Surge-type glaciers:

controls, processes and distribution

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Abstract

Glacier surging is an internally triggered instability. Surge-type glaciers periodically alternate between long periods of slow flow (the quiescent phase) and short periods of fast flow (the surge phase). Surging yields down-glacier transport of mass and often results in large and sudden glacier advances. The surging phenomenon has always challenged the notion of normality in glacier flow dynamics. The mechanisms of surging remain poorly understood. Observation of different surge behaviors across the world has been used as evidence for the development of glacier type-specific surge models that lack transferability and representativeness. Although only about 1% of the entire glacier population has been observed to surge, the surge phenomenon questions the completeness of our understanding of glacier dynamics.

This thesis uses different perspectives to gain a new understanding on the global, regional and local controls on surging and reconcile the mechanisms of surging under a single model. Through a geodatabase of surge-type glaciers, datasets of climate and glacier geometry variables and a global distribution model we explore the controls on the non-random distribution of surge-type glaciers on a global scale. The highest densities of surge-type glaciers are found in a well-defined climatic envelope bounded by temperature and precipitation thresholds, while glacier geometry exerts a second-order control on their distribution. We introduce the enthalpy cycle model which relates flow oscillations to imbalances between enthalpy gains and losses. Enthalpy balance is satisfied outside of the optimal surge envelope, in cold and dry or warm and wet regions. However, the intermediate conditions of the optimal surge envelope prevent enthalpy balance to be reached, yielding dynamics cycling of glacier flow.

Thermal switch models have been used to explain surging of polythermal glaciers. We reconstruct the evolution of the thermal regime of six glaciers in Svalbard from existing and new data. The large and thick surge-type glaciers of our sample do not return to a cold-based conditions between surges, demonstrating that thermal switching cannot apply to surges of large glaciers in Svalbard. On the other hand, the thin and mostly cold glaciers display evidence of former warm-based thermal regimes, showing that switches in climate can make glaciers go in and out of surging. We demonstrate that the concept of enthalpy cycling can explain surge and surge-like behavior in Svalbard.

Finally, we investigate the role played by local controls on the initiation and development of the surges of two large polythermal glaciers in Svalbard. First, passive seismics and DEM differencing enabled the reconstruction of the chronology of events that led to the catastrophic surge of the Nathorstbreen glacier system. Removal of backstress by the failure of the frozen glacier terminus

triggered the catastrophic collapse of one of the tributaries of the glacier system, source of unusual seismic activity. Secondly, the upward propagating surge of Svalbard tidewater glacier Aavatsmarkbreen is understood in terms of changes in the force balance. Glacier retreat and thinning caused a rapid steepening of the glacier snout, which in turn increased the driving stresses substantially. Development of crevasse fields during the late quiescent and surge phases allowed transfer of surface meltwater to the bed, increasing basal water storage and causing ice acceleration. The increase in driving stress and surface-to-bed drainage both contributed to basal enthalpy production, and controlled the pattern of surge evolution.

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Chapter 1: Introduction

1.1 Motivation

Glaciers have been defined as 'natural climate-meters' in the IPCC Fifth Assessment report. They not only act as passive indicators of climatic changes but also contribute actively to the global climatic balance (IPCC, 2013). Observations have shown that glaciers across the world have been shrinking since the end of the Little Ice Age, and that the rate of mass loss has been increasing since the 1980s (Leclercq and others, 2011). Together with thermal expansion, glaciers have made a major contribution to global sea level rise over the 21st century (Cazenave & Le Cozannet, 2014). There is high confidence that glaciers and ice sheets will keep losing mass even without any further changes in climate (Cubasch and others, 2013).

Estimations of future sea level rise suffer from two main issues. First, they are based on a small proportion of the world's glaciers, and second they rarely include a dynamic response of the glaciers to changing climatic conditions. Improving the integration of glacier dynamics in future estimates of glacier mass loss must therefore originate from a better understanding of glacier dynamics. Glacier calving and glacier surging are at the crux of this challenge. While calving rates are predicted to make an increasing contribution to sea level rise, directly related to increasing air and water temperatures, surging glaciers behave in a more unpredictable way, and episodically discharge large volumes of ice on land or in the oceans.

The surge phenomenon was defined by Kamb and others (1985) as one of the 'outstanding unsolved problems of glacier mechanics', and remains so today. Several decades of studies have so far been unsuccessful at identifying what makes glaciers surge. The study of glacier surging aims not only to better understand the processes behind the unsteady flow of these glaciers but also requires a comprehensive reassessment of the physical laws of glacier flow. Beyond investigating glacier dynamics, the study of surging comprises numerous strands. The cyclic behavior of surge-type glaciers, alternating between fast and slow flow, has often been compared to the unstable behavior of ice streams (Clarke, 1987 ; Bindschadler, 1997). Surging of ice streams is thought to have played a major role in extension of palaeo ice sheets (Boulton and others, 1977 ; Andreassen and others, 2014).

The behavior of surging glaciers is characteristically decoupled from climate trends. Still today in a period of global glacier recession, glaciers surge in many parts of the world. Surging glaciers complicate the investigation of glacier response to climate variability. The hazards that surging glaciers represent are non-negligible. Although they tend to surge regularly, predicting the comparing the evolution of the thermal structure of a representative set of Svalbard glaciers, the processes taking place throughout the surge cycle are questioned. The investigation of small and thin, large and thick, tidewater and land-terminating glaciers could reveal whether glaciers can switch in and out of surging cycling in response to climatic changes.

Finally, the processes taking place during two recent surges in Svalbard are studied using a variety of techniques. Passive seismics and elevation differencing allow the reconstruction of the chronology of events that led to the large surge of the Nathorstbreen glacier system. The aim is to assess the role played by changes in the force balance in the triggering and development of a surge, and to show how passive seismics can provide a unique insight into the mechanisms of surging. In a second example, the surge of a tidewater glacier is monitored through feature-tracking of SAR imagery, combined with measurements of elevation changes, crevasse mapping and calculation of the driving stress. This approach aims to uncover the mechanisms taking place during the upward propagating surges of tidewater glaciers in Svalbard.

1.3 Outline

An overview of the thesis is provided in Chapter 1, followed by the relevant theoretical background in Chapter 2. Chapter 3 focuses on the definition, distribution and characteristics of surge-type glaciers, along with details of the main surge models. As this work assesses surging both on a global and regional perspective, Chapter 4 follows this progression: in a first part we review the global distribution of surge-type glaciers and variations in surge characteristics across regions, and secondly we introduce the archipelago of Svalbard, home to the densest cluster of surge-type glaciers on Earth. The main techniques used for this work are described in Chapter 5. Chapter 6 summarizes the four articles included in this thesis. Chapter 7 concludes this work and suggests future work perspectives. References can be found in Chapter 8.

Finally, Chapter 9 displays the four articles included in this thesis in their entirety: Article I: Sevestre, H. and D.I. Benn (2015) Climatic and geometric controls on the global distribution of surge-type glaciers: implications for a unifying model of surging. *Journal of Glaciology ;* Article II: Sevestre, H., Benn, D.I., Hulton, N.R.J., Baelum, K. (In Review) Thermal structure of Svalbard glaciers and implications for thermal switch models of glacier surging. *Journal of Geophysical Research – Special Issue on Surging and Ice Streaming* ; Article III: Sevestre, H., Köhler, A., Benn, D.I., Nuth, C., Luckman, A., Weidle, C. (In Prep) Seismic detection of a catastrophic glacier surge and Article IV: Sevestre, H., Benn, D.I., Luckman, A., Nuth, A., Kohler, J., Lindbäck, K., Pettersson, R. (In Prep) A tidewater glacier surge initiated at the terminus: Aavatsmarkbreen, Svalbard.

occurrence of a surge still remains elusive. The advance of a surging glacier can result in river damming and disruption to roads, trails, snowmobile routes, etc. Large quantities of ice discharged on land or in the ocean represent a significant danger, particularly in coastal areas and where dense maritime traffic occurs. The end of the surge phase is often marked by the release of huge volumes of turbid water causing floods and destruction. Predicting the occurrence of surges could improve protection of communities and traffic.

Many questions remain partially or completely unanswered in the study of surging. To build on Jiskoot (1999), some of the main issues that need further addressing concern the controls on surging and the mechanisms taking place before, during and after a surge:

- What are the controls on the global distribution of surge-type glaciers?
- Why, within the same cluster, some glaciers surge while others do not?
- What role does climate play in the distribution and occurrence of surges?
- Can one surge mechanism be applied to all surge-type glaciers, regardless of their thermal regime?
- Can normal glaciers become surge-type glaciers, and vice versa?
- What is the role of thermal regime in the surges of polythermal glaciers in the Arctic?
- Why do tidewater glaciers surge differently from that of land-terminating glaciers?
- What is the surge trigger?

This thesis aims to tackle these issues by taking a novel approach in the study of the surge phenomenon.

1.2 Aims and objectives

Investigating the controls on the non-random distribution of surge-type glaciers could unlock a new understanding of the surge phenomenon. This work aims to identify these controls by using a global inventory of surge-type glaciers, along with global datasets of modelled climatic data and information on glacier geometry. The climatic distribution of surge-type glaciers and geometry properties across the climatic spectrum could lead to the launch a new theory as to why glaciers surge in some parts of the world, while they do not surge in other regions.

Secondly, this work aims to test and evaluate the thermal switch mechanism, which is typically used to explain surges of polythermal glaciers in Svalbard and other Arctic regions. By

Chapter 2: Scientific background

The uniqueness of surge-type glaciers can only be grasped if the characteristics and processes of ‘normal’ glacier flow are reviewed first. Basics of glacier flow are defined in section 2.1, before moving on to the concept of balance velocities in section 2.2. Glacial hydrology and water storage are described in section 2.3. Finally, section 2.4 focuses on ice temperature and thermodynamics.

2.1 How do glaciers flow?

2.1.1 The force balance

Glacier flow is a matter of balance between stresses that drive the flow and stresses that acts against it. All glaciers move under their own weight, or more particularly under the horizontal gradient in gravitational potential, called the driving stress. In equilibrium, the driving stress is balanced by resisting stresses, namely the basal shear stress (that acts along a surface, in this case the bed), lateral drag (on the glacier's sides) and the longitudinal stress gradients. Basal drag can be negligible in the case of ice shelves, or ice streams, where most of the resistance occurs at the sides. Longitudinal stress gradients exert pushes or pulls, driving or resisting the flow. In the force balance, the efficiency of the driving stress at driving the flow is compared to that of the resisting stresses at restricting it. The force balance of a glacier in equilibrium is zero.

2.1.2 Ice deformation and fracturing

Under stresses, ice either deforms or fractures. Fracturing occurs when the strength of the ice is overcome by pulling stresses. It is a major process in the flow of tidewater glaciers, ice shelves and ice streams. Ice creep, on the other hand, is a much more widespread and efficient process of ice motion.

The structure of a crystal of ice can be compared to a deck of card. Molecules can easily glide on top of each other when deformation occurs along the crystal basal planes. However non-basal plane glide requires much higher stresses, and is often referred to as ‘hard glide’. Linear defects in the structure are thought to largely facilitate slip along basal planes. Under stress, polycrystalline ice responds by creeping. In addition to movements of dislocations within crystals and glide of the crystals on top of one another, crystal growth and recrystallization also contribute to ice deformation.

Flow laws relate the rate of ice deformation to stress. They are essential to model flow dynamics and in the study of many glaciological mechanisms. Glen's flow law, today's most widely employed flow law is based on extensive laboratory experiments (Glen, 1955). It quantifies the rate of ice deformation at the secondary creep phase, and relates a dominant shear stress to the rate of ice deformation $\dot{\varepsilon}$. It must be acknowledged that Nye (1957) made the first applications of the power law, which led to the subsequent adoption of the following form:

$$\dot{\varepsilon}_{xz} = A{\sigma_e}^{n-1}\,\tau_b \qquad \textbf{(Eq. 1)}$$

where τ_b is the basal shear stress, and σ_e the effective stress that incorporates all the stress components. Parameter n is defined as the exponent. Its value is governed by the creep mechanism operating. Results from laboratory experiments point to a value of 3. Rate parameter A relates to the viscosity of the ice. Its value can be calculated by the Arrhenius relation (Hooke, 1981), and relates to the ice temperature, fabric, water content, density and grain size. One of the main implications of Glen's flow law is that ice does not deform linearly in response to stress. It is a distinct property of non-Newtonian (or non-linear viscous) materials. In addition, it only relates stress to strain rate; a more general constitutive relation would also include fracturing and elastic deformation.

2.1.3 Basal processes

Basal motion occurs by a combination of ice creep, sliding and deformation of the substrate. As challenging as the investigations of subglacial processes can be, large advances were made using subglacial laboratories and by studying recently deglaciated areas.

By observing that the presence of protuberances over hard beds was not preventing the flow of glaciers, Weertman (1957) described two processes by which ice can move over and around bed bumps: regelation and enhanced creep. Regelation (or re-freezing) occurs around small obstacles. High pressures on the upstream side of bumps cause a lowering of the pressure melting point. Ice ultimately melts, and meltwater travels to areas of lower pressure on the downstream side of bumps. There, the pressure melting point is raised and ice refreezes. Latent heat thus released is conducted through the obstacle and further enhances ice melt upstream (Weertman, 1964 ; Lliboutry, 1968 ; Lliboutry, 1987 ; Kamb, 1970). Field observations by Kamb and Lachapelle (1964) and Cohen (2000) have confirmed the occurrence of regelation. Enhanced creep relates to the changes in ice viscosity in response to conductive stresses. These stresses are high on the upstream side of large bumps, lowering the viscosity of the ice, and increasing (in a non-linear fashion) deformation rates.

The size of the obstacle determines the favored process. As the transfer of latent heat is more efficient on small bumps than on large obstacles, Weertman (1964), Nye (1969, 1970) and Kamb (1970) were able to define a critical obstacle size of 0.5 m below which regelation occurs, and otherwise enhanced creep is favored. Although as suggested by Lliboutry (1993), the two mechanisms undoubtedly work in combination.

Short-term and significant variations of velocity observed by Iken and Bindschadler (1986) exposed the incompleteness of Weertman's theory of sliding. Processes other than ice deformation and changes in driving stress must be taking place at the bed. Lliboutry (1968) was the first to observe the formation of cavities at the interface between ice and a rough bed. Cavities forming on the lee side of obstacles can be filled with water, reducing basal drag. They expand when more water is delivered than can be discharged, and decoupling occurs when the water pressure exceeds the ice overburden pressure. The driving stress is then only supported in areas where the ice is in contact with the bed.

Glacier flow over a soft bed is radically different. Substrate deformation and sliding (ploughing) over the till can contribute to ice flow. Boulton (1979) and Boulton and Hindmarsh (1987) showed that in Iceland, deformation of the top few centimeters of the substrate was contributing from 80% to 90% of the ice motion. The Boulton-Hindmarsh model states that the strain rate of granular materials increases as the basal drag becomes more important that the yield strength of that material, but decreases as basal water pressures top the ice overburden pressures. On the other hand, Kamb (1991), and Engelhardt and others (1990) showed that under Ice Stream B, the till failed completely past a threshold in applied shear stress (Coulomb-plastic rheology).

Water plays a central role in till deformation and sliding of ice over its substrate. Low water pressures encourage ice infiltration in the pore spaces of the substrate. To the contrary, high water pressures prevent infiltration and allow ice to slide over the substrate.

2.2 Balance velocities

Glaciers constantly have to balance rates of accumulation and ablation, and variations in their force balance. They naturally tend to modulate their flow velocity in order to match mass gain up-glacier with mass loss downstream, and therefore maintain a more or less fixed geometry over long time scales. This outlines the concept of balance velocity. High turnover glaciers typically found in wet maritime environments tend to have higher balance velocities than glaciers in more continental, drier and cooler regions. Balance velocities are also strongly influenced by the glacier geometry. Glaciers with large catchments and narrow outlets will tend to flow faster. Thermal regime, the type of basal substrate and bed topography also influence the balance velocities.

