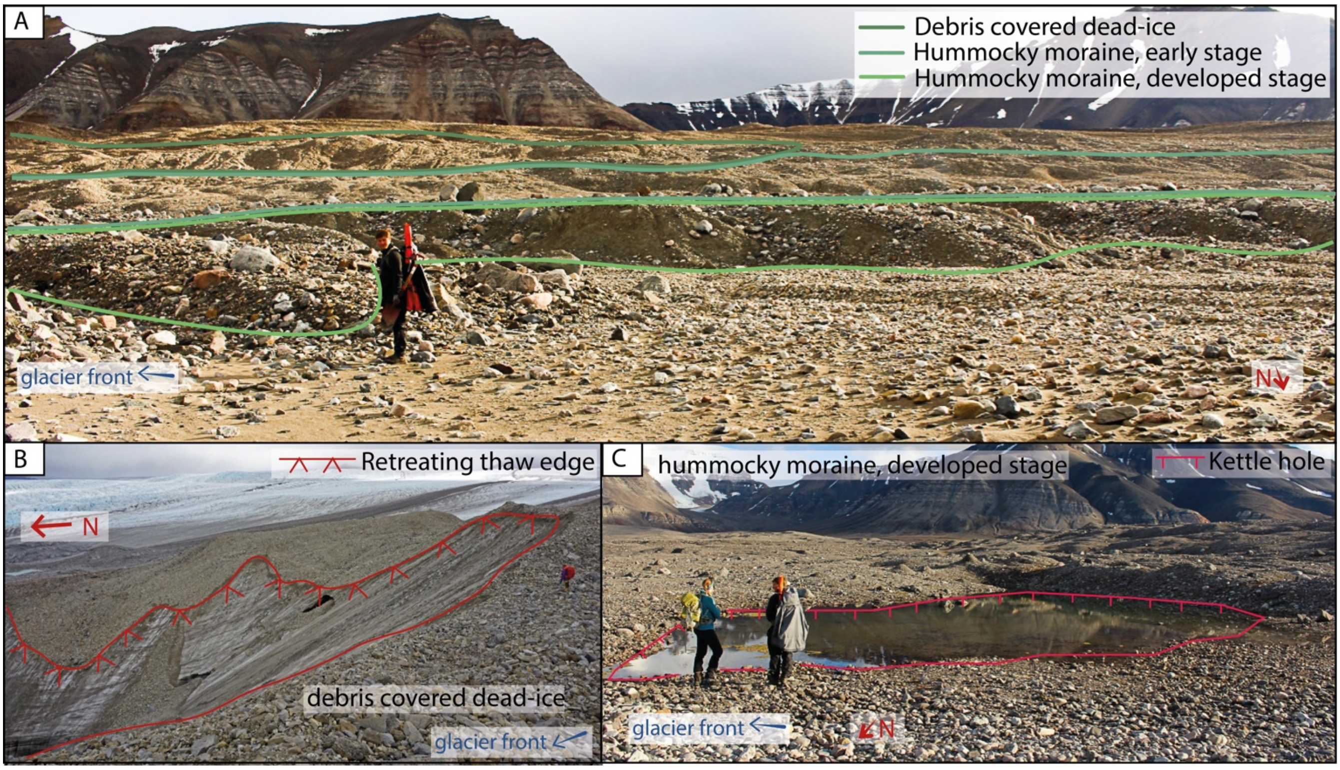


__Figure 24.__ Supraglacial landforms. A) Hummocky moraine in three stages. Furthest up the debris covered dead ice is located, next step is hummocky moraine, early stage and hummocky moraine, developed stage appears in the lower part of the image. B) Retreating thaw edge in the debris covered dead ice. C) Large kettle hole in hummocky moraine, developed stage, southern shore.

5.3.4 Retreating thaw edge

Description

A total number of 43 sharp edges occur in the debris covered dead-ice terrain (Figs 13, 24b). The length varies from 32 to 208 m. The landform has only been observed on the southern shore. It appears in the transition zone between glacier ice and debris covered dead-ice, where meltwater cuts down and cliffs of former debris covered ice are exposed.

Interpretation

The edges are interpreted as *retreating thaw edges*, where active backwasting takes place (Krüger & Kjær 2000; Schomacker & Kjær 2008). Supraglacial meltwater streams have cut down into the debris covered dead-ice and thereby previously buried glacier ice has been exposed. The location of these edges is expected to change rapidly. Therefore, the location of the landforms observed in the field is expected not to match with the location on the map, since the latter is based on aerial imagery from 2009. And the glacier has since then retreated.

5.3.4 Hummocky moraine(s)

The landforms described in this section represent three stages of dead-ice development (Benn & Evans 2010) and the differentiation between the three stages has been made on the degree of degradation.

Description

Debris covered dead-ice

This landform is found only on the southern shore, where it covers an area of ~ 1.5 km^2 (Figs 12, 13, 24a, b). It occurs along the southern most part of the terminal moraine. The area is elevated above the rest of the forefield and the surface is covered by angular to subrounded debris. Steps have formed in the surface parallel to ice flow direction, forming a tribune down towards the sea. A few small kettle holes and mud boils are observed on the surface. In the most ice proximal part of the area, supraglacial streams have cut down and previously buried, clean glacier ice is exposed. The debris layer is ~ 0.5 m thick. Active backwasting takes place (Fig. 24b)

Hummocky moraine, early stage

North of the *debris covered dead-ice* an area mapped as *hummocky moraine, early stage* covers ~ 0.8 km^2. The surface material is coarse. The area is dominated by slumping, debris flows, relict drainage streams and fall sorting. Emptied kettle holes are observed in the western part of the area. The topography is very irregular (Fig. 24a). Faintly pronounced ridges are observed on the aerial image in the western part of the area. The debris cover is assumed to be of similar thickness as that observed in *debris covered dead-ice*.

Hummocky moraine, later stage

This landform is found on both shores and it covers an area of ~ 0.66 km^2 in total, ~ 0.35 km^2 on the northern (Fig. 11) and ~ 0.3 km^2 on the southern shore (Figs 12, 13, 21a, c). Low relief mounds and kettle holes are spread over the area. The area is cut and eroded by drainage channels and some slumping activity occurs on the surface.

A pit was dug in the area very close to the terminal moraine on the northern shore (Figs 11, 25). There two units were observed. The lower one composing a diamict given the lithofacies code DmM(m_2)3 and the upper one composing a different diamict given the lithofacies code DmC(c)2, according to Krüger and Kjær (1999). Fissility was observed in the lower unit and the boundary to the upper unit is conform, undulating with a sand lense in the contact.

Interpretation

The debris covered dead-ice is the most premature stage of the group of landforms and the hummocky moraine found furthest away from the glacier front is the most mature stage (Hambrey *et al.* 1997; Krüger & Kjær 2000; Kjær & Krüger 2001; Schomacker & Kjær 2008; Benn & Evans 2010). This differentiation of the three stages is based on the amount of slumping activity seen on the surface. The amount of ice is assumed to be highest in stage one and the least in stage three. No quantitative measurements have been made, but the map can work as background for estimation of dead-ice degradation in the future. The large amount of debris covered dead-ice is suggested to mainly originate from the tributary glacier, Gerritbreen (de Geer 1910; Tyrrell 1922a). The pattern in the surface morphology represent the old pattern that was created, when Gerritbreen suddenly advanced and pushed the medial moraines of Nordenskiöldbreen. The sinuous form left by the push (de Geer 1910; Tyrrell 1922a) can still be traced in the surface morphology.

The units in the section are on the background of the observed fissility in the lower unit, the conform boundary and presence of sorted material in the contact zone interpreted to represent subglacial till and melt-out till, respectively. The section illustrates how the landscape has developed with time and the previously supra- and englacial material now rests directly upon the subglacial material. The melt-out till dominates the surface material in *hummocky moraine, later stage*.

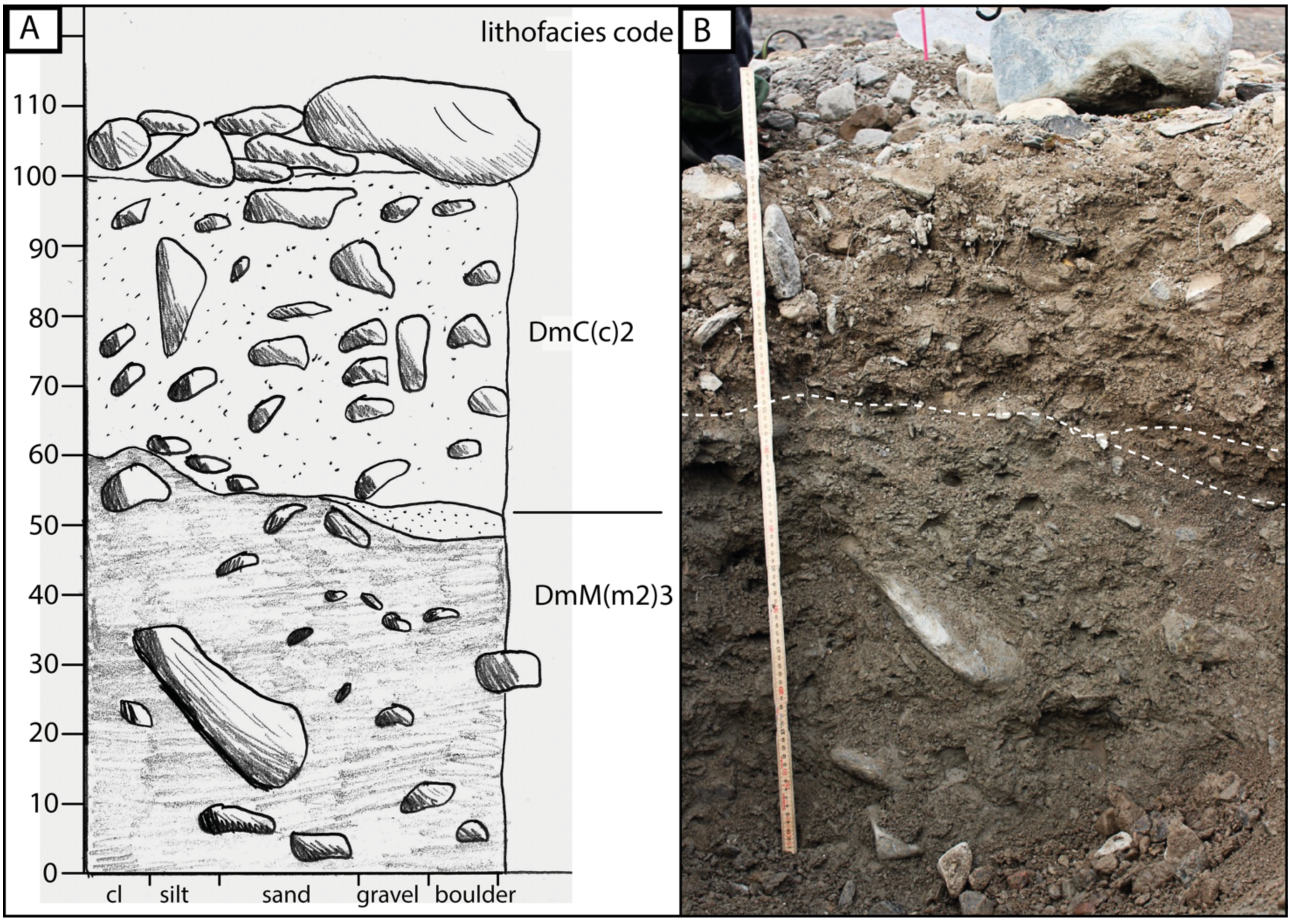


Figure 25. *Log of section 3, for location see Fig. 11**.** A) The lower unit comprises a subglacial till and the upper unit comprises a melt-out till, characteristic for the surface material of hummocky moraine, developed stage. The boundary between the two units is conform and a sand lense is observed in the contact. B) Photo of the section, the sand lense is surrounded by a dashed line.*

5.3.5 Medial moraines

Description

An area of ~ 0.56 km^2 is mapped with this label: ~ 0.02 km^2 occur in the northern part of the glacier and ~ 0.53 km^2 occur in the southern part of the glacier (Fig. 10).

Debris bands parallel to glacier flow are observed on the glacier surface. The debris covered part is elevated above the surrounding glacier ice (Fig. 26). The debris bands in the southern part of the glacier can be traced back to the nunataks, Ferrierfjellet and Terrierfjellet, located ~ 6.5 km NE of the glacier front. The debris bands in the northern part are short and narrow, and they taper off after less than 1 km.

Interpretation

The bands of debris are interpreted to be medial moraines, with material mainly originating from the nunataks (Tyrrell 1922b; Hambrey & Glasser 2003). They form when the glacier ice flows by the nunataks and carries off material (Fig. 2). The debris septa prevents the ice underneath from melting, and the medial moraines appear therefore elevated above the surrounding glacier ice (Hambrey & Glasser 2003; Benn & Evans 2010). The short medial moraines in the northern part of the glacier are interpreted to originate from material being scraped off of a subglacial bedrock obstacle.

