Faculty of Science and Technology

Glacial history and Geomorphology of Trygghamna, western Svalbard

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Master thesis in Quaternary Geology (GEO-3900)
May 2017

Abstract

This study investigates Trygghamna, a small fjord on the western coast of Spitsbergen, Svalbard. In order to investigate its glacial history, high-resolution aerial images and swath bathymetry are used to produce a detailed geomorphological map of the area, with a focus on the Neoglacial extent of the three largest glaciers in the fjord, Protektor-, Harriet- and Kjerulfbreen. The landforms are classified into; subglacial, supraglacial, ice-marginal, glaciofluvial, proglacial landforms and extra-marginal surface cover. The ice-margins were reconstructed from ~1900, based on historical and geomorphological data together with aerial and satellite images.

The Neoglacial maximum extent of the glaciers was reached around 1900, or the culmination of the Little Ice Age in Svalbard. Harriet- and Kjerulfbreen are considered to have exhibited surge behavior based on the presence of crevasse squeeze ridges (CSRs), which are considered unique for surging glaciers. The maximum extent is therefore, to some degree, related to a surge-event. Ice-marginal reconstructions and historical data suggest that the glaciers have been in overall retreat since then. It is more complex to determine if Protektorbreen surged as different factors in the glacier's environment affect the formation and preservation potential of the landforms. The landform assemblages in the forefields do not show a good correspondence to previously published landsystem models for surge-type glaciers. Landsystem models should therefore be used with precaution when identifying undocumented surge-type glaciers. This investigation highlights the contrast in the record between terrestrial and marine environments of the glacier forefields in Trygghamna. Therefore it demonstrates the importance of incorporating evidence from both terrestrial and marine archives when reconstructing past glacial history, due to dynamic glacial behavior in different environments.

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1. Introduction

1.1 Motivation

Glaciers are important for paleoclimate reconstructions due to their sensitivity to changes in climate and the imprints they leave behind can further be used to reconstruct the glacial history (Oerlemans, 2005). Investigations of those glacial fingerprints can reveal information about both ice-marginal changes and past glacial dynamics (Evans, 2003). After the termination of the last major glacier advance marking the peak of the Little Ice Age (LIA) some 100 years ago in Svalbard, glaciers have undergone retreat and negative mass balance (Oerlemans, 2005; Nuth et al., 2013), exposing extensive areas of glaciated landscape (Ingólfsson, 2011). The Arctic region is extremely sensitive to changes in climate and experiences changes of greater amplitude than regions at lower latitudes (IPCC et al., 2007; Miller et al., 2010). Therefore, investigations have focused on glaciers in high Arctic settings and their relation with climate to be able to predict how they will respond to future changes.

One of the challenges when reconstructing glacial history and dynamics are surge-type glaciers, which are glaciers that experience cyclic flow alternating between slow and rapid flow on a timescale of a few years to several decades. Glacier surges have received considerable attention since generally described by Meier and Post (1969) and first review of surging glaciers in Svalbard by Liesøl (1969). Even though the causes and mechanisms behind surges are still debated, they are considered to have a non-direct respond to climate and therefore they are not suitable for climate reconstruction. This complicates the direct correlation between glacier advances and changes in temperature and precipitation. It is therefore important to identify surge-type glaciers and understand the reason behind individual glacier advances (Meier and Post, 1969; Lefauconnier and Hagen, 1991; Yde and Paasche, 2010; Farnsworth et al., 2016).

Surge-type glaciers often produce diagnostic glaciological (Copland et al., 2003) and geomorphological features, both in marine and terrestrial environment. If the surge-event was not documented with historical data and glaciological data is absent, landsystem models have been developed to identify antecedent surge events (Sharp, 1988a; Evans and Rea, 1999, 2003; Ottesen et al., 2008). In the glacial environments there are, however, several processes affecting the formation and preservation potential of landforms. Investigations in variable glacial settings have therefore led to the development of modified landsystem models that demonstrate different assemblage than in the original ones (Brynjólfsson et al., 2012, 2014; Schomacker et al., 2014; Brynjólfsson, 2015; Flink et al., 2015).

Surge-type glaciers are also important for the understanding of glacial dynamics (Murray et al., 2003). The landforms formed by surge-type glaciers contain information about the processes occurring at the ice-bed interface (Kjær et al., 2006; Ottesen et al., 2008; Ingólfsson et al., 2016) and give an insight into the interaction between climate and glacier dynamics (Evans and Rea, 1999, 2003; Brynjólfsson, 2015). Surging glaciers have also been considered a modern analogue to terrestrial paleo-ice streams and surging ice-sheet lobes (Evans and Rea, 1999, 2003; Ottesen et al., 2008; Schomacker et al., 2014).

1.2 Aims and approach

The aim of this thesis is to use both high-resolution aerial images and swath bathymetry to map the terrestrial and marine forefields of glaciers in Trygghamna, a fjord on western Svalbard. The terrestrial part of the map was further verified with fieldwork. In addition to investigating the surface cover, the composition of key landforms and sub-bottom profiles from the bathymetry were examined to be able to determine their genesis. Ice-marginal change was reconstructed based on historical data; old maps and photographs together with aerial and satellite images.

The evidence for the maximum Neoglacial extent of glaciers in Trygghamna and their subsequent retreat is discussed based on the results of the geomorphological mapping and ice-marginal reconstructions. The pre-Neoglacial history is fragmentary in the geomorphology but sub-bottom profiles give some insight. The geomorphological and historical data is further used to highlight past glacial dynamics and thermal regimes and to test whether or not the glaciers have surged in the past.

This research sets the stage for discussing the differences and similarities between the geomorphology of the terrestrial and marine forefields; which controls do ice dynamics and processes in the proglacial and ice-marginal environment have on formation of landforms and their preservation potential. The landform assemblage in Trygghamna is compared to previously published landsystem models and the use of landsystem models will be discussed and how applicable it is to use them to infer past glacial dynamics.

1.3 Surge-type glaciers

Surge-type glaciers are distinguishable from non-surging ones by quasi-periodic oscillation of increased flow during the active phase, which lasts from several months to few years, followed by slower flow rates during the quiescent phase, which last tens to hundreds of years. Maximum velocities during the active phase are usually one to two magnitudes greater than during the quiescent phase (Meier and Post, 1969; Thorarinsson, 1969; Kamb et al., 1985; Raymond, 1987; Sharp, 1988b; Murray et al., 2003). During the quiescent phase ice builds up in a reservoir area in the upper part of the glacier. Due to difference in the mass in the upper and lower part after a surge the ice surface gradient increases until the initiation of the active phase. Once that stage is reached the mass is transferred down to the receiving area resulting in a rapid ice lowering in the reservoir area. Simultaneously the receiving area thickens and often results in a significant advance of the ice margin (Sund et al., 2009; Benn and Evans, 2010) (Figure 1, 2B). That is albeit not always the case and surges, usually in tidewater glaciers, with no advances of the glacier terminii have been recorded (Sund et al., 2009, 2014). The surge usually initiates up glacier and propagates downwards but examples of it propagating upwards have also been recorded (Murray et al., 2003; Benn and Evans, 2010).
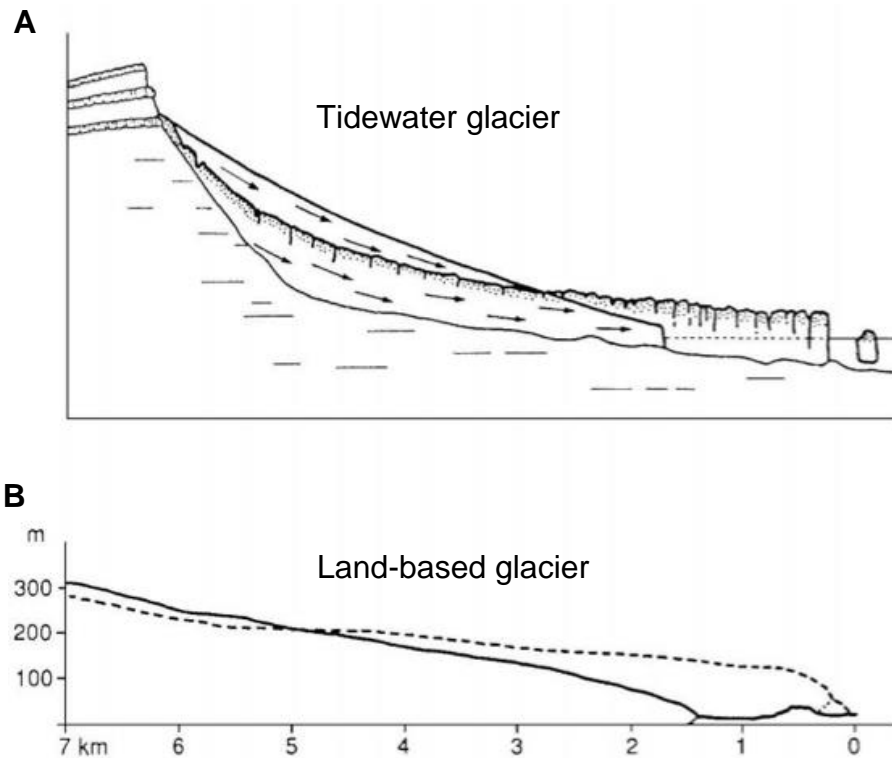
Figure 1. *A schematic illustration of the geometric changes of glaciers during a surge where the reservoir area is lowered and simultaneously mass is moved to the receiving area. A) Tidewater glacier, the irregular line marks the surface after a surge (Liestøl, 1989). B) Land-based glacier, the dashed line marks the surface after a surge but the whole before (Hagen 1987, modified by Lefauconnier and Hagen, 1991).*

1.3.1 Mechanism

Initially the reason for surging was considered related to internally driven oscillations in conditions at the glacier bed rather than external forcing (Meier and Post, 1969; Sharp, 1988b). However, the triggering mechanism for a surge is yet not fully understood. The *hydrologic switch* model was developed to explain the changes in the basal hydrology often observed during surges, especially for temperate glaciers. The drainage changes from efficiently subglacial conduits system during the quiescent phase to an inefficient linked cavity system during the active phase leading to higher storages of water during the surges that increases the basal sliding (Kamb et al., 1985; Raymond, 1987). The switching between thermal regimes at the glacier bed is considered more valid for polythermal glaciers. The bed is cold in the quiescent phase while ice builds up in the reservoir area. The thicker and steeper ice will warm the basal ice up to the pressure-melting point. The warmer ice will lead to increased melting and faster motion resulting in a surge-event. The glacier gets thinner and the glacier bed converts back to being cold based (Fowler et al., 2001; Sevestre et al., 2015). These different theories have been combined by the theory of enthalpy cycling. For the glacier to maintain a steady-state the enthalpy gain from glacier flux needs to be evened out with heat conduction and/or meltwater discharge. If rates of energy accumulation and energy dissipation are unequal the glacier behavior will become unsteady. According to this model, surges are the result of dynamic instabilities but relate to environmental forcing (Sevestre and

Benn, 2015; Sevestre et al., 2015). It is as well considered that glaciers can alter from surging to non-surging glaciers due to changes in climate, which can affect mass accumulation in the reservoir area and the thermal regime. This shift has been shown with examples of smaller glaciers that surged during the LIA but are cold based today. With the ongoing negative mass balance they are not considered to surge again in the near future under the present climate (Dowdeswell et al., 1995; Sevestre et al., 2015).

1.3.2 Distribution
Surging glaciers tend to cluster in certain areas and are most common to occur within a climatic envelope of mean annual temperature of ca. 0-10°C and annual precipitation of 200-2000 mm. Major clusters occur in Svalbard, Alaska-Yukon, Arctic Canada, parts of West Greenland, Iceland, Novaya Zemlya and number of mountain ranges in central Asia (Sevestre and Benn, 2015). Glacier geometry has also been investigated in terms of their geographic distribution. They generally tend to be longer and with larger areas than non-surging glaciers, although this is inconsistent between regions (Clarke et al., 1986; Sevestre and Benn, 2015). Glacier surges have been described from variable environments; land terminating outlet, cirque and valley glaciers and tidewater glaciers. Both temperate and polythermal glaciers can surge (Murray et al., 2003).

1.3.3 Surge-type glaciers in Svalbard
The number of glaciers in Svalbard exhibiting surge behavior is still unknown but estimates range from 13-90% of all glaciers (Jiskoot et al., 1998; Lefauconnier and Hagen, 1991). Recent work favors that their density is high but the number is still not clear (Sund et al., 2009; Farnsworth et al., 2016). During the LIA, surges are also thought to have been much more common than at present, which complicates determining their quantity (Liestøl, 1969; Dowdeswell et al., 1995; Sevestre et al., 2015; Lovell and Boston, 2017).
Surges in Svalbard differ from most other regions. The duration of the active phase is considered longer in Svalbard, 4-10 years, compared to other locations in the world where it is usually only 1-3 years. A typical surge-cycle is around 20-40 years in most areas (Benn and Evans, 2010). The surge cycles are generally longer in Svalbard and can be up to several hundred years (Dowdeswell et al., 1991; Benn and Evans, 2010). Shorter cycles do occur and until recently it was considered to be down to 40 years at Tunabreen (Flink et al., 2015). However, monitoring of Tunabreen shows that its velocities increased during the winter of 2016, indicating that a surge might be initiating 15 years earlier than expected from its surge cycle (Borstad, 2017). This highlights how dynamic the surge-type glaciers are and the importance of future monitoring of them to increase our understanding. The ice velocity is lower in Svalbard, ranging from 1.3 to 16 m day^{-1} (Figure 2E) compared to 40-60 m day^{-1} for Variegated glacier, Alaska, (Kamb et al., 1985) and up to 120 m day^{-1} in Brúarjökull, Iceland (Thorarinsson, 1969). Surge-type glaciers in Svalbard tend to be longer and with a steeper surface gradient and are most likely to be situated on sedimentary rocks (Hamilton and Dowdeswell, 1996; Jiskoot et al., 2000).

1.3.4 Observations
Surge events have been recorded in the historical data by observations of changes at the ice front and rapid advances (Meier and Post, 1969; Liestøl, 1969, 1988; Kamb et al., 1985;

Hagen et al., 1993; Björnsson et al., 2003). Remoteness of areas, such as Svalbard, does make direct observations of surge events relatively rare (Dowdeswell et al., 1991; Lønne, 2016). During a surge and shortly after it, glaciers can easily be recognized by glaciological evidence such as heavily crevassed surfaces, increased velocities during the surge and advanced terminii (the latter, in some cases) (Figure 2A-E). Looped moraines are among the most obvious evidence and form when a tributary glacier surges into the trunk glacier during its quiescent phase (Copland et al., 2003) (Figure 2B). Geometric changes have also been observed on the glacier surface (Sund et al., 2009) (Figure 1, 2D). These features can be examined by historical data; old photographs and documentation, aerial and satellite images. After these features have vanished the identification of surge-type glaciers becomes more problematic. The identification of past surges based on glaciological features can be complex as surge behavior varies greatly between glaciers (Meier and Post, 1969; Murray et al., 2003) and the features can also appear on non-surging glaciers, often associated with fast flowing tidewater glaciers (Copland et al., 2003). Surge-type glaciers also produce diagnostic landform assemblages but the most widely used landform to identify them are crevasse squeeze ridges (CSRs). They and the landsystem models are presented in the following sections (*Section 1.3.5 & 1.4).

1.3.5 Crevasse squeeze ridges (CSRs)
Crevasse squeeze ridges (CSRs) are considered a unique landform for glaciers that exhibit surge behavior (Figure 3A, B). They have been identified from a number of terrestrial glacier forefields over the years (Evans and Rea, 1999; 2003; Rea and Evans, 2011; Kjær et al., 2008; Schomacker et al., 2014). The first detailed study was conducted by Sharp (1985a, b) at Eyjabakkajökull, Iceland. He described them as 1-2 m high ridges that can be up to several hundred meters long. The ridges could be traced into the down-wasting glacier front, where they occurred as dykes of debris in crevasses. The ridges were aligned in a cross-cutting pattern that corresponded well to the radial and transverse crevasses observed on the glacier. The ridges are composed of matrix-supported diamict and often occur on fluted till plains. Cross-cutting ridges have also been observed in the submarine setting in front of surging tidewater glaciers (Figure 3C, D). They are often referred to as rhombohedral or geometrical ridge network (Ottesen and Dowdeswell, 2006; Ottesen et al., 2008; Flink et al., 2015; Lovell et al., 2015). To avoid confusion, they will be termed CSRs in both settings in this study.
The formation of CSRs is still discussed (Rea and Evans, 2011; Ingólfsson et al., 2016) but they are generally thought to be formed by upward infilling of saturated sediments into basal crevasses that form in association with the longitudinal and extensional stress during the surge. The process happens towards the termination of the surge and subsequent meltout (Lovell et al., 2015). Several models have been developed to explain their formation and deposition (Sharp, 1985a; Bjarnadóttir, 2007, Lovell et al., 2015). Ben-Yehoshua (2017) created a conceptual model based on the appearance of CSRs on different subsurface in Trygghamna.
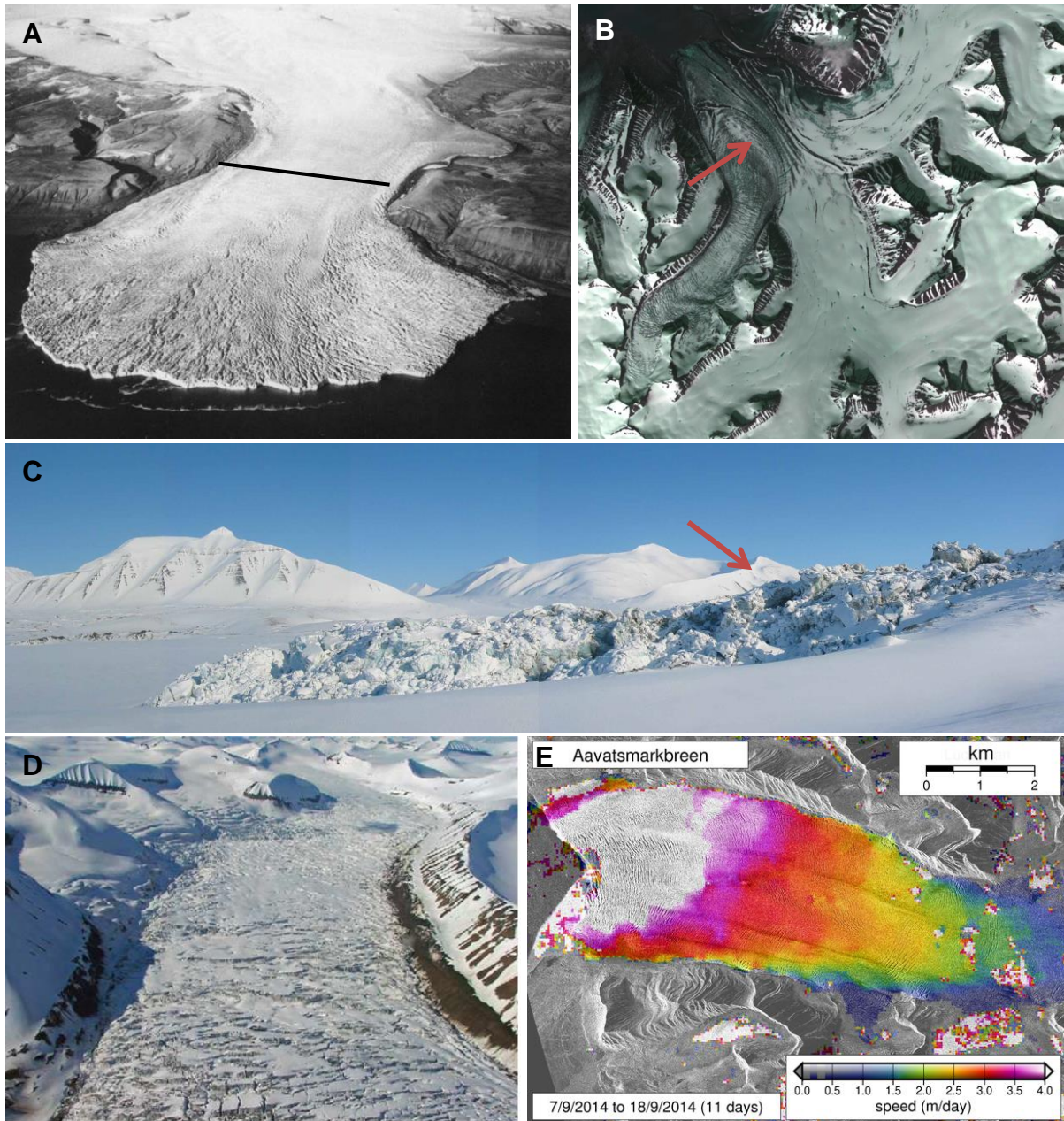
Figure 2. *Glaciological features formed by surging. A) Advance in glacier terminii of Freemanbreen, Svalbard. The line indicates the approximate position prior to the surge (Hagen et al., 1993). B) A looped moraine (arrow) formed when tributary glacier, Skobreen, surged into Paulabreen on Svalbard (NPI, 2005). C) The surging Paulabreen advanced into the fjord, forming a surge bulge (arrow) (photo: D. Benn, 2005). D) Increased crevassing of the glacier surface and the draw down of the reservoir area during the surge in Skobreen, Svalbard (photo: L. Kristensen, 2005). E) Higher velocities, up to 4.5 m day^{-1} were detected during the surge of Aavatsmarkbreen in 2014 (A. Luckman, 2015).*
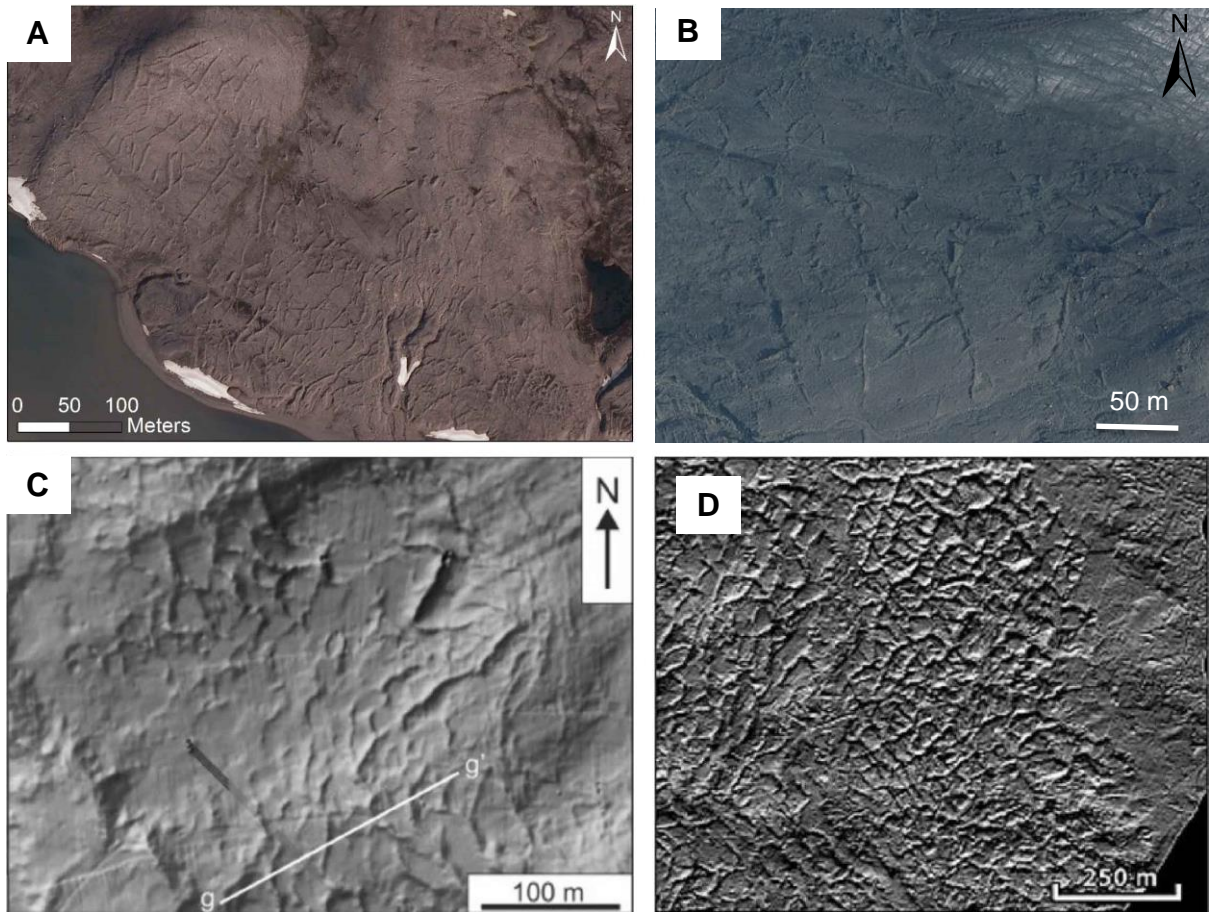
Figure 3. Examples of CSRs from remote sensing. A) Terrestrial network in front of Kjerulfbreen, Trygghamna (the study area for this thesis) (NPI, 2009). B) Terrestrial ridges in front of Aavatsmarkbreen, western Spitsbergen (NPI, 2009). C) Subaqueous ridges in the forefield of Tunabreen, Svalbard (Flink et al., 2015). D) Subaqueous ridges in Yoldibukta, Svalbard (Ottesen and Dowdeswell, 2006).

1.4 Landsystem models

Sediment and landforms are usually produced in assemblages, reflecting the range of processes in certain environments. Investigations on different spatial scales are a useful tool on how sediments, landforms and landsystems are connected and the effects the landscape and glacial dynamics have on the depositional processes (Benn and Evans, 2010). To illustrate this, conceptual landsystem models have been developed for variable glacial settings. They are considered a useful tool for the reconstruction of former glacier environments and glacial dynamics. However, they are based on environments with only one type of glacier and are therefore simplified models as glaciers often occur in more complex systems (Evans, 2003). The development of the landscape will further change with time. High-resolution surveys on recently deglaciated forefields are therefore of a high importance to monitor the landscapes (Schomacker and Kjær, 2008).

Expanding research on glaciers in variable environments and with different glacial dynamics has demonstrated that different factors in the glacier's environment affect the landform assemblage. Modified landsystem models have thus been developed for surge-type glaciers in different environments, both terrestrial and marine, to demonstrate their variability (Ottesen and Dowdeswell, 2006; Ottesen et al., 2008; Schomacker et al. 2014; Brynjólfsson et al., 2012, 2014; Brynjólfsson, 2015; Flink et al., 2014). In the following sections a short summary of landsystem models is presented which are considered appropriate for this investigation.

1.4.1 Terrestrial glaciers

Polythermal glaciers
Glasser and Hambrey (2003) presented a landsystem model for polythermal terrestrial glaciers in maritime high-arctic setting (Figure 4). The model is based on Midtre and Austre Lovénbreen in western Svalbard. Midtre Lovénbreen may have surged in the early 19th century (Liestøl, 1988; Hansen, 2003), although this is considered unlikely by Hambrey et al. (2005). This demonstrates the complications utilizing landsystem models to identify glacier types as will be discussed later. The landform assemblage associated with surge-type glaciers are described separately in the following sections.

Three zones with characteristic landforms have been identified in their forefields of polythermal glaciers. An outer moraine ridge marks the termination of the forefield. The ridge is usually high, with arcuate shape and ice-cored. On its proximal side are so called moraine-mound complexes that consist of aligned hummocks or mounds with variable morphology. Supraglacial debris stripes often drape the area. The inner zone lies between the moraines and the glaciers front and consists of diamict with variable quantities of supraglacial derived debris stripes, geometrical ridge networks, streamlined ridges/flutes and foliation-parallel ridges. Proglacial streams commonly erodes the landscape.
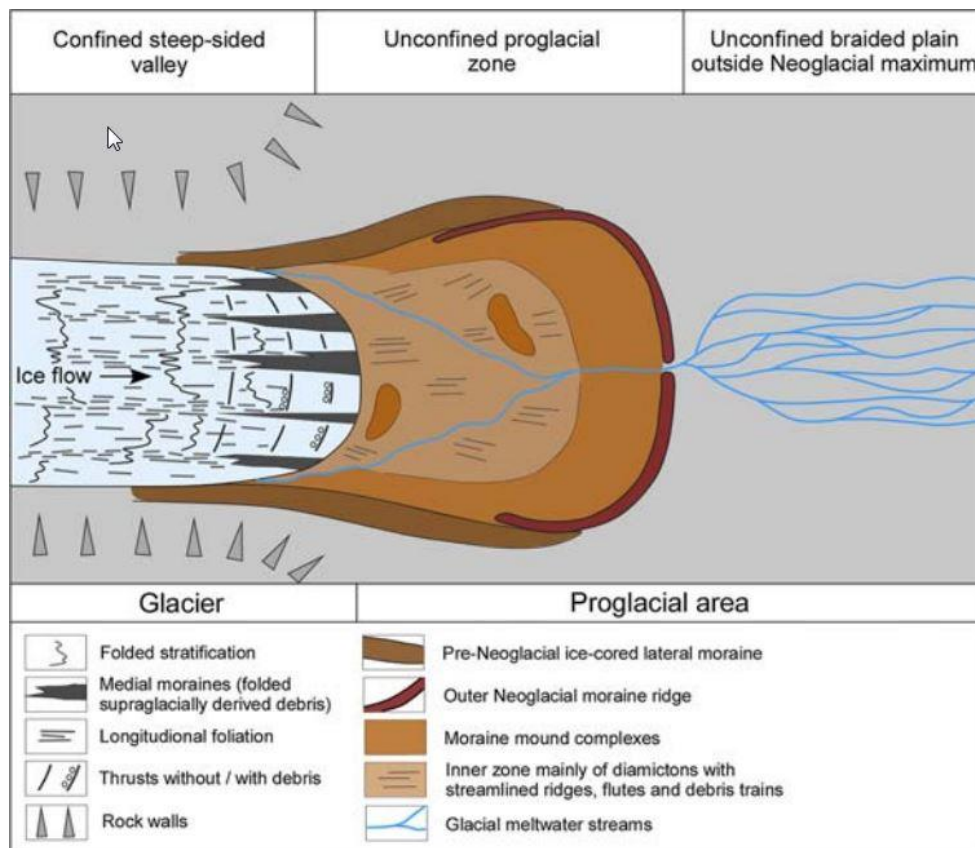
Figure 4. A landsystem model for terrestrial Svalbard polythermal glacier (from Glasser and Hambrey (2003) and modified by Ingólfsson (2011)).

Surge-type glaciers

Two conceptual surge-type landsystem models have been proposed for nontopographically constrained terrestrial outlet glaciers based on geomorphological, sedimentological and remote sensing data (Evans and Rea, 1999, 2003; Schomacker et al., 2014) (Figure 5). Evans and Rea (1999, 2003) investigated glaciers in Svalbard, Iceland and Alaska and identified three overlapping zones, each consisting of characteristic landform assemblage. Schomacker et al. (2014) based their model on Eyjabakkajökull, a frequently surging glacier in Iceland, which agrees well with the earlier Evans and Rea model, except for some modifications of the locations of landforms.

The outer zone (A) consists of glaciotectonic end moraine, consisting of deformed pre-surge sediments. Blow-out structures are often present at the distal side. Hummocky moraine, usually comprised of till, is associated with the end moraine and located at its proximal side. An active, channelized outwash plain and inactive, pitted outwash deposited on stagnant ice make up the intermediate zone (B). Overridden end moraines and patches of hummocky moraine and till plain can also be present. The inner zone (C) consists of subglacial till, flutes, drumlins, CSRs and concertina eskers (Evans and Rea, 1999, 2003; Schomacker et al., 2014).
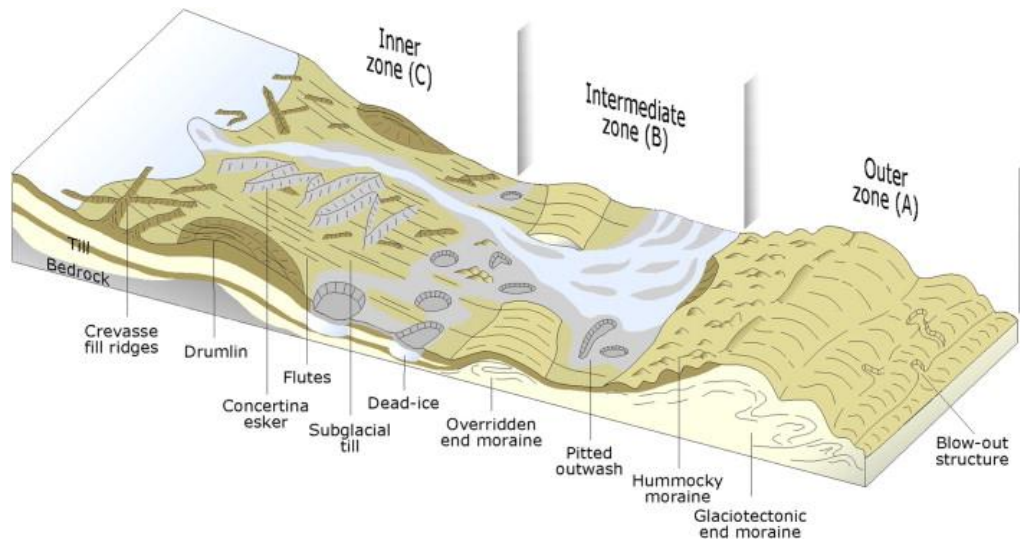
Figure 5. A landsystem model of non-topographically constrained terrestrial outlet glacier based on observations from Eyjabakkajökull (Schomacker et al., 2014, partly based on Evans and Rea, 1999, 2003).

Surge-type glaciers in Drangajökull, northwest Iceland, are confined within valleys and differentiate thus from the environment described above. Brynjólfsson et al. (2014) and Brynjólfsson (2015) developed landsystem model for topographically constrained outlet glaciers (Figure 6). The model is based on detailed geomorphological mapping of three surge-type glaciers in Drangajökull. The sediment cover in the area is generally thin and contains coarse grained basal till and locally weathered bedrock. Ice molded bedrock outcrops are common. Extensive sandurs cover the valleys, therefore the fluvial erosion is high and the preservation potential of landforms low. End moraines mark the maximum extent of the surges and are frequently eroded and fragmented. Flutes also occur but eskers, kame terraces, pitted outwash and hummocky moraines are rarely occurring. CSRs and concertina eskers have not been observed in the forefields.
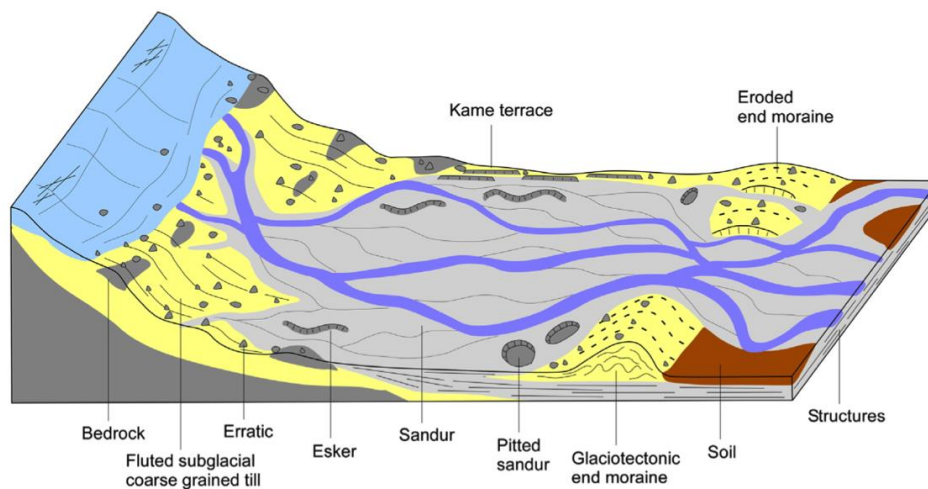
Figure 6. A landsystem model of surge-type valley glacier (strongly topographically confined) based on observations from the Drangajökull outlets (Brynjólfsson, 2015).

Brynjólfsson et al. (2012) developed a landsystem model for surge-type cirque glaciers in north Iceland from Búrfellsjökull and Teigarjökull, on Tröllaskagi peninsula (Figure 7) based on geomorphological, geological and remote sensing data from landforms and sediments. Due to the alpine landscape around the glaciers the sediments are usually coarse grained, angular to sub-angular, interpreted to be supra- and/or englacial material originated from rockfalls from steep mountain sides surrounding the glaciers. Few moraines occur in front of both glaciers, each connected to the maximum extent of a certain surge. They are generally relatively small, irregular and asymmetric. Hummocky moraines are common at the proximal slope of the end moraines. Their surface consists of coarse debris of angular cobbles and boulders. Buried dead-ice and signs of active melting is common in them as well as at other locations in the forefields. Small-scale CSRs are present in the model, consisting of coarse grained diamict. Few, poorly preserved flutes occur in the forefield. Their dimension is smaller than in other settings. Low amplitude ridges, extending from the ablation zone onto the glacier forefield have been interpreted as medial moraines.
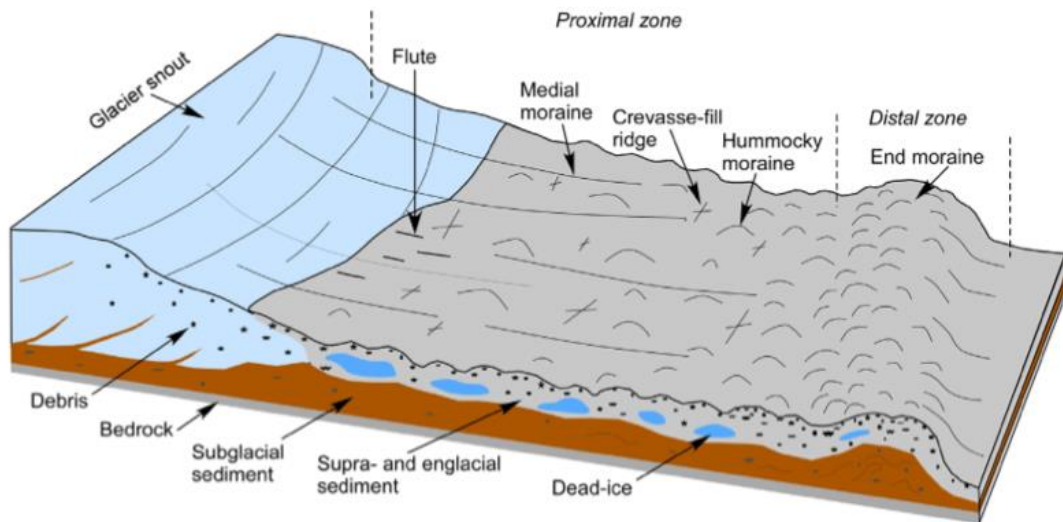
Figure 7. A landsystem model of surge-type valley glacier (strongly topographically confined) based on observation from Tröllaskagi, north Iceland (Brynjólfsson et al., 2012).

1.4.2 Tidewater glaciers

A number of models have been proposed for tidewater glaciers, surging and non-surging. Landsystem models have been developed for the submarine geomorphology based on glaciers in Svalbard, using marine-geophysical evidence (Figure 8B). Ottesen and Dowdeswell (2006) and Ottesen et al. (2008) describe landforms from both known and indented surge-type glaciers and their relative age relationship. Flink et al. (2015) presented a model for glaciers that undergo surges frequently where the landforms could be directly linked to the surge event. The assemblage is rather similar but the model by Flink et al (2015) is more complex and the landforms modified and overprinted due to the multiple surges. The landforms described are presented in the order of their formation based on their cross-cutting relationship (Ottesen and Dowdeswell, 2006; Ottesen et al., 2008). The outer fjord consists of

terminal moraine with glacigenic debris flow at the distal side. Glacial lineations, CSRs, eskers and retreat moraines are located at the inner fjord. In the model by Flink et al. (2015) the esker is absent but several terminal moraines are present due to the multiple surges and overridden moraine.

Plassen et al. (2004) developed a landsystem model for sedimentation of tidewater glaciers in high Arctic settings based on both surge-type and non-surging glaciers in Svalbard (Figure 8A). The morphology is described as well as the sediment packages. The model is similar to the surge-type model except CSRs, streamlined glacial lineations and eskers are not present. The sediment descriptions include glacimarine infill sediments between the transverse ridges and distal stratified glacimarine sediments occur distal to the debris lobes and sometimes extend above and beneath them. The distal sediments are usually composed of stratified clayey silt with low amount of ice-rafted debris (IRD).
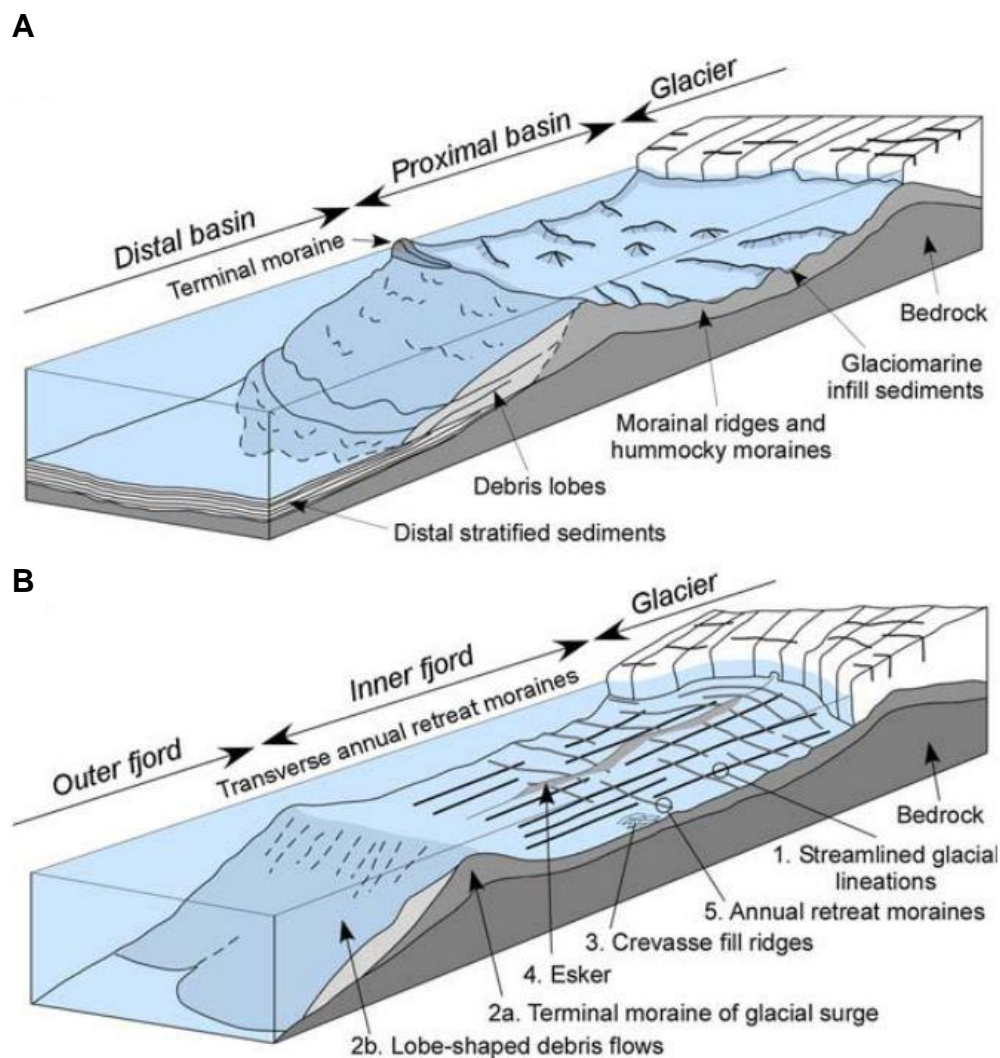
Figure 8. Tidewater glaciers. A) A landsystem model for proglacial sedimentation in front of polythermal tidewater glaciers in Svalbard (Plassen et al, 2004). B) Landsystem model for landforms in front of surge-type tidewater glaciers. (Ottesen et al., 2008, partly based on Ottesen and Dowdeswell, 2006). (Modified by Ingólfsson, 2011).