However, glaciers often depart from steady-state and are either growing or shrinking. Changes in the balance between driving and resisting stresses cause short to long-term changes in glacier flow velocities. For example, the driving stress can increase when the surface profile of a glacier steepens, either by melting at the terminus or unusual accumulation in its upper parts; and resisting stresses can be modulated by variations in subglacial water storage and pressure.

2.3 Glacial hydrology and its influence on glacier dynamics

2.3.1 Water supply and plumbing system

Meltwater is produced at the surface from snow/ice melt, within the glacier when ice deformation causes friction between ice grains, and under the glacier from geothermal heating and basal friction. Supraglacial melt is the greatest source of meltwater although it varies spatially over the glacier surface and in quantity seasonally. Internal and basal melt is a steadier water supply although volumes can change drastically over long timescales. Rain and groundwater also feed the glacier's hydrological system. Water can be routed from the surface to the bed, and also from the bed to the surface depending on the processes that drive or resist water flow.

2.3.2 Drainage systems and storage

The controls that the glacier hydrological system exerts on ice motion depend on supply, drainage and storage, and connections between the supraglacial, englacial and subglacial systems.

Meltwater produced in the melt season can either percolate through the firn in the accumulation zone, or runoff on bare ice in the ablation zone. Refreezing in the firn releases latent heat, which in turn progressively brings the layer to the pressure melting point. Any further increase in temperature then leads to melting. Drainage through the firn is very inefficient compared to runoff on bare ice. Supraglacial meltwater cuts channels through the glacier surface that can incise deep over one melt season, as long as channel incision rates are greater than surface ablation rates (Gulley and others, 2009). Water can also be stored in ponds on the glacier surface in areas of gentle slope.

Four types of englacial drainage systems connect the surface with the bed. Moulins form when supraglacial meltwater exploits a fracture on the glacier surface. Sustained water supply will allow the moulin to grow and deepen. Hydrofracturing of crevasses occurs when the pressure that water exerts onto the crevasse walls overcomes the strength closing the crevasse (Röthlisberger & Lang, 1987). It allows rapid downwards propagation and connection to the basal plumbing network (Das and others, 2008 ; Stevens and others, 2015). This process has been shown to contribute to a great extent to the variations in flow of the Greenland ice sheet (Zwally and others, 2002 ; van de Wal and others, 2008). Incision of supraglacial meltwater channels also connects supraglacial meltwater to the englacial and subglacial networks (Fountain & Walder, 1998 ; Gulley and others, 2009). Finally, Gulley and Benn (2007) have shown that water can also easily exploits lines of a different permeability to that of the ice, such as debris-filled fractures, crevasse traces, and other similar fractures. Storage within temperate glaciers can occur in conduits, crevasses or fracture network, as observed on Storglaciären by Fountain and others (2005), while blockages of incised meltwater channels commonly occur in polythermal glaciers (Gulley and others, 2009).

Subglacial drainage systems have a huge influence on ice dynamics. They can either be channelized when water is discharged through an efficient network of conduits, or form distributed systems that are relatively inefficient at discharging water. Channelized systems are divided into R- (or Röthlisberger) channels incised between the bed and the ice, N- (or Nye) channels cut into bedrock, or sediments and large tunnel valleys. In times of high water input, high water pressures in the channels force the water out, while low water inputs cause the water to migrate towards the channels. Tunnel size also adjusts to the water intake, making the drainage efficiency of the network increase through the melt season.

Distributed subglacial drainage systems on hard beds are divided into thin water films and linked cavities. Thin water films are maintained when energy gained (geothermal activity) or produced at the bed (strain, frictional heating) is greater than energy conduction through the ice. They are found at the interface between the ice and the bed, and have the most limited ability to transport meltwater. As seen in section 2.1.3, linked cavities exerts a large control on flow speed, reducing basal drag to areas of ice-bed contact. Large linked cavities can progressively turn into an efficient, channelized drainage system as a result of high fluxes. Kamb (1987) suggested this could be a mechanism leading to the termination of glacier surges.

Distributed systems over or within soft beds can discharge large amounts of water (Alley and others, 1986). Water can be stored in the substrate pores and advected during shearing of the subglacial layer. It can also flow through the pores of the substrate under a hydraulic gradient. The volume discharged depends on the permeability and thickness of the aquifer. A water-saturated horizon has a strong influence on basal sliding and sediment deformation. Films at the ice-sediment interface can form when the water supply exceeds the availability of pore spaces. Finally, water can travel in shallow channels or “canals” within the substrate. Such systems are expected to be present below gently sloping ice streams. Subglacial drainage systems largely control flow rates, and in turn, ice motion can alter the drainage systems. Such feedback mechanisms remain poorly understood.

2.4 Ice temperature and thermodynamics

Ice temperature exerts a strong control on flow dynamics. Switches in basal temperature can result in a dramatic acceleration of glaciers, and are thought to be involved in the surging of polythermal glaciers (Fowler and others, 2001).

2.4.1 Controls on ice temperature

Changes in ice temperature can take place at the ice surface, englacially and subglacially. At the glacier surface, energy exchanges with the atmosphere can raise or reduce ice temperature. In the accumulation zone in particular, latent heat released by the refreezing of meltwater can significantly increase the ice temperature and eliminate the cold winter wave. Once the firn is raised to the melting point, any excess in heat will produce melting. In winter, snow insulates the glacier, reducing heat loss. Conduction defines the transfer of heat from warm to cold areas in the glacier. It is most efficient where steep thermal gradients occur. Advection of ice transfers temperate ice produced in the accumulation zone downstream, increasing temperatures at depth, whereas advection of cold ice originating from the interior of ice sheets reduces temperatures. Ice deformation produces most heat at depth, where the shear stress is largest. At the glacier bed, geothermal heat needs to be greater than conduction in order to raise the ice temperature to the melting point. Other processes such as frictional heating and regelation can significantly warm basal ice, as well as sliding and subglacial sediment deformation.

2.4.2 Thermal regimes

Glaciers can be divided into three categories based on their thermal structures. Temperate glaciers are composed of ice at the pressure melting point, except for a shallow surface layer a few meters thick subjected to seasonal changes in temperature. Between 0.1% and 2% of water can be found between grains (Lliboutry, 1976). For a glacier to be fully temperate, the cold winter wave has to be eliminated either by release of latent heat in the accumulation zone, or more efficiently by ablation in the summer. The thickness of the cold wave depends on winter temperatures and insulation provided by the snowpack. Therefore temperate glaciers are most likely to occur where snow accumulation and ablation rates are high, such as in temperate-maritime climates.

Cold glaciers are frozen to their beds. Heat production is less efficient than heat loss, maintaining the ice temperature below the pressure melting point. Cold glaciers are typically found in cold and dry environments such as the Dry Valleys in Antarctica.

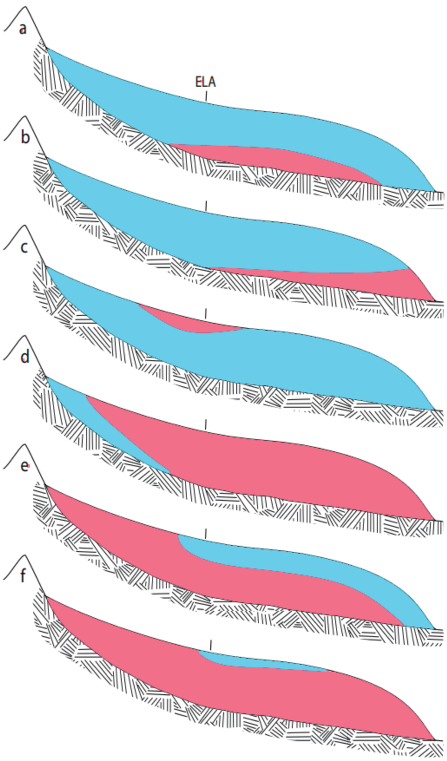

Figure 1: thermal structures of idealized valley glaciers. Red indicates temperate ice, blue indicates cold ice. Modified from Pettersson (2004).

Polythermal glaciers contain temperate and cold ice. The proportion of both ice types varies, creating a spectrum of thermal structures spanning from predominantly cold glaciers with a limited temperate basal layer, to predominantly warm with cold ice close to the surface. Six main types of polythermal valley glaciers have been described by Blatter and Hutter (1991) and Pettersson (2004) (Fig. 1), although the spectrum of polythermal structure shall not be limited to these 6 examples. Types a and b are found in cold environments where negligible melt occurs in the accumulation area.

Strain heating raises parts of the bed to the pressure melting point. Latent heat release of meltwater in the low accumulation area produces a limited core of warm ice (type c). In regions where temperate ice is produced by the same process, and the cold wave is ablated in the summer, the glacier becomes predominantly temperate, with cold ice still being produced in the uppermost accumulation area (type d). In type e, the winter cold ice is not successfully removed in the ablation zone. Glaciers of this type are widespread in regions like Svalbard where snowpacks are thin and melt rates moderate. Finally, glaciers of type f are found where summer ablation is important and strips the cold wave from the ablation zone in the summer.

2.4.3 Feedback mechanisms between ice temperature and dynamics

Ice temperature and glacier flow dynamics are interrelated. Changes in ice temperature exert a strong influence on ice dynamics, while ice dynamics can, through feedback mechanisms, gradually modify a glacier's thermal structure (Benn & Evans, 2010).

Positive feedback mechanisms are thought to be involved in the trigger of glacier surges. Glacier acceleration yields enhanced frictional heating at the glacier base. This is turn begins to warm the ice, and eventually causes melting. Increasing volumes of meltwater promote faster sliding until decoupling occurs. The mechanism stops when frictional heating at the bed is not efficient enough to keep feeding the system.

Negative feedback mechanisms can occur in response to changes in geometry of the glacier. The mass continuity equation relates a change in glacier thickness through time to the difference between the mass balance and the flux of a quantity of ice. If the lower part of the glacier starts accelerating, dynamic thinning occurs making the glacier more vulnerable to conduction from the bed to the surface. Inversely, if more ice is transported down-glacier, thickening occurs and conduction is limited.

2.4.4 Enthalpy

Modelling the thermodynamical properties of polythermal glaciers is challenging as they are composed of a fraction of cold ice, and a fraction of temperate ice. These two ice types do not respond the same way when submitted to a change in heat content. Raising the heat content of cold ice results in a change in temperature, while raising the heat content of temperate ice yields a change in water content (Aschwanden & Blatter, 2005, 2009). This difference in ice properties can be understood within the framework of enthalpy. In a glaciological context, enthalpy is the 'internal energy' of a glacier.

Enthalpy is produced when a glacier flows downslope and gravitational potential energy is converted into thermal energy. Radiative and turbulent heat fluxes at the surface can produce or discharge enthalpy. Internal and basal strain heating increase enthalpy, as well as latent heat release and geothermal heat fluxes. Enthalpy gains can be dissipated at the glacier surface and bed by runoff, and through the glacier by conduction or by calving.

For a glacier to remain in steady-state, not only it must flow at its balance velocity to balance mass accumulation with mass loss, but it must also maintain equilibrium between enthalpy gains and enthalpy losses. Any surplus or deficit in mass and enthalpy will have consequences on the glacier dynamics. At the glacier bed, increased strain heating can raise the ice temperature and further produce basal meltwater. If the meltwater and the rising temperature cannot be evacuated by conduction or runoff as fast as they are generated, positive feedbacks will cause the glacier to accelerate above its balance velocity, advecting ice faster than can be produced. To the contrary, if heat is lost more rapidly by conduction or runoff faster than it is produced, the glacier will decelerate and ice will accumulate within the system.

Chapter 3: Surge-type glaciers

This chapter focuses on surge-type glaciers, their dynamics, distribution and mechanisms. The characteristics of the surge cycle are defined in Section 3.1. The contemporary distribution of these glaciers and variations in surge behaviors observed across the world are detailed in section 3.2. Section 3.3 focuses on the identification of surge-type glaciers, along with the relevant techniques, and evidence to look for. Section 4.4 summarizes the results of statistical studies investigating controls on regional distributions of surge-type glaciers. Finally, models of surge mechanisms are presented in Section 3.5.

3.1 Definition

The spectrum of glacier flow velocities stretches from slow-moving cold-based glaciers in the Dry Valleys of Antarctica or Arctic Canada to extremely rapidly-flowing outlet glaciers in Greenland or ice streams in parts of Antarctica. Flow behavior spans from more or less constant flow vs to velocity pulses. Surge-type glaciers have the ability to periodically switch between long periods of continuously slow flow (the passive - or quiescent phase) and short periods of very fast flow (the active - or surge phase). Surging is defined as an internally triggered instability, as opposed to other pulsating behaviors such as the patterns of advance and retreat observed on Greenland outlet tidewater glaciers, which are influenced by bed morphology and climate (Meier & Post, 1969 ; Sharp, 1988). The surge cycle is composed of the quiescent and the surge phase. Its length tends to be more or less constant for each surge-type glacier (Meier & Post, 1969).

The quiescent phase typically lasts from a couple to several decades, and is characterized by flow speeds below balance velocities. Seasonal variations in velocity can still occur during quiescence (Abe & Furuya, 2015 ; Burgess and others, 2013). Ice builds-up in the 'reservoir zone' up-glacier, while flow is restricted in the lower reaches of the glacier, altering its longitudinal profile. 'Mini-surges' or 'wavy surges' can be detected a few years prior to the beginning of the main surge, travelling down the glacier at great speeds (few hundreds of meters per hour) (Kamb & Engelhardt, 1987 ; Dolgoushin & Osipova, 1978). The thickening continues, forming a clear front between the stagnant ice down-glacier and the increasingly active ice up-glacier (Clarke and others, 1984). The bulge or ice front steepens until the surge is triggered (Meier & Post, 1969 ; Raymond, 1987).

During the surge, the glacier velocity can increase by a factor of ten. These high velocities are maintained from a few months to a few years only. The changes in glacier geometry during quiescence are rapidly reversed as mass is transferred down-glacier into a 'receiving zone'. The thick surge front travels down-glacier, causing compression as it moves into more stagnant ice, and extension behind it. Drawdown of the glacier surface in the reservoir zone leaves ice hanging on the valley sides (Post & LaChapelle, 1971). As the ice stretches, the glacier becomes intensively crevassed. The down-glacier propagation of the surge may reach beyond the glacier limit in fast and dramatic advances. Exceptions to this pattern are some tidewater glaciers in parts of Iceland, Greenland and Svalbard. There, surges have been observed to begin at the calving front and propagate up-glacier (Björnsson and others, 2003 ; Rolstad and others, 1997 ; Dowdeswell & Benham, 2003 ; Murray and others, 2003b, 2012 ; Pritchard and others, 2003, 2005).

The mechanisms taking place during a surge are still poorly understood, although it is clear that ice deformation cannot solely account for the surge velocities. A combination of sliding and substrate deformation at the glacier bed must be involved (Raymond, 1987).

3.2 Distribution of surge-type glaciers and variations in surge behaviors

Meier and Post (1969) noted that surging can affect glaciers of all types, and can occur in almost all climatic environments. One of the most fascinating facts about the population of surge-type glaciers is its non-random distribution both on the global and regional scale (Raymond, 1987). During the past 100 years or so, surging has occurred in two main 'superclusters' namely the Arctic Ring, and western Central Asia. Within the Arctic Ring, surge-type glaciers are clustered in clusters in Alaska, Yukon Territory, Arctic Canada, parts of Greenland, Iceland, Svalbard and Novaya Zemlya (Post, 1969 ; Fischer and others, 2003 ; Copland and others, 2003 ; Jiskoot and others, 2002, 2003 ; Grant and others, 2009 ; Citterio and others, 2009 ; Yde & Knudsen, 2007). In western Central Asia, glaciers have been observed to surge in the Karakoram, Pamirs and western Tien Shan (Copland and others, 2009, 2011 ; Hewitt, 1969, 1998 ; Kotlyakov and others, 2008 ; Osipova and others, 1998). A small number of surge-type glaciers have been reported in the Caucasus, parts of the Andes, Russian high Arctic, Kamchatka and Tibet (Kotlyakov, 1996 ; Kotlyakov and others, 2004 ; Casassa and others, 1998 ; Espizúa, 1986 ; Zhang, 1992 ; Yafeng and others, 2010 ; Dowdeswell & Williams, 1997 ; Dolgoushin & Osipova, 1975). The total population of surge-type glaciers is thought to represent about 1% of the global population of glaciers (Jiskoot and others, 1998).