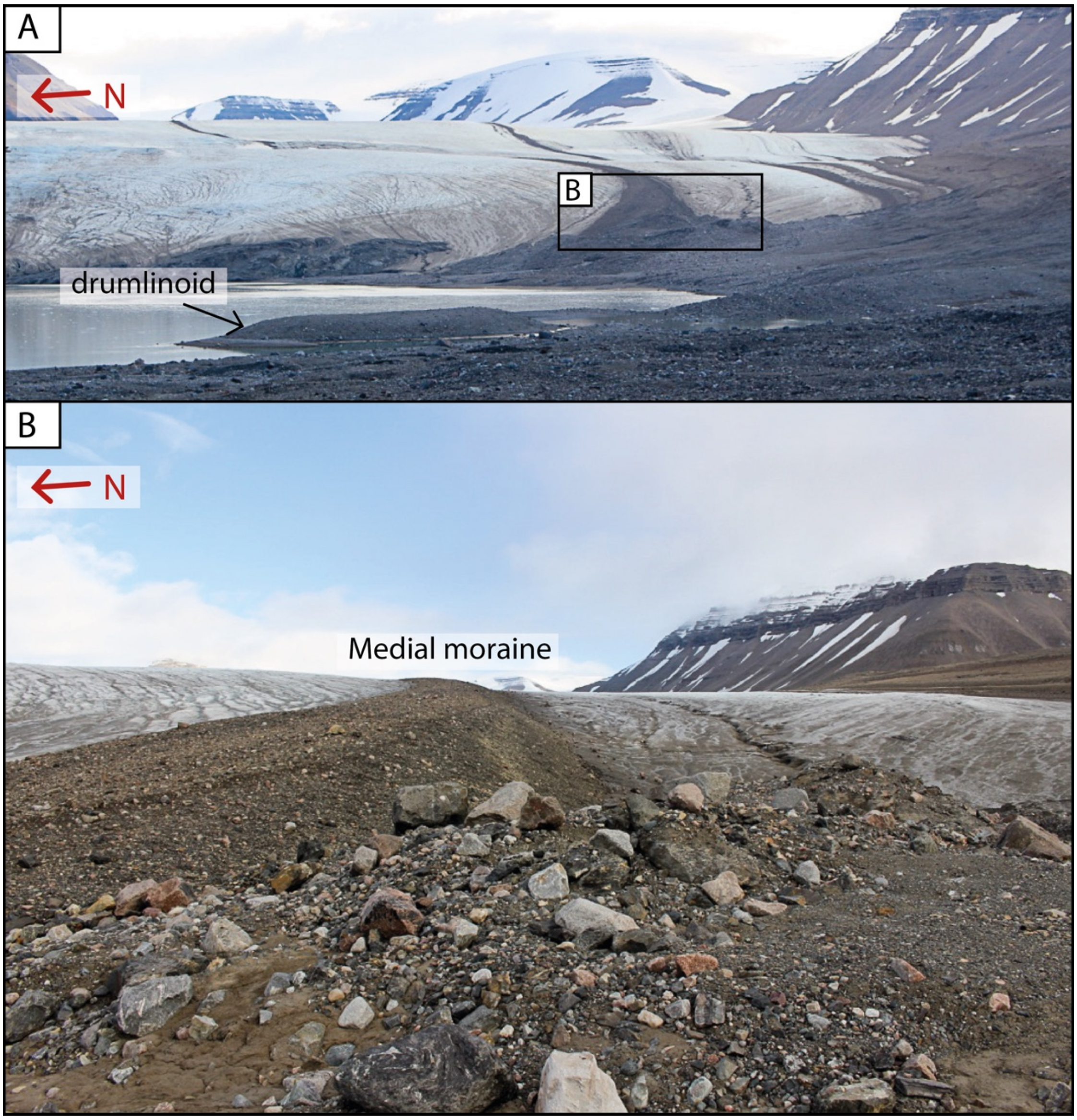


Figure 26. *Medial moraine. A) The southern shore, viewed from the terminal moraine. Drumlinoid no. 19 is seen central in the photo. The medial moraines are seen on the glacier surface. B) Close up of the medial moraine, originating from Ferrierfjellet. The surface of the medial moraine is elevated above the surrounding glacier ice.*

5.3.6 Erratics

Description

Large boulders are randomly spread on the entire forefield (Figs 10, 27). A number of erratics are also found outboard of the terminal moraine on the southern shore (Figs 10, 27b). They vary in size and lithology. Some of the boulders are of similar lithology as the basement exposed in the forefield, but exotic lithologies, that are not found in near proximity to the glacier, also occur (Fig. 27c). Most of the boulders are subrounded to rounded.

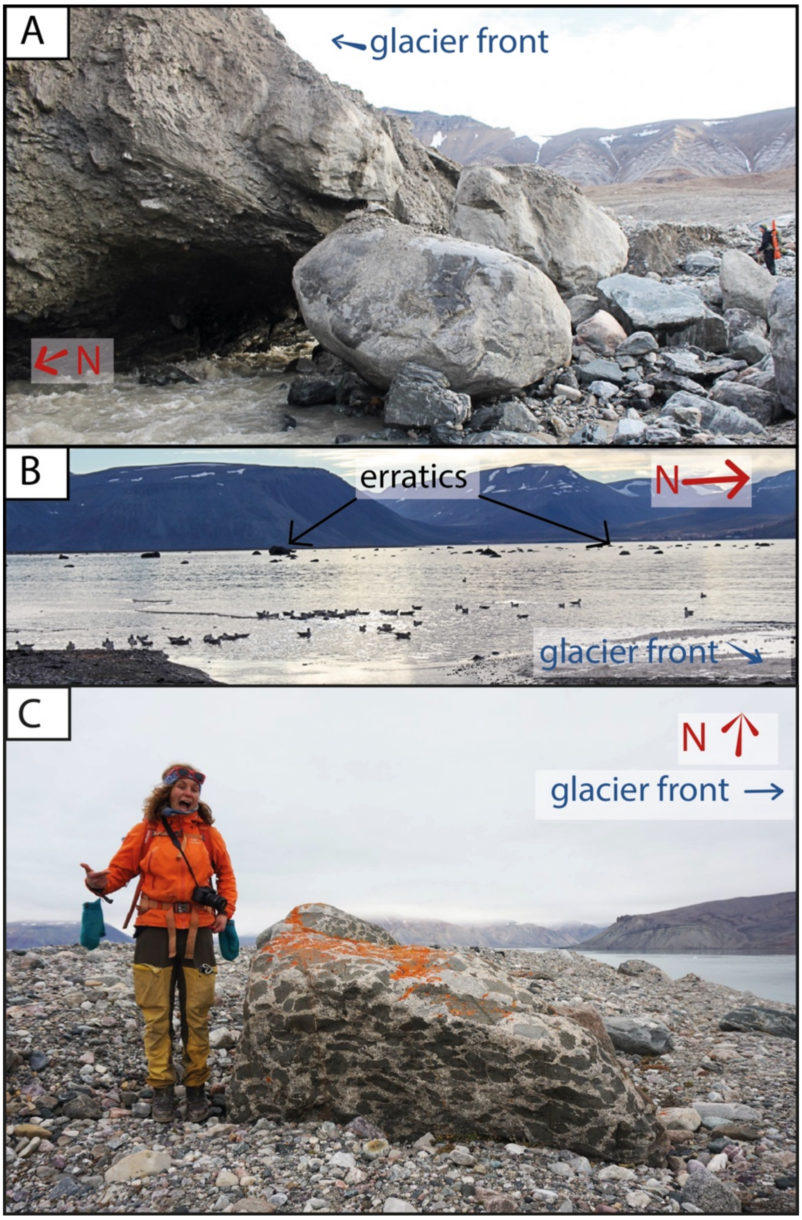


Figure 27. *Erratics in the forefield. A) Erratics in the proximal part of the southern shore forefield. B) Erratics outboard of the terminal moraine, southern shore. C) The origin of the erratic of interesting lithology is unknown, however, it has been suggested to be either a tectonic breccia or a lithified gravity flow deposit (W.K. Dallmann 2016, personal communication). The roundness of the erratic indicates glacial transport. Photo 21C: Nina Friis, 2015.*

Interpretation

The boulders are interpreted to be erratics and to originate from supra- or englacial transport. The supgraglacial transport is documented in old photos (Fig. 4), where large erratics resting on the glacier surface were observed. Erratics are also described in old literature (Cole 1911). The erratics found outboard the terminal moraine are interpreted to have been deposited by an advance prior to the LIA.

5.4 Proglacial landforms

5.4.1 Distal fjord bottom

Description

About ~ 2.5 km^2 are mapped with this label. The area is situated in the distal part of the basin, at water depths of ~ 160 m.b.s.l. (Fig. 15). It is characterised by low acoustic response (Fig. 5).

Interpretation

Due to the low acoustic response seen in the bathymetrical data set, the area is interpreted to be covered by finegrained material that has settled from suspension. This in accordance with the interpretation by Plassen *et al.* (2004).

5.4.2 Ice berg plough marks

Description

A total number of 275 of furrows in the seafloor have been mapped. They occur in the SW corner of the map (Figs 10, 15). The furrows vary in length from ~ 5 to ~ 178 m, the width is up to ~ 4 m and they are up to ~ 2 m deep. They are observed as small scars in the seafloor at shallow water depths and have multiple directions.

Interpretation

The features are interpreted as ice berg plough marks, made by ice bergs originating from Nordenskiöldbreen (Baeten *et al.* 2010). Due to the shallow water depth it can be argued that some of them also originate from sea ice. The scour marks can be related to both advance and retreat of Nordenskiöldbreen, in either case an increase in calving rate is expected (Dowdeswell & Forsberg 1992; Warren 1992; Baeten *et al.* 2010).

5.5 Glaciofluvial landforms

5.5.1 Outwash fans

Description

This landform covers area of ~ 0.9 km^2 in total: ~ 0.3 km^2 on the northern shore (Fig. 11) and ~ 0.6 km^2 on the southern shore (Figs 12, 13). Outboard of the terminal moraines on both northern and southern shore large fan-shaped deposits occur (Fig. 28). Smaller fans at different locations are observed in the forefield (Fig. 10)

The surfaces of the fans slope gently towards the sea. The biggest fan is located on the southern shore, immediately west of the terminal moraine (Figs 12, 28b). The surface material comprises sand, gravel and small pockets of fine-grained material. The material is sorted and longitudinal bars appear. Active meltwater streams occur in some parts of the surfaces, however, in other parts dry channels occur.

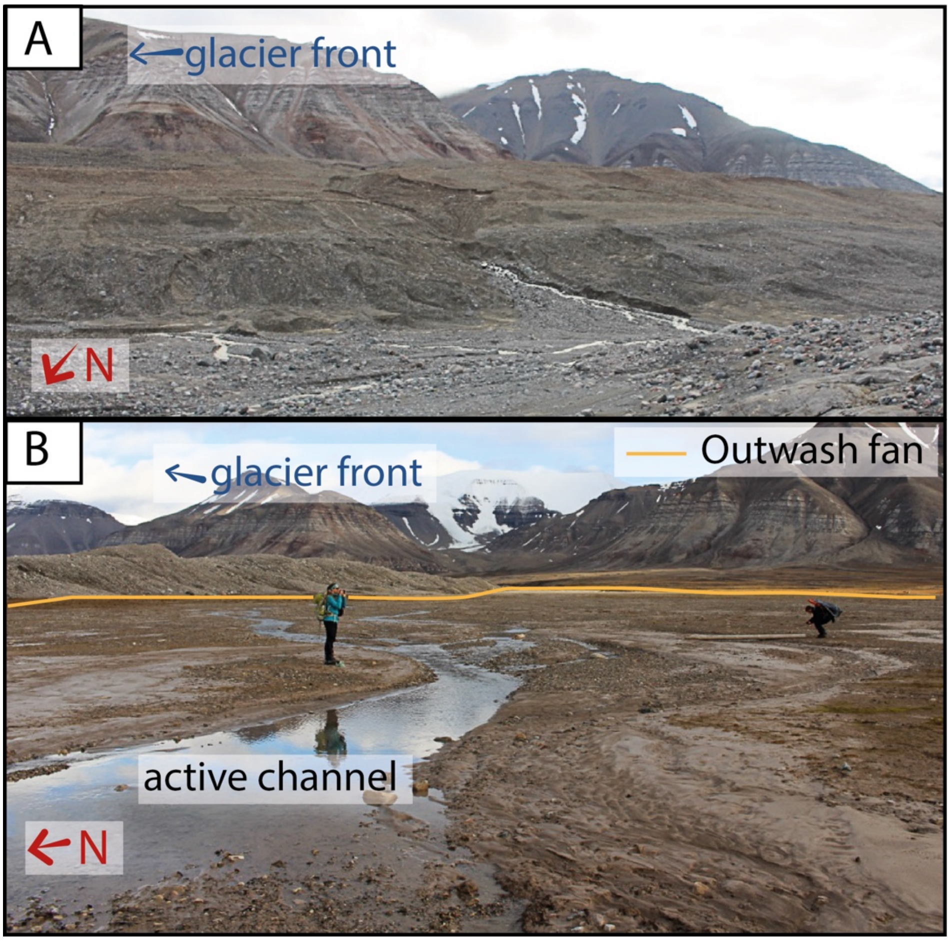


Figure 28. *Outwash fans. A) Meltwater channel cutting through the hummocky terrain, southern shore. B) Outwash fan with braided river system outboard of the terminal moraine on the southern shore. The terminal moraine is seen in the upper left corner of the photo.*

A natural section occurs immediately outboard of the terminal moraine on the southern shore (Fig. 29). The section is located adjacent to a clear edge in the terrain. A log from there reveals two units (Fig. 29). The lower one, ~ 180 cm thick, consists of sorted gravelly-sand and sandy-gravel. Clasts up to 20 cm in diameter and small pockets of sand occur. The clasts appear slightly imbricated. Shell fragments are found in the lower part of the section. The boundary to the upper unit is conform and the upper unit comprises interbedded layers of coarse and fine sands. At the surface, a ~ 5 cm thin soil horizon with roots was found.

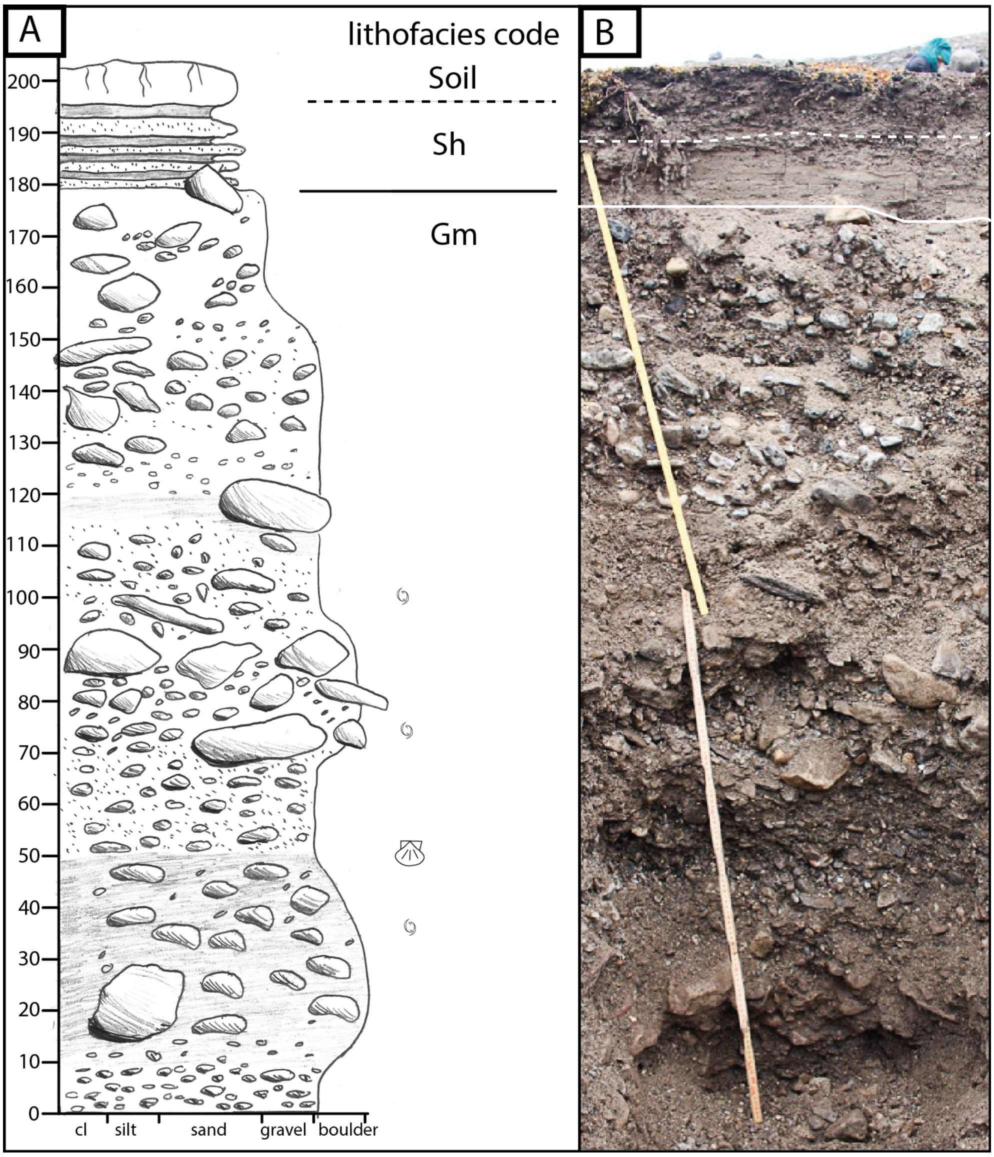


Figure 29. Log of section 4, for location see Fig. 12. The section is located where the outwash fan outboard the terminal moraine meets the terminal moraine on the southern shore. A)Two units are observed here; the lower consisting of gravelly sand with a faint stratification. Shellfragments and a single whole shell occur. The upper unit comprises interbedded layers of coarse and fine sands. B) Photo of the section. The thin soil-horizon is seen at the top.