2. Setting

2.1 Regional setting of Svalbard

2.1.1 Physical geography

The Svalbard archipelago is located on the north-western corner of the Barents Sea Shelf, between 74°-81°N and 10°-35°W (Figure 9). The archipelago consists of a large number of islands, covering in total 60.667 km^2. The main islands are Spitsbergen, Nordaustlandet, Barentsøya, Edgeøya, Kongs Karls Land, Prins Karls Forland and Bjørnøya, the largest one being Spitsbergen (Dallmann, 2015). The archipelago is highly glaciated and in the late 2000s the glacier coverage was ca. 57% of the land area or 33 775 km^2 (Nuth et al., 2013). The distribution of the glaciers is controlled by a combination of topography and climate and is most extensive on the western and eastern coast of Spitsbergen (Dallmann, 2015). The equilibrium-line altitude (ELA) increases on a transect from west to east over Spitsbergen, reflecting the distribution of precipitation (Hagen et al., 1993; Humlum, 2002) (Figure 10). Many different types of glaciers can be found, cirque and valley glaciers and ice-caps and ice-fields. Tidewater glaciers drain two-thirds of the glaciated area but land-terminating glaciers are also frequent (Nuth et al., 2013; Dallmann, 2015). Most of the glaciers are polythermal, containing a mixture of cold and temperate ice. They are temperate in the accumulation area and deep down in the ablation area but the margins of the glacier and the upper parts of the ablation area are cold-based (Dowdeswell, 1984; Petterson, 2004). Smaller glaciers tend to be entirely cold-based because their entire mass is below pressure-melting point (Hagen et al., 1993). Svalbard's glaciers are also well known for their surging behavior (Sevestre and Benn, 2015) although their number is debated (Hagen et al., 1993; Jiskoot et al., 2000).

The landscape and fjords in Svalbard are most recently shaped by Pleistocene glaciations in combination with bedrock geology. Due to weathering, fluctuations of the glaciers front and permafrost the landscape has been modified significantly during the Holocene (Dallmann, 2015). Fingerprints from the Quaternary glaciations, both in terrestrial and marine environment, can be used to reconstruct the glacial history (Ingólfsson, 2011). Several glacial and periglacial landforms related to permafrost are widespread, such as rock glaciers, pingos, ice wedges and patterned ground. Permafrost is continuous on Svalbard, and can be found everywhere outside the glaciers, and its thickness ranges from <100 m near the coast up to >500 m in the mountains (Humlum et al., 2003; Dallmann, 2015). Periglacial processes can alter landforms and surface over time due to solifluction and melting of dead-ice (Schomacker and Kjær, 2008). Ongoing warming will lead to increase in the annual thaw (active layer) (Humlum et al, 2003).
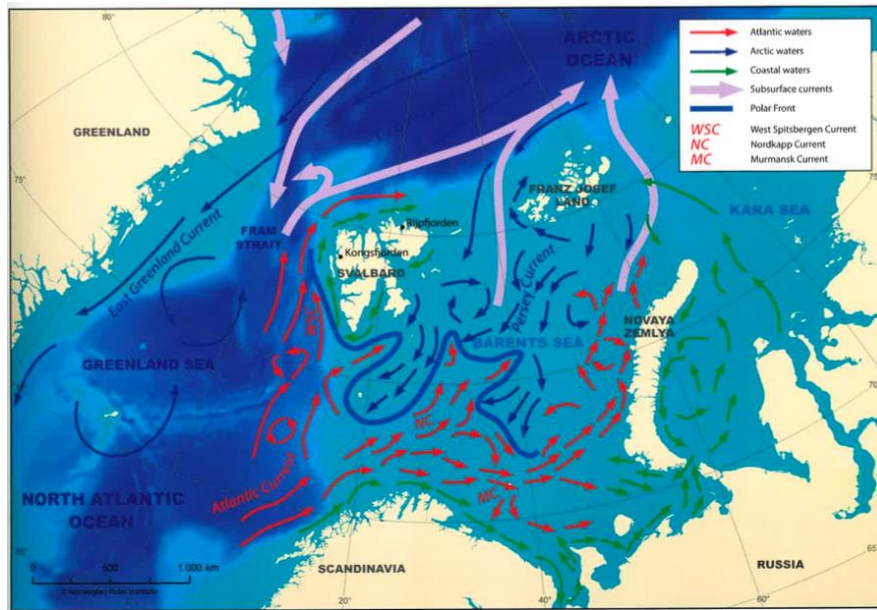
Figure 9. *Location of Svalbard archipelago on the north-western corner of the Barents Sea Shelf and the ocean currents surrounding it. The West Spitsbergen Current (WSC) brings warm water up to the west coast but the Persey Current delivers cold polar water to the east (Dallmann, 2015).*
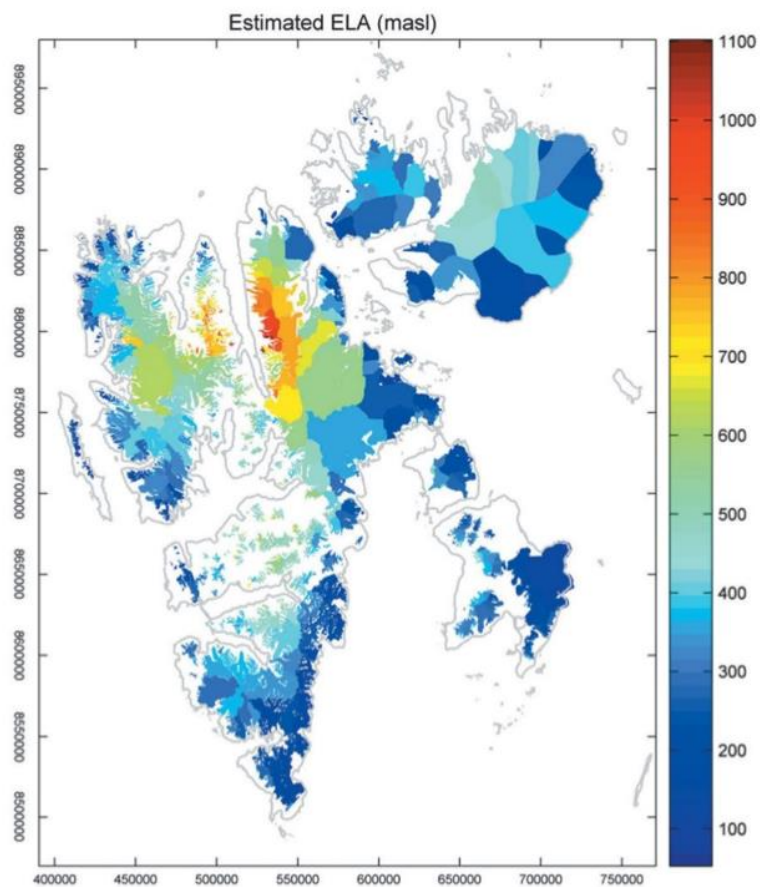
Figure 10. *The Svalbard archipelago is heavily glaciated or ~57%. The different colors indicate the equilibrium line altitude (ELA), increasing on a transect from west to east (Köning et al., 2014).*

2.1.2 Climate and Oceanography

Svalbard is extremely sensitive to changes in the climate due to its location. The archipelago is situated in the main transport path for air masses and ocean currents into the Arctic basin, which explains the relatively mild climate (Hanssen-Bauer et al., 1990; Dickson et al., 2000). The West Spitsbergen current (WSC) brings warm and high-salinity water from the Gulf Stream up to the west coast, creating the northernmost open water. The eastern coast is dominated by cold, fresh polar water from the Arctic basin (Figure 9). The two large-scale air currents over the North Atlantic mainly control the climate in Svalbard with the low pressure system over Iceland and high pressure system over Greenland and the Arctic Ocean. They result in mild air transported towards Svalbard from lower latitudes. Large temperature differences can occur between them and their variation causes great fluctuations in the climate (Steffensen, 1982; Hanssen-Bauer et al., 1990).

The Arctic land areas have recently experienced stronger warming, compared to other regions on Earth (IPCC, 2007) due to positive feedbacks such as changes in sea ice and snow cover (Miller et al., 2010; Serreze and Barry, 2014). The feedbacks amplifiy the surface air temperature response to climate forcing, therefore termed the "Arctic amplification". Paleoclimate reconstructions in the Arctic indicate that climate shifts occurred with greater changes during the Holocene as well (Hald et al., 2007; Miller et al., 2010).

Meteorological observations began in 1911 on Svalbard, with the first permanent weather station in Green Harbour (Finneset in Grønfjorden). The number of stations increased and they were relocated over the years due to large climate gradients over the archipelago (Nordli, 2010). Incomplete measurements have been carried out at Svalbard airport and Longyearbyen from 1911 until today and measurements from Isfjord Radio exist between 1934-1976, except from 1941-1946 when it was destroyed during World War II (Steffensen, 1982; Hansen-Bauer et al., 1990; Nordli, 2010; Førland et al., 2011) (Figure 12A). Combining several series and homogenizing them, a composite temperature series for Svalbard airport/Longyearbyen has been developed (Nordli, 2010) (Figure 15). The climate on Svalbard is classified as polar tundra climate, where at least one month has a mean air temperature above 0°C. Areas close to the outer coast on western Svalbard have stronger maritime influence but the climate is dryer towards the central part (Hanssen-Bauer et al., 1990). Isfjord Radio is considered to have maritime climate (Serreze and Barry, 2014). Winter temperature is therefore generally lower (2-5°C) at Svalbard airport than Isfjord Radio but summer temperature higher (1-2°C) (Hanssen-Bauer et al., 1990).

2.1.3 Quaternary glaciations

The Svalbard-Barents Sea Ice sheet has repeatedly grown and disintegrated over the last 2.6 MY, reflected by large trough–mouth fans on the western and northern shelf and glaciomarine sediments. However, the preservation potential increases with younger glaciations and therefore the reconstruction of the Late Weichselian ice-sheet is more complete (Svendsen et al., 2004; Vorren et al., 2011). The ice sheet reached the shelf edge during the Last Glacial Maximum around 24 ka BP (Landvik et al., 1998; Jessen et al., 2010), with fast flowing ice-streams draining the ice-sheet trough fjords and less active ice covering the inter-fjord areas (Landvik et al., 2005; Ottesen et al., 2007; Landvik et al., 2014). The ice-marginal retreat initiated from the outer shelf around the western margin of Svalbard ca. 20.5

ka BP. (Jessen et al., 2010). Subsequently the ice sheet began to thin (Hormes et al., 2013). The deglaciation continued and between 15 and 12 ka BP. the ice sheet retreated rapidly from the western shelf towards the fjord mouths and the present coastline (Jessen et al., 2010; Hormes et al., 2013) (Figure 11). The deglaciation continued from the major fjords into the inter-troughs and terrestrial environment (Hormes et al., 2013; Ingólfsson and Landvik, 2013; Landvik et al., 2014).
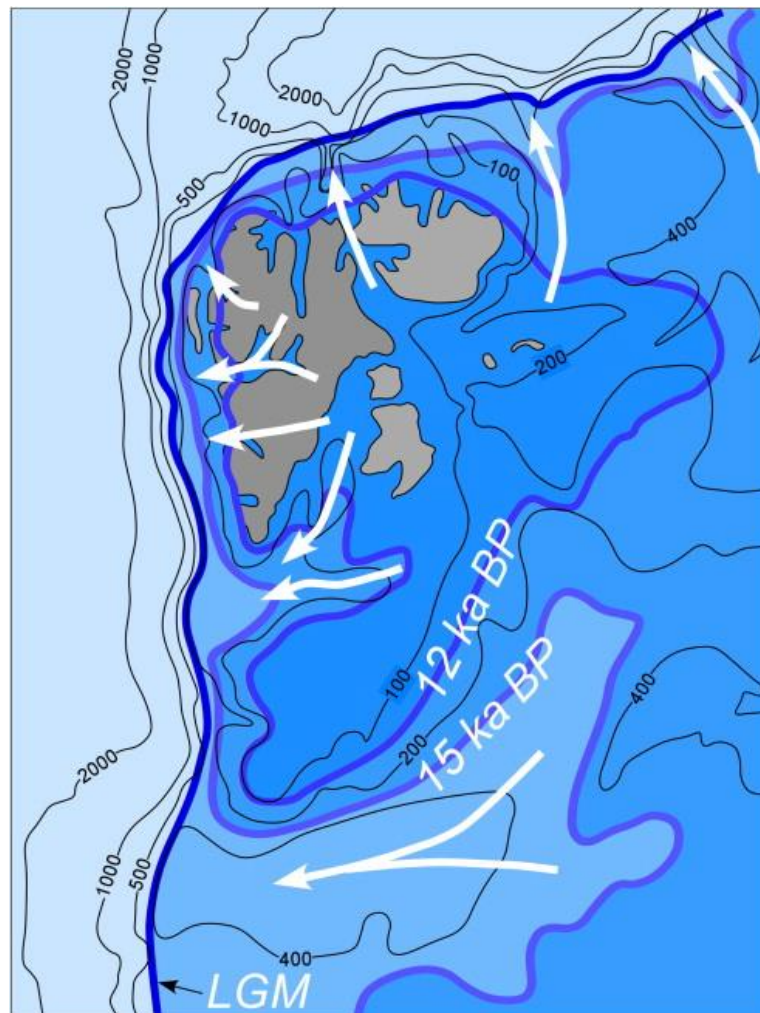
Figure 11. *Reconstruction of the margins of the Svalbard-Barents Sea ice sheet during the Late Weichselian from LGM and trough the deglaciation, at 15 and 12 ka BP. White arrows are conceptual flow lines (ice-streams) based on geomorphological evidences (from Landvik et al., 1998, modified by Ingólfsson and Landvik et al., 2013).*

The marine and terrestrial archives can both be used to reconstruct the extent and dynamics of past glaciations and deglaciations (Ingólfsson, 2011). Streamlined landforms, orientated in the former ice-flow direction, with occasionally grounding zone wedges that formed during standstills of the retreat are indicator for fast flowing ice. The inter ice-stream areas were dominated by submarine landforms orientated mainly transverse to the ice-flow, formed by the maximum event and slowly retreating margin (Landvik et al., 2005; Ottesen et al., 2005, 2007; Ottesen and Dowdeswell, 2009; Ingólfsson, 2011; Landvik et al., 2014). The terrestrial

record is more fragmented due to glacial erosion and long hiatuses but glacial drift, erratics and striations are among the evidences for ice extent and flow (Ingólfsson, 2011; Landvik et al., 2013, 2014). The stratigraphic record also reflects Late Quaternary glaciations by repeated regressional sequences, which form due to regional glaciations causing isostatic depressions that are later followed by regression due to the isostatic rebound (Ingólfsson, 2011). The pattern of emergence of the former load of the ice-sheet is one of the tools used to constrain the timing of deglaciation and extent and thickness of the ice sheet by investigating the pattern of postglacial emergence. The pattern indicates that the maximum loading during the LGM increased towards east (Forman et al., 1990, 2004). Glacimarine sediments from the fjords that were deposited during early deglaciation further reveal the deglaciation pattern and its timing (Ó Cofaigh et al., 2001; Hogan et al., 2011).

The glacier history during the Holocene is not that well understood, as the distribution of data is uneven, the resolution is often coarse and poor chronology. Climate reconstructions on Svalbard during the Holocene, prior to the instrumental period, are few, though increasing (Birks, 1991; van der Bilt et al., 2015; Røthe et al., 2015). The Holocene Thermal Optimum (HTO) is recorded by the abundance of thermophilous molluscs, lower amount of ice rafted debris (IRD) and changes from glacial proximal to distal deposits in marine sediments, indicating warmer conditions compared to present climate. The timing of it is varied through Svalbard but is thought to have occurred between 11.2-5.5 ka BP (Salvigsen et al., 1992; Salvigsen, 2002; Jessen et al., 2010). Based on that and data from lake archives and moraines the glaciers are suggested to have been smaller than current ones or even absent during the period (Svendsen and Mangerud, 1997; Reusche et al., 2014; van der Bilt et al., 2015).

The evidence for the behavior of the glaciers during Late-Glacial / Early Holocene (LGEH) in Svalbard are often inconsistent or lacking (Svendsen and Mangerud, 1997; Mangerud and Landvik, 2007; Forwick and Vorren, 2010). However, increasing investigations based on moraines and glacial deposits suggests an early Holocene readvance of glaciers around Svalbard (Salvigsen et al., 1990; Svendsen et al., 1996; Lønne, 2005; Forwick, 2005; Forwick and Vorren, 2010; van der Bilt et al., 2015; Farnsworth et al., 2017a). An Early Holocene readvance is described shortly after 9.5 ka years BP at Esmarkbreen, just east of Trygghamna, (Salvigsen et al., 1990) and in Bolterdalen, close to Longyearbyen (Lønne, 2005). The deposits often reach far outboard of the Neoglacial maxima. That is not in agreement with the traditional view (Mangerud and Svendsen, 2007) that the maximum Holocene extent occurred during the LIA (Farnsworth et al., 2017a).

The onset of the Neoglacial is recorded by the regrowth and advance of glaciers on the west coast of Spitsbergen around 4.5-3 ka BP. Glaciers are thought to have existed since that time but an increase in summer temperatures during the Medieval Warm Period (MWP) around 1200 AD (Divine et al., 2011) and *in situ* plants retrieved from below Longyearbreen dated to 1100 years, indicate that glaciers were smaller during that time than at present (Humlum et al., 2005). Few advances during the period can as well be seen with different sets of moraines and in lake sediments (Werner et al., 1993; Svendsen and Mangerud, 1997; Reusche et al., 2014; van der Blit et al., 2015; Røthe et al., 2015). The maximum Holocene extent of the glaciers is generally thought to have occurred during the Little Ice Age (LIA) (Svendsen and Mangerud, 1997), initiating around 13th or 14th century (Werner, 1993; van der Bilt et al., 2015; Røthe et al., 2015). Extensive, often ice-cored terminal moraines mark the maximum

extent of the glaciers during the period (Werner, 1993; Glasser and Hambrey, 2003; Reusche et al., 2014). At least two sets of advances have been described, ca. 1300 and 1900 AD (Werner, 1993; van der Bilt et al., 2015; Røthe et al., 2015).

Increase in temperature led to the termination of the LIA around 1920 (Steffensen, 1982; Humlum, 2003; Divine et al., 2011). The net mass balance has been negative and the terminus retreated for majority of the glaciers since then (Hagen et al., 1993; Nuth et al., 2013). Extensive areas of glacigenic landforms have been exposed due to the retreat (Ingólfsson, 2011). However, some glaciers have exhibit surge behavior during the 20th and 21th century with rapid advances and increased flow rates, so called surges (Hagen et al., 1993).

2.2 Study area: Trygghamna

Trygghamna (78°14.5'N − 13°51.0'E) is a 6 km long and 2 km wide fjord located on the outer part of northern Isfjorden, between Värmlandryggen and Protektorfjellet on southern Oscars II land. Five glaciers are located in the fjord today (Norwegian Polar Institute (NPI), 2003; 2017) (Figure 12A, B; Table 1).
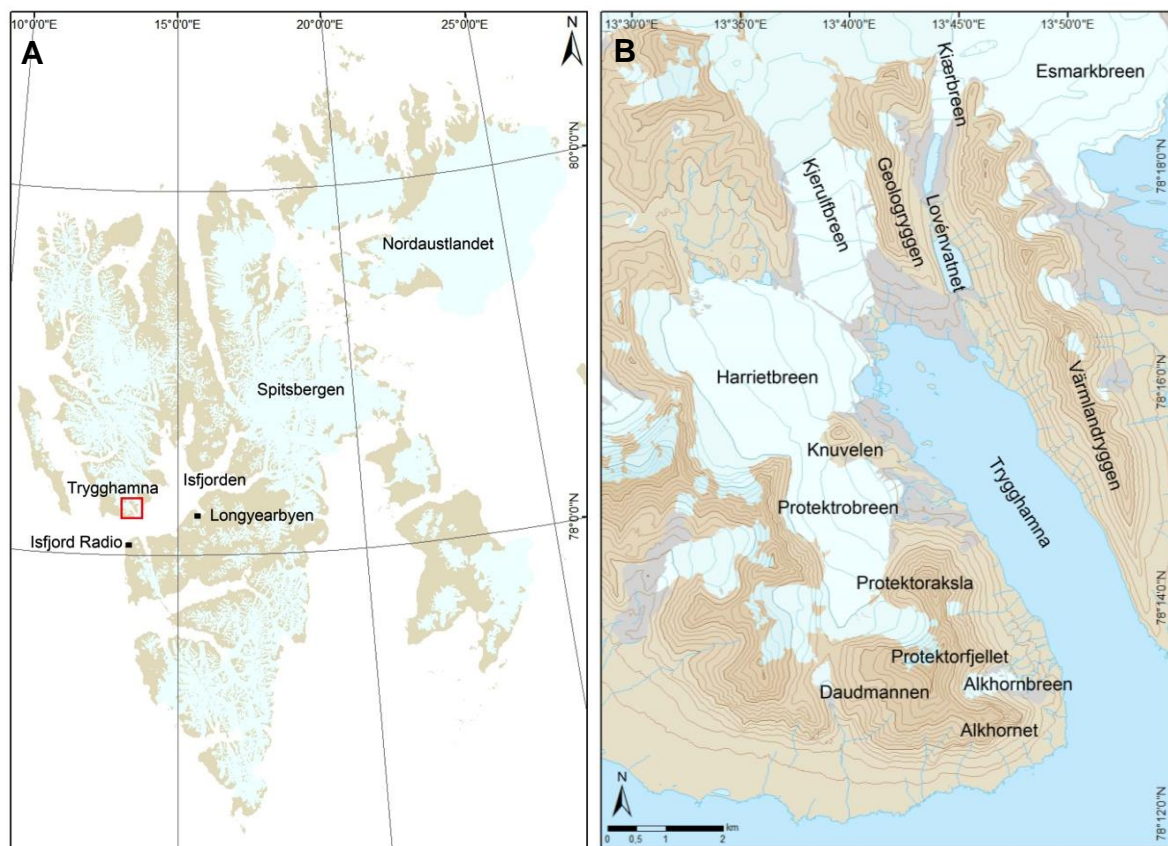
Figure 12. *Location of study area, Trygghamna. A) Svalbard archipelago. Marked on the figure are the study area, Trygghamna, and Isfjord Radio and Longyearbyen where the weather data originates from. B) The study area, Trygghamna. Place names referred to in the text are marked. Source: Svalbardkartet (NPI), 2017 (http://svalbardkartet.npolar.no).*

2.2.1 Bedrock geology

The western side of Trygghamna consists of low-grade metamorphic rocks and the east side consists of deformed Mesozoic and Paleozoic rocks. The oldest bedrock is on the south-western side, from the Late Neoproterozoic and consists of low-grade metamorphic phyllite, quartzite and carbonates from the Løvliebreen Formation and Alkhornet Formation (St. Jonsfjorden Group). The formations are bounded by a normal fault on the southern side of Protektorbreen forefield and reach towards Alkhornet and around the glaciers towards Kjerulfbreen. An outcrop of metagabbro, most likely from Neoproterozoic age, is visible above Protektorbreen. Outcrops of organic rich sandstone and shale from Orustdalen Formation (Billefjorden group) of Early Carboniferous age are visible in the forefields of Protektor-, Harriet- and Kjerulfbreen and on the islands. These rocks also make up Knuvelen and Geologryggen. The strata on the eastern side are almost vertically tilted, similar to Festingen on the other side of Isfjorden. Multicoloured sandstone, shale and conglomerate of Tårkanten Formation (Gipsdalen Group) of Late Carboniferous age are in front of Kiærbreen and around Lovénvatnet. The bedrock gets younger towards the east and Värmlandryggen consists of carbonate rocks, evaporites, sandstone and shale from Early Permian to Early Triassic. The area lies within the West Spitsbergen Foldbelt, resulting in intense folding and faulting (Dallmann, 2015; NPI, 2017) (Figure 13).
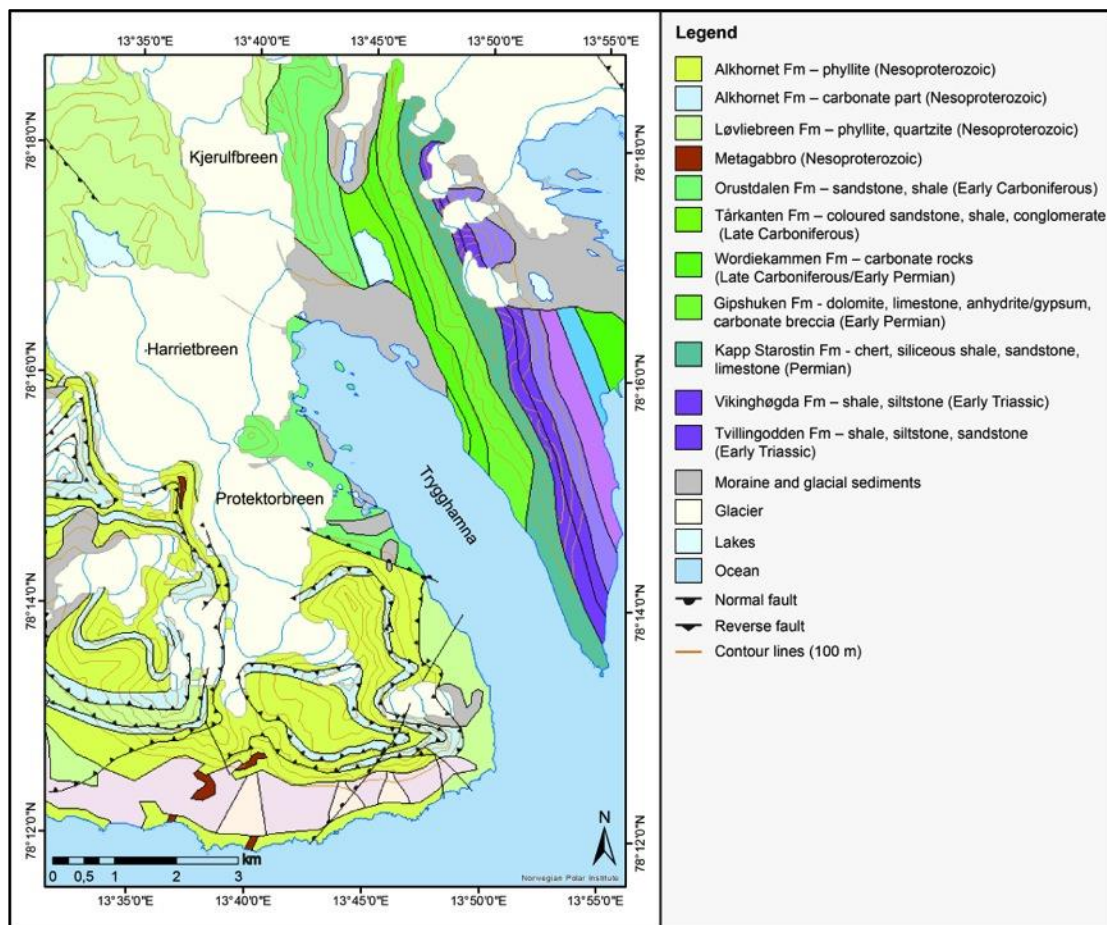
Figure 13. *Bedrock map of Trygghamna. Note that bedrock outside Trygghamna is not marked on the legend and legend begins from oldest to youngest (almost west to east on the map). Source: Svalbardkartet (NPI), 2017 (http://svalbardkartet.npolar.no).*

2.2.2 Glacial history

Previous work in Trygghamna based on sub-bottom profiles indicates that three moraines are located in the outer part of the fjord (Figure 14). They are all considered to be from the Late-glacial / Early-Holocene although an exact age constrain for these landforms is lacking. Based on the stratigraphy, the one furthest out is the oldest, the youngest is located in the middle and the inner most was deposited between them (Forwick, 2005; Forwick and Vorren, 2010). The youngest moraine was interpreted to be from an Early Holocene readvance that has also been described from other areas in Isfjorden (Salvigsen et al., 1990; Lønne, 2005; Farnsworth et al., 2017a). Stratified glacimarine sediments formed during deglaciation and mass-transport deposits formed by slope failure have also been described (Forwick and Vorren, 2012).

Few studies existed on the present day glaciers and their forefields in Trygghamna until recently. They suggest that at least some of the glaciers were more dynamic in the past and that they exhibited surge behavior (Wallin, 2016; Ben-Yehoshua, 2017).
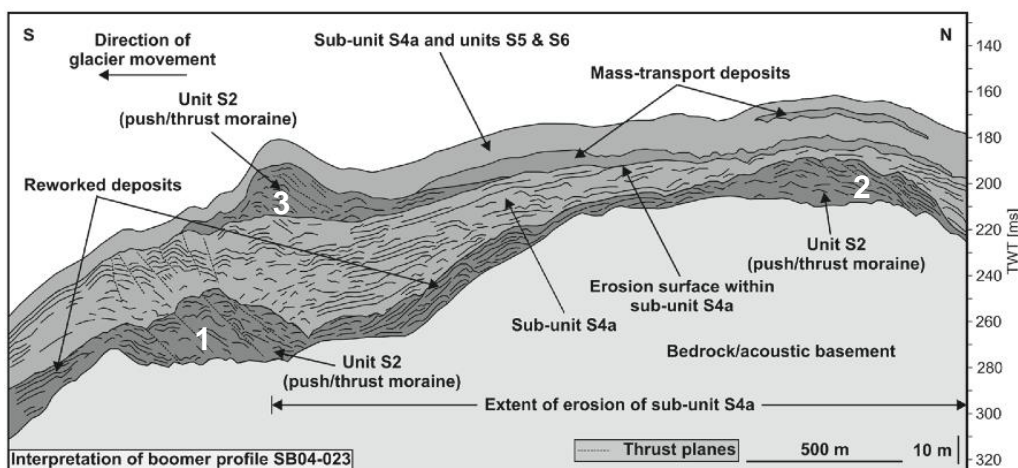
Figure 14. An interpretation of a boomer profile from the outer part of Trygghamna. Three moraines were interpreted from it. They are all considered to be deposited during the Late-Glacial / Early-Holocene. The relative age based on the stratigraphy: 1-oldest, 2-second youngest, 3-youngest (Modified from Forwick and Vorren, 2010).

Out of the five glaciers in the fjord, three are larger valley glaciers that are more prominent in the landscape. Protektorbreen has two small catchment areas, under Daudmannen and Protektorfjellet and flows down between the nunatak, Knuvelen, and Protektoraksla. West of Knuvelen, Harrietbreen is connected to the northern part of it. It is the only glacier in the fjord that has tidewater front today. A medial moraine marks the junction between it and Kjerulfbreen, which flows down from Geologpasset between Geologryggen and a mountain ridge from Krokfjellet. Two smaller glaciers are also located in the fjord. Kiærbreen is a small tributary glacier from Esmarkbreen, east of Geologryggen. Alkhornbreen is the smallest glacier remaining. It is located in a cirque between Protektorfjellet and Alkhornet (GLIMS, 2012; NPI, 2003; 2016) (Figure 12B). Further information is summarized in Table 1. In this study the emphasis will be on the three large glaciers in the fjord, Protektor-, Harriet- and Kjerulfbreen but the two smaller, Alkhorn- and Kiærbreen, will be described briefly.

Table 1. *Description of the five glaciers located in Trygghamna (GLIMS, 2012; NPI, 2003;*
2017).

Glacier	Coordinates	Area (km²)	Length (km)	Type	Thermal regime	Flow direction
Protektorbreen	78°10.2′N 18°06.5′E	~6.79	< 3	valley glacier	mostly cold-based	NE
Harrietbreen	78°16.0'N 13°37.0'E	~9.75	~3	valley (partly grounded calving front)	polythermal	E to EES
Kjerulfbreen	78°16.5'N 13°40.0'E	~8	<5	valley	polythermal	S to SE
Kiærbreen	48°18.3N' 13°45.0'E	~1.23	< 2	valley	cold-based	S
Alkhornbreen	78°12.9'N 13°49.5'E	0.51	~1	cirque	cold-based	E

2.2.3 Weather data

The following sections are based on data from both the composite Svalbard airport/Longyearbyen series and the Isfjord Radio station due to scarcity of observations and availability of data from Trygghamna. Due to its coastal location Isfjord Radio is considered to be more representative for the study area than Svalbard airport. However, Isfjord Radio is more exposed to precipitation and wind than Trygghamna.

Temperature
The mean annual air temperature at Svalbard airport/Longyearbyen was -6.7°C during 1961-1990 and -4.6°C during 1981-2010. At Isfjord Radio it was -4.7°C from 1951-1975. Significant warming has occurred since then, with a linear increase of 2.5°C at Svalbard airport/Longyearbyen in the period 1912-2011 (Steffensen, 1982; Hanssen–Bauer et al., 1990; Førland et al., 2011). The most prominent event occurred around 1920 when the mean annual temperature changed from -9°C to -4°C within 5 years at Svalbard airport/Longyearbyen. The event was also observed at Isfjord Radio (Steffensen, 1982; Humlum, 2003). This was followed by a winter cooling (0.9-1.8°C per decade) between 1943 and 1965. From 1966-2011 the temperature increased during all seasons at both stations, especially during spring and winter (Førland et al., 2011) (Figure 15).

Precipitation
Precipitation measurements are more complicated due to high wind speeds causing blowing and drifting snow (Steffensen, 1982; Humlum, 2002). Overall precipitation is quite low on Svalbard and measures only 190 mm per year on average from 1961-1990 at Svalbard airport. However, precipitation varies much more locally than air temperature (Førland et al., 2011) and at Isfjord Radio the annual precipitation is 435 mm per year (1951-1975), or more than twice the airport precipitation. This demonstrates how the coastal areas receive more precipitation than the inland (Steffensen, 1982; Humlum, 2002). All stations on Svalbard do, however, show an increase in annual precipitation through the observed periods (Førland et al., 2011) (Figure 16).
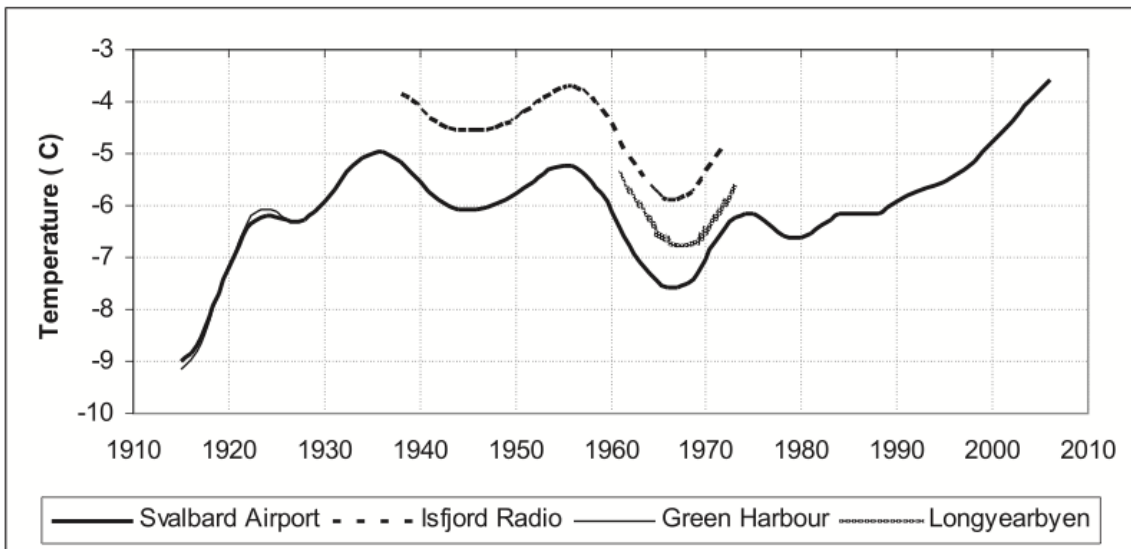
Figure 15. *Annual temperature development from the composite Svalbard airport series and three stations in the Isfjorden area. Note higher temperatures at Isfjord Radio than the other stations (Nordli, 2010).*
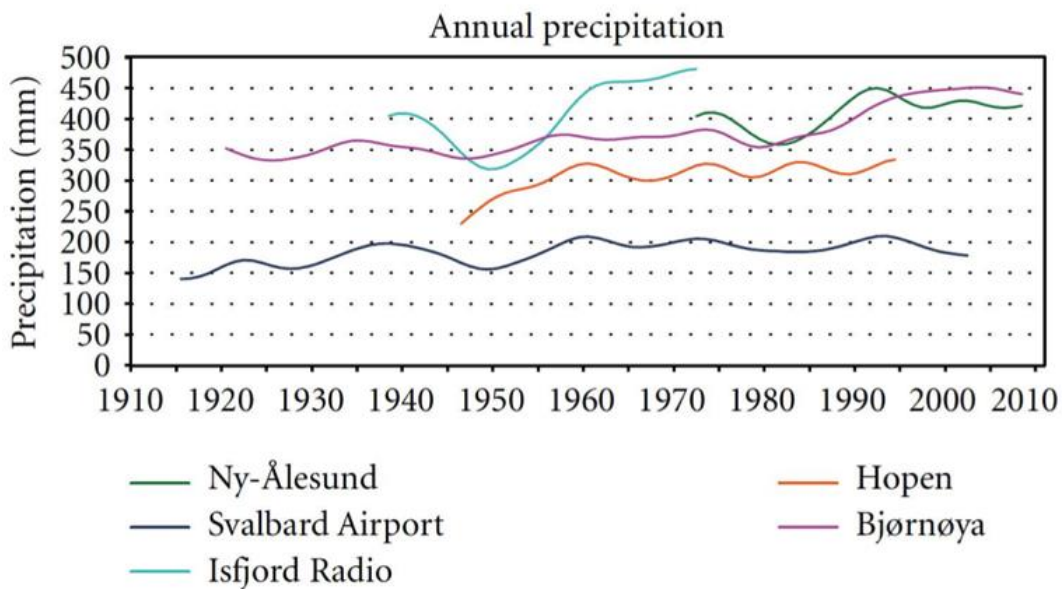
Figure 16. *Annual precipitation development at weather stations in Svalbard. Note higher precipitation at Isfjord Radio than Svalbard airport (Førland et al., 2011).*

2.2.4 Exploration history of Trygghamna

Shortly after the discovery of Spitsbergen in 1596 by William Barents, whalers began to arrive (Conway, 2012). Many of the coastal areas in Svalbard acquired their names from that period. Trygghamna was visited frequently by whalers, both for hunting and for shelter and safe harbour (Norwegian Hydrographic Service and Norwegian Polar Institute, 2016). Its first name is thought to be *Behouden (Safe) haven* after Van Muyen when he sought refuge there in 1612, or by Poole in 1610. It is termed *Poopy bay* or *Niches cove* in Baffin (1613). The latter converted to *Port Nick* by Heley in 1617, which seems to be the most common English name. However, *Safe heaven* is thought to be the most widely used name (Conway, 2012). The name was later translated into Norwegian as *Trygghamna* (Vahl, 1927) and appears in The Place-Names in Svalbard (Hoel, 1942).