Equally important to the known distribution of surge-type glaciers are glacierized regions where surge-type glaciers do not exist today. These are the Brooks Range and the Southern Coastal range in Alaska, the contiguous states of the USA, northernmost and southernmost Greenland, Pyrenees, European Alps, mainland Scandinavia, Franz Josef Land, Himalayan Range, Bhutan, New

Zealand, Northern Andes, and Verkhoyansk in Russia. Although Wellman (1982) explained the former geometric changes of Fisher glacier through surging, no surges have been directly observed in Antarctica.

Evidence shows that the population of surge-type glaciers might have migrated in the past following climatic patterns. The European Alps hosted at least one surge-type glacier in Austria. Vernagtferner underwent five successive surge-like advances between the 17th and 19th century (Hoinkes, 1969). Other studies reveal that surge behaviors can be modulated by climate. In Iceland Striberger and others (2011) traced the surges of Eyjabakkajökull over the past 1700 years showing a clear reduction in the length of the surge cycle in times of high precipitation rates and low temperatures, and an increase when the conditions were less favorable. In the Karakoram, a sustained rise in precipitation has apparently led to the “Karakoram anomaly”: a sudden increase in the number of surge occurrences (Hewitt, 2005). In Svalbard, Dowdeswell and others (1995) interpreted a reduction in the frequency and occurrence of surges in Svalbard as a consequence of climate change, although recent data contradict these conclusions as more than 15 glaciers are currently surging on the archipelago (pers. Com. Luckman, April 2015).

Surge behavior varies greatly between regions and within individual surge-clusters. Two contrasting types of surge behavior have traditionally been recognized: the Alaska-type surge and the Svalbard-type surge, which are thought to be a reflection of the predominant thermal regimes found in these regions. However, in reality a wide spectrum of behaviors actually exists between and beyond these two classes. The shortest surges have been observed in Iceland, Alaska, Yukon Territory and in the Pamirs, where they rarely exceed 4 years in duration (Osipova and others, 1998 ; Thorarinsson, 1964, 1969 ; Dolgoushin & Osipova, 1975 ; Post, 1969). These clusters have correspondingly short quiescent phases. Average surge velocities are highest in Alaska and Iceland (Eisen and others, 2001 ; Kamb and others, 1985 ; Björnsson and others, 2003). The longest surge cycles take place in Greenland, Svalbard and Arctic Canada (Liestøl, 1969 ; Dowdeswell and others, 1995 ; Hagen and others, 1993 ; Weidick, 1988 ; Jiskoot and others, 2003). Glaciers in the Karakoram are typically found between these two groups, with moderately long surges (3 to 6 years) and quiescent phases from 15 to 70 years (Copland and others, 2011 ; Hewitt, 1998). Heterogeneities in the evolution of surges are also clear within this region (Quincey and others, 2015). The development of the surge cycle also varies from one region to another. Murray and others (2003b) showed that the surge of Monacobreen was characterized by a multiyear phase of steady acceleration, and terminated with a gradual deceleration, as opposed to rapid initiation and termination for the surge of Variegated glacier in Alaska (Kamb & Engelhardt, 1987 ; Kamb and others, 1985).

3.3 Identification of surge-type glaciers

The identification of surge-type glaciers is complicated by the changes in geometry, appearance and behavior the glacier undergoes throughout the surge cycle. A glacier in full surge is relatively easy to identify, and the surge dynamics can leave distinct geomorphological evidence that enables identification during the quiescent phase. Surge-type glaciers can be identified using a suite of glaciological and geomorphological evidence. Some features have a stronger diagnostic power than others, and the most reliable identifications are based on a combination of surge-indicative features. Copland and others (2003, 2011) and Grant and others (2009) have made comprehensive lists of features for the identification of surge-type glaciers.

Among the clearest glaciological evidence of surging lies an increase in surface flow velocity by an order of magnitude or more. Rare in-situ velocity measurements were collected during surges of Variegated and Trapridge glaciers (Clarke and others, 1984 ; Kamb and others, 1985). Today, remote sensing techniques are favored to detect such velocity changes (Mansell and others, 2012 ; Burgess and others, 2012 ; Dowdeswell and others, 1999 ; Luckman and others, 2002 ; Fischer and others, 2003 ; Joughin and others, 1996). Very rapid changes in length can be indicative although not conclusive of surging, as not all surges result in an advance of the glacier (Braun and others, 2011 ; Mansell and others, 2012). Intense crevassing and sheared margins are commonly detected on actively surging glaciers (Meier & Post, 1969). The evolution of the crevasse pattern can give information the timing of the surge and stress patterns (Dowdeswell & Benham, 2003 ; Hodgkins & Dowdeswell, 1994). Measuring elevation changes can allow identification of surging glaciers, in particular where sudden thickening of the lower reaches of glaciers coincides with up-glacier thinning (Meier & Post, 1969 ; Paterson, 1994 ; Melvold & Hagen, 1998 ; Nuth and others, 2010 ; Bevington & Copland, 2012). Finally, looped moraines, surface foliation and potholes can be identified using systematic visual interpretation of airborne or spaceborne images (Copland and others, 2003, 2011 ; Dowdeswell and others, 1991 ; Hamilton & Dowdeswell, 1996 ; Post, 1969).

During quiescence, identification of surge-type glaciers has to rely on geomorphological features. Surge-type glaciers produce consistent landform-sediment assemblages that cannot be observed on normal glaciers in steady-state (Evans & Rea, 2003). A comprehensive model of the surge landsystem was described by Evans and Rea (2003), based on landforms previously identified by Sharp (1985a, 1985b), Croot (1988a, 1988b), and Knudsen (1995). Detailed mapping of basal ice sequences and of internal structures have been applied to reconstruct past changes in flow dynamics of the Tellbreen glacier in Svalbard (Lovell and others, 2015).

3.4 Controls on the distribution of surge-type glaciers: statistical studies

The unique distribution of surge-type glaciers has motivated a series of studies investigating the connections between surging and a variety of attributes. Post (1969) provided the first qualitative assessment of the role played by various parameters over the distribution of surge-type glaciers in western North America. Successively, univariate and multivariate regression techniques have investigated regional clusters such as the St Elias mountains, Yukon Territory, (Clarke and others, 1986 ; Clarke, 1991), western North America (Wilbur, 1988), Pamirs (Glazyrin and others, 1987) and Svalbard (Hamilton, 1992 ; Hamilton & Dowdeswell, 1996 ; Jiskoot and others, 1998, 2000 ; Atkinson and others, 1998), Iceland (Hayes, 2001), east Greenland (Jiskoot and others, 2003) and Karakoram (Barrand, 2002 ; Barrand & Murray, 2006).

Glacier length, size and morphology have been tested in almost all clusters. In the Yukon Territory, Svalbard and Karakoram, long glaciers are most likely to surge. Area correlates well with surging in the Karakoram and Iceland. In east Greenland, complex glaciers (characterized by long perimeters relative to their size) are most likely to surge. As suggested by Clarke and others (1986) and Jiskoot and others (2000), the tendency for longer glaciers to surge can be related to the increasing vulnerability of the drainage system to instability and collapsing with length. Longer glaciers also tend to spread over several lithological boundaries, further affecting the subglacial drainage system. Glacier length could also be a proxy for mass balance (Budd, 1975 ; Raymond, 1987), hypsometry (Glazyrin and others, 1987 ; Wilbur, 1988), subglacial conditions (Post, 1969 ; Clarke, 1991) or thermal regime (Murray and others, 2000). Statistics on glacier slope produced mixed results. In Svalbard and Yukon, surge-type glaciers tend to have relatively steep slopes, while in eastern Greenland glaciers with a low slope are more likely to surge. Clarke (1991) demonstrated that slope is only a by-product of its inverse relationship with length. Finally, glacier aspect varies tremendously between regions, and could reflect topographic effects on mass balance.

Paterson (1994) noted that the global distribution of surge-type glaciers appears to be confined to new mountain ranges undergoing rapid erosion. Surging glaciers were observed over sedimentary, volcanic and metamorphic lithologies in north-western America. In Svalbard, Hamilton (1992) found that the probability for surging increased for glaciers underlain by sedimentary rocks, although this was the case for 80% of the glaciers in his sample. In the same region, Jiskoot and others (1998) showed that surging was well-correlated with young fine grained sedimentary substrate.

The thermal regime of surge-type glaciers has been comprehensively investigated in Svalbard. Jiskoot and others (2000), based on a sample of 137 glaciers, showed that a polythermal regime was more conductive to surging, as suggested by Bamber (1987) and Macheret and Zhuravlev (1982).

However, the influence of the thermal regime is not clear, but cold-based glaciers have never been observed to surge. The thermal regime could have an effect on the drainage system and the energy balance of a glacier.

Interestingly, climate has never been fully investigated as a potential control on the global distribution of surge-type glaciers. Post (1969) noted that surging glaciers in Alaska and in the Yukon Territory were found across a spectrum of climates from sub-maritime to continental, and concluded that no specific climate conditions are in favor of surging. However, Budd (1975) suggested that the accumulation rate and the bed profile are the primary controls on surging.

3.5 Surge models

Since the first observations of surges at the beginning of the 20th century, theories have been developed to explain the mechanisms behind surging. Field observations and modelling efforts contributed to the development of such models (Clarke, 1976 ; Harrison, 1972 ; Fowler, 1987 ; Clarke and others, 1986 ; Kamb and others, 1985). Today, the theories on surging have considerably improved, although they lack representativeness.

Early theories rapidly recognized surging glaciers as out-of-balance with their environments (De Geer, 1910). Tarr and Martin (1914) and Nielsen (1937) related surging to tectonic activity and volcanism. However, extensive observations by Post (1969) and Thorarinsson (1964, 1969) swiftly contradicted these ideas. The fact that surging had been witnessed on glaciers of all types found in almost all tectonic and climatic environments was the strongest argument against external controls on surging. New models therefore aim to explain surging based on internally triggered instabilities. The main models are described below.

3.5.1. The thermal switch mechanism

The thermal switch or thermal instability mechanism has a long history. Robin (1955) argued that switches in temperature at the bed of glaciers from cold to warm could trigger surges. Below the thickening reservoir zone, feedback mechanisms between increasing shear stresses and ice deformation would progressively bring the base to the pressure melting point, causing a surge. Clarke (1976) dismissed this mechanism as it would yield longer surge cycles than were actually witnessed. Schytt (1969), based on observations of polythermal glaciers in Svalbard, proposed that the cold ring detected along the glaciers margins could act as a dam for meltwater, and that enhanced water pressure in times of strong melting would lead to surging. However, large volumes of water at the cold-warm ice transition could not be detected on Trapridge and Variegated glaciers (Bindschadler, 1997 ; Clarke and others, 1984). Moreover, the surge bulge found on Trapridge glacier was not located at this thermal transition.

Extensive field observations on Bakaninbreen, Svalbard, revived this theory and led to the development of the ‘thermal switch mechanism’ (Murray and others, 2000 ; Fowler and others, 2001). According to this model, surges of polythermal glaciers occur in response to changes at the base from cold to warm. During the quiescence of a cold-based glacier, mass starts to accumulate in the reservoir zone. The longitudinal profile of the glacier is progressively altered, yielding enhanced ice deformation. A positive feedback between ice motion and heat production (caused by strain heating) takes place. Eventually ice reaches the melting point and any excess heat contributes to melting it. Cold ice surrounding the core of warm ice, and the underlying permafrost prevent meltwater from escaping. Rising water pressure reduces basal drag, promoting sliding. The surge is eventually triggered and propagates down-glacier as stresses are transferred from the surge area to the surrounding ice, facilitating ice deformation. Infiltration of water at the ice-bed interface and into the ice significantly contributes to sliding. Surge termination occurs when the water is able to escape, via faults into the ice or through the permafrost.

The application of the thermal switch mechanism is limited to polythermal glaciers. However, Bindeschadler and others (1976) observed that fully temperate glaciers could also surge. This led to the development of a model specific to temperate glaciers: the hydrologic switch mechanism.

3.5.2 The hydrologic switch mechanism

Röthlisberger (1969) first suggested that a switch in the subglacial drainage system, from efficient tunnels to inefficient distributed networks could lead to surging. Thorarinsson (1969) also emphasized the role of water in the surges observed in Iceland.

The development of the hydrologic switch mechanism is mostly based on observations made during the surge of Variegated glacier, Alaska (Kamb and others, 1985). Field measurements revealed the presence of a low-pressure, efficient subglacial drainage system during quiescence, and a much more dispersed flow with high pressures during the surge. Peaks in water pressure corresponded to uplifts and accelerations of the glacier. At surge termination, large volumes of turbid water were released, coinciding with a drop in water pressure and in ice surface.

Kamb and others (1985) suggested that the surge of Variegated was sustained by extensive cavitation over a hard bed and ice bedrock separation, and that its trigger and termination were caused by a switch in the drainage system. Surging would initiate when a conduit system collapses into linked cavities. According to Kamb (1987), high flow velocities and low hydraulic gradient keep the linked cavity system stable. On the other hand, instability might arise when water inputs are important, enlarging the connection between cavities and eventually switching the drainage system to conduits. Although processes of surge termination are clear, this model does not satisfactorily explain how a conduit system initially turns into linked cavities. Finally, the last requirement for a linked cavity system to efficiently cause decoupling of ice from the bed is a large amount of water. Fatland and Lingle (2002) and Lingle and Fatland (2003) suggested that this water could originate from englacial storage in voids, and that a surge could be initiated as soon as this water reaches the subglacial drainage system. The main limitation to Kamb (1987)'s model relates to the bed of the glacier. Cavities require a hard bed, although Richards (1988) detected soft, deformable sediments under Variegated glacier, as it is the case under numerous other temperate glaciers (Harrison & Post, 2003 ; Truffer & Harrison, 2006). A surge model based on a linked cavity system, therefore, may not be appropriate in these cases.

Problems exist with both the hydrological and thermal switch models. Although surges of temperate and polythermal glaciers differ in terms of magnitude and development, they produce similar dynamics. Most interestingly, surges of both glacier types have occurred within the same surge cluster of the St Elias mountains (Frappe-Seneclauze & Clarke, 2007), suggesting an underlying dynamical unity. Further, as stated in Frappe-Seneclauze and Clarke (2007): "*surge models should be able to explain the entire spectrum of surge speed and structure observed within a geographical cluster, irrespective of the thermal regime of the glaciers*." Development of such a model is one of the aims of this thesis.

Chapter 4: Study areas

This thesis investigates surging on a global, regional and local scale. Section 4.1 introduces the surging phenomenon on a global scale. In section 4.2 the surge cluster of Svalbard is presented, along with a sample of six glaciers representative of the spectrum of glacier geometries and thermal regimes found in the region. Finally, glaciers investigated in two case studies are described in section 4.3.