Interpretation

The sorted, coarse-grained material and fan-shaped deposits are interpreted as outwash deposits (Miall 1977; Anderson 1989; Maizels 1993). The low-angled fans can further be interpreted as glaciofluvial outwash fans (Kjær *et al.* 2008; Brynjólfsson *et al.* 2014; Rubensdotter 2015). The active meltwater channels seem to migrate according to meltwater supply, and thereby the activity of the fans is defined. The most active outwash fans are found outboard of the terminal moraine and in fornt of the ice-margin.

The sediment in the section is interpreted to be glaciofluvial deposits and the top of the pit is interpreted to represent a former base level for the drainage.

5.5.2 Active and relict meltwater channels

Description

Channels are observed in the terrestrial part of the forefield (Figs 11 – 13, 28). Their general orientation is transverse to ice flow. Several of the channels initiate in single apices and spread out in a braided pattern downslope, following the topography. Running water is observed in the channels outboard of the terminal moraine and also in the most proximal channels. During a heavy rain-event all the stream channels on the big fan on the southern shore were observed to be filled with water. Single channels that occur in the hummocky moraines are either dry or have a very little amount of water in them.

Interpretation

The channels are interpreted as meltwater channels (Ashworth & Ferguson 1986; Rubensdotter 2012; Schomacker *et al.* 2012). The channel activity migrates according to meltwater outlet from the glacier and the channels mimic the glacier front. The sets of meltwater channels have followed the ice retreat, and the dry channels are interpreted as relict meltwater channels or seasonal channels, active only during spring melt.

5.5.3 Sinuous shaped gravel ridges

Description

Several sinuous shaped ridges occur on the southern shore (Figs 13, 30d, e) and a single very small occurs on the northern shore (Figs 11, 30a, b, c). The ridges cover an area of ~0.06 km^2 and are up to

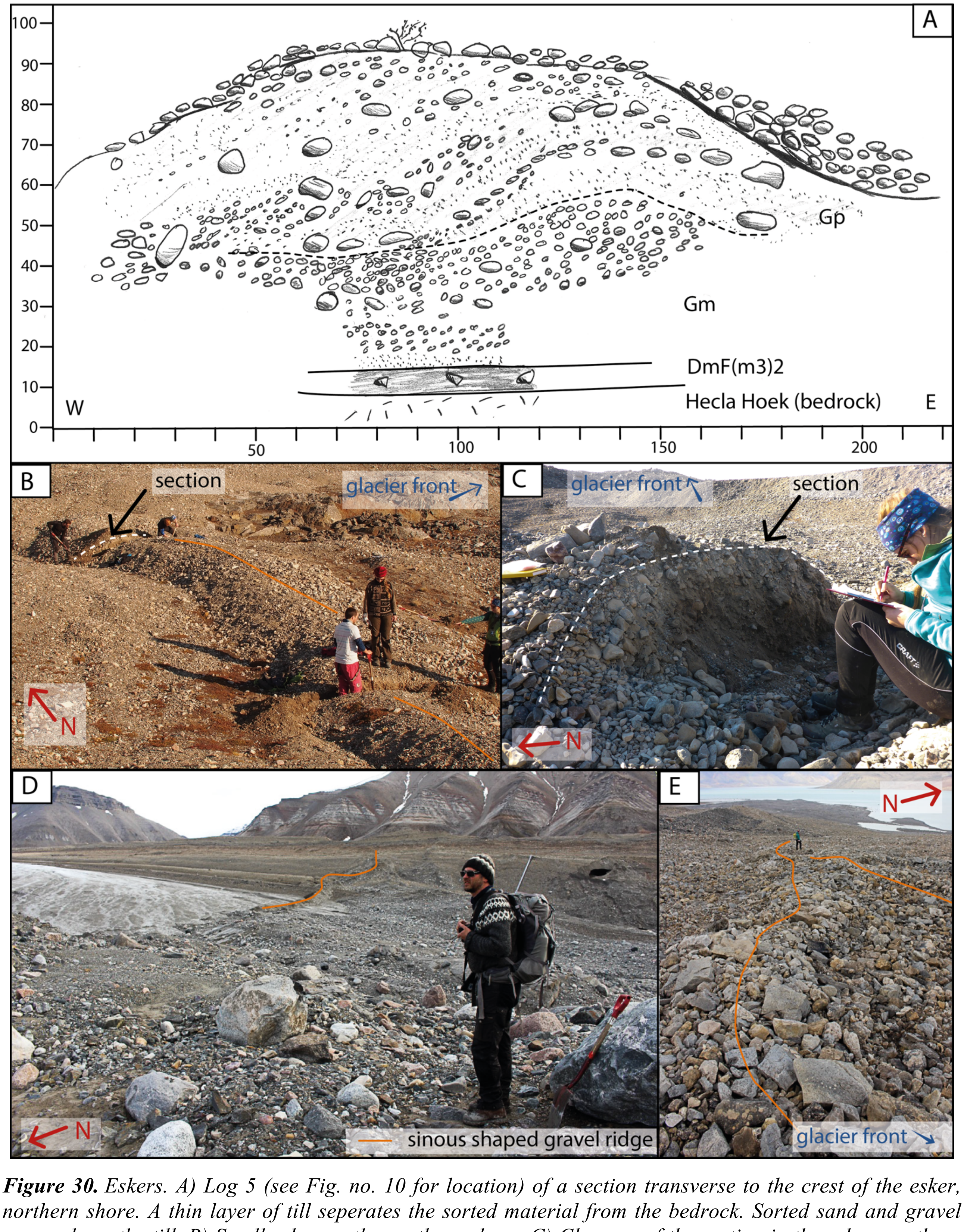


Figure 30. *Eskers. A) Log 5 (see Fig. no. 10 for location) of a section transverse to the crest of the esker, northern shore. A thin layer of till seperates the sorted material from the bedrock. Sorted sand and gravel occur above the till. B) Small esker on the northern shore. C) Close up of the section in the esker, northern shore, D) Supraglacial esker on the southern shore, crest is marked with orange colour. E) Supraglacial esker on the southern shore, crest is marked with orange. Photos: 21A, Ólafur Ingólfsson, 2014, 21C, Hannah Marshall, 2014.*

1.5 m high and ~ 2 m wide. They occur in the south-eastern corner of the forefield where they overprint the debris covered dead-ice, and the one on the northern shore is situated ~ 100 m south of drumlinoid no. seven. The orientation of the landform on the northern shore matches the orientation of the nearby meltwater channels. The ridges on the southern shore continue for up to ~ 500 m. A topographical divide (Fig. 30e) occurs at the highest point, there the sinuous ridges follow the descending slopes respectively towards north and south. The surface material consists of subrounded to angular blocks of varied lithologies.

A section was cleared in the ridge on the northern shore (Fig. 30a, b, c), transverse to the crest of the ridge. There, the landform was observed to consists of sorted to well-sorted sediments of sandy gravel and gravelly sand. A faint cross bedding was observed. Below the sorted material in the section, a thin layer of diamict occurred and separated the landform from the bedrock.

Interpretation

The ridges are interpreted as eskers based on their sinuous shape, the internal composition of the ridge on the northern shore and the surface material (Banerjee 1975). The northern example is interpreted as an englacial esker (Banerjee 1975), as it can not be traced on the surface of the glacier ice on the aerial image from 1956 (Fig. 22). The landforms on the southern can suggestively have a supraglacial origin (Banerjee 1975), as present day parallels can be found in the supraglacial channels in the dead-ice terrain (Figs 13, 20c).

The reason for mapping the two types of ridges in the same landform category is the similarity in shape and sedimentology. Their origin is interpreted to be a combination of deposition by running water and slope processes in ice-marginal/supraglacial meltwater channels. Some material has been deposited by the meltwater and the rest by slope processes where surface material has slid down and ended above the glaciofluvial material. The present day supraglacial meltwater channels (Figs 13, 24b) are suggested as analogues to explain this. Similar landforms have been described at the forefield of Holmströmbreen, Ekmanfjorden, Svalbard and are characteristic for dead-ice terrain with a high amount of debris cover (Boulton *et al.* 1999; Schomacker & Kjær 2008).

5.6 Periglacial landforms

5.6.1 Weathered bedrock

Description

Outboard of the lateral moraine on the northern shore dark bedrock with a strongly weathered surface makes up ~ 0.05 km^2(Figs 10, 31a) The full area of this landscape is not included in the map. No glacial lineations were observed.

Interpretation

The strongly weathered surface and the location outboard of the terminal moraine indicate that this part of the bedrock has not been covered by the LIA advance of Nordenskiöldbreen. The bedrock is interpreted to have been exposed to physical and chemical weathering processes for longer time than the bedrock found inboard of the terminal moraine (Porter 1975; Ballantyne 2002).

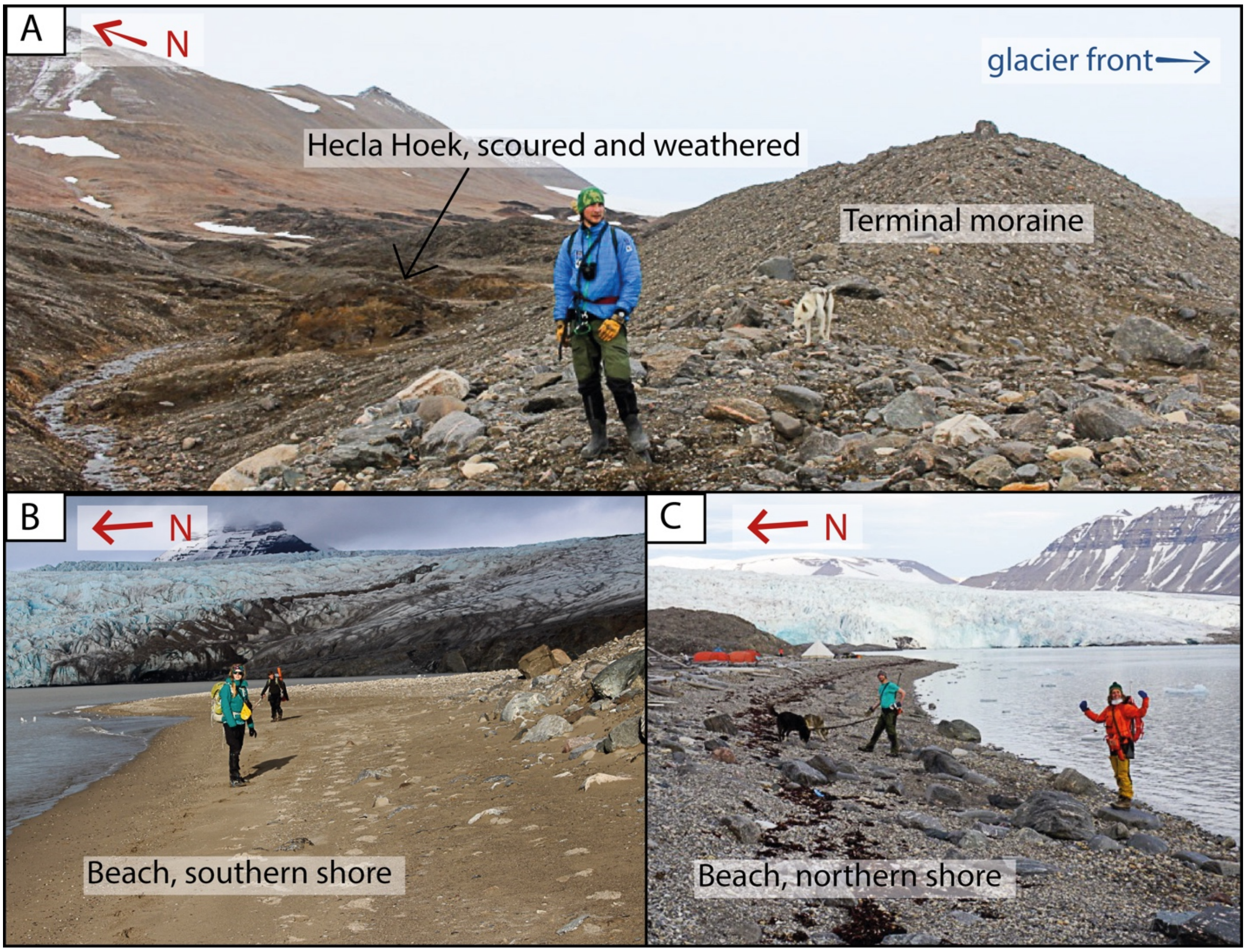


Figure 31. *Periglacial and coastal landforms. A) The scoured and weathered bedrock is observed outboard of the terminal moraine on the northern shore. B) Beach on the southern shore, close to the glacier front. Medial moraine material reaches the sea at the glacier front. C) Beach on the northern shore with coarser material. Photo: 28C, Nina Friis, 2014.*

5.7 Coastal landforms

5.7.1 Beaches, modern and Holocene

Description

A total area of ~ 0.010 km^2 are mapped as this landform, ~ 0.028 km^2 on the northern shore (Fig. 11), ~ 0.071 km^2 on the southern shore (Figs 12, 13). In addition, ~ 0.012 km^2 of similar material occurs ~ 800 m north of the present shoreline located ~ 34 m.a.s.l. (Fig. 11).

The landform appears as a continuous band of finegrained material along the shoreline (Fig. 31a, b). The width of the band varies from ~ 1 to ~50 m on the northern shore and from ~ 7 to ~60 m on the southern shore. The band slopes towards the sea with less than 1° and the material is sorted by wave activity. The bands are widest at the point where the outwash fans meet the sea. The largest difference between the northern and the southern shore is that the northern coastline is dominated by coastal bedrock cliffs. The northern shoreline is half the length of the southern shoreline. Close to the glacier front on the southern shore, cliffs of bedrock extend down to the sea. Following the coast towards west the beach is narrow the first kilometre, where waves have eroded into the glacial deposits. From the spit the band with fine material increases in width and the relief of the adjacent terrain gets lower.