The whaling activity ceased due to over-exploitation of the whale population, and instead Russian Pomors started trapping stations on Svalbard in early 18th century, later followed by the Norwegians (Storås, 1989; Dallmann, 2015; Norwegian Hydrographic Service and Norwegian Polar Institute, 2016). Remains of Russian settlement, dated from the middle of the 18th century, are visible at the inlet of Trygghamna. Two important resources were in the fjord, the bird cliff Alkhornet and the front of Kjerulf glacier, which was noted to have been much closer to the settlement than today. Eggs and down were collected around Alkhornet and sea-birds, seals, fish and other mammals were hunted that were frequent in the waters close to the glacier front (Liljequist, 1993; Storå, 1989). Svalbard became a very popular place for scientific pioneers in the 19th and early 20th century, with numbers of research and expeditions (Dallmann, 2015). Trygghamna was both a target for scientific research and the use of shelter (Liljequist, 1993). A topographical expedition, led by Gunnar Isachsen (1915), mapped the ice-marginal positions around Spitsbergen, including Trygghamna, between 1909-1910. At that time the large glaciers at the head of the fjord coalesced and reached significantly further out in the fjord. The glacier was termed Glacier Kjerulf. Lovénvatnet and Knuvelen are also present on the map (Figure 17). Olaf Holtedahl, a Norwegian scientist, also took a part in that expedition and published a geological map of Spitsbergen, including the glacier margins (1912) and De Geer visited Trygghamna in 1910 to investigate and map the bedrock and the large glaciers (Liljequist, 1993). Other expeditions that visited Trygghamna took photographs of the glaciers during the period 1890-1908 (Nordenskiöld, 1892; Hamberg, 1905; Halldin, 1908) (Figure 18, 47A-B; 48A; 49A). None of the expeditions did though describe the glaciers in any details and no advances or increased crevassing of the glacier fronts was described during that time. A flight campaign from NPI in 1936 took oblique aerial images of the fjord (Figure 19). On a map of Isfjorden from NPI (1992) the glacier margins from 1936 and 1968 are marked together with Lovénvatnet, Knuvelen and an ice dammed lake by Kjerulfbreen.
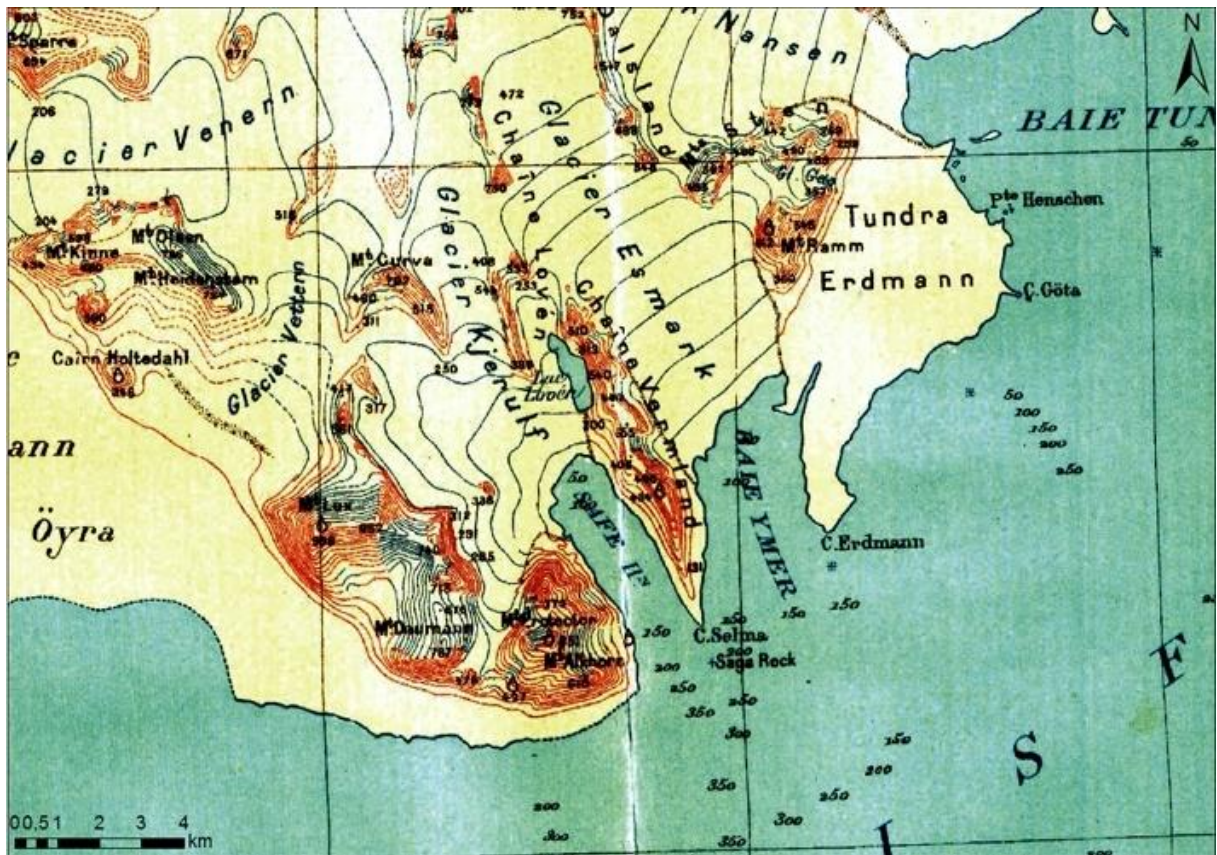
Figure 17. *Topographical map, including ice front positions, of Trygghamna in 1909/10. The glaciers at the head of the fjord coalesced and were termed Glacier Kjerulf (Isachsen, 1915).*

Figure 18. *A panorama of Trygghamna taken in 1908 by Halldin, where the glaciers are significantly larger than today.*
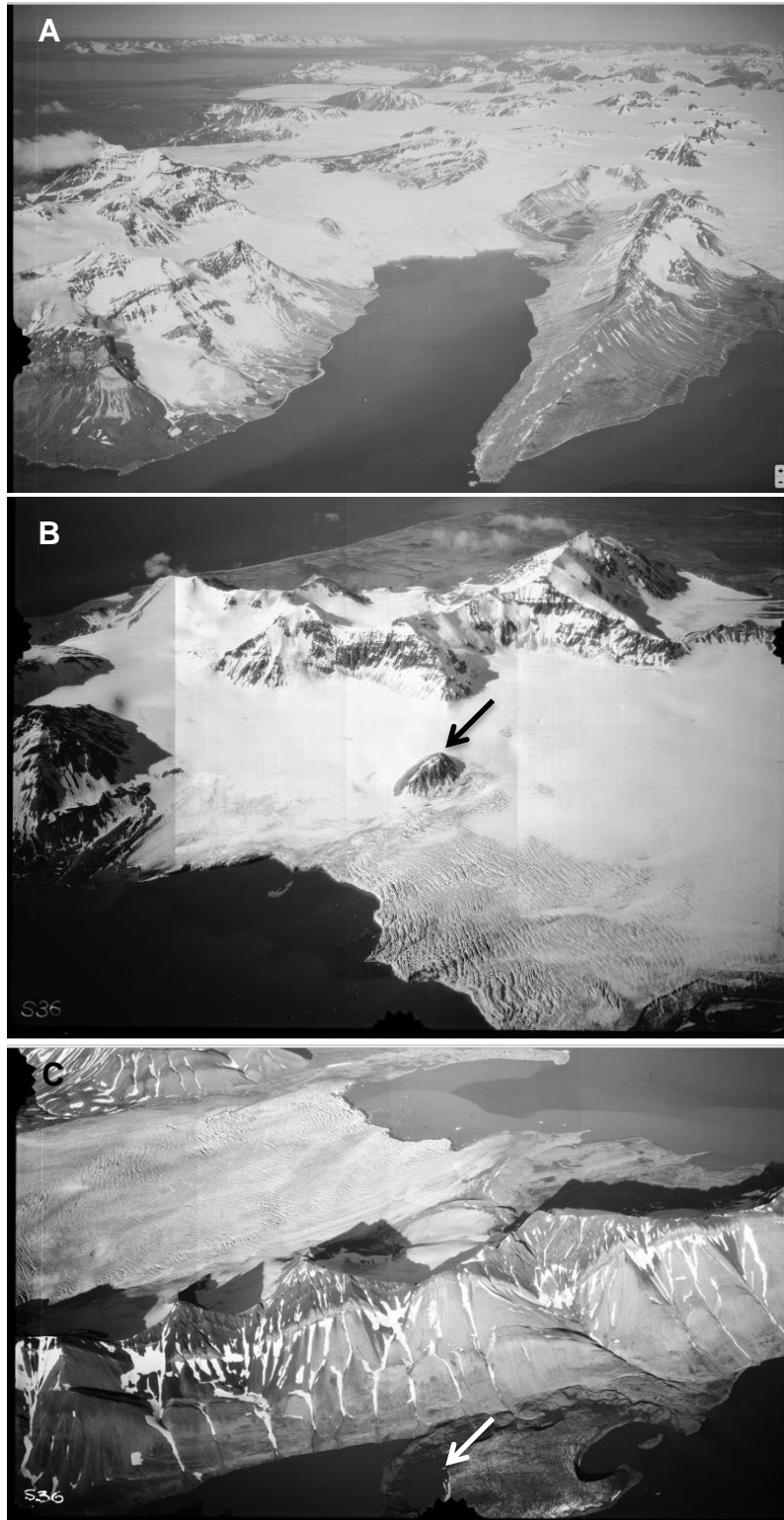
Figure 19. *Oblique aerial images from NPI taken in 1936. A) An overview of Trygghamna. All glaciers reached much further out than at present. B) The glacier margin is close to the Protektorbreen's terminal moraine and the present coastline but Harriet- and Kjerulfbreen have an active tidewater front. Knuvelen, the nunatak, is sticking out (arrow). C) Terrestrial based margin of Kjerulfbreen is close to the terminal moraine. The ice-marginal lake is visible (arrow) and Lovénvatnet north of the moraine (NPI, 1936).*

3. Material and methods

3.1 Geomorphological mapping

The geomorphological record in both the terrestrial and marine forefields was mapped based on the material presented below using remote sensing techniques and field work. The spatial reference system used is WGS84/UTM 33N.

Aerial images

Aerial images of Trygghamna were acquired from Norwegian Polar Institute (NPI) to be able to map geomorphological features, investigate the area prior to field work and reconstruct ice-marginal positions. The images are taken during their flight campaigns in 1936 and 2009 (Figure 19, 20; Table 2). The 1936 images were acquired from elevation of 3500 m and are oblique and gray-scale. The 2009 images are orthorectified with pixel resolution between 0.4-0.5 m and in color.

Digital elevation model (DEM)

A digital elevation model (DEM) from 2009 with 5 m resolution, provided by NPI, was utilized to make contour lines and hillshades of the area (Table 2). The model is generated from stereophotogrammetry from the 2009 aerial images.

Submarine data

The Norwegian Hydrographic Service collected swath bathymetric data of the seafloor morphology in Trygghamna in two parts. The depth data are reproduced according to permission No 13/G706. The outer part of the bathymetric data (outside the terminal moraine; ~9 km^2) was acquired in 2000 by Kongsberg EM-1002 multibeam echo-sounder. The grid size is 5 m furthest out but 3 m closer to the moraine. In 2007 the shallower part of the bathymetric data (inside the terminal moraine; ~ 4 km^2) was collected using a Kongsberg EM-3000 multibeam echo-sounder. The grid size is 1 m (Figure 20; Table 2).

Sub-bottom acoustic data were collected using a 0.5-8 kHz (5 ms) chirp pulse along profiles from the outer part of the fjord in 2015, during a AG-339 cruise on Viking Explorer. The data were acquired with Edge Tech 2000 CSS chirp 0.5-12 kHz sub-bottom profiler. Two profiles were used in this study (Figure 20; Table 2). The resolution is 8-20 cm and in typical soft sediments in can penetrate through 150-200 m (2000 CSS technical specification sheet). The sub-bottom acoustic chirp data were not processed.

Remote sensing and softwares

Analysis of the aerial images was conducted with ESRI ArcGIS 10.4 and 10.3 to produce the terrestrial part of the geomorphological map. Where the aerial images are overlapping (263 and 943; 943 and 941) they were viewed in stereoscopic view using the ERDAS IMAGINE 2015 with the SAFA extension for ArcMAP. Outside the overlapping areas the aerial images were analyzed without the stereoscopic view.

The bathymetric data was viewed in the free software iView4D in 3D, where the different landforms were measured and described. The mapping was conducted in ESRI ArcGIS 10.4. The sub-bottom profiles from the outer part were examined in the software Discover to investigate sediment structures and thickness, used to further support classification of

landforms. There are no direct observations of the sediment composition and only sub-bottom profiles from the outer part. The sediment thickness of the shallower part is therefore unknown and interpretations of the landforms are solely based on their geomorphology.

The zoom level while mapping was set to 1:1500, for both the terrestrial and marine parts. Polygons, polylines and point symbols were used to represent features on the map based on their classification. As the final product was a combined map of the terrestrial and marine forefields, it was important to use the same mapping style in both settings. Slightly different terminology and symbols are used between the environments based on its differences. Two overview maps were produced, showing the inner and outer parts, to be able to present them with higher zoom level and resolution. The final scale of the overview maps was in 1:23 000 and 1:17 000. Zoomed in maps were as well produced of their forefields, where smaller landforms are mapped in greater detail in the scale 1: 5500 and 1:8000 The finalizing of the maps was carried out in Adobe Illustrator CC 2017.

Field investigations
Field campaigns were carried out for 12 days in July/August 2015 and 10 days in July 2016. The field camp was located in the forefield of Protektorbreen and a zodiac was used for transport to the other site. The main purpose with the field work was to ground verify the terrestrial map. The main focus was on the three larger glaciers at the head of the fjord. The first field season focused on getting familiar with the area, which had only been briefly investigated on the aerial images before. During the second season key sites were chosen to focus on to finalize the ground verification of the map. Field mapping was carried out with the guidance of Hubbard and Glasser (2005). Geomorphological features were described, documented and photographed. Their distribution was noted together with their orientation using Silva Clinometer compass. Surface sediment in the forefields was described; size, roundness, lithology, amount and striations. Investigations on the composition were carried out on key sites according to Krüger and Kjær's (1999) legend for glacial sediments. Clast fabric was not measured since the sections were usually clast rich and the clasts were contact, which can affect the fabric measurements (Kjær and Krüger, 1998). A handheld Garmin GPS 64s was used for acquiring locations and tracks when needed and a Panasonic CF-19 Toughbook was used in the field to map directly.

3.2 Ice-marginal reconstruction
The retreat history of the glaciers from 1909/10 has been reconstructed using historical data, oblique- and orthorectified aerial images (*section 3.1*), satellite images and GLIMS Glacier Database (Table 2). The historical data and satellite images were georeferenced. The spatial reference system used is WGS84/UTM 33N. The data was imported into ESRI ArcGIS 10.4 to map the positions. Due to quality, the satellite images were only used to map Protektor-, Harriet- and Kjerulfbreen The final scale of the map was in 1:40 000. The final production of the map was carried out in Adobe Illustrator CC 2017.

Satellite images
Satellite images from 1976 were examined in order to reconstruct glacier marginal positions. For the final version two images from 1990 and 2016 were used. Multispectral images were obtained by USGS LandsatLook Viewer and collected with Landsat satellites (4-5 TM and 8

OIL) (Table 2). Only satellite images with less than 20% cloud cover and taken during daylight were searched for. When two margins were too closely spaced to show difference or did not show that the glacier had readvance, the clearer one was chosen.

GLIMS (Global Land Ice Measurements from Space) Glacier Database

The GLIMS Glacier Database is a project monitoring glaciers around the world. The information is gathered by optical satellite instrument, such as SPOT-5 and ASTER. On Svalbard the work is done in cooperation with NPI. The 2007 margin for all glaciers in Trygghamna were utilized in this project (Table 2).

Historical data

A topographical map from 1909/10, including glacier margin positions of Spitsbergen (Isachsen, 1915) and photographs from Nordenskiöld (1890), Hamberg (1898) and Halldin (1908) are used to reconstruct the glacier margin and glaciological features (Figure 17; 18; 48A; 49A). The glacier margins from 1936 and 1968 are shown on a map of Isfjorden from NPI (1992) (Table 2).

Table 2. A list of all the data utilized for geomorphological mapping and ice-marginal reconstruction. The information is given when available.

Date	Source	Type	Resolution/scale	Number –id
1890	Nordenskiöld (1892)	Photographs	-	-
1898	Hamberg (1905)	Photographs	-	-
1908	Halldin (unpublished)	Photographs	-	-
1909/10	Isachsen (1915)	Historical map	1:200 000	-
1936	NPI	Oblique aerial	-	s36_0751 s36_0752 s36_1943
1936	NPI (1992)	Historical map	1:100 000	Temakart 16, sheet B9G
1968	NP1 (1992)	Historical map	1:100 000	Temakart 16, sheet B9G
19.09.1990	USGS	Satellite image-Landsat 4-5 TM	30 m pixel size	LT52170041990231KIS00
2000	Norwegian Hydrographic Service	Multibeam data	5 ma and 3 m grid size	-
2007	GLIMS Glacier Database	Glacier information	-	103197 (Kjerulfbreen) 103196 (Harrietbreen) 03195 (Protektorbreen) 103189 (Alkhornbreen) 103264 (Kiærbreen)
2007	Norwegian Hydrographic Service	Multibeam data	1 m grid size	-
2009	NPI	Vertical aerial (orthorectified)	0.4-0.5 m pixel size	13822/263; 264; 941; 943
2009	NPI	Digital elevation model (DEM)	5 m pixel size	138222
2015	UNIS (AG-339)	Chirp sub-bottom profile	-	VE15-2108-001; 003; 004
25.07.2016	USGS	Satellite image – Landsat 8 OLI	30 m pixel size	LC82170042016207LGN00
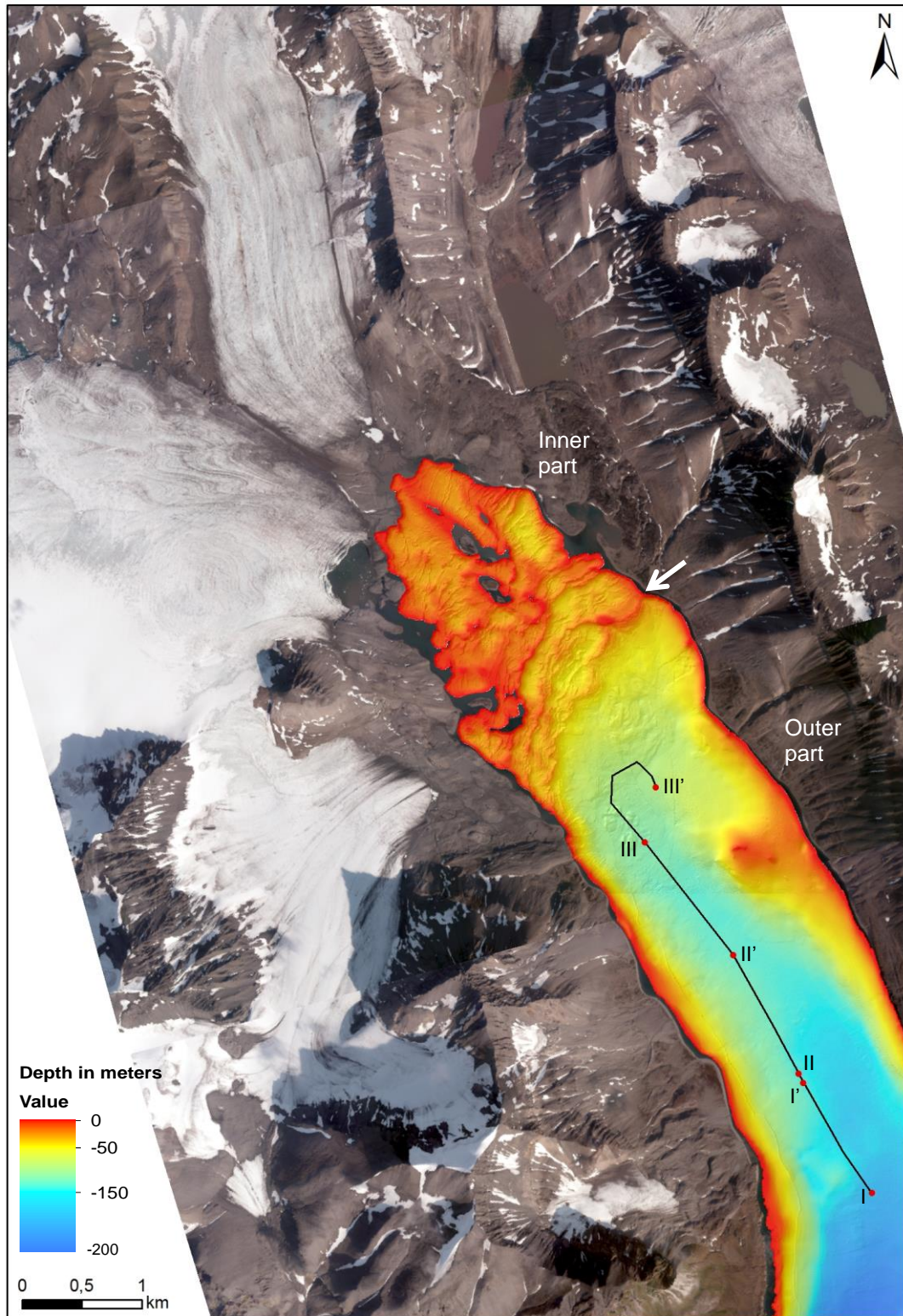
Figure 20. *The combined aerial images from NPI (2009) and swath bathymetric dataset from the Norwegian Hydrographic Service (2000, 2007) were used to produce the geomorphological map. The aerial images were viewed in stereoscopic view. The white arrow points at the terminal moraine, which divides the submarine setting into inner (shallower) and outer (deeper) part. The black lines represent the location of the two sub-bottom profiles and the red dots the examples of them shown on Figure 46.*

4. Results

A composite geomorphological map, including the terrestrial and marine forefields, has been produced for Trygghamna covering an area of ~67 km^2 in total, ~5 km^2 terrestrial glacier forefields, ~13 km^2 submarine environment and ~49 km^2 beyond outboard that (Figure 22, 23). As the main focus is on the Neoglacial extent of the three largest glaciers, Protektor-, Harriet- and Kjerulfbreen they were mapped in most detail (Figure 24, 25, 26). The forefields cover and area of; 1.63 km^2, 0.79 km^2, 2.58 km^2, respectively. The legend for all of the maps is presented in Figure 21. The landforms in the forefield of Alkhornbreen and outboard of the forefields were not mapped in as high resolution and only the terminal moraine for Kiærbreen (Figure 23; 50). The surface cover between the forefields is generally classified as extra-marginal surface cover but specific landforms have been mapped. The record for the Neoglacial advance is the most prominent but little evidence for former events, within and outboard the forefields, was observed that are included in the maps.

In the first two sections the geomorphology will be described separately for the terrestrial and marine forefields because of difference in the landform assemblage in the two environments and landforms with similar genesis have different appearance. Landforms will be described and interpreted below and then further discussed in chapter 6. In addition, the compositions for key landforms are described. They are classified after their origin into; subglacial, supraglacial, ice-marginal, glaciofluvial, proglacial landforms and extra-marginal surface cover. As the forefields of Alkhorn- and Kiærbreen were not examined in detail they are not included in this.

In *section 4.3* the sub-bottom data from the outer part of the bathymetry is presented in addition to the geomorphology. It is used to support the interpretation of landforms in *section 4.2* when possible.

Ice-marginal positions were mapped from 1909/10 and are described in *section 4.4* for all of the glaciers (Figure 50). The amount of retreat was measured between each positions and the pattern of retreat discussed. In addition, the glaciological features observed on old photographs are described.

▲	Exposed bedrock		Rock glacier
∩	Hummocks		Medial moraine
∪	Solifluction lobes		Hummocky moraine
∪	Kettle holes		Terminal moraine (terrestrial)
⌐·⌐	Flutes		Terminal moraine (subaqueous)
⫽	Crevasse squeeze ridges (CSRs)		Debris flow apron
↘	Glacial striations		Fluted till plain
⌐	Esker		Fluted till plain with CSRs
⫶	Supraglacial debris bands		Glacilacustrine sediments
\	Medial moraine		Beach material
⤬	Moraine ridge		Debris covered bedrock
╱	Transverse ridge		Exposed bedrock
⌒	Lage-scale transverse ridge (subaqueous)		Fluvial outwash plain
⌇	Small-scale transverse ridge (subaqueous)		Glacigenic material (subaqueous)
—	Trimline		Distal glacimarine sediments
----	Shoreline		Extra-marginal surface
∿	Active fluvial channels		Modern beach
∿	Seasonal or inactive fluvial channels		Lakes
⁝⁝	Ploughmarks		No bathymetric data, fjord
••	Pockmark		
—•	Debris flow tracks		
⫶⫶	Mass-transport deposit		
⌢	Slide scar		

Figure 21. *Legend for all of the maps below.*
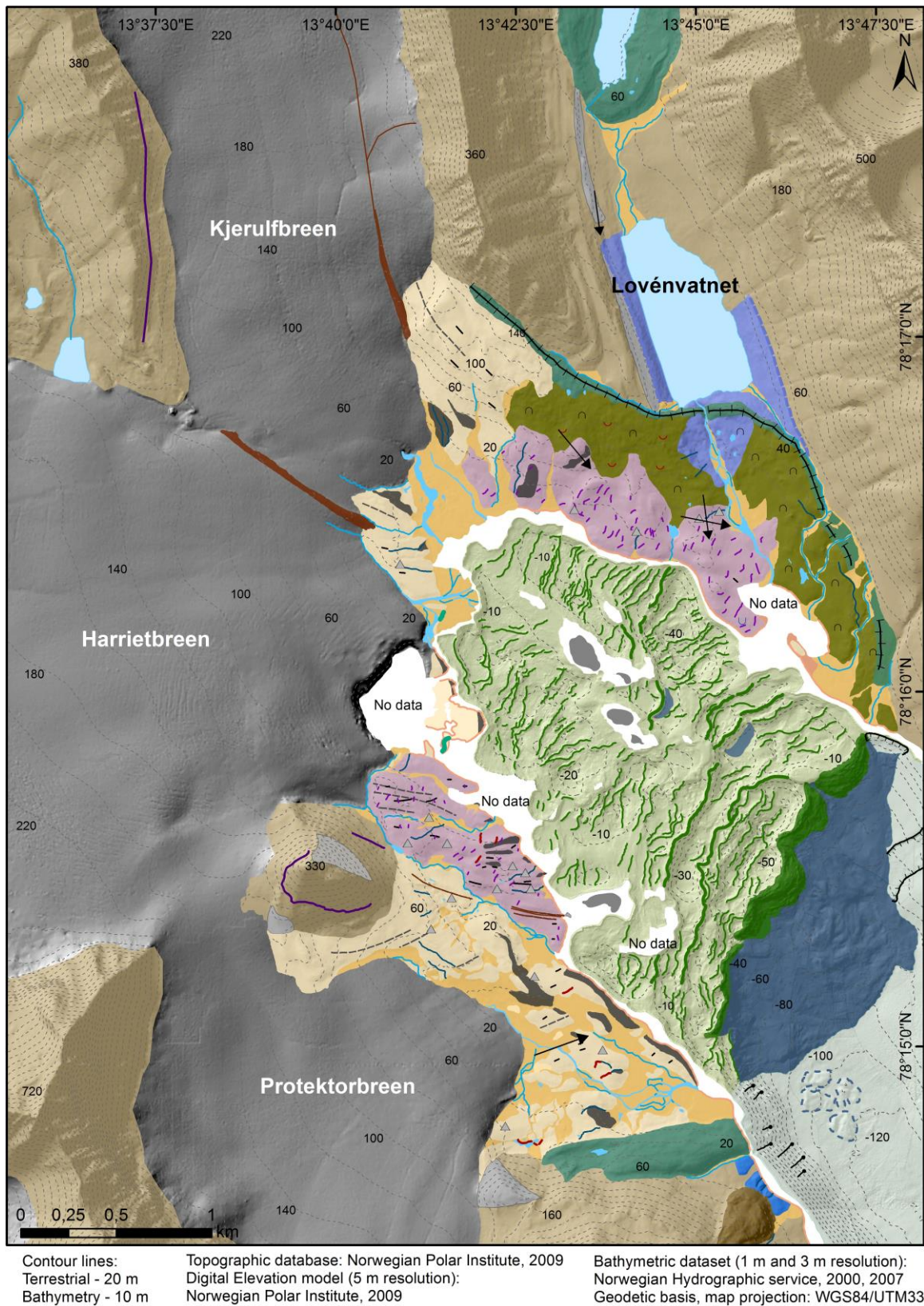
Figure 22. *A geomorphological map of the inner part of Trygghamna in the scale 1:23 000. The forefields of the three large glaciers in Trygghamna, Protektor-, Harriet- and Kjerulfbreen, and the inner part of the bathymetry are shown. The end of the terminal moraine in front of Kiærbreen is also visible and Lovénvatnet. See supplementary data for full version of the map.*
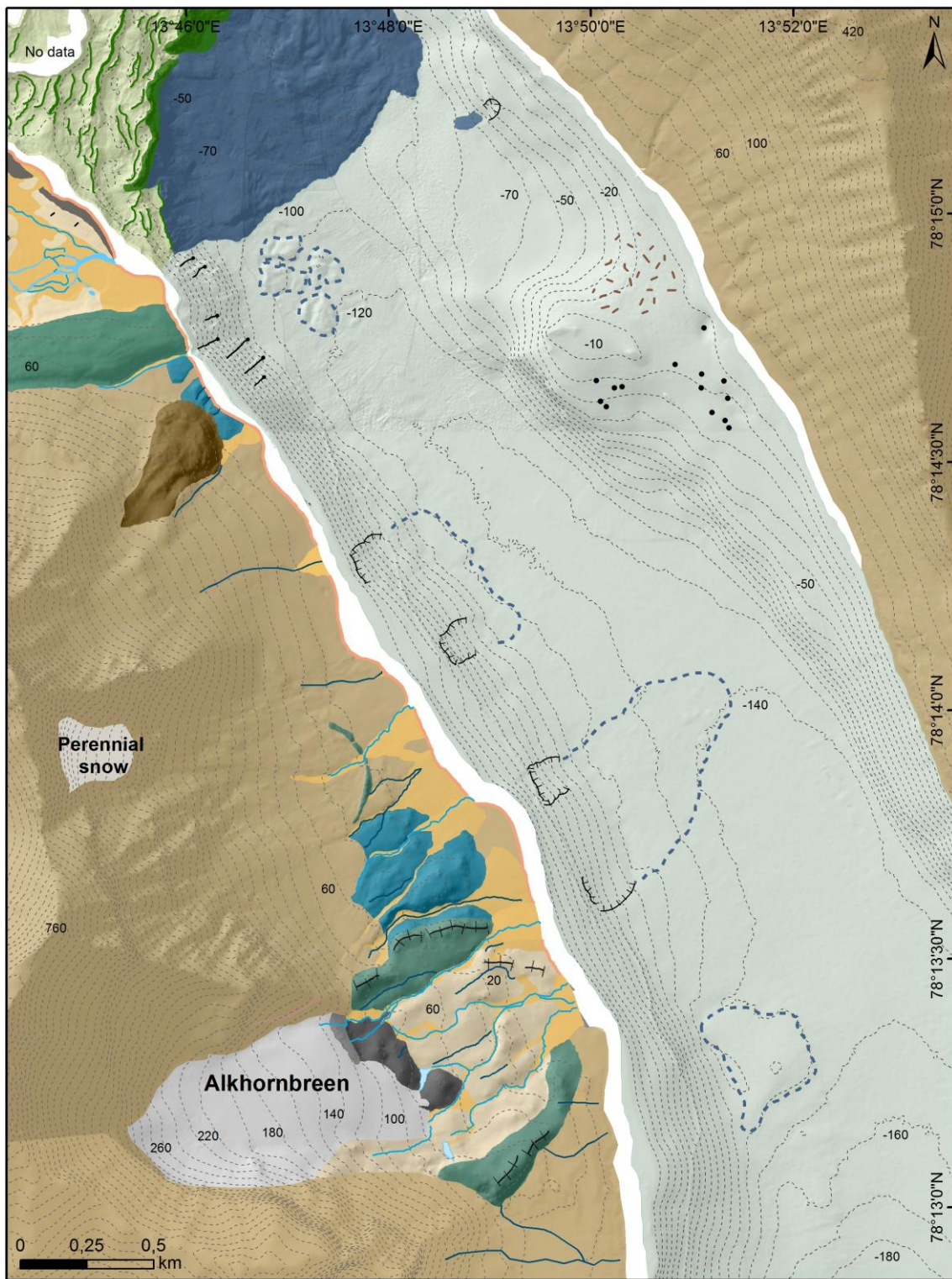
Figure 23. *A geomorphological map of the outer part of Trygghamna in the scale 1:17 000. The forefield of Alkhornbreen is shown and the outer part of the bathymetry. See supplementary data for full version of the map.*
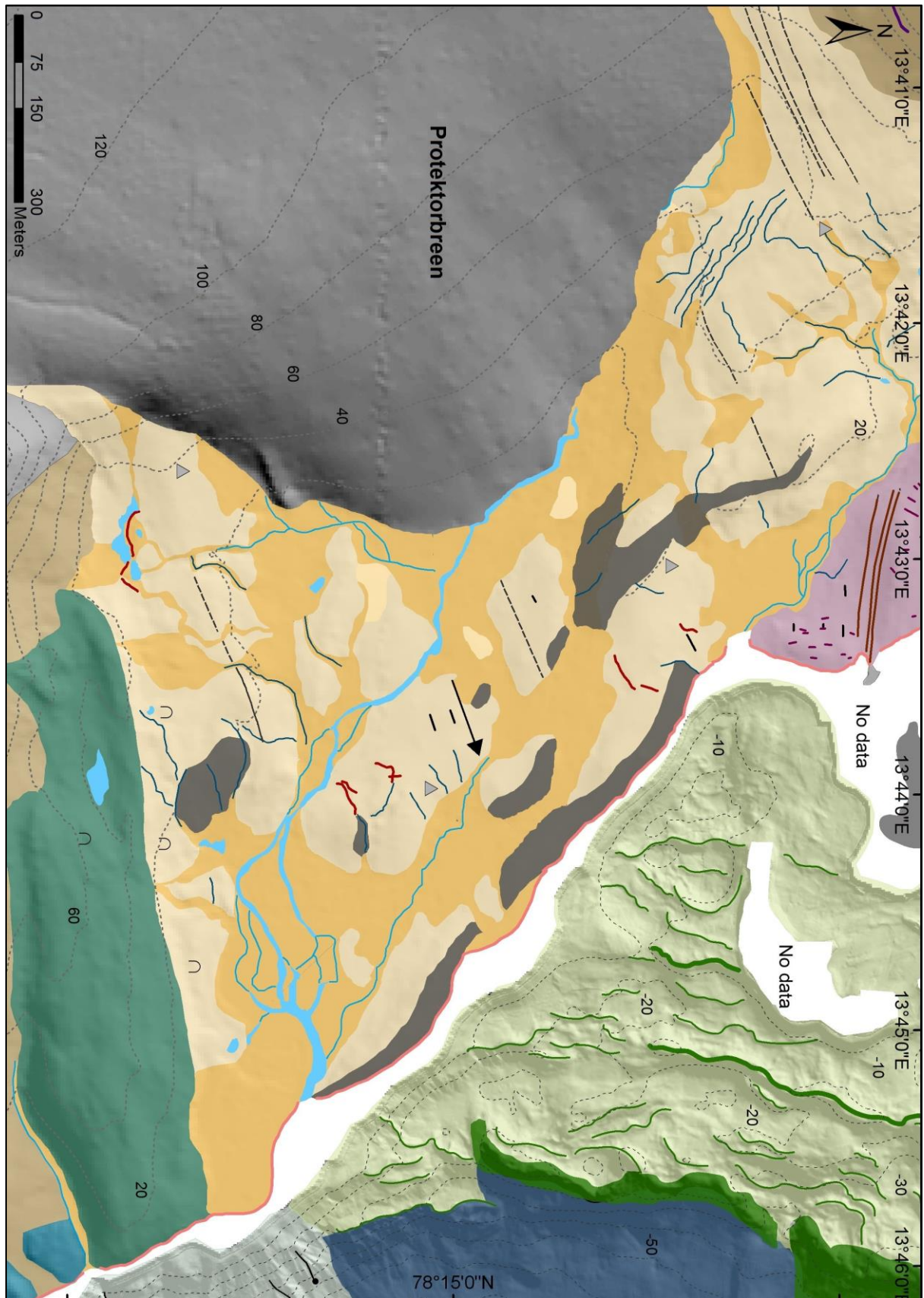
Figure 24. *Map section of the forefield of Protektorbreen in the scale 1:8000. Terrestrial contour lines are 20 m and bathymetric contours are 10 m.*
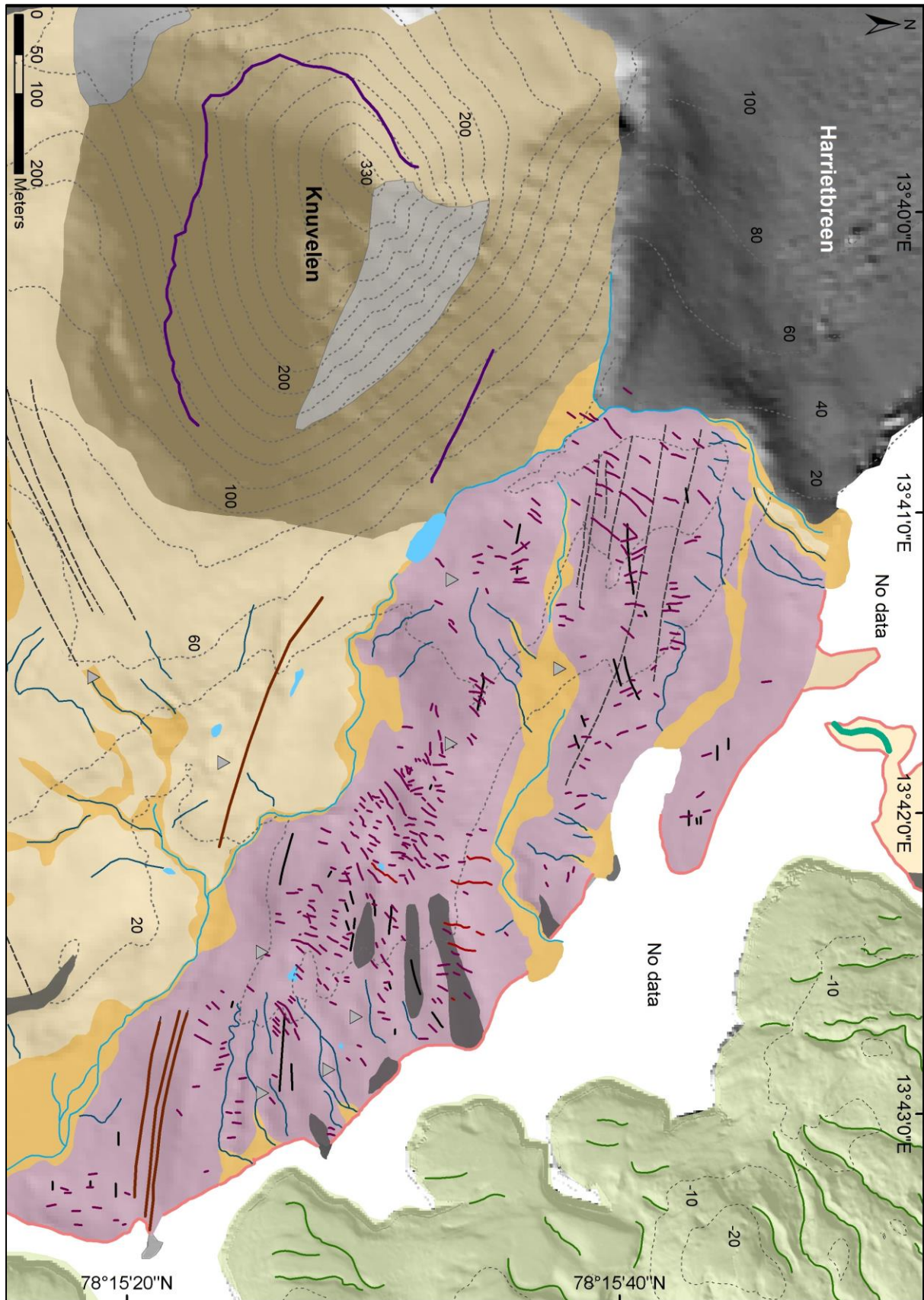
Figure 25. *Map section of the forefield of Harrietbreen in the scale 1:5500. Terrestrial contour lines are 20 m and bathymetric contours are 10 m.*
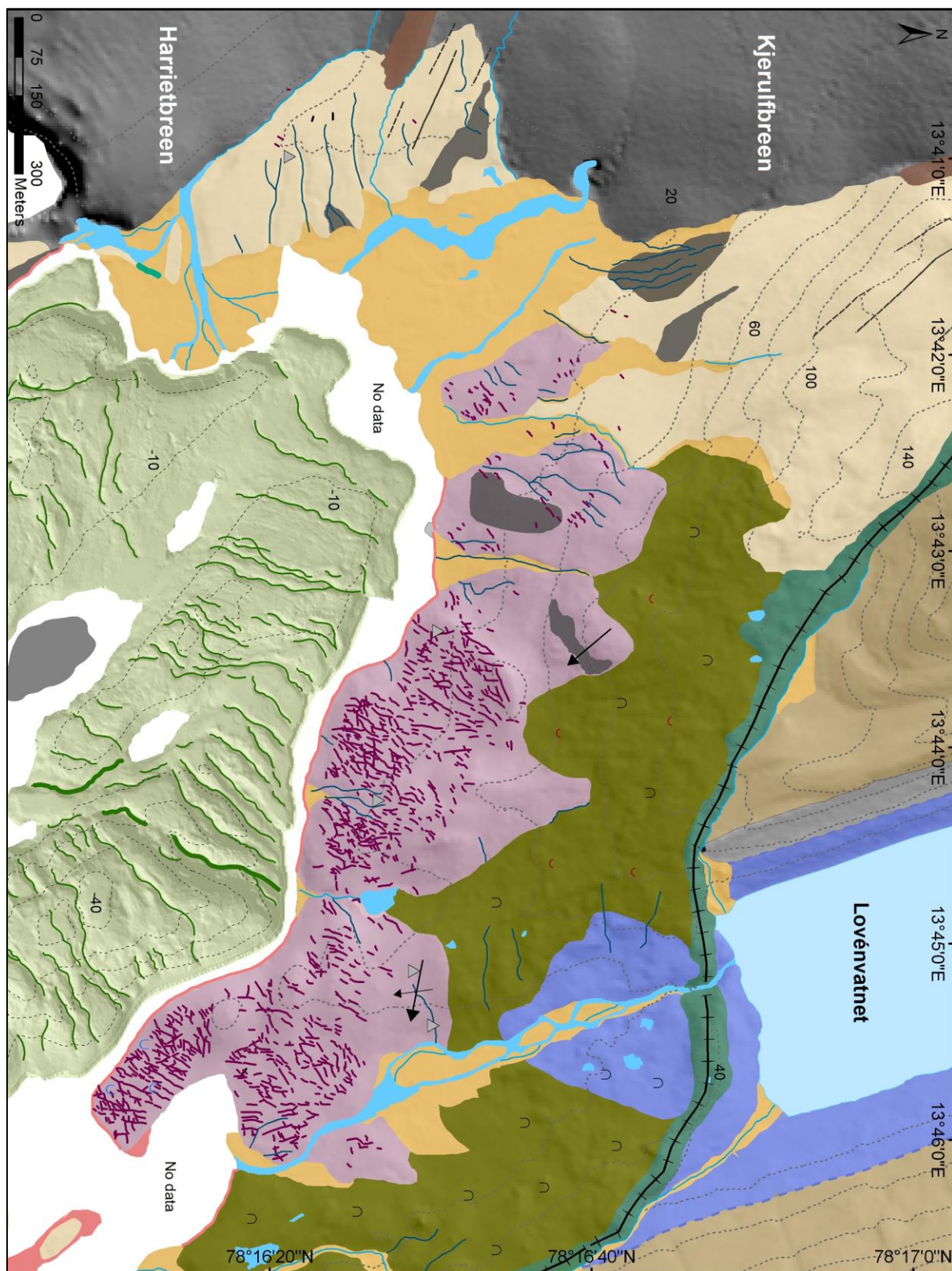
Figure 26. Map section of the forefield of Kjerulfbreen in the scale 1:8000. Terrestrial contour lines are 20 m and bathymetric contours are 10 m.

4.1 Geomorphology: Terrestrial environment

4.1.1 Subglacial landforms
The subglacial surface is composed of fluted till plain, with and without CSRs, scoured and streamlined bedrock.

Streamlined diamict cover: Fluted till plain
Description
The surface cover in the glacier forefields is composed of a clast rich diamict. The matrix consists of finer material but the abundance of it is generally not high. The clasts are subangular to subrounded, coarse gravel to boulders. The largest ones are up to 2 m in diameter although they generally vary from 10 to 50 cm. Light and brownish sandstone, carboniferous shale and phyllite are the most frequent lithology. The sediment cover is generally discontinuous so bedrock is often exposed. The lithology and its abundance are though variable between the forefields as will be described in the following sections. Little vegetation is present in all the forefields. Striations are occasionally present on sandstone and shale clasts but are most prominent in the forefield of Kjerulfbreen (Figure 28C). The fluted till plain (with and without CSRs) covers an area of ~2.47 km^2 in total.

The sediment cover is the most discontinuous in the forefield of Protektorbreen (total area of ~0.86 km^2) as bedrock is most frequently exposed (Figure 24; 32B; 38A). Angular to subangular phyllite is more widespread than in the other forefields (Figure 27A, B; 34C). Below Knuvelen is a prominent change in the surface cover over to cobbles to large boulders of light colored sandstone, angular to subrounded. The cobbly surface reaches all the way to the shore with decrease in grain size. The fluted till plain in front of Harrietbreen covers a total area of ~0.59 km^2. The most prominent lithology is sandstone but areas with higher abundance of phyllite occurs (Figure 27C, D; 29B; 34A). The fluted till plain in front of Kjerulfbreen covers a total area of ~1.1 km^2. Even less phyllite is present there and the frequency of subrounded clasts is higher. Larger sandstone boulders, similar as below Knuvelen, are present close to the glacier margin (Figure 27F; 28A, C; 29; 36A).

Natural sections occurred at few locations in the forefield of Kjerulfbreen. Sedimentological work by Ben-Yehoshua (2017) demonstrated that they consist of matrix-supported diamict. The matrix is compact clay to sand. It is clast-rich, with angular to subrounded clasts. The clasts were of various lithology and striations were common on the clasts. No structures were observed and as the clasts were usually in contact, fabric analysis was not conducted. Shell fragments were occasionally observed. The sections were generally not thicker than 30 cm. The diamict overlies loose gravel that is up to 2 m thick or bedrock. The upper boundary to the overlying CSR was vague but the lower boundary to the underlying layer was sharper. The lower gravel unit consisted of loose, sorted gravel with high abundance of shell fragments. The clasts are mostly rounded, with a flattened shape (Figure 28A, B).

Sub-parallel lineations are frequent on the diamict surface in the forefields of the three glaciers. Most have vague appearance and can only be observed on the aerial images (Figure 27E). Few are, however, more prominent and can easily be distinguished from the surrounding both in the field and on aerial images. The most prominent lineations are marked on the map. They range from 3-97 m in length, up to 1 m wide and with a low relief or less than 0.3 m high. Their surface often consists of the same material as the surrounding diamict

cover, except with smaller clast size. Some of the lineations have more washed surface than the surrounding and a few of them have a boulder at their head.

