Dynamics of surging tidewater glaciers in Tempelfjorden Spitsbergen

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Preface

This Master’s thesis is Anne Flink’s degree project in Physical Geography and Quaternary Geology at the Department of Physical Geography and Quaternary Geology, Stockholm University. The Master’s thesis comprises 60 credits (two terms of full-time studies).

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Cover Picture: Tunabreen Glacier 2004, NERC (National Environmental Research Council)
Abstract

Terrestrial glacial geomorphology has long been used to evaluate the extent, chronology and dynamics of former glaciers and ice sheets. New marine geophysical methods provide an opportunity to study the glacial submarine morphology of modern continental shelves and fjord systems. This makes it possible to study landform assemblages in the submarine settings that are often better preserved than their terrestrial counterparts.

This study focuses mainly on the recent surge history of the tidewater glacier Tunabreen, which calves into Tempelfjorden in Western Spitsbergen. Tunabreen is a small outlet glacier of the Lomonosovfonna ice cap and has experienced several surges and terminal retreats during the last century. The multiple surge events have most likely removed or reworked landform assemblages created by earlier surges, resulting in a complex geomorphological imprint on the bed of Tempelfjorden.

Tunabreen has left a specific morphological imprint on the sea floor, consisting of ice flow-parallel lineations and generally flow-transverse retreat moraines. Comparison of retreat moraines mapped from high resolution multibeam bathymetric data and glacier terminal positions, established using remote sensing imagery suggest that the moraines in the inner part of Tempelfjorden are annually formed recessional moraines, formed during winter still stands of the glacier margin or during its minor readvances. Although detailed reconstruction of glacier surge dynamics based solely on the landform distribution is challenging, it is evident that Tunabreen has experienced fast flow during surges and semiannual retreat of the margin after the surges.

The main achievements of this study are a spatial reconstruction of the dynamics of Tunabreen, which has experienced three surges during the last hundred years. Together with the Little Ice Age surge of the adjacent von Postbreen, four recent surges have been recorded in Tempelfjorden since 1870, which distinguishes the study area from earlier studied Svalbard tidewater surge glacier settings, where the glaciers have been known to surge only once or twice. However a detailed understanding of surge triggering mechanisms and their role in controlling the dynamics of the tidewater glaciers in Svalbard is still poor and requires further investigations. Svalbard, where most of the small outlet glaciers are believed to be of surge type, is an excellent natural laboratory for such investigations.
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1. Introduction

Marine ice margins play an important role when it comes to understanding the connection between climate, marine processes and ice dynamics. It is vital to understand the behavior of floating ice tongues and marine ice margins in a world subjected to global climate warming. Ice shelves and floating glacier tongues control the stability of the attached inland ice sheet through their interaction with the ocean by, for example, contributing to faster glacier flow, higher mass wasting and enhanced sea level rise (Joughin and Alley 2011, Holland et al. 2008, Shepherd and Wingham 2007). Despite this, there still exists a lack in knowledge when it comes to fully understanding marine ice marginal processes. In the 4th IPCC report, the representation of marine ice marginal processes in contemporary numerical models was considered so uncertain that ice marginal processes were excluded from the assessment. To improve and evaluate numerical models it is important to co-evaluate model output and geological evidence.

Marine ice margins are investigated both in Antarctica and Greenland (Scheuer et al. 2006, Jakobsson et al. 2011, Holland et al. 2008), but can also be studied in more accessible sites such as Svalbard, where geological evidence related to marine ice margins and dynamics can be studied on a smaller scale. This can for example be done by investigating smaller surging tidewater glaciers. Sund et al. (2009) define a glacier surge as a cyclical velocity fluctuation, which develops due to internal changes in the glacier system. Therefore, a surge reflects internal changes in the glacier system rather than changes in climate. The internal dynamics of surge type glaciers are however still poorly understood and several processes have been proposed as surge triggering mechanisms.

Surging tidewater glaciers leave a characteristic morphological imprint on the seafloor, which can be studied with marine geophysical methods. The often well-preserved glacial morphology of the seafloor provides a possibility to reconstruct the former extent and dynamics of the glaciers. It provides an opportunity to investigate the link between submarine landform assemblages and observed surge dynamics in order to better understand the mechanisms behind surge events, when co-evaluated with satellite and/or modeled data.
2. Aim of study

I have studied the small tidewater glacier Tunabreen in Tempelfjorden (see fig. 8), Western Spitsbergen focusing on the submarine glacial morphology and the surge event records established from marine geophysical and remote sensing data. The aim of the project is to map the glacial morphology in Tempelfjorden near the margin of the tidewater glacier Tunabreen and a land-terminating glacier of von Postbreen. The further objective of the study is to map the recent terminal fluctuations of Tunabreen, in order to link the submarine landforms to observed glacier margin fluctuations. The surge history of the glacier has been evaluated by combining submarine records and terrestrial geomorphological data. The submarine sea floor of Tempelfjorden has been studied by the use of marine geophysical methods, while the surge history of Tunabreen has been examined by the use of terrestrial morphological and ice margin data derived from satellite images.

The main questions investigated are:

- Which landform assemblages can be found at the margin of surging tidewater glaciers?
- How are different submarine landforms related to glacial surge events?
3. Background

3.1. Quaternary glaciations of Svalbard

Glacier studies on Svalbard have been conducted for more than a hundred years and there exists a long tradition in glacier research on the islands (Österholm 1978). Already during the late 19th century, Nordenskjöld (1866) proposed that presently ice free parts of Svalbard had been covered by an ice sheet during the Weichselian period. In 1900, De Geer (1900) suggested that the Scandinavian ice sheet had been connected to the Svalbard ice sheet by extensive sea ice. He was the first to propose that a large marine based ice sheet had once covered Barents Sea. During the 1960s, several Swedish expeditions were launched to Svalbard, with the aim to study the glacial history of the area (Österholm 1978). The expeditions conducted ice movement studies of the Weichselian ice sheet, and confirmed that large parts of Svalbard were covered by an ice sheet, during the Weichselian (Schytt et al. 1968, Hoppe et al. 1969, Österholm 1978).

The extent and chronology of the Svalbard-Barents Sea Ice Sheet (SBSIS) has been discussed for decades (Svendsen et al. 2004, Ingolfsson and Landvik, 2013). The suggestions have ranged from a region completely covered by grounded glacier ice to an area with almost no glaciers (Vorren et al. 2011). Recent evidence suggests that Svalbard and the adjoining Barents Sea have been repeatedly covered by a marine based ice sheet during the Quaternary (Laberg et al. 2010, Ingolfsson and Landvik, 2013). Submarine geomorphological records prove that the ice sheet reached the continental shelf break in the Barents Sea during full glacial conditions (Ottesen et al. 2007).

According to Svendsen et al. (2004), at least four prominent glaciations have been recorded in the Northwestern Barents Sea during the last 160 000 years. According to Sejrup et al. (2008), the ice sheet reached its largest extent during the Saalian period, when it covered large areas in the north (Fig. 1). The Saalian glacial maximum in Western and Central Europe occurred during marine isotope stage (MIS) 6 (Svendsen et al. 2004). According to Wohlfarth et al. (2008) it is most likely that the Svalbard Barents Sea ice sheet coalesced with the Scandinavian ice sheet during periods of its maximum extent. Svendsen et al. (2004), confirm that this happened at least during MIS 6.

The extent of the Eurasian ice sheet in Late Weichselian (25 – 5 ka) has been reconstructed by Svendsen et al. (2004). Its extent peaked during the Last Glacial Maximum (LGM) approximately -20 ka, when it covered the entire Scandinavia, northern Europe, British Isles, NW Russia, and the Svalbard-Barents Sea, (Fig. 2). According to Landvik et al. (1998), the north western parts of the Barents Sea experienced extensive glaciation during the LGM. Marine evidence from the shelves around Spitsbergen further demonstrates that the area experienced extensive
glaciation during the Late Weichselian (Ottesen et al. 2007, Ottesen and Dowdeswell 2009, Dowdeswell et al. 2010). Svendssen et al. (2004) however point out that the southern and eastern limits of the LGM Eurasian ice sheet are still debated.

Figure 1: Reconstruction of the maximum ice extent during the late Saalian, around 160 – 140 ka. The green stippled line shows the maximum ice extent during the Quaternary (Svendssen et al. 2004).

Figure 2: Reconstruction of the late Weichselian ice sheet during LGM (Svendssen et al. 2004).
More recently, the extent and chronology of the Eurasian and Barents Sea ice sheets have been modeled by numerical ice sheet models, forced by global insolation and sea level changes (Siegert and Dowdeswell 1999, Siegert et al. 2001, Siegert and Marsiat 2001). Siegert and Marsiat (2001) focused on the Late Weichselian Eurasian ice sheet behavior, by comparing atmospheric general circulation model (AGCM), with an ice sheet model to assess the Eurasian climate during LGM. Siegert et al. (2001) used a numerical ice sheet model to reconstruct the thickness and dynamics of the Eurasian ice sheet throughout the whole Weichselian period. The results showed that four major glacial advances existed during the Weichselian and that the extent of the advances became progressively larger.

The ice sheet models used in the above mentioned studies belong to the old generation of shallow-ice-approximation ice sheet models, so called zero order models. The zero order models are however insufficient, when for example ice streams shall be simulated. Therefore simulations of coupled ice sheet, ice shelf and ice sheet-ice stream systems, require higher order models (Kirchner et al. 2011). This is predominantly important if the results of the models are aimed to complement spatial ice flow reconstructions, based on higher resolution geophysical or geological data.

Modeled data show that fast ice streams flowed through the bathymetric troughs in Svalbard (Siegert et al. 2001). However, zero order simulations may falsely give the impression of ice streams in topographic troughs, because of the higher local ice thickness (Kirchner et al. 2011). Submarine geomorphological features, such as large trough mouth fans, provide evidence of fast flowing ice streams in the Barents Sea area (Andreassen et al. 2008). According to Sejrup et al. (2000), it is likely that up to 20 large ice streams drained the Svalbard-Barents Sea ice sheet during the Late Weichselian glaciation. The ice streams switched on and off repeatedly and discharged large amounts of sediment to the through mouth fans at the shelf break (Ottesen et al. 2007). The paleo-ice streams left geomorphological evidence in the cross shelf troughs, in the form of mega-scale glacial lineations (MSGL), several kilometers long lateral moraines and large ice-marginal grounding-zone accumulations (Sejrup et al. 2000).