4.1 Global analyses

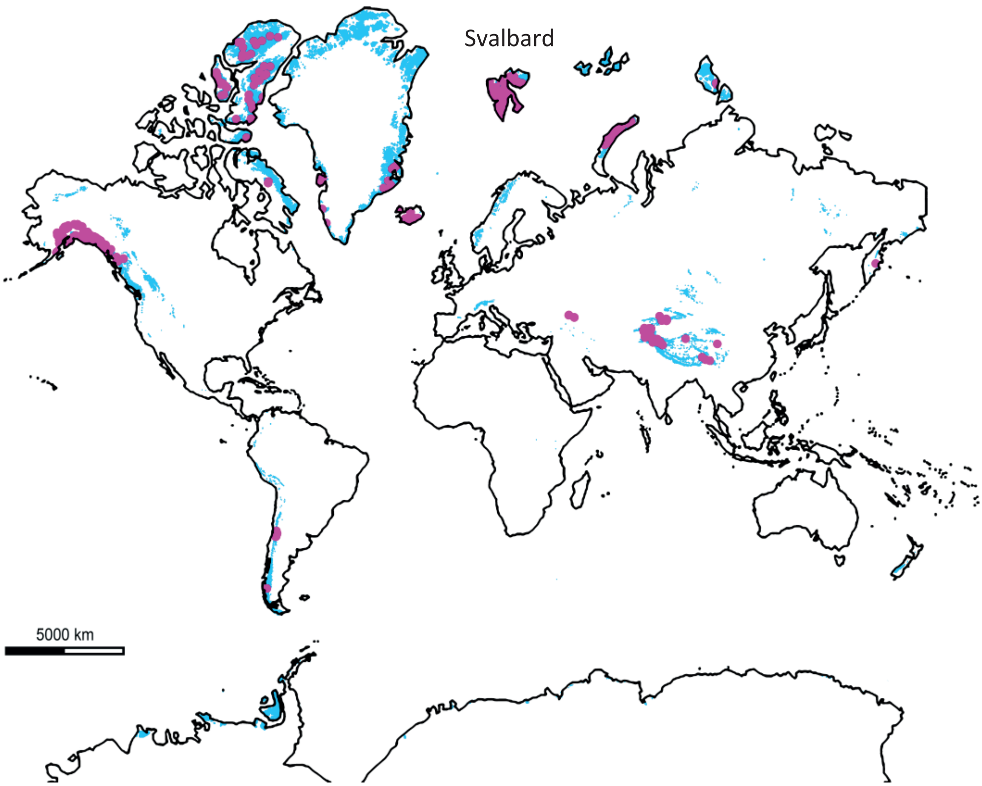

Figure 2: Global distribution of surge-type glaciers (pink ellipses) based on the geodatabase of surge-type glaciers, and normal glaciers (in blue, from the Randolph Glacier Inventory (RGI) v4, Pfeffer and others (2014)).

The global population of surge-type glaciers is estimated to represent about 1% of the total number of glaciers in the world (Jiskoot and others, 1998), which corresponds to a little under 2000 glaciers according to the last estimate of the total population of glaciers from the RGI v4 (Pfeffer and others, 2014) (Fig. 2). Today, most of the main clusters of surge-type glaciers have been investigated by the means of statistical studies, remote sensing techniques or field measurements. To date, studies have focused on individual clusters, and findings related to one cluster might not be transferrable to other regions. Statistical studies for example, have emphasized the differences between surge-type and normal glaciers within clusters instead of focusing on common properties of all surge-type glaciers. As expected, regional and local controls seem to have a strong influence on surging style and magnitude, but region-specific factors may overshadow the global controls on surging. Exploring the controls on the non-random distribution of surge-type glaciers may unlock a new understanding of the surge phenomenon.

Fortunately, a wide range of tools are now available to study surging on a global scale. A century of observations and identifications of surge-type glaciers has built an accurate picture of where surge-type glaciers are found, and where they do not exist. High-resolution datasets allow correlations of surging with glaciological and environmental parameters on a global scale, which was not possible before. The knowledge gathered from the global distribution of surge-type glaciers can then be applied to improve the understanding of regional surge dynamics.

4.2 Regional analyses: the High Arctic archipelago of Svalbard

The Norwegian archipelago of Svalbard lies in the High Arctic between 74 - 81°N and 10 - 35°E. It is composed of four main islands, the largest being Spitsbergen, followed by Nordaustlandet, Edgeøya and Barentsøya. Svalbard is 57% covered by glaciers (Nuth and others, 2013), and a total of 1615 individual glaciers are registered in the RGI v4 (König and others, 2013 ; Pfeffer and others, 2014).

4.2.1 Environmental settings

Located at the confluence of contrasting ocean currents and air masses, Svalbard displays a unique climate sensitivity (Ahlmann, 1953 ; Lamb, 1977). The Arctic Ocean is connected to the North Atlantic Ocean through the Fram Strait, a deep gateway between Greenland and Svalbard. The archipelago lies on an emerged part of the Barents Sea Shelf. Conditions are relatively mild in Svalbard despite its high latitude. The northernmost extremity of the North Atlantic Drift flows along the western coast of Svalbard, and is characterized by warm and high-salinity water, while to the east of the archipelago cold and low-salinity polar water flows south (Humlum and others, 2003 ; Svendsen and others, 2002 ; Saloranta & Svendsen, 2001). There is therefore a strong contrast between the mild marine climate of the south-west of the archipelago and the more Arctic conditions of the north east.

This is reflected in the strong presence of sea ice for most of the year along the east coast, and its absence on the south west coast (with the exception of the formation of fjord ice in winter). Svalbard also lies on one of the major gateways for atmospheric heat and moisture transport in the Arctic Basin, due to its location in the North Atlantic cyclone track (Tsukernik and others, 2007 ; Dickson and others, 2000). Extreme temperature changes on the order of >20°C can occur within a few hours, resulting in occasional rain events, even in the middle of winter.

Svalbard has the longest meteorological record of the Arctic extending from 1911 until today (Førland and others, 1997), representative of the conditions in central Spitsbergen (Nordli and others, 2014). Mean annual temperatures underwent a very strong increase around the 1920s, reaching -5°C at sea level in central Spitsbergen. This marked the end of the Little Ice Age (LIA) in Svalbard. A decrease of about 4°C occurred in the 1960s before the temperature steadily rose, reaching the present values 4 to 5°C higher than at the end of the LIA. From 2004 to 2014, temperatures registered at Longyearbyen airport are the coldest in the month of March (mean = -12°C) and the highest in July (mean = 7.1°C) (eklima.no), while precipitation is at the lowest from February to June (10.5 mm w.eq per month), and at the highest from July to January (20 mm w.eq per month). The weather during the winter season is mainly influenced by the Siberian high, a strong and cold anticyclone, while summer is primarily characterized by low pressure systems passing across the archipelago (Humlum and others, 2003). The mean annual temperature at Longyearbyen airport is -3°C, and the mean annual precipitation is around 195 mm (eklima.no). Precipitation rates are highest towards the east and west coasts (Sand and others, 2003), and decrease from south to north, while the central parts of the archipelago are the driest (Winther and others, 1998).

Geologically, Svalbard spans from Precambrian Heckla Hoek formations to young Tertiary rocks (Hjelle, 1993). The archipelago is crisscrossed by major striking fault zones oriented N-S to NNW-SSE. Permafrost is found almost all over the archipelago with thicknesses between 60-90 m (Humlum and others, 2003) to up to 450 m in high areas with limited snow cover (Liestøl, 1977).

4.2.2 Glacier types and evolution since the Little Ice Age

There is no such thing as a typical Svalbard glacier, as a wide range of glacier types can be found in the archipelago. Cirque and valley glaciers are abundant in central Spitsbergen (Nordenskiöldland and Andrée Land), while ice fields and ice caps make up most of the glacierized area. Ice fields are defined as large ice masses divided into individual glaciers by topography such as mountain ridges and nunataks. Three ice fields cover Spitsbergen in the south-east, north-east and north-west. The two largest ice caps, Austfonna and Vestfonna, make 40% of the glacierized area of Svalbard, and are found in Nordaustlandet. Smaller ice caps are located in the south-east of the archipelago on

Edgeøya and Barentsøya. Both tidewater and land-terminating glaciers co-exist on the archipelago. All tidewater margins are grounded (Dowdeswell, 1989).

In Svalbard, the LIA is thought to have ended at the beginning of the 20th century. Glaciers started retreating in the 1920s as a consequence of an increase in summer temperatures. Between 1936-38 and 1990, glaciers were thinning at low elevations, while the interior of ice caps and ice fields was thickening (except on Prins Karls Forland) (Nuth and others, 2007). The most negative annual geodetic mass balances over this period occurred in the south and in the west (Prins Karls Forland, Nathorstland and Wedel Jarlsberg Land), while the less negative balances were measured on the east coast and central Spitsbergen. Comparisons to the other arctic regions reveal that, considering the changes in glacier area and geodetic mass balance, Svalbard has the most negative balance of the Arctic (Nuth and others, 2010).

The thermal regime of Svalbard glaciers spans from completely cold for small thin land terminating glaciers (Bælum & Benn, 2011 ; Etzelmüller and others, 2000) to polythermal for glaciers typically thicker than 120-130 m (Dowdeswell and others, 1984). Temperate ice is produced in the accumulation zone of glaciers by latent heat release and is advected down-glacier. Strain heating maintains a warm core, while cold ice is found near the surface and at the front of land-terminating glaciers (Macheret & Zhuravlev, 1982 ; Björnsson and others, 1996 ; Bamber, 1987).

Land-terminating glaciers tend to flow at speeds below 10 m per year (Hagen and others, 2003), while larger tidewater glaciers can move much faster, some reach velocities of a couple of meters per day (Lefauconnier and others, 1994). Today, 68% of the glacierized area of Svalbard drains through tidewater glaciers (Nuth and others, 2013).

4.2.3 Surging in Svalbard

A large proportion of the population of glaciers in Svalbard are of surge-type, although the exact number of surge-type glaciers has always been widely debated. Estimations span from 13% to 34% and to 90% of surge-type glaciers on the archipelago (respectively Jiskoot and others (2000), Hamilton & Dowdeswell (1996), Lefauconnier & Hagen (1991)), making Svalbard the region with the highest proportion of surge-type glaciers in the world. The archipelago is therefore one of the best laboratories to study the surging phenomenon.

Svalbard has always been a fertile ground for exploration. Numerous expeditions have made substantial observations on glacier dynamics through the 1800s, and early 1900s. Field notes, maps, drawings or photographs constitute early records that today are crucial to understand the current behavior of glaciers. The very first recorded surges (interpreted later as such) took place on Recherchebreen in c. 1839 (French expedition *La Recherche*), and on Basin 3 of Austfonna between 1850-1878 (Swedish expedition of Adolf Erik Nordenskiøld). Liestøl (1969) and Schytt (1969) realized early that a large concentration of surge-type glaciers can be found on Svalbard.

Surges in Svalbard have often been described as 'muted' or 'sluggish' compared to surges in Alaska (Jiskoot, 1999 ; Dowdeswell and others, 1991). The surge cycle is indeed longer, and velocities experienced during the surge phase tend to be slower. Typically the surge cycle spans from 30 years on Tunabreen (Hagen and others, 1993 ; Blaszczyk and others, 2009) to 500 years for Bråsvellbreen (Dowdeswell and others, 1991). The average surge cycle duration is actually close to 60-70 years for the limited number of glaciers that have surged more than twice on Svalbard. The duration of the surge phase is on average around 6 years (Lefauconnier & Hagen, 1991 ; Hagen and others, 1993) but longer surges up to 12 years in duration are not rare (Nuth and others, 2007 ; Blaszczyk and others, 2009). The long duration of the surge cycle in Svalbard has been attributed to the low accumulation rates on Svalbard glaciers (Dowdeswell and others, 1991).

Liestøl (1969, 1988) gathered evidence of a very large number of surges towards the end of the LIA. The rapidly changing conditions could have caused a wave of surges or surge-like advances. However, evidence is sparse, and some surges could have been confused with a large glacial extent in response to favorable LIA conditions (Lefauconnier & Hagen, 1991). Dowdeswell and others (1995) predicted that, in a context of climate warming causing more negative mass balances, fewer and fewer surges should occur in Svalbard. This is definitely not the case as today in 2015 no less than 15 glaciers are simultaneously surging (pers. Com. Luckman, April 2015).

4.2.4 Thermal regime of Svalbard glaciers

Regional mechanisms of surging were investigated through a sample of six glaciers (Fig. 3):

Von Postbreen (78.45°N; 17.82°E) is a large (168 km^2, König and others (2013)) polythermal surge-type land-terminating glacier draining into Tempelfjorden, central Spitsbergen (Fig. 3). Its last surge dates from 1870 (De Geer, 1910). The glacier has been in quiescence ever since, with both of its streams presently moving at low speeds. One of its tributaries, Bogebreen surged in 1980 (Dowdeswell and others, 1984). In the same size category is Kongsvegen (78.79°N; 13.15°E) another large (108 km^2) surge-type glacier located near Ny Ålesund (Fig. 3). Today, the glacier is partly water-, partly land-terminating. Kongsvegen flows into Kongsfjorden joined to the neighboring tongue of fast-flowing tidewater glacier Kronebreen. Together they are responsible for three large advances in 1800, 1869 and c. 1948 (Lefauconnier, 1987 ; Liestøl, 1988 ; Voigt, 1965 ; Woodward and others, 2002). Only the last surge is confidently attributed to Kongsvegen. Today the glacier is quiescent and moves at velocities between 1.4 to 3.6 m a^{-1} (Melvold & Hagen, 1998b) while retreating at 250 m a^{-1} (Lefauconnier and others, 1994).

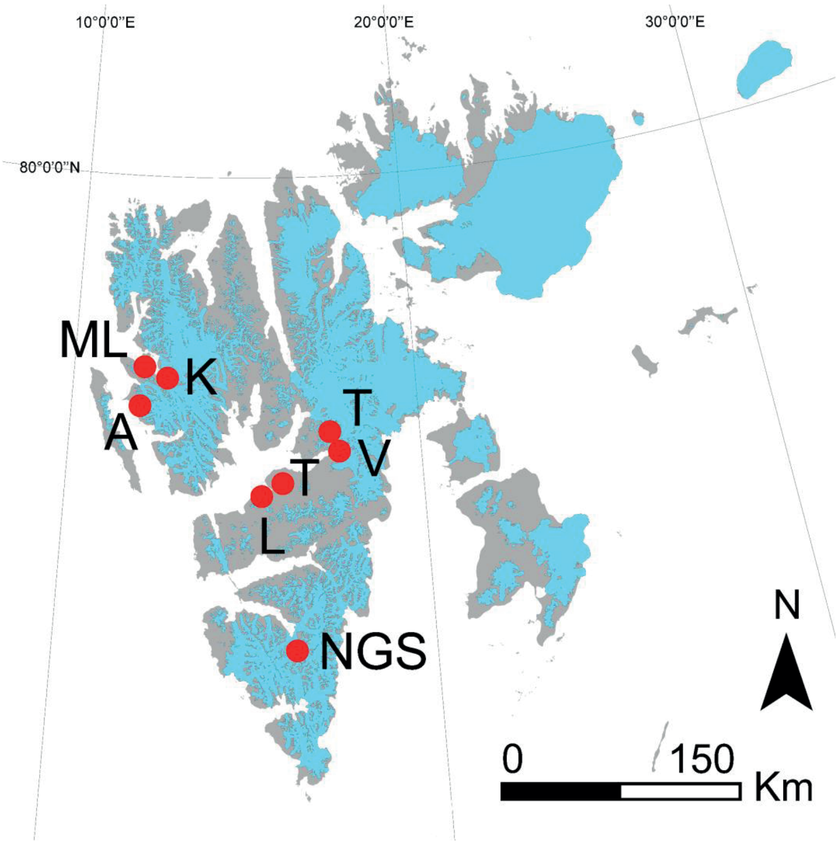

Figure 3: Archipelago of Svalbard and location of studied glaciers. ML: Midtre Lovénbreen, A: Aavatsmarkbreen, K: Kongsvegen, L: Longyearbreen, Te: Tellbreen, Tu: Tunabreen, V: Von Postbreen, NGS: Nathorstbreen glacier system. Glacier outlines from the RGI v4 (Pfeffer and others, 2014).

Tunabreen (78.56°N; 17.61°E) is another large (163 km^2) surge-type tidewater glacier draining Filchnerfonna and Lomonosovfonna into Tempelfjorden (Fig. 3). Its tongue is separated from that of Von Postbreen by an ice-cored moraine. It is one of the few glaciers in Svalbard that has been observed to surge three times since the end of the LIA: in the 1930s, 1970s and from 2003-2005. Today the glacier is quiescent and fast flow velocities occur at the calving front (1 m d^{-1}) (Flink, 2013 ; Flink and others, 2015).