In the forefield of Protektorbreen they usually appear very vague with average orientation of WWS-EEN (250°-70°) (Figure 27A, B). Boulder train was also observed at couple of places in the forefield. The lineations are however much more prominent in the forefield of Harrietbreen and around 40 were mapped. The peninsula in front of the glacier is covered with very small lineations, <0.5 m in length, and thus only the most prominent ones mapped. Their main orientation is around W-E (270°-90°) but some of them curve slightly northwards (27C, D). Those are usually situated on a slope and follow the topography. Some cross-cut debris band and CSRs (Figure 31A, B). In the forefield of Kjerulfbreen only one individual lineation has been mapped (Figure 27F), as they were almost impossible to identify in the field because of how obscure they were. The aerial images show on the other hand a streamlined surface with sub-parallel lineations. Their orientation is roughly NW-SE (310°-130°) (Figure 27E). Proximal in the forefield they are more prominent and longer lineations lying almost parallel with debris bands were observed.

Interpretation
The diamict cover is interpreted as subglacial till based on its characteristics. It is poorly sorted, contains all grain sizes, the clasts are predominantly subrounded to -angular with variable lithology and striations are present on some of them (Boulton, 1970; Krüger and Kjær, 1999). Sedimentological analysis by Ben-Yehoshua (2017) further supports this. However, the more angular phyllite is considered to have a supraglacial origin (section *4.1.2*). The underlying loose gravel is interpreted as Holocene beach gravel, formed during higher sea level after the deglaciation (Salvigsen et al., 1990; Forman et al., 1990, 2004). The difference in lithology, roundness and grain size is dependent on the source area and transport path (Benn and Evans, 2010). This explains the difference between the forefields and is in accordance to the bedrock map, where phyllite is dominant on the western side but sandstone on the eastern side (Figure 13). The amount of sediment cover is conditional on the availability of the sediment in the basin and is considered vital for the formation of landforms (Brynjólfsson et al., 2014; Lønne, 2016). As the sediment cover was discontinuous, bedrock was often exposed and the sections studied show relatively thin cover (< 30 cm) (Ben-Yehoshua, 2017), it is suggested to be rather thin. The thickness is however, to a degree controlled by the topography and areas with thicker deposits in depressions are therefore likely.

The glacial lineations, orientated parallel to the ice-flow direction are interpreted as flutes. They are thought to form during subglacial deformation of till, suggesting ice-bed coupling. Based on that and the similarity to the surrounding subglacial till plain they are considered to consists of till. Their direction reflects the former direction of the ice-flow. However, their variable direction in the forefield of Harrietbreen is interpreted to reflect the effect topography has on their formation as they are often located on a slope. They are easily degraded by water and wind and therefore have low preservation potential, explaining their often obscure appearance in the field (Kjær et al., 2006). They can form in front of both surging and non-surging glaciers and due to that cannot be used independently as diagnostic landforms for surge-type glaciers (Evans and Rea, 1999, 2003).
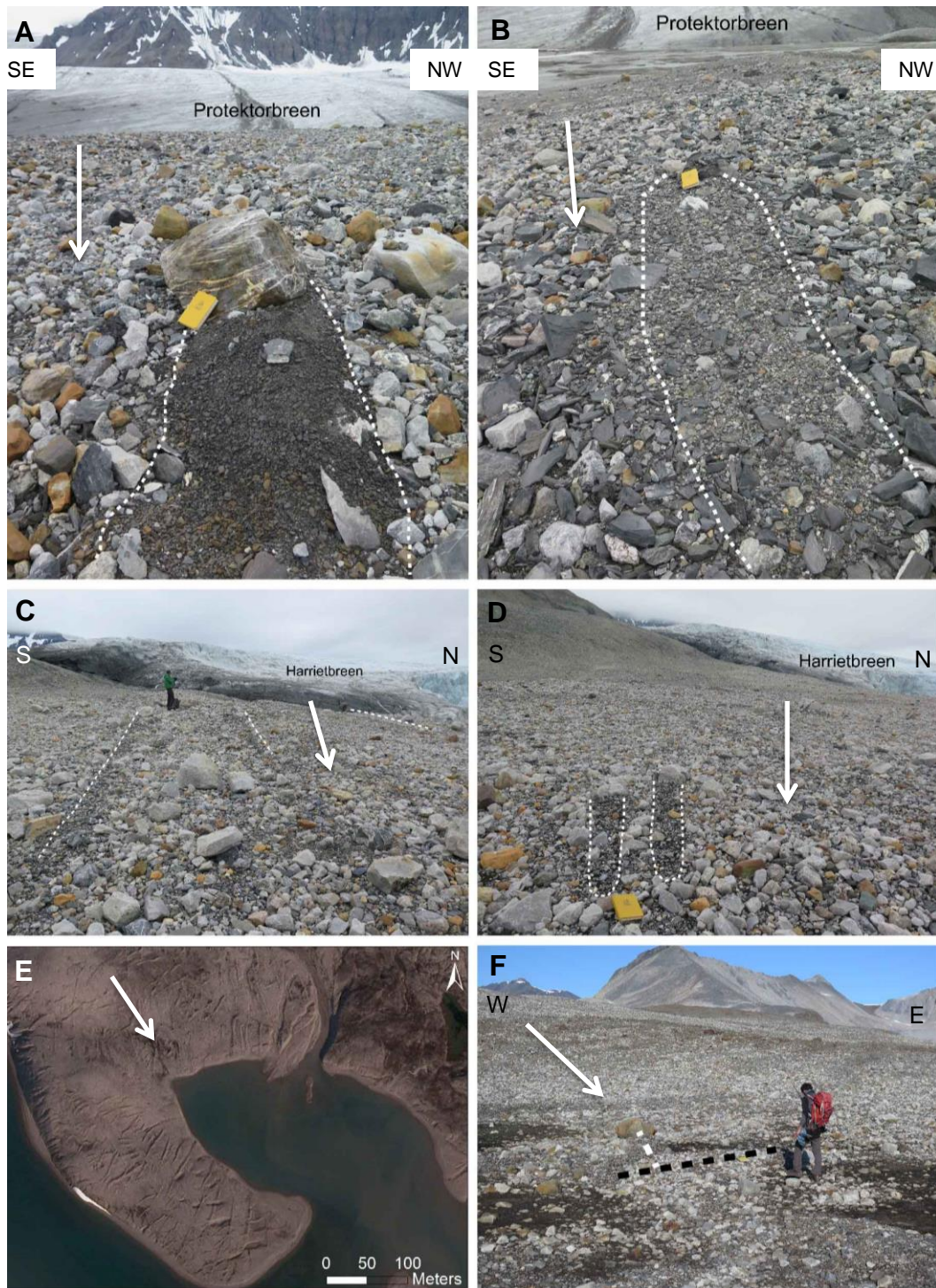
Figure 27. *Fluted till plain. A) A vague flute initiating from a small boulder in the forefield of Protektorbreen. The flute consists of finer grained diamict than the surrounding till plain. B) One of few prominent flutes initiating from a boulder in the forefield of Protektorbreen. Note high frequency of (sub)angular phyllite. C) Sub-parallel flutes are frequent on the till plain in front of Harrietbreen. D) Mini-flutings on the peninsula in Harrietbreen's forefield. E) An aerial image of the peninsula in the forefield of Kjerulfbreen. The till plain is covered by streamlined glacial lineations and CSRs, generally transverse to the former glacier flow are also very prominent. Kettle holes are also present. F) The only prominent flute in the forefield of Kjerulfbreen (white line), cross-cut by a CSR (black line). More CSRs are located in the background. The direction of generalized glacier flow is indicated with an arrow.*
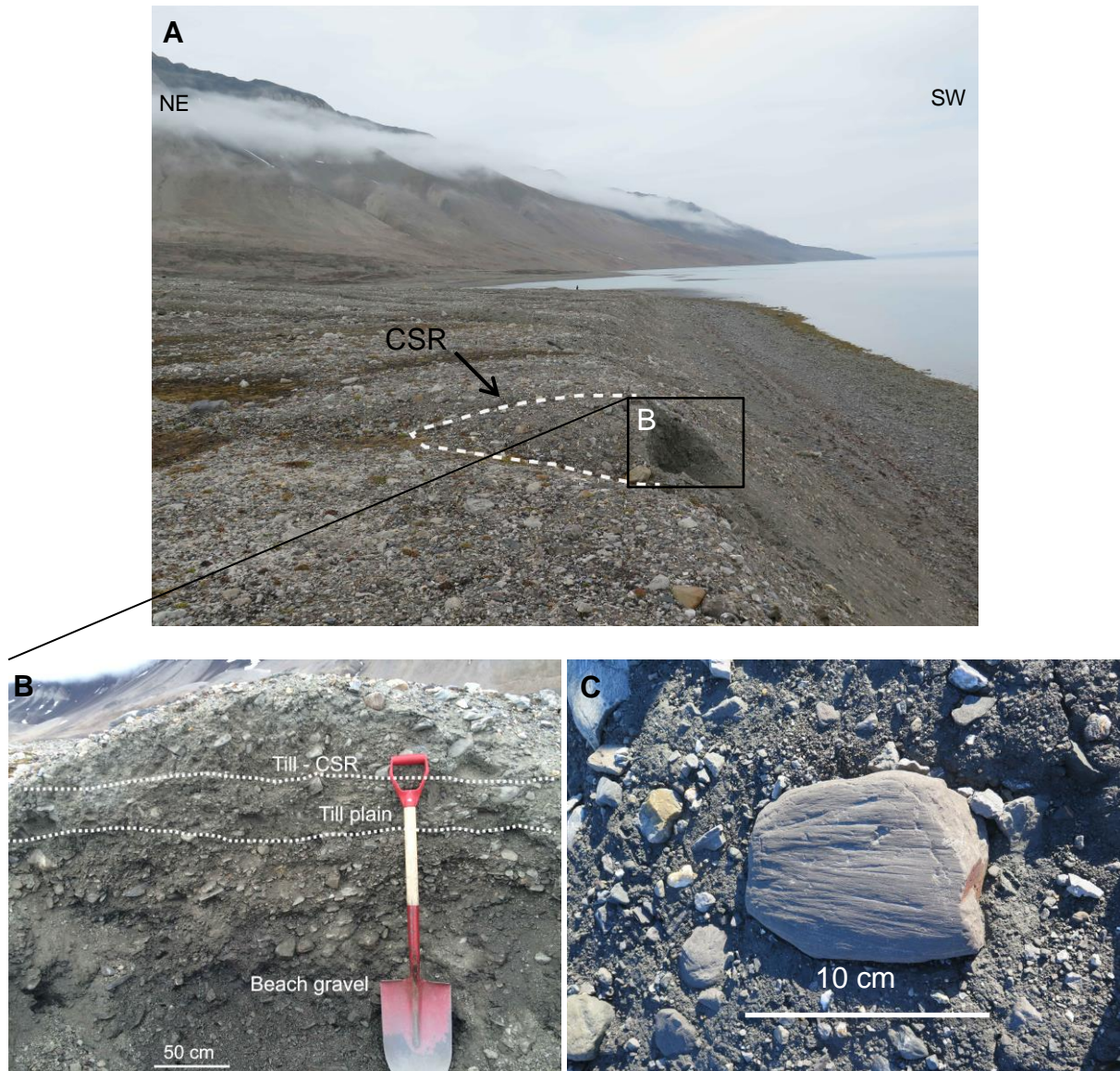
Figure 28. *Till plain in the forefield of Kjerulfbreen. A) Sediment cliffs by the coast eroding CSRs. The ridge is ~1 m long and the cliff ~2.5 m high. B) The section of the remnants of the CSR (Figure A). The top consists of till (CSR), vague boundary separates it from the subglacial till plain. Sharper boundary is to the underlying loose beach gravel that reaches all the way down the cliffs (photo credit: Daniel Ben-Yehoshua). C) Striated, subrounded sandstone clast on the subglacial till plain. Fine matrix and clast are around it.*

Bedrock: Debris covered and exposed bedrock
Description
Areas of bedrock are frequently exposed under the thin diamict cover, and where extensive fluvial erosion has occurred. This is shown as debris covered bedrock and exposed bedrock on the map with a point symbol or a polygon. In total ~0.16 km^2 were mapped as debris covered bedrock and ~0.17 m^2 as exposed bedrock (often occurring outboard the forefields).

The exposed bedrock is most prominent in front of Protektorbreen, covering in total ~0.079 km^2, where fluvial erosion has revealed large areas of bedrock all over the forefield, and in the distal part it is commonly exposed at the lee side of elongated hills, orientated, NW-SE

(Figure 24; 32B; 37B). In the distal part of the Harrietbreen forefield, covering in total ~0.021 km^2, bedrock is also visible at the lee side of streamlined hills with W-E orientation. The proximal part is characterized by similar streamlined topography except with orientation NW-SE (Figure 25). Several areas with debris-covered bedrock are present in the forefield of Kjerulfbreen, covering in total ~0.061 km^2. The bedrock is frequently exposed at the lee side of hills or associated with fluvial erosion. Dolerite dyke lies parallel to the western side of Lovénvatnet with erratics of quartz-rich sandstone on top (Figure 26; 29A). Bedrock is also observed on the islands in the fjord sticking out from the discontinuous debris cover.

Polished surface and striations were noted on the bedrock on few places in the forefields of Protektor- and Kjerulfbreen and close to Lovénvatnet. In the forefield of Protektorbreen they have an average direction of WWS-NNE (70°) (Figure 24; 29C). In the forefield of Kjerulfbreen two main directions were observed, roughly NW-SE (102°/130°) and N-S (188°). Where both directions were observed, N-S appeared deeper and wider, underlying the smaller and lighter striations (Figure 26; 29D). At the northern side of Lovénvatnet striations with the direction roughly N-S (155°) were observed.
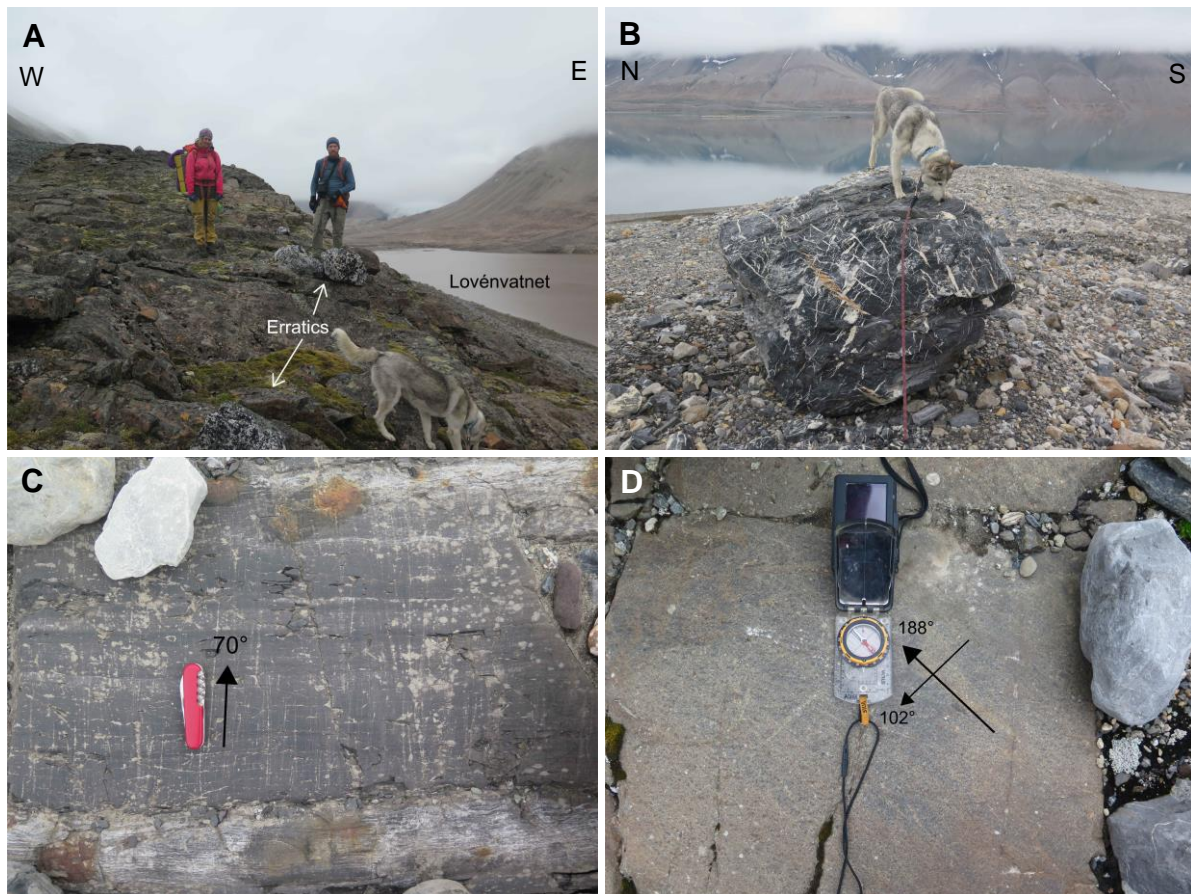
Figure 29. *Scoured bedrock. A) Dolerite dyke on the western side of Lovénvatnet with erratics on top. B) A subangular phyllite boulder (~1.5 m high) located on one of the bedrock hills in the forefield of Harrietbreen. C) Glacial striations in front of Protektorbreen, average direction is WSW-NEN (70°). D) Glacial striation with two directions in the forefield of Kjerulfbreen, NW-SE (102°; lighter) and N-S (188°; deeper and wider). The latter was only observed at this location and on the western side of Lovénvatnet but striations with NW-SE, were found at few locations.*

Interpretation

The bedrock hills are interpreted to be partly shaped by subglacial erosion based on their often scoured and polished surface. Thin layer of subglacial till was often deposited on top of them. Exposed bedrock is very prominent in all of the forefields and the topography therefore considered to be highly controlled by it. The curved shape of the hills in the Harrietbreen forefield is suggested to be due to the former morphology of resistant bedrock rather than glacier flow, indicating that the bedrock is fairly resistant.

Outboard of the areas with prominent fluvial erosion, the bedrock often shows evidence of glacial erosion in the form of striations and polished surfaces (Benn and Evans, 2010). The direction of the striations correlates fairly well to the present day glacier flow for Protektorbreen (SW-NE) and one of the sets from of Kjerulfbreen (NW-SE). The deeper striations in the forefield of Kjerulfbreen and by Lovénvatnet are suggested to pre-date the Neoglacial from time when the area has been covered with ice (Salvigsen et al., 1990; Forwick and Vorren, 2010). The direction of the striae indicates that ice has been flowing down through the valley occupied by Kiærbreen and Lovénvatnet today.

Geometrical ridges: Crevasse squeeze ridges (CSRs)
Description

A dense network of geometrical ridges covers most of the fluted till plain in the forefields of Harriet- and Kjerulfbreen (Figure 22). Around 1200 features were mapped in total, on an area of ~0.97 km^2. They are usually orientated transverse to the former ice-flow direction of the individual glaciers but few occur oblique to it. They range from 0.5-100 m in length but the average length is around 15 m. They are usually about 1 m wide and 0.5 m high but examples of 7 m wide and 2 m high do occur. Due to their often small dimension many are not recognized on the aerial images as the zoom level is set to 1:1500. The number of ridges is therefore considered much higher than the ones mapped.

In the forefield of Harrietbreen less than 400 ridges have been mapped on an area of ~0.37 km^2 (Figure 25). Their network is not as dense and complex and most of them are aligned N-S (Figure 31C) but patches with NE-SW orientation are also common. Usually they occur as single ridges but ridges intersecting each other and flutes were also observed. At the intersection the features usually fade out, making it difficult to examine the cross-cut relationship. Two locations did though reveal ridges on top of flutes (Figure 31A, B). They are as common on all slope angel and aspects, unlike Kjerulfbreen.

The network is denser in the forefield of Kjerulfbreen where over 800 have been mapped on an area of ~0.58 km^2 (Figure 26; 28E). They form a network where the ridges often seem to intersect at the end, forming complex patterns with variable orientation. The dominating orientation is however NNE-SSW or between 10°-40°. They are located on the till plain close to the coast and reach from the peninsula mid-way to the glacier, where their frequency starts to decrease and are absent closest to the glacier front. Their pattern is also denser on the lee side of bedrock hills (Figure 30A, B). Few examples of ridges lying directly on bedrock where observed (Figure 30E). On one location the ridges form a complex cross-cutting network where they are orientated in most direction. Their surface is hummocky and higher mounds of debris occur at the intersection (Figure 30D).

The ridges are almost absent in the forefield of Protektorbreen (Figure 24). Twelve ridges

have though been mapped on Protektorbreen's side, next to the medial moraine dividing the two forefields. Their network is not as dense as in the other forefields and they have an orientation of N-S (Figure 32A). Very vague ridges were also observed on aerial images in the forefield of Protektorbreen at two locations on the proximal side of bedrock hills. The ridges were not observed in the field. Their orientation is NW-SW (Figure 32B).

To examine their composition 10 cross sections in total were cleared and described from the forefields of Kjerulf- and Harrietbreen by Ben-Yehoshua (2017). Only the main observations from his work will be discussed here. The ridges consist of a matrix-supported diamict with a mixture of sand, silt and clay. The fraction of clay and silt is high (20-35%). It is compact and clast-rich, and even though it is matrix-supported the clasts are often in contact. Clasts are angular to rounded and up to 30 cm in diameter. Striations were noted on high fraction of the clasts, 58-74%. Shell fragments occur sparsely in the sediment. No structures were observed or preferred orientation of clasts but it was not possible to make fabric analyses because of the touching of clasts. The ridges were found resting on the underlying till plain, beach gravel (Figure 28B; 30C) or bedrock (Figure 30E).

Interpretation

These ridges are interpreted as crevasse squeeze ridges (CSRs). Network of diamict ridges, mainly orientated transverse or oblique to the former ice-flow direction are common in forefields of surge-type glaciers. CSRs are one of the main diagnostic criteria for surge-type glaciers (Sharp et al., 1985a; Evans and Rea, 1999, 2003; Schomacker et al., 2014).

Their formation is generally thought to be by upward infilling of saturated sediments into basal crevasses that form in association with the longitudinal and extensional stress during the surge. The process happens towards the termination of the surge and subsequent meltout (Rea and Evans, 2011; Lovell et al., 2015). The complex network of CSRs in the forefield of Kjerulfbreen indicates that the base of the glacier was heavily crevassed in multiple directions. The CSRs were observed lying on the subglacial till plain, flutes, beach gravel and directly on bedrock. Ben-Yehoshua (2017) therefore argued that the ridges are formed upglacier and transported englacially until the glacier stagnanted and the ridges melted out. Other studies have observed similar processes on Svalbard (Sobota et al., 2016).

Their preservation potential is usually fairly high in front of surge-type glaciers because of the stagnant glacier that wastes away in situ after the surge (Evans and Rea, 1999, 2003). However, fluvial erosion and burial by supraglacial sediments can decrease the preservation because of their small dimension. The abundance and characteristics of the sediment in the glacier's forefields is also a vital factor for their formation (Brynjólfsson et al., 2012, 2014; Brynjólfsson, 2015; Lønne, 2016). This will be discussed further in chapter 6 and the reason for the almost lack of CSRs in the forefield of Protektorbreen.

The diamict is interpreted as till, similar as the surrounding till plain based on the same characteristics (Krüger and Kjær, 1999). Sedimentological investigations are supportive of that. The shell fragments further indicate that it contains reworked material from the underlying beach gravel (Ben-Yehoshua, 2017).
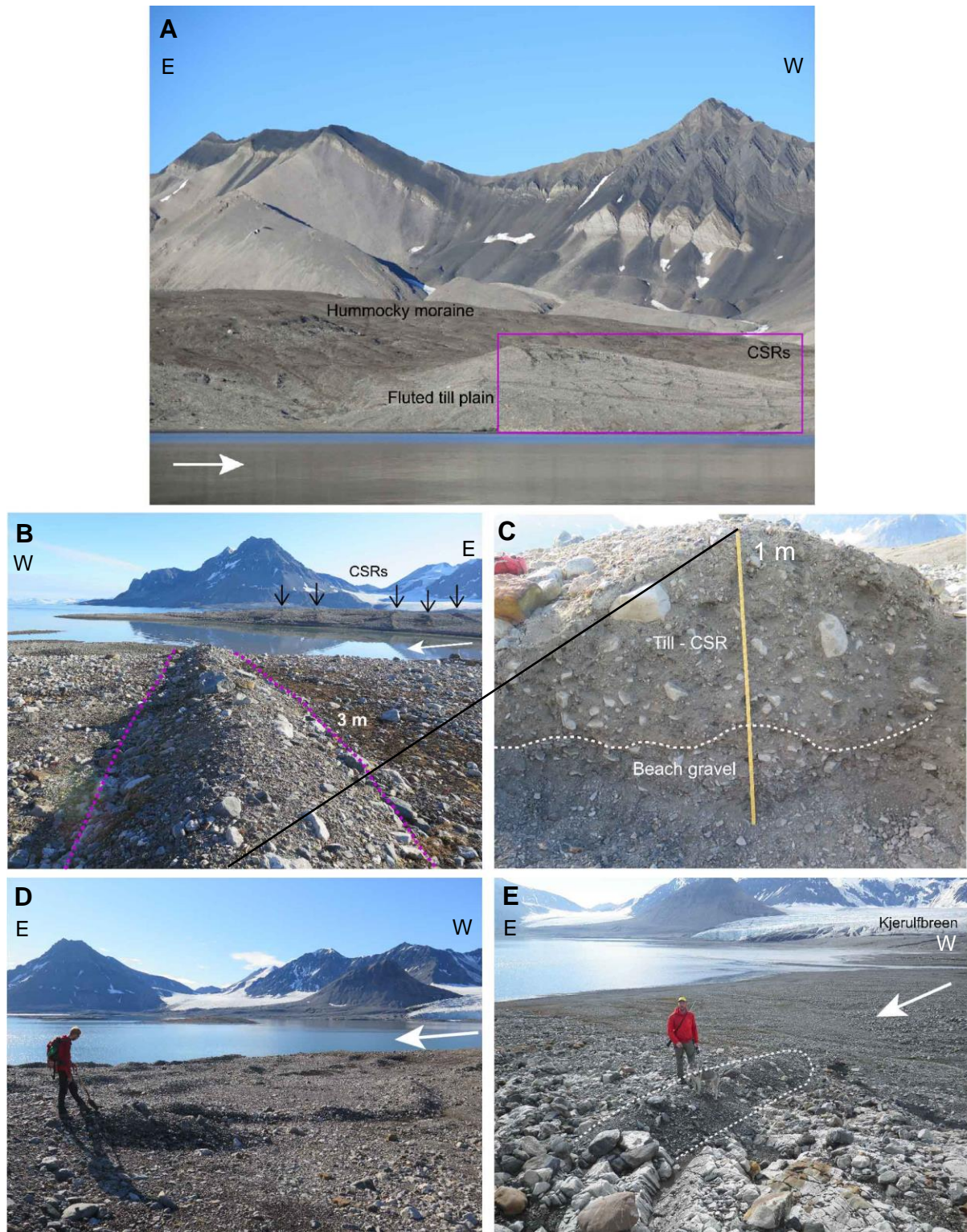
Figure 30. *CSRs in the forefield of Kjerulfbreen. A) A set of CSRs located on a lee side of a bedrock hill. They are located on a fluted till plain but no ridges or lineations are observed on the hummocky moraine above. The length of the box is ~0.5 km. B) CSRs by the bay (3 m long.) C) A cross section of the CSR on figure B). It consists of compacted, clast rich till with loose beach gravel underneath. The measuring stick is 1 m long. D) A complex network of cross-cutting ridges. Higher mounds occur at the intersection. E) CSR lying directly on bedrock (photo credit: Daniel Ben-Yehoshua). Large arrows indicate the generalized direction of glacier flow.*
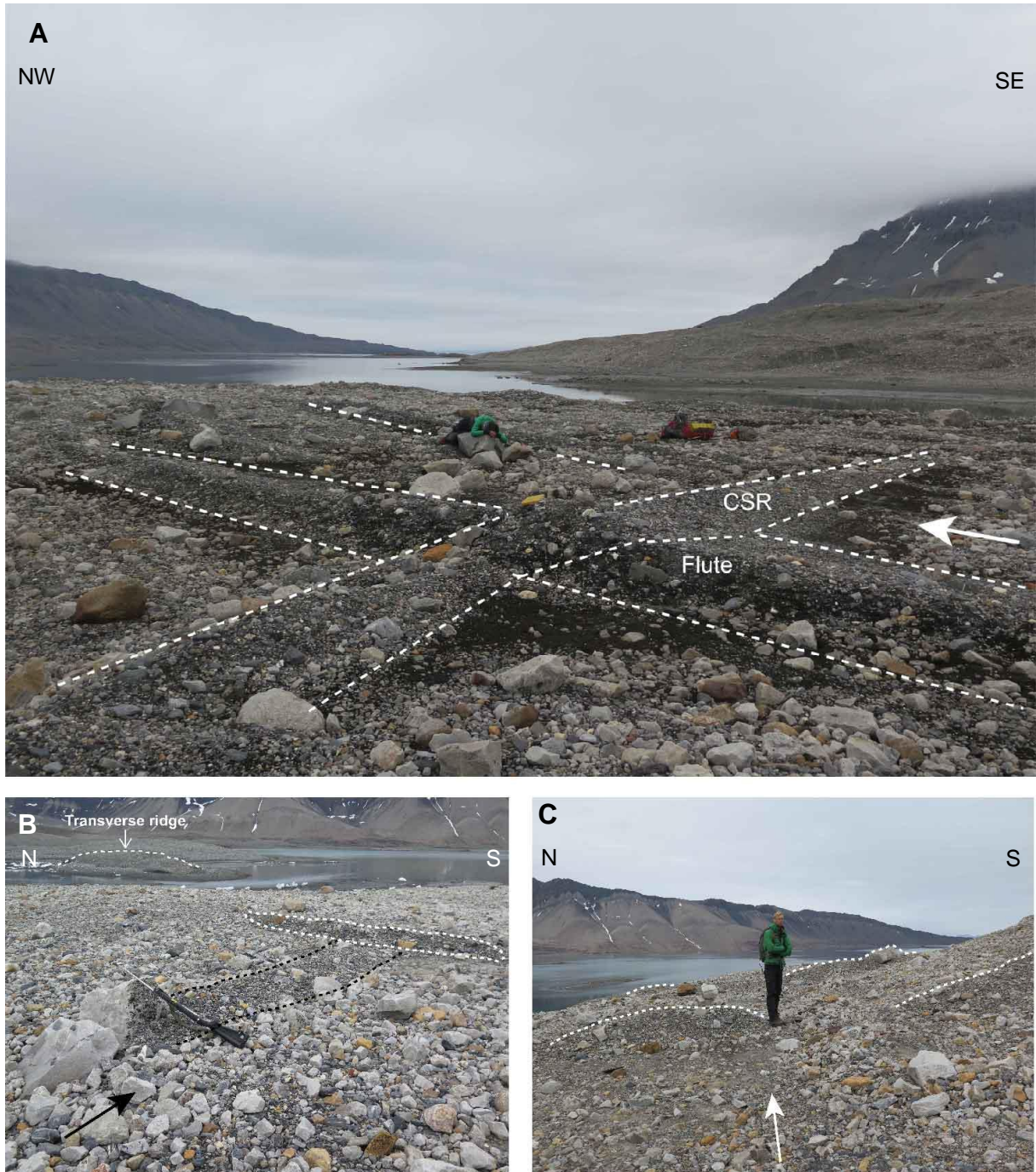
Figure 31. *CSRs and flutes in the forefield of Harrietbreen. A) One of few examples of the perpendicular cross-cut of a CSR and flute, where the CSR lies on top of the flute. B) Cross-cut of a CSR and a flute. No obvious difference could be detected between the two landforms, except their orientation. In the background is the transverse ridge right in front of Harrietbreen. The rifle is ~0.7 m high. C) Set of CSRs orientated transverse to the glacier flow direction. Note the large boulder in one of them. Large arrows indicate the generalized direction of glacier flow.*

Figure 32. *Aerial image of the forefield of Protektorbreen. A) The medial moraine marks the former intersection of the glaciers and the present day forefields (white line). Few CSRs are located on Protektorbreen's side close to it. The zoomed in box further shows them. Supraglacial debris bands are marked with black lines. B) Eskers can be seen going up the proximal side of a topographic hill, evolving into abandoned fluvial channels on their distal side that has cut down through the bedrock (red box). Very vague transverse ridges are seen on both the bedrock hills (red and purple box). On both images, high amount of fluvial activity surrounds the fluted till plain and bedrock is frequently exposed. Arrows indicate the generalize direction of the glacier flow (NPI, 2009).*

4.1.2 Supraglacial landforms

Hummocky moraine
Description
Changes in the surface cover were observed in the forefield of Kjerulfbreen adjacent to the terminal moraine (Figure 26). The whole area is dominated by small hummocks (1-5 m in height), small streams and lakes and higher frequency of vegetation than on the till plain. Striations were seen on some of the clasts but no lineations or ridges are present on the surface. This hummocky area covers ~0.80 km^2 in total. Two main areas were observed with slightly different characteristics, divided by the large river.

On the western side of the large river (in total ~0.33 km^2) the surface cover consists of a diamict with coarse, subangular to subrounded debris, ~5-30 cm in diameter, and matrix of silt to fine sand. The diamict is located on a gentle slope with lobes of sediments flowing downwards and low relief hummocky surface. Flatter areas with water-laid sediments and higher amount of vegetation are located between the hummocks. Fall sorting of coarse particles is associated with the hummocks (Figure 33A, B).

On the eastern side of the river (in total ~0.47 km^2) hummocks are present with higher relief than on the other side, especially proximal to the terminal moraine. Here also fall sorting and backslumping are associated with the hummocks. Their surface generally consists of a diamict. Clast rich diamict is most widespread but reddish clay rich diamict is also present. Shell fragments are often found on the surface and occasionally unidentified species of whole shells. Vague platforms are located between the hummocks and distal to the terminal moraine. Their surface often consists of sorted sand or gravel with patches of water-laid sediments, higher frequency of vegetation and small lakes (Figure 33C, D, E). Bedrock is frequently exposed where fluvial erosion has reached it (Figure 38B).

Interpretation
The area is interpreted as hummocky moraine based on the hummocky surface, solifluction lobes, fall sorting, backslumping and water-laid sediments. These features all indicate that re-sedimentation has occurred. Higher amount of vegetation and lakes and small streams further indicates that it is more water available in the area than on the till plain (Hambrey et al., 1997; Kjær and Krüger, 2001; Schomacker and Kjær, 2008). The wide range of material is interpreted to be a reworked subglacial till, glacilacustrine sediments and fluvial outwash and pre-Neoglacial beach and marine sediments and till. Their formation has been connected to thrusting, where material is brought into en- and supraglacial positions. No thrusting features were though detected from sections or from their morphology (Sharp, 1985a, 1988a; Hambrey et al., 1997). Substantial amount of sediments can as well be brought upwards from the glacier bed by intrusion trough crevasses. The stagnation of the glacier further leads to the formation of hummocky moraine, as the debris on the glacier surface can preserve the underlying ice if it is thick enough (Evans and Rea, 1999, 2003). Due to their dead-ice origin their morphology can be modified over time due to degradation (Schomacker and Kjær, 2007, 2008). Although dead ice was not observed it is not known if dead ice is still present in some of the highest hummocks but all the resedimentation possesses indicate that a significant amount has already degraded (Kjær and Krüger, 2001). Hummocky moraine

appears in front of both surging and non-surging glaciers and due to that cannot be use independently as diagnostic for surge-type glaciers (Evans and Rea, 1999, 2003).
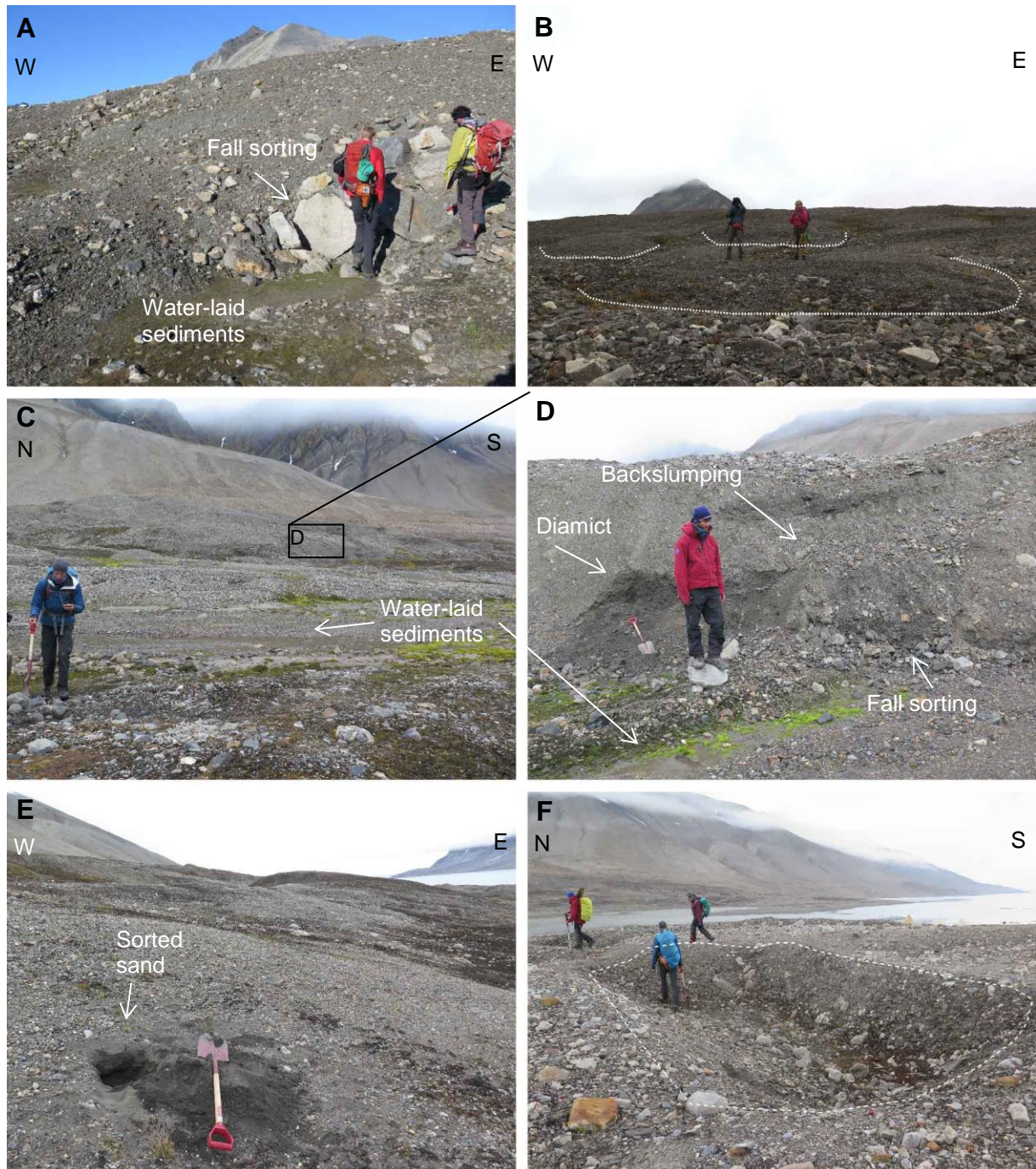
Figure 33. Hummocky moraine in the forefield of Kjerulfbreen. A) Hummock on the western side of the river with fall sorting at the edge and water-laid sediments in front with higher amount of vegetation. B) Solifluction lobes (white dashed line) flowing downhill from the terminal moraine on the western side of the river. C) Hummocks in the background with flatter platforms in front of them. D) A distinct edge on one of the hummocks (Figure C). Fall sorting and backslumping is present. The surface of the small section consisted of diamict. Water-laid sediments and vegetation are present. E) Sorted sand or gravel were common on the flatter platforms and smaller hummocks. The shovel is ~1 m high. F) A kettle hole on the peninsula (Figure 27E). Vegetation and finer materials can be seen in the bottom.

Depressions in the till plain: Kettle holes
Description

At the peninsula in front of Kjerulfbreen eight depressions occur with almost a rectangular shape. They have a dimension up to 20 m long, 7 m wide and 2 m deep. Their sides are covered with coarser material than the surrounding till plain and no cracks observed on them. Vegetation and finer material only appear at the bottom of the depressions (Figure 26; 27E; 33F).

Interpretation

These depressions are interpreted as kettles holes formed by the melt-out of isolated buried blocks of glacier ice. As the glacier got stagnant and retreated ice blocks can get buried under the debris from the glacier surface that will prolong their melting. When they have melted completely, depressions are left behind that often fill of water shortly after. That explains the finer sediments and vegetation in their bottom (Benn and Evans, 2010).

Medial moraine
Description

At the joint of the forefields of Harriet- and Protektorbreen, right below Knuvelen, are three parallel debris bands. They are very distinct close to the coast but get vaguer closer to Knuvelen. The first debris band consists of brownish sandstone with smaller clasts size than the surrounding lighter sandstone and the next two consist of cobbles of phyllite. The clasts are angular to subrounded. The debris bands have an orientation close to W-E (Figure 25; 34A, B). Supraglacial debris can also be seen on the glacier surface, at the intersection between Harriet- and Kjerulfbreen and at the eastern margin of Kjerulfbreen (Figure 22). On both places, darker debris than the surrounding till plain can be traced from them down onto the till plain. In total ~0.059 km^2 were mapped.

Interpretation

These debris bands have all been interpreted as medial moraines based on their location between both the active and former junction of the glaciers, and at their margin.