Ice streams are an important dynamical feature of an ice sheet, since they affect ice sheet mass balance and stability (Winsborrow et al. 2010). According to Stokes and Clark (2001) an ice stream can be defined as an artery of fast flowing ice in an ice sheet or ice cap, which is often bordered by highly crevassed slower ice flow on both sides (Cogley et al. 2011). Understanding ice stream dynamics is important for understanding surging glaciers, through analogy reasoning, since both surging glaciers and ice streams display the behavior of fast ice flow. The main difference between ice streams and surging glaciers is the scale, the boundary conditions and the environment they are embedded in. Glaciers can occur in different settings than
ice streams, for example surging tidewater glaciers and surging mountain glaciers unconnected to an ice sheet or ice cap.

Figure 3: The deglaciation of the North Western Eurasian Arctic (Landvik et al. 1998).

The deglaciation of the North Western Eurasian Arctic began at around 15.4 – 13.3 ka and can be traced in the $\delta^{18}$O record from the central Arctic Ocean (Mangerud et al. 1998). Figure 3 shows the deglaciation of the Svalbard-Barents Sea area. According to Knies et al. (2000) Svalbard experienced a very rapid deglaciation between 14 – 10 ka. Svendssen et al. (1996) point out that the deglaciation of the Svalbard area preceded in a stepwise manner. Figure 4 provides a reconstruction of the stepwise deglaciation which occurred in Isfjorden and Western Spitsbergen.

Around 13 ka, an increase in sediment flux to the fjords suggests increased melting due to a warmer climate. According to Mangerud et al. (1992), the coastal areas of Spitsbergen became ice free at around 13 ka. This relatively warm stage was interrupted by a climatic cooling and glacial readvance at around 12.4 ka, which correlates with the Older Dryas readvance of the Fennoscandian ice sheet (Svendssen et al. 1996). According to Vorren et al. (2011), the deglaciation pattern of the Spitsbergen fjords is still debated and it has been postulated that the glacier fronts in Isfjorden were located far back in the inner fjords during the Younger Dryas cold period (Mangerud et al 1992), which would provide no geomorphological evidence of a younger Dryas readvance in the area.
3.2. Present day glacier cover of Svalbard

According to Solheim et al (1991), nearly 60% of Svalbard is covered by glaciers, with more glacier-covered areas in its northern and the eastern parts. The ice caps and glaciers of the Svalbard archipelago are to a large extent influenced by the extent and distribution of sea ice (Ottesen et al. 2008). Sea ice has a big influence on Svalbard glaciers and ice caps, since it prevents warm ocean currents from reaching the shores and creates a buffer for tidewater glaciers. In turn, the sea ice conditions are influenced by ocean currents, climate and the Arctic and North Atlantic Oscillations. Most of Svalbard’s glaciers are of sub-polar type (Solheim et al. 1991), which means that they have a polythermal regime (Murray et al. 2003).

Local topographic and climatic factors influence the location, size and dynamics of Svalbard’s glaciers. Svalbard is located in a climatically sensitive area, where air masses and ocean currents of different thermal character meet (Humlum et al. 2007). According to Dowdeswell et al. (1995) the climate of Svalbard is influenced by the relatively warm ocean currents which reach far north at the coast, as well as by atmospheric depression tracks, which meet polar air masses and contribute to the influence of varying air masses. Wind is an important agent in Svalbard due to the lack of high vegetation. This contributes to significant redistribution of snow during
the winter months. Local wind directions vary due to topographical channeling effects, but several meters of snow can accumulates on the lee sides of mountain slopes and on the glaciers (Eckerstorfer and Christiansen, 2011).

Many of Svalbard’s glaciers and ice caps are of surge type, (see Fig. 5, for the location of surge-type glaciers) (Murray et al. 2003, Hagen et al. 2005). Up to 90% of the glaciers in Svalbard have been described as surge-type glaciers by (Lefauconnier and Hagen 1991), whereas others, like Jiskoot et al. (1998) have defined only 13% of the Svalbard’s glaciers as surge-type glaciers. Leufaconnier and Hagen (1991) determined the approximate percentage of surge glaciers in Svalbard by mapping former terminal positions from aerial images, satellite data and literature search. According to Jiskoot et al. (1998), many of the glaciers defined as surge glaciers during the earlier study, rather reflect the maximum glacier extent during LIA and do not provide evidence for glacier surges.

According to Sund et al. (2009), it is more likely that the number of surge-type glaciers is closer to the maximum estimate of 90%, due to the long quiescent phases in Svalbard. Many glaciers in Svalbard have been observed to surge only once, but long quiescent periods may occur between surge events (Ottesen et al. 2008).
Glaciers that have not been observed to surge might therefore still be of surge-type, but with exceptionally long quiescence periods (Sund et al. 2009). In Svalbard many smaller outlet glaciers are connected to larger ice caps. A surge advance of one of the glaciers can therefore affect the glacier upstream and cause a surface lowering of the whole glacier system (Hagen et al. 2005).

Hagen et al. (2005) emphasizes that glacier surging is not related to climatic variations, but the length of the quiescence phase can be affected by climate. According to Sund et al. (2009) climate might accentuate the surge event, since surges in general become less pronounced, when the mass of the glaciers decreases. Hagen et al. (2005) points out that even if the net mass balance is zero a surge type glacier can experience changes in its longitudinal surface profile which shows that other mechanisms than the climate drive glacier surging. Sund et al. (2009) however emphasizes that a glacier surge may accelerate the climate-induced thinning under changing climatic conditions, since a larger part of the glacier is transported to lower elevation.

The glaciers on Svalbard were close to their Holocene maximum extent during the Little Ice Age (LIA), between the 18th to the late 20th century (Solheim et al. 1991). Since the end of the LIA they have been experiencing mass loss, due to a general increase in temperatures since the 20th century (Dowdeswell et al. 1997). The mass losses of the coastal glaciers, which terminate in the fjords, occur mostly by calving. In Svalbard, iceberg calving is most active during the summer, when sea ice is absent (Ottesen at el. 2008).

### 3.3. Mechanisms of glacial surges

A number of possible processes controlling the glacier surges can be identified. These can be divided into the following seven groups (Classification suggested by the author):

- Subglacial bed character
- Thermal control
- Subglacial drainage system
- Mass balance
- Calving margins, changes in heat flux from ocean waters
- Glacier geometry
- Time static processes

Surge-type glaciers are found in certain geographical locations, whilst they are completely absent in other areas (Jiskoot et al. 2000). This suggests that, specific factors are controlling the spatial distribution of surge glaciers and therefore specific conditions are required for surges to occur (Hamilton and Dowdeswell 1996). Since
surges are not primarily climatically controlled, other factors must be controlling them.

Glacier flow and surge behavior are controlled, among others, by changes in geothermal heat flux, bed roughness, permeability, pore water pressure, erodibility and ground water (Jiskoot et al. 2000). Jiskoot et al. (1998) however state that the surge-type glaciers on Svalbard overlie more than one type of lithology. Therefore mechanisms affected by lithology, such as pore water pressure and erodibility seem to not be of major importance for surge behavior.

3.3.1. Subglacial bed character

The mechanisms that initiate a surge in a glacier are still not completely understood and several mechanisms have been suggested as the main triggers of glacier surges. Jiskoot et al. (2000) suggest a thermally controlled soft bed mechanism as the main trigger, in which fast glacier flow requires a lubricating basal layer of meltwater or saturated sediments. It is most likely that the sedimentary factors, such as grain size and porosity play an important role in the behavior of surge-type glaciers. According to Winsborrow et al. (2011) subglacial geology might play an important role for ice flow, since it determines the thickness of subglacial till and the sedimentary properties of the substratum. Ice flow might be enhanced by a layer of basal soft sediments, since soft sediments reduce basal resistance to ice flow. According to Jiskoot et al. (2000) glaciers overlying fine-grained sedimentary lithologies have been shown to be more susceptible to surging. Ottesen and Dowdeswell (2006) further state that fast glacier flow in Svalbard, can be linked to deformation of soft, glacimarine fjord sediments.

However, the soft bed mechanism and the subglacial bed character is not a sole explanation for surge-events, since it would imply that all surge-type glaciers require similar basal sediments. Neither does it fully explain why surges occur at a specific time.

3.3.2. Thermal control

Jiskoot et al. (2000) consider the subglacial bed character important, but point out that the thermal properties of a glacier have the greatest influence to surging in Svalbard. Thermal properties influence ice rheology, basal hydrology and can have great influence on flow dynamics. The polythermal glaciers on Svalbard appear to be conductive to surging. But thermal regime data does not exist for most of Svalbard glaciers, which makes it difficult to investigate this hypothesis (Jiskoot et al. 2000).

The propagation of a surge front, which has been observed during several Svalbard surges, suggests that surges may be thermally controlled (Murray et al. 2003). A thermal boundary between warm and cold based glacier ice might cause flow instabilities, since differences in thermal properties within the ice can affect basal
hydrology, ice rheology, and substrate mobility, which might further affect shear stress and basal movement (Jiskoot et al. 2000). This could provide an explanation to why polythermal glaciers surge. Polythermal glaciers are frozen to their beds during the quiescence phase. When ice starts to build up and the pressure melting point is reached, basal melt water leads to elevated pore water pressures at the bed and the glacier starts to surge (Murray et al. 2003).

### 3.3.3. Subglacial drainage system

The subglacial hydrology is undoubtedly important when it comes to ice flow. The subglacial drainage system in combination with basal soft sediments might influence glacier dynamics and could explain surge events. According to Solheim et al. (1991), the subglacial drainage system plays the most important role in surge development, since glacier surging involves buildup of large amounts of pressurized water at the bed or in permeable sediments under the glacier. Christoffersen et al. (2005) have shown that the surge of Elisebreen, in Northwestern Svalbard, occurred because of elevated pore water pressure in a thin layer of thawed water saturated till.

In the section above the geometry of the subglacial hydrological system has been unspecified, but different hydrological systems have been observed. These are the linked cavity system and the channel/tunnel flow system. It is possible that different subglacial systems are active in separate regions. Murray et al. (2003) for example propose that the glaciers on Svalbard act differently during a surge cycle, compared to the glaciers in Alaska. According to Kamb (1987) a change from a basal tunnel based system to a linked cavity system has been proposed as the main mechanism for surging in Alaska. A linked cavity system leads to high basal water pressures, which enhances the glacier flow and lead to surging (Jiskoot et al. 2000). The rapid surge termination on Variegated glacier in Alaska, could be explained by a collapse of the linked cavity system, which would lead to a change in subglacial hydrological systems and reduced ice flow (Kamb 1987).