The small isolated area of similar material found further inland on the northern shore exhibits vertical sections. There, bimodal sediments with shell fragments incorporated occur.

Interpretation

Due to the coastal location, appearance of wave washed material and occurrence of shell fragments, the bands of fine-grained material are interpreted to be beaches, dominated of reworked glacial deposits (Gilbert 1983; Hart & Plint 1989; Powell & Molnia 1989; Alexanderson & Murray 2012). The shape of the spit on the southern shore indicates that material is transported along shore in a direction away from the glacier (Alexanderson & Murray 2012). The small area found ~800 m inland on the northern shore is interpreted to be beach deposits of older age, suggestively early Holocene (Klysz *et al.* 1989).

5.8 Retreat history

Twenty-eight ice front positions have been reconstructed from different data sources (Figs 7, 32, Table 2). The sources are historical maps, historical aerial images, modern orthorectified aerial images and satellite images.

Most terrestrial Svalbard glaciers have debris covered snouts, therefore the exact ice front positions are difficult to determine (Etzelmüller 2000; Rachlewicz *et al.* 2007). Clean ice border, as defined by Rachlewicz *et al.* (2007), has been applied in this study. The border between active debris-covered ice and debris-covered dead-ice on the southern shore of Adolfbukta was difficult to determine, which leads to uncertainties in this area.

From 1896 until 2015 Nordenskiöldbreen retreated ~ 2357 m in the northern part of the basin and ~ 3551 m in the southern part of the basin. The spacing between the retreat lines is closest in two periods: 1) from 1960 to 1965 and 2) from 2002 to 2015, as more data have been obtained from those periods. The lines in northern basin are closer spaced compared to the southern basin. There is a notable difference in retreat between the terrestrial and marine part. The ice positions that are reconstructed from data with evidence from both archives all are horse-shoe shaped, in accordance with the observation by Slater (1925).

Retrettøya has been exposed in the centre of the bay since 1961. Thereafter, the shape of the lines can be described as double horse shoes, with the hinges resting on both shores and on Retrettøya (Fig. 32). The glacier retreated ~ 314 m in the northern part of the basin and ~ 754 m in the southern part between 1990 and 2002.

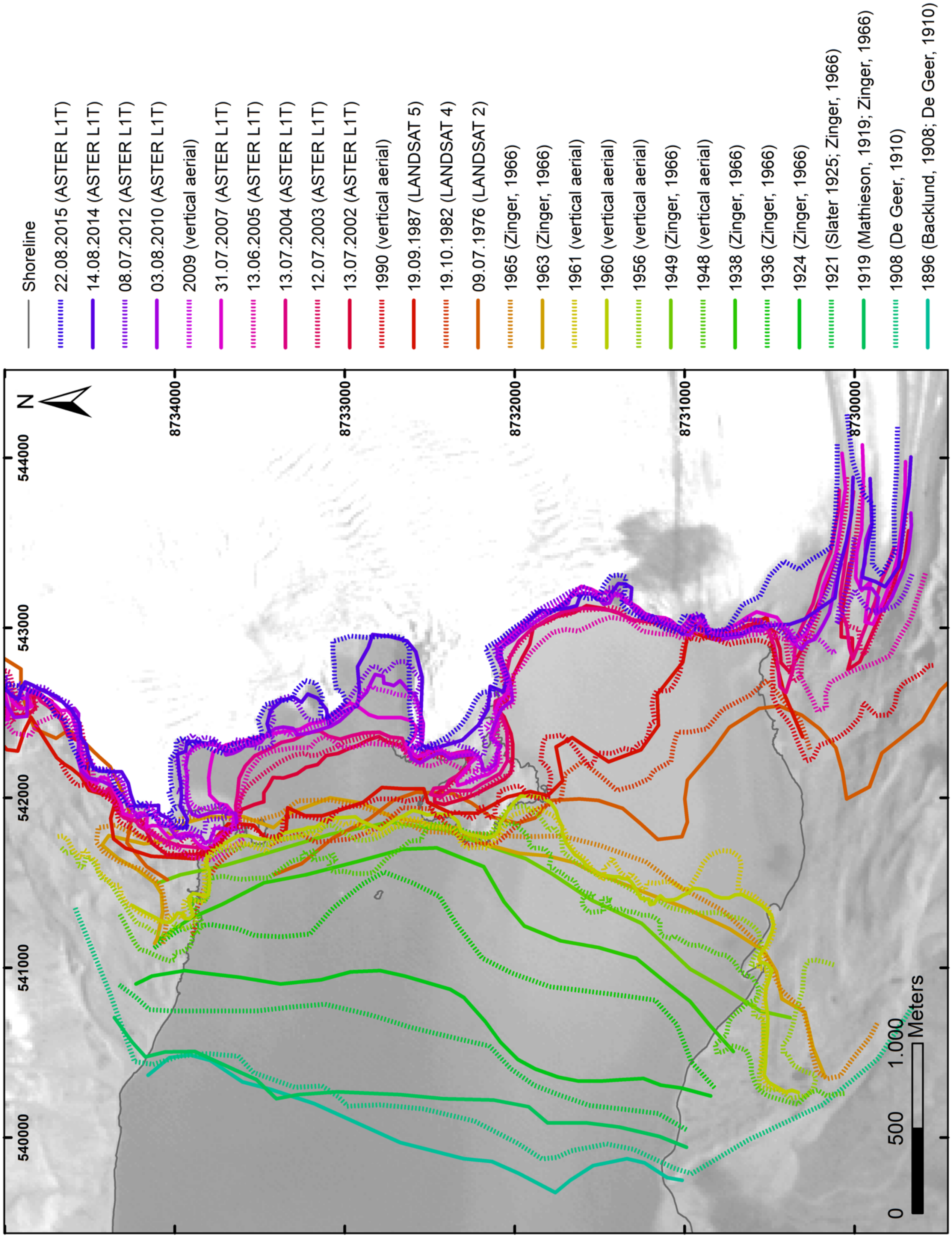


Figure 32. *Twenty-eight reconstructed icefront positions for Nordenskiöldbreen from 1896 until 2015.*

6. Discussion

Nordenskiöldbreen has produced a wide spectre of glacial "fingerprints". A focus in this section will be on the landforms. Many landforms are argued to show *equifinality* (Chisholm 1967; Haines-Young & Petch 1983; Goudie 2004). *Equifinality* means that morphology alone cannot explain the origin of a landform (Goudie 2004). Therefore, the following features mapped in Nordenskiöldbreen forefield will be discussed more thoroughly: *subglacial till plain, glacigenic material, sculptured, streamlined landforms as flutes and drumlinoids, terminal moraine and transverse ridges.* The preservation potential of the landforms in their respective archives and retreat history of the glacier is discussed in relation to this.

The next focus will be on how the presence of certain landforms reveals information about thermal regime and dynamics of Nordenskiöldbreen. *Presence* and *absence* of landforms is critical for the interpretation of the glacier forefield, and therefore the preservation potential in the marine and terrestrial realm is discussed. Also, the implications for glacier dynamics and basal thermal regime will be discussed, as this is of greater importance for the scope of this study than the exact genesis of the individual groups of landforms. Finally, the landsystem model approach and how it applies to Nordenskiöldbreen will be discussed.

6.1 Equifinality of landforms

6.1.1 Subglacial till plain

Presence of shell-fragments in the terrestrial substratum, classified as *subglacial till* (Friis 2015), indicates at least partially marine origin of the material. The subglacial till plain hosts, as described, flutes and drumlinoids. Similar landforms appear in the marine record in the area labelled as: *glacigenic material, sculptured*. Therefore, the latter is suggested to represent a continuum of the *subglacial till plain* found on land. The material is not verified as subglacial till in the subaqueous part (with sediment cores), which prohibits this final interpretation. Plassen *et al.* (2004) argues that the pattern observed inboard the LIA terminal moraine on an early bathymetrical data set from Adolfbukta, often is interpreted as till.

The thickness of the till deposits in both the terrestrial and marine part may vary throughout the till plain (Kjær *et al.* 2003). Nordenskiöldbreen is still actively calving and transferring sediments to Adolfbukta (Szczuciński *et al.* 2009). From the acoustic profile (Fig. 18b), the subaqueous record is observed to be draped by a thin cover of glaciomarine material. It is interpreted to be fine-grained material, that has settled from suspension since Nordenskiöldbreen started to retreat from the LIA moraine (Baeten *et al.* 2010). Although the fine-grained material has not been a focus in this thesis, it indicates that the surface expression of the landforms in the fjord environments will decrease with time (Elverhøi *et al.* 1983; Dowdeswell & Dowdeswell 1989). The time window for a snap shot of well-pronounced surface expressions of the glacially shaped landforms on the seafloor is limited by high sedimentation rates (Dowdeswell & Dowdeswell 1989; Szczuciński *et al.* 2009).

6.1.2 Drumlinoids

The research on drumlinoids has been dictated by the search for a unifying theory of formation process. Subglacial processes are problematic to survey due to the inaccessibility of the subglacial environment (Hart & Rose 2001; Rose & Hart 2008; Benn & Evans 2010). Formation theories of drumlinoids have therefore been based on the landform appearance and internal composition of drumlinoids found in deglaciated areas (Hart 1997; Johnson *et al.* 2010). In addition surveys beneath a West Antarctic ice stream have identified drumlinoids at the bed (King *et al.* 2007; King *et al.* 2009). Hart (1997) narrowed the theory of drumlinoid genesis down to a continuum of three processes: 1) depositional, 2) deformational and 3) erosional, all associated with net erosion of the subglacial bed. However, Dowling (2016) argues that drumlinoids often show equifinality – they can form by a set of processes.

Dowling (2016) focuses on *why, how and where* drumlinoids form. A very important outcome from his work is that drumlinoids and streamlined bedforms in general can be considered to be results of a positive feedback cycle, whereby glacier flow is enhanced (Dowling 2016). Local to regional factors are further argued to be of great importance for drumlinoid formation and a principal physical theory is not expected to be able to catch this variety. The only known active drumlinoid field in the world is situated at Múlajökull, Iceland (Johnson *et al.* 2010). There, local factors are considered to be very important for drumlinoid formation. Dowling argues that the Múlajökull forefield can not be used as an analogue for past ice sheet margins. The flowset differs from an ice-marginal flowset, as Múlajökull is a surging glacier. However, it can be argued that both environments are ice-margin environments. It is also possible that the former ice-sheet margins were much more dynamic than previously thought, and that some ice-margins had surging lobes. This study has shown that Nordenskiöldbreen is a highly dynamic glacier, and it is hard to imagine that larger ice masses behaved in a less dynamic way. It is all a matter of scale.

Dowling (2016, pp. 19) writes that *"a deforming till layer"* in many cases can be assigned to be the key process for drumlinoid formation, *but not the only one,* and that the subglacial system can be described as a chaotic system, which is hard to parameterise.

Friis (2015) investigated the formation of the drumlinoids in the Nordenskiöldbreen forefield. The drumlinoids were shown to be composed of pre-existing sediments and draped with a thin cover of subglacial traction till – i.e. *a deforming till layer*. The traction till is interpreted to have formed under high hydraulic pressure, which has been connected to the presence of a frozen glacier snout (Moran 1980) and further linked to drumlinoid formation (Patterson & Hooke 1995). This is assumed to be a plausible explanation for the formation of drumlinoids in the terrestrial part of the Nordenskiöldbreen forefield, but fails to explain the presence of drumlinoids in the subaqueous record. Due to the marine terminus, the front is unlikely to have been frozen there, and the snout must have been buoyant (Moran 1980; Benn *et al.* 2007; Friis 2015). Relatively high subglacial water pressures is expected at grounded glacier margins in fjords but this is expected to be much lower than the one caused by a frozen snout on land (Benn & Evans 2010; Dowdeswell & Fugelli 2012).

If a glacier advances over an undulating bedrock surface, over-deepenings can be created (Alley *et al.* 2003). There, pore water will be forced to stream uphill due to the impermeable bedrock which will lead to an increase in the subglacial hydraulic pressure (Cook *et al.* 2006). Over-deepenings in the bedrock surface upglacier of the terminus has been documented from Múlajökull, Iceland and suggested to be an important factor for drumlinoid formation (Johnson *et al.* 2010; Jónsson *et al.* 2014; Ingólfsson *et al.* 2016). A GPR survey of Nordenskiöldbreen and Lomonosovfonna revealed bedrock heights below sea level, ~ 2 and ~6 km upglacier from the ice margin (Petterson 2010, unpublished,van Pelt *et al.* 2013; Fig. 6). Variations in the bedrock topography are also seen in the newly exposed forefield, which shows that over-deepenings previously existed in near proximity to the glacier front and more over-deepenings are likely to exist further upglacier.

A take-home message from the study of Nordenskiöldbreen forefield is that the drumlinoids there formed in proximity to the ice front, and the formation is assumed to have been influenced by the presence of a deforming till layer and over-deepenings created by bedrock topography, they are formed during advance and indicate fast ice flow. Maybe more drumlinoids will show up, if/when Nordenskiöldbreen retreats further.

6.1.3 Terminal moraine

Benn and Evans (2010, pp. 488 - 513) distinguish between terrestrially and subaqueous formed terminal moraines. The terrestrial terminal moraines are divided into four different groups; proglacial glacitectonic landforms, push and squeeze moraines, dump moraines and ice-marginal aprons and latero-frontal fans and ramps. The different moraines are formed either purely by deposition, deformation of pre-existing sediments or a mixture of both processes (Boulton & Eyles 1979; van der Wateren 1995; Bennett 2001; Benn & Evans 2010; Aber *et al.* 2012).

The classification of the terrestrial part of the terminal moraine at Nordenskiöldbreen is hard to do, without a section with exposure of the internal structure. However, the surface morphology itself can reveal indications of the genesis. The double crested appearance in the western-most part of the ridge, on the northern shore (Figs 11, 20a), may be a result of thrusting of the material in front of the glacier during advance (Bennett 2001). Thrusting in active polythermal ice experiencing strong longitudinal compression is common at the transition from warm-based interior to cold-based ice at the margin (Hambrey *et al.* 1997). The two crests are assumed to represent two individual thrust ridges (Evans & Benn 2001). The décollement plane is assumed to have been where there is a difference in rheology between two sediment bodies (Sollid & Sørbel 1988; Etzelmüller 2000; Etzelmüller & Hagen 2005). This difference can be represented by a change in sedimentology, a bedding plane or a permafrost boundary (Benn & Evans 2010).

The moraines along the flanks of the forefield are single-crested and are therefore interpreted to have experienced less compressional push than the most terminal part (Bennett *et al.* 2000; Benn & Evans 2010). The transition from terminal to lateral moraine in the eastern hinges of the ridge on both shores is interpreted to be gradual (Boulton & Eyles 1979).

The subaqueous moraines form where glacier grounding lines are stable for a longer period (Powell & Molnia 1989; Fischer & Powell 1998).These types of moraines are called *morainal banks* and a subaqueous fan is normally associated with the morainal bank (Powell 1981). Various processes are suggested to be responsible for the material delivery; supraglacial debris slumps, calve dumping, en- and subglacial melt-out of debris bands, iceberg melt-out and push-up of unfrozen sediment from beneath the glacier margin (Powell 1981; Powell & Molnia 1989). On the distal side of the morainal bank, grounding-line fans build out. In the Nordenskiöldbreen, these are suggested to be represented by the debris flow lobes, observed in the subaqueous part.