The debris bands between Harriet- and Protektorbreen have been interpreted as remnant of an interaction medial moraine formed by the lateral compression caused by the confluence of the two glaciers. They generally consist of medial debris septa with variable clasts characteristics and are considered to be both of supraglacial and basal origin, transported in englacial position to the surface and with longitudinal foliations (Eyles and Rogerson, 1978; Hambrey et al., 1999). The shape of the debris bands is dependent on the relative strength of the flow units (Eyles and Rogerson, 1978). The orientation is clearly different from supraglacial debris bands in both forefields (Figure 22), indicating that flow from both glaciers affected that area to some extent. However, since the orientation is closer to the ice-flow from Harrietbreen it agrees with what is seen in the submarine environment, that Harrietbreen was more active than Protektorbreen during their last advance. The vague color difference that can be traced down from the medial moraine between Harriet- and Kjerulfbreen and at the side of Kjerulfbreen is as well interpreted as the remnant of the medial moraine. As their preservation potential is low they are not always easy to distinguish it in the field after the glaciers have retreated (Glasser and Hambrey, 2003).
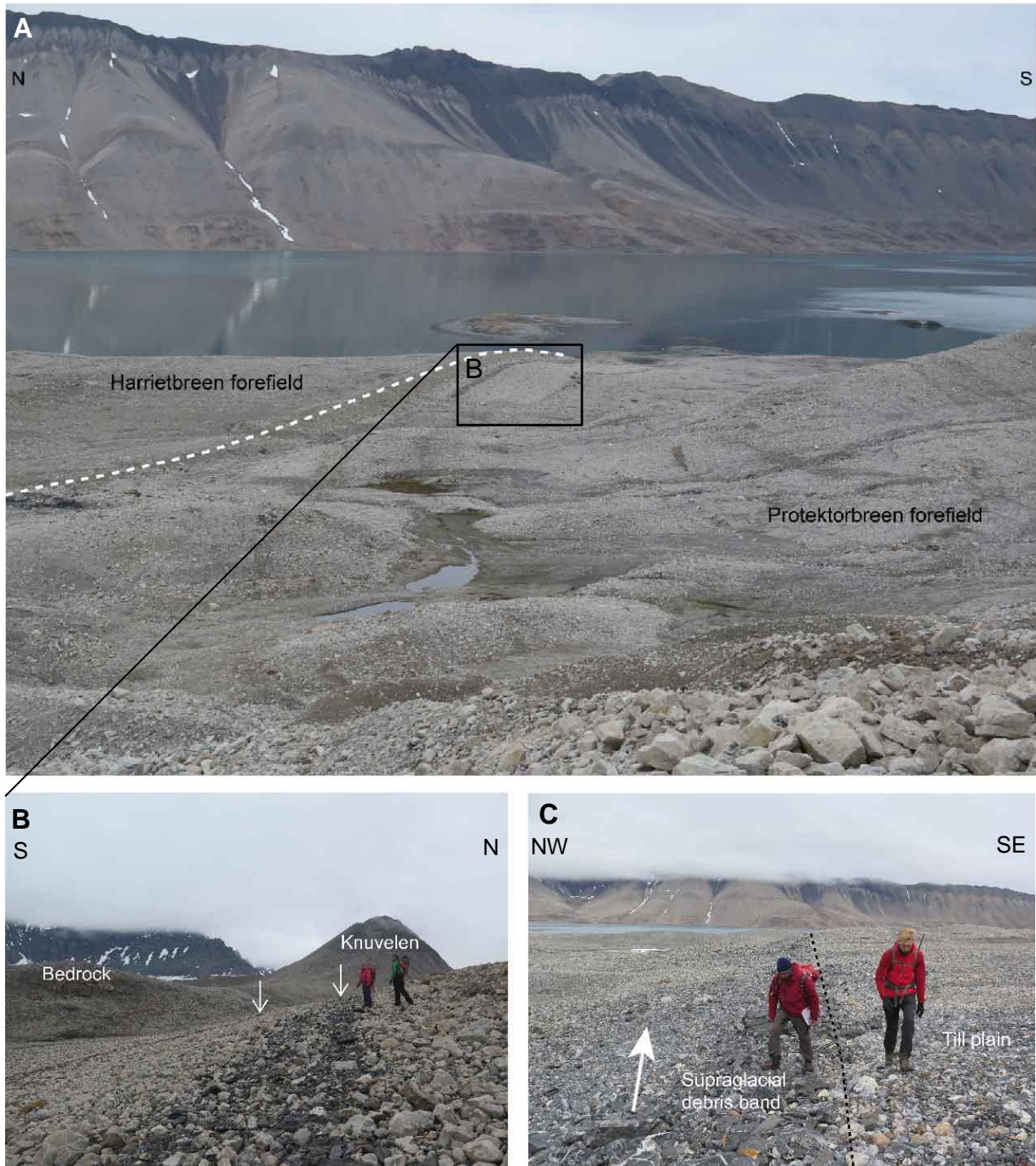
Figure 34. *Medial moraine and supraglacial debris bands. A) Large, angular sandstone boulders are located under Knuvelen at the intersection between Protektor- and Harrietbreen forefields (black dashed line). Below are three prominent supraglacial debris bands, interpreted as interaction medial moraine. Supraglacial debris bands are also seen in the forefield of Protektorbreen. B) Two of the supraglacial debris bands in the medial moraine. Bedrock is exposed at the less side of topographic high in the background. C) Supraglacial debris band in the forefield of Protektorbreen. The person to the left stands on supraglacial material, consisting of more angular clasts of phyllite but the one the right on less angular clasts with variable lithology interpreted to be subglacial till plain. The large arrow indicates the generalized direction of the glacier flow.*

Supraglacial debris bands
Description
Debris bands with different sediment surface than the underlying till plain and orientated approximately in the ice-flow direction are present in all three forefields (Figure 22). They appear darker on the aerial images and the one examined in the field consisted of angular to subangular cobbles of phyllite (Figure 34A, C; 37A). The bands are around 0.3-1 m wide and can be up to 500 long. Sometimes the bands cover wider areas but for easier visualization they are drawn as lines on the map. They can often be traced to supraglacial debris cover on the glacier, which was observed to have the same composition (Figure 38A). Some of the debris bands on the glaciers could be traced all the way to the debris source but others started in the middle of the glacier. Piles of supraglacial material were also visible. They are most widespread in the forefield of Protektorbreen, with orientation WWS-EEN. In the forefield of Harrietbreen they have an orientation of WWN-EES and are also located close to the glacier margin although they extend further out. In the forefield of Kjerulfbreen they have an orientation of NNW-SSE and only occur close to the margin.

Interpretation
Theses band are interpreted as supraglacial derived debris bands as they consist of angular phyllite and can often be traced to supraglacial material on the glacier surface. They usually occur in longitudinal sets on the glaciers and reflect the location of persistent debris sources and ice-flow lines. Their composition can therefore often be traced to the parent material. They can be derived from local rockfall, glacier erosion from the mountain side or elevated basal debris. Once the glacier retreats the debris bands melts out in the forefield, although their preservation potential is poor (Glasser and Hambrey, 2003).

4.1.3 Ice-marginal landforms

Terminal moraine
Description
Prominent ridges are located adjacent to the mountain sides in the forefields of Protektor- and Harrietbreen. No ridge was observed in the forefield of Harrietbreen (Figure 22).

An extensive ridge with roughly W-E orientation delimits the southern margin of Protektorbreen's forefield. The moraine ridge covers an area of ~0.22 km^2, is up to 60 m high, over 1 km long and almost 300 m wide (Figure 24; 35A). An investigation of the surface cover revealed that it consists of clast rich diamict with subangular to subrounded clasts, cobbles up to small boulders of variable lithology. The matrix is a mixture of sand, silt and clay but most fines have been washed away from the surface. The matrix alters though a bit between locations and parts contain angular to subangular boulders of sandstone (Figure 35B, C). The internal structure of the ridge was not examined since natural sections did not exist and it was not possible to excavate any because of the coarse grain size. The thickness of the surface cover is therefore unknown. The northern and middle part the surface can be described as hummocky and cracks and sediment lobes are visible (Figure 35C). Few small ridges are present on it, usually orientated parallel to the terminal ridge but oblique ridges also occur. They consist of higher frequency of phyllite than the surrounding and cracks are visible on the surface. Vegetated areas and solifluction are present between them and the

hummocks. The rest of the surface is covered with linear structures that lie mostly parallel to the terminal ridge, although they curve slightly (Figure 35B). A small lake is located there as well and abandoned channels, outwash plain and small hummocks are in connection to the moraine (Figure 37B).

In the forefield of Kjerulfbreen, a prominent ridge lies adjacent to the mountain side and the southern side of Lovénvatnet. It curves towards the mountain side in east, where it fades into a lower relief ridge. In total the extensive part of the ridge covers an area of ~0.19 km^2, is over 2 km long, up to 100 m wide and 30 m higher than the surrounding. It is orientated NW-SE in the proximal part but curves towards W-E next to Lovénvatnet (Figure 26; 36A). The surface cover consists of a diamict, with smaller grain size and lower frequency of angular clasts than the ridge described above. As well higher frequency of matrix is visible on its surface. Few cracks were noted along its side (Figure 36B). A section on the southern side of the moraine, south of Lovénvatnet reveled matrix supported diamict. The matrix was composed of silt to coarse sand. Few clasts were observed, generally rather small (<5 cm). Higher frequency of clasts was on the top of the section and the bottom, that ranged up to 20 cm. Clasts were subangular to subrounded and striations observed on them. Lenses of sorted gravel and sands were observed on few places. Shell fragments were rather common and one intact shell was found. Slumping was around the section and above it (Figure 36C).

Interpretation

The ridges are interpreted as terminal moraines and considered to mark the maximum Neoglacial extent of the glaciers (Glasser and Hambrey, 2003) (Figure 47A, B). In high alpine environments the dumping of large amounts of debris around the glacier margin forms large lateral moraines (latero-frontal dump moraines). The debris can originate from both supraglacial materials on the glacier and from local rockfall near mountain sides. They usually consist of coarse debris with high angularity and little matrix. Based on their position and coarse, angular surface cover they are at least partly considered to form during similar processes. The reason for the larger grain size and higher frequency of angular boulders on the Protektorbreen moraine is the local lithology and proximity to the mountain side (Boulton and Eyles, 1979; Benn and Evans, 2010). The shell fragments and layers of gravel and sand in the Kjerulfbreen moraine, indicates that the material contains reworked beach gravel as was described from the hummocky moraine and underlying the till plain and CSRs.

The hummocky surface, cracks and the large dimension are all indicative that the one in front of Protektorbreen is ice-cored to a large extent. The cracks, solifluction lobes and hummocky surface under Kjerulfbreen moraine indicate that it is ice-cored to some extent and dead ice-melting has occurred but based on the size difference it is considered to contain less ice (Schomacker and Kjær, 2008). The curved ridges and linear structures on Protektorbreen moraine are suggested to be old flow structures from the glacier, preserved in englacial debris bands sitting inside the dead-ice. As the dead-ice slowly melts these bands become visible and as well get modified as the sides creep downslope. This also explains why they have a slightly different surface cover than the surrounding (Ole Humlum, personal communication).
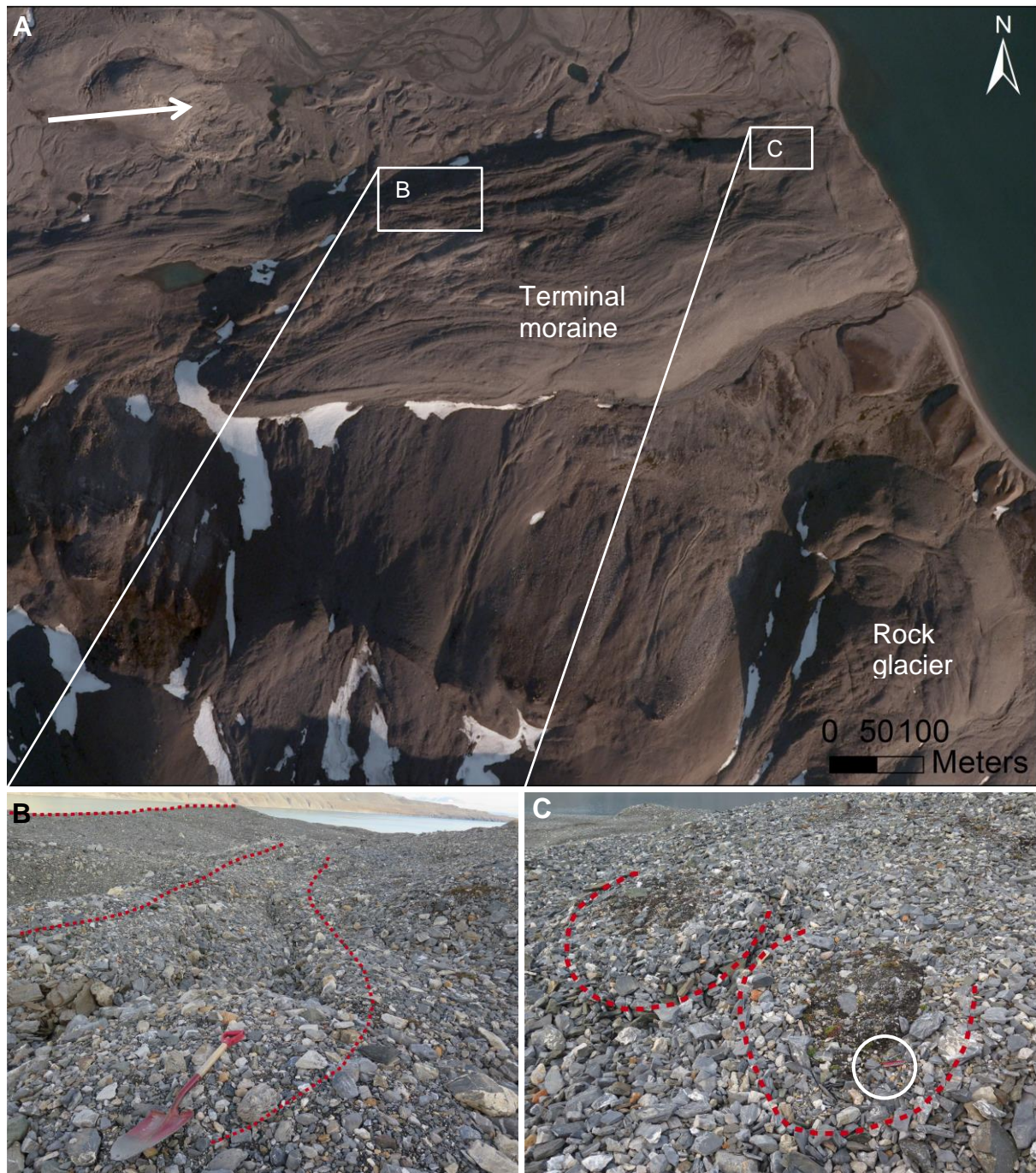
Figure 35. *Terminal moraine - Protektorbreen. A) An aerial image of the ice-cored terminal moraine and rock glacier outboard the forefield. Note the curved ridges and lineations on the surface. The arrow indicates the generalized direction of glacier flow (NPI, 2009). B) The surface cover contains higher frequency of coarse, angular clasts than the till plain and less matrix. Slightly curved parallel, ridges and lineations are seen (red lines). The shovel is ~1 m high. C) Small solifluction lobes on the side of the moraine. Note the knife for scale (~7 cm).*
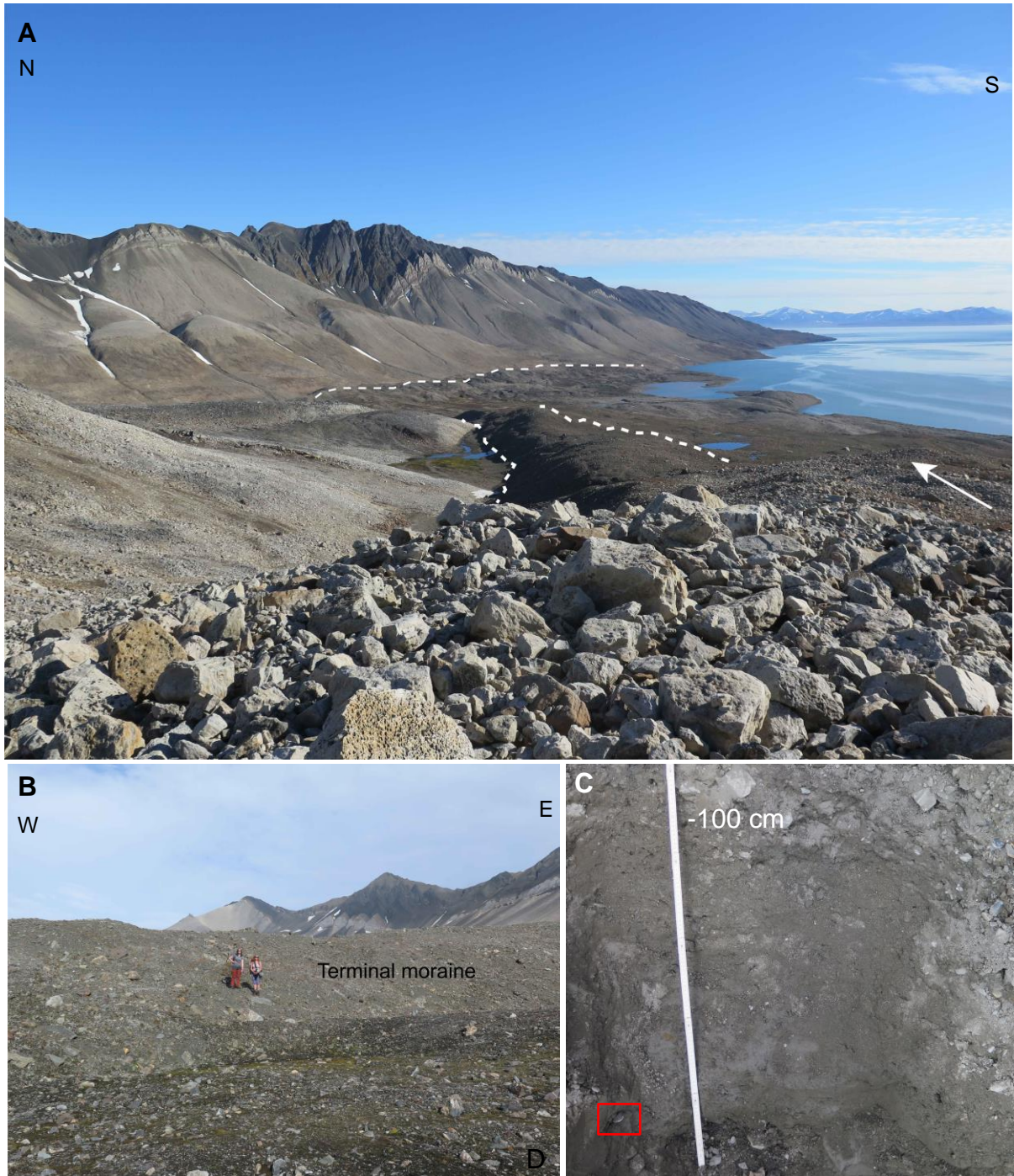
Figure 36. *Terminal moraine - Kjerulfbreen. A) Sandstone boulders are located close to the glacier margin, classified as part of the till plain. The terminal moraine (white dashed line) delimitates the forefield and low relief hummocks, composed of diamict are on its southern side. The arrow indicates the generalized direction of glacier flow. B) The terminal moraine was easily distinguished from the hummocky moraine by its ridge shape and lack of vegetation. C) A section in the terminal moraine. It consists of a matrix-supported diamict with layers of sorted sand and gravel. Shell fragments are common on one whole shell was found (red box). No structures were observed. Slumping was on the sides and above it.*
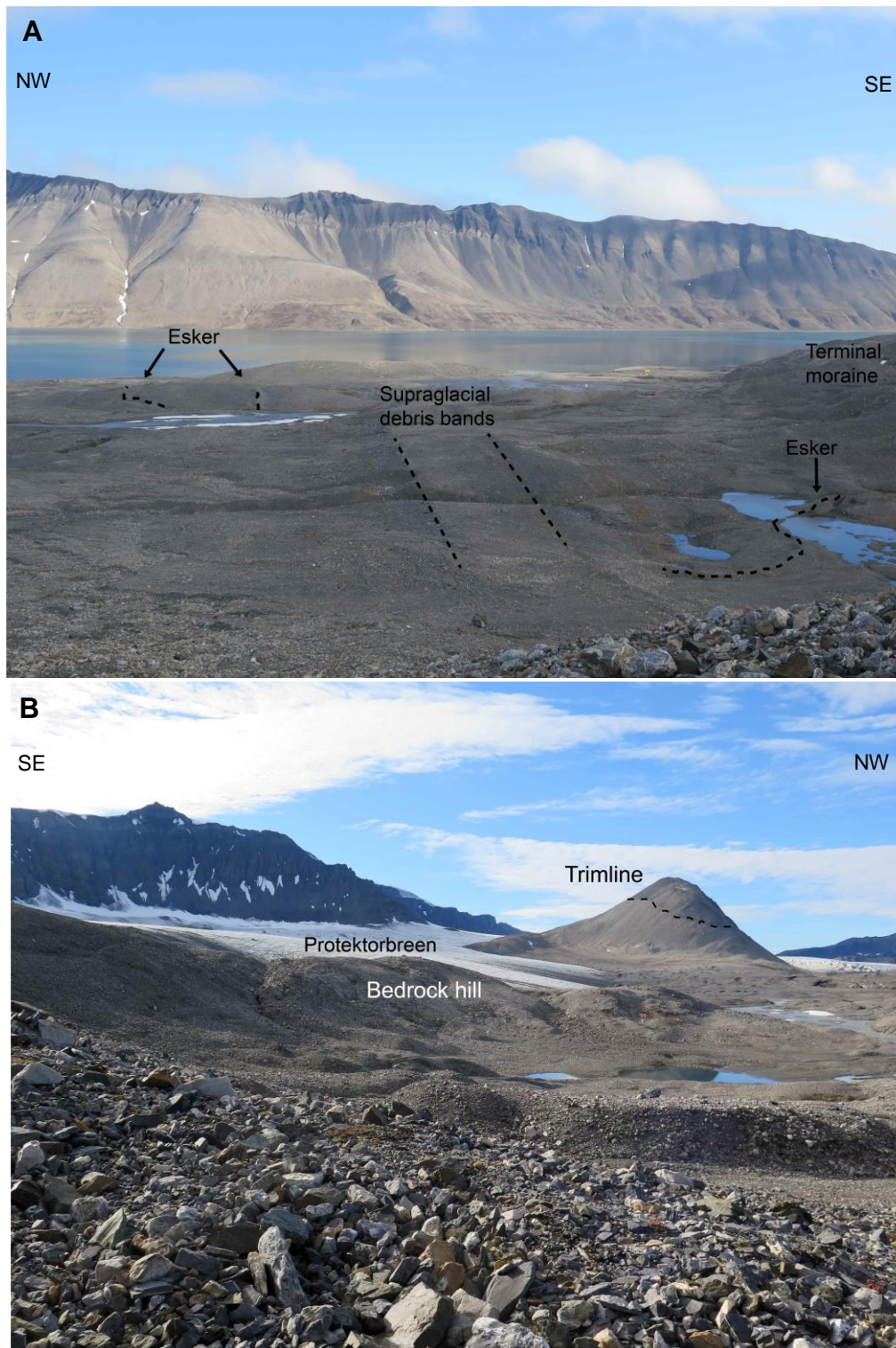
Figure 37. *An overview of the forefield of Protektorbreen. A) The southern part seen from the glacier margin. Prominent change can be seen between the fluted till plain and the darker and more angular supraglacial debris bands. Two eskers can be traced going up a bedrock hill and one affected by fluvial erosion. B) Debris covered bedrock hill, dissected by inactive fluvial channels is located in the middle of the figure and hummocky terrain is located in front of it. The trimline on Knuvelen is in the background.*

Transverse ridge: Push moraine
Description
Two ridges, 40 and 90 m long and 20 m wide were observed on the aerial images, transverse to the glacier margin of Harrietbreen. One ridge is located on a small spit connected to the till plain right in front of the margin (Figure 25; 26; 31B) and cannot be traced further on to the till plain. The other one is on the northern side of the Harrietbreen margin. It is surrounded by fluvial outwash plain and streams and is partly eroded. As they are located on an area only accessible by a boat they were not examined in detail.

Interpretation
The ridges are interpreted as small push moraines formed during minor readvance and therefore indicate that this part of the margin was still depositing or pushing material during their deposition (Evans and Twigg, 2002). No other similar ridges are observed in the terrestrial setting but the fluvial activity around them suggests that more ridges might have formed that have been eroded. However, a set of transverse ridges will be described in the submarine environment (*section 4.2.2*) and their connection discussed later.

Trimline
Description
Clear and sharp changes in rock color can be seen on the south-eastern, 140-280 m a.s.l, and north-eastern, 60-100 m a.s.l., sides of Knuvelen and north of Kjerulfbreen, 180-280 m a.s.l. The slope below the line has fresher appearance but is more weathered above it (Figure 25; 37B).

Interpretation
These limits have been interpreted as trimlines. They form by the upper limits of valley glaciers that erode the slope beneath them, leaving a distinct difference in their appearance (Ballantyne et al., 1997). They are therefore considered to be from the maximum Neoglacial extent and demonstrate the vertical loss of ice since then (Figure 47A).

4.1.4 Glaciofluvial landforms

Fluvial activity
Description
The amount of fluvial activity and erosion is variable between the forefields but in total ~1.03 km^2 have been mapped. The features classified under fluvial activity are fluvial outwash plains and active, seasonal or abandoned fluvial channels. Numerous channels are observed but only a fraction was mapped. The surface material of the fans consists of sorted sand and gravel, generally with higher frequency of rounded clasts than on the surrounding till plain. Active and inactive channels are located on it. Individual channels are usually orientated transverse to the glacier front or follow the topography. As previously mentioned, bedrock is often exposed on the fluvial outwash plains and in the abandoned fluvial channels.

The fluvial activity is most extensive in the forefield of Protektorbreen where ~0.47 km^2 or close to 50% of the forefield is marked with fluvial outwash plain (Figure 24). Active fluvial drainage is high although it is mostly concentrated into one big stream. A set of sub-parallel

channels is located on the northwestern margin of the glacier front. Individual channels generally occur in connection to fluvial outwash plains (Figure 38A).

Fluvial activity and erosion is not as dominant in the forefield of Harrietbreen and covers ~0.13 km^2 (Figure 25). Most of the water drains trough two active channels situated in the forefield, transverse to the glacier margin and at the western margin of the forefield, where a small lake forms during melting seasons. Numerous seasonal or abandoned channels are also present associated with fluvial outwash. A large river and fluvial outwash plain, building out as a delta in the ocean, are also situated on the northern side of the glacier between Harriet- and Kjerulfbreen.

In the forefield of Kjerulfbreen the fluvial activity is limited to restricted areas. In total it covers ~0.43 km^2 (Figure 26). Large stream drains Lovénvatnet and the area with hummocky moraine to the ocean. Fluvial outwash plain and channels cover the whole front of the glacier. Set of sub-parallel channels is located on exposed bedrock at the eastern margin of the glacier. Fluvial outwash plains are located at several locations by the beach, cutting the fluted till plain. In most cases they are seasonal or abandoned. Few abandoned channels with SW-NE orientation are located on the western side of the large stream. On the eastern side of it are number of deep, abandoned fluvial channels, orientated roughly N-S. Some active streams and outwash plain are located in connection to them (Figure 38B).
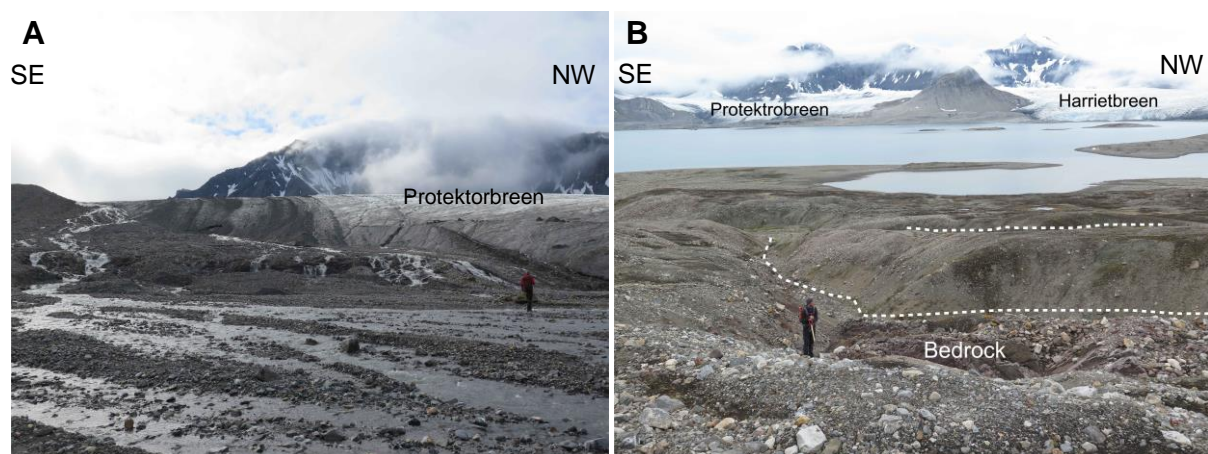
Figure 38. *Fluvial activity. A) High fluvial activity at the southern margin of Protektorbreen, as in most of the forefield. Supraglacial debris can be seen on the glacier surface. B) Deep seasonal or abandoned meltwater channels frequently dissect the hummocky moraine in the forefield of Harrietbreen and the other forefields, often exposing bedrock (white dashed lines).*

Interpretation

Glaciofluvial deposits are common in glacial landscape due to seasonal melting but are variable after the environments. They can be deposited trough sub- and englacial conduits and by supra- and proglacial streams, that carries great quantities of sediments with them on their way to the ocean (Benn and Evans, 2010). The most active outwash fans and channels are in the forefield of Protektorbreen and around the intersection between Harriet- and Kjerulfbreen, which indicates high amount of meltwater flowing between the junction of the glaciers. The seasonal or abandoned channels are interpreted to be either subglacial and/or

lateral meltwater channels. Lateral meltwater channels often form at the margins of polythermal or cold based glaciers, when water cannot penetrate trough the frozen bed (Dyke et al., 1993).

Sinuous ridges: Eskers
Description
Several sinuous ridges were observed in the forefields of Protektor- and Harrietbreen (Figure 24; 25).

In the forefield of Protektorbreen they appear in three areas. The one at the southern side of Protektorbreen's margin is orientated W-E and around 0.5-2 m high and 30-40 m long. Fluvial erosion has eroded parts of it (Figure 39A). Ridges are also located at the proximal side of a bedrock hill, going up it (Figure 37A). They are orientated roughly SE-NW but one smaller cross-cuts it almost perpendicular. The ridges are up to 60 m long, 1.5 m high and 1 m wide. At their distal side they are connected to a relatively deep, abandoned fluvial channel that has cut through the diamict cover down to the bedrock (Figure 32B). The surface cover has finer matrix than the surrounded diamict and cracks are visible on the surface. Sections in the ridges revealed that they are composed of 5-10 cm horizontal layers of sorted sediments, matrix-supported, silt to medium sand, and clast-supported, fine gravel to 30 cm boulders. The clasts are subangular to rounded (Figure 39D).

In the middle of the forefield of Protektorbreen a 50 m long, 2 m wide and 0.5 m wide ridge (Figure 24) was mapped with a surface cover consisting of subrounded to well rounded cobbles of sandstone and phyllite. The cobbles ranged from 2-15 cm in diameter. No fine matrix was observed. Its orientation is SW-NE. At its distal side the ridge evolved into a circular mound up to 1 m high. Beyond that it changed into a smaller ridge, consisting of finer grained rounded clasts going up a topographic hill towards north. On the western side of the ridge is a smaller mound of finer grained, rounded clast (Figure 39E, F).

In the forefield of Harrietbreen six features were mapped (Figure 25). They all lie on a gentle slope with an orientation of roughly N-S. They are up to 50 m long, 1 m high and wide. Detailed investigation of their composition was not conducted but their surfaces consist of sand and subrounded to rounded gravel to cobbles (Figure 39B).

Interpretation
The ridges are interpreted to be eskers formed by glaciofluvial activity based on their sinuous form and sorted sediments. They are infillings of ice-walled river channels and can form in sub-, en- or supraglacial positions (Warren and Ashley, 1994). They have variable morphology and internal structure, as topography and ice thickness also plays an important role for their formation.

The eskers in the forefield of Protektorbreen are interpreted to be subglacial or englacial features, based on their stratified sand and gravel layers. If water flows in over-pressurized conduits under the glacier they can flow upwards, explaining the ridges going up the topographic hills. They are as well often correlated with Nye channels eroded into the sediment and bedrock as was observed (Benn and Evans, 2010). The cracks observed on one of them indicated dead-ice melting below the deposits, which can modify the esker

morphology over time. That also indicates that it was deposited englacial (Evans and Twigg, 2002; Ólafsdóttir, 2011).
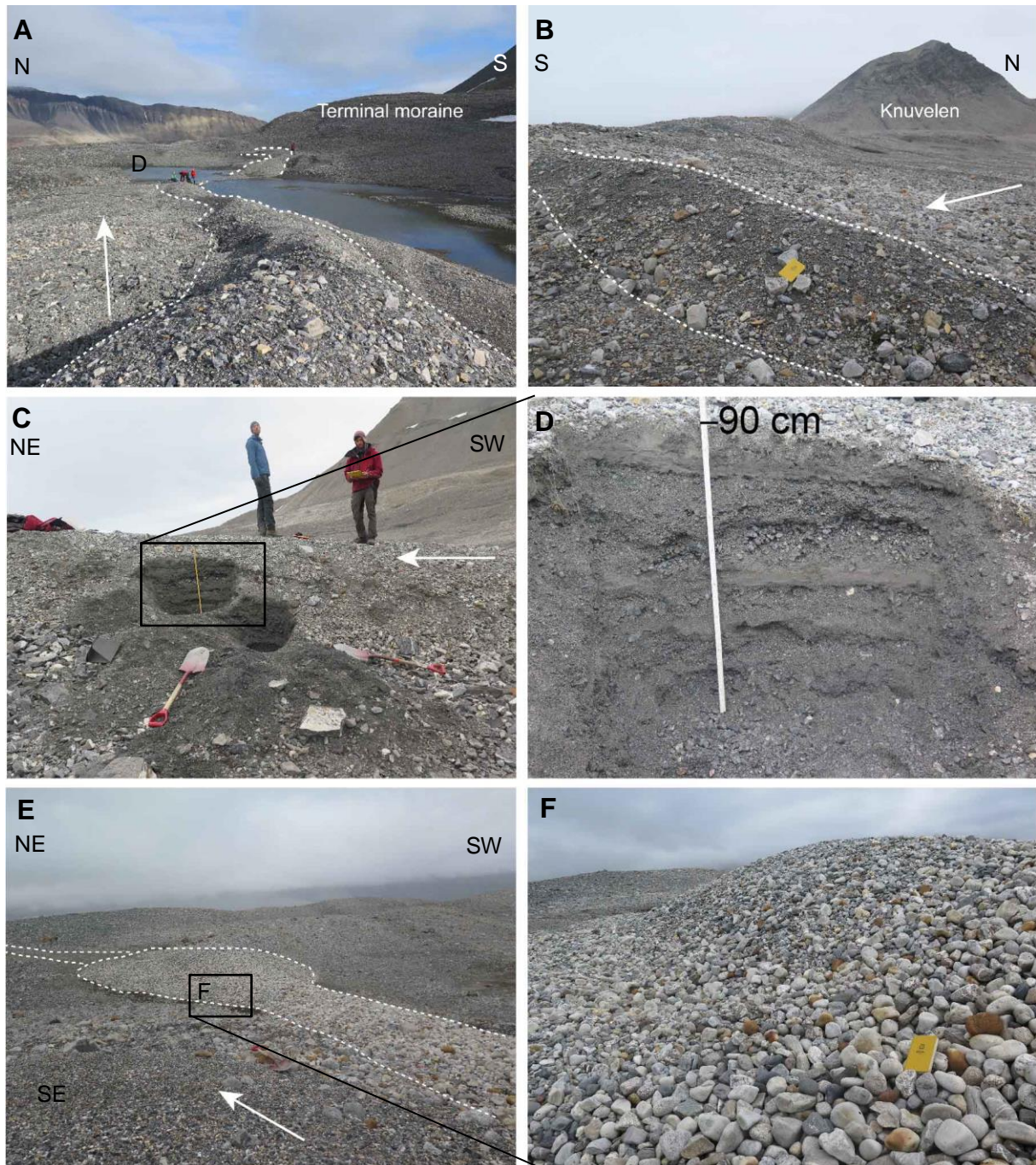
Figure 39. *Sinuous eskers. A) Esker with a sinuous shape, eroded by fluvial activity at the southern margin of Protektorbreen. B) Supraglacial esker located at a gentle slope in the forefield of Harrietbreen. C) An esker in the forefield of Protektorbreen going up a topographic hill. D) A cross section of that esker. It consists of horizontal layers of sorted sand and gravel, mostly subrounded to rounded. E) A ridge evolving into a circular mound of subrounded to well rounded cobbles in the forefield of Protektorbreen. F) A close up of the mound showing the sorted rounded cobbles. The note book is 20 cm long and the shovel ~1 m. The arrows indicates the generalized glacier flow direction.*

The roundness and sorting of cobbles in the ridge and mound in the forefield of Protektorbreen (Figure 39E, F) indicate that it was formed by fluvial erosion (Warren and Ashley, 1994). However, this ridge does not have the common morphology and composition of an esker. The larger clast size and amount of roundness indicate higher amount of energy transporting the debris than in the other eskers.

The eskers in the forefield of Harrietbreen are orientated more or less transverse to its margin. Their morphology at first sight looks very similar as the CSRs. However, they are longer than the average CSRs, with meandering form and consist of sorted gravel and sand material but not diamict as is in the CSRs. Based on that they were classified as eskers. Eskers that are orientated transverse to the ice margin are common for former ice-walled channels located in stagnant glacier margins. These landforms are termed supraglacial eskers (Boulton et al., 1999). As the glacier front was heavily crevassed due to the surge it is considered likely that some of them were also filled with sediments derived from supraglacial streams.

Glacilacustrine sediments and shorelines
Description
Sediment cliffs at the western side of the river contain clast-supported layers of coarse sand to gravel with occasional cobbles. The gravel and cobbles are angular to subrounded and often have a flattened shaped (Figure 40B). Below the cliffs and reaching over to the eastern side of the river, similar material was observed with a hummocky surface, vegetation and patches with fine sand. Another terrace was located above the cliffs, cut by abandoned fluvial channels with larger grain size at their end (Figure 26; 40A). In total ~0.086 km^2 were mapped.

A set of horizontal terraces was located around Lovénvatnet. The number of linear horizontal terraces ranged from 2-9, depending on location. The highest one was up to 50 m a.s.l. and could be traced over most of the eastern side of the lake. The surface cover of the terraces consists of coarse washed gravel similar as seen on the current lake beach. Cobbles of sandstone from the local bedrock outcrops are often visible (Figure 41A-C). No datable material was found.

Interpretation
When examining the aerial images from 1936 (Figure 40C) an ice-marginal lake is visible between the northern part of the ice margin and the lateral moraine. The drainage of the lake is visible adjacent to the glacier margin towards south. The location of the sediment cliffs agrees well with the extent of the lake at the time of the image. Based on that the sediments have been interpreted as glacilacustrine sediments formed by an ice-marginal lake during glacier retreat when meltwater becomes trapped between the ice-margin and local topography, the terminal moraine in this case (Benn and Evans, 2010). Often these lakes develop supraglacially and contain stratified sediments and mass-flow diamict. The hummocky terrain is furthermore an indicator of dead-ice melting and that at least part of the lake was developed supraglacially (Schomacker and Kjær, 2008). The inactive fluvial channels further indicate drainage towards the lake from the glacier and from it towards the ocean.

It is difficult to determine the extent of the sediments as the image is only capturing the size of the lake at a specific time and no other evidence exists on it evolution trough time and as well the hummocky terrain surrounding it. The minimum extent was thus estimated but further suggested that glacilacustrine sediments might extend further.

The horizontal terraces are interpreted as former shorelines of Lovénvatnet, indicative of higher lake levels in the past. Kjerulfbreen partly blocked the drainage during its maximum extent, which would have resulted in higher water level. It might as well have provided meltwater into the lake.
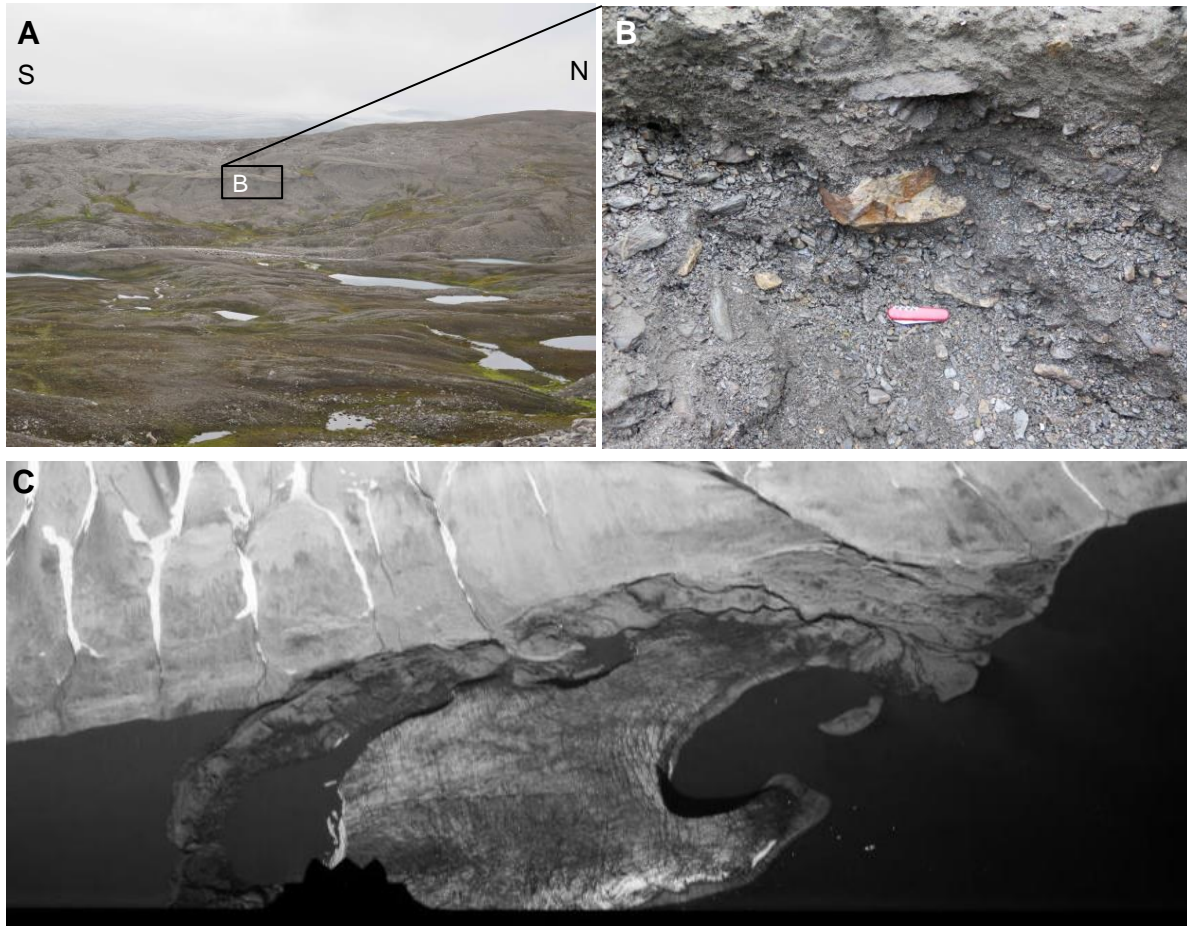
Figure 40. Glacilacustrine sediments. A) Sediment cliffs with hummocky terrain below it are interpreted to be glacilacustrine sediments formed by and ice marginal lake during ice retreat. Inactive fluvial channels are visible above the cliffs with larger grains size. B) An example of the sorted layers of sand and gravel observed in the sediment cliffs. The knife is ~7 cm long. C) The ice-marginal lake is visible on the 1936 oblique aerial images and the lateral drainage channels.
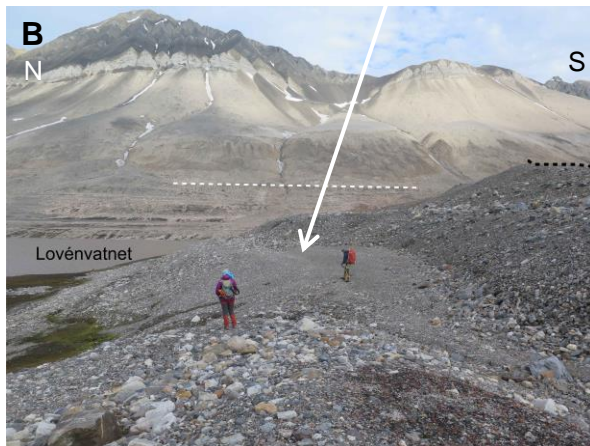
Figure 41. *Shorelines by Lovénvatnet. A) At least three shorelines (white dashed lines) can be seen at the distal side of the Kjerulfbreen terminal moraine (black dashed line). Behind the moraine is an area with glacilacustrine sediments. B) The first two shorelines and top of the terminal moraine (black dashed line). In the distance, the shorelines on the eastern side of the lake can be seen. The highest on (white line) reached ~50 m a.s.l. C) The surface cover of the shorelines consists of coarse washed gravel similar as seen on the current beach.*

4.1.5 Extra-marginal surface

All superficial sediment that extends beyond the forefields of the three large glaciers and the terminal moraine of Alkhornbreen are mapped as extra-marginal surface (Figure 22, 23). In total ~48.8 km^2 were mapped. The map unit includes bedrock and locally weathered bedrock, thin slope deposits, colluvial fans and avalanche tracks (Figure 42A, B). The dolerite dyke observed on the western side of Lovénvatnet was though mapped as it is considered important for the retreat of the glaciers and shorelines and striations were observed on it

(Figure 29A).