Midtre Lovénbreen (78.88°N; 12.03°E) is one of the most studied glaciers in Svalbard. The glacier is a relatively small (5.2 km^2) land-terminating glacier found close to Ny Ålesund (Fig. 3). It was identified by Liestøl (1988) as a surge-type glacier although other studies have since then disagreed with this classification (Jiskoot and others, 2000 ; King and others, 2008). The glacier is currently retreating and flows at velocities under 7.3 m a^{-1} (Rippin and others, 2005).

Small land-terminating glaciers Tellbreen (78.25°N; 16.17°E) and Longyearbreen (78.17°N; 15.46°E) are located in central Spitsbergen (Fig. 3). Tellbreen is a small (3.9 km^2), thin (<100 m, Bælum & Benn (2011)) cold valley glacier, thinning and slow-moving (<1 m a^{-1}) (AGF212, 2014). No surges have ever been observed on this glacier although recent studies of its basal ice and internal structures have uncovered evidence of faster flow in the past (Lovell and others, 2015). Longyearbreen is slightly smaller (2.9 km^2) but thicker (140 m, Etzelmüller and others (2000)) than Tellbreen. It is flowing between 1 and 4 m a^{-1} and experiencing negative balance (Hagen and others, 2003). Similarly to Tellbreen, Longyearbreen has never been identified as a surge-type glacier (Humlum and others, 2005).

4.3 Individual case studies

The chronology of two surges was reconstructed in detail: the surge of the Nathorstbreen glacier system (NGS) that started in 2008, and the surge of Aavtsmarkbreen (2013-2015) (Fig. 3).

The Nathorstbreen glacier system (77.30°N ; 16.67°E) is a very large glacier (430 km^2) composed of four main branches flowing into Van Keulenfjorden in southern Spitsbergen. A large surge of the glacier system is thought to have occurred in 1870 based on the comparison of maps made during expeditions 1861 and 1905 (Liestøl, 1973). This produced an advance of 12 km into the fjord. Over the last century, the glacier retreated over 17.5 km (Sund and others, 2014). An advance of the combined terminus of the NGS started to occur in 2008-2009, and marked the beginning of one of the largest surges ever recorded in Svalbard since the 1930s (Sund and others, 2014).

Aavatsmarkbreen (78.70°N 12.29°E) is a large (73 km^2) tidewater glacier found on the north west coast of Spitsbergen, south of Ny Ålesund (Fig. 3). The glacier has had two previous advances, one between 1909 and 1936-38 (Niewiarowski, 1982 ; Lankauf, 1999), and in the early 1980s (Lankauf, 1999). The glacier has lost over 1.3 km in length from 1990 and 2006 (Grześ and others, 2008). After a relatively short quiescence of 30 years, the glacier started accelerating in May 2013, and produced a two year long upward propagating surge.

Chapter 5: Material and methods

An overview of the different techniques used in this research is presented in this Chapter. These techniques were applied by the author unless otherwise specified. Section 5.1 describes the process used to compile the global geodatabase on surge-type glaciers, and section 5.2 elaborates on the model used to reproduce the distribution of surge-type glaciers in combination with the geodatabase. Mapping of the thermal regime of glaciers is described in section 5.3. Section 5.4 focuses on the various remote sensing techniques employed, and finally section 5.5 is dedicated to the use of passive seismics in the study of surging glaciers.

5.1 Compilation of the global geodatabase on surging

The first step in the investigation of global controls on the distribution of surge-type glaciers is the compilation of a world-scale geodatabase on surge-type glaciers (Article I). For this work, a combination of 305 peer-reviewed publications, field observations and historical reports were used to inventory known surge-type glaciers, with their precise location and details about their behavior. No new observations were made.

Table 1: Structure of the geodatabase

Table A	**Table B**	**Table C**
Country – Region	Surge ID	Reference ID
RGI ID	RGI ID	Date
Glacier name	Glacier name	Region studied
Latitude (WGS84)	Surge onset date	Complete reference
Longitude (WGS84)	Surge termination date	Data
Area	Evidence	Technique
Length	Study surge index	Years of documentation
Min. elevation	Harmonized surge index	Surge index description
Max. elevation	Area affected by the surge	Surge identification criteria
Slope	Glacier advance	
Aspect	Peak surge velocities	
	Surge interval	
	Quiescence interval	
	References ID	

The geodatabase is divided into three tables (Table 1). All glaciers included in the database are listed in Table A, along with glacier location and a suite of key information on glacier geometry. Tributaries are included as separate entries when their behavior differs from that of the main trunk. Table B compiles all the evidence that led to the identification of a glacier as surge type. In the case of a direct observation of a surge, information on surge timing and magnitude is provided. Finally, all references are provided in Table C. Out of the 2317 glaciers composing the geodatabase, 1148 surges were directly observed. A total of 186 glaciers have been observed to surge more than once.

Integrity of the geodatabase was preserved by evaluating every entry. To be included in the database, glaciers had to display either periodical changes in flow velocity, advances asynchronous to neighboring glaciers or combinations of glaciological and geomorphological evidence. Secondly, a surge-index was implemented in order to rank the quality of the identifications based on the evidence presented.

To allow comparison with normal glaciers, every surge-type glacier is associated with its Randolph Glacier Inventory ID. To do so, a spatial join was performed by combining the centerpoint of surge-type glaciers with glacier outlines of the RGI. As a result, tributaries were merged with their main trunk, resulting in a total of 1430 individual surge-type glaciers. The RGI was first released in February 2012 and can downloaded from: http://www.glims.org/RGI/ (Arendt & others., 2012).

The potential of this geodatabase goes beyond this work as it enables to study the surge phenomenon on the global, regional and individual scale. It also allows spatio-temporal analyses of surge trends, magnitudes, periodicities, as well as inter- and intra-cluster comparisons as all the glaciers are georeferenced.

5.2 Investigating controls on global distribution of surge-type glaciers with Maxent

Finding the controls on the distribution of surge-type glaciers has been a strong focus in glaciology for the past 30 years. The development of statistical tools such as univariate and multivariate regression techniques have enabled the investigation of the relationships between a variety of glaciological and environmental controls and surging (Jiskoot and others, 2000 ; Hamilton & Dowdeswell, 1996 ; Clarke and others, 1986 ; Clarke, 1991 ; Barrand & Murray, 2006). However, controls on the global distribution of surge-type glaciers had not been investigated before this work.

We used the Species Distribution Model Maxent version 3.3.3k (Phillips and others, 2006 ; Phillips & Dudík, 2008) to investigate the global distribution of surge-type glaciers. Maxent aims to maximize entropy of a species in a geographic domain of a species in relation to a set of environmental variables (James and others, 2015). The model is based on two components. First, the presence data (i.e. known locations of the species) defines constraints on the probability distribution. Second, the model uses maximum entropy as a probability distribution to fulfil these constraints. Maxent originally starts by assuming a perfectly uniform probability before drifting from this distribution only to the extent that it is forced to by the constraints (Merow and others, 2013). The selected distribution is therefore only informed by the prior data.

To investigate a species' distribution Maxent requires a dataset of presence locations and a suite of environmental variables (the predicators). The global geodatabase on surge-type glaciers is in this case the list of presence locations. Environmental variables consist of mean annual temperature and mean annual precipitation extracted from ERA-Interim (ERA-I) reanalysis data (European Centre for Medium-Range Weather Forecasts, Dee and others (2011)), glacier area from the RGI v3.2 (Arendt & others., 2012) and length, slope, and range derived from a global dataset of glacier centerlines and the ASTER GDEM v2 (Machguth & Huss, 2014 ; Kienholz and others, 2014). To ensure spatial consistency between climate and geometry datasets, glacier geometry values of all glaciers present in every ERA-I cells were averaged. The same grid was therefore used for ERA-I and glacier geometry data.

Maxent's main output is a map of the probabilities of presence of surge-type glaciers across the glacierized regions. The model output is evaluated using the AUC (area under the receiver operating curve) (Hanley & McNeil, 1982). It discriminates between presence and background points. An AUC of 0.5 qualifies the model as no better than random, while values over 0.5 show a higher predictive power. Two series of runs were performed and compared. First, the model was trained on 75% of the presence data and tested on the other 25%. And second, a 10-fold cross validation was used, testing 10% of the training data at each repetition (Radosavljevic & Anderson, 2014). Importance of the different variables in building the model was analyzed by a Jackknife test (Wu, 1986). Finally, response curves show how climate and glacier geometry variables influence the probabilities of presence.

5.3 Investigation of the thermal regime of surge-type glaciers in Svalbard

Ground penetrating radar (GPR) has been widely used to investigate the thermal regime and to measure the thickness of Svalbard glaciers (Hagen & Sætrang, 1991 ; Björnsson and others, 1996 ; Macheret & Zhuravlev, 1982 ; Macheret and others, 1985 ; Kotlyakov & Macheret, 1987 ; Dowdeswell and others, 1984 ; Bamber, 1987 ; Drewry and others, 1980). The propagation of electronic waves emitted by ground penetrating radar is influenced by the ice dielectric properties. Cold ice is transparent to electromagnetic waves, whereas small volumes of liquid water contained in temperate ice cause strong scattering, making the two easily distinguishable (Navarro & Eisen, 2009). Comparisons of borehole temperature measurements and GPR data in Svalbard showed that a continuous internal reflecting horizon (IRH) corresponds to the cold-temperate transition surface (CTS) (Hagen & Sætrang, 1991 ; Björnsson and others, 1996).

For this work, data were collected on a number of glaciers chosen for their representativeness of the different types of glaciers found in Svalbard: tidewater and land-terminating glaciers, as well as large and thick, small and thin glaciers, quiescent and surging glaciers (Articles II and IV). Here we describe the GPR system and processing sequence used in Article II. A complete Malå system was employed, fitted with “rough terrain” antennae with center frequencies between 25 and 100 MHz. Ideally, higher frequencies (>345 MHz) are most suitable for detecting the CTS (Pettersson, 2005), thus the frequencies we use provide a minimum thickness for the CTS. The Malå system is limited to 267 m in depth, but the focus is to map the CTS, which is generally shallower. Transmitter and receiver are oriented in line, and the configuration of the antennae only allows for common offset surveys. A GPS was connected to the system for positioning of the survey lines in the vertical and horizontal. Typically GPS coordinates were recorded every 5-25 traces. In the field, the system is installed on a snowscooter, and the antenna is pulled as the scooter moves at a constant speed close to or below 20 km h^{-1}.

Data were processed using a combination of two softwares. ReflexW version 7.5 (Sandmeier Scientific Software) was used for the main processing, while Petrel version 6.1.7601 (Schlumberger) allowed 3D interpolation of the results. Pre-processing of the raw data consisted of correcting errors in GPS positioning, time zero adjustment and removal of repeated traces. Main processing steps included Dewow to suppress unwanted low frequencies, dynamic correction and implementation of a constant time increment between all traces. Data were migrated in order to move the reflections and diffractions to their true position. A kirchhoff migration with a velocity of 0.167 m ns^{-1} was used on a bracket of 150 traces. A bandpass butterworth filter was applied to reduce noise caused by the migration, and finally a linear gain function strengthened the signal with depth. Bed and CTS were then manually picked on every single radargram. The software Petrel was then used to interpolate the picked data and provide a unique representation of the thermal structure of the glaciers.

Uncertainties in this technique can be divided into system-specific limitations, data processing errors, and positioning errors. System-specific limitations are related to the vertical and horizontal resolution of the antenna chosen for each survey. Data processing and more particularly manual picking of reflectors introduce uncertainties. Migration quality and cross-over errors were assessed by measuring offsets between the bed reflector of valley-parallel and valley-transverse lines. Finally, errors in positioning were accounted for. In the early surveys (Tellbreen, Von Postbreen, Kongsvegen and Midtre Lovénbreen) a Garmin handheld GPS with an accuracy of 15 m in the horizontal and in the vertical was used, while in the recent surveys a Trimble SPS885 GPS was preferred, characterized by an accuracy of 30 cm both in the horizontal and vertical.

5.4 Remote sensing techniques

A variety of remote sensing techniques were used in this work. They were essential to derive glacier velocities on a high temporal and spatial resolution, measure elevation changes, and map crevasse propagation during surging.

5.4.1 Extracting velocities from feature-tracking of TerraSAR-X images

Feature-tracking of synthetic aperture radar (SAR) images has been widely used to derive glacier surface velocities (Luckman and others, 2006 ; Fallourd and others, 2011 ; Floricioiu and others, 2009). The TerraSAR-X satellite, launched in 2007, provides its own source of illumination making in independent of sun light. X-band (9.65 GHz) SAR images penetrate rain and cloud, allowing year-round monitoring of the Polar Regions. Strength of the backscatter depends on the wavelength, humidity and roughness of the surface.

A. Luckman applied feature-tracking to derive glacier velocities of Aavatsmarkbreen throughout its surge (Article IV). The technique consists of matching patches between repeat pass pairs of images and converting the displacement in velocity (Strozzi and others, 2002 ; Luckman and others, 2006). Glaciers commonly display surface features such as crevasses, debris, making tracking particularly efficient. Errors in the technique arise when the surface features change too significantly for matching. Quality of the results can be assessed by tracking over off-glacier “zero-displacement” areas.

5.4.2 Measuring elevation changes

Differencing digital elevation models (DEMs) consists of subtracting DEMs collected at different epochs. Significant changes in elevation occur during the surge cycle. Measuring these elevation changes enables the detection of patterns of mass accumulation in the reservoir zone prior to surging, and during the surge itself, drawdown up-glacier and thickening of the receiving zone.

C. Nuth employed a large variety of DEMs in Articles III and IV. To investigate the surge of the NGS (Article III), four different DEMs were employed: a DEM from the Norwegian Polar Institute built from aerial photographs collected in 1990 (20 m resolution), a SPOT5 HRS DEM (resolution of 5 m and 10 m in the vertical) from 2008 and ASTER DEMs from 2003 and 2010 (30 m resolution) (Article III). DEM differencing followed the sequence presented in Nuth & Kääb (2011), with co-registration of the datasets to WGS 1984 UTM zone 33, and correction of elevation-dependent biases and instrumentation related errors.

DEMs were also used to detect elevation changes of Aavatsmarkbreen during its surge (Article IV). Glacier longitudinal profiles were extracted from four different DEMs to reconstruct changes in elevation during the two decades prior to surging.

5.4.3 Crevasse mapping on SAR images

As explained above, backscatter strength is affected by the dielectric properties and texture of the reflecting material. During the surge of the NGS (Article III), high backscatter values corresponded to pervasively crevassed ice, as opposed to low backscatter values over smoother ice. A total of 1468 ENVISAT ASAR C-band Wide Swath Mode scenes collected between 1st January 2007 and 31st December 2010 were acquired. Each scene was classified according to backscatter intensity in order to monitor the propagation of highly crevassed ice. Training sets collected in highly crevassed areas displayed an average value of -5 dB. This threshold was used to identify highly crevassed ice on all the scenes. The spatial propagation of crevassing was then analyzed to understand surge onset and its development. Classification was only applied to the lower reaches of the NGS in order to avoid changes in reflection due to the presence of snow and firn. Nevertheless, backscatter values reached a minimum every summer when meltwater was present on the glacier surface. The technique worked best when the glacier surface remained frozen, from the months of October until June. Short drops in winter backscatter values occurred during rain on snow events, while short peaks corresponded to short-lived increases in temperature.

SAR images can also be used for manual mapping of glacier surface features. This technique was implemented in the study the surge of Aavatsmarkbreen (Article IV). In this case, TerraSAR-X images were chosen for the mapping. Their high spatial resolution (2 m) and low penetration of the signal (<1 m in snow) make them particularly suitable for mapping. Three winter scenes were chosen in order to avoid difficulties in observing features due to surface melt.

5.5 Passive seismics

This technique has been applied to study ice quakes emerging during stick-slip events (Weaver & Malone, 1979 ; Deichmann and others, 1979 ; Ekström and others, 2003), iceberg calving (Walter and others, 2010 ; O'Neel & Pfeffer, 2007), and crevasse opening (Mikesell and others, 2012). In Svalbard, a few studies have employed temporary networks of seismometers to study surge mechanisms (Stuart and others, 2005), and calving activity (Köhler and others, 2012 ; Vinogradov and others, 2015 ; Cichowicz, 1983). This technique fits well with the challenging monitoring of surging glaciers, as seismometers can remotely and safely sense activity occurring tens of kilometers away.