The unifying factor between terrestrial and subaqueous end moraines is that the outermost moraine always will be termed *terminal moraine*, delimiting the ice advance. Benn and Evans (2010, pp. 488) argues that terminal moraines can be divided into *frontal* and *lateral* components, which fits well with the observations from Nordenskiöldbreen forefield. The picture gets complex if there has been several advances and older moraines have been overridden by new ice. Thereby, it can be difficult to determine which moraine is the youngest.

The presence of permafrost is argued to amplify the size of the terminal moraines in front of many glaciers in Svalbard (Etzelmüller 2000; Kjær *et al.* 2003; van der Meer 2004; Humlum *et al.* 2007). Another factor influencing the size of the terminal moraine may be the amount of available sediment (Hunter & Smith 2001; van der Meer 2004). In the fjord environment just the presence of an end moraine can affect the glacier retreat: similar to exposed bedrock, the moraine can function as a pinning point and thereby limit calving activity and retreat (Ottesen & Dowdeswell 2006; Benn *et al.* 2007). On the basis of the above discussion, it can be argued that terminal moraines show equifinality.

6.1.4 Transverse ridges

In the subaqueous part of the Nordenskiöldbreen forefield an extensive set of transverse ridges is mapped (Figs 10, 14, 18). This type of ridges frequently in the literature are called *De Geer Moraines (DGM)* after Gerhard de Geer who first described such landforms in Sweden (de Geer 1889). He suggested annual origin of the ridges, as the distance between the crests of the ridges coincided with the annual retreat rate obtained from varve chronology in the area (Lindén & Möller 2005). de Geer therefore termed them *annual moraines*.

The genesis of DGMs have been discussed throughout time (de Geer 1889; Lundqvist 2000; Lindén & Möller 2005; Lindén 2006; Bouvier *et al.* 2015). There are two dominating theories of the genesis; squeezing of sediment into submarginal crevasses (Hoppe 1959) and push of sediment either by a small winter advance or by a faster flowing ice during the summer (de Geer 1940). Hoppe (1959) argued from field evidence from Norrbotten, Sweden, that two to three ridges were formed each year. The term *annual moraine* therefore was misleading, and he renamed them *De Geer Moraines* (Lindén & Möller 2005). The sub-annual origin is supported by Lindén and Möller (2005).

The descriptive term *transverse ridges* is used in this thesis, as DGMs have been assigned a certain genesis. And the ridges found in Nordenskiöldbreen forefield are interpreted as recessional moraines, that have formed during momentary halts in the grounding line retreat (Lindén & Möller 2005). Whether their origin is annual or sub-annual is less important.

Similar transverse ridges have been observed in the neighbouring fjord of Billefjorden, Tempelfjorden, where they have been deposited by Tunabreen glacier (Flink 2013; Flink *et al.* 2015).

There, they are interpreted to have formed in the ice-marginal environment as a result of small winter advances and stabilisation due to the presence of sea ice (Flink *et al.* 2015). The small advances are suggested related to the imbalances caused by longitudinal extension due to calving at the ice front (Benn *et al.* 2007; Flink *et al.* 2015). Flink *et al.* (2015) match the transverse ridge location to the ice front position revealed from satellite images and conclude that they are annual retreat moraines. However, this method is limited by the satellite image resolution and resultant conclusions may not be as reliable as indicated. The ridge shapes clearly match the ice front, but their annual origin can be debated (Lindén & Möller 2005). It is obvious, however, that they represent retreat.

The author of this thesis participated in an excursion in Sweden (Nordqua 2015) where some of the classical DGMs were visited. The consensus within the excursion participants was that both the squeezing (Hoppe 1959) and pushing (de Geer 1940) genesis could be possible for the ridges found there.

The transverse ridges in the Nordenskiöldbreen forefield can suggestively be crevasse squeeze ridges, but they are assumed to be recessional ridges as only a few, very faint terrestrial examples are found, and since no surges have been observed at Nordenskiöldbreen - even though being possible (Jiskoot *et al.* 2000; Baeten *et al.* 2010; Farnsworth *et al.* 2016).

The formation process for the transverse ridges from (Flink *et al.* 2015) is adapted to the ridges in the Nordenskiöldbreen forefield. Flink *et al.* (2015) argues that the transverse ridges in Tempelfjorden are annual. This conclusion is problematic, as the resolution of the satellite images is 2 – 30 m and the bathymetry data is 5 x 5 m grid and landforms smaller than this are not included in the survey.

The transverse ridges overprint the streamlined bedforms (flutes and drumlinoids) and I argue that the most important implication of the presence of transverse ridges is their indication of *dynamic glacier retreat*. Thereby the inferred annual retreat rate of 170 m yr^{-1} during the deglaciation of outer Billefjorden (Baeten *et al.* 2010) is questioned, as the basis of the calculated rate is the appearance of transverse ridges.

Based on the discussion from Nordqua 2015, the research from Tempelfjorden (Flink *et al.* 2015), my own study from Nordenskiöldbreen and pre-existing (de Geer 1889, 1940; Zilliacus 1989; Bouvier *et al.* 2015) literature, it is argued that large parts of the research on transverse ridges, similar to the drumlinoid research, have been biased by the search for a definitive formation theory. Lindén and Möller (2005) leave the question open. They suggest that DGMs can form during both summer and winter, but that their formation is closely related to calving.

6.2 Preservation potential

A recently deglaciated forefield might contain a significant amount of dead-ice, and the landscape is therefore expected to change rapidly (Ballantyne 2002). The most prominent changes are relief adjustments and changes in the sedimentological characteristics of surface and near-surface sediments (Ballantyne 2002). The mass movement processes are expected to slow down according to the increasing distance to the ice margin and slope angles in the moraine will decrease and stabilize with time.

Terrestrial landforms are subject to postglacial erosion by nivation and deflation processes and their preservation potential therefore time constrained (Chirm 1980; Christiansen 1998). In the subaqueous record, the expression of the landforms on the seafloor surface is restricted by the postglacial sedimentation rate (Cowan & Powell 1991). The landforms are expected to be buried by finegrained sediments in a limited amount of time. However, the preservation potential in the marine record is from a stratigraphical (opposite to a geomorphological) point of view, high. This enhances the importance of geomorphological mapping. It is important to catch a snapshot of the landscape while the surface expression in both records still is high.

Right after initiation of the glacier retreat it is assumed that backwasting has been the main process for ice-decay (Schomacker 2007). This process will dominate as long as enough meltwater is available to facilitate new exposures of dead-ice (Schomacker & Kjær 2008). When the amount of dead-ice has decreased and thereby also the relief, the landscape will settle and downwasting will take over (Etzelmüller & Hagen 2005; Schomacker 2007; Schomacker & Kjær 2008; Benn & Evans 2010). Downwasting is suggested to be very slow in permafrost environments (Etzelmüller & Hagen 2005; Humlum *et al.* 2007). The higher relief observed closer to the glacier may indicate that the ice content is higher and less downwasting has occurred there.

Dead-ice landforms have a short life-time, compared to landforms with no ice, as they will be ever changing as long as they still contain ice (Schomacker *et al.* 2006; Schomacker 2007). When the internal ice has disappeared the landforms will continue to degrade due to regular erosion processes. The three stages of hummocky terrain are good examples of that. The supraglacial eskers occurring in the dead-ice terrain also have low preservation potential due to their supraglacial origin. The debris surface will mimic the dead-ice surface and the esker shape may be incorporated in the dead-ice topography. Therefore, it is important to map glacier forefields soon after deglaciation/glacier retreat to preserve the imprints of the glacier in a time-transgressive medium – a geomorphological map. Thereby will crucial evidence, in form of landforms, for past glacier dynamics be preserved.

6.3 Retreat history

Rate of glacial advance or retreat have been documented as key factors for landscape evolution and relief of glacier forefields for glaciers of both terrestrial and marine termination (Boulton 1972; Bennett *et al.* 1996; Sletten *et al.* 2001; Plassen *et al.* 2004). The retreat of Nordenskiöldbreen has been reconstructed from historical data. Ice front positions can be problematic to determine; as most onshore glaciers have debris covered snouts (Bennett *et al.* 1996; Etzelmüller & Hagen 2005; Humlum *et al.* 2007; Schomacker & Kjær 2008). Clean ice border has been used here, but is not giving the real picture, but more an indicator of landscape development.

Nordenskiöldbreen has been in overall retreat since 1896. Fig. no. 32 shows the reconstructed ice front positions from 1896 to 2015. The reconstructed positions of the ice front correlates with the presence of pinning points, where the ice has been able to 'rest' (Joughin *et al.* 2008). It is therefore argued that the exposed crystalline bedrock has worked as 'thresholds' for the glacier during retreat. A larger component of bedrock is exposed on the northern shore, compared to the southern shore. Whether this is related to general bedrock relief or just is a function of the southern shore being covered by extensive amounts of debris, still remains an open question.

The pattern of the meltwater channels, outwash fans and kame terraces also reveal interesting aspects of the retreat history. Several outwash fans have repeatedly cut through the landscape on the northern shore, initiating in north and fanning out towards south. On the southern shore one big outwash fan *outboard* of the terminal moraine and a few smaller fans in the glacier proximal area have been observed (Figs 12, 13). From this observation it is inferred, that glacial meltwater has played a different role on the respective shores. Lateral meltwater channels have been argued to delineate sequential positions of ice front positions for polythermal glaciers, where meltwater has been hindered to penetrate frozen glacier snouts (Dyke 1993). This is applicable for the pattern of the outwash fans on the northern shore, and the retreat pattern seen from the old aerial images supports (Figs 7, 31). The kame terraces on the northern shore mimics the glacier front and the different terrace steps represent temporary glacier halts.

Large volumes of meltwater also have occurred on the southern shore – indicated by the size of the outwash fan there. The debris-covered dead-ice is suggested to have obstructed the meltwater and let it outboard of the terminal moraine (van der Meer 2004). The amount of debris-covered dead-ice in the present marginal area is limited, and therefore the path of the water is less obstructed and small outwash fans are found there.

The retreat is expected to be enhanced by the presence of a reverse bedrock slope (Krimmel & Vaughn 1987; Warren 1992). This is conditioned by whether the bedrock exposed at the glacier front represents the closing fjord head or if the fjord waters extend further west (behind the present glacier front). The likelihood of the latter is implied by the presence of bedrock heights below sea level upglacier (Fig. 6).

Right behind the front and again ~ 2.5 km upglacier the surface of the glacier is heavily crevassed, with mainly transverse crevasses. Crevasses are *strain made visible* (Benn & Evans 2010, pp. 134) and form in regions of extending flow (Benn & Evans 2010). Close to the terminus, the formation of crevasses is connected to the longitudinal extension caused by the floatation when the glacier reaches the fjord (Benn *et al.* 2007). During fieldwork, it was observed that large parts of the ice front today rests on bedrock and the near-future evolution of the crevasses in the frontal zone will be interesting to follow. The location of crevasses further upglacier is suggested to be linked to a higher relief of the bedrock underneath, which also will lead to extensional flow of the glacier (Benn & Evans 2010). This is supported by the GPR survey by Petterson 2010, unpublished data (Fig. 6), where irregular bedrock topography was identified. Another question is if Nordenskiöldbreen still can be classified as a tidewater glacier, as only a limited part of the front still is buoyant.

6.4 Dynamics and thermal regime

Rachlewicz *et al.* (2007) argue that glaciers are sensitive recorders of climate change and Glasser and Hambrey (2001, pp. 697) state that "*glacier beds provide crucial information concerning past and present ice dynamics and thermal regime*". Dynamics, thermal regime and climate is part of a long chain of parameters with cross-cutting feedback mechanisms, that are not entirely understood (Benn & Evans 2010).

From this combined study of the Nordenskiöldbreen forefield, differences and similarities between the marine and terrestrial part of a Svalbard glacier have been highlighted. A difference in basal thermal regime between the two different archives is inferred from the landform record.
The streamlined bedforms, drumlinoids and flutes, are indicative of warm based ice (Stokes & Clark 2002). The appearance of streamlined bedforms indicate that the terrestrial part of Nordenskiöldbreen has been polythermal with a narrow, frozen snout and warm based interior at the LIA maximum extent/during advance. The width of the frozen snout during the LIA maximum position is assumed

to match the width of the terminal moraine (Figs 10 -15). The absence of recessional moraines on land and the presence of permafrost in Svalbard disfavours the glacier to have been warm-based throughout (Sollid & Sørbel 1988; Sollid & Sørbel 1994; Björnsson *et al.* 1996). Due to cold air temperatures, terrestrial glaciers in Svalbard need to have a thickness exceeding 100 m before warmbased conditions can arise (Murray & Porter 2001; Christoffersen *et al.* 2005). From historical images the height of the ice cliff of Nordenskiöldbreen at the LIA maximum is assumed not to have exceeded 100 m (Slater 1925). Differently, as discussed in section 6.1.3, the snout of the marine terminus is not assumed to have been frozen to its bed, and the geomorphology points towards a warm-based glacier in the marine part.

When Nordenskiöldbreen started retreating and thereby also thinning, the cold was able to affect the entire ice thickness in the terrestrial part of the glacier, and the width of the cold based margin is assumed to have increased. The gradient of the glacier front today is gentle as opposed to 1922 (Fig. 4). The well-preserved flutes and drumlinoids (deposited during advance) implies a cold-based retreat on land, as cold-based glaciers are weak erosive agents (Davis *et al.* 2006; Benn & Evans 2010) This is in accordance with observations from other glaciers with terrestrial terminus in Svalbard (Glasser & Hambrey 2001; Hansen 2003).

The appearance of recessional ridges on the seafloor indicates that Nordenskiöldbreen, even though in retreating phase, acts very dynamic in the marine part. Small winter advances might occur, but the overall mass-balance is negative (van Pelt *et al.* 2012). The more dynamic behaviour in the marine-based part is suggested to be related to the active calving front (Benn *et al.* 2007; Flink *et al.* 2015).

On the map (Fig. 10), an offset between the terrestrial and marine part of the terminal moraine in the northern part of the forefield is observed. This can suggestively be explained by the presence of a small, floating front, with a grounding line located behind the buoyant front during the LIA maximum extent (Benn *et al.* 2007; Joughin *et al.* 2008).

The grounding line/terminal moraine in the fjord mimics the bedrock topography, and it is suggested that the ice mass at the slope break could not 'reach down' to the bottom of the basin and instead was buoyant westwards of this point - if it extended beyond it at all (Fig. 18).