Beach material was observed at several locations and mapped on the western side. The material was found at two locations, south of the terminal moraine in Protektorbreen's forefield and north of Alkhornbreen's moraine. They reached up to 60 m a.s.l. and vague platforms can be seen in them. They consisted of sorted sand and gravel and their surface was vegetated compared to the surroundings (Figure 42A). Shell fragments were observed on the surface. The sandy gravel surface is interpreted to be the remains of raised beaches from higher sea level in the area following the deglaciation (Forman et al., 2004). Alkhornbreen moraine partly overlies the material indicating their relative age.
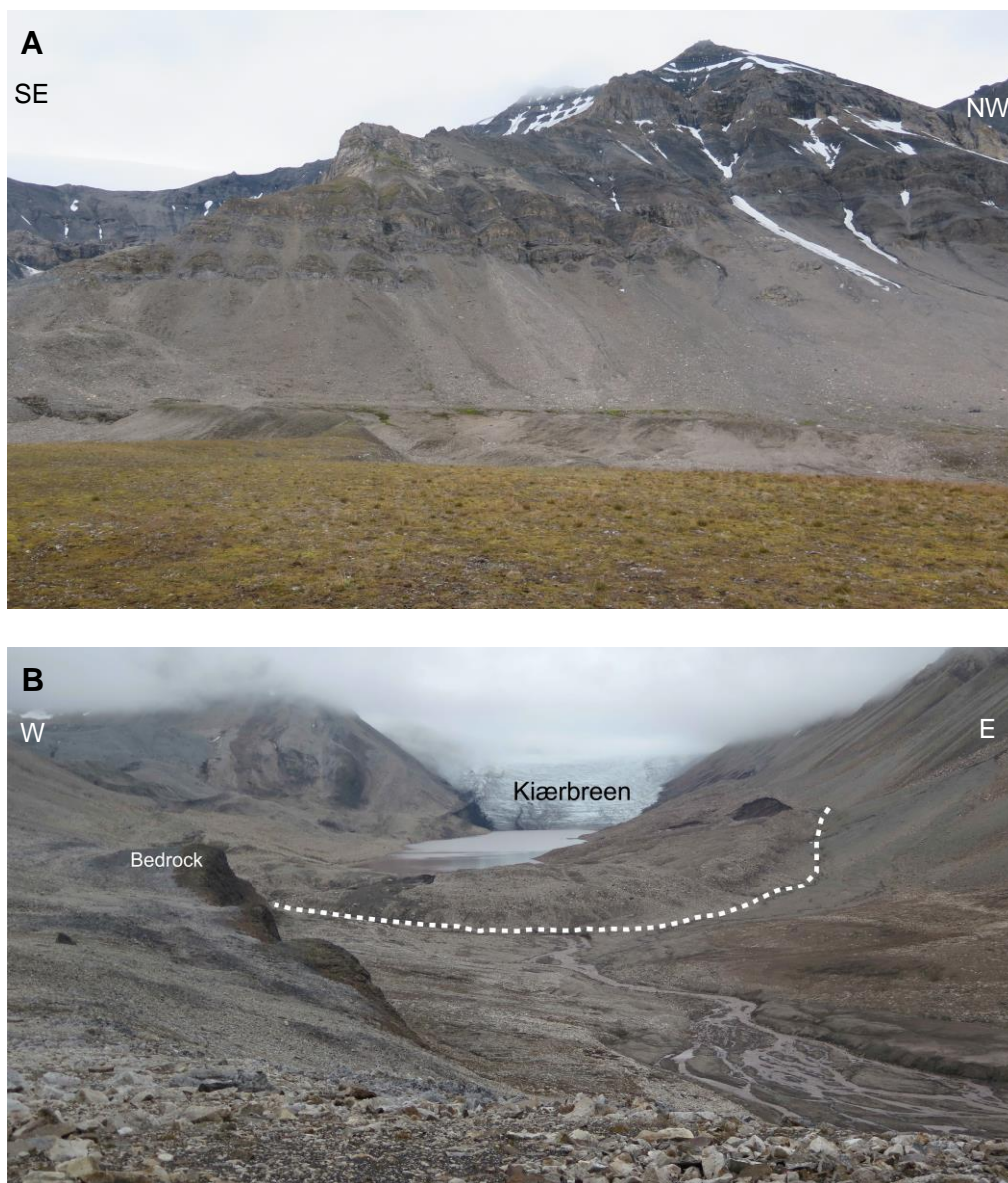
Figure 42. Raised beaches, extra-marginal sediments and Kiærbreen. A) Raised beach platforms on the western side. They are more vegetated than the surrounding area. B) Kiærbreen's terminal moraine and the moraine dammed lake in the background. The dolerite dyke stands out of the surrounding due to its resistance. In situ weathered bedrock, thin slope deposits and avalanche tracks can be seen on the mountain slopes in all figures.

4.2 Geomorphology: Submarine environment

The bathymetry data is divided into two parts by the terminal moraine; the inner, shallower part and the outer, deeper part. The inner part is deepest on the lee side of the terminal moraine, down to ~50 m b.s.l. It can be divided into two basins, a deeper NE basin, ~15-40 m b.s.l. and shallower SW basins, ~10-25 m b.s.l. Sets of transverse ridges are prominent in the inner part. Five islands, consisting of bedrock covered by glacial drift, are located in the fjord and no bathymetry data exists around them due to shallow water. Around the coastline and several other areas also had too low water depth to collect data. No data therefore exist from these locations and these are marked with a white polygon on the map. Topographic highs, ~1-3 m b.s.l. are also common, often in association with the islands and no data, and can be seen by the contour lines on the map. The topographic highs and islands have fairly smooth surface and a relatively flat top. The outer part of the fjord is more featureless than the inner one. It has rather steep side slopes and one basin, ranging from 100 m b.s.l. closest to the moraine down to 200 m b.s.l., furthest out in the fjord. On the E-side a ~500 m long and wide topographic high is situated 60-90 m higher than the surrounding, with 3-20 m water depth (Figure 20).

4.2.1 Subglacial landforms

Seafloor (inner part): Glacigenic material (inner part)
Description
Relatively smooth areas occur between the transverse ridges and topographic highs in the inner part of the fjord, covering an area of ~3.2 km^2 (Figure 22).

Interpretation
The inner part of the seafloor is interpreted to be draped with glacigenic material as it is known from historical data and the geomorphology that the glaciers reached further out in the fjord during their Neoglacial maximum. However, studies from the inner parts of other high-Arctic fjords have shown that a layer of fine grained glacimarine sediment from suspension, of unknown thickness, is likely to drape over the glacial landforms and glacigenic material after the retreat of the glaciers (Plassen et al., 2004; Flink et al., 2015, 2017; Allaart, 2016; Fransner et al., 2017). The thickness of the draped fine-grained unit is considered rather high due to high sedimentation rates because of the meltwater from the glaciers in Trygghamna (Plassen et al., 2004; Forwick and Vorren, 2010). The surface is drawn as glacigengic material to separate it from the outer part (*distal glacimarine sediments*) but draped with finer material from suspension. Sedimentary or acoustic data from the inner part of the fjord would be needed to conclude this.

4.2.2 Ice-marginal landforms

Major transverse ridge: Terminal moraine
Description
A large NE-SW orientated ridge almost crosses the fjord transverse to its axis (Figure 22). The ridge is located 2.2-2.5 km from the glacier fronts at the head of the fjord and ~500 m from the 1909/10 position (Figure 50). It covers an area of ~0.18 km^2. The transverse ridge is

semi-continuous and can be divided into three main parts. The NE part is ~350 m long and up to 150 m wide and has a slightly asymmetric profile, ~10 m high on the proximal side with 13-19° slope but ~25 m high on the distal side with 10-12° slope. It is mostly single crested and has a crescent shape. This connection of the profile is connected to a topographic high that further extends into a large-scale transverse ridge on the proximal side of the terminal moraine. The middle part is longest, ~900 m, and up to 150 m at the widest part and consists of two large lobes, one with a crescent shape and one smaller with straighter shape. It is up to ~20 m high at the NE part but only ~5 m at the SW part. It has an asymmetric profile, with a steeper proximal slope (17-24°) and smoother distal slope (10-13°). The SW part has the smallest dimensions, ~500 m long, up to 40 m wide and 10 m high. This section is very different from the other parts and has a symmetric profile, 20-30° and is curvi-linear in planform. The water depth is shallowest over the NE part, ~6-10 m, but down to ~30 m in the middle of the fjord (Figure 44; 45C).

Interpretation

The ridge is interpreted to be a continuation of the terrestrial terminal moraine onto the seafloor and represent the last maximum extent of Harriet- and Kjerulfbreen. They formed by the advancing glacier terminus that pushed marine muds and subglacial sediments in front of it or by sediment accumulation during longer still stands (Boulton et al., 1996). Similar subaqueous terminal moraines have been described from Svalbard in front of non-surging ones (Plassen et al., 2004) as well as surge-type glaciers, except in the latter they are often multi-crested and with larger dimension (Ottesen and Dowdeswell, 2006; Ottesen et al., 2008; Flink et al., 2015, 2017). Their massive size and asymmetric cross profiles are similar to the terrestrial thrust moraines formed in front of surge-type glaciers (Boulton et al., 1999; Evans and Rea, 1999, 2003). The reason for the non-continuous part and difference in dimension is suggested to be because Protektorbreen was less active than the others during the time of its formation and had limited access to sediments compared to the others. Fluvial erosion is also considered as a factor due to the proximity of the river outlet.

Large- and small-scale semi-transverse ridges: Retreat moraines
Description

Semi-transverse ridges are prominent in the inner part of the fjord. They have been divided into large- and small-scale ridges based on their dimension (Figure 22).

Small-scale ridges: Around 300 closely spaced, sub-parallel ridges are located in the inner part of the fjord. Their orientation is generally transverse to the fjord axis (NE-SW) but differ slightly around topographic highs and closer to the shore. They range from 7-760 m in length, with 124 m in average. They are ~5-20 m wide and ~0.5-4 m high. Their cross-profile varies greatly, from symmetric to asymmetric profiles with both steeper proximal and distal side. The slope angle is also highly variable (6-20°) (Figure 45A, B). They are curvi-linear in planform. They are most common in the deepest parts, closest to the terminal moraine and in the two basins. They are larger and with regularly spacing in the NE, deeper basin, than in other areas where they have much more irregular spacing and orientation (Figure 43A, B).

Large-scale ridges: Several curvi-linear segments were mapped in the inner part, ranging from 33 m to over 1000 m in length. They are ~20–50 m wide and ~4-12 m high. The ridges

both appear to be symmetric and asymmetric in profile, with a steeper proximal side than the distal. They make up almost continuous line on two places close to the terminal moraine that crosses the fjord. The ridges are often intertwined with topographic highs and islands. The alignment is usually NE-SW but curves around topographic highs. They are also wider, with a flatter top than the small one (Figure 45A, B). The ridges have a fairly good correlation to the 1909/10 and 1936 ice marginal positions (Figure 50).
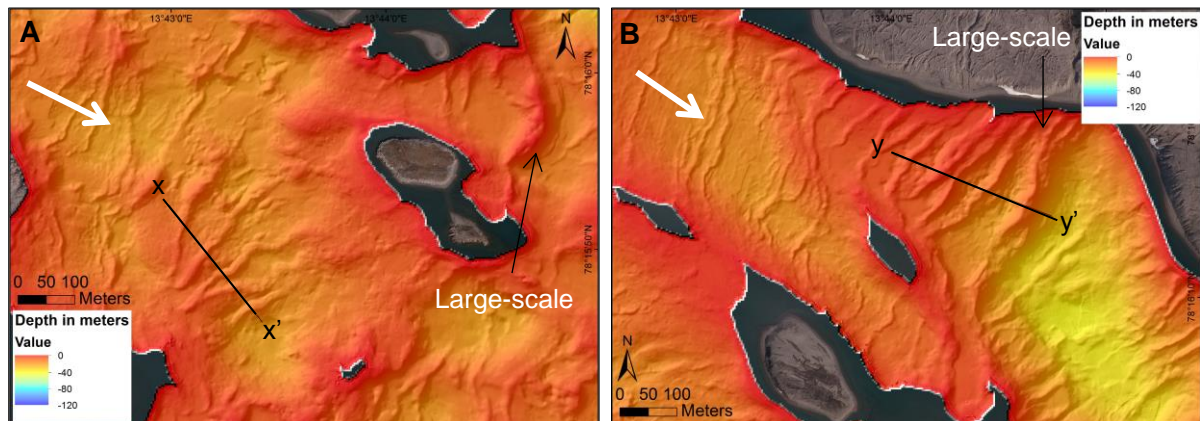
Figure 43. Large- and small-scale retreat moraine ridges in the submarine environment. A) Transverse ridges in the SW basin in the inner part of the fjord with rather irregular spacing and orientation. B) Transverse ridges in the NE basin in the inner part of the fjord. They have rather regular spacing compared to on other places. Both small and large-scale ridges are present on both images. Note the correlation of the ridges to the islands and topographic highs. The black lines indicate location of seafloor profiles (Figure 43A, B). The white arrow indicates the generalized direction of glacier flow.

Interpretation

The semi-transverse ridges are interpreted as retreat moraines formed during winter still-stands or minor readvances based on their morphology and correlation to the mapped ice-marginal positions. Similar ridges have been described in front of tidewater glaciers in Svalbard, both surging and non-surging. Therefore they cannot be used solely to determine if the glacier surged in the past (Ottesen and Dowdeswell, 2006; Ottesen et al., 2008; Flink et al., 2015). Similar ridges termed De Geer moraines have been described. They are thought to form by subglacial push during temporary halts in the grounding line retreat during deglaciation (Lindén and Möller, 2004) although their genesis is still discussed (Bouvier et al., 2015). As their profile both appeared asymmetric and symmetric it is not possible to conclude that they were formed during push (Boulton et al., 1996). The small size of small-scale ridges and their curvi-linear planform also made it difficult to determine their cross-profile.

The reason for the size difference could be a combination of several factors. Either the large-scale ridges were formed by longer still-stands of the glacier margin or readvances as described by Flink et al. (2015), which destroyed former ridges and incorporated them into the larger ridge. The correlation to topographic highs and islands could also be the reason for their larger dimension. The large-scale ridges could also be surge terminal moraines formed

by smaller surges following the maximum extent as is seen in the model by Flink et al. (2015). As there is no indication for a surge after the maximum extent the ridges are rather considered to be formed during the retreat of the glacier (*section 5.2.2*).

Velocity and calving measurements in front of Tunabreen, Svalbard, by Flink et al. (2015) demonstrated how the glacier experienced a small advance during the winter months due to lower calving rates. These ridges can often be correlated to annual positions and can therefore be termed annual retreat moraines. 12 to 18 ridges are located between the large-scale transverse ridges that have been correlated with 1909/10 and 1936 ice marginal positions.

Sediment lobe at the distal side of ridges: Debris flow apron
Description

Large, lobate accumulations of sediments are located at the distal side of the terminal moraine (Figure 22). The lobe begins at ~30-40 m b.s.l. and reaches down to ~100 m b.s.l. and has a maximum length of ~0.7 km, width of 1.6 km and covers an area of ~0.9 km^2. The lobe has a gentle slope, which flattens out towards the bottom (Figure 45C). The surface is rather irregular with blocks and mounds. Several lobes are visible within it. The lobe gets less extensive towards the SW (Figure 44). Smaller and less distinct lobes are located at the distal side of several of the larger ridges, up to 60 m long.

The sub-bottom profile crosses over the outer most part of the lobe (Figure 20) and shows up to 10 m thick acoustically homogenous, transparent sediment mass, with relatively irregular surface compared to laminated sediments outside it. It corresponds to acoustic facies F.3 (Table 3; Figure 46C).

Interpretation

The sediment lobes are interpreted to be glacigenic debris-flow deposits formed during the maximum Neoglacial extent of the glaciers based on its morphology and location at the distal side of the terminal moraine. This interpretation is in agreement with Forwick (2005) who associated the deposits with the latest advance, during the LIA or a recent surge event. Their formation is thought to be the result of combination of ice push and quasicontinuous slope failure of glacigenic material brought subglacial by the glacier to the terminal moraine (Ottesen et al., 2008; Kristensen et al., 2009). The acoustically homogenous and transparent sediment mounds further support their genesis (Plassen et al., 2004; Hogan et al., 2011). Similar lobes are described from other fjords in Svalbard, both surging and non-surging glaciers (Plassen et al., 2004; Ottesen and Dowdeswell, 2006; Ottesen et al., 2008; Flink et al., 2015, 2017). The stacked lobes indicate high sediment input, which could be related to a surge-event (Flink et al., 2017). The difference in dimension indicates that the SW side received less material, which is in agreement with the morphology of the terminal moraine.

The smaller lobes at the large-scale retreat moraines are interpreted to be debris-flow lobes as well, supporting that they formed during longer still-stands or re-advance of the glacier.
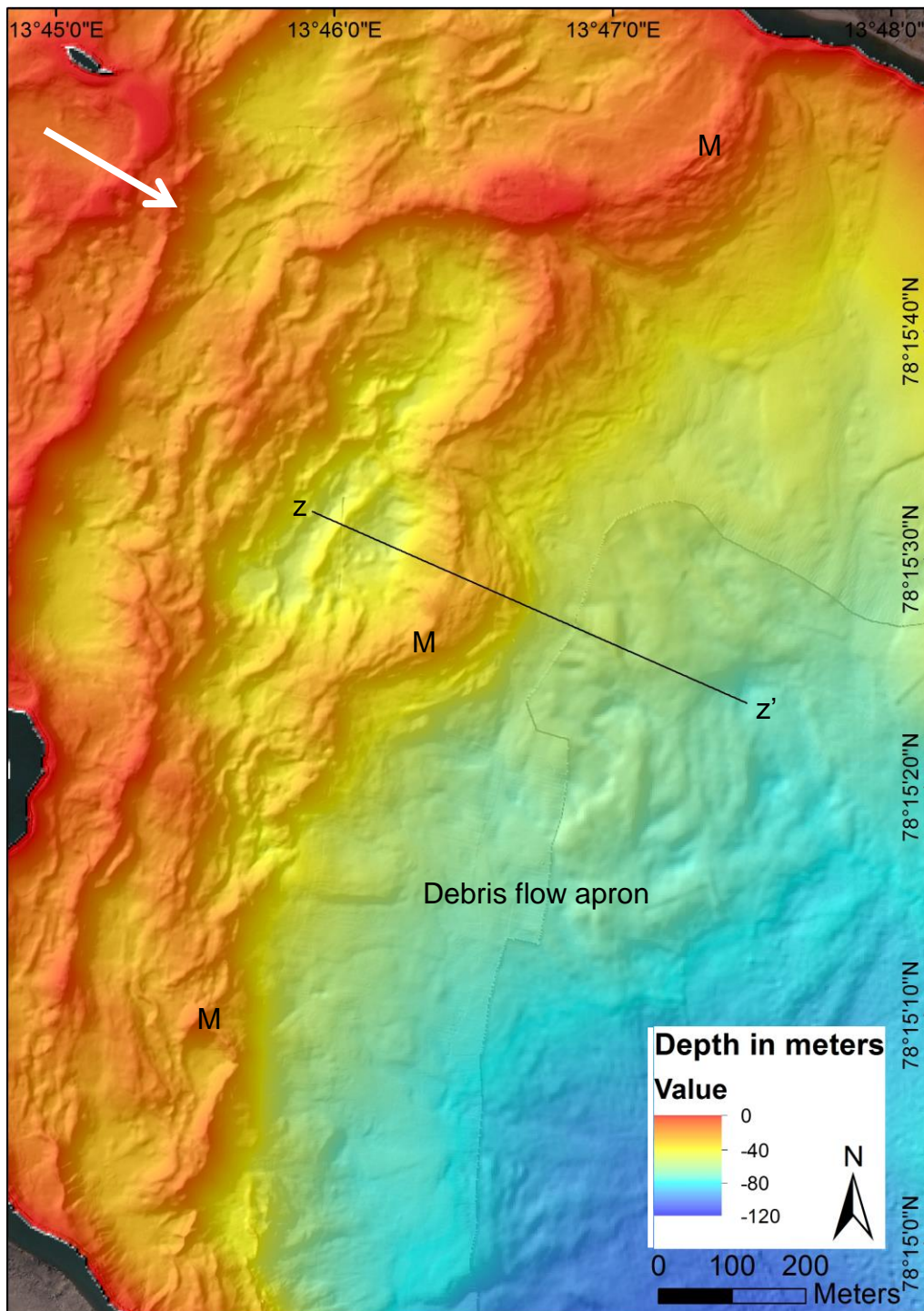
Figure 44. *Large terminal moraine (marked with M) with debris flow apron on its distal side and retreat moraines on the proximal. The terminal moraine is larger and has a crescent shape in the NE part but becomes more curvi-linear and smaller towards the SW. The debris flow apron has several stacked lobes and irregular surface towards the end. Retreat moraines, small and large-scale, are on the proximal side. Their correlation with the topographic highs is visible. The black line indicates the location of the seafloor profile (Figure 45C). The arrow indicates the generalized direction of glacier flow.*
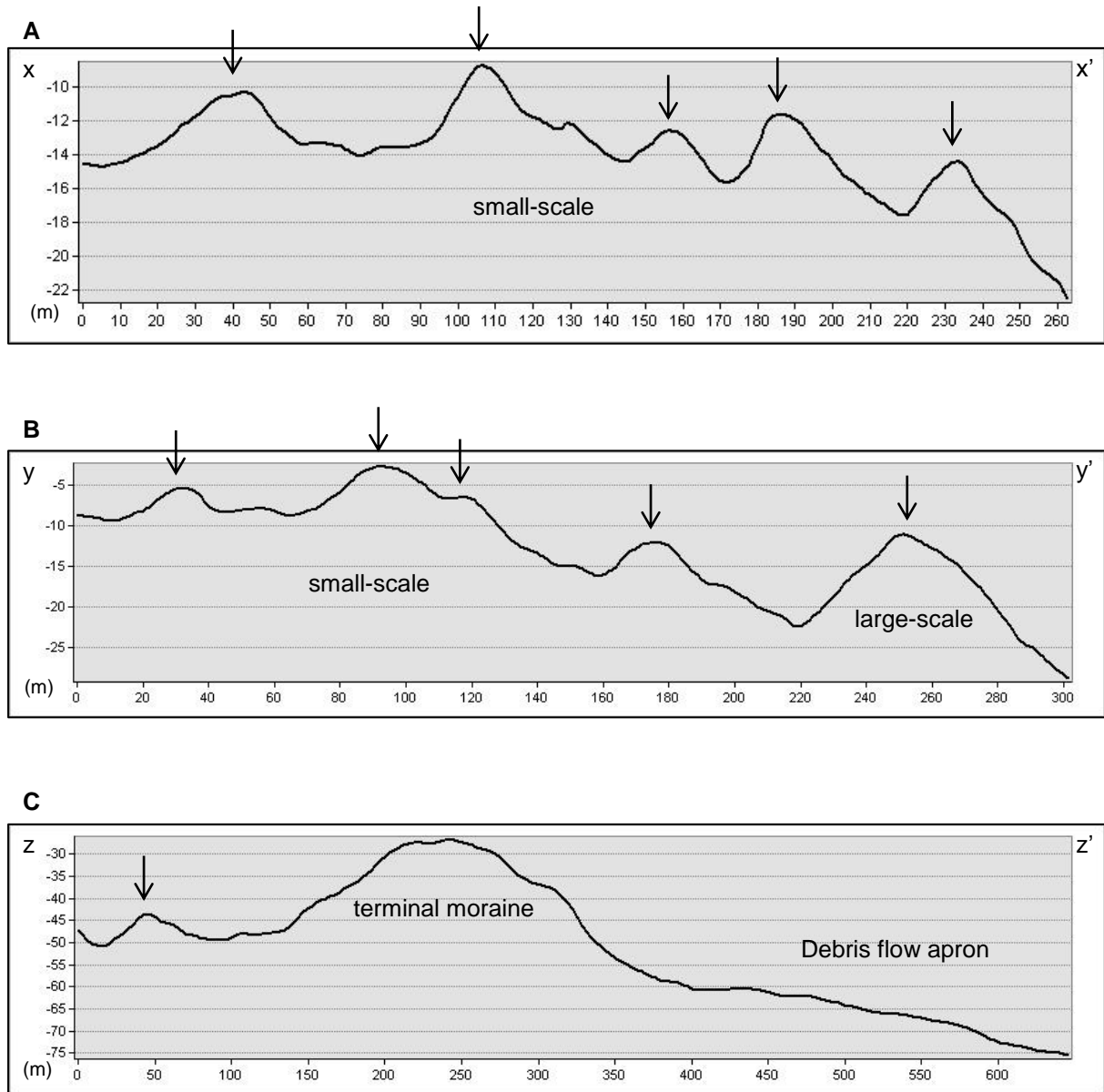
Figure 45. *Examples of submarine landforms from seafloor profiles based on the bathymetric data. A) Small-scale retreat moraines with variable cross-profiles, symmetric and asymmetric. B) Small-scale and one large-scale retreat moraines. Difference can be seen between the dimension of large-scale and small-scale. C) The terminal moraine, with slightly asymmetric cross-profile, and the debris flow apron on its distal side. A small-scale retreat moraine is located proximal to the terminal moraine. The arrows point at transverse ridges. Ice flow is approximately from left to right.*

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4.2.3 Proglacial landforms

Seafloor (outer part): Distal marine sediments
Description
Large sections of the sub-bottom profiles reveal stratified sediments distal to the debris flow apron. The sediment package is up to 30 m thick, corresponding to acoustic facies F.2 (Table 3; Figure 46A-C). The sediments follow the underlying topography on two locations. Acoustically homogeneous and transparent sediments drape the stratified sediments and are up to 3 m thick. At some places the distal marine sediments seem to be almost absent but are thicker in depressions. This unit corresponds to acoustic facies F.1 (Table 3; Figure 23; 46A-C).

Interpretation
Stratified sediments have been described distal of the terminal moraine in Trygghamna (Forwick, 2005; Forwick and Vorren, 2010) (Figure 14) and in other fjords on Svalbard and interpreted as distal glacimarine sediments, deposited since last deglaciation (cf. Ó Cofaigh & Dowdeswell, 2001; Plassen et al., 2004; Hogan et al., 2011).
Fairly high sedimentation rates have been estimated in outer Trygghamna during the Holocene, supporting that the glacimarine sediments are covered by fine grained material from suspension (Forwick, 2005; Hogan et al., 2011). The thickness varies though based on the topography. As the geomorphological map only shows the surface cover, it has been mapped as distal marine sediments covering in total ~9 km^2.

Slide blocks/lobes: Mass-transport deposits
Description
In the outer part of the fjord, vaguely outlined sediments lobes occur at three to four locations on the western side and one at the eastern side (Figure 23). These lobes are not as well defined as the debris-flow lobe at the distal side of the terminal moraine. They lie right under the fjord slopes and slide-scars are above them all except for the outer most. As the lobes are very vague and the resolution fairly low, the lines are only representing roughly the area of the lobes and the slide blocks.
On the western side the lobes are ~350-600 m wide and ~150-600 m long but ~50 m wide and ~100 m long on the eastern side. The first and third lobe on the western side, counted from the moraine, appear more as a blocky form rather than a lobe. The slide blocks closest to the moraine are up to 150 m wide, 100 m long and 8 m high. The blocks are smaller closer to the fjord mouth, up to 80 m wide, 40 m long and 2 m high. A vague lobe was observed around them and therefore it is mapped as a lobe rather than individual slide blocks. The sub-bottom profiles cross two of them. A small change in elevation is seen on both landforms and irregular mounds of 1-3 m thick, non-stratified and transparent sediment. The landform corresponds to acoustic F.3 (Table 3; Figure 46B, C).
A similar landform was also observed at the mouth of the fjord. It appears though more as a mound in front of a depression below the termination of the slope, rather than a lobe. No slide-scars were observed above that either. On the sub-bottom profile an increase in elevation can be seen. A ~5 m thick sediment package with vague laminations is located on top but below that no sub-bottom reflectors can be seen (Figure 46A).

Interpretation

The vague sediment lobes and blocks are interpreted as sediments that originate from down-slope mass-transport deposits based on their irregular surface and non-stratified, transparent sediments on the acoustic profiles and slight increase in elevation. The closeness to steep slopes and slide-scars further supports this interpretation. Forwick and Vorren (2010, 2012) (Figure 14) likewise described blocks and mass-transport deposits on the western side of Trygghamna. The triggering mechanism is thought to be a combination of slope failure on steep slopes and proximity to fluvial outwash. Retreat of glaciers during the deglaciation further exposes steep slopes and therefore increases the risk of slope-failure (Benn and Evans, 2010). The innermost could be a combination of sediments originating from the sides and the debris flow apron based on its location.

Slope depressions: Slide-scars
Description

Several depressions occur along the steep sides of the outer fjord, often with obvious scars around them. The scars vary from ~200-1000 m in length. Either they have a prominent U-shape around the depressions or only around part of them.

Interpretation

The depressions along the fjord sides are interpreted as slide-scars formed after down-slope mass-transport deposits due to slope failure as was discussed above (Forwick and Vorren, 2012).

Debris-flow tracks
Description

On the western side of the fjord, below the terminal moraine in Protektorbreen's forefield are around 20 scours that cut the slope. They range from 55 to 113 m in length.

Interpretation

These scours are interpreted as sediment debris-flow track, formed due to the unstable sediment on steep slopes. The instability could be triggered from several processes as was discussed above (Forwick and Vorren, 2012).

Elongated depressions: Iceberg ploughmarks
Description

Slightly elongated depressions occur at the surface of the proximal side of the topographic high on the NE side of the fjord. They are not that deep in profile, up to ~0.5 m and less than 5 m wide. Their shape is very chaotic, making it impossible to measure their actual length. The depressions occur at water depth from 25 to 10 m.

Interpretation

These depressions are interpreted as iceberg ploughmarks formed by icebergs calving of the glaciers, either during advance or retreat, and later grounded on the proximal side of the topographic high. Baeten et al. (2010) suggested that fresh appearance and shallow water

depths of iceberg ploughmarks in Billefjorden were formed by small icebergs during the Late Holocene. The same scenario might have been in Trygghamna during the last advance/retreat cycle of the glaciers. Sea ice could though as well have formed them due to the shallow water depths in the area, <10 m b.s.l.

Shallow seafloor depressions: Pockmarks
Description
Circular and elliptical in plan view, depressions are located on the distal side of the large topographic high outside the terminal moraine on the NE side of the fjord. Most of these depressions are located on a gentle slope, at 35 to 15 m b.s.l. They are fourteen in total and twelve of them are circular, ~10 m in diameter and ~0.5 m deep. The remaining two are elliptical and much larger, with maximum length of 90 m, 45 m in width and 4 m deep. They all have smooth sides with gentle slopes except for one with sharper and steeper edges.

Interpretation
These depressions are interpreted as pockmarks based on their morphology. They can both be single circular and elliptical or composite (Judd and Hovland, 2007) and are as well classified after their formation, e.g. fault-strike pockmarks, iceberg scour pockmarks, current-modified pockmarks. Forwick et al. (2009) described an elongated trough as a pockmark trough that has evolved from several pockmarks. The same mechanism could have occurred for the elliptical depressions. Pockmarks have been described in the neighboring fjords, Ymerbukta and Grønfjorden, although with much larger dimension (Forwick et al., 2009). Those pockmarks are suggested to have formed by the releasing of thermogenic gas through the sub-seafloor, controlled by bedrock stratigraphy and tectonic lineaments (Roy et al., 2014, 2015). The focus of this study is not the formation of pockmarks but the formation is suggested to be similar as organic-rich bedrock and faults are located in the fjord (Figure 13). Previous authors did however not find any evidences of them in Trygghamna when studying the bathymetry. Their data did however not cover the entire fjord as this study does, which is thought to be the reason it was not mentioned.

4.3 Acoustic sub-bottom facies of outer Trygghamna

Sub-bottom profiles from chirp data in the outer part of the fjord (Figure 20) are used to support the interpretation of landforms. The acoustic sub-bottom facies can be divided into four different types (Table 3). F.1 – F.3 are described in *section 4.2.2* and *4.2.3*. In addition evidences for landforms buried by sediments were observed, F.4. Examples of the chirp data can be seen in Figure 46A-C. Facies F.1, distal marine sediments, drape most of the outer part of the seafloor and is therefore present on the map (Figure 23).

Table 3. *Acoustic sub-bottom facies interpreted from chirp data in the outer part of Trygghamna.*

Acoustic facies	Description	Interpretation
F.1 – Distal marine sediments	Acoustically homogenous, transparent sediments, draping the seafloor. Up to 3 m thick. Strong and continuous top reflector but locally diffused bottom reflector.	Fine grained Holocene muds formed by distal sedimentation (cf. Hogan et al., 2011).
F.2 – Distal glacimarine sediments	Parallel or sub-parallel, often diffuse multiple sub-bottom reflectors.	Distal glacimarine sediments, deposited since last deglaciation (cf. Ó Cofaigh & Dowdeswell, 2001; Plassen et al., 2004; Hogan et al., 2011).
F.3 – Mass-transport deposits /debris flow apron	Acoustically transparent sediments forming irregular mounds. Located adjacent to steep slopes. Range from 1-10 m in thickness.	Down-slope mass-transport deposits and debris flow apron (cf. Plassen et al., 2004; Hogan et al., 2011)
F.4 - Moraine	Acoustically transparent with a strong upper reflector. Sub-bottom reflectors below are absent. With a mound shape. Up to 10 m high and >100 m wide.	Subglacial till or bedrock are acoustically impenetrable (cf. Plassen et al., 2004; Hogan et al., 2011). These landforms have previously been interpreted as push moraines (Forwick 2005; Forwick and Vorren, 2010). This will be discussed in *section 5.1.1*.
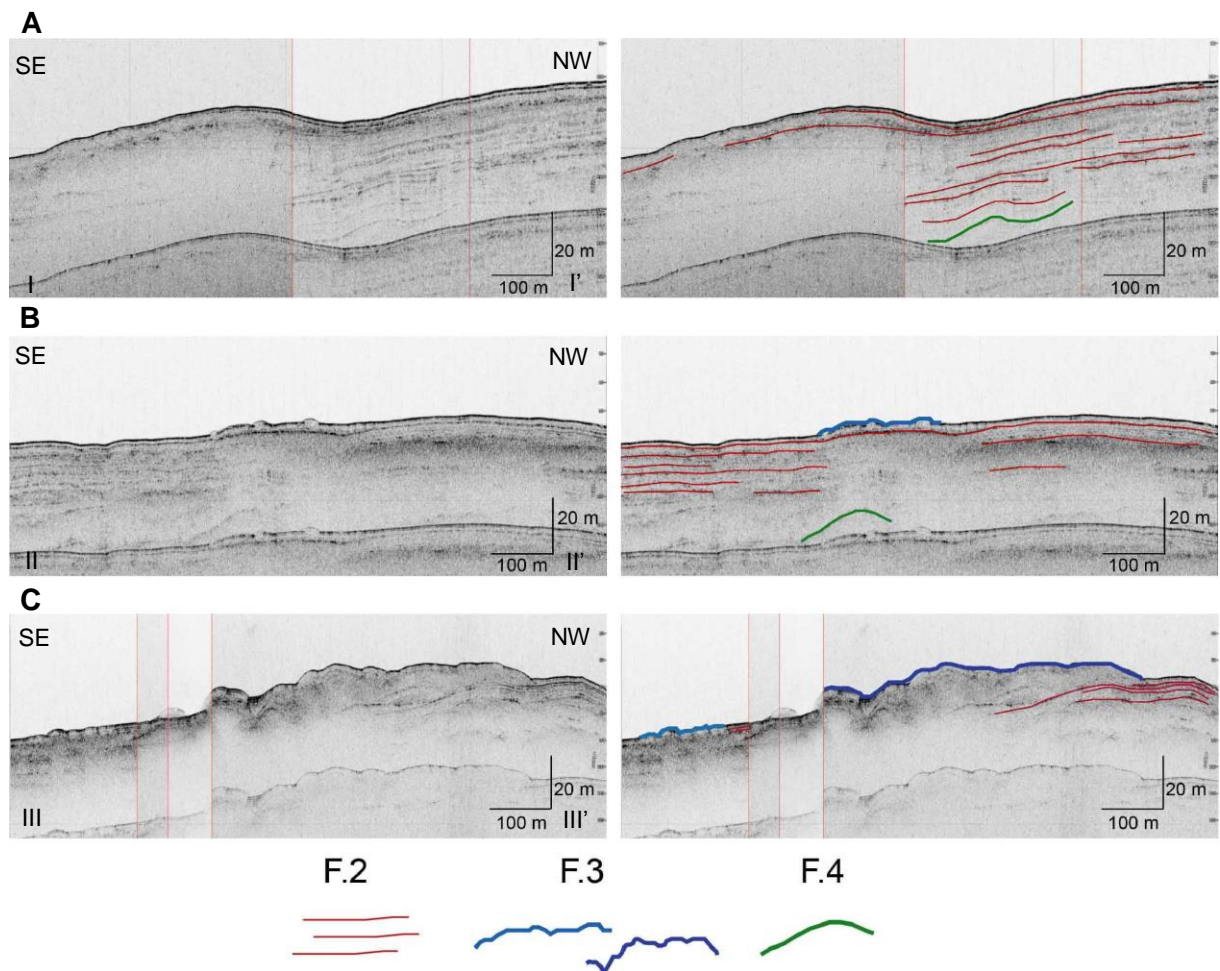
Figure 46. *Sub-bottom profiles from chirp data in the outer part of the fjord. The original data and interpretation are displayed together. Location for the profiles can be seen on Figure 20 and the legend is below. A) The topographic rise at the mouth of the fjord has weak stratification on top but acoustically transparent underneath it. Stratified distal glacimarine sediments are located outboard it. Below them a strong upper reflector has been interpreted as a moraine. B) Mass-transport deposits have more irregular surface and acoustically transparent sediments than the stratified glacimarine sediments. Another possible moraine can be seen. C) The lowest part of the debris flow apron has a very irregular surface and up to 10 m thick non-stratified sediments compared to the stratified distal glacimarine sediments outboard and partly underneath it. Distal to it the inner most mass-transport deposits can be seen. All of the profiles contain various thicknesses, <3 m, of distal marine sediments (F.1). They have not been marked on the profiles due to their low thickness but are considered to drape most of them.*

4.4 Ice-marginal reconstructions

Ice-marginal reconstructions can be a useful tool to examine the retreat history of glaciers (Liestøl, 1988; Oerlemans, 2005). The ice-marginal positions from 1909/10 were mapped to be able to reconstruct the glacier retreat, with a focus on if any signs of advances could be detected and to compare it with the geomorphological data. Six yearly positions were mapped in total for the three larger glaciers but only two for the smaller ones since the resolution was not sufficient enough. The terminal moraine in the submarine environment was also mapped. The terminal moraines were in addition mapped for Alkhorn-, Kiærbreen and below the perennial snow on the western side that used to be a small glacier (Figure 48A, B; 49A, B; 50). The quality of the reconstructions is however based on the resolution of the data. In Trygghamna the former coalesced glacier margin is for example easier to reconstruct than the terrestrial margins due to its debris cover fronts and less distinct outlines. The timing of the images (darkness, clouds, snow and sea-ice cover) is also a reason that only six positions were presented on the final map. None of the other margins that were examined showed any signs of advance and were therefore not concluded on the map.

When searching for historical data the aim was to find the oldest available data revealing the ice-marginal positions, to obtain the longest possible record. Even though a great number of hunters and expeditions sought shelter and carried out scientific work in Trygghamna, little documentation of the glaciers were found. The search is also constrained by the multiple names that Trygghamna had in the past and also that the many of the documentation are not in English or Scandinavian. Based on descriptions from the Russian Pomors the glacier front reached further out in the fjord during the 18th century than today although we don't know the extent (Storå, 1989). This historical account is in agreement with the map from Isachsen (1915) (Figure 17) and on old photographs from 1890-1908 (Nordenskiöld, 1892; Hamberg, 1905; Halldin, 1908), where the tidewater fronts of the three larger glaciers all coalesced and the terrestrial margins were at their terminal moraines (Figure 18; 47A-B; 48A; 49A). Based on that the terrestrial glacier margins were all located adjacent to their terminal moraines between 1890 and 1909/10.

The measured retreat of the glaciers can be seen in Table 4. The former connection of the three large glaciers complicates the measurements and they are only made when thought applicable. The 1909/10 margin is located 0.5-0.6 km inside the subaqueous terminal moraine. The measurements for the terrestrial margins are more complicated as discussed before but the aerial images from 1936 show that all of the margins have retreated from their terminal moraines at that time. The positions show a continuous retreat after 1909/10. Kjerulfbreen has retreated the most, followed by Harrietbreen. The retreat was fastest until 1968 for Kjerulfbreen but until 2007 for Harrietbreen. Kjerulf- and Protektorbreen are both completely land-terminating today. Harrietbreen still has a grounded calving front and is therefore at the boundary between tidewater and land-terminating glacier. As mention in the submarine environment (*section 4.2.2*), comparison of the ice-marginal positions to the bathymetry data reveals that the 1909/10 and 1936 margins coincide well with large-scale retreat moraines in the bathymetry. The shape of the positions is best described as concave but points out around the islands (Figure 50).