Regional seismic networks have numerous advantages as they allow continuous monitoring of large areas over several years. A network of three permanent stations can be found on Spitsbergen, separated by about 100 km. The main station was installed at Longyearbyen in 1992 (SPITS array). It consists of 9 instruments and has an aperture of a kilometer. Two smaller stations with single seismometers are located at Ny Ålesund and Hornsund. The stations started operating respectively in 1994 and 2007 (see Köhler and others (2015) for further description of the network specificities).

The regional network of Spitsbergen can detect tectonic activity, calving, surging-related calving, and most interestingly, surging on land. A large peak of seismic activity was detected from October 2008 to May 2009 in the vicinity of Nathorstbreen glacier system (NGS) (Article III). A. Köhler collected, processed and analyzed the seismics data. The signature of the seismic events were different from that of calving as they displayed no seasonality, lacked low frequencies and originated from one of the tributaries of the NGS.

Data pre-processing commonly consists of selecting a small number of strong seismic events with clear seismic phase arrivals for localization. A total of 96 events were selected from the SPITS array (magnitude >0.5) and from Hornsund station (magnitude >1). Traditional methods of localization of seismic events detected on three or more seismic stations are inadequate for events detected at regional distances. In this instance, event localization was performed by measuring the time delay between P and S wave, calculating the backazimuth (station to source direction) and using the BARENTS3D regional velocity model for wave propagation velocities (NORSAR, 2014 ; Köhler and others, 2015). Errors in localizations were estimated using the mean and standard deviation of backazimuth and distance. Two different techniques were applied to calculate the backazimuth from the Longyearbyen and Hornsund stations, resulting in two different error ellipses.

Local magnitudes of the seismic events were calculated based on a magnitude-amplitude (power law) relation based on regional tectonic events recorded from one of the seismometers in Longyearbyen. The local magnitude was then converted to seismic moment using an empirical relation.

Chapter 6: Summary of articles and key results

Article I: Climatic and geometric controls on the global distribution of surge-type glaciers: implications for a unifying model of surging

Surge-type glaciers represent about 1% of the world's glacier population (Jiskoot and others, 1998). Despite this relatively small proportion, surge-type glaciers have attracted a great deal of attention over the past few decades. Their periodic behavior, alternating between fast and slow flow is one of the largest enigmas in glaciology. The issue of surging goes beyond the fascination for the phenomenon. We cannot be satisfied with our present understanding of glacier flow if we cannot provide a clear explanation as to why some glaciers surge, while most do not. Only an innovative approach could shine a new light on the surge phenomenon. One of the most puzzling facts about surge-type glaciers is their abundance in some regions, while they are completely absent in others. We speculate that the non-random distribution of surge-type glaciers holds the key to a better understanding of the surge phenomenon. In this paper, the role of climate on the global distribution of surge-type glaciers is examined, while glacier geometry is investigated as a possible secondary, local-scale control.

A global inventory of surge-type glaciers is compiled. Climate data were extracted from the ERA-Interim reanalysis dataset (Dee and others, 2011). We focused on two main variables, namely mean annual temperature and mean annual precipitation averaged from January 2000 to December 2009. For every glacier in the world, whether it is normal or of surge-type, mean annual temperature and mean annual precipitation were extracted based on location of the glacier centerpoint. Glacier area was derived from the RGI outlines (Arendt & others., 2012), while length, slope and range were obtained from Machguth and Huss (2014) centerline dataset and the ASTER GDEM v2. Maxent, a species distribution model is employed to reproduce the distribution of surge-type glaciers based on a limited set of variables, and to investigate variable importance in their distribution.

Analysis of the climatic distribution of surge-type glaciers reveals that they are concentrated in a narrow climatic envelope bounded by temperature and precipitation, which we identified as the 'optimal surge envelope' (Fig. 4). Interestingly, there is a near perfect overlap in climatic characteristics of the two surge superclusters (the 'Arctic Ring' and western Central Asia). Surge-type glaciers are longer and have greater areas than normal glaciers across all regions, especially at the cold and dry end of the climatic spectrum (Fig. 6). Maxent accurately reproduced the distribution of the main surge clusters based on four variables: mean annual temperature and mean annual precipitation, and mean glacier length and surface slope (Fig. 9). High probabilities of presence were attributed over the major surge clusters; however more marginal clusters such as the Andes, Caucasus and Arctic Canada are under-predicted.

The results are interpreted in terms of a new enthalpy cycle model. For a glacier to remain close to equilibrium, the glacier must maintain balance velocities, and energy gains must be matched by energy losses. At the cold and dry end of the climatic spectrum, surge-type glaciers are scarce, but are noticeably larger than normal glaciers. On small glaciers, low basal enthalpy production is easily dissipated by heat conduction. Large and thick glaciers have higher basal enthalpy production, and their thickness reduces heat losses to the atmosphere. Therefore, large glaciers are less likely to find a balance between enthalpy production and dissipation. At the warm and wet end of the climatic spectrum, densities of surge-type glaciers are very low. Glaciers in these environments have a high turnover, thus enthalpy production is significant. High runoff evacuates enthalpy gains via efficient drainage networks. Finally, in the optimal surge envelope, enthalpy gains cannot be fully dissipated by enthalpy losses. In temperate glaciers this causes growing storage of water, while in polythermal glaciers, ice is progressively brought to the pressure melting point, and eventually melts, contributing to storage of meltwater.

Glacier geometry exerts a second-order control on enthalpy balance. Large, long and branchy glaciers are higher enthalpy producers than small glaciers. Enthalpy gains are less easily evacuated by conduction on thick glaciers. Drainage systems are more vulnerable to instabilities under long glaciers. This hierarchy in the controls on the distribution of surge-type glaciers is further reflected by Maxent's results. The model predicted the distribution of surge-type glaciers well based on climatic variables only, but failed to do so based solely on glacier geometry variables.

Our results also suggest that many other local-scale controls such as thermal regime, bed substrate, or subglacial topography must be involved in the mechanisms of surging. The enthalpy cycle model redefines surging as a way glaciers respond to their climatic and topographic environments.

Article II: Thermal structure of Svalbard glaciers and implications for thermal switch models of glacier surging

Two main mechanisms are used to explain glacier surging today: the hydrologic switch mechanism for temperate glaciers (Kamb, 1987), and the thermal switch mechanism for polythermal glaciers (Fowler and others, 2001). We test the validity of the thermal switch mechanism for surges in Svalbard. The archipelago is home to a dense population of surge-type glaciers. By comparing published data with new GPR data collected for this work, we measure the evolution of the thermal structure of surge-type glaciers through time. Finally, we reflect on the possibility for glaciers to go in and out of surge cycling by focusing on small, cold, low-activity glaciers which display strong evidence of past fast-flow dynamics.

The thermal regime of six glaciers is investigated. The glaciers were chosen for their representativeness, surge history and availability of previously collected data. Among the large glaciers of our sample, Kongsvegen and Tunabreen are marine-terminating and Von Postbreen is land-terminating. All three have a known surge history and are currently in quiescence. Their thermal structure was mapped in the past (Hagen & Sætrang, 1991 ; Björnsson and others, 1996 ; Bamber, 1987). The three other glaciers are small and land-terminating. Midtre Lovénbreen is known to display a polythermal structure with a narrow core of temperate ice at its base (Björnsson and others, 1996), and has a disputed surge history (Hambrey and others, 2005 ; King and others, 2008). The remaining two, Longyearbreen and Tellbreen are cold and have not previously been identified as surge-type glaciers (Etzelmüller and others, 2000 ; Bælum & Benn, 2011).

The six study sites were revisited between 2009 and 2015 with a Malå GPR system. Antennae with center frequencies between 25 and 100 MHz were used to distinguish cold from temperate ice (Table 1). The main processing of the radar data was achieved using ReflexW (Sandmeier Scientific Software). The same processing sequence was used for bed and CTS mapping (Table 2). Picking of these horizons was done manually. A second software, Petrel (Schlumberger) allowed interpolation of the picked lines, improving the visualization of the thermal structure and thickness of the glaciers. A detailed estimation of the errors related to the GPR system, data localization and data processing is provided.

New data reveals that the large surge-type glaciers of our sample are remain almost fully warm-based during their quiescence. Von Postbreen (Fig. 1) is a typical polythermal glacier of type e while Kongsvegen (Fig. 2) is a variation of type e in which the frozen terminus is removed by calving (Pettersson, 2004). Since previous studies, the shallow cold layer has expanded horizontally up-glacier and has deepened (Björnsson and others, 1996). The tidewater glacier Tunabreen (Fig. 3) is structurally very similar to Kongsvegen although its last surge in 2003 brought the CTS closer to the surface (Bamber, 1987). Midtre Lovénbreen is a polythermal glacier of type e in the process of cooling (Figures 4,5). Finally, Tellbreen and Longyearbreen are entirely cold (Figures 6, 7). Studies have revealed that Tellbreen and Midtre Lovénbreen display evidence of formerly more dynamical flow (Glasser & Hambrey, 2001 ; Hambrey and others, 2005 ; Lovell and others, 2015). On Longyearbreen, we detected a large number of elongated internal reflectors extending from the bed to the surface, which corresponds to arcuate features on the glacier surface (Fig. 8).

Glaciological and geomorphological features show that small glaciers of our sample were more active in the past. We propose that these glaciers underwent thermal cycling (from cold to warm) when the conditions were optimal during the LIA. This resulted in a sudden and temporary dynamical switch to fast flow, similar to the thermal switch mechanism proposed by Fowler and others (2001). Larger glaciers such as Tunabreen and Kongsvegen do not display a frozen terminus, and together with Von Postbreen remain fully warm-based during quiescence. Thermal switches therefore cannot be involved in surge initiation on large and thick glaciers.

We explain the former fast-flow behavior of small glaciers as well as the cyclic surges of large glaciers with the concept of enthalpy cycling (presented in Article 1). Under the present climate, small and thin glaciers are low enthalpy producers. Conduction and runoff of meltwater efficiently balance enthalpy production. However, during the LIA, favorable conditions led to sustained mass accumulation. Enthalpy production rose as the glaciers were thickening. Net enthalpy gains yielded positive feedbacks at the bed, ultimately toppling the glaciers into above balance velocity flow regimes. On the other hand, large glaciers currently undergo periodic surges. As these glaciers are warm-based, enthalpy cycling is manifested by variations in production and drainage of meltwater. The efficiency of conductive heat losses is reduced by the glacier thickness. Studies have shown that glaciers with high altitude accumulation basins still accumulate mass (Nuth and others, 2007). The glacier flow rates must remain below balance velocities to allow build-up of mass. As potential energy is increasing, basal enthalpy production rises, encouraging meltwater production. If meltwater cannot be discharged by efficient drainage networks, pressure builds-up in the system, eventually triggering a surge.

We argue that enthalpy cycling can explain the full spectrum of surging behavior in Svalbard. Our findings also provide a new explanation on the proportion of surge-type glaciers on the archipelago. During the LIA, surges or surge-like instabilities affected a very large number of small glaciers in Svalbard, agreeing with the estimation of Lefauconnier and Hagen (1991) for the period prior to the 20th century. However, under the present conditions, about 20% of all glaciers of Svalbard surge (Article I)

Article III: Seismic detection of a catastrophic glacier surge

At the end of 2008 – early 2009 the permanent regional seismic network of Spitsbergen detected a burst of activity in the vicinity of the Nathorstbreen glacier system (NGS). Signals recorded were significantly different from that of calving or tectonic activity. The NGS was then initiating one of the largest surges recorded on Svalbard since the 1930s (Sund and others, 2014). Interestingly, seismic activity completely ceased three months later, despite continuation of the surge until 2013. In this study, we located the origin of the seismic signals, and with a combination of DEM differencing and mapping crevasse propagation we reconstructed the chronology of events leading to the surge.

The NGS is composed of four glaciers flowing as a combined front into Van Keulenfjorden in southern Spitsbergen (from west to east: Zawadzkibreen (Z), Polakkbreen (P), Nathorstbreen-Ljosfonn (N) and Dobrowolskibreen (D)). The NGS extends from 944 m a.s.l. to sea level (König and others, 2013). The glacier has been in quiescence since its last surge in 1870 (Liestøl, 1973 ; De Geer, 1910). Radio echo-soundings from 1980 show a polythermal regime with mostly temperate ice and a frozen terminus (Dowdeswell and others, 1984).

Records of seismic events were collected from the permanent stations of Longyearbyen (SPITS array, 90 km north of the NGS) and Hornsund (40 km south). The first signals were detected in October 2008, but the main period of activity occurred from early January to May 2009. A total of 14 000 low-magnitude events were detected. Only the 96 of the strongest and clearest events were used for localization, and tributary Z was revealed as the main source of the signals (Fig. 1).

DEM differencing allowed reconstruction of the chronology of events leading to the surge. Between 1990 and 2003 mass was building-up at high elevations on tributaries Z, P and D. From 2003 to 2008 mass was transported downstream on Z and D. The mass transfer stopped a few kilometers from the glacier front, at the limit between temperate and cold ice (Fig. 1).

ENVISAT ASAR C-band wide swath mode scenes were used to map crevasse propagation before and during the surge. In the ablation area of the glacier, when melting did not occur, backscatter strength was mostly dominated by surface texture. Highly crevassed regions were characterized by backscatter values greater than -5 dB. A total of 1468 scenes were classified based on backscatter intensity (Fig. 3). Between October 2008 and January 2009, crevassing rapidly covered the entire lower reaches of D, indicating activation of a surge. During the main period of seismic activity, linear zones of crevassing formed around the frozen terminus, and between the different branches of the NGS. The onset of the surge of the NGS therefore coincided with the fracturing and collapse of the frozen terminus. Crevassing also expanded along the lateral margins of Z and between Z and its two small tributaries Rozyckibreen (R) and Biernawskibreen (B).

Comparison of high-resolution optimal images between September 2008 and July 2009 showed the formation of unusually large chasms in these areas, up to 100 m wide and 50 m deep (Figures 1 and 6). Furthermore, long rifts formed along the margins of Z.

Similarly to the surge of Aavatsmarkbreen (Article IV), we conclude that the surge of the NGS is the result of multi-year processes. The advance of the NGS was eventually triggered by the failure of its frozen tongue. As a result, stresses were suddenly transferred to the margins of Z and regions of high bed strength. Seismic signals originated from the catastrophic collapse of Z, while terminus advance did not lead to detectable seismicity. The absence of signals during the main phase of the surge suggests glide over a weak bed with little energy release. The surge of the NGS is analogous to glacier speed-ups following the collapse of the Larsen ice shelves. Removal of backstress causes sudden glacier acceleration in both cases.

Article IV: A tidewater glacier surge initiated at the terminus: Aavatsmarkbreen, Svalbard

Aavatsmarkbreen is a surge-type glacier found in north-western Spitsbergen (Fig. 1). After three decades of recession and thinning, the glacier underwent a two year-long acceleration reaching velocities nine fold those during quiescence. The speed-up was initiated at the front and propagated to regions higher up-glacier. This behavior relates to other observations of surges of tidewater glaciers in Svalbard such as Osbornebreen (Rolstad and others, 1997), Perseibreen (Dowdeswell & Benham, 2003), Monacobreen and Fridtjovbreen (Murray and others, 2003a, 2003b). In this study, we used a rich data set of glacier velocities, bed topography and surface elevation changes to reconstruct the evolution of the glacier before and during the surge, and investigated what drives the upward propagating surges of tidewater glaciers in Svalbard.

Feature tracking of TerraSAR-X imagery was used to derive the glacier velocity field c. every 11 days from late 2012 to June 2015. The glacier front position during quiescence was mapped on a series of Landsat images. The evolution of the crevasse pattern was studied through high-resolution optical imagery in 2009-2011 and TerraSAR-X SAR images during the surge. Surface elevation profiles were extracted along the glacier centerline from 1990 to 2010, and the 2015 surface profile and bed topography were obtained from a helicopter-borne GPR survey performed in May 2015. The combination of surface longitudinal profiles and bed topography enabled the calculation of the driving stress from 1990 to 2015.