Numerous Svalbard glaciers are documented surge-type glaciers and much research has been directed towards surging glaciers, as their behaviour and dynamics are not fully understood (Clarke *et al.* 1986; Dowdeswell *et al.* 1991; Bennett *et al.* 1999; Larsen *et al.* 2006; Flink 2013; Sevestre *et al.*

2015). Surge-type glaciers show fluctuations on centennial to decadal scale (Benn & Evans 2010) and several theories for surge triggering have been suggested (Kamb *et al.* 1985; Clarke *et al.* 1986; Björnsson 1998; Fowler *et al.* 2001). The different theories have recently been tied together by the theory relating to the enthalpy balance of a glacier (Sevestre & Benn 2015; Farnsworth *et al.* 2016).

It is unknown if Nordenskiöldbreen has exhibited surge-type behaviour. Jiskoot et al. (2000) interpreted the glacier to have exhibited surge-type behaviour based on model results and 1936 oblique aerial images indicating a heavily crevassed glacier terminus. Jiskoot et al. (2000) supported their interpretation with reference to looped moraines and a heavily crevassed surface on historical aerial images. However, the landform record does not show any diagnostic evidence for Nordenskiöldbreen being a surge-type glacier (Figs 10 - 15). The streamlined bedforms do indicate fast flow, but neither crevasse squeeze ridges nor concertina eskers, potential evidence for past surge activity (Knudsen 1995; Farnsworth *et al.* 2016), have been identified. This could be due to not only constraints on formation but preservation potential. Despite the early map by de Geer (1910) (Fig. 7) and descriptions of the landscape (Tyrrell, 1922a, b) argues that the looping was caused by an advance of the tributary glacier, Gerritbreen. The looping is not very extensive. As of yet, no smoking gun has been identified, indicating surge-activity for Nordenskiöldbreen. Absence of surge landforms does not categorically mean that a given glacier has not surged in the past (Ingólfsson *et al.* 2016) , and Brynjólfsson et al. (2014) showed that surge-diagnostic evidence was lacking in the landform record of documented surging outlet glaciers in Iceland.

From a number of surveys of glaciers with terrestrial termini, Glasser and Hambrey (2001) showed that Svalbard glaciers have several dynamic modes. They argue that changes in ice thickness, dynamics and sedimentation are driven by changes in mass-balance, which is driven by climate. The shift from polythermal to frozen-bed conditions occurs within a short timeframe in receding Svalbard glaciers. Sevestre (2015) argues that this applies to small glaciers, but not larger glacier systems.

What will happen with Nordenskiöldbreen in the near-future? From field observations of the ice front in 2015 it is seen that the ice front now almost entirely rests on bedrock. The flow is thereby constrained by the bedrock. It can be questioned if Nordenskiöldbreen will develop into a surge-type glacier, since it cannot discharge mass in form of calving anymore. Contrary, the possible presence of a reverse bedrock slope behind the front might be able to enhance retreat – if the created basin still is connected to the fjord.

6.5 Landsystem model approach – how applicable?

As described in the introduction, conceptual models for the geomorphology of glacier forefields, so-called landsystem models, have been developed for terrestrial and marine terminating glaciers of both surge- and non-surge type (Evans & Rea 1999; Glasser & Hambrey 2001; Plassen *et al.* 2004). The landsystem model is a tool, which incites a holistic approach where the whole suite of landforms in the glacier forefield is assessed, and thereafter the glacier type is identified.

The landforms in the forefield of Nordenskiöldbreen have shown a very mixed glacial signal. The glacier both has a terrestrial and marine margin, it has an actively calving front (which maybe is about to change), it has been polythermal during advance, the terrestrial front has been coldbased during retreat, the marine part warmbased and on the southern shore signals of surge from the tributary glacier, Gerritbreen, occur. It is therefore problematic to apply the existing landsystem models to the landform record found there. The landsystem models are good tools for hypothesis testing but in the Nordenskiöldbreen case, the landsystem model approach is not applicable – at least not for identification of glacier type. Many other glaciers in Svalbard show similar signals due to recent retreat and it is therefore stressed that landsystem models should be used cautiously for Svalbard glaciers.

The surge-signal seen in the Nordenskiöldbreen forefield is suggested to be the surge imprint of the tributary glacier, Gerritbreen. This linkage is allowed due to historical maps and descriptions of the area made right after the surge had happened (see 3.2.1). If the landscape morphology had not been captured, the surge-signal of Gerritbreen would have been let out, and important knowledge about the dynamics of the Nordenskiöldbreen glacier system missed. This further supports the importance of geomorphological mapping and snapshot capturing of glacier forefields.

The landsystem models are first of all biased due to the number of glacier observations they are based on. The observations are few and they are often limited to glaciers that are in close proximity to infrastructure (*e.g.* Midtre Lovénbreen, Ny Ålesund, Glasser and Hambrey (2003)). They are simplified models where a *few* data points (glaciers) are used to generalize a certain behaviour for a *large* number of glaciers. This is problematic as glaciers are 'individual creatures' and their behaviour and dynamics depend on a range of factors such as the condition of the bed, resistant forces, location of the glacier, amount of received precipitation and type of termination. The landsystem models have also have evolved with time and more and more details have been added. Brynjólfsson *et al.* (2014) found no diagnostic landform evidence of the historically reported surges of the outlet glaciers of Drangajökull ice cap, and the coarse substratum and lack of soft, deformable sediment is suggested as a reason therefore. This points out the last problem with the landsystem models: the substratum and its effect on glacial behaviour is not incorporated in the existing landsystem models.

7. Conclusions

- By taking a holistic approach, the full LIA imprint and the post LIA-landscape evolution in both the marine and terrestrial archive of Nordenskiöldbreen is captured and preserved in a snap shot – a geomorphological map.
- A comprehensive literature review has gathered all the pre-existing literature on Nordenskiöldbreen.
- The glacier has retreated further in the fjord than on land, and most retreat is seen in the southern part of the basin. The retreat has been controlled by bedrock pinning points.
- Nordenskiöldbreen is diverse and has components of all Svalbard type glaciers. It is a polythermal, tidewater, calving, outlet glacier with both terrestrial and marine termination, with surge imprints in the southern part.
- The glacier has shown different dynamic behaviour in the marine based part compared to the terrestrial part, and the preservation potential is different in the two archives.
- The pre-existing landsystem models are not applicable for the Nordenskiöldbreen landsystem.
- The study contributes to an improved understanding of Svalbard glaciers and their response to climate fluctuations and is a part of the research project: "Holocene history of Svalbard Ice Caps and Glaciers" (see Research in Svalbard (RIS) database at: http://www.researchinsvalbard.no/project/7567).

8. References

Aber, J. S., Croot, D. G. & Fenton, M. M. 2012: *Glaciotectonic landforms and structures*. 182 pp. Springer Science & Business Media.

Alexanderson, H. & Murray, A. S. 2012: Luminescence signals from modern sediments in a glaciated bay, NW Svalbard. *Quaternary Geochronology 10*, 250-256.

Alley, R., Lawson, D., Larson, G., Evenson, E. & Baker, G. 2003: Stabilizing feedbacks in glacier-bed erosion. *Nature 424*, 758-760.

Anderson, M. P. 1989: Hydrogeologic facies models to delineate large-scale spatial trends in glacial and glaciofluvial sediments. *Geological Society of America Bulletin 101*, 501-511.

Ashworth, P. & Ferguson, R. 1986: Interrelationships of channel processes, changes and sediments in a proglacial braided river. *Geografiska Annaler. Series A. Physical Geography*, 361-371.

Backlund, H. G. 1908: Observations dans le Spitzberg central. Carte du Spitzberg occidental entre Klaas Billen Bay et le Storfiord avec la partie Sud du massif du Chydenius. Échelle 1 : 168 000. *Missions scient. p. 1. mesure d'un arc de méridien au Spitzberg, 1899 - 1901. Miss. russe. T. 2. Sect. 9. B. 2*, 32 pp.

Baeten, N. J., Forwick, M., Vogt, C. & Vorren, T. O. 2010: Late Weichselian and Holocene sedimentary environments and glacial activity in Billefjorden, Svalbard. *Geological Society, London, Special Publications 344*, 207-223.

Balescu, S. & Lamothe, M. 1994: Comparison of TL and IRSL age estimates of feldspar coarse grains from waterlain sediments. *Quaternary Science Reviews 13*, 437-444.

Ballantyne, C. K. 2002: Paraglacial geomorphology. *Quaternary Science Reviews 21*, 1935-2017.

Banerjee, I. 1975: Nature of esker sedimentation. *Society of Economic Paleontologists and Mineralogists, Special Publication 23*, 132-154.

Benn, D. & Evans, D. 2010: Glaciers and Glaciation. 802 pp. *Arnold, London*.

Benn, D. I. 1994: Fluted moraine formation and till genesis below a temperate valley glacier: Slettmarkbreen, Jotunheimen, southern Norway. *Sedimentology 41*, 279-292.

Benn, D. I., Warren, C. R. & Mottram, R. H. 2007: Calving processes and the dynamics of calving glaciers. *Earth-Science Reviews 82*, 143-179.

Bennett, M., Huddart, D., Hambrey, M. & Ghienne, J. F. 1996: Moraine development at the high-arctic valley glacier Pedersenbreen, Svalbard. *Geografiska Annaler. Series A. Physical Geography*, 209-222.

Bennett, M. R. 2001: The morphology, structural evolution and significance of push moraines. *Earth-Science Reviews 53*, 197-236.

Bennett, M. R., Hambrey, M. J., Huddart, D., Glasser, N. F. & Crawford, K. 1999: The landform and sediment assemblage produced by a tidewater glacier surge in Kongsfjorden, Svalbard. *Quaternary Science Reviews 18*, 1213-1246.

Bennett, M. R., Huddart, D., Glasser, N. F. & Hambrey, M. J. 2000: Resedimentation of debris on an ice-cored lateral moraine in the high-Arctic (Kongsvegen, Svalbard). *Geomorphology 35*, 21-40.

Björnsson, H. 1998: Hydrological characteristics of the drainage system beneath a surging glacier. *Nature 395*, 771-774.

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

Błaszczyk, M., Jania, J. A. & Hagen, J. O. 2009: Tidewater glaciers of Svalbard: Recent changes and estimates of calving fluxes. *Pol. Polar Res 30*, 85-142.

Boulton, G. 1970a: On the deposition of subglacial and melt-out tills at the margins of certain Svalbard glaciers. *Journal of Glaciology 9*, 231-245.

Boulton, G. 1970b: On the origin and transport of englacial debris in Svalbard glaciers. *Journal of Glaciology 9*, 213-229.

Boulton, G. 1976: The origin of glacially fluted surfaces Á/observations and theory. *Journal of Glaciology 17*, 287-309.

Boulton, G. & Eyles, N. 1979: Sedimentation by valley glaciers: a model and genetic classification. *Moraines and varves 33*, 11-23.

Boulton, G., Van der Meer, J., Beets, D., Hart, J. & Ruegg, G. 1999: The sedimentary and structural evolution of a recent push moraine complex: Holmstrømbreen, Spitsbergen. *Quaternary Science Reviews 18*, 339-371.

Boulton, G. S. 1972: Modern Arctic glaciers as depositional models for former ice sheets. *Journal of the Geological Society 128*, 361-393.

Bouvier, V., Johnson, M. D. & Påsse, T. 2015: Distribution, genesis and annual-origin of De Geer moraines in Sweden: insights revealed by LiDAR. *GFF 137*, 319-333.

Briner, J. P. 2007: Supporting evidence from the New York drumlin field that elongate subglacial bedforms indicate fast ice flow. *Boreas 36*, 143-147.

Broster, B. E. & Hicock, S. R. 1985: Multiple flow and support mechanisms and the development of inverse grading in a subaquatic glacigenic debris flow. *Sedimentology 32*, 645-657.

Brown, R. R. 1920: Recent developments in spitsbergen 1. *The Scottish Geographical Magazine 36*, 111-116.

Brynjólfsson, S., Schomacker, A. & Ingólfsson, Ó. 2014: Geomorphology and the Little Ice Age extent of the Drangajökull ice cap, NW Iceland, with focus on its three surge-type outlets. *Geomorphology 213*, 292-304.

Chirm, T. 1980: Glacier balances in the dry valleys area, Victoria Land, Antarctica.

Chisholm, M. 1967: General systems theory and geography. *Transactions of the Institute of British Geographers*, 45-52.

Christiansen, H. H. 1998: Nivation forms and processes in unconsolidated sediments, NE Greenland. *Earth Surface Processes and Landforms 23*, 751-760.

Christoffersen, P., Piotrowski, J. A. & Larsen, N. K. 2005: Basal processes beneath an Arctic glacier and their geomorphic imprint after a surge, Elisebreen, Svalbard. *Quaternary Research 64*, 125-137.

Clark, C. D., Hughes, A. L., Greenwood, S. L., Spagnolo, M. & Ng, F. S. 2009: Size and shape characteristics of drumlins, derived from a large sample, and associated scaling laws. *Quaternary Science Reviews 28*, 677-692.

Clarke, G. K., Schmok, J. P., Ommanney, C. S. L. & Collins, S. G. 1986: Characteristics of surge-type glaciers. *Journal of Geophysical Research: Solid Earth (1978–2012) 91*, 7165-7180.

Clayton, L. 1964: Karst topography on stagnant glaciers. *Journal of Glaciology 5*, 107-112.

Cleland, C. E. 2013: Common cause explanation and the search for a smoking gun. *Geological Society of America Special Papers 502*, 1-9.

Cole, G. A. 1911: Glacial features in Spitsbergen in relation to Irish geology. *Proceedings of the Royal Irish Academy. Section B: Biological, Geological, and Chemical Science 29*, 191-208.

Conway, W. M. 1898: An Exploration in 1897 of Some of the Glaciers of Spitsbergen. *Geographical Journal*, 137-151.

Cook, S. J., Waller, R. I. & Knight, P. G. 2006: Glaciohydraulic supercooling: the process and its significance. *Progress in physical geography 30*, 577-588.

Cowan, E. A. & Powell, R. D. 1991: Ice-proximal sediment accumulation rates in a temperate glacial fjord, southeastern Alaska. *Geological Society of America Special Papers 261*, 61-74.

D'Andrea, W. J., Vaillencourt, D. A., Balascio, N. L., Werner, A., Roof, S. R., Retelle, M. & Bradley, R. S. 2012: Mild Little Ice Age and unprecedented recent warmth in an 1800 year lake sediment record from Svalbard. *Geology 40*, 1007-1010.

Dahl, R. 1965: Plastically sculptured detail forms on rock surfaces in northern Nordland, Norway. *Geografiska Annaler. Series A. Physical Geography*, 83-140.

Dallmann, W. K., Piepjohn, K. & Blomeier, D. 2004: Geological Map of Billefjorden, Central Spitsbergen, Svalbard: With Geological Excursion Guide. Norsk Polarinstitutt.