The linked cavity system hypothesis does however not provide a good explanation for surges in Svalbard, since the termination of Svalbard surges has been observed to be very gradual (Murray et al. 2003). Eskers have also been found in Svalbard surge landform assemblages, further pointing towards a subglacial channel flow system (Ottesen et al. 2008).

### 3.3.4. Mass balance

It has been suggested that there is a relation between surging glaciers and mass balance (Dowdeswell et al. 1995). Local topoclimatic effects such as wind patterns, snow distribution, topographic shading and solar insolation, might influence the glacier mass balance and impose a control on surging (Jiskoot et al. 2000). Changes in
mass balance are related to glacial geometry and might affect glacier length and the steepness of the glacier’s slopes. It is however unlikely that differences in glacier mass balance alone control surge events.

3.3.5. Calving margins, changes in heat flux from ocean waters
Surge-type glaciers may end on land or in water. So influence from ocean currents is undoubtedly not the only mechanism controlling glacier surges. However it might be an important mechanism when it comes to the dynamics of tidewater glaciers. According to Winsborrow et al. (2011) calving margins are an effective way of drawing down ice from up glacier and might therefore affect the surface profile of the glacier. Tidewater glaciers may also be affected by ocean currents, and calving might be enhanced due to oceanographically controlled factors, such as changes in heat transport by ocean currents. Oceanographic factors may lead to undercutting of the glacier terminus by melting at or below the waterline (Benn et al. 2007).

3.3.6. Glacier geometry
According to Jiskoot et al. (2003) there is not only a correlation between the probability of surge and fine-grained sedimentary lithologies, but also between, glacier length and the steepness of the glaciers slope. Hamilton and Dowdeswell (1996) tested several geometric and environmental factors on the glaciers in Svalbard and agree that the length of the glaciers, was the only geometric factor that could be related to surging. The statistically conducted study showed that there is an increasing probability to surge with increasing glacier length. According to Jiskoot et al. (2000) this might be explained by the fact that longer glaciers are more sensitive to variations in mass movement between the upper and lower parts of the glacier.

Research of surge glaciers in other regions also shows that a relation seems to exist between glacier length and its surge potential. According to Clarke (1991) surge type glaciers in Canada tend to be longer and have lower average slopes than normal glaciers. This could be explained by the fact that subglacial drainage systems of long glaciers with gentle slopes are more vulnerable to surging. Glacier geometry is however closely linked to mass balance and glacier length could therefore be a proxy for other factors, such as mass balance or thermal conditions at the bed (Jiskoot et al. 2000).

3.3.7. Time static processes
The six surge triggering mechanisms listed above are mechanisms that can change at the same timescales as the surge takes place and even activate feedback-processes. But processes that are considered 'static' on the timescales of the surge can also be considered as mechanisms affecting or triggering a surge. These processes are associated with large scale morphological features, such as valley shapes or bedrock features which do not change during the surge.
Murray et al. (2003) propose that controls on glacier flow can be divided into two groups. The first group is controls that might change during a surge, for example the basal hydrological system, basal sedimentary composition and thermal structures. The second group comprises of controls that would remain constant throughout the surge event, for example bedrock features or valley shapes. According to Murray et al. (2003), the local pattern of strain rate and velocity on Monacobreen was observed to be consistent throughout a surge event during the 1990s. Therefore the factors controlling the spatial distribution of velocity remained fixed throughout the surge event. This suggests that the glacier bed comprises of sticky spots with relatively low velocities in comparison to zones with higher velocities, which remain in the same spatial locations. This would also imply that basal bedrock features are the main controls on small scale glacier velocity changes.

### 3.4. Surge initiation

A typical glacier surge, in Svalbard has been observed to start with an approximately one year long period of steady acceleration, which is followed by a shorter period of rapid acceleration (Murray et al. 2003). According to Sund et al. (2009), glacier surge can be divided into three stages. In the first stage, a surge event can be recognized by an increase in the amount of crevasses, which indicate an increase in flow rates and rapid changes in stresses (Hagen et al. 2005). A build-up towards a surge is further indicated by a surface thickening in the accumulation area and a lowering of the ablation area (Sund et al. 2009). In stage two, surging is initiated when the slope reaches a critical value and sliding increases quickly (Jiskoot et al. 2000). The surge results in a large flux of ice from the higher parts of the glacier towards the terminus, which, in stage 3, results in a surface lowering of the accumulation area and an advance of the glacier front (Hagen et al. 2005). In Svalbard the end of the fast flow phase is very gradual (Murray et al. 2003), which differs from studies of surging glaciers in other regions, such as Alaska, where surge termination has been observed to be very rapid (Kamb 1987).

After surge termination the glacier enters the quiescent phase, the period between surges (Sund et al. 2009). According to Cogley et al. (2011), a quiescent phase can be defined as a period when the flow velocities are lower than in a non-surge type glacier and the ice discharge is too small to maintain the glaciers longitudinal profile. This further implies that surge type glaciers do not experience periods of steady state. It is most likely that the glaciers in Svalbard are polythermal and frozen to their beds during the quiescence phase. This is supported by Murray et al (2003), who have suggested that surging contributes to changes in the polythermal regime. According to Hagen et al. (2005) the longitudinal surface profile of the glacier changes gradually during the quiescence period, when the glacier experiences a
Conflicting views exist regarding, the area of a glacier where surge is initiated. These differences can most likely be attributed to different observation periods and methods. According to Sund et al. (2009), the surges of both terrestrially terminating and tidewater glaciers are initiated in the upper parts of the glacier. The surges of tidewater glaciers in Svalbard have, according to Murray et al. (2003) however, been observed to initiate in the lower part of the glacier. This would imply that there is a difference in surge initiation between land terminating and tidewater glaciers. Murray et al. (2003) base their conclusions on observations of the tidewater glacier, Monacobreen, in Northern Spitsbergen, where the surge seems to have started simultaneously over the lower parts of the glacier and then propagated up glacier. Sund et al. (2009) however, points out that the area of surge initiation can be misjudged and that surge initiation in the upper part of a glacier can be overlooked due to, for example, the observation period. According to Sund et al. (2009) there exists no evidence for surge initiation at the terminus of tidewater glaciers. There should therefore be no difference between the surge initiation of land terminating and tidewater glacier.

### 3.5. Sedimentological evidence of surge events

It is difficult to directly observe the sedimentological processes that take place beneath contemporary glaciers and ice sheets. Because of this, studies of the sea floor in front of tidewater glaciers play an important role for understanding processes at the glacier bed. Landforms are preserved on the ocean floor without significant modification, since they are not subjected to subaerial erosion (Ottesen et al. 2008). Subaerial processes, such as erosion and periglacial activity, do not operate on the ocean floor (Ottesen and Dowdeswell 2006). Seafloor sediments can however be reworked by iceberg-keel ploughing, currents and mass wasting on the seafloor (Ottesen and Dowdeswell, 2006).

The characteristic sea floor morphology of surging tidewater glaciers commonly displays a suite of distinct glaciogenic structures. These structures include terminal moraines, which record the maximum extent of the latest surge and are formed through ice push and reworking of material in the front of the advancing glacier (Ottesen et al. 2008). Much of the materials building up the moraines, are suspended sediments, brought by melt water, to the glacier front during the surge event. Slumping is, according to Solheim (1991), a common process at the slope of the terminal moraines and occurs mainly at the distal slope. Slumping can be defined as a downward intermittent movement of rock debris. It usually occurs due to the removal of buttressing soil at the foot of a slope or as a gravity flow, i.e. a fluid
flowing due to the forces of gravity alone. According to Ottesen and Dowdeswell (2006) other characteristic landform assemblages are large streamlined, glacial lineations, lobe shaped debris flows, and transverse annual retreat moraines. The flow-parallel lineations are characteristic features beneath surge type glaciers. According to Ottesen et al. (2008) eskers, long ridges deposited by subglacial or englacial streams, can also be found under tidewater glaciers, but in bathymetric data from glaciers, in northern Spitsbergen eskers have been rare or absent.

\[\text{Figure 6: Landform assemblage model for surging tidewater glaciers in Svalbard, based on swath bathymetry from Van Keulenfjorden and Rindersbukta, Spitsbergen (Ottesen et al. 2008).}\]

The configuration and retreat history of tidewater glaciers in Svalbard, as well as the associated sedimentary processes, such as slumping, gravity flow and mass wasting have been established based on marine geological and geophysical data. These surveyed glaciers include Paulabreen in van Mijenfjorden, Nathorstbreen in van Keulenfjorden and Nansenbreen, Borebreen and Wahlenbergbreen in Isfjorden, (Ottesen et al. 2008, Ottesen and Dowdeswell 2006). From these data Ottesen et al. (2008) have constructed a conceptual land system model, which is presenting the characteristic landforms created by surge glaciers (Fig 6). They have concluded that the bathymetric studies from the fjords in Spitsbergen (see figure 7 for the location of the glaciers) show many similarities regarding landform assemblages. Similar crosscutting relations, i.e. glacial lineations crosscut by moraines, have also been observed at the different study sites (Ottesen et al. 2008).

The landform model shows that landforms are most likely produced in a characteristic time sequence (Ottesen et al. 2008). The glacial lineations form during fast ice flow, when the glacier surges, whilst rhombohedral crevasse fill ridges form
just after the surge (Solheim 1991). The terminal moraine records the maximum extent of the glacier front and indicates a longer still-stand of the glacier. The outer extent of the terminal moraine is often characterized by lobe shaped debris flow deposits. Transverse annual retreat moraines form after the surge has stagnated and the glacier front retreats back into the fjord (Ottesen et al. 2008). The landform model further indicates that the surge zone contrasts strongly from the area outside the surge moraines, which is characterized by normal marine processes and iceberg plough marks (Solheim 1991).