Old photographs further reveal information about the glacier surface. Photographs from 1890 and 1908 (Figure 18) show how the three large glaciers have a high tidewater ice front and slightly undulating surface. Both of the features are most prominent on Protektorbreen. The snow cover on the glaciers is though still rather high based on the snow cover on the surrounding ground, potentially obscuring features on the glaciers. However, patches with crevasses can be seen on all of them. They are most extensive on Protektorbreen (Figure 47A, B). The ice-marginal lake is not visible on the 1908 photographs but has formed in 1936 (Figure 19; 47B).
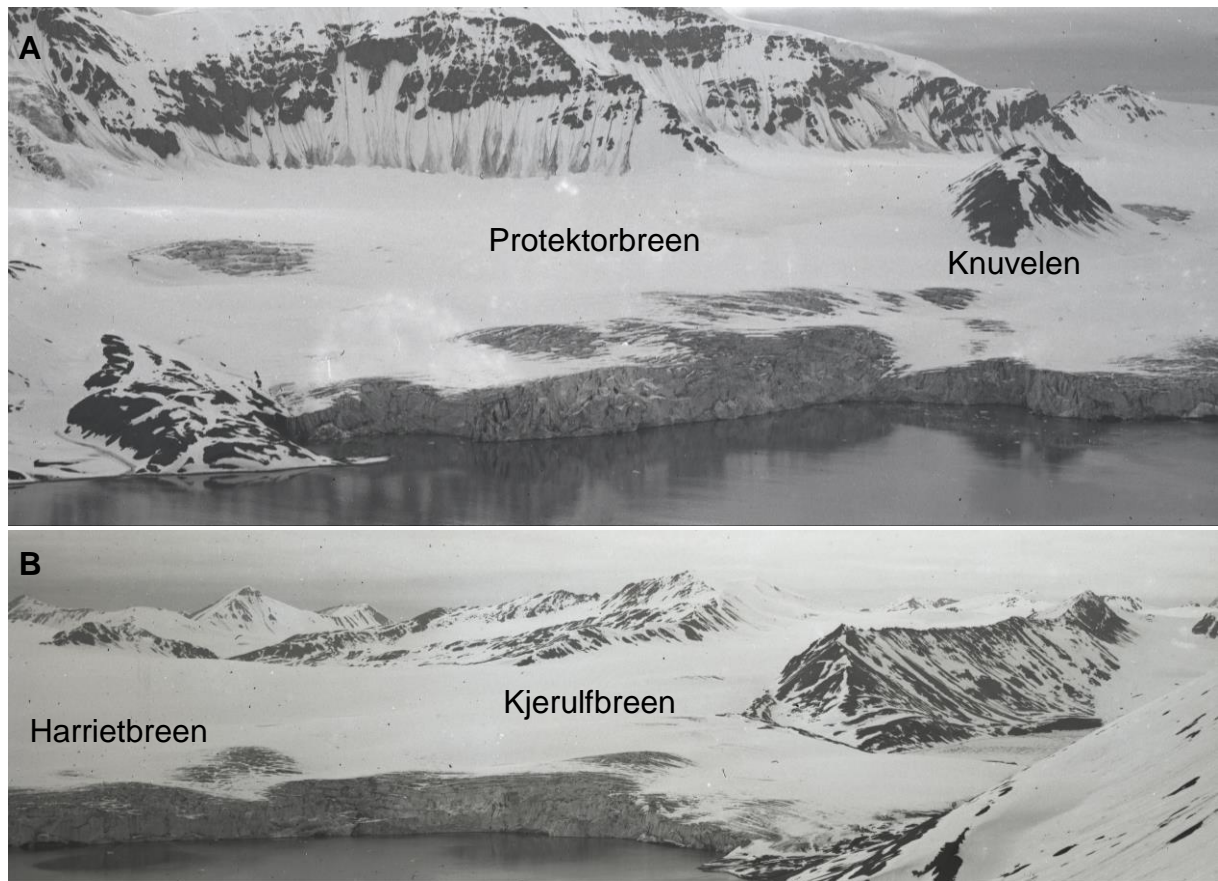
Figure 47. Ice-marginal reconstructions and glaciological features on photographs from 1908. A) Protektorbreen had an especially high ice front and patches with extensive crevasses. The glacier surface is at similar location and the height of the trimline (Figure 37B). B) Harriet- and Kjerulfbreen also had rather high ice front. Some areas with extensive crevasses can as well be seen. On both images the terrestrial margins of the glaciers are located at the terminal moraines (Halldin, 1908).
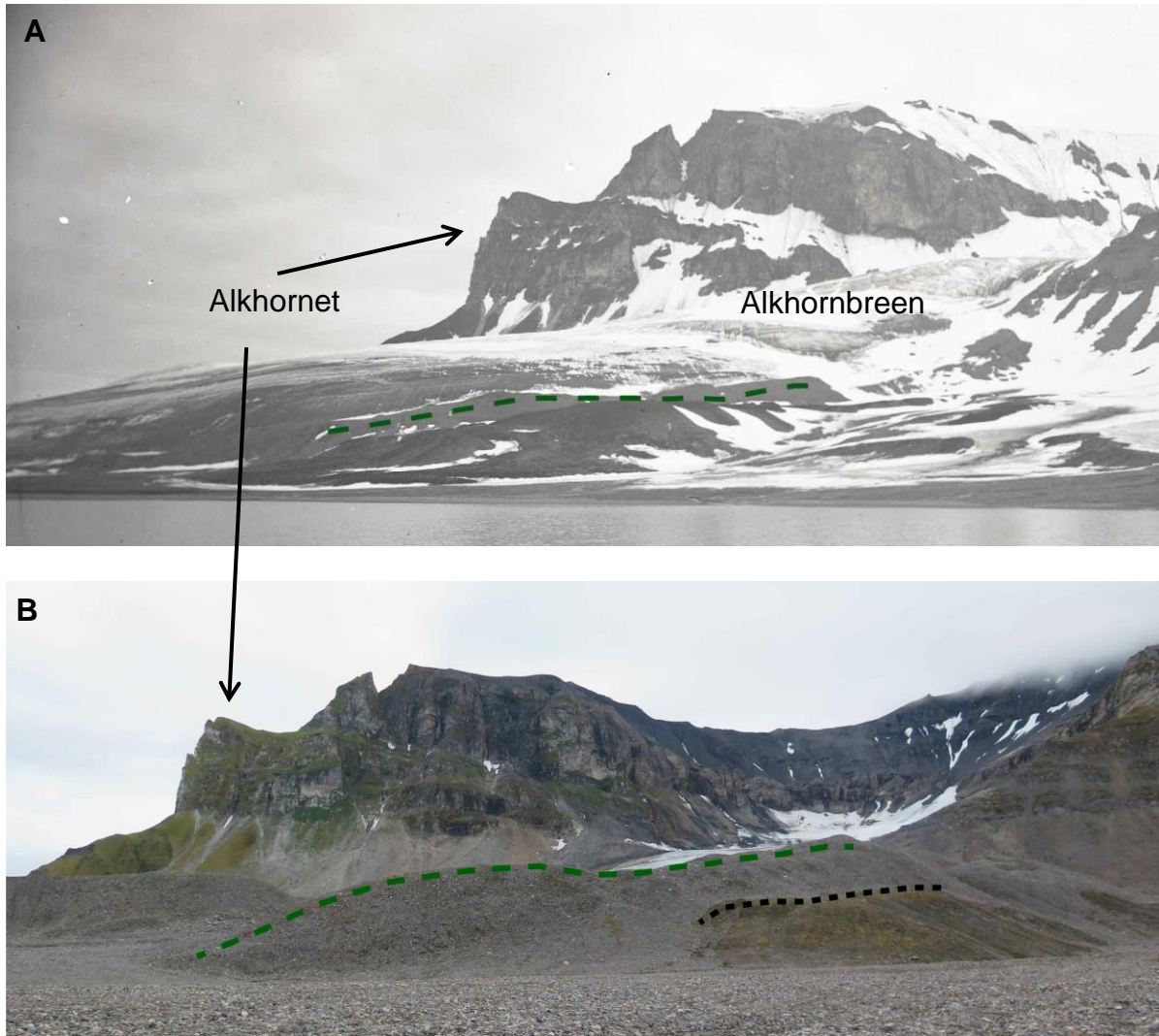
Figure 48. *Ice-marginal reconstructions - Alkhornbreen. A) Old photograph from 1898 (Hamberg, 1905). The glacier reached all the way to the ocean and its terminal moraine (green line). The platforms on its sides are the remants of raised beaches. B) Alkhornbreen in summer 2016. The terminal moraine (green line) of Alkhornbreen overlying remains of raised beach platform (black line). The difference in vegetation is significant.*

A

B
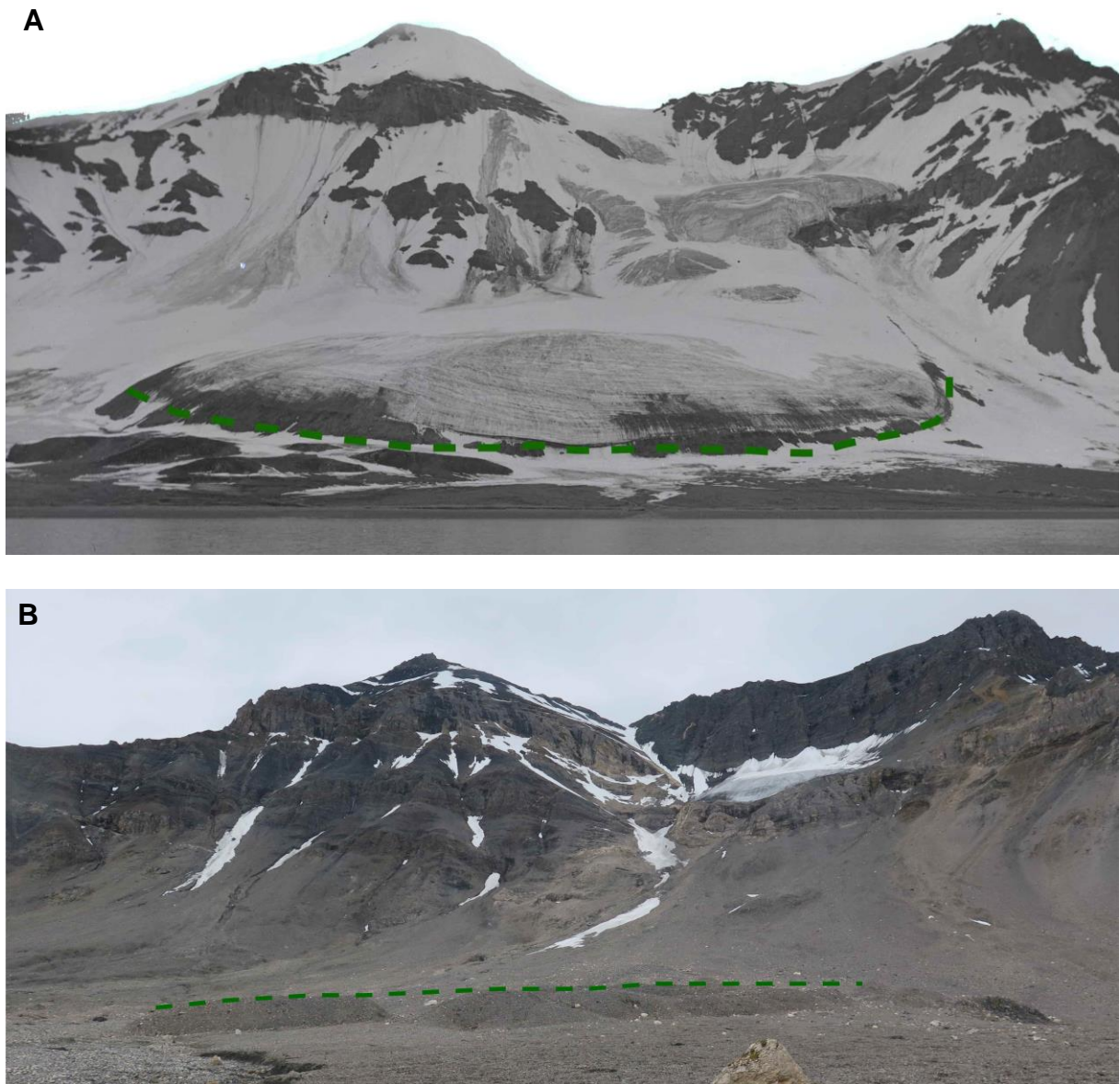
Figure 49. *Ice-marginal reconstructions – unnamed glacier. A) Old photograph from 1898 (Hamberg, 1905). Small hanging glacier was than located on the western side of Trygghamna. It reached all the way to its terminal moraine (green line). B) The terminal moraine today and the remanats of the glacier.*

Table 4. *Measured ice-marginal retreat in km for each glacier from the terminal moraine to 2016 when applicable.*

		Year					
		Terminal moraine-1909/10	1909/1910-1936	1936-1968	1968-1990	1990-2007	2007-2016	Terminal moraine - 2016
Glaciers	Alkhornbreen	<-	0.1	<-	0.5	->	-	**>0.6**
	Protektorbreen	<-	0.25	0.6	->	0.1-0.2	0.2	**~1.1**
	Harrietbreen	0.5	0.6	0.5	0.2	0.5	<0.1	**2.4**
	Kjerulfbreen	0.5	0.7	0.9	0.2	0.3	0.2	**2.8**
	Kiærbreen	<-	0.1	<-	1	->	-	**>1.1**
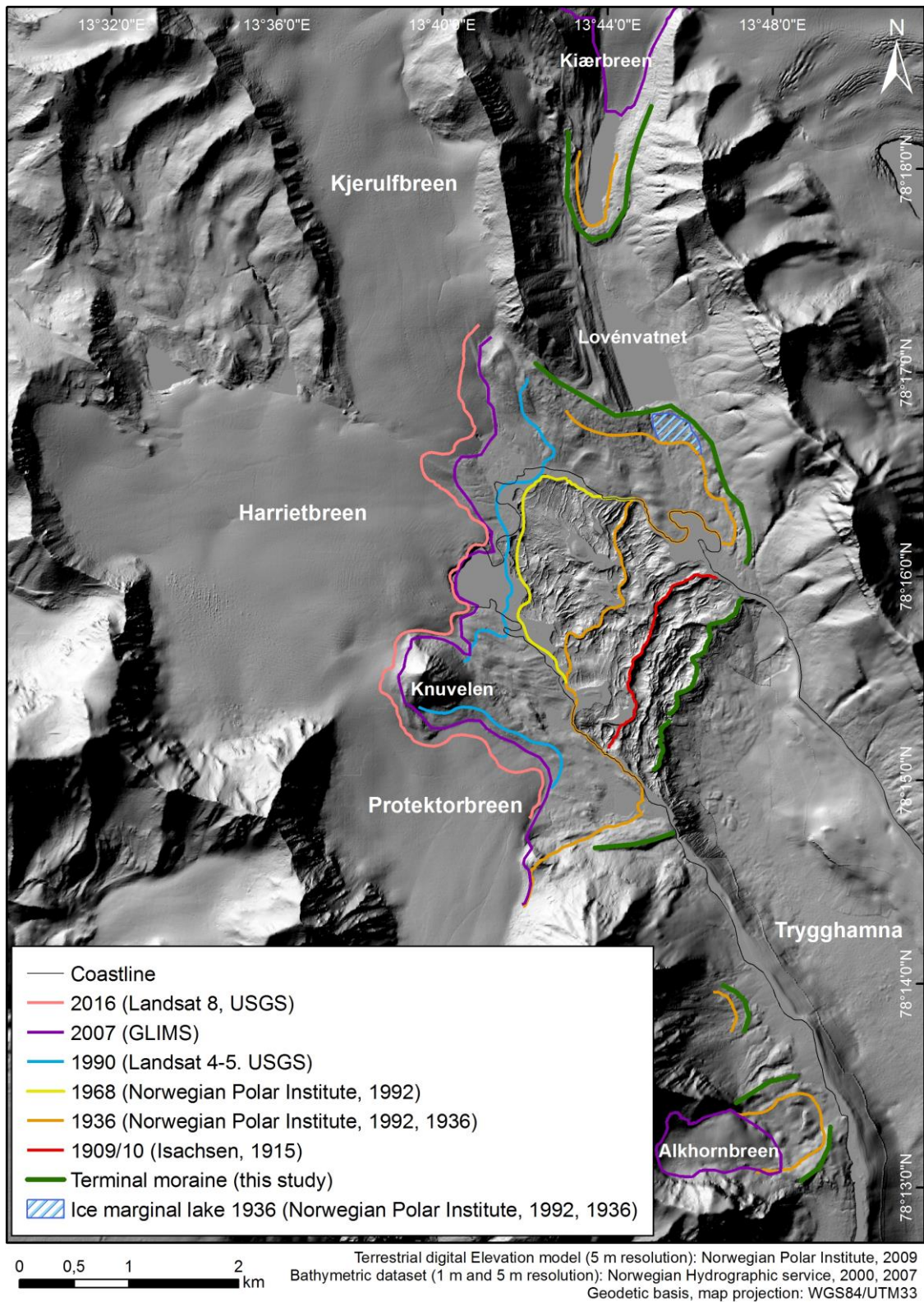
Figure 50. *Ice-marginal reconstructions from 1909/10 until today based on historical data, satellite and aerial images. The terminal moraine marks the maximum extent in all forefields. The correlation between the 1909/10 and 1936 margins with the large-scale retreat moraines can be seen.*

5. Discussion

In Trygghamna, the most prominent geomorphological record is from the last glacial advance/retreat cycle, i.e. the LIA. This is generally the trend across Svalbard as the glacial imprints tend to be biased towards the youngest event (Landvik et al., 2014). The Late-Glacial / Early-Holocene record is more fragmented.

Based on the geomorphological map, it can be argued that at least two of the glaciers have exhibited surge behavior. The landform assemblage in Trygghamna is further compared with previously published landsystem models. Historical data is used to discuss the timing of the event.

Different factors in the glaciers environment are considered important factors controlling the formation of landforms and their preservation potential. This could result in the absence of landforms.

The application of the landsystem models is discussed. The risks by using them to identify undocumented surges are highlighted but at the same time the importance for understanding surge-type glaciers in the greater context and the reason behind individual advances.

A clear difference was observed in the landforms assemblage between the terrestrial and marine environments due to implications by their thermal regime.

5.1 Glacial history of Trygghamna

5.1.1 Late-Glacial / Early-Holocene
Glacial striations with orientation approximately N-S and erratics (Figure 22) are considered the only clear evidence observed in the terrestrial part for the advance of the glaciers prior to the Neoglacial. This can be argued as they are either outside the maximum extent or their direction differs significantly from the Neoglacial ones. Their orientation indicates a southwards flowing tributary glacier, draining from the current Esmarkbreen down through the valley where Kiærbreen and Lovénvatnet are located today and merged with the ice existing in Trygghamna. Based on marine data sets, a large ice-stream drained Isfjorden during the Late Weichselian (Ottesen et al., 2007; Landvik et al., 1998) and topographical constrained local ice-flow existed in Trygghamna (Forwick and Vorren, 2010). Glacial striae, erratics and till form previous studies, further indicate that the terrestrial area around Trygghamna was completely glaciated during the Late Weichselian (Salvigsen et al., 1990).

Two moraines are located in the outer part of Trygghamna based on the sub-bottom profiles in this study and their locations correspond to previously described push/thrust moraines (Forwick, 2005; Forwick and Vorren, 2010) (Figure 14; 46A, B). Three moraines were described in that study. However, their resolution penetrates deeper down, as it is from a boomer profile, than the chirp profile from this study. That on the other hand has higher resolution, likely explaining the differences observed. The moraines were earlier interpreted to be from the Late-Glacial / Early-Holocene and represent readvances or halts during the regional deglaciation. Based on the stratigraphy their relative age could be assumed (Forwick and Vorren, 2010) (Figure 51). This is in accordance with other studies from Svalbard that have suggested readvances during that time (Salvigsen et al., 1990; Svendsen et al., 1996;

Lønne, 2005) and recent work indicates the importance of investigating those (Farnsworth et al., 2017a).

Higher sea level is expected due to the isostatic depression following the regional deglaciation. The marine limit rises from 48 m a.s.l. west of Trygghamna to 60 m a.s.l. east of it, with a minimum age around 10 ka, constraining the timing of the deglaciation for the area (Forman et al., 1990, 2004; Salvigsen et al., 1990). Remains of raised beaches platforms in Trygghamna indicate that relative sea level was at least up to ~60 m a.s.l. (Figure 23). The marine limit therefore correlates well with the surrounding area.

5.1.2 Neoglacial extent

The terminal moraines and trimlines mark the maximum extent of the glaciers during the Neoglacial (Glasser and Hambrey, 2003) (Figure 50). Based on the morphology of the terminal moraine in the submarine environment, Harriet- and Kjerulfbreen were the dominant agents during its deposition. More localized flow is indicated by the terrestrial evidence, glacial striations, flutes and supraglacial debris bands. The slight offset between them and the current ice-flow can be explained by local topography influencing the ice-flow.

Based on the ice-marginal reconstruction the tidewater part reached its maximum position prior to 1909/10 in the submarine environment. Old photographs and historical map further reveal that the terrestrial margins of all five glaciers and a small hanging cirque glacier on the western side, were located adjacent to their terminal moraines ~1900 (Nordenskiöld, 1892; Hamberg, 1905; Halldin, 1908) (Figure 50; 51). On the 1936 aerial image all of the terrestrial margins have retreated from their terminal moraine (Figure 19A-C).

Evidences for Neoglacial advances have been reported from all over Isfjorden and most glaciers are considered to have reached their maximum extent during the LIA (Hagen et al., 1993; Svendsen and Mangerud, 1997). Pre-LIA Neoglacial moraines have been described on Svalbard. They are however still poorly constrained in time. Differences in vegetation and lichen cover on moraines indicate that some glaciers were roughly at their LIA maximum extent or outboard of it during the Early Neoglacial (Werner, 1993). This model has partly been supported with cosmogenic exposure ages from moraines (Reusche et al., 2014). Pre-LIA advances have also been described from lake sediments in northwestern Svalbard (van der Bilt, 2015; Røthe et al., 2015). Another complication regarding the maximum Neoglacial extent is the common occurrence of surge-type glaciers in Svalbard and that many LIA moraines have been correlated to surge-events (Lefauconnier and Hagen, 1991; Hagen et al., 1993). The surging history for Trygghamna will be discussed in *section. 5.2*.

Due to the lack of an exact time constraint for the moraine formation in Trygghamna, only their minimum age can be assumed. However, based on the general record from the surrounding fjords and the close proximity of the ~1900 glacier margins to the terminal moraines, they are considered to be from the LIA. It can though not be concluded for how long the glacier margins were there or if pre-LIA advances reached the same extent.

5.1.3 Twentieth century retreat

All glaciers are thought to have been in overall retreat following the ~1900 and 1909/10 position based on the ice-marginal reconstructions and historical data (Figure 50). However, only six positions were mapped due to the resolution and timing of the satellite images. It is therefore acknowledge that the resolution is sparse and smaller readvances or longer halts in

recession could have occurred between the positions.

The pattern of the retreat is further preserved in the ice-marginal and geomorphological record. High abundance of retreat moraines in the submarine environment reveals the morphology of the grounded glacier margin during its retreat (Flink et al., 2015) (Figure 22). The correlation between the positions of the retreat moraines and mapped ice-marginal positions to topographic highs and islands, which are considered to have a bedrock origin, highlights how the glacier rested for a longer period around them (Figure 22; 50). Similar scenarios have been documented at other tidewater glaciers in Svalbard, where the topography acted as pinning points during the retreat by providing a temporary stability for the ice front during retreat (Maclachlan et al. 2010; Sund et al., 2011; Allaart, 2016). The islands and most of the forefields are thought to consist of sandstone and shale (NPI, 2016) (Figure 13) and its appearance in the terrestrial record, which has a large control on its topography, indicates that it is fairly resistant. A dolerite dyke was also observed on the western side of Lovénvatnet that is not present on the bedrock geology map of the area. That and more dykes that haven't been observed could extend into the submarine environment and provide even more resistance. The bedrock on land is considered to have had the same effect on the retreat. The regional geology and topography thus play a vital role for the glacier's retreat. The concave shape of the ice-marginal positions and that the tidewater margins retreated faster than the terrestrial ones (Figure 50) indicates how the glaciers acted differently during the retreat in the submarine than terrestrial setting due to the positive feedback from the ocean (Anderson and Ashley, 1991)

During the retreat an ice-marginal lake was formed in the forefield of Kjerulfbreen as seen on old aerial images and presence of glacilacustrine sediments (Figure 26). The lake is not visible on the 1908 images. The extensive abandoned fluvial channels and outwash plains can further be used to trace the draining of the lake adjacent to the terminal moraine. Other abandoned channels in the forefields can further be used to show the sequential retreat of the glaciers as has been done previously by Dyke (1993) at cold-based margins. However, due to almost complete absence of retreat ridges in the terrestrial environment the retreat cannot be reconstructed based on them.

Retreat moraines can sometimes be connected to annual winter positions of the glacier front and have therefore been termed annual retreat moraines (Evans and Twigg, 2002). Often it is difficult to connect them to former positions of the glacier front due to lack of historical data, low preservation potential and their incompleteness. Therefore, it cannot always be proven that they formed annually (Jónsson et al., 2014, Flink et al., 2015). The 1909/10 and 1936 margins coincide well with large-scale retreat moraines (Figure 50). However, counting of small-scale retreat moraines between those two positions did not demonstrate that they were formed annually in this setting. The explanation could be that longer still-stands were controlled by the subglacial topography.

5.2 Surging glaciers in Trygghamna

5.2.1 Evidences for past surges

Harriet- and Kjerulfbreen are considered to have exhibited surge behavior based on the presence of CSRs that are thought unique for surging glaciers (Sharp 1985a; Evans and Rea, 1999, 2003; Rea and Evans, 2011; Schomacker et al., 2014) (Figure 25; 26). This is in agreement with recent work from the area (Wallin, 2016; Ben-Yehoshua, 2017) and a remote sensing survey, utilizing solely the presence of CSRs to classify surge-type glaciers (Farnsworth et al., 2016).

The landform assemblage does now show a good correspondence to any of the presented models for this study but can be correlated to some extent to all of them. The abundance of CSRs in the terrestrial forefields of Harriet- and Kjerulfbreen together with other landforms correlates best with the surge outlet glacier landsystem model but also shows similarity to the other surge-type landsystem models. Fluted till plain and hummocky moraine are present in the outlet and cirque glacier model. Supra- and englacial debris are present in the cirque glacier model. Concertina eskers were not observed in neither of the forefields but are presented in all of the models except the valley glacier model, where it only rarely occurs (Evans and Rea, 1999, 2003; Brynjólfsson et al., 2012, 2014; Schomacker et al., 2014; Brynjólfsson, 2015). The landform assemblage in Trygghamna also shows similarities to the polythermal model, however it differs by the presence of CSRs that are not in the model (Glasser and Hambrey, 2003). Glaciotectonised end moraines are common in front of surge-type glaciers and in areas with permafrost (Evans and Rea, 1999, 2003). No obvious thrust features were observed from the morphology or the one section from the terminal moraine in Kjerulfbreen's forefield. It cannot be excluded that thrusting was one of the mechanism for their formation.

The submarine environment shows a fairly good correspondence to the published surge-type tidewaters models except that CSRs and streamlined lineations were not observed. Both major terminal moraines and retreat moraines occur in front of both surging and non-surging glaciers. Based on the submarine environment alone, it could therefore not be concluded that the glaciers surged in the past (Ottesen and Dowdeswell, 2006; Ottesen et al., 2008; Flink et al., 2015) (Figure 22). However, monitoring of surges with both terrestrial and marine sectors has demonstrated that the surge is likely to affect the whole glacier (Sund et al., 2009; Flink et al., 2015; Lovell et al., 2015; Sobota et al., 2016). The possible reason for the lack of CSRs is discussed in *section 5.3*.

To determine if Protektorbreen has exhibited surge behavior is more complex. Only few CSRs were observed in the forefield and some had vague appearance (Figure 24). Based on their presence alone it can be argued that Protektorbreen has surged in the past (Farnsworth et al., 2016). However, it has also been implied by Ben-Yehoshua (2017) that the ridges close to Harrietbreen's forefield could be the result of intense crevassing due to lateral drag from the difference in velocities when Harrietbreen surged. The landform assemblage is a combination of the surge-type valley and cirque models but resemble to the polythermal model is also high. Due to the extensive fluvial outwash plains and fluvial erosion in the forefield of Protektorbreen the landform assemblage correlates best with the valley glacier model. The abundance of supra- and englacial debris fits though better with the cirque glacier model and

polythermal model. CSRs are not present in the valley and polythermal model but occur rarely in the cirque glacier model (Glasser and Hambrey, 2003; Brynjólfsson et al., 2012, 2014; Brynjólfsson, 2015). What can be concluded for the former surge behavior of Protektorbreen will be discussed further in *sections 5.2.2 and 5.3*.

The high glacier fronts and undulating surface on photographs from 1908 indicate that a mass displacement has occurred from the upper to lower part for all three glaciers. It can be argued that it was a sudden movement and the glacier surface could therefore not adjust to a lower gradient profile (Sund et al., 2009). Patches with crevasses further indicate that rapid or strong changes in stress had occurred resuling from high velocities or the effect of the underlying topography on the glacier surface. These features all indicate that the advance was rather rapid. More detailed study on them would be necessary to conclude if it was due to a surge or a climate drive advance (Copland et al., 2003; Sund et al., 2009).

5.2.2 Timing of surging

Surges were documented from a number of glaciers in the neighboring fjords and on western Spitsbergen around the termination of LIA and soon after it (Liestøl, 1969, 1988; Hagen et al., 1993; Dowdeswell et al., 1995). This list is though not complete and a great amount of glaciers that are considered to have exhibit surge behavior in the past have been identified in addition to the recorded surges (Sund et al., 2009; Farnsworth et al., 2016). However, ice-marginal reconstructions in this study, dating back to 1909/10, do not indicate any advances during the retreat from the glaciers in Trygghamna (Figure 50). Based on that the traffic in the fjord was fairly high during that time due to its location and easy access, it is considered likely that if a surge occurred after the LIA it would have been noted and documented. The location of the major terminal moraine outside the 1909/10 position also supports that the timing of the surge from Harriet- and Kjerulfbreen took place prior to that (Figure 51). The maximum Neoglacial extent for Trygghamna is therefore considered to be, at least partly, related to the surge even. According to that and no historical descriptions found of a surge event in the fjord, it is suggested that the glaciers have not undergone a major surge after 1900. The lack of surge terminal moraines inside the major terminal moraine further supports this interpretation (Flink et al., 2015). However, other scenarios can be expected. It is possible that the terminus did not advance as has been described from other surges and the surge was therefore not documented or that the surged ceased before reaching it final stage and did not leave any geomorphic imprints (Sund et al., 2009, 2014). It can therefore not be excluded that the glaciers were more dynamic during the last century but due to enigmatic evidences it was not documented.

The fairly straight shape of the medial moraines between the glaciers indicates that flow speeds have been similar during their formation (Copland et al., 2003) (Figure 22). However, the orientation of the remains of the interaction medial moraine between Harriet- and Protektorbreen's forefields corresponds better with Harrietbreen's flow direction. Harriet- and Kjerulfbreen are thus considered to have been more active during the latest advance. The morphology of the terminal moraine and the non-discontinuity between the subaqueous terminal moraine and the terrestrial ice-cored moraine in the forefield of Protektorbreen further supports that. If we would consider the three glaciers all to have exhibited rapid advances in the past, this would suggest that Protektorbreen surged first and the other two

followed. The difference in timing of the advances could be the result of un-balanced equilibrium conditions due to loss of back-stress as other glaciers retreat. Similar setting has been suggested in other fjords on Svalbard (Farnsworth et al., 2017b).

5.3 Formation and preservation potential of landforms

5.3.1 Formation
One of the controlling factors for the formation of landforms, such as CSRs, end moraines and flutes, is the availability of sediments and deformable beds (Brynjólfsson et al., 2014). Lønne (2016) described that glaciers in land-terminating basins with poorly deformable substrate were not likely to form proglacial moraines because of the coarse grained till. Thrust moraines can also only form in areas with sufficient sediment (Evans and Rea, 1999; 2003). The till cover in all of the forefields was rather thin and discontinuous and bedrock was frequently exposed. In the forefield of Protektorbreen it was especially thin, where only few CSRs were observed. The till there was generally rather coarse grained with lack of finer material. The characteristics of the till in the forefield of Protektorbreen are thus considered unfavorable for the formation of CSRs and flutes. This apparent lack of abundant sediment could as well explain the possible absence of thrust moraine in the forefields.

5.3.2 Preservation potential
In the valley landsystem model it was argued that the absence of CSRs and indistinct appearance of end moraines could be explained by intensive fluvial erosion, lowering their preservation potential (Brynjólfsson et al., 2014; Brynjólfsson, 2015). The abundance of en- and supraglacial debris is as well likely to bury smaller landforms such as CSRs as has for example been explained in the cirque model (Brynjólfsson et al., 2012). Fluvial outwash is rather limited and is mostly located proximal to the glacier fronts of Harriet- and Kjerulfbreen. However, fluvial activity is much more intense in the forefield of Protektorbreen where fluvial outwash plains and active and abandoned fluvial channels are widespread together with supraglacial debris (Figure 24). The intense fluvial erosion and high amount of supra- and englacial debris in the forefield of Protektorbreen therefore lowers the preservation potential of CSRs and other landforms.

Subaqueous CSRs have usually a relatively low relief with an average height of 2 m but often less than 1 m. Therefore they are likely to be draped and eventually buried by glaciomarine sediments with time (Flink et al., 2017). Sedimentation rates have been estimated to be between 0.027 and 0.1 cm a^{-1} in Isfjorden during the Mid and Late-Holocene. However, higher rates have been estimated in outer Trygghamna due to comparatively high precipitation and low-grade metamorphic rock that is relatively easily eroded by water (Forwick, 2005). The amount of meltwater coming from the glaciers and Lovénvatnet indicates that it is also high in the inner part. The timing of the surge is suggested to be prior ~1900 and thus glaciomarine sediment has been accumulating for around 100 years in the distal part. To bury a 2 m high ridge for 100 years the sedimentation rates would need to be ~2 cm a^{-1}, which is considered well possible compared to other fjords in Svalbard. In the forefield of Tunabreen, a surge-type glacier in Svalbard, the sedimentation rates are 3.8 cm a^{-1} close to the glacier front (Forwick et al., 2010).

The CSRs are as well thought to form during the termination of the surge (Lovell et al., 2015) but the retreat moraines during the subsequent retreat (Ottesen et al., 2008; Flink et al., 2015). They could therefore easily overprint the CSRs during their formation and/or incorporate them into the moraines. Likewise, Ottesen et al. (2008) discussed the similarities between CSRs and retreat moraines and the difficulties in identifying between them. The dimension of the retreat moraines, in comparison to the previously described CSRs supports that they were rather formed during retreat. It can though not be excluded that CSRs are present but due to their small dimension they were not observed. It is therefore not concluded if CSRs were formed in the submarine environment or if they were not observed due to their small dimension or were not preserved due to their low preservation potential.

Another important aspect is that the preservation potential is usually considered much greater in the marine part due to no fluvial erosion and permafrost degradation, resulting in more complete landform assemblage (Evans and Rea, 1999, 2003; Schomacker and Kjær, 2007; Ottesen et al., 2008; Flink et al., 2015). The record from Trygghamna however demonstrates that if CSRs were formed in the submarine environment the preservation potential of CSRs is higher in the terrestrial part. That would also be caused by difference between the stagnant terrestrial glacier front preserving the landforms versus the active marine based front that might overprint landforms (*section 5.5*). The fluvial activity is as well rather restricted in the forefield of Harriet- and Kjerulfbreen, which would result in little erosion, compared to Protektorbreen, where CSRs were not observed. This emphasizes the difference between environments and the importance in incorporating both terrestrial and marine data when possible.

5.4 Application of landsystem models

The surge-type landsystem models have been used to identify previously undocumented surge-type glaciers, where historical data and direct observations are lacking (Ottesen and Dowdeswell, 2006; Evans et al., 2010; Evans, 2011; Grant et al., 2009; Brynjólfsson et al., 2012). CSRs have also been used solely as a simple method for the same purpose (Farnsworth et al., 2016). The identification of new surge-type glaciers could improve our understanding of the relationship between glacial dynamics and climate (Evans and Rea, 1999, 2003; Brynjólfsson, 2015) and the reason behind individual glacier advances, to be able to conclude if it was driven directly by climate (Lefauconnier and Hagen, 1991; Yde and Paasche, 2010; Farnsworth et al., 2016). Surging glaciers have also been considered as a modern analogue to terrestrial paleo-ice streams and surging ice-sheet lobes (Evans and Rea, 1999, 2003; Evans et al., 2007, Kjær et al. 2008; Ottesen et al., 2008; Schomacker et al., 2014) with similar landform assemblage although of different scale (Kjær et al., 2008). They have been used to identify traces of paleo-ice streams (Evans et al., 1999, 2008) but further investigations are needed to demonstrate if it is possible to scale the landsystem model (Ingólfsson et al., 2016).

The terrestrial and marine forefields of Protektor-, Harriet- and Kjerulfbreen do not show a good correspondence to the landsystem models presented in this study but can be correlated to different factors in all of them. This is considered to be because the glaciers in

Trygghamna fit into different glacial classifications, valley- and tidewater glaciers. The glaciers have also evolved through time from being tidewater glaciers to eventually become land terminating. The models are on the other hand based on more simplified static environments. The influence the glaciers have on each other is as well a factor that needs to be taken into account, especially if they exhibited different dynamics and on different timescale. The CSRs in Protektorbreen's forefield might be an example of that if they were formed by the surge of Harrietbreen.

Many of the landforms can form in front of both surging and non-surging glaciers and therefore it has been emphasized that it is the landforms assemblage instead of individual landforms that are indicative of a surge (Ottesen et al., 2008), although CSRs and concertina eskers are thought to be diagnostic for surge-type glaciers. The low preservation potential and limiting factors for forming landforms also indicates that the absence of evidence is not the evidence of absence (Benn and Evans, 2010; Ingólfsson et al., 2016) as might be the case with Protektorbreen. This problem is highlighted by the polythermal model by Glasser and Hambrey (2003) that was considered by them as a non-surging glacier. However, other investigations suggested that it was a surge-type glacier despite the lack of CSRs as historical data and glaciological features were in favor of that. Another example is terrestrial retreat moraines. They are generally only present in the fjord surge-type models. Their presence has been used as evidence that a glacier is non-surging (Kjær et al., 2003; Brynjólfsson et al., 2012). They have however been described in the forefields of terrestrial surge-type glaciers (Johnsson et al., 2010, Jónsson et al., 2014) demonstrating that exceptions can occur and highlights that there is a certain danger in using them for the identification of undocumented surge-type glaciers.

Despite frequent correspondence between the models exceptions occur supporting the effect of the glacier's environment (Ottesen et al., 2008; Brynjólfsson et al., 2012, 2014; Brynjólfsson, 2015; Lønne, 2016). Ingólfsson et al. (2016) addressed that one of the outstanding research question for surge-type glaciers is if the landform-sediment assemblages is variable after the glacier environment and how the surges affect the sediment distribution. The development of landsystem models in different environments will increase our understanding of this difference and add to our knowledge of surge-type glaciers.

5.5 Implications for thermal regime

The three larger glaciers all show evidences of a former warm based thermal regime by the presence of flutes and striations during the LIA advance (Clark, 1999). The set of parallel lateral meltwater channels at the terrestrial margins of the glaciers and in their forefields is indicative of cold based margins during their retreat (Dyke et al., 1993) as in the majority of Svalbard's glaciers that are polythermal, with warm based interior but cold based front (Hagen et al., 1993; Petterson, 2004). The transverse ridges in front of Harrietbreen indicate that part of the glacier front was still active during their deposition. A subglacial tunnel under the medial moraine between the glaciers further supports a polythermal origin for Harriet- and Kjerulfbreen (Benn and Evans, 2010). Protektorbreen is however most likely close to being entirely cold based today based on ground penetrating radar from 2014/15 that only

showed restricted areas of temperate ice in the deeper parts and its appearance (Andy Hodson, personal communication) coinciding with the work of Sevestre et al. (2015) from Svalbard that show how glaciers can switch from being polythermal to cold based due to loss of mass. If it is expected that they behave similar to many small glacier in Svalbard, they will change into cold based glaciers due to their loss of ice thickness and not considered to be able to surge under present climatic conditions (Dowdeswell et al., 1991; Sevestre et al., 2015).

The geomorphological record demonstrates how the glaciers behaved differently between the terrestrial and marine environments during the retreat as a result of differences in the thermal regime. The transverse ridges have been interpreted in this study to form during the overall retreat of the glaciers. Therefore they indicate that the ice-margin was not frozen to its bed and must have been active in the submarine environment. The almost lack of transverse ridges, high preservation of the CSRs in the terrestrial part and the lateral meltwater channels indicate, on the contrary, that the terrestrial glacier margin must have been cold based and inactive during the retreat to be able to preserve the landforms. During the quiescent phase of a surge, the terrestrial glacier margin becomes stagnant and wasted down in situ. The surge behavior that has been suggested in Trygghamna for at least two glaciers would thus further lead to the high preservation of landforms. Retreat moraines are rarely found in the terrestrial forefields of surge-type glaciers (Evans and Rea, 1999, 2003). A similar scenario was described for the tidewater glacier Nordenskiöldbreen in Svalbard by Allaart (2016), where terrestrial and marine archives were combined. Previous studies have mostly focused on one of the environment. With the ongoing retreat of glaciers in Svalbard, areas with combined terrestrial and marine forefields will get revealed. This increases the opportunities to investigate the connection between the areas in close setting to examine the behavior of glaciers in variable setting.
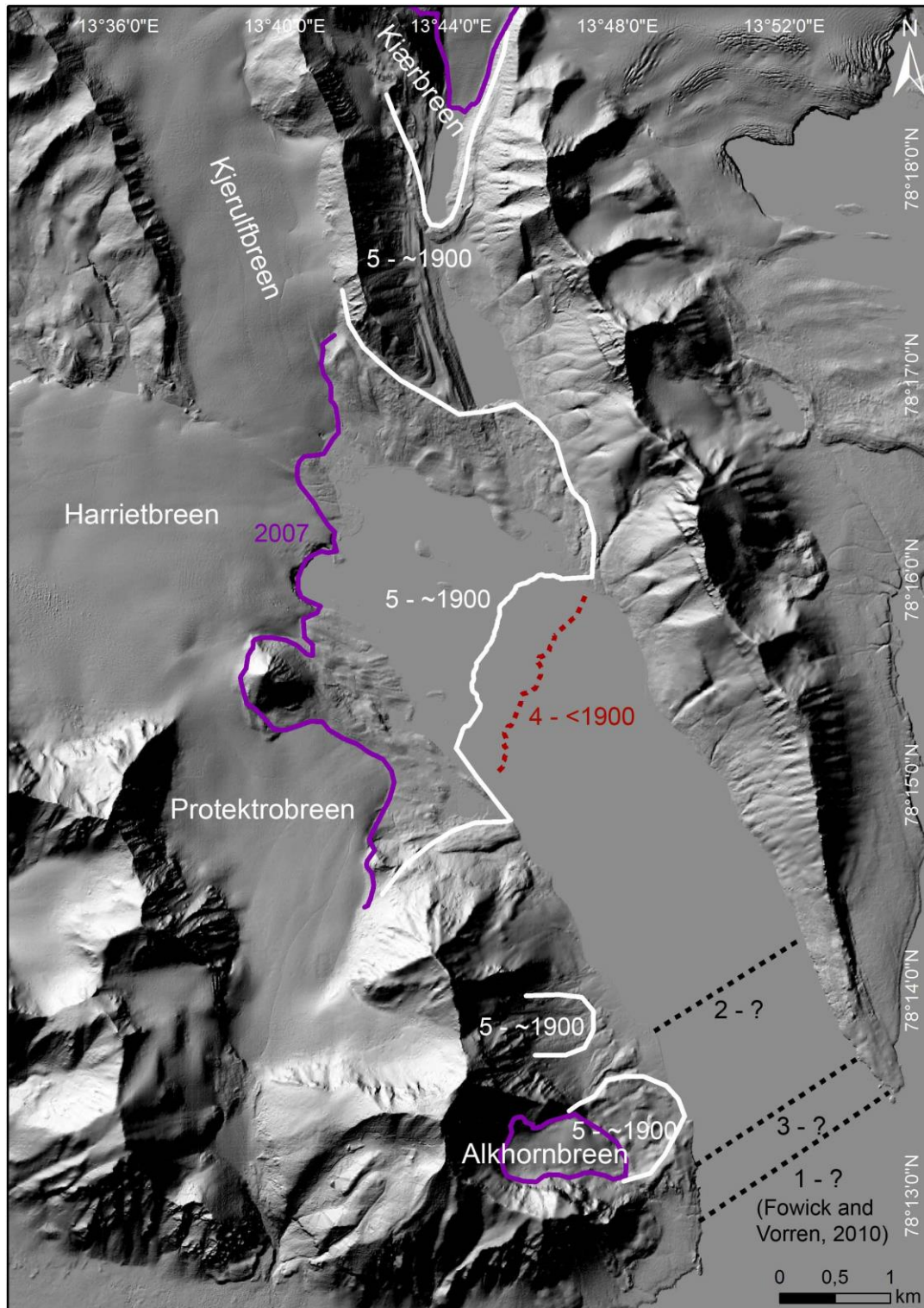
Figure 51. *Generalized reconstruction of the glacier fluctuations in Trygghamna based on this study and Forwick and Vorren (2010) (Figure 14). The numbers represent the order of the events. As there is no age constrains on events 1-3, it is marked with a question mark but the relative age is based on the stratigraphy. Event 4 (red line) is the suggested surge for Harriet- and Kjerulfbreen, considered to occur prior to 1900 AD. Event 5 (white line) is the culmination of the LIA, ~1900 AD. The white and red lines are considered to be the Neoglacial maximum extent. The purple line demonstrates the glacier margins in 2007(GLIMS).*

6. Further studies

This study was able to use the geomorphology and ice-margins to reconstruct the glacial history and suggest a more dynamic behavior for the glaciers in the past. However, due to limitations of the project there are certain uncertainties that could be improved with further studies in the area, presented in the following section.