Davis, P. T., Briner, J. P., Coulthard, R. D., Finkel, R. W. & Miller, G. H. 2006: Preservation of Arctic landscapes overridden by cold-based ice sheets. *Quaternary Research 65*, 156-163.

de Geer, G. 1889: Ändmoräner i trakten mellan Spånga och Sundbyberg. *Geologiska Föreningens i Stockholm Förhandlingar 11*, 395-397.

de Geer, G. 1910: Guide de l'excursion au Spitsberg. *Excursion A1. Stockholm, Sweden: XI Inter– national Geological Congress*, 305-310 pp.
de Geer, G. 1940: *Geochronologia suecica principles*. 367 pp. Almqvist & Wiksells, Stockholm.
den Ouden, M., Reijmer, C., Pohjola, V. A., van de Wal, R., Oerlemans, J. & Boot, W. 2010: Stand-alone single-frequency GPS ice velocity observations on Nordenskiöldbreen, Svalbard. *The Cryosphere 4*, 593-604.
Donnelly, R. & Harris, C. 1989: Sedimentology and origin of deposits from a small ice-dammed lake, Leirbreen, Norway. *Sedimentology 36*, 581-600.
Dowdeswell, J. & Fugelli, E. 2012: The seismic architecture and geometry of grounding-zone wedges formed at the marine margins of past ice sheets. *Geological Society of America Bulletin 124*, 1750-1761.
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. A. & Dowdeswell, E. K. 1989: Debris in icebergs and rates of glaci-marine sedimentation: observations from Spitsbergen and a simple model. *The Journal of Geology*, 221-231.
Dowdeswell, J. A., Elverhfi, A. & Spielhagen, R. 1998: Glacimarine sedimentary processes and facies on the polar North Atlantic margins *Quaternary Science Reviews 17*, 243-272.
Dowdeswell, J. A. & Forsberg, C. F. 1992: The size and frequency of icebergs and bergy bits derived from tidewater glaciers in Kongsfjorden, northwest Spitsbergen. *Polar Research 11*, 81-91.
Dowling, T. 2016: The drumlin problem: streamlined subglacial bedforms in southern Sweden. *Quaternary Sciences Department of Geology.* Lund University. Doctoral dissertation. 175 pp.
Dyke, A. S. 1993: Landscapes of cold-centred Late Wisconsinan ice caps, Arctic Canada. *Progress in physical geography 17*, 223-247.
Elverhøi, A., Lønne, Ø. & Seland, R. 1983: Glaciomarine sedimentation in a modern fjord environment, Spitsbergen. *Polar Research 1*, 127-150.
Elverhøi, A., Svendsen, J. I., Solheim, A., Andersen, E. S., Milliman, J., Mangerud, J. & Hooke, R. L. 1995: Late Quaternary Sediment Yield from the High Arctic Svalbard Area. *The Journal of Geology 103*, 1-17.
Etzelmüller, B. 2000: Quantification of thermo-erosion in pro-glacial areas-examples from Svalbard. *Zeitschrift für Geomorphologie, NF*, 343-361.
Etzelmüller, B. & Hagen, J. O. 2005: Glacier-permafrost interaction in Arctic and alpine mountain environments with examples from southern Norway and Svalbard. *Geological Society, London, Special Publications 242*, 11-27.
Evans, D. & Benn, D. 2001: Our glacial legacy: Earth's giant bulldozers. *Geography Review 14*, 29-35.
Evans, D., Phillips, E., Hiemstra, J. & Auton, C. 2006: Subglacial till: formation, sedimentary characteristics and classification. *Earth-Science Reviews 78*, 115-176.
Evans, D. J. & Rea, B. R. 1999: Geomorphology and sedimentology of surging glaciers: a land-systems approach. *Annals of Glaciology 28*, 75-82.
Evans, D. J., Twigg, D. R., Rea, B. R. & Orton, C. 2009: Surging glacier landsystem of Tungnaárjökull, Iceland. *Journal of Maps 5*, 134-151.
Farnsworth, W. R., Ingólfsson, Ó., Retelle, M. & Schomacker, A. 2016: Over 400 previously undocumented Svalbard surge-type glaciers identified. *Geomorphology 264*, 52-60.
Fischer, M. P. & Powell, R. D. 1998: A simple model for the influence of push-morainal banks on the calving and stability of glacial tidewater termini. *Journal of Glaciology 44*, 31-41.
Flink, A. 2013: Dynamics of surging tidewater glaciers in Tempelfjorden, Spitsbergen. *Department of Physical Geography and Quaternary Geology.* Stockholm University. Master's thesis. 58 pp.
Flink, A. E., Noormets, R., Kirchner, N., Benn, D. I., Luckman, A. & Lovell, H. 2015: The evolution of a submarine landform record following recent and multiple surges of Tunabreen glacier, Svalbard. *Quaternary Science Reviews 108*, 37-50.
Forwick, M. & Vorren, T. O. 2009: Late Weichselian and Holocene sedimentary environments and ice rafting in Isfjorden, Spitsbergen. *Palaeogeography, Palaeoclimatology, Palaeoecology 280*, 258-274.

Forwick, M. & Vorren, T. O. 2010: Stratigraphy and deglaciation of the Isfjorden area, Spitsbergen. *Norwegian Journal of Geology/Norsk Geologisk Forening 90*, 163-179.
Fowler, A., Murray, T. & Ng, F. 2001: Thermally controlled glacier surging. *Journal of Glaciology 47*, 527-538.
Frazer, R. 1922: The topographical work of the Oxford University Expedition to Spitsbergen (1921). *The Geographical Journal 60*, 321-334.
Frenot, Y., Gloaguen, J., Cannavacciuolo, M. & Bellido, A. 1998: Primary succession on glacier forelands in the subantarctic Kerguelen Islands. *Journal of Vegetation Science 9*, 75-84.
Friis, N. 2015: Stratigraphy and sedimentary properties of drumlinoid landforms in the forefield of Nordenskiöldbreen, Svalbard. *Faculty of Science.* University of Copenhagen, København. Master's thesis. 57 pp.
Førland, E. J., Benestad, R. E., Flatøy, F., Hanssen-Bauer, I., Haugen, J. E., Isaksen, K., Sorteberg, A. & Ådlandsvik, B. 2010: *Klimautvikling i Nord-Norge og på Svalbard i perioden 1900–2100: klimaendringer i norsk Arktis: NorACIA delutredning 1*. 52 pp. Norwegian Polar Institute.
Garwood, E. J. 1899: Additional Notes on the Glacial Phenomena of Spitsbergen. *Quarterly Journal of the Geological Society 55*, 681-691.
Gilbert, R. 1983: Sedimentary processes of canadian arctic fjords. *Sedimentary Geology 36*, 147-175.
Glasser, N. & Hambrey, M. 2003: Ice-marginal terrestrial landsystems: Svalbard polythermal glaciers. *Glacial Landsystems. Arnold, London*, 65-88.
Glasser, N. F. & Hambrey, M. J. 2001: Styles of sedimentation beneath Svalbard valley glaciers under changing dynamic and thermal regimes. *Journal of the Geological Society 158*, 697-707.
Goudie, A. 2004: Encyclopedia of geomorphology. *2*, 321 - 323.
Hagen, J. O., Kohler, J., Melvold, K. & Winther, J. G. 2003: Glaciers in Svalbard: mass balance, runoff and freshwater flux. *Polar Research 22*, 145-159.
Hagen, J. O., Liestøl, O., Roland, E. & Jørgensen, T. 1993: *Glacier atlas of Svalbard and Jan Mayen*. pp. Norwegian Polar Institute.
Haines-Young, R. H. & Petch, J. 1983: Multiple working hypotheses: equifinality and the study of landforms. *Transactions of the Institute of British Geographers*, 458-466.
Hambrey, M. J. & Glasser, N. F. 2003: The role of folding and foliation development in the genesis of medial moraines: examples from Svalbard glaciers. *The Journal of Geology 111*, 471-485.
Hambrey, M. J., Huddart, D., Bennett, M. R. & Glasser, N. F. 1997: Genesis of ‘hummocky moraines’ by thrusting in glacier ice: evidence from Svalbard and Britain. *Journal of the Geological Society 154*, 623-632.
Hansen, S. 2003: From surge-type to non-surge-type glacier behaviour: midre Lovénbreen, Svalbard. *Annals of Glaciology 36*, 97-102.
Hanssen-Bauer, I., Solas, M. & Steffensen, E. 1990: The climate of Spitsbergen. . *DNMI-Rapport 39*, 40.
Hart, B. & Plint, A. 1989: Gravelly shoreface deposits: a comparison of modern and ancient facies sequences. *Sedimentology 36*, 551-557.
Hart, J. & Rose, J. 2001: Approaches to the study of glacier bed deformation. *Quaternary International 86*, 45-58.
Hart, J. K. 1997: The relationship between drumlins and other forms of subglacial glaciotectonic deformation. *Quaternary Science Reviews 16*, 93-107.
Holmgren, S. U., Bigler, C., Ingólfsson, Ó. & Wolfe, A. P. 2010: The Holocene–Anthropocene transition in lakes of western Spitsbergen, Svalbard (Norwegian High Arctic): climate change and nitrogen deposition. *Journal of Paleolimnology 43*, 393-412.
Hoppe, G. 1959: Glacial morphology and inland ice recession in northern Sweden. *Geografiska Annaler 41*, 193-212.
Hoppe, G. & Schytt, V. 1953: Some observations on fluted moraine surfaces. *Geografiska Annaler*, 105-115.
Hormes, A., Gjermundsen, E. F. & Rasmussen, T. L. 2013: From mountain top to the deep sea–Deglaciation in 4D of the northwestern Barents Sea ice sheet. *Quaternary Science Reviews 75*, 78-99.
Humlum, O. 2005: Holocene permafrost aggradation in Svalbard. *Geological Society, London, Special Publications 242*, 119-129.

Humlum, O., Christiansen, H. H. & Juliussen, H. 2007: Avalanche-derived rock glaciers in Svalbard. *Permafrost and Periglacial Processes 18*, 75-88.
Hunter, L. E. & Smith, G. W. 2001: Morainal banks and the deglaciation of coastal Maine. *Special papers - Geological Society of America*, 151-170.
Ingólfsson, Ó. 2011: Fingerprints of Quaternary glaciations on Svalbard. *Geological Society, London, Special Publications 354*, 15-31.
Ingólfsson, Ó., Benediktsson, Í. Ö., Schomacker, A., Kjær, K. H., Brynjólfsson, S., Jónsson, S. A., Korsgaard, N. J. & Johnson, M. D. 2016: Glacial geological studies of surge-type glaciers in Iceland—Research status and future challenges. *Earth-Science Reviews 152*, 37-69.
Jakobsson, M., Ingólfsson, Ó., Long, A. J. & Spielhagen, R. F. 2014: The dynamic Arctic. *Quaternary Science Reviews 92*, 1-8.
Jessen, S. P., Rasmussen, T. L., Nielsen, T. & Solheim, A. 2010: A new Late Weichselian and Holocene marine chronology for the western Svalbard slope 30,000–0 cal years BP. *Quaternary Science Reviews 29*, 1301-1312.
Jiskoot, H., Murray, T. & Boyle, P. 2000: Controls on the distribution of surge-type glaciers in Svalbard. *Journal of Glaciology 46*, 412-422.
Johnson, M. D., Schomacker, A., Benediktsson, Í. Ö., Geiger, A. J., Ferguson, A. & Ingólfsson, Ó. 2010: Active drumlin field revealed at the margin of Múlajökull, Iceland: a surge-type glacier. *Geology 38*, 943-946.
Jónsson, S. A., Schomacker, A., Benediktsson, Í. Ö., Ingólfsson, Ó. & Johnson, M. D. 2014: The drumlin field and the geomorphology of the Múlajökull surge-type glacier, central Iceland. *Geomorphology 207*, 213-220.
Joughin, I., Howat, I., Alley, R. B., Ekstrom, G., Fahnestock, M., Moon, T., Nettles, M., Truffer, M. & Tsai, V. C. 2008: Ice-front variation and tidewater behavior on Helheim and Kangerdlugssuaq Glaciers, Greenland. *Journal of Geophysical Research: Earth Surface 113*, 11.
Kamb, B., Raymond, C., Harrison, W., Engelhardt, H., Echelmeyer, K., Humphrey, N., Brugman, M. & Pfeffer, T. 1985: Glacier surge mechanism: 1982–1983 surge of Variegated Glacier, Alaska. *science 227*, 469-479.
Karczewski, A., Borówka, M., Gonera, P., Kasprzak, L., Kłysz, P., Kostrzewski, A., Lindner, L., Marks, L., Rygielski, W. & Stankowski, W. 1990: Geomorphology–Petuniabukta Billefjorden Spitsbergen 1: 40 000. *Adam Mickiewicz University, Poznań*.
King, E. C., Hindmarsh, R. C. & Stokes, C. 2009: Formation of mega-scale glacial lineations observed beneath a West Antarctic ice stream. *Nature Geoscience 2*, 585-588.
King, E. C., Woodward, J. & Smith, A. M. 2007: Seismic and radar observations of subglacial bed forms beneath the onset zone of Rutford Ice Stream, Antarctica. *Journal of Glaciology 53*, 665-672.
Kjær, K. H., Korsgaard, N. J. & Schomacker, A. 2008: Impact of multiple glacier surges—a geomorphological map from Brúarjökull, East Iceland. *Journal of Maps 4*, 5-20.
Kjær, K. H. & Krüger, J. 2001: The final phase of dead-ice moraine development: processes and sediment architecture, Kötlujökull, Iceland. *Sedimentology 48*, 935-952.
Kjær, K. H., Krüger, J. & van der Meer, J. J. 2003: What causes till thickness to change over distance? Answers from Mýrdalsjökull, Iceland. *Quaternary Science Reviews 22*, 1687-1700.
Kleman, J. 1990: On the use of glacial striae for reconstruction of paleo-ice sheet flow patterns. *Geografiska Annaler. Series A. Physical Geography*, 217-236.
Klysz, P., Lindner, L., Marks, L. & Wysokinski, L. 1989: Late Pleistocene and Holocene relief remodelling in the Ebbadalen-Nordenskioldbreen region in Olav V Land, central Spitsbergen. *Polish Polar Research 10*, 277-301.
Knudsen, Ó. 1995: Concertina eskers, Bruarjökull, Iceland: an indicator of surge-type glacier behaviour. *Quaternary Science Reviews 14*, 487-493.
Kohler, J., James, T., Murray, T., Nuth, C., Brandt, O., Barrand, N., Aas, H. & Luckman, A. 2007: Acceleration in thinning rate on western Svalbard glaciers. *Geophysical Research Letters 34*, 5.