The 2013-2015 surge was preceded by retreat, thinning, and steepening of the lower 4 km of the glacier (Fig. 2). No thickening could be detected prior to the surge, nor could we detect the presence of a surge bulge on the glacier surface. The glacier underwent a progressive acceleration between the early 2000s and the surge. This acceleration was reflected by the apparition of narrow transverse crevasses between 2009 and 2011 (Fig. 5). By the end of 2012 large extensional crevasses had already largely expanded up-glacier (Fig. 6). In May 2013 just as melting was initiated, the glacier terminal zone started to accelerate, and by the end of the meltseason velocities reached 2.5 m d^{-1} (Figures 3,4). While flow stabilized through the following autumn and winter, velocities and crevassing propagated up-glacier. A second step-change in velocities occurred again during the meltseason, this time with a larger impact on the dynamics. The velocities reached 4.5 m d^{-1} at the glacier front, and high velocities could be measured over the lower 9 km of the glacier. A steady decline followed, and by May 2015 the glacier returned to pre-surge velocities (Fig. 7).

We demonstrate that the surge of Aavatsmarkbreen was caused by two processes. First, through the years leading to the surge, sustained steepening progressively increased the driving stress. Strain heating was enhanced at depth, causing the glacier to slowly accelerate and large extensional crevasses to expand up-glacier. The surge was triggered at the onset of the 2013 meltseason. Routing of surface meltwater to the bed through the expanding crevasse field reduced bed strength and facilitated glacier motion. The same mechanism took place the following summer, this time doubling the speeds of the glacier and mobilizing a greater glacier area. At the end of the meltseason velocities steadily decreased through the following year, reflecting slow release of subglacial water.

This surge can be understood in terms of enthalpy production/dissipation (presented in Article I). In the late quiescent phase, strain heating, hence enthalpy increased. Enthalpy production in a polythermal glacier with a warm base is expressed as a rise in meltwater production. Steepening continued, indicating that the glacier kept flowing below its balance velocities. Limited enthalpy production is likely to have been evacuated by an efficient drainage system. However, addition of large volumes of meltwater through crevasses suddenly increased water storage (hence enthalpy) at the glacier bed, and eventually triggered the surge. The behavior of Aavatsmarkbreen reflects the interplay between global and local factors on surging (Articles I and II) and changes in the force balance (Article III).

Chapter 7: Conclusions and future perspectives

Although a small number of glaciers in the world have been observed to surge, the lack of a solid understanding of their behavior uncovers weaknesses in our knowledge of glacier dynamics. For this reason, surge-type glaciers have been regarded as 'abnormal'. They are often excluded from modelling efforts aiming to quantify glacier's response to climate forcing. This thesis addresses some of the essential questions related to surging, and by combining new data with innovative techniques, provides a novel perspective as to why glaciers surge.

This research shows that the non-random distribution of surge-type glaciers is primarily governed by climate. Glacier geometry variables on the other hand, act as a second-order control on surging and allow glaciers to surge outside of the optimal surge envelope. These results led to the introduction of the enthalpy cycle model. Enthalpy cycling can explain why some regions are prone to surging. This concept expresses how glaciers interact with their environment, and unites the behavior of normal and surge-type glaciers under the same framework. The model of enthalpy cycling can be applied to explain surging of both temperate and polythermal glaciers. In Svalbard polythermal glaciers remain warm-based throughout their surge cycle, demonstrating that surging of large glaciers does not occur through thermal switches. However, cold and thin glaciers that today do not surge display strong evidence of past fast flow dynamics that can be explained by thermal cycling during the Little Ice Age. The present behavior of large polythermal surge-type glaciers, and the former dynamics of cold and thin glaciers can also be understood within the enthalpy cycle model. Finally, case studies have shown that surges in Svalbard are the result of multi-year processes. Failure of the frozen terminus of the Nathorstbreen glacier system produced the largest advance recorded in Svalbard since the 1930s, and triggered the catastrophic collapse of one of its tributaries. In turn this produced a three month long period of unusual seismic activity. Sustained steepening of the terminus of tidewater glacier Aavatsmarkbreen produced a chain reaction that led to gradual increase in velocities and expansion of the crevasse field. It is ultimately the transfer of surface meltwater to the bed that triggered a two year long upward propagating surge. Surges in Svalbard are therefore the result of the combination of global controls on surging (climate), and controlled by local factors (thermal regime, geometry).

This new understanding of surge mechanisms and surge behaviors exposes exciting new areas of research. The geodatabase on surge-type glaciers could be used to refine the analyses of regional distributions of these glaciers. Combined to high-resolution climatic data and regional datasets of glacier geometry, bed topography and substrate lithology, regional and local controls on surging could be identified. Field data are scarce and challenging to obtain, especially on surging glaciers, but they are an absolute necessity to further improve our understanding of the surge mechanisms. New technologies such as wireless sensors for example are perfectly suited to monitor glaciers going into surging. Further field studies could be designed to quantify the energy balance of neighboring surge-type and normal glaciers in order to test the enthalpy cycle model. Eventually such field data would provide a strong basis for modelling of glacier thermodynamics.

Chapter 8: References

Abe, T. and M. Furuya 2015. Winter speed-up of quiescent surge-type glaciers in Yukon, Canada. *The Cryosphere*, **9**: 1183-1190.

AGF212 2014. Field Report 2014 - Snow and Ice Processes. The University Centre in Svalbard.

Ahlmann, H.W. 1953. *Glacier variations and climatic fluctuations.* New York, American Geograhical Society.

Alley, R.B., D.D. Blankenship, C.R. Bentley and S.T. Rooney 1986. Deformation of till beneath ice stream B, West Antarctica. *Nature*, **322**: 57-59.

Andreassen, K., M.C.M. Winsborrow, L.R. Bjarnadóttir and D.C. Rüther 2014. Ice stream retreat dynamics inferred from an assemblage of landforms in the northern Barents Sea. *Quaternary Science Reviews*, **92**: 246-257.

Arendt, A.A. and a. others. 2012. Randolph Glacier Inventory - A Dataset of Global Glacier Outlines: Version 3.2. In Space, G.L.I.M.f., *ed.*, Boulder Colorado, USA.

Aschwanden, A. and H. Blatter 2005. Meltwater production due to strain heating in Storglaciären, Sweden. *Journal of Geophysical Research: Earth Surface*, **110**(F4).

Aschwanden, A. and H. Blatter 2009. Mathematical modeling and numerical simulation of polythermal glaciers. *Journal of Geophysical Research: Earth Surface*, **114**(F1).

Atkinson, P., H. Jiskoot, R. Massari and T. Murray 1998. Generalized linear modelling in geomorphology. *Earth Surface Processes and Landforms*, **23**: 1185-1195.

Bamber, J.L. 1987. Internal reflecting horizons in Spitsbergen glaciers. *Annals of Glaciology*, **9**: 5-10.

Barrand, N. 2002. Controls on glacier surging in the Karakoram Himalaya. (Master thesis University of Leeds.)

Barrand, N.E. and T. Murray 2006. Multivariate Controls on the Incidence of Glacier Surging in the Karakoram Himalaya. *Arctic, Antarctic & Alpine Research*, **38**(4): 489-498.

Benn, D. and D.J.A. Evans 2010. *Glaciers and Glaciation, 2nd Edition.* Hodder Arnold.

Bevington, A. and L. Copland 2012. Characteristics of the last five surges of Lowell Glacier, Yukon, Canada, since 1948. *Journal of Glaciology*, **60**(219): 113-123.

Bindeschadler, R., W.D. Harrison, C.F. Raymond and C. Gantet 1976. Thermal regime of a surge-type glacier. *Journal of Glaciology*, **16**(74): 251-259.

Bindschadler, R. 1997. Actively surging West Antarctic ice streams and their response characteristics. *Annals of Glaciology*, **24**: 409-414.

Björnsson, H., Y. Gjessing, S.-E. Hamran, J.O. Hagen, O. Liestøl, F. Pálsson and B. Erlingsson 1996. The thermal regime of sub-polar glaciers mapped bu multi-frequency radio-echo sounding. *Journal of Glaciology*, **42**(140): 23-32.

Björnsson, H., F. Pálsson, O. Sigurðsson, G.E. Flowers, C.F. Raymond and K. Van der Veen 2003. Surges of glaciers in Iceland. *Annals of Glaciology*, **36**: 82-90.

Blaszczyk, M., J.A. Jania and J.O. Hagen 2009. Tidewater glaciers of Svalbard: Recent changes and estimates of calving fluxes. *Polish Polar Research*, **30**(2): 85-142.

Blatter, H. and K. Hutter 1991. Polythermal conditions in Arctic glaciers. *Journal of Glaciology*, **37**(126): 261-269.

Boulton, G.S. 1979. Processes of glacier erosion on different subtrata. *Journal of Glaciology*, **23**: 15-38.

Boulton, G.S. and R.C.A. Hindmarsh 1987. Sediment deformation beneath glaciers: rheology and sedimentological consequences. *Journal of Geophysical Research*, **92**(B9): 9059-9082.

Boulton, G.S., A.S. Jones, K.M. Clayton and M.J. Kenning 1977. A British Ice Sheet model and patterns of glacial erosion and deposition in Britain. In Shotton, F.W., *ed. British Quaternary Studies*, Oxford, Clarendon Press, 231-246.

Braun, M., V.A. Pohjola, R. Pettersson, M. Möller, R. Finkelnburg, U. Falk, D. Scherer and C. Schneider 2011. Changes of glacier frontal positions of Vestfonna (Nordaustlandet, Svalbard). *Geografiska Annaler*, **93**(4): 301-310.
Budd, W.F. 1975. A first model for periodically self-surging glaciers. *Journal of Glaciology*, **14**(70): 3-21.
Burgess, E., R.R. Forster, C.F. Larsen and M. Braun 2012. Surge dynamics on Bering Glacier, Alaska, in 2008-2011. *The Cryosphere*, **6**: 1251-1262.
Burgess, E., C.F. Larsen and R.R. Forster 2013. Summer melt regulates winter glacier flow speeds throughout Alaska. *Geophysical Research Letters*, **40**: 6160-6164.
Bælum, K. and D.I. Benn 2011. Thermal structure and drainage system of a small valley glacier (Tellbreen, Svalbard), investigated by ground penetrating radar. *The Cryosphere*, **5**: 139-149.
Casassa, G., L.E. Espizúa, B. Francou, P. Ribstein, A. Ames and J. Alean 1998. Glaciers in South America. In Haeberli, W., M. Hoelzle and S. Soter, *eds. Into the second century of worldwide glacier monitoring: prospects and strategies.*, Paris, 125-146. (Studies and Reports in Hydrology 56).
Cazenave, A. and G. Le Cozannet 2014. Sea level rise and its coastal impacts. *Earth's Future*, **2**(2): 15-34.
Cichowicz, A. 1983. Icequakes and glacier motion: The Hans glacier, Spitsbergen. *Pure Applied Geophysics*, **121**(1): 27-38.
Citterio, M., F. Paul, A.P. Ahlstrom, H.F. Jepsen and A. Weidick 2009. Remote sensing of glacier change in West Greenland: accounting for the occurrence of surge-type glaciers. *Annals of Glaciology*, **50**(53): 70-80.
Clarke, G.K.C. 1976. Thermal regulation of glacier surging. *Journal of Glaciology*, **16**(74): 231-250.
Clarke, G.K.C. 1987. Fast glacier flow: ice streams, surging, and tidewater glaciers. *Journal of Geophysical Research*, **92**(B9): 8835-8841.
Clarke, G.K.C. 1991. Length, width and slope influences on glacier surging. *Journal of Glaciology*, **37**(126): 236-246.
Clarke, G.K.C., S.G. Collins and D.E. Thompson 1984. Flow, thermal structure, and subglacial conditions of a surge-type glacier. *Canadian Journal of Earth Sciences*, **21**(2): 232-240.
Clarke, G.K.C., J.P. Schmok, C.S.L. Ommanney and S.G. Collins 1986. Characteristics of surge-type glaciers. *Journal Of Geophysical Research*, **91**(B7): 7165-7180.
Cohen, D. 2000. Rheology of ice at the bed of Engabreen, Norway. *Journal of Glaciology*, **44**: 315-325.
Copland, L., S. Pope, M.P. Bishop, J.F. Shroder, P. Clendon, A. Bush, U. Kamp, Y.B. Seong and L.A. Owen 2009. Glacier velocities across the central Karakoram. *Annals of Glaciology*, **50**(52): 41-49.
Copland, L., M.J. Sharp and J.A. Dowdeswell 2003. The distribution and flow characteristics of surge-type glaciers in the Canadian High Arctic. *Annals of Glaciology*, **36**(1): 73-81.
Copland, L., T. Sylvestre, M.P. Bishop, J.F. Shroder, Y.B. Seong, L.A. Owen, A. Bush and U. Kamp 2011. Expanded and Recently Increased Glacier Surging in the Karakoram. *Arctic, Antarctic, and Alpine Research*, **43**(4): 503-516.
Croot, D.G. 1988a. Morphological, structural and mechanical analysis of neoglacial ice-pushed ridges in Iceland. In Croot, D.G., *ed. Glaciotectonics: forms and processes*, Rotterdam, Balkema, 33-47.
Croot, D.G. 1988b. Glaciotectonics and surging glaciers, a correlation based on Vestspitsbergen, Svalbard, Norway. In Croot, D.G., *ed. Glaciotectonics: forms and processes*, Rotterdam, Balkema, 49-61.
Cubasch, U., D. Wuebbles, D. Chen, M.C. Facchini, D. Frame, N. Moahowald and J.-G. Winther 2013. Observations: Atmosphere and Surface. In Stocker, T.F., D. Qin, G.-K. Plattner, M. Tignor, S.K. Allen, J. Boschung, A. Nauels, Y. Xia, V. Bex and P.M. Midgley, *eds. Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*, Cambridge, United Kingdom and New York, NY, USA, Cambride University Press, 159-254.