To be able to combine the glacial history of Trygghamna into greater context to Spitsbergen, an age constraint from the landforms would be needed. Datable material from the raised beaches and a sediment core from the outer part of the submarine environment would provide an estimate of the timing of regional deglaciation in the area (Forman, 1990; Forman et al., 2004; Hogan et al., 2011). That, together with more detailed geophysical investigations of the seafloor, would give a better knowledge about the genesis of the moraines observed in the acoustic data and could add to our understanding of the deglaciation history in Svalbard, as recent work suggests that ice retreat was more dynamic than previously though (Farnsworth et al., 2017a). Sub-bottom profiles proximal to the terminal moraine could provide a better insight into the formation of the retreat moraines and the dynamics of the glaciers during the retreat. Profiles could as well reveal the thickness of glacigenic sediment and if it is possible that the CSRs have been buried underneath glaciomarine sediments (Flink et al., 2017). A sediment core from Lovénvatnet could provide a better age constrain of the timing of Neoglacial advances and the climate variability during the Holocene as has been done by other glacial fed lakes in Svalbard (van der Bilt et al., 2015; Røthe et al., 2015). The surge signal might even be capture in the lake by higher sedimentation rates, which hopefully would provide age constrain for it (Striberger et al., 2011). Detail sedimentological investigations on landforms, such as the CSRs, terminal and hummocky moraines, would be needed to determine further about their genesis and if thrusting occurred. Recent work has highlighted the importance of detailed studies on the internal structure of landforms in relation to the glacier dynamics and paleoglaciology. Excavation of large sections would be the best approach (Hambrey et al., 1997; Kjær and Krüger, 2001; Schomacker and Kjær, 2008; Benediktsson et al., 2008, 2009, 2010). The vague ridges in the forefield of Protektorbreen should be examined in more detail, if they are visible in the field.

The old photographs provide an example of what could be done by further investigations on the glaciological features. Detailed work on photographs and aerial images from different time is suggested (Sund et al., 2009, 2014) as different lines of evidence can be seen from the geomorphic and glaciological imprints. This approach highlights the importance of combining the two data sets, when appropriate, to study the former behavior of the glaciers.

To understand the mechanism of surge-type glaciers and their behavior a close monitoring through a surge cycle and the process occurring after it would be needed. As has been stated the landscape alters quickly in the dynamic glacier's forefields. Time series of aerial photographs shortly after a known surge could reveal landforms as the glacier front melts. The glaciers also retreat during their quiescent phase and new landforms could potentially be revealed in the future (Brynjólfsson et al., 2014, Brynjólfsson, 2015). Dead-ice melting can also result in severe changes in the landscape of ice-cored landforms, which can develop into areas of hummocky landscape with time (Schomacker and Kjær, 2007, 2008; Ólafsdóttir,

2011). A geomorphological map is therefore only capturing a snapshot from the time it was made. Ongoing monitoring of known surge-type glaciers, preferably of a full surge cycle and the changes in the landscape with time, is therefore important for further research of surge-type landsystem models in different environments (Brynjólfsson, 2015, Lønne, 2016).

Further investigation is needed to determine what caused those glaciers advances, both during deglaciation and the Neoglacial. As surge behavior is implied during the LIA in Trygghamna more study is needed to determine the details of the former advances and if they were driven by a climate controlled event or a more dynamic behavior. This emphasizes the importance in identifying surge-type glaciers and gaining better knowledge about their behavior, both during recent times but also during the Holocene.

7. Summary and conclusions

- The terrestrial parts of all the glaciers were located adjacent to their terminal moraine in ~1900. The tidewater margin was located ~0.5 km inside it in 1909/10. All margins had retreated in 1936. The maximum Neoglacial extent was therefore reached during the culmination of the LIA. Ice-marginal reconstructions do not indicate any advances during the subsequent retreat.

- This study and Forwick and Vorren (2010) have described in total, three halts or readvances during the regional deglaciation based on acoustic data. Their relative age is based on stratigraphy.

- Harriet- and Kjerulfbreen exhibited surge-behavior based on the presence of CSRs. It cannot be concluded with certainty if Protektorbreen has surged in the past as other factors could result in the absence of landforms. The surge is suggested to have occurred prior to ~1900 based on ice-marginal reconstruction in combination with historical data. Further studies are recommended to investigate glaciological features on old photographs and the present day glacier surface, which might give a further insight into the glacial dynamics and the timing of events.

- Important factors affecting the landform assemblage are the basinal conditions; quantity of sediments, their grain size and substratum as well as glacial dynamics and morphology of the surroundings. Preservation potential of landforms can also modify the landscape significantly because of often high fluvial erosion, dead-ice melting and burial by other landforms and sediments. The absence of evidence is therefore not an evidence of absence.

- The previously published landsystem models did not show a good correspondence to the glacier forefields in Trygghamna as they fit into different glacial classifications. This emphasize that they can serve as a good guideline identifying undocumented surge-type glaciers although they should be used with precaution due to the different factors mentioned above.

- The difference between the geomorphological record in the marine and terrestrial environments indicates different glacial dynamics and preservation potential. This highlights that studies reconstructing glacier dynamics should, where appropriate incorporate evidence from both archives.

- Monitoring of surge-type glaciers and the modification of the landscape, in variable environments, is important to improve landsystem models and our understanding of the interaction between the formation of landforms, climate, glacier dynamics and environment.

8. References

Allaart, L. (2016). Combining terrestrial and marine glacial archives: A geomorphological map of the Nordenskiöldbreen forefield, Svalbard. MSc. Thesis, Norwegian University of Science and Technology, Trondheim.

Anderson, J.B. & Ashley, G.M. (eds.) (1991). Glacial marine sedimentation; Paleoclimatic significance. Geological Society of America Special Paper 261. 232 p.

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*(1), 207-223.

Baffin, W. (1881). *The Voyages of William Baffin, 1612-1622* (Vol. 63). London: Printed for The Hakluyt Society.

Ballantyne, C. K. (1997). Periglacial trimlines in the Scottish Highlands. *Quaternary International*, *38*, 119-136

Benediktsson, Í. Ö., Ingólfsson, Ó., Schomacker, A., & Kjær, K. H. (2009). Formation of submarginal and proglacial end moraines: Implications of ice- flow mechanism during the 1963-64 surge of Brúarjökull, Iceland. *Boreas*, *38*, 440–457.

Benediktsson, Í. Ö., Möller, P., Ingólfsson, Ó., van der Meer, J. J. M., Kjær, K. H., & Krüger, J. (2008). Instantaneous end moraine and sediment wedge formation during the 1890 glacier surge of Brúarjökull, Iceland. *Quaternary Science Reviews*, *27*(3–4), 209–234.

Benediktsson, Í. Ö., Schomacker, A., Lokrantz, H., & Ingólfsson, Ó. (2010). The 1890 surge end moraine at Eyjabakkajökull, Iceland: a re-assessment of a classic glaciotectonic locality. *Quaternary Science Reviews*, *29*(3), 484–506.

Benestad, R.E., Hanssen-Bauer, I., Skaugen, T.E. and Førland, E.J. (2002). *Associations between sea-ice and the local climate on Svalbard*. Oslo: Norwegian Meteorological Institute Report No. 07/02, 1-7.

Ben-Yehoshua, D. (2017). Crevasse-Squeeze Ridges in Trygghamna, Svalbard. MSc. Thesis, University of Iceland, Reykjavik.

Benn, D. I. and Evans, D. J. A. (2010). *Glacier and Glaciations* (Second edition). London: Hodder Education.

Bilt, W. G. M. Van Der, Bakke, J., Vasskog, K., Andrea, W. J. D., Bradley, R. S., & Ólafsdóttir, S. (2015). Reconstruction of glacier variability from lake sediments reveals dynamic Holocene climate in Svalbard. *Quaternary Science Reviews*, *126*, 201–218. http://doi.org/10.1016/j.quascirev.2015.09.003

Birks, H. H. (1991). Vegetational history and climatic change in west Spitsbergen - plant macrofossils from Skardtjørna, an Arctic lake. *The Holocene*, *1*, 209–218.

Bjarnadóttir, L.R., 2007. Ice-marginal Environments of Surging Glacier: Distribution, Formation and Morphological Evolution of Flutes and Crevasses Cast Ridges at Brúarjökull, Iceland. M.S. Thesis, University of Iceland, Reykjavik.

Björnsson, H., Pálsson, F., Sigurdsson, O., & Flowers, G. E. (2003). Surges of glaciers in Iceland. *Annals of Glaciology*, *36*, 82–90.

Borstad, C. (2017, February 15). Tunabreen may be surging decades earlier than expected. Retrieved from: http://www.unis.no/tunabreen-may-surging-decades-earlier-expected/

Boulton, G. S. (1970). On the deposition of subglacial and melt-out tills at the margins of certain Svalbard glaciers. *Journal of Glaciology*, *9*(56), 231-245.

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

Boulton, G. S., van der Meer, J. J. M., Beets, D. J., Hart, J. K., & Ruegg, G. H. J. (1999). The sedimentary and structural evolution of a recent push moraine complex: Holmstrombreen, Spitsbergen. *Quaternary Science Reviews*, *18*, 339–371.

http://doi.org/10.1016/S1571-0866(04)80103-9

Boulton, G. S., van de Meer, J. J. M., Harts, J., Beets, D., Ruegg, G. H. J., van der Wateren, F. M., & Jarvis, J. (1996). Till and moraine emplacement in a deforming bed surge - an example from a marine environment. *Quaternary Science Reviews*, *15*(95), 961–987.

Bouvier, V., Johnson, M. D., & Påsse, T. (2015). Distribution , genesis and annual-origin of De Geer moraines in Sweden : insights revealed by LiDAR. *GFF*. http://doi.org/10.1080/11035897.2015.1089933

Brynjólfsson, S. (2015). *Dynamics and Glacial History of the Drangajökull Ice Cap, Northwest Iceland.* Norwegian University of Science and Technology, Trondheim. (Doctoral dissertation, 233 pp).

Brynjólfsson, S., Ingólfsson, Ó., & Schomacker, A. (2012). Surge fingerprinting of cirque glaciers at the Tröllaskagi. *Jökull*, *62*, 153–168.

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. http://doi.org/10.1016/j.geomorph.2014.01.019

Clark, C. D. (1999). Glaciodynamic context of subglacial bedform generation and preservation. *Annals of Glaciology*, *28*(1), 23-32

Clarke, G.K.C., Schmok, J.P. & Ommaney, C.S.L. (1986). Characteristics of Surge- type glaciers. *Journal of Geophysical Research,* 91, 7165-7180.

Copland, L., Sharp, M. J., & Dowdeswell, J. A. (2003). The distribution and flow characteristics of surge-type glaciers in the Canadian High Arctic. *Annals of Glaciology*, *36*, 73–81. http://doi.org/10.3189/17275640378181630

Conway, M. (2012). *No man's land: a history of Spitsbergen from its discovery in 1596 to the beginning of the scientific exploration of the country*. Cambridge University Press.

Dallmann, W. K. (ed). (2015). *Geoscience Atlas of Svalbard. Norwegian Polar Institute Report Series*, (148)

Dickson, R. R., Osborn, T. J., Hurrell, J. W., Meinke, J., Blindheim, J., Adlandsvik, B., Vinje, T., Alekseev, G. & Maslowski, W. (2000). The Arctic Ocean response to the North Atlantic Oscillation. *Journal of Climate*, *13*(15), 2671–2696. http://doi.org/10.1175/1520-0442(2000)013<2671:TAORTT>2.0.CO;2

Divine, D., Isaksson, E., Martma, T., Meijer, H. A. J., Moore, J., Pohjola, V., Roderik, S.W., Wal, van de & Godtliebsen, F. (2011). Thousand years of winter surface air temperature variations in Svalbard and northern Norway reconstructed from ice-core data Svalbard. *Polar Research*, *30*, 1–12. http://doi.org/10.3402/polar.v30i0.7379

Dowdeswell, J. A. (1984). *Remote sensing studies of Svalbard glaciers* (Doctoral dissertation, University of Cambridge).

Dowdeswell, J. A., Hamilton, G. S., & Hagen, J. O. (1991). The duration of the active phase on surge-type glaciers: contrasts between Svalbard and other regions. *Journal of Glaciology*, *37*(127), 388–400.

Dowdeswell, J. A., Hodgkins, R., Nuttall, A., Hagen, J. O., & Hamilton, G. S. (1995). Mass balance change as a control on the frequency and occurrence of glacier surges in Svalbard, Norwegian High Arctic. *Geophysical Research Letters*, *22*(21), 2909–2912.

Dyke, A. S. (1993). Landscapes of cold-centred Late Wisconsinan ice caps, Arctic Canada. *Progress in Physical Geography*, *17*(2), 223-247.Evans, D. J. A. (2003). Introduction to glacial landsystems, *Glacial Landsystems.* Arnold, London, 1-10.

Evans, D. J. A. (2003). Introduction to glacial landsystems. *Glacial Landsystems: London, UK, Arnold,* 1-11.Evans, D. J. A. (2011). Glacial landsystems of Sátujökull, Iceland:

A modern analogue for glacial landsystem overprinting by mountain icecaps. *Geomorphology*, *129*(3–4), 225–237. http://doi.org/10.1016/j.geomorph.2011.01.025

Evans, D. J. A., Clark, C. D., & Rea, B. R. (2008). Landforms and sediment imprints of fast glacier flow in the southwest Laurentide Ice Sheet. *Journal of Quaternary Science*, *23*(3), 249–272.

Evans, D. J. A., Lemmen, D. S., & Rea, B. R. (1999). Glacial landsystems of the southwest Laurentide ice sheet: Modern Icelandic analogues. *Journal of Quaternary Science*, *14*(7), 673–691.

Evans, D. J., & Rea, B. R. (1999). Geomorphology and sedimentology of surging glaciers: a land-systems approach. *Annals of Glaciology*, *28*(1), 75-82.

Evans, D. J. A., & Rea, B. R. (2003). Surging glacier landsystem. *Glacial Landsystems. Arnold, London*, 259-288.

Evans, D. J. A., & Twigg, D. R. (2002). The active temperate glacial landsystem : a model based on Breiðarmerkurjökull and Fjallsjökull, Iceland. *Quaternary Science Reviews*, *21*, 2143–2177.

Evans, D. J. A., Twigg, D. R., & Orton, C. (2010). Sátujökull glacial landsystem, Iceland. *Journal of Maps*, *6*(1), 639–650. http://doi.org/10.4113/jom.2010.1129

Evans, D. J. A., Twigg, D. R., Rea, B. R., & Shand, M. (2007). Surficial geology of the Brúarjökull surging glacier landsystem. *Journal of Maps*, *3*(1), 349–367.

Eyles, N., & Rogerson, R. J. (1978). A framework for the investigation of medial moraine formation: Austerdalsbreen, Norway, and Berendon Glacier, British Columbia, Canada. *Journal of Glaciology*, *20*(82), 99-113.

Farnsworth, W. R., Ingólfsson, Ó., Retelle, M., & Schomacker, A. (2016). Over 400 previously undocumented Svalbard surge-type glaciers identified. *Geomorphology*, *264*, 52–60. http://doi.org/10.1016/j.geomorph.2016.03.025

Farnsworth, W. R., Ingólfsson, Ó, Noormets, R., Allaart, L., Alexanderson, H., Henriksen, M. & Schomakcer A.(in review). Dynamic Holocene glacial history of St. Johnsfjorden, Svalbard. *Boreas*.

Farnsworth, W. R., Ingólfsson, Ó., Allaart, L. Håkansson, L. & Schomacker A. (in review).Advances in deglaciation on Svalbard. *Geology*.

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. http://doi.org/10.1016/j.quascirev.2014.11.006

Flink, A. E., Noormets, R., Fransner, O., Hogan, K. A., & Oregan, M. (2017). Past ice flow in Wahlenbergfjorden and its implications for late Quaternary ice sheet dynamics in northeastern Svalbard. *Quaternary Science Reviews*, *163*, 162–179. http://doi.org/10.1016/j.quascirev.2017.03.021

Forman, S. L. (1990). Post-glacial relative sea-level history of northwestern Spitsbergen, Svalbard. *Geological Society of America Bulletin*, *102*(November), 1580–1590.

Forman, S. L., Lubinski, D. J., Ingólfsson, Ó., Zeeberg, J. J., Snyder, J. A., Siegert, M. J., & Matishov, G. G. (2004). A review of postglacial emergence on Svalbard , Franz Josef Land and Novaya Zemlya, northern Eurasia. *Quaternary Science Reviews*, *23*, 1391–1434. http://doi.org/10.1016/j.quascirev.2003.12.007

Forwick, M. (2005). *Sedimentary processes and palaeoenvironments in Spitsbergen fjords*. University of Tromsø. (Doctoral dissertation, 173 pp).

Forwick, M., Baeten, N. J., & Vorren, T. O. (2009). Pockmarks in Spitsbergen fjords. *Norwegian Journal of Geology*, *89*(1/2), 65-77

Forwick, M., Vorren, T.O., Hald, M., Korsun, S., Roh, Y., Vogt, C., Yoo, K.-C., (2010). Spatial and Temporal Influence of Glaciers and Rivers on: the Sedimentary

Environment in Sassenfjorden and Tempelfjorden, Spitsbergen. In: Geological Society of London, Special Publications, vol. 44, pp. 163-183.

Forwick, M., & Vorren, T. O. (2010). Stratigraphy and deglaciation of the Isfjorden area, Spitsbergen. *Norwegian Journal of Geology*, *90*, 163–179.

Forwick, M., & Vorren, T. O. (2012). Submarine mass wasting in Isfjorden, Spitsbergen. In *Submarine mass movements and their consequences* (pp. 711-722). Springer Netherlands.

Fowler, A. C., Murray, T., & Ng, F. S. L. (2001). Thermally controlled glacier surging. *Journal of Glaciology*, *47*(159), 527–538.

Fransner, O., Noormets, R., Flink, A.E. Hogan, K.A., Ó Regan, M.A. & Jakobsson, M., 2017. Glacial landforms and their implications for glacier dynamics in Rijpfjorden and Duvefjorden, northern Nordaustlandet, Svalbard. Journal of Quaternary Science 32, 437-455.

Førland, E. J., Benestad, R., Hanssen-bauer, I., Haugen, J. E., & Skaugen, T. E. (2011). Temperature and Precipitation Development at Svalbard 1900 − 2100, *Advances in Meteorology*, 2011. http://doi.org/10.1155/2011/893790

Glasser, N. F., & Hambrey, M. J. (2003). Ice-marginal terrestrial landsystems: Svalbard polythermal glaciers. *Glacial Landsystems. Arnold, London*, 65-88.

GLIMS, and National Snow and Ice Data Center. (2005, updated 2012). GLIMS Glacier Database, Version 1. [Glacier values]. Boulder, Colorado USA. NSIDC: National Snow and Ice Data Center. doi: http://dx.doi.org/10.7265/N5V98602 (23.01'17).

Grant, K. L., Stokes, C. R., & Evans, I. S. (2009). Identification and characteristics of surge-type glaciers on Novaya Zemlya, Russian Arctic. *Journal of Glaciology*, *55*(194), 960–972.

Hagen, J. O. (1987). Glacier surge at Usherbreen , Svalbard. *Polar Research*, *5*, 239–252. http://doi.org/10.3402/polar.v5i2.6879

Hagen, J. O., Liestøl, O., Roland, E., & Jørgensen, T. (1993). *Glacier Atlas of Svalbard and Jan Mayen*. Oslo: Norsk Polarinstitutt.

Hald, M., Andersson, C., Ebbesen, H., Jansen, E., Klitgaard-Kristensen, D., Risebrobakken, B., Salomonsen, G. R.,Sarnthein, M. & Telford, R. J. (2007). Variations in temperature and extent of Atlantic Water in the northern North Atlantic during the Holocene. *Quaternary Science Reviews*, *26*(25), 3423-3440.

Hamberg. (1905). Karte der Baie recherché und van Keulen Bay auf Spitzbergen wahrend der von A. G. Nathorst geleiteten Schwedischen Polarexpedition 1898 photogrammetrisch aufgenommen und konstruiert von axel Hamber. Scale 1:100 00. −A. Hamberg, Astronomische, photogrammetrisch und erdmagnetische Arbeiten der von A. G. nathorst geleiteten schwedischen Polarexpedition 1898. Uppsala 1905. [Pl. 1] (vet-ak. Handl. N. F. 39. No. 6.)

Hambrey, M. J., Huddart, D., & Bennett, M. R. (1997). Genesis of ' hummocky moraines ' by thrusting in glacier ice : evidence from Svalbard and Britain. *Journal of the Geological Society*, *154*, 623–632. http://doi.org/10.1144/gsjgs.154.4.0623

Hambrey, M. J., Bennett, M. R., Dowdeswell, J. A., Glasser, N. F., & Huddart, D. (1999). Debris entrainment and transfer in polythermal valley glaciers. *Journal of Glaciology*, *45*(149), 69–86.

Hambrey, M. J., Murray, T., Glasser, N. F., Hubbard, A., Hubbard, B., Stuart, G., Hansen, S. & Kohler, J. (2005). Structure and changing dynamics of a polythermal valley glacier on a centennial timescale: Midre Lovénbreen, Svalbard. *Journal of Geophysical Research: Earth Surface, 110*(F1).

Hamilton, G. S., & Dowdeswell, J. A. (1996). Controls on glacier surging in Svalbard. *Journal of Glaciology*, *42*(140), 157–168.

Hansen, S. (2003). From surge-type to non-surge-type glacier behaviour: Midre Lovénbreen, Svalbard. *Annals of Glaciology*, *36*(1), 97-102.

Hanssen-Bauer, M., Kristensen, S., & Steffensen, E. L. (1990). *The climate of Spitsbergen*. Oslo: Norwegian Meteorological Institute Report No. 39/90.

Hogan, K. A., Dix, J. K., Lloyd, J. M., Long, A. J., & Cotterill, C. J. (2011). Seismic stratigraphy records the deglacial history of Jakobshavn Isbræ , West Greenland. *Journal of Quaternary Science*, *26*, 757–766. http://doi.org/10.1002/jqs.1500

Holtedahl, O. (1912). Geologische Karte uber Das gebiet ostlich des vorlandsundes auf Spitsbergen.

Hoel, A. (1942). Skrifter om Svalbard og Ishavet (Nr. 80).Oslo: *The place-names of Svlabard. Norges Svalbard- og Ishavs-underskøkelser.*

Hormes, A., Gjermundsen, F. E., & Lander, T. (2013). From mountain top to the deep sea e Deglaciation in 4D of the northwestern Barents Sea ice sheet. *Quaternary Science Reviews*, *75*, 78–99. http://doi.org/10.1016/j.quascirev.2013.04.009

Hubbard, B., & Glasser, N. F. (2005). *Field techniques in glaciology and glacial geomorphology*. John Wiley & Sons

Humlum, O. (2002). Modelling late 20th-century precipitation in Nordenskiöld Land, *56*, 96–103.

Humlum, O., Elberling, B., Hormes, A., Fjord-, K., Hansen, O. H., & Heinemeier, J. (2005). Late-Holocene glacier growth in Svalbard , documented by subglacial relict vegetation and living soil microbes. *The Holocene*, *3*, 396–407.

Humlum, O., Instanes, A., & Sollid, J. L. (2003). Permafrost in Svalbard: A review of research history, climatic background and engineering challenges. *Polar Research*, *22*(2), 191–215. http://doi.org/10.1111/j.1751-8369.2003.tb00107.x

Ingólfsson, Ó. (2011). Fingerprints of Quaternary glaciations on Svalbard. *Geological Society, London, Special Publications*, *354*(1), 15–31. http://doi.org/10.1144/SP354.2

Ingólfsson, Ó., Benediktsson, Í. Ö., Schomacker, A., Kjær, K. H., Brynjólfsson, S., Jónsson, S., 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. http://doi.org/10.1016/j.earscirev.2015.11.008

Ingólfsson, Ó., & Landvik, J. Y. (2013). The Svalbard-Barents Sea ice-sheet: Historical, current and future perspectives. *Quaternary Science Reviews*, *64*, 33–60.

IPCC, 2007. Climate Change (2007). The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change. In: Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K.B., Tignor, M., Miller, H.L. (Eds.). Cambridge University Press, Cambridge, United Kingdom and New York, 996 pp.

Isachsen. (1915). Ocean Glacial arctique. Spitsberg. [Partie Nord-Ouest.] Carte dressee sous la direction du captiaine Gunnar Isachsen par l'expedition Norvegienne. 1909-1910. Echelle 1:200 000. – Aiso in: Gunnar Isachsen. Travaux topographiques de l'expedition Iscahsen. 1909-1910. Kristiania 1915. [Annexure.] (Vid.-selsk. Skr. I. 1915. No. 7.

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*(9–10), 1301–1312. http://doi.org/10.1016/j.quascirev.2010.02.020

Jiskoot, H., Boyle, P., & Murray, T. (1998). The incidence of glacier surging in Svalbard: evidence from multivariate statistics. *Computers and Geosciences*, *24*(4), 387–399. http://doi.org/10.1016/S0098-3004(98)00033-8

Jiskoot, H., Murray, T., & Boyle, P. (2000). Controls on the distribution of surge-type glaciers in Svalbard. *Journal of Glaciology*, *46*(154), 412-422.

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. http://doi.org/10.1016/j.geomorph.2013.11.007

Judd, A., & Hovland, M. (2007). Seabed fluid flow—the impact on geology. *Biology and the Marine Environment*

Kamb, B., Raymond, C. F., Harrison, W. D., Engelhardt, H., Echelmeyer, K. A., Humphrey, N., Brugman, M. M. & Pfeffer, T. (1985). Glacier Surge Mechanism : 1982-1983 Surge of Variegated Glacier , Alaska. *Science*, *227*(4686), 1982–1983.

Kjær, K. H., & Krüger, J. (1998). Does clast size influence fabric strength?. *Journal of Sedimentary Research*, *68*(5).

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. M. (2003). What causes till thickness to change over distance? Answers from Mýrdalsjökull, Iceland. *Quaternary Science Reviews*, *22*(15–17), 1687–1700.

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*(1), 5–20.

Kjær, K. H., Larsen, E., Meer, J. Van Der, Ingólfsson, Ó., Kruger, J., Benediktsson, Í. Ö., Knudesen, C. G. & Schomacker, A. (2006). Subglacial decoupling at the sediment / bedrock interface : a new mechanism for rapid flowing ice, *25*, 2704–2712. http://doi.org/10.1016/j.quascirev.2006.06.010

Kristensen, L., Benn, D. I., Hormes, A., & Ottesen, D. (2009). Mud aprons in front of Svalbard surge moraines: Evidence of subglacial deforming layers or proglacial glaciotectonics? *Geomorphology*, *111*(3–4), 206–221.

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*(3), 386–402. http://doi.org/10.1111/j.1502-3885.1999.tb00228.x

König, M., Nuth, C., Kohler, J., Moholdt, G., & Pettersen, R. (2014). A digital glacier database for Svalbard. In *Global Land Ice Measurements from Space* (pp. 229-239). Springer Berlin Heidelberg.

Landvik, J. Y., Alexanderson, H., Henriksen, M., & Ingólfsson, Ó. (2014). Landscape imprints of changing glacial regimes during ice-sheet build-up and decay: a conceptual model from Svalbard. *Quaternary Science Reviews*, *92*, 258–268.

Landvik, J. O., Bondevik, S., I, A. E., Fjeldskaar, W., Mangerud, J. A. N., Salvigsen, O., Siegert, M. J., Svendsen, JI. &Vorren, T. O. (1998). The Last Glacial Maximum of Svalbard and the Barents Sea Area : Ice Sheet Extent and Configuration. *Quaternary Science Reviews*, *17*(97), 43–75.

Landvik, J. Y., Brook, E. J., Gualtieri, L. Y. N., Linge, H., Raisbeck, G., Salvigsen, O., & Francoise, Y. (2013). Be exposure age constraints on the Late Weichselian ice-sheet geometry and dynamics in inter-ice-stream areas, western Svalbard. *Boreas*, *42*, 43–56. http://doi.org/10.1111/j.1502-3885.2012.00282.x

Landvik, J. Y., Ingólfsson, Ó., Mienert, J., Lehman, S. J., Solheim, A., Elverhøi, A., & Ottesen, D. (2005). Rethinking Late Weichselian ice-sheet dynamics in coastal NW Svalbard. *Boreas*, *34*, 7–24. http://doi.org/10.1080/03009480510012809

Lefauconnier, B., & Hagen, J. O. (1991). *Surging and Calving Glaciers in Eastern Svalbard.* Oslo: Norsk Polarinstitutt, 116.

Liestøl, O. (1969). Glacier surges in west Spitsbergen. *Canadian Journal of Earth Sciences*, *6*(4), 895-897.Liestøl, O. (1989). *Kompendium i glasiologi.*

Liestøl, O. (1988). The glaciers in the Kongsfjorden area , Spitsbergen. *Norsk Geografisk Tidsskrift - Norwegian Journal of Geography*, *42*, 231–238.Liestøl, O. (1988). The glaciers in the Kongsfjorden area , Spitsbergen. *Norsk Geografisk Tidsskrift - Norwegian Journal of Geography*, *42*, 231–238.

Liljequist, G. H. (1993). *High Latitudes: A History of Swedish Polar Travels and Research.* Stockholm: The Swedish Polar Research Secretariat. [Polarforskningssekretariatet].

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*(2), 113–133. http://doi.org/10.1002/jqs.902

Lovell, H., & Boston, C. M. (2017). Glaciotectonic composite ridge system and surge-type glaciers: an updated correlation based on Svalbard , Norway. *Springer*, *3*(2), 1–16. http://doi.org/10.1007/s41063-017-0028-5

Lovell, H., Fleming, E. J., Benn, D. I., Hubbard, B., Lukas, S., Rea, B. R., ... & Flink, A. E. (2015). Debris entrainment and landform genesis during tidewater glacier surges. *Journal of Geophysical Research: Earth Surface*, *120*(8), 1574-1595. http://doi.org/10.1002/2015JF003509.Received

Lønne, I. (2005). Faint traces of high Arctic glaciations : an early Holocene ice-front fluctuation in Bolterdalen , Svalbard. *Boreas*, *34*, 308–323. http://doi.org/10.1080/03009480510012971

Lønne, I. (2016). A new concept for glacial geological investigations of surges, based on High-Arctic examples (Svalbard). *Quaternary Science Reviews*, *132*, 74–100. http://doi.org/10.1016/j.quascirev.2015.11.009

Maclachlan, S. E., Howe, J. A., & Vardy, M. E. (2010). Morphodynamic evolution of Kongsfjorden-Krossfjorden, Svalbard, during the Late Weichselian and Holocene. *Geological Society, London, Special Publications*, *344*(1), 195-205.

Mangerud, J., & Landvik, J. Y. (2007). Younger Dryas cirque glaciers in western Spitsbergen : smaller than during the Little Ice Age. *Boreas*, *36*, 278–285. http://doi.org/10.1080/03009480601134827

Meier, Mark, F., & Post, A. (1969). What are glacier surges? *Canadian Journal of Earth Science*, *6*(4), 807–817.

Miller, G. H., Alley, R. B., Brigham-Grette, J., Fitzpatrick, J. J., Polyak, L., Serreze, M. C., & White, J. W. C. (2010). Arctic amplification: Can the past constrain the future? *Quaternary Science Reviews*, *29*(15–16), 1779–1790. http://doi.org/10.1016/j.quascirev.2010.02.008

Murray, T., Strozzi, T., Luckman, A., Jiskoot, H., & Christakos, P. (2003). Is there a single surge mechanism? Contrasts in dynamics between glacier surges in Svalbard and other regions. *Journal of Geophysical Research-Solid Earth*, *108*(B5, 2237). http://doi.org/10.1029/2002JB001906

Nordenskiöld, G. (1892). Redogörelse för den svenska expeditionen till Spetsbergen 1890 : med sex taflor och en karta. Stockholm.

Nordli, Ø. (2010). The Svalbard airport temperature series. *Bulletin of Geography*, (3), 5–25.

Norwegian Polar Institute (2003). *The place names of Svalbard*. (Rapportserie nr. 122). Tromsø: Norwegian Polar Institute.

Norwegian Polar Institute (2017). Svalbardkartet: http://svalbardkartet.npolar.no

Norwegian Hydrographic Service and Norwegian Polar Institute. (2016). The Norwegian Pilot, Volume 7. Sailing directions, Svalbard and Jan Mayen. (3rd edition).

Nuth, C., Kohler, J., König, M., Deschwanden, A. Von, Hagen, J. O., Kääb, A.,Moholdt, G. & Pettersson, R. (2013). Decadal changes from a multi-temporal glacier inventory of. *The Cryosphere*, *7*, 1603–1621. http://doi.org/10.5194/tc-7-1603-2013

Oerlemans, J. (2005). Extracting a Climate Signal from 169 Glacier Records. *Science*, *308*(5722), 675–677.

Ohta, Y., Hjelle, A., Andresen, A., Dallmann, W.K. & Salvigsen, O. 1992: Geological Map Svalbard 1:100,000, sheet B9G Isfjorden. Norsk Polarinstitutt Temakart Nr. 16.

Ottesen, D., & Dowdeswell, J. A. (2006). Assemblages of submarine landforms produced by tidewater glaciers in Svalbard, *111*, 1–16. http://doi.org/10.1029/2005JF000330

Ottesen, D., & Dowdeswell, J. A. (2009). An inter – ice-stream glaciated margin : Submarine landforms and a geomorphic model based on marine-geophysical data from Svalbard. *Geological Society of America Bulletin*, *121*(11–12), 1647–1665. http://doi.org/10.1130/B26467.1

Ottesen, D., Dowdeswell, J. A., Benn, D. I., Kristensen, L., Christiansen, H. H., Christensen, O., Hansen, L., Lebesbye, E., Forwick, M. & Vorren, T. O. (2008). Submarine landforms characteristic of glacier surges in two Spitsbergen fjords. *Quaternary Science Reviews*, *27*(15–16), 1583–1599. http://doi.org/10.1016/j.quascirev.2008.05.007

Ottesen, D., Dowdeswell, J. A., Landvik, J. Y., & Mienert, J. (2007). Dynamics of the Late Weichselian ice sheet on Svalbard inferred from high-resolution sea-floor morphology. *Boreas*, *36*, 286–306. http://doi.org/10.1080/03009480701210378

Ottesen, D., Dowdeswell, J. A., & Rise, L. (2005). Submarine landforms and the reconstruction of fast-flowing ice streams within a large Quaternary ice sheet : The 2500-km-long Norwegian-. *Geological Society of America Bulletin*, *117*(7–8), 1033–1050. http://doi.org/10.1130/B25577.1

Ó Cofaigh, D., & Dowdeswell, J. A. (2001). Laminated sediments in glacimarine environments : diagnostic criteria for their interpretation. *Quaternary Science Reviews*, *20*, 1411–1436.

Ólafsdóttir, E. (2011). *Myndun og þróun krákustígsása við Eyjabakkajökul*. MSc thesis. University of Iceland.

Pettersson, R. (2004). *Dynamics of the cold surface layer of polythermal Storglacier*. Stockholm University. (Doctoral dissertation).

Plassen, L., Vorren, T. O., & Forwick, M. (2004). Integrated acoustic and coring investigation of glacigenic deposits in Spitsbergen fjords. *Norsk Polarinstitutt*, *23*(1), 89–110.

Raymond, C. F. (1987). How do glaciers surge? A review. *Journal of Geophysical Research*, *92*(B9), 9121–9134. http://doi.org/10.1029/JB092iB09p09121

Rea, B. R., & Evans, D. J. A. (2011). An assessment of surge-induced crevassing and the formation of crevasse squeeze ridges. *Journal of Geophysical Research: Earth Surface*, *116*, 1–17. http://doi.org/10.1029/2011JF001970

Report of the Svalbard Commissioner, concerning the Claims to Land in Svalbard Part. I-II. A. Text, Copenhagen 1927. B. Maps, prepared by De norske Svalbardekspedisjoner. Oslo 1927.

Reusche, M., Winsor, K., Carlson, A. E., Marcott, S. A., Rood, D. H., Novak, A., Roof, S., Retelle, M., Werner, A., Caffee, M. & Clark, P. U. (2014). 10Be surface exposure ages on the late-Pleistocene and Holocene history of Linnébreen on Svalbard. *Quaternary Science Reviews*, *89*, 5–12.

Roy, S., Hovland, M., Noormets, R. and Olaussen, S., 2015. Seepage in Isfjorden and its tributary fjords, West Spitsbergen. Marine Geology 363, 146-159.

Roy, S., Senger, K., Braathen, A., Noormets, R., Hovland, M. and Olaussen, S., 2014. Fluid migration pathways to seafloor fluid seepage in inner Isfjorden and Adventfjorden, Svalbard. *Norwegian Journal of Geology* 94, 99-119.

Røthe, T. O., Bakke, J., Vasskog, K., Gjerde, M., Andrea, W. J. D., & Bradley, R. S. (2015). Arctic Holocene glacier fluctuations reconstructed from lake sediments at Mitrahalvøya, Spitsbergen. *Quaternary Science Reviews*, *109*, 111–125. http://doi.org/10.1016/j.quascirev.2014.11.017

Salvigsen, O. (2002). Radiocarbon-dated Mytilus edulis and Modiolus modiolus from northern Svalbard: climatic implications. *Norsk Geografisk Tidsskrift-Norwegian Journal of Geography*, *56*(2), 56-61

Salvigsen, O., Elgersma, A., Hjort, C., Legerlund, E., Liestøl, O., & Svensson, N. (1990). Glacial history and shoreline displacement on Bohemanflya, Spitsbergen, Svalbard. *Polar Research*, *8*, 261–273.

Salvigsen, O., Forman, S. L., & Miller, G. H. (1992). Thermophilous molluscs on Svalbard during the Holocene and their paleoclimatic implications. *Polar Research*, *11*(1), 1–10. http://doi.org/10.3402/polar.v11i1.671

Schomacker, A., Benediktsson, Í. Ö., & Ingólfsson, Ó. (2014). The Eyjabakkajökull glacial landsystem, Iceland: Geomorphic impact of multiple surges. *Geomorphology*, *218*, 98–107. http://doi.org/10.1016/j.geomorph.2013.07.005

Schomacker, A., & Kjær, K. H. (2007). Origin and de-icing of multiple generations of ice-cored moraines at Brúarjökull, Iceland. *Boreas*, *36*(4), 411-425.

Schomacker, A., & Kjær, K. (2008). Quantification of dead-ice melting in ice-cored moraines at the Holmströmbreen, Svalbard. *Boreas*, *37*, 211–225. http://doi.org/10.1111/j.1502-3885.2007.00014.x

Serreze, M. C. & Barry, R. G. (2014. *The Arctic Climate System* (2 ed.). USA: Cambridge University Press. 404 pp.

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*(228), 646–662. http://doi.org/10.3189/2015JoG14J136

Sevestre, H., Benn, D. I., Hulton, N. R. J., & Bælum, K. (2015). Thermal structure of Svalbard glaciers and implications for thermal switch models of glacier surging. *Journal of Geophysical Research : Earth Surface*, *120*, 2220–2236. http://doi.org/10.1002/2015JF003517.

Sharp, M. J. (1985a). Landform Ridges-A Landform Type Characteristic of Surging Glaciers ? *Geografisker Annaler, Series A, Physical Geography*, *67*(3–4), 213–220.

Sharp, M. (1985b). Sedimentation and Stratigraphy at Eyjabakkajökull-An Surging Glacier. *Quaternary Research*, *24*, 268–284.

Sharp, M. (1988a). Surging glaciers : geomorphic effects. *Progress in Physical Geography*, *12*(4), 533–559.

Sharp, M. (1988b). Surging glaciers : behaviour and mechanisms. *Progress in Physical Geography*, *12*, 349–370. http://doi.org/10.1177/030913338801200302

Sobota, I., Weckwerth, P., & Nowak, M. (2016). Surge dynamics of Aavatsmarkbreen , Svalbard , inferred from the geomorphological record. *Boreas*, 1–17. http://doi.org/10.1111/bor.12160

Steffensen, E. (1982). *The climate at Norwegian Arctic Stations.* Oslo: Norwegian Meteorological Institute Report No. sept. 1982 (5).

Striberger, J., Björck, S., Benediktsson, Í. Ö., Snowball, I., Uvo, C. B., Ingólfsson, Ó., & Kjær, K. H. (2011). Climatic control of the surge periodicity of an Icelandic outlet glacier. *Journal of Quaternary Science*, *26*(6), 561-565

Storå, N. (1989). Svalbard-centre and periphery: Aspects of cultural research. *Acta

Borealia, *6*(1), 10-20.

Sund, M., Eiken, T., Hagen, J. O., & Kääb, A. (2009). Svalbard surge dynamics derived from geometric changes. *Annals of Glaciology*, *50*(52), 50–60. http://doi.org/10.3189/172756409789624

Sund, M., Eiken, T., & Rolstad, C. D. (2011). Velocity structure , front position changes and calving of the tidewater glacier Kronebreen , Svalbard. *The Cryosphere*, *5*, 41–73. http://doi.org/10.5194/tcd-5-41-2011

Sund, M., Lauknes, T. R., & Eiken, T. (2014). Surge dynamics in the Nathorstbreen glacier system, Svalbard. *The Cryosphere, 8*(2), 623.

Svendsen J.I., Alexandersson H., Astakhov V., Demidov J., Dowdeswell J.A., Henriksen M., Hjort C., Houmark-Nielsen M., Hubberten H., Ingólfsson Ó., Jakobsson M., Kjær K., Larsen E., Lokrantz H., Luunka E.P., Lysa A., Mangerud J., Maslenikova O., Matioushkov A., Murray A., Möller P., Niessen F., Saarnisto M., Siegert M., Stein R. & Spielhagen R. (2004). Ice sheet history of northern Eurasia. *Quaternary Science Reviews 22*, 1229– 1271.

Svendsen, J. I., Elverhøi, A., & Mangerud, J. (1996). The retreat of the Barents Sea Ice Sheet on the western Svalbard margin. *Boreas*, *25*, 244–256.

Svendsen, J. I., & Mangerud, J. (1997). Holocene glacial and climatic variations on Spitsbergen, Svalbard. *The Holocene*, *7*(1), 45–57.

Thorarinsson, S. (1969). Glacier surges in Iceland, with special reference to the surges of Brúarjökull. *Canadian Journal of Earth Sciences*, *6*(4), 875-882.

Vahl, M. Report of the Svalbard Commissioner, concerning the Claims to Land in Svalbard Part. I-II. A. Text, Copenhagen 1927. B. Maps, prepared by De norske Svalbardekspedisjoner. Oslo 1927. *Geografisk Tidsskrift*, *32*.

Vorren, T. O., Landvik, J. Y., Andreassen, K., & Laberg, J. S. (2011). *Glacial History of the Barents Sea Region. Quaternary Glaciations - Extent and Chronology* (1st ed., Vol. 15). Elsevier Inc. http://doi.org/10.1016/B978-0-444-53447-7.00027-1

Wallin, Å. C. (2016). *Investigating methods for paleo surge-type glaciers of highly dynamical ice flows in Trygghamna, west Spitsbergen*. MSc. Thesis, University of Stockholm.

Warren, W. P., & Ashley, G. M. (1994). Origins of the ice-contact stratified ridges (eskers) of Ireland. *Journal of Sedimentary Research*, *64*(3).

Werner, A. (1993). Holocene moraine chronology, Spitsbergen, Svalbard: lichenometric evidence for multiple Neoglacial advances in the Arctic. *The Holocene*, *2*, 128–137.

Yde, J. & Paasche, Ø. (2010). Reconstructing Climate Change : Not All Glaciers Suitable. *EOS*, *91*(21), 189–196.