According to Solheim (1991), the rhombohedral ridges are particularly characteristic of surging glaciers. Ottesen et al. (2008) suggests that the rhombohedral ridges form through squeeze-up of debris into basal crevasses during the early stages of ice stagnation after the last surge event. These basal crevasses form in grounded glaciers if the water pressure is nearly equal to the ice overburden pressure. High water pressures are related to fast flow and have been observed, in ice streams and surging glaciers (Kristensen et al. 2009). When sediments are thrusted into basal crevasses and the crevasse fill ridges form, high water pressures are required (Solheim 1991). The rhombohedral ridges are therefore characteristic of surge events and fast glacier flow (Kristensen et al. 2009).
4. Study area

Tunabreen is an outlet glacier of the small Lomonosovfonna ice cap, located in Central Western Spitsbergen. Tunabreen calves into the around 14 km long and up to 5 km wide Tempelfjorden, which is located at the Eastern part of Isfjorden (Fig. 8). Tempelfjorden has a surface area of nearly 57 km² (Forwick et al. 2010). The maximum water depth of Tempelfjorden is around 110 m in the central and outer parts of the fjord. At the head of the fjord Tunabreen terminates with a tidewater ice cliff and coalesces with the larger land terminating glacier, von Postbreen. According to Forwick et al. (2010) Tempelfjorden can be divided into two basins, which are a larger outer main basin and a smaller glacier-proximal basin. The basins are separated by a moraine ridge that was deposited by von Postbreen during the 1870 surge. The maximum water depth of the glacier proximal basin is around 70 m.

During the deglaciation of Isfjorden, the fronts of Tunabreen and von Postbreen retreated stepwise into the fjord (Forwick et al. 2010). During the Younger Dryas, the tributary fjords were still covered by glacier ice. According to Forwick and Vorren (2009), it is probable that the glaciers around Isfjorden were small during early Holocene. Svendssen et al. (1996) suggests that there were probably no tidewater glaciers in Isfjorden, during early and mid-Holocene. A regrowth of the glaciers calving into Tempelfjorden occurred first during late Holocene and culminated in the LIA advance, when von Postbreen reached its maximum Holocene extent during AD 1870. During the end of the LIA period, in the 1920s, air temperatures in Svalbard
rose, by almost 5°C in a short time period. In general the glacier mass balance has been negative since the end of the LIA (Dowdeswell et al. 1995). According to Yoshioka et al. (2006) Tunabreen has been experiencing a general decrease in mass since the 1930s. The terminus of the glacier has retreated and the thickness of the terminal area has decreased.

The drainage basins of Tunabreen, von Postbreen and the nearby river Sassenelva are the major sedimentary sources in the area draining into Tempelfjorden. The influence of Tunabreen and von Postbreen, on the sedimentary environment of Tempel- and Sassenfjorden has been varying in time, which confirms that the glaciers have been fluctuating during Holocene (Forwick et al. 2010). The bedrock geology of the Tunabreen drainage basin area comprises mainly of Late Palaeozoic sandstones and shales. The northern basin is dominated by carbonates and evaporitic rocks. Clastic sedimentary rocks, silicified carbonates and dolerites can also be found in the area (Dallmann et al. 2002). In Svalbard many of the tidewater glaciers flow over sedimentary beds, which contain fine grained glacimarine fjord sediments, which are conductive to surging. The glaciers are commonly fed from large interior drainage basins and have high velocities due to a high mass flux (Ottesen et al. 2008).

The variation in sea ice extent is according to Forwick et al. (2010) an important factor contributing to the climatic sensitiveness of the area. This is mainly due to the large inter-annual variations in the location of the sea ice edge and variations in the timing of melt and sea ice break up. The sea ice in the western Spitsbergen fjords usually forms during November and persists until April to July (Forwick et al. 2010), but some years, fjords can be completely sea ice free. For example in the warm winter of 2012 most of Tempelfjorden was sea ice free (personal observation).

In historical times glacier surges on Tunabreen have been recorded to occur in the 1930s, 1970s and between 2002 – 2005 (Forwick et al. 2010). The adjacent von Postbreen experienced a surge in 1870, whilst the smaller Bogerbreen surged in 1980 (Hagen et al. 1993). According to Yoshioka et al. (2006) the surface of Tunabreen has been observed to experience a general lowering since the 1930s. Since 1933 the surface has been observed to decrease in height from 120 m. a.s.l. to 90 m a.s.l. in 1993, with an increase of 5 m during the surge in 2004. During the most recent surge, with maximum reached in 2004, the calving front of Tunabreen advanced with around 1.4 km. The longest recent advance occurred during the 1930s surge, when the glacier front advanced with around 3 km (Plassen at al. 2004). The glacier terminus did however not, during any surge event, advance further out than during the previous surge (Yoshioka et al. 2006). See figure 13 for the frontal position of Tunabreen during different years.
5. Methods

5.1. Mapping the seabed

This study is based on analysis of multibeam data and analysis of satellite and aerial images. The main data in the study consists of multibeam swath bathymetric data from the Norwegian Hydrographic Service, which is a part of the Norwegian Mapping Authority, and from UNIS. The raw swath bathymetric data have been processed and gridded prior to this study at the University Centre in Svalbard (UNIS).

The multibeam echosounder technique provides an opportunity to study landforms which are located below the sea surface. The Kongsberg EM-3002 multibeam echosounder system was used for acoustic mapping of the seabed depth and surface character. It emits several individual beams, resulting in high-resolution images of the seafloor landforms and surface sediment variations. The bathymetric data used in the study consists of multibeam data from the inner part of Tempelfjorden (see figure 9). The dataset covers the innermost c. 5 km of the fjord to the present margin of Tunabreen. Nearly the whole area in the inner part of Tempelfjorden has therefore been mapped with the multibeam echosounder, except for the areas nearest to the calving glacier front, and the shallowest areas on both sides of the fjord.

The processed multibeam bathymetric data were imported into the ArcGIS program suite, provided by ESRI. Mapping and interpretation of landforms has been conducted in the ArcMap software, which allows change of factors such as color and hillshade, to get a clearer view of the submarine landforms.

Landform analysis was conducted from both color shaded and black and white bathymetric images. A large scale color shaded bathymetric image reveals principally the largest landforms and the highest relief. Landforms which span a large range of water depths are therefore difficult to visualize in the color shaded bathymetric image and can be better visualized if the color scale of the bathymetric image is tuned to the geomorphological features of interest. Using a grey scale image however gives a better overall representation of the landforms and structures on the seabed. Because of this the interpretation of landforms was mainly done from a black and white bathymetric image, since this gives a clearer contrast between depth differences on the seafloor.

Most of the landforms mapped in this study are only a couple of meters in height, which requires high precision, during the mapping procedure. The small size of the landforms therefore contributes to an error margin in the mapping process. The large depth difference in the fjord makes it difficult to visualize the landforms from the color shaded bathymetric data, since the height of them is very small in comparison with the depth of the fjord. Some areas of the multibeam data also display a chaotic structure, with no clear landform assemblages. To assess the height differences on
the seafloor and the heights of different landforms, profiles of the seabed can be taken in ArcMap. This can be done by creating a 2D line that is mapped as a profile graph, see figure 11 in the results section. The 2D graphs can provide a clearer view of the depth differences on the seafloor and allow for the assessment of landform height and width. Profile graphs also make it possible to better assess the distance between crests of glacial lineations and moraine ridges. It is further possible to create a 3D image of the full fjord bathymetry with the ArcScene software, to get an overview of the seabed morphology of the fjord.

5.2. Data sets used for mapping the fluctuations of the glacier terminus

5.2.1. Terminal positions mapped during earlier studies

The position of the glacier terminus in Tempelfjorden has been evaluated from the 1870 surge event until the year of 2011. The frontal position of the glacier has been mapped from different data sets. The 1870, 1830 and 1966 positions have been mapped by reference from earlier existing data (De Geer 1910, Plassen et al. 2004, Yoshioka et al. 2006). The position of the glacier front during the 1930s surge has originally been mapped by Yoshioka et al. (2006) from existing photographs taken from boats visiting the front of Tunabreen in 1930. The terminal position in 1966 has been revised from an earlier study conducted by Plassen et al. (2004). The earlier studies were based on photographs, maps issued by the Norsk Polarinstittutt, as well as aerial photographs from Norsk Polarinstittutt (Yoshioka et al. 2006, Plassen et al. (2004). The mapped maximum position of the 1870 surge was derived from (De Geer 1910).

5.2.2. ASTER satellite images

The more recent terminal positions of Tunabreen have mainly been evaluated by the use of ASTER (Advanced Spaceborne Thermal Emission and Reflection Radiometer) satellite data. The ASTER images provide good, high resolution data of the glacier terminus. The images have a higher resolution than the Landsat data used in the study, and provide a better option for mapping the glacier front. ASTER images of the study area were however only available for specific years during the 2000s, which is why they were not used solely.

The ASTER images were used to map the terminal position in the years 2002, 2003, 2004, 2005 and 2010. The images were provided from the US Geological Survey (see Table 1 for data source) and were obtained as TIFF files. The ASTER image files were added into the GIS ArcMap software, where the glacier terminal positions were mapped. The ASTER data was obtained during the summer months. The 2002 and 2005 images were taken in June, the 2003 and 2004 in July and the 2010 image was taken in the beginning of August. The ASTER data was provided with the WGS 84,
UTM zone 33 N reference coordinate system and were matched without difficulties with the multibeam data.

5.2.3. Landsat satellite images

Landsat images were used for evaluating the terminal position during the 1970’s and 1980’s. The terminal positions of Tunabreen in the years 1975 and 1988, in figure 13, have been mapped from Landsat satellite images, provided by the US Geological Survey (Table, 1). The 1975 Landsat image was obtained during the winter period in the month of April and the 1988 image was obtained in June. The three first bands of the images were inserted into GIS and merged to create the satellite image. Landsat images for several years during the 1970s and 1980s have been studied, but most of the data was of too poor quality to be used for more than a visual estimation of the terminal fluctuations. Some of the Landsat images, which were studied, were not of high quality. The images contained shadows, sea ice and some were simply blurry. However the glacier front could be mapped for the years 1975 and 1988, despite of sea ice and shadows. The poor quality of the data most likely leads to an error of up to a couple of meters in the mapping process.

Landsat data did not exist for all years during the 1970s and 1980s. Landsat data for the 1971 maximum surge extent, mapped in Plassen et al. (2004) was not obtained from satellite images during this study. The mapped 1971 position in Plassen et al. (2004) however corresponds well with the earliest Landsat data obtained for the area, from the year 1975. Newer Landsat data from 2000 was also used, together with the ASTER data. The 2011 terminal position in figure 14 was for example mapped from Landsat data, obtained in April 2011. Landsat data obtained in July was also used for evaluating the terminal position in the year 2006. A Landsat image from 2002 was used as background in data presentation (Fig. 9 – 14).