Kostrzewski, A., Kaniecki, A., Kapuściński, J., Klimczak, R., Stach, A. & Zwoliński, Z. 1989: The dynamics and rate of denudation of glaciated and non-glaciated catchments, central Spitsbergen. *Polish Polar Research 10*, 317-367.
Krimmel, R. M. & Vaughn, B. H. 1987: Columbia Glacier, Alaska: changes in velocity 1977–1986. *Journal of Geophysical Research: Solid Earth 92*, 8961-8968.
Krüger, J. & Kjær, K. H. 1999: A data chart for field description and genetic interpretation of glacial diamicts and associated sediments… with examples from Greenland, Iceland, and Denmark. *Boreas 28*, 386-402.
Krüger, J. & Kjær, K. H. 2000: De-icing progression of ice-cored moraines in a humid, subpolar climate, Kötlujökull, Iceland. *The Holocene 10*, 737-747.
Krüger, J. & Thomsen, H. H. 1984: Morphology, stratigraphy, and genesis of small drumlins in front of the glacier Myrdalsjökull, South Iceland. *Journal of Glaciology 30*, 94-105.
König, M., Nuth, C., Kohler, J., Moholdt, G. & Pettersen, R. 2014: A digital glacier database for Svalbard. *Global Land Ice Measurements from Space*, 229-239 pp. Springer.
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.
Lawson, D. E. 1979: Sedimentological Analysis of the Western Terminus Region of the Matanuska Glacier, Alaska. *CRREL Report 79-9*, 122.
Lindén, M. 2006: Glaciodynamics, Deglacial Landforms and Isostatic Uplift During the Last Deglaciation of Norrbotten, Sweden. *LUNDQUA Thesis 54*, 29.
Lindén, M. & Möller, P. 2005: Marginal formation of De Geer moraines and their implications to the dynamics of grounding-line recession. *Journal of Quaternary Science 20*, 113-133.
Lundqvist, J. 2000: Palaeoseismicity and De Geer Moraines. *Quaternary International 68–71*, 175-186.
Maher, H. & Braathen, A. 2011: Løvehovden fault and Billefjorden rift basin segmentation and development, Spitsbergen, Norway. *Geological Magazine 148*, 154-170.
Maizels, J. 1993: Lithofacies variations within sandur deposits: the role of runoff regime, flow dynamics and sediment supply characteristics. *Sedimentary Geology 85*, 299-325.
Mathieson, J., Campbell, A. F., Cowan, G. M. & Scottish Spitsbergen, S. 1919: Spitsbergen: Map of Klaas Billen Bay & Temple Bay. J. Bartholomew & Son, Edinburgh.
Menzies, J. 1979: A review of the literature on the formation and location of drumlins. *Earth-Science Reviews 14*, 315-359.
Miall, A. D. 1977: Lithofacies types and vertical profile models in braided river deposits: a summary. *Fluvial Sedimentology Memoir 5*, 597 - 604.
Moran, S. 1980: Glacier-bed landforms of the prairie region of North America. *Journal of Glaciology 5*, 458-476.
Murray, T. & Porter, P. R. 2001: Basal conditions beneath a soft-bedded polythermal surge-type glacier: Bakaninbreen, Svalbard. *Quaternary International 86*, 103-116.
Nilsen, F., Cottier, F., Skogseth, R. & Mattsson, S. 2008: Fjord–shelf exchanges controlled by ice and brine production: the interannual variation of Atlantic Water in Isfjorden, Svalbard. *Continental Shelf Research 28*, 1838-1853.
Nuth, C., Kohler, J., Aas, H., Brandt, O. & Hagen, J. 2007: Glacier geometry and elevation changes on Svalbard (1936–90): a baseline dataset. *Annals of Glaciology 46*, 106-116.
Odell, N. E. 1922: Geological notes from the Oxford expedition to Spitsbergen. *The Geographical Journal 60*, 424-426.
Odell, N. E. 1927: Preliminary Notes on the Geology of the Eastern Parts of Central Spitsbergen; with special reference to the Problem of the Hecla Hook Formation. *Quarterly Journal of the Geological Society 83*, 147-162.
Ottesen, D. & Dowdeswell, J. 2006: Assemblages of submarine landforms produced by tidewater glaciers in Svalbard. *Journal of Geophysical Research: Earth Surface 111*, 16.
Ottesen, D., Dowdeswell, J., Benn, D., Kristensen, L., Christiansen, H., Christensen, O., Hansen, L., Lebesbye, E., Forwick, M. & Vorren, T. 2008: Submarine landforms characteristic of glacier surges in two Spitsbergen fjords. *Quaternary Science Reviews 27*, 1583-1599.

Patterson, C. J. & Hooke, R. L. 1995: Physical environment of drumlin formation. *Journal of Glaciology 41*, 30-38.
Plassen, L., Vorren, T. O. & Forwick, M. 2004: Integrated acoustic and coring investigation of glacigenic deposits in Spitsbergen fjords. *Polar Research 23*, 89-110.
Porter, S. C. 1975: Weathering rinds as a relative-age criterion: Application to subdivision of glacial deposits in the Cascade Range. *Geology 3*, 101-104.
Powell, R. D. 1981: A model for sedimentation by tidewater glaciers. *Annals of Glaciology 2*, 129-134.
Powell, R. D. & Molnia, B. F. 1989: Glacimarine sedimentary processes, facies and morphology of the south-southeast Alaska shelf and fjords. *Marine Geology 85*, 359-390.
Rachlewicz, G. & Szczuciński, W. 2008: Changes in thermal structure of permafrost active layer in a dry polar climate, Petuniabukta, Svalbard. *Polish Polar Research 29*, 261-278.
Rachlewicz, G., Szczuciński, W. & Ewertowski, M. 2007: Post‒"Little Ice Age" retreat rates of glaciers around Billefjorden in central Spitsbergen, Svalbard. *Polish Polar Research 28*, 159-186.
Rose, J. 1987: Drumlins as part of glacier bedform continuum. *Drumlin symposium*, 103-116 pp.
Rose, K. C. & Hart, J. K. 2008: Subglacial comminution in the deforming bed: Inferences from SEM analysis. *Sedimentary Geology 203*, 87-97.
Rubensdotter, L. 2012: Landforms and sediments: Todalen, upper Gangdalen and upper Bødalen, Svalbard, Map in scale 1:25 000. NGU, Trondheim.
Rubensdotter, L. 2015: Landforms and sediments : Bjørndalen - Vestpynten, Svalbard, Map in scale 1:10 000. NGU, Trondheim.
Ruszczyńska-Szenajch, H. 2001: Lodgement till" and "deformation till. *Quaternary Science Reviews 20*, 579-581.
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.
Salvigsen, O. & Høgvard, K. 2006: Glacial history, Holocene shoreline displacement and palaeoclimate based on radiocarbon ages in the area of Bockfjorden, north-western Spitsbergen, Svalbard. *Polar Research 25*, 15-24.
Schomacker, A. 2007: Dead-ice under different climate conditions: processes, landforms, sediments and melt rates in Iceland and Svalbard. *LUNDQUA Thesis 59*, 25.
Schomacker, A. 2008: What controls dead-ice melting under different climate conditions? A discussion. *Earth-Science Reviews 90*, 103-113.
Schomacker, A., Benediktsson, Í. Ö. & Ingólfsson, Ó. 2014: The Eyjabakkajökull glacial landsystem, Iceland: Geomorphic impact of multiple surges. *Geomorphology 218*, 98-107.
Schomacker, A., Benediktsson, Í. Ö., Ingólfsson, Ó., Friis, B., Korsgaard, N. J., Kjær, K. H. & Keiding, J. K. 2012: Late Holocene and modern glacier changes in the marginal zone of Sólheimajökull, South Iceland. *Jökull 62*, 111-130.
Schomacker, A. & Kjær, K. H. 2008: Quantification of dead-ice melting in ice-cored moraines at the high-Arctic glacier Holmströmbreen, Svalbard. *Boreas 37*, 211-225.
Schomacker, A., Krüger, J. & Kjær, K. H. 2006: Ice-cored drumlins at the surge-type glacier Brúarjökull, Iceland: a transitional-state landform. *Journal of Quaternary Science 21*, 85-93.
Sevestre, H. 2015: Surge-type glaciers: controls, processes and distribution. *Series of dissertations 1671*, 53.
Sevestre, H. & Benn, D. I. 2015: Climatic and geometric controls on the global distribution of surge-type glaciers: implications for a unifying model of surging. *Journal of Glaciology 61*, 646-662.
Sevestre, H., Benn, D. I., Hulton, N. R. & 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.
Shaw, J. 1983: Drumlin formation related to inverted melt-water erosional marks. *Journal of Glaciology 29*, 461-479.
Shaw, J. & Kvill, D. 1984: A glaciofluvial origin for drumlins of the Livingstone Lake area, Saskatchewan. *Canadian Journal of Earth Sciences 21*, 1442-1459.
Shaw, J., Kvill, D. & Rains, B. 1989: Drumlins and catastrophic subglacial floods. *Sedimentary Geology 62*, 177-202.

Slater, G. 1925: Observations on the Nordenskiöld and neighboring glaciers of Spitsbergen, 1921. *The Journal of Geology*, 408-446.
Sletten, K., Lyså, A. & Lønne, I. 2001: Formation and disintegration of a high-arctic ice-cored moraine complex, Scott Turnerbreen, Svalbard. *Boreas 30*, 272-284.
Smalley, I. J. & Unwin, D. J. 1968: The formation and shape of drumlins and their distribution and orientation in drumlin fields. *Journal of Glaciology 7*, 377-390.
Smith, A., Murray, T., Nicholls, K., Makinson, K., Aðalgeirsdóttir, G., Behar, A. & Vaughan, D. 2007: Rapid erosion, drumlin formation, and changing hydrology beneath an Antarctic ice stream. *Geology 35*, 127-130.
Sollid, J. L. & Sørbel, L. 1988: Influence of temperature conditions in formation of end moraines in Fennoscandia and Svalbard. *Boreas 17*, 553-558.
Sollid, J. L. & Sørbel, L. 1994: Distribution of glacial landforms in southern Norway in relation to the thermal regime of the last continental ice sheet. *Geografiska Annaler. Series A. Physical Geography*, 25-35.
Stokes, C. R. & Clark, C. D. 2002: Are long subglacial bedforms indicative of fast ice flow? *Boreas 31*, 239-249.
Stokes, C. R., Spagnolo, M. & Clark, C. D. 2011: The composition and internal structure of drumlins: Complexity, commonality, and implications for a unifying theory of their formation. *Earth-Science Reviews 107*, 398-422.
Stott, F. 1936: The marine foods of birds in an Inland Fjord Region in West Spitsbergen: part 1. Plankton and in Shore Benthos. *The Journal of Animal Ecology*, 356-369.
Strzelecki, M. 2011: Schmidt hammer tests across a recently deglacierized rocky coastal zone in Spitsbergen-is there a" coastal amplification" of rock weathering in polar climates? *Polish Polar Research 32*, 239-252.
Sugden, D. E., Glasser, N. & Clapperton, C. M. 1992: Evolution of large roches moutonnées. *Geografiska Annaler. Series A. Physical Geography*, 253-264.
Svendsen, J. I. & Mangerud, J. 1997: Holocene glacial and climatic variations on Spitsbergen, Svalbard. *The Holocene 7*, 45-57.
Szczuciński, W. & Zajączkowski, M. 2012: Factors controlling downward fluxes of particulate matter in glacier-contact and non-glacier contact settings in a subpolar fjord (Billefjorden, Svalbard). *Int Assoc Sedimentol Spec Publ 44*, 369-386.
Szczuciński, W., Zajączkowski, M. & Scholten, J. 2009: Sediment accumulation rates in subpolar fjords–Impact of post-Little Ice Age glaciers retreat, Billefjorden, Svalbard. *Estuarine, Coastal and Shelf Science 85*, 345-356.
Tyrrell, G. W. 1921: Geographical observations in Spitsbergen, 1919 and 1920. *The Scottish Geographical Magazine 37*, 227-242.
Tyrrell, G. W. 1922a: I. The Glaciers of Spitsbergen. *Transactions of the Geological Society of Glasgow 17*, 1-49.
Tyrrell, G. W. 1922b: X.—The Pre-Devonian Basement Complex of Central Spitsbergen. *Transactions of the Royal Society of Edinburgh 53*, 209-229.
van der Meer, J. J. 2004: *Spitsbergen Push Moraines: Including a Translation of K. Gripp: Glaciologische und Geologische Ergebnisse Der Hamburgischen Spitzbergen-Expedition 1927*. 200 pp. Elsevier, London.
van der Wateren, D. 1995: Processes of glaciotectonism. *Glacial environments Vol. 1. Modern glacial environments: Processes, dynamics and sediments*, 309-335.
van Pelt, W., Oerlemans, J., Reijmer, C., Pettersson, R., Pohjola, V. A., Isaksson, E. & Divine, D. 2013: An iterative inverse method to estimate basal topography and initialize ice flow models. *The Cryosphere 7*, 987-1006.
van Pelt, W. J. J., Oerlemans, J., Reijmer, C. H., Pohjola, V. A., Pettersson, R. & van Angelen, J. H. 2012: Simulating melt, runoff and refreezing on Nordenskiöldbreen, Svalbard, using a coupled snow and energy balance model. *The Cryosphere 6*, 641-659.
Warren, C. R. 1992: Iceberg calving and the glacioclimatic record. *Progress in physical geography 16*, 253-282.

Waters, C. N., Zalasiewicz, J., Summerhayes, C., Barnosky, A. D., Poirier, C., Gałuszka, A., Cearreta, A., Edgeworth, M., Ellis, E. C., Ellis, M., Jeandel, C., Leinfelder, R., McNeill, J. R., Richter, D. d., Steffen, W., Syvitski, J., Vidas, D., Wagreich, M., Williams, M., Zhisheng, A., Grinevald, J., Odada, E., Oreskes, N. & Wolfe, A. P. 2016: The Anthropocene is functionally and stratigraphically distinct from the Holocene. *science 351*.

Werner, A. 1993: Holocene moraine chronology, Spitsbergen, Svalbard: lichenometric evidence for multiple Neoglacial advances in the Arctic. *The Holocene 3*, 128-137.

Wiles, G. C., Calkin, P. E. & Post, A. 1995: Glacier fluctuations in the Kenai Fjords, Alaska, USA: an evaluation of controls on iceberg-calving glaciers. *Arctic and Alpine Research*, 234-245.

Wolfe, A. P., Hobbs, W. O., Birks, H. H., Briner, J. P., Holmgren, S. U., Ingólfsson, Ó., Kaushal, S. S., Miller, G. H., Pagani, M. & Saros, J. E. 2013: Stratigraphic expressions of the Holocene–Anthropocene transition revealed in sediments from remote lakes. *Earth-Science Reviews 116*, 17-34.

Wordie, J. M. 1921: Present-day conditions in Spitsbergen. *The Geographical Journal 58*, 25-45.

Wright, R. & Anderson, J. B. 1982: The importance of sediment gravity flow to sediment transport and sorting in a glacial marine environment: Eastern Weddell Sea, Antarctica. *Geological Society of America Bulletin 93*, 951-963.

Zilliacus, H. 1989: Genesis of De Geer moraines in Finland. *Sedimentary Geology 62*, 309-317.

Zinger, Y. M., et al 1965: Issledovanieye lednikov Shpitsbergena sovetskoy ekspeditsiyey letom 1965 goda [Study of Spitsbergen glaciers by the Soviet expedition in the summer of 1965]. *Materialy Glyatsiokologicheskikh. Kronika. Obsuzhdeniya. 12*, 13.

9. Appendix

A digital full-size version of the geomorphological map is available at:

https://www.dropbox.com/sh/gsegmhdkmfpknhz/AAB3arD2OZv1lPh42n_jJk7Na?dl=0