The main difficulty with the older Landsat data was that it did not completely match the multibeam data. The multibeam and ASTER data were both provided with the WGS 84, UTM zone 33 N, geodetic system. The Landsat data that was obtained before 1984 was not georeferenced to the WGS 84 system. The Landsat data had therefore been converted to WGS 84, UTM zone 33N, but an offset appeared when the data was converted. This resulted in an offset of a couple of meters to kilometers, between the Landsat images and the multibeam data in the GIS. Each image had a different offset, where some nearly matched the multibeam data. The image from 1975, which was used for mapping the terminal position, was therefore manually rectified to be comparable with the multibeam data.

5.2.4. Aerial photograph

The 2004 surge extent was also evaluated from an aerial photograph, which provided a good control of the terminal position. The aerial image from 2004 has together with
satellite data, been used to map the terrestrial glacial landforms adjacent to Tempelfjorden. These landforms are the lateral moraines on both sides of the fjord. The aerial image further show structures on the surface of the glacier, providing a good overview of a highly crevassed surface at the front of Tunabreen, during the 2004 surge. The aerial image has been taken in the end of July 2004, from a height of 3200 m a.s.l. The aerial photograph has been provided by NERC (Natural Environmental Research Council), UK.

5.2.5. Annual changes of the terminus from ASAR data

To evaluate the annual changes of the terminus, the glacier frontal fluctuations were mapped annually between 2006 – 2010 at the beginning of June and October. These two months were chosen to evaluate how large a difference in terminal position took place during the summer ablation season and if this in some way would provide an explanation for the genesis of the recessional moraines on the seafloor.

ASAR (Advanced Synthetic Aperture Radar) images in Wide-Swath-Mode (WSM) were used to map the annual changes of the terminus. The ASAR images where taken by the European Space Agency (ESA) Envisat satellite. The data were provided by Dough Benn at UNIS. The front of the glacier was mapped in ArcGIS. The terminal position was mapped separately for the months of October and June. The mapped terminal positions at the end of the winter season and at the end of the summer season were then compared.

The data set consists of images taken since 2006, with almost monthly intervals. During 2006 images were however not available for the summer months, which is why the position of the terminus was only mapped during October in 2006. The months of June and October were partly chosen because they show the terminal position just in the beginning of the summer ablation season and in autumn, at the end of the ablation season. The main reason why these months were chosen was because the data quality exceeded the quality of the images taken during most other months. The ASAR images taken in August were for example more blurry than the October images. The August 2008 image was unreadable, which would have made it difficult to estimate the exact position of the glacier front during this year.

The images taken in October and June were however neither of high quality. The satellite images did not have a high spatial resolution, but were considered useful for mapping the frontal position of a tidewater glacier since the spatial resolution was considered to be good enough for this purpose. The resolution and blurriness of the images however provides an error margin of at least a couple of meters in the mapping procedure.

The ASAR images were provided as TIFF raster files and matching well with each other, but without any georeferencing. Because of this they could not be merged into ArcMap together with the multibeam data. Due to this the images were matched
Dynamics of surging tidewater glaciers in Tempelfjorden, Spitsbergen with the multibeam and other satellite data by manual matching. To get a better view of annual terminal changes a longer time series data set would however be useful. Table 1, provides a summary of the different data types and data sources used for mapping of the terminal positions of Tunabreen.

Table 1: Data types/sources used for mapping the terminal position of Tunabreen.

<table>
<thead>
<tr>
<th>Year</th>
<th>Surge</th>
<th>Data type</th>
<th>Data source</th>
</tr>
</thead>
<tbody>
<tr>
<td>1870</td>
<td>x</td>
<td>Lit.</td>
<td>De Geer 1910</td>
</tr>
<tr>
<td>1930</td>
<td>x</td>
<td>Lit.</td>
<td>Yoshioka et al. 2006</td>
</tr>
<tr>
<td>1966</td>
<td></td>
<td>Lit.</td>
<td>Plassen et al. 2004</td>
</tr>
<tr>
<td>2002</td>
<td></td>
<td>Landsat</td>
<td>NASA Landsat Program, 2002, Landsat TM, scene LE72130042002164AGS00, USGS, Sioux Falls, 04/16/2012</td>
</tr>
<tr>
<td>2007</td>
<td></td>
<td>ASAR</td>
<td>ASAR Wide Swath Mode (ENVISAT.ASA.WSM__1P). European Space Agency</td>
</tr>
<tr>
<td>2008</td>
<td></td>
<td>ASAR</td>
<td>ASAR Wide Swath Mode (ENVISAT.ASA.WSM__1P). European Space Agency</td>
</tr>
<tr>
<td>2009</td>
<td></td>
<td>ASAR</td>
<td>ASAR Wide Swath Mode (ENVISAT.ASA.WSM__1P). European Space Agency</td>
</tr>
</tbody>
</table>
6. Results

The exact position and extent of the subsea landforms in the innermost part of Tempelfjorden have not been mapped prior to this study. The more distal subsea landforms have however been studied earlier (Plassen et al. 2004, Forwick et al. 2010). The landforms that have been described during earlier research are the landforms distal to the 1870 von Postbreen terminal extent. These are the debris flow lobes, the thrust moraines and the von Postbreen terminal moraine.

Plassen et al. (2004) used acoustic data (seismic and side-scan sonar) to study seafloor structures in Tempelfjorden. From these data they described the von Postbreen terminal moraine, and the debris flow lobe located distal to the moraine. The exact position and full extent of the landforms were however not mapped. Forwick et al. (2010) used multibeam data to study the seafloor morphology in Tempelfjorden. The landforms described during the study were the two debris flow lobes and the thrust moraines. The exact position and full extent of the thrust moraines and debris flow lobes were however not mapped. The focus of the study was on the landforms located in the outer part of Tempelfjorden. Following sections give a description of the submarine landform in Tempelfjorden, and the recent fluctuations of the glacier margin.

6.1. General setting

The fronts of Tunabreen and the adjacent von Postbreen have had several advance and retreat episodes during the last hundred years. Four recent surges have been recorded in Tempelfjorden, which distinguishes the study area from earlier studied Svalbard tidewater surge glacier settings, where the glaciers have been known to surge only once or twice (Ottesen and Dowdeswell 2006, Ottesen et al. 2008). The multiple surge events have resulted in a complex pattern of landforms in the fjord and adjacent coastal areas.

Swath bathymetric data from Tempelfjorden provides a good overview of the seafloor morphological features. Figure 9 shows the bathymetry of tempelfjorden superimposed on a Landsat satellite image. The inner part of Tempelfjorden covered by this study is a deep through with large differences in seafloor depth, ranging from just one meter below sea level at the sides of the fjord, to almost 90 m of depth in the deepest parts of the basin. The colour scheme, in figure 9, showing the seafloor depth below sea level has been adapted to show more of the height differences in the shallower parts of the fjord.

The multibeam bathymetric images can be seen as a digital elevation model, where a complex pattern of submarine elevations and small basins can be observed. The most
Dynamics of surging tidewater glaciers in Tempelfjorden, Spitsbergen

recent surge events, which are the focus of this study have however only been affecting the seafloor in the innermost parts of Tempelfjorden. A zoomed in picture of Tempelfjorden can be seen in figure 10. In the zoomed in picture the most recently formed moraines can be more clearly recognised. The moraines are only a few meters in height and are clearly reaching from one side of the fjord to the other. The moraines are superimposed on distinct glacial lineations in the innermost part of the fjord.

Figure 9: Swath bathymetry image showing the seafloor structures in the innerpart of Tempelfjorden. The delineated area shows the main study area.

The depth differences of the seafloor in the inner part of Tempelfjorden can be noted from figure 10, where the darker blue parts represent glacially scoured depressions which are around 60 m deep. The yellow areas lie on 20 – 30 m depth and the red areas are only around 10 m deep. The profile lines A – B and C – D, in figure 10 are represented as graphs in figure 11. The graphs clearly show the depth differences along the drawn profiles. In both profiles the deepest parts lie around 45 – 50 m below sea level, while the shallowest parts of the seafloor are just 20 to 35 m below sea level. The moraines in the innermost part of Tempelfjorden can be seen as small, around 2 to 4 m high peaks in the graphs (see arrows in Fig. 11). The 2004 surge moraine is however between 6 – 8 m high and is represented by the largest crest in the profiles.
6.2. Submarine landforms in Tempelfjorden

The landforms found at the inner fjord are most likely related to the more recent surges, but the morphological record is not apparent and older landforms or partly eroded older landforms might be present. The landforms related to the recent 2004 surge are well preserved and will be described at the end of this section.
Figure 12 provides a summary of the landforms found in Tempelfjorden. The left hand side panel in figure 12 is a conceptual figure, showing a spatial landform reconstruction, while the right hand side figure shows the multibeam data superimposed on a Landsat image from 2002. The submarine moraine found in the location of the 1930s surge extent has been mapped with a dashed line and question mark, since the total extent and position of the ridge has not been possible to map from the bathymetric data. Small ridges on both sides of the fjord however suggest that this is a moraine ridge related to the 1930s surge. The position of the moraine coalesces well with the maximum position of Tunabreen during the 1930s surge, as revised from Yoshioka et al. (2006). The mapped landforms in Tempelfjorden have however not been dated, and only a relative chronology has been established, with younger landforms on top of older ones.

![Figure 12: Landform assemblage in the inner part of Tempelfjorden. The surge moraines have been marked with arrows. See Fig, 9 for the main study area.](image)

### 6.2.1. Terminal moraine system- von Postbreen surge 1870

The terminal moraine recording the maximum Little Ice Age extent in Tempelfjorden is located around 4 to 4.5 km from the present glacier front (Fig. 12). This terminal position corresponds with the von Postbreen surge in 1870 as mapped by De Geer (1910). The terminal moraine has been deposited transverse to the fjord axis and has a height of around 5 to 10 m and a width of around 150 to 200 m. According to Plassen et al. (2004) the terminal moraine has a thickness of up to 70 m. The moraine is around 3.4 km long and extends perpendicularly through the fjord. On the proximal side of the terminal moraine glacial lineations occur whereas a debris flow has been deposited on the distal slope. According to Plassen et al. (2004) three
more moraine ridges are deposited proximal to the terminal moraine. Their distribution is however not continuous throughout the fjord.

6.2.2. Subaerial moraines

The proximal side of the clearly defined debris flow lobe corresponds well with the 1870 subsea terminal moraine. The terminal moraine on the other hand corresponds well with the subaerial moraines seen on both sides of the fjord (Fig. 12). The subaerial lateral moraine on the southern edge of the fjord corresponds especially well with the maximum position of the 1870 surge, as suggested by Plassen et al. (2004). The lateral moraine at the northern margin of the fjord also shows a correlation to the 1870 surge event. The landform system, with the subsea terminal moraine and the subaerial lateral terminal moraines records the maximum LIA extent of the von Postbreen.

6.2.3. Debris flow lobe

A large lobe-shaped feature can be observed in the center of the swath bathymetric image. This lobe is clearly defined in the subsea data and has a slope towards the opening of the fjord. The lobe is located beyond the transverse terminal moraine ridge related to the 1870 surge of von Postbreen. The depth difference between the upper parts of the lobe, which is at around 30 m below sea level to the front of the lobe at 80 m below sea level, is 50 m.

The lobe-shaped feature is most likely a debris flow lobe (Plassen et al. 2004). The debris flow has been deposited beyond the outermost ridge associated with the surge of von Postbreen in 1870. According to Ottesen and Dowdeswell (2006) these types of debris flow lobes are most likely made up from debris which has been extruded from the tidewater glacier terminus during a surge event. The timing of the debris flow is however difficult to assess, since it could also have been deposited after the surge, when the buttressing effect of the glacier front disappeared or even later.

According to Forwick et al. (2010) two superimposed debris lobes exist in Tempelfjorden. Beneath the clearly defined debris flow lobe occurs another lobate feature, which is not as clearly defined as the 1870 von Postbreen lobe. The second lobe extends further to the west but the morphology of the lobe is less clearly outlined. Because of this, the lobe has been mapped with dashed line in figure 12. The presence of the debris flow lobe, extending downslope from the terminal moraine ridge, fits well into the landform assemblage model by Ottesen et al. (2008). The debris flow lobe is a characteristic landform feature in a surging glacier setting (see section 3.5). The two lobate features most likely represent different events, where the second, less clearly defined lobe is older than the lobe initiating from the
1870 terminal moraine. The ridges on top of the second lobe confirm that this is not an occurrence of multiple mass wasting events related to the von Postbreen surge.

6.2.4. Thrust moraine

Several curved and irregular ridges occur on top of the second lobe (Fig. 12). The ridges display an irregular pattern and do not clearly extend from one side of the fjord to the other, in contrast to the more glacier proximal ridges. According to Forwick et al. (2010) sparker, i.e. a form of acoustic sound data reveals not several ridges, but one multicrested ridge, which is around 1.5 km wide, up to 3 km long. This ridge is up to 13 m high and has been interpreted as a thrust moraine (Forwick et al. 2010). It is a large morphological feature, but it’s structure is not clearly defined from the bathymetric data. The proximal crest seems to be a sharp feature, almost 10 m high, but the more distal crests are less high and sloping towards the mouth of the fjord.

The ridge has been interpreted as a thrust moraine due to the difference of the ridge in comparison with the recessional moraines found in the area. The thrust moraine is a larger feature, with greater thickness than recessional moraines and the crests are located with less spacing between one another (Forwick et al. 2010). The moraines on top of the second lobe have however not been thoroughly studied. The evidences presented in Forwick et al. (2010) leading to their interpretation as thrust moraines are rather weak and they could as well be recessional moraines common to this setting.

6.2.5. Glacial lineations

Streamlined linear landforms exist in several places on the seabed of Tempelfjorden. They can be found just in front of the present glacier terminus, and as clusters up flow from the debris flow lobe (Fig. 12). Streamlined glacial lineations also occur further out in Tempelfjorden according to Forwick et al. (2010) and are related to events taking place before the late Holocene maximum, i.e. LIA. Streamlined glacial lineations are a product of fast flowing ice (Ottesen and Dowdeswell 2006). This suggests that Tempelfjorden has been modified by fast glacier flow during multiple time episodes.

The linear features in the inner part of Tempelfjorden occur in two sets (Fig. 12). The distinct more innermost lineations are most likely related to the 2004 surge event, whilst the outer ones are more chaotically placed and not so clearly defined in the geomorphological record. The linear features are oriented parallel to the fjord axis. They are on average around 500 m long, with some of the innermost ones being more than a kilometer long. The relatively limited lateral extent of the linear features in the outer part of the study area could be explained by the history of Tunabreen, where several glacier advance and retreat events have erased, or partly destroyed.
and modified older landforms. The more coherent set of linear features in the central part of the bathymetric data coverage (Fig. 12) seems to be related to the 1930 surge event, but this cannot be supposed with certainty, since no dating of landforms has been carried out.

The glacial lineations in Tempelfjorden are related to surge events. They have formed subglacially during ice advance and were formed parallel to the ice flow. The streamlined bedforms can be created on a variety of scales (Ottesen and Dowdeswell 2006), which can be seen in the geomorphological record in Tempelfjorden. The height of the lineations is around 10 m, even though there are also ice flow parallel features of merely, 3 – 5 m high. The distance between lineations is fairly uniform, being around 150 to 300 m between each landform. These ice flow parallel streamlined landforms can be interpreted as glacial lineations.

6.2.6. Annual retreat moraines

Behind the 2004 surge terminal moraine occur a number of relatively closely spaced transverse moraine ridges. The ridges are located perpendicular to the fjord axis and subparallel to each other. The ridges follow the bathymetric contours and bend towards the wall of the fjord at the fjord margins. They are superimposed on the streamlined glacial lineations at the base of the fjord. This suggests that the ridges are the youngest landforms in Tempelfjorden.

The ridges are clearly visible in the bathymetric data and form crests spanning through the floor of the fjord. These ridges can be interpreted as annual retreat moraines. The interpretation is based on the existence of a large number of ridges and due to the relatively regular spacing between these ridges. This interpretation is consistent with other studies of recessional moraines in Svalbard (Ottesen et al. 2008). The character of these ridges also fits into the description of annually created recessional push moraines presented by Ottesen and Dowdeswell (2006), who point out that the abundance of ridges, the regular spacing between ridges, and the cross cutting of glacial lineations are characteristic features of annual retreat moraines.

The average spacing between the ridges in the inner part of Tempelfjorden is around 300 m. The height of the ridges is around 2 – 4 m, except for the 2004 moraine which is a larger feature. This gives the ridges the same proportions as earlier studied annual retreat moraines in Svalbard (Ottesen and Dowdeswell 2006). The 2004 moraine is between 6 – 8 m high. This distinguishes the 2004 moraine from the retreat moraines. This suggests that the larger moraine records the maximum surge extent, while the smaller ridges were formed during semiannual glacial retreat.

The moraines have most likely been deposited during halts or minor re-advances of the glacier terminus, during a period of general glacier retreat. Earlier these types of ridges have been interpreted as series of annual push ridge moraines. The push ridges form each winter, when the terminus of the tidewater glacier experiences a
minor re-advance during the winter period, even though the glacier undergoes a general retreat face (Ottesen and Dowdeswell 2006).

6.3. Recent fluctuations of the glacier front, comparison between marine evidence and remote sensing data.

The terminal fluctuations of Tunabreen and von Postbreen can be followed in maps, satellite, and aerial images. The mapped terminal positions in figure 13 confirm that both of the glaciers terminating into Tempelfjorden have been advancing and retreating in a complex pattern since the LIA. The LIA glacier extent is represented by the lateral subaerial moraines and can also be observed in the subsea data, where it is represented by the 1870 von Postbreen terminal surge moraine (Fig. 12). After the LIA advance von Postbreen retreated out of the fjord (Plassen et al. 2004), whilst Tunabreen experienced three subsequent surges during the following century.

During the 1930s Tunabreen experienced a surge. The extent of the 1930s surge is not marked by a single distinct terminal moraine in the subsea record (Fig. 12). However, a broader and more complex moraine belt can partly be observed. A group of glacial lineations can also be observed just up flow of the terminal surge extent. The lineations and the entire transverse ridge-through complex just east of the 1930 line on the map in figure 13 could represent a ‘marginal zone’ of the 1930 surge maximum. Without any absolute dating it is however difficult to evaluate the time of landform formation and therefore the lineations could also have formed during earlier surge events.

After 1930, Tunabreen retreated and reached almost as far back as its present extent. In 1966 the front was located near the 2002 terminal position. During the 1970s the glacier surged again, reaching its maximum extent in 1971 (Plassen et al. 2004). This surge extent has been recorded in the subsea data by a distinct terminal moraine, which corresponds well to the mapped 1975 extent of the glacier front (Fig. 13). The terminal position of the 1971 surge also corresponds well with the 1975 position (Fig. 13; Plassen et al. 2004), which implies that there were no large changes in the location of the terminus during the immediate years following 1971. No glacial lineations related to the 1970s surge can be clearly detected in the subsea data. The sea floor between the terminal 1970s moraine and the 2004 surge extent displays a chaotic morphology, with large depth differences due partly to rugged bedrock relief. After the 1970s surge, Tunabreen retreated again.
The 2004 surge episode is the most clearly recorded event in the subsea data. Evidence of this surge can be seen as a clear cluster of ice flow-parallel glacial lineations, and terminal moraines. The 2004 surge created a surge landform system, with similarities to the landform assemblage model in Ottesen et al (2008).

Graph 1 provides a simplified view of the terminal fluctuations of Tunabreen. The distance from the reference point, which is chosen as the position of the 2002 moraine, has been measured along a reference measurement line and is displayed on the x-axis. The graph in panel A shows the terminal fluctuations from the 1870 terminal position to today. The panel B shows the more recent fluctuations of the terminus. From the panel B it is clear that the surge of the glacier margin was very rapid, whereas the retreat of the terminus was more gradual, with the terminus retreating on average 100 to 150 m/y.
The recent position of the glacier terminus, mapped from satellite images can be seen in figures 14 and 15. In figure 14 the annual extent of the frontal position of Tunabreen has been mapped from 2002 to 2011, and is presented in relation with the submarine moraines in Tempelfjorden. In 2002 the terminus is located far back in the fjord. It advances nearly 2 km during the 2004 surge event. During the years following 2004 the glacier front starts retreating, with a rate of around 150 m annually. The overall retreat rate is variable being one of the highest (c. 200 m) in 2009.

Figure 14 suggest that there is a spatial relation between the position of the glacier front and the position of the submarine moraines. The 2004 glacier terminal extent corresponds well with one of the most distinct moraines. In 2005, 2007, 2008 and 2009 the terminal position also seems to correlate well with submarine moraines. This confirms that there is a correlation between the frontal position of the glacier and the position of the moraine ridges. The close relation between glacier terminus and moraine ridge position would confirm Ottesen and Dowdeswell (2006) hypothesis that the ridges are annual push moraines created during a still stand or minor winter re-advances of the glacier terminus.
Figure 14: Ice terminal position, shown for the years 2002 – 2011 (black lines), in relation with the submarine moraines (red lines) in the inner part of Tempelfjorden. The terminal positions have been mapped from ASTER/Landsat/ASAR images.

Figure 15 shows the position of the glacier front in June, at the beginning of the summer melt season and in October, at the end of the summer melt season. The annual frontal positions have been mapped for two months, June and October from 2006 to 2010. For June 2006 ASAR data was not available. The terminal position in October is represented by red and orange colors, whilst the June terminal position is represented by blue and green colors. The mapped terminal positions show that the glacier retreat takes place during the short summer season. It can further be observed that the glacier front has been retreating around 50 to 100 m every summer season during the recent years. The winter re-advances have not been considerable. However it is not straightforward to relate the exact position of the glacier terminus with the subsea morphology because the satellite images only give a picture of the situation at the sea level.
Dynamics of surging tidewater glaciers in Tempelfjorden, Spitsbergen

Figure 15: Glacier terminal position shown for June and October, during the years 2006 – 2010. For the year 2006 only the frontal position in October has been mapped, since no ASAR data existed for June 2006.

The mapped frontal positions demonstrate a complex picture of glacier dynamics. After each of the three surges during the last decennia the front of Tunabreen has retreated further back into the fjord, whilst every subsequent advance has been less extensive than the previous one. The fluctuations of the terminus have generated a complex assemblage of subsea landforms. Most of the surge type glaciers in Svalbard, investigated using subsea data are known to have surged only once or twice (Ottesen and Dowdeswell 2006, Ottesen et al. 2008). This suggests that Tunabreen in Tempelfjorden is one of the most frequently surging glaciers with four recorded, recent surge events. The multiple glacier advance and retreat episodes in Tempelfjorden have resulted in a complex landform assemblage, where landforms of previous surges have been erased or modified by the subsequent surge events. The most distinct data set in the bathymetric record therefore consist of the landforms from the most recent, 2004 surge. Since every subsequent surge has not reached the maximum extent of the previous one, a complex picture of younger surge events superimposed on older ones has been preserved in Tempelfjorden.
7. Discussion

7.1. The submarine morphology of Tempelfjorden

Marine geophysical data is an important source of information in glacial geomorphology, yielding high-resolution images of sea floor landforms. The submarine landforms in Tempelfjorden are well preserved because they have not been exposed to post-depositional erosional processes as their terrestrial counterparts.

Many submarine landforms are clearly visible and easily mapped from the multibeam data, such as, for instance, the debris flow lobe distal to the 1870 terminal moraine. Mapping landforms from multibeam bathymetric data is however not always straightforward. The interpretation of the distinguished morphologies is one of the key aspects of mapping and could be a significant error source. Also the spatial extent of some landforms is difficult to map exactly. This applies for example to the less distinct debris flow lobe, which has been mapped with a dashed line due to the somewhat uncertain spatial extent of the lobe. The 1930s surge moraine is another example of a difficult to map landform. The submarine ridges at both sides of the fjord suggest that they are a part of the 1930s terminal moraine. The exact position and spatial extent of the moraine is however not clear from the multibeam data. The position of the moraine has therefore been mapped with a dashed line. This makes the interpretation skills of the mapper the most important factor for determining the spatial extent of the landform and also the most important error factor of the mapping process.

The subsea morphology in Tempelfjorden is complex in comparison with previously studied surging tidewater glacier settings in Svalbard. The landform assemblages found in Tempelfjorden are not completely consistent with earlier mapped, surge type, tidewater landform assemblages in Svalbard (Ottesen and Dowdeswell 2006, Ottesen et al. 2008. This can be explained by the fact that the innermost part of Tempelfjorden has been affected by at least three surges since the end of the LIA. These subsequent surges have erased and reworked the landform assemblages created by the earlier surge events. This has resulted in complex seabed morphology with several surge events preserved in the morphological record. This complex history of Tempelfjorden makes it difficult to discover and map landforms generated by the earlier surge events.

The complex history of terminal fluctuations in Tempelfjorden can provide an explanation to why the submarine morphological setting looks like it does. It can for example provide an explanation to why glacial lineations can only be found in some specific parts of the fjord, and why some surge related landforms cannot at all be found in the multibeam data. The rhombohedral crevasse fill ridges, mapped and described by Ottesen et al. (2008), (fig. 6) from other Svalbard tidewater surge
settings, cannot be found in Tempelfjorden, even though it is evident that the glaciers terminating into Tempelfjorden have been experiencing surges. The lack of rhombohedral crevasse fill ridges might be explained by the many surges which have modified earlier landforms in the fjord. For rhombohedral ridges to form the glacier would also need to remain stable and stagnant. It is possible that Tunabreen has been so active that even if seasonal winter stabilization occurs, the potentially formed ridges have been destroyed during the following melt-season glacier activity.

During a surge, fluid pressure is equal or nearly equal to ice overburden pressure. Basal crevasses form if the fluid pressure is near or equal to the ice overburden pressure. The basal crevasses can become filled with subglacial material during the early part of the quiescent phase, when the glacier sinks to its bed after the surge has stagnated (Ottesen et al. 2008). Since the fluid pressure is unknown during the initial phase after surge stagnation, it is not possible to determine which maximal glacier ice thickness would have enabled rhombohedral crevasse fill ridges to form in Tempelfjorden. If the ice overburden pressure is too large in relation to the fluid pressure, no ridges can be formed, which could be the case after the 2004 surge.

Many submarine landform assemblages, which have been observed in Svalbard surging glacier settings can also be found in Tempelfjorden. These landforms are for example the lobe shaped debris flow, which is found distal to the 1870 surge terminal moraine, the glacial lineations and the assemblage of retreat moraines in the inner fjord. These landforms and their relation to each other provide information about the surge events. The crosscutting relation between the glacial lineations in the inner part of the fjord and the retreat moraines confirm the relative chronology of deposition, where the glacial lineations are created during fast ice flow when the glacier surges out to the fjord. The moraines superimposed on the glacial lineations must therefore have formed when the glacier started to retreat back into the fjord from its maximum surge extent.

### 7.2. Evidence for annually formed retreat moraines

Figure 14 in the results section indicates that there seems to be a spatial relation between the position of the glacier front and the position of the submarine moraines. To fully evaluate this relationship, it would be advised to gain more data in the future by continuing the measurements of terminal fluctuations in relation to submarine data. In the future dating of the landforms would further help to establish an absolute time of their formation.

The merged bathymetric and satellite data show that the 2004 surge extent corresponds well with one of the most distinct moraines on the seabed. The 2005 glacial extent also corresponds to a submarine moraine. In the years 2007, 2008 and
2009 the terminal position also seems to correlate fairly well with moraines on the bed of the fjord. This would imply that there is an annual mechanism behind the formation of the moraines.

In 2006, the glacier terminal position does however not seem to correspond with any distinct moraine ridge. This complicates the interpretation, if the moraines are, as suggested by Ottesen and Dowdeswell (2006) annual winter push moraines. The absence of moraines correlating with the 2006 glacier terminal positions might be due to glacial dynamics, but it can also be that the moraine is too indistinct to show in the bathymetric data. However poorly distinguishable ridge-like feature in the swath data seems to correspond to the position of the 2006 glacier margin. Therefore, the glacier seems to not have experienced any winter advance during the winter of 2006/2007, since the 2006 October and 2007 June terminal positions are nearly identical (Fig. 15). The fact that the glacier terminus did not advance during the winter months, might be explained by differences in climate. A warm or less precipitation-rich winter could for example result in the terminus being nearly stagnant. The fact that the terminus of Tunabreen seems to not have advanced in the winter of 2006/2007 does however not explain the lack of a moraine in the position of the terminus, since many of the other moraines seem to have been created during similar winter still stands. The winter re-advances of the glacier front, observed in figure 15 in the results section have not been major. In the winter of 2008 the glacier experiences a winter advance, but most years the terminus seems to have been stagnant during the winter months.

The close relation between the position of the submarine moraines and the glacier frontal positions suggest that a causal relationship exists between them. Table 2, confirms that the glacier has been grounded during the entire 2004 surge advance and retreat cycle, therefore the grounding line and the above sea level terminal positions can be regarded as being almost the same, even though there can exist a minor offset. Therefore it seems evident that the retreat moraines proximal to the 2004 surge moraine are annually created push ridge moraines.

The surge landform assemblage for the 2004 surge does however differ from the landform assemblages related to the earlier surges in Tempelfjorden. There seems to be a clear difference between the 1930s, 1970s surges and the 2004 surge. No evidence of recessional moraines related to the 1930s or 1970s surges can be observed in the submarine landform record. This might first and foremost be explained by the chaotic morphology left by a glacier which has experienced multiple surge events during the last hundred years. It might also be explained by differences in floating condition of the glacier margin during or after the surge. It is however difficult to determine the floating condition during the 1930s and 1970s surges, since the total ice thickness can only be assumed. The depth of the sea floor at the 1970s
surge position is maximum 50 m, but the sea floor lies mostly at 40 m depth, which implies that the front of the glacier was most likely grounded both during and after the 70s surge.

The mapped glacier terminal positions after the 1970s surge however indicate that the glacier front was stagnant during a couple of years after it had reached the maximum terminal position in 1971, as revised from Plassen et al. (2004). The maximum position in 1971 corresponds well to the 1975 terminal position mapped from a satellite image. The glacier front seems to have experienced major retreat first in the late 1970s and early 1980s. This implies that the 1970 glacier retreat might have differed from the retreat after the 2004 surge. Whether this difference would have been due to internal glacier dynamics or climate related is difficult to evaluate.

7.3. Evaluation of the floating condition

In the results section it has been assumed that Tunabreen has no floating tongue and that the position of the terminus above sea level is equal to the grounding line. This assumption has been made to simplify the relation between the position of the submarine moraines and the position of the glacier terminus. It is however possible that the position of the glacier tongue relative to the grounding line has changed in time. This is particularly possible if the total height of glacier ice at the terminus has been fluctuating through time.

The floating condition of Tunabreen in historical times can be crudely evaluated by comparing the relation between the total thickness of glacier ice at the terminus and the depth of the sea floor. Through Archimedes principle, it can be determined that, at a certain bathymetric depth, a glacier with a certain ice thickness becomes floating. Archimedes principle enables to determine the depth of the ice/sea interface \( H_{\text{ice/sea}} \), given the total height of the meteoric ice \( H \), the density of meteoric ice \( \rho \) and the density of shallow sea water \( \rho_{\text{sw}} \):

\[
H_{\text{ice/sea}} = z_s - \frac{\rho}{\rho_{\text{sw}}} H
\]

Where \( z_s \) is the sea surface in relation to the bathymetry expressed in m. The average density of meteoric ice \( \rho \) is 0.9167 g/mL and the average density of shallow marine water \( \rho_{\text{sw}} \) is 1.025 g/mL. \( H \) is the total ice thickness. Only a single layer of meteoric ice has been considered in the equation, which therefore does not take into account refrozen marine water. The results, near the range of the present total ice thickness of around 80 m, have been summarized in Table 2. The numbers in the table show the heights of the water column between the seafloor and the underside of the ice (Fig. 16). The zero values in the table indicate that the glacier is grounded.
Figure 16: Sketch explaining the floating condition in a tidewater glacier setting. Zs and $H_{\text{ice/sea}}$ are measured in m from the sea depth at the coordinates (x,y) and are positive values. If $H_{\text{ice/sea}}$ is null (or negative), then the distance between the glacier and the sea depth is null and the glacier is grounded.

Table 2: The distance between the sea bed and the glacier boundary ($H_{\text{ice/sea}}$) as a function of depth below sea surface Zs and the total ice thickness H. Null values indicate that the glacier is grounded.
According Machenko et al. (2012) the height of Tunabreen above the sea ice was almost 40 m in the winter of 2011. The ice thickness was measured from the calving ice cliff. The average depth of the ocean bathymetry just in front of the present terminal position is around 40 m. By summarizing the two values we can obtain that the total ice thickness at present is almost 80 m. It can be assumed that the total ice thickness has been fairly constant during the recent surge event. Since Tempelfjorden is on average around 40 m deep, and not more than maximum 50 m deep, at its most inner part, the glacier must have been grounded during the entire 2004 surge event. The glacier would also have been grounded during the terminal retreat and the position of the ice cliff above sea level would therefore have been equal to the grounding. The presence of annual retreat moraines further confirm that the glacier must have been grounded during both advance and retreat.

If the total ice thickness has continuously been 80 m, a water depth of at least 75 m would be needed for the glacier to become floating. This would imply that the glacier has been grounded during the entire last hundred years, since the bathymetry in the inner part of Tempelfjorden is at maximum 60 m deep. But it can also be assumed that the total ice thickness has changed during surge events. If the total ice thickness and the height of the ice surface above sea level thin, the glacier may become floating in deeper waters. If the total ice thickness decreases to 70 m, a sea depth of only 65 m is required for the front to possibly become floating. If we assume that the ice thickness at the front of the glacier at some point in time was only 60 m, a sea depth of only 55 m would possibly have been enough for the front to become floating. This would also imply that an ice cliff of only 6 m protruded above the sea surface.

Under a surge event, the glacier front advances rapidly, which means that the ice thickness of the terminus decreases. When the surge ends and the ice flux from the interior basin decreases the glacier can experience further thinning at the terminus. This dynamic ice flux leads most likely to surge glaciers experiencing large differences in terminal ice thickness. It can therefore be possible that the front of Tunabreen has been floating during or just after earlier surge events. This might especially have occurred during the 1930s surge. There exists a deep depression in the bathymetry just before the 1930 surge extent. The sea floor depth is more than 50 m in the central parts of the fjord. The maximum depth is 60 m. If the total ice thickness was less than during the 2004 surge the front of the glacier could have been floating. Floating conditions might especially have occurred during the beginning of the retreat phase, when the terminus retreated into deeper water at the same time as the front of the glacier experienced thinning due to a decreased ice flux. This might also explain why no distinct annual retreat moraines are found in relation to the 1930s surge extent. It is therefore possible that the front was grounded during advance and floating during retreat.
7.4. Tidewater glacier surge mechanism

The mechanisms, which trigger and control a glacier surge, are still poorly understood. Surge triggering mechanisms remain obscure partly due to the difficulties of obtaining data from the interior of glaciers and from the glacier bed. The geomorphology created by a surge glacier provides one way of evaluating the dynamics and history of surge events. The question is if submarine landform assemblages can provide enough information to evaluate glacier surge dynamics. It is a common understanding that glacial lineations for example provide evidence of fast glacier flow during initial (advancing) stages of surge. Annual retreat moraines provide information about the retreat phase after the surge. The presence of eskers in the landform assemblage also provides information about the subglacial hydrology.

However a detailed surge dynamics, and especially surge triggering mechanisms are very difficult, if not impossible to establish from morphological evidence alone. The geomorphological data provides limited information about the internal mechanics of the glacier, such as the position of the glacier margin and the presence of crevasses at its base. Geomorphological records also provide information about the character of the glacier advance and retreat enabling a spatial reconstruction and establishing a relative chronology of landform generating processes during a surge event.

Glacier surges on Svalbard differ from surges in other parts of the world, which imply that the surge mechanism in Svalbard might be different from surge mechanisms in other parts of the world (Jiskoot et al 1998). The long surge duration of the glaciers in Svalbard suggests that a distinct surge mechanism is operating in Svalbard. The long quiescent phases could be ascribed to a slower mass accumulation in the Svalbard region compared to more temperate regions. Surge mechanisms might also be different in polythermal glacier settings compared to a temperate glacier setting. This might imply that surge triggers and mechanisms on Svalbard are different from surge mechanisms in other parts of the world, where glaciers have a temperate regime. Tunabreen differs however from the typical Svalbard surge glacier by having seemingly shorter surge durations as well as shorter quiescent phases.

The surge glaciers on Svalbard are located in a continuous permafrost setting (Humlum et al. 2003). The interaction between the glaciers and the permafrost might therefore also affect the surge dynamics. Svalbard glaciers are of polythermal type, which means that the outer parts of the glacier are frozen to its bed and might therefore overlie permafrost (Solheim 1991). Permafrost conditions are not likely to affect the dynamics of tidewater glaciers, which terminate into fjords, but it has been suggested that permafrost might exist even under the fjords of Svalbard (Kristenssens et al. 2009). Permafrost underlying a surge glacier enhances basal sliding. Ice
velocities could for example be enhanced by pressurized basal water, which gets trapped between the permafrost layer and the base of the moving ice leading to higher basal water pressures and further enhancing glacier flow. Thinning of the glacier during a surge would on the other hand lead to refreezing at the base and slower ice velocities.

The tidewater glaciers in Svalbard are mainly located on soft sediments of glacimarine origin, which distinguishes them from surge type glaciers in other parts of the world. Fine-grained glacimarine sediments offer little resistance to glacier flow, especially when they are unfrozen. Glacier surges in Svalbard can therefore most likely be at least partly explained by the soft bed mechanism. The soft bed mechanism provides an explanation to why there seems to be a difference between surge behavior of glaciers located in regions with deformable beds and glaciers overlying hard beds. Unfortunately it is still difficult to measure and understand the mechanisms of bed deformation.

Like many glaciers in Svalbard Tunabreen overlies soft fjord sediments. The soft bed mechanism could therefore be an important surge trigger in Tempelfjorden. Changes in thermal regime might also affect surge dynamics and glacier hydrology in Svalbard. Changes in the subglacial drainage system together with the deformable soft sediments can provide a plausible hypothesis to why Tunabreen surges. Subglacial water trapped at the glacier bed leads to increased sediment deformability, which in turn results in increased ice flow velocities. Water trapped beneath the glacier ice and the bed could explain the rapid advance of the terminus during a surge event. During the 2004 surge the rapid terminal advance, points to pressurized conditions at the glacier bed. The bathymetric data from Tempelfjorden does however not provide support for a single surge mechanism. Most likely glacier surges do not develop due to a single surge mechanism. It is more likely that several mechanisms work together during a surge. Feedback mechanisms between different glacial, geological and climate processes are also important, but extremely complex and presently poorly understood.
8. Conclusions

Multibeam bathymetric data, satellite and aerial imagery from Tempelfjorden and adjacent areas have been used to reconstruct the glacial morphological setting and evaluate the past extent and dynamics of Tunabreen. The results suggest that Tunabreen is a more dynamic glacier than many other Svalbard surge-type glaciers, since it has experienced three surges during the last century. Four surges since the Little Ice Age have been recorded in Tempelfjorden, compared to other Svalbard tidewater surging glacier settings, where the glaciers have been known to surge only once or twice. The multiple surge events in Tempelfjorden have resulted in a complex pattern of landforms on the seafloor.

The 2004 surge is the most clearly distinguishable event in the submarine morphology. Tunabreen has been grounded during the entire 2004 surge event, which suggests that the position of the terminus above the sea level has been nearly identical to the position of the grounding line. The glacier may however have been partly floating during earlier surge events. No distinct submarine annual retreat moraines are found in relation to the 1930s and 1970s surge extents, which might be explained by a difference in floating condition.

The retreat after the 2004 surge has left behind distinct annual retreat moraines, which can be correlated to the position of the retreating terminus of Tunabreen. The moraines have most likely formed during winter terminal still stands or minor re-advances. However to fully evaluate the spatial relationship between submarine moraines and the position of the glacier terminus more data must be obtained by continued measurements of the terminal fluctuations in relation to the submarine data. Dating of the landforms would further help to establish an absolute chronology of the events which have taken place in Tempelfjorden.

An interdisciplinary approach combining geological, oceanographic, glaciological, hydrological and modeling studies is clearly required to improve our understanding of the mechanisms and processes associated with the glacier evolution and their behavior during surge events.
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