Das, S., I. Joughin, M.D. Behn, I.M. Howat, M.A. King, D. Lizarralde and M.P. Bhatia 2008. Fracture propagation to the base of the Greenland ice sheet during supraglacial lake drainage. *Science*, **320**: 778-781.
De Geer, G. 1910. Guide de l'excursion au Spitzberg, Stockholm. *XIe Congres Geologique Internationale*, 23.
Dee, D.P., S.M. Uppala, A.J. Simmons, P. Berrisford, P. Poli, S. Kobayashi, U. Andrae, M.A. Balmaseda, G. Balsamo, P. Bauer, P. Bechtold, A.C.M. Beljaars, L. Van de Berg, J. Bidlot, N. Bormann, C. Delsol, R. Dragani, M. Fuentes, A.J. Geer, L. Haimberger, S.B. Healy, H. Hersbach, E.V. Hólm, L. Isaksen, P. Kållberg, M. Köhler, M. Matricardi, A.P. McNally, B.M. Monge-Sanz, J.-J. Morcrette, B.-K. Park, C. Peubey, P.d. Rosnay, C. Tavolato, J.-N. Thépaut and F. Vitart 2011. The ERA-Interim reanalysis: configuration and performance of the data assimilation system. *Quaternly Journal of the Royal Meterological Society*, **137**: 553-597.
Deichmann, N., J. Ansorge and H. Röthlisberger 1979. Observations of glacier seismicity on Unteraargletscher. *Journal of Glaciology*, **23**(89): 409.
Dickson, R.R., T.J. Osborn, J.W. Hurrell, J. Meincke, J. Blindheim, B. Adlandsvik, T. Vinje, G. Alekseev and W. Maslowski 2000. The Arctic Ocean Response to the North Atlantic Oscillation. *Journal of Climate*, **13**: 2671-2696.
Dolgoushin, L.D. and G.B. Osipova 1975. Glacier surges and the problem of their forecasting. *Snow and Ice-Symposium*, Moscow, IAHS-AISH, 292-304.
Dolgoushin, L.D. and G.B. Osipova 1978. Balance of a surging glacier as the basis for forecasting its periodic advances. *Mater. Glyatsiologicheskik Isslef. Khronica Obsuzhdeniya*, **32**: 260-265.
Dowdeswell, J.A. 1989. On the nature of Svalbard icebergs. *Journal of Glaciology*: 224-234.
Dowdeswell, J.A. and T.J. Benham 2003. A surge of Perseibreen, Svalbard, examined using aerial photography and ASTER high resolution satellite imagery. *Polar Research*, **22**(2): 373-383.
Dowdeswell, J.A., D.J. Drewry, O. Liestøl and O. Orheim 1984. Air-borne radio echo sounding of sub polar glaciers in Spitsbergen. *Norsk Polarinstitutt. Srifter*, **182**: 41.
Dowdeswell, J.A., G.S. Hamilton and J.O. Hagen 1991. The duration of the active phase on surge-type glaciers: contrasts between Svalbard and other regions. *Journal of Glaciology*: 388-400.
Dowdeswell, J.A., R. Hodgkins, A.-M. Nuttall, J.O. Hagen and G.S. Hamilton 1995. Mass balance change as a control on the frequency and occurrence of glacier surges in Svalbard, Norwegian High Arctic. *Geophysical Research Letters*, **22**(21): 2909-2912.
Dowdeswell, J.A., B. Unwin, A.M. Nuttalla and D. Wingham, J. 1999. Velocity structure, flow instability and mass flux on a large Arctic ice cap from satellite radar interferometry. *Earth and Planetary Science Letters*, **167**(3-4): 131-140.
Dowdeswell, J.A. and M. Williams 1997. Surge-type glaciers in the Russian High Arctic identified from digital satellite imagery. *Journal of Glaciology*, **43**(145): 489-494.
Drewry, D.J., O. Liestøl, C.S. Neal, O. Orheim and B. Wold 1980. Airborne radio-echo sounding of glaciers in Svalbard. *Polar Record*, **20**: 261-275.
Eisen, O., W.D. Harrison and C.F. Raymond 2001. The surges of Variegated Glacier, Alaska, U.S.A., and their connection to climate and mass balance. *Journal of Glaciology*, **47**(158): 351-358.
Ekström, G., M. Nettles and G.A. Abers 2003. Glacial Earthquakes. *Science*, **302**(5645): 622-624.
Engelhardt, H., N. Humphrey, B. Kamb and M. Fahnestock 1990. Physical conditions at the base of a fast-moving Antarctic ice stream. *Science*, **248**: 57-59.
Espizúa, L.E. 1986. Fluctuations of the Rio del Plomo glaciers. *Geografiska Annaler*, **68(A)**(4): 317-327.
Etzelmüller, B., R.S. Ødegard, G. Vatne, R.S. Mysterud, T. Tonning and J.L. Sollid 2000. Glacier characteristics and sediment transfer system of Longyearbreen and Larsbreen, western Spitsbergen. *Norsk Geografisk Tidsskrift - Norwegian Journal of Geography*, **54**: 157-168.
Evans, D.J.A. and B.R. Rea 2003. Surging glacier landsystem. In Evans, D.J.A., *ed. Glacial landsystems*, Routledge, 544.
Fallourd, R., O. Harant, E. Trouve, J.-M. Nicolas, A. Walpersdorf, J.-L. Mugnier, J. Serafini, D. Rosu, L. Bombrun, G. Vasile, N. Cotte, F. Vernier, F. Tupin, L. Moreau and P. Bolon 2011. Monitoring Temperate Glacier Displacement by Multi-Temporal TerraSAR-X Images and Continuous GPS

Measurements. *IEEE Journal of Selected Topics in Applied Earth Observations and Remote Sensing* **4**(2): 372-386.
Fatland, D.R. and C.S. Lingle 2002. InSAR observations of the 1993-95 Bering Glacier (Alaska, USA) surge and a surge hypothesis. *Journal of Glaciology*, **48**(162): 439-451.
Fischer, A., H. Rott and H. Bjornsson 2003. Observation of recent surges of Vatnajökull, Iceland, by means of ERS SAR interferometry. *Annals of Glaciology*, **37**(1): 69-76.
Flink, A.E. 2013. Dynamics of surging tidewater glaciers in Tempelfjorden Spitsbergen. (Master thesis Stockholm University.)
Flink, A.E., R. Noormets, N. Kirchner, D.I. Benn, A. Luckman and H. Lovell 2015. The evolution of a submarine landform record following recent and multiple surges of Tunabreen glacier, Svalbard. *Quaternary Science Reviews*, **108**: 37-50.
Floricioiu, D., M. Eineder, H. Rott, N. Yague-Martinez and T. Nagler 2009. Surface velocity and variations of outlet glaciers of the Patagonia Icefields by means of TerraSAR-X. *Geoscience and Remote Sensing Symposium, 2009 IEEE International, IGARSS*, Cape Town, IEEE.
Fountain, A., R.W. Jacobel, R. Schlichting and P. Jansson 2005. Fractures as the main pathways of water flow in temperate glaciers. *Nature*, **433**(7026): 618-621.
Fountain, A.G. and J. Walder 1998. Water flow through temperate glaciers. *Reviews of Geophysics*, **36**: 299-328.
Fowler, A.C. 1987. A theory of glacier surges. *Journal of Geophysical Research*, **92**(B9): 9111-9120.
Fowler, A.C., T. Murray and F.S.L. Ng 2001. Thermally controlled glacier surging. *Journal of Glaciology*, **47**(159): 527-538.
Frappe-Seneclauze, T.P. and G.K.C. Clarke 2007. Slow surge of Trapridge glacier, Yukon Territory, Canada. *Journal of Geophysical Research*, **112**(F3).
Førland, E.J., I. Hanssen-Bauer and P.Ø. Nordli 1997. Climate statistics and longterm series of temperature and precipitation at Svalbard and Jan Mayen.
Glasser, N.F. and M.J. Hambrey 2001. Styles of sedimentation beneath Svalbard valley glaciers under changing dynamic and thermal regimes. *Journal of the Geological Society*, **158**: 697-707.
Glazyrin, G.Y., G.M. Kamnyanskiy, A.B. Mazo, V.K. Nozdryukhin and A.N. Salamatin 1987. Mekhanizm podvizhki lednika Abramova v 1972-1975 (Mechanism of the Abramova Glacier surge in 1972-1975). *Mater. Glyatsiol. Issled. Khron. Obsuzhdeniya.*, **60**: 84-90.
Glen, J.W. 1955. The creep of polycrystalline ice.
Grant, K.L., C.R. Stokes and I.S. Evans 2009. Identification and characteristics of surge-type glaciers on Novaya Zemlya, Russian Arctic. *Journal of Glaciology*, **55**(194): 960-972.
Grześ, M., M. Król and I. Sobota 2008. Glacier geometry change in the Forlandsundet area (NW Spitsbergen) using remote sensing data. In Tijm-Reijmer, C.H., *ed. The Dynamics and Mass Budget of Arctic Glaciers. Extended Abstracts*, Obergurgl (Austria). IASC Working group on Arctic Glaciology, 50-52.
Gulley, J. and D.I. Benn 2007. Structural control of englacial drainage systems in Himalayan debris-covered glaciers. *Journal of Glaciology*, **53**: 399-412.
Gulley, J., D.I. Benn, A. Luckman and D. Müller 2009. A cut-and-closure origin for englacial conduits on uncrevassed parts of polythermal glaciers. *Journal of Glaciology*, **55**(89): 66-80.
Hagen, J.O., O. Liestøl, E. Roland and T. Jorgensen 1993. Glacier atlas of Svalbard and Jan Mayen. *Norsk Polarinstitutt. Meddelelser*, **129**.
Hagen, J.O., K. Melvold, F. Pinglot and J.A. Dowdeswell 2003. On the net mass balance of the glaciers and ice caps in Svalbard, Norwegian Arctic. *Arctic, Antarctic and Alpine Research*, **35**(2): 264-270.
Hagen, J.O. and A. Sætrang 1991. Radio-echo soundings of sub-polar glaciers with low-frequency radar. *Polar Research*, **9**(1): 99-107.
Hambrey, M.J., T. Murray, N.F. Glasser, A. Hubbard, B. Hubbard, G. Stuart, S. Hansen and J. Kohler 2005. Structure and changing dynamics of a polythermal valley glacier on a centennial timescale: Midre Lovénbreen, Svalbard. *Journal of Geophysical Research*, **110**(F01006).

Hamilton, G.S. 1992. Investigations of surge-type glaciers in Svalbard. (PhD thesis University of Cambridge.)
Hamilton, G.S. and J.A. Dowdeswell 1996. Controls on glacier surging in Svalbard. *Journal of Glaciology*, **42**(140): 157-168.
Hanley, J.A. and B.J. McNeil 1982. The meaning and use of the Area under a Receiver Operating Characteristic (ROC) curve. *Radiology*, **143**(1): 29-36.
Harrison, W.D. 1972. Reconnaissance of Variegated glacier: thermal regime and surge behaviour. *Journal of Glaciology*, **11**(63): 455-456.
Harrison, W.D. and A.S. Post 2003. How much do we really know about glacier surging? *Annals of Glaciology*, **36**: 1-6.
Hayes, K. 2001. Controls on the incidence of glacier surging. (Unpublished Ph.D. disseration University of Leeds.)
Hewitt, K. 1969. Glacier surges in the Karakoram Himalaya (Central Asia). *Canadian Journal of Earth Sciences*, **6**(4): 1009-1018.
Hewitt, K. 1998. Recent glacier surges in the Karakoram Himalaya, south central Asia. *EOS Electronic Supplement*.
Hewitt, K. 2005. The Karakoram anomaly? Glacier expansion and the 'elevation effect', Karakoram Himalaya. *Mountain Research & Development*: 332-340.
Hjelle, A. 1993. *Geology of Svalbard.* Oslo.
Hodgkins, R. and J.A. Dowdeswell 1994. Tectonic processes in Svalbard tide-water glacier surges: evidence from structural glaciology. *Journal of Glaciology*, **40**(136): 553-560.
Hoinkes, H.C. 1969. Surges of the Vernagtferner in the Otztal Alps since 1599. *Canadian Journal of Earth Sciences*, **6**(4): 853-861.
Hooke, R.L.B. 1981. Flow law for polycrystalline ice is glaciers: comparison of theoretical predictions, laboratory data, and field measurements. *Reviews of Geophysics*, **19**(664-672).
Humlum, O., B. Elberling, A. Hormes, K. Fjordheim, O.H. Hansen and J. Heinemeier 2005. Late-Holocene glacier growth in Svalbard documented by subglacial relict vegetation and living soil microbes. *The Holocene*, **15**(3): 396-407.
Humlum, O., A. Instanes and J.L. Sollid 2003. Permafrost in Svalbard: a review of research history, climatic background and engineering challenges. *Polar Research*, **22**(2): 191-215.
Iken, A. and R.A. Bindschadler 1986. Combined measurements of subglacial water pressure and surface velocity at Findelengletscher, Switzerland: conclusions about drainage system and sliding mechanism. *Journal of Glaciology*, **32**: 101-119.
IPCC 2013. Climate change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change.
James, A.M., C. Burdett, M.J. Mccool, A. Fox and P. Riggs 2015. The geographic distribution and ecological preferences of the American dog tick, Dermacentor variabilis (Say), in the U.S.A. *Medical and Veterinary Entomology*, **29**(2): 178-188.
Jiskoot, H. 1999. Characteristics of surge-type glaciers (Unpublished PhD Thesis). University of Leeds.)
Jiskoot, H., P. Boyle and T. Murray 1998. The incidence of glacier surging in Svalbard: evidence from multivariate statistics. *Computers & Geosciences*, **24**(4): 389-399.
Jiskoot, H., A. Luckman and T. Murray 2002. Controls on surging in East Greenland derived from a new glacier inventory. *NSIDC Report GD-30*: 62-63.
Jiskoot, H., T. Murray and P. Boyle 2000. Controls on the distribution of surge-type glaciers in Svalbard. *Journal of Glaciology*, **46**(154): 412-422.
Jiskoot, H., T. Murray and A. Luckman 2003. Surge potential and drainage-basin characteristics in East Greenland. *Annals of Glaciology*, **36**(1): 142-148.
Joughin, I., S. Tulaczyk, M. Fahnestock and R. Kwok 1996. A mini-surge on the Ryder Glacier, Greenland observed via satellite radar interferometry. *Science*, **274**: 1525-1530.
Kamb, B. 1970. Sliding motion of glaciers: theory and observation. *Reviews of Geophysics and Space Physics*, **8**: 673-728.

Kamb, B. 1987. Glacier surge mechanism based on linked cavity configuration of the basal water conduit system. *Journal of Geophysical Research*, **92**: 9083-9100.
Kamb, B. 1991. Rheological nonlinearity and flow instability in the deforming bed mechanism of Ice Stream motion. *Journal of Geophysical Research*, **96**: 585-595.
Kamb, B. and H. Engelhardt 1987. Waves of accelerated motion in a glacier approaching surge: the mini-surges of Variegated glacier, Alaska, U.S.A. *Journal of Glaciology*, **33**(113): 27-46.
Kamb, B. and E.R. Lachapelle 1964. Direct observation of the mechanism of glacier sliding over bedrock. *Journal of Glaciology*, **5**: 159-172.
Kamb, B., C.F. Raymond, W.D. Harrison, H. Engelhardt, K.A. Echelmeyer, N. Humphrey, M.M. Brugman and T. Pfeffer 1985. Glacier surge mechanism: 1982-1983 surge of the Variegated Glacier, Alaska. *Science*, **227**(4686): 469-479.
Kienholz, C., J.L. Rich, A.A. Arendt and R. Hock 2014. A new method for deriving glacier centerlines applied to glaciers in Alaska and northwest Canada. *The Cryosphere*, **8**: 503-519.
King, E.C., A.M. Smith, T. Murray and G.W. Stuart 2008. Glacier-bed characteristics of midtre Lovénbreen, Svalbard, from high-resolution seismic and radar surveying. *Journal of Glaciology*, **54**(184): 145-156.
Knudsen, O. 1995. Concertina eskers, Brúarjökull, Iceland: an indicator of surge-type glacier behaviour. *Quaternary Science REviews*, **14**: 487-493.
Kotlyakov, V.M. 1996. Variations of Snow and Ice in the past and at present on a Global and Regional Scale. UNESCO.
Kotlyakov, V.M. and Y.Y. Macheret 1987. Radio echo-soundings of sub-polar glaciers in Svalbard: some problems and results of Soviet studies. *Annals of Glaciology*, **9**: 151-159.
Kotlyakov, V.M., G.B. Osipova, D.G. Tsvetkov and J. Jacka 2008. Monitoring surging glaciers of the Pamirs, Central Asia, from space. *Annals of Glaciology*, **48**: 125-134.
Kotlyakov, V.M., O.V. Rototaeva and G.A. Nosenko 2004. The September 2002 Kolka Glacier Catastrophe in North Ossetia, Russian Federation: evidence and analysis. *Mountain Research & Development*, **24**(1): 78-83.
Köhler, A., A. Chapuis, C. Nuth, J. Kohler and C. Weidle 2012. Autonomous detection of calving-related seismicity at Kronebreen, Svalbard. *The Cryosphere*, **6**: 393-406.
Köhler, A., C. Nuth, J. Schweitzer, C. Weidle and S.J. Gibbons 2015. Dynamic glacier activity revealed through passive regional seismic monitoring on Spitsbergen, Svalbard. *Polar Research*, **in press**.
König, M., C. Nuth, J. Kohler, G. Moholdt and R. Pettersen 2013. A digital glacier database for Svalbard. Ch. 10. *Global Land Ice Measurements from Space*, Praxis-Springer.
Lamb, H.H. 1977. *Climate: present, past and future. Vol 2.* London, Methuen.
Lankauf, K.R. 1999. Retreat of the Aavatsmark glacier (Kaffiöyra Region, Oscar II Land, Spitsbergen) in XX Century. In Studies, P.P., *ed. XXVI Polar Symposium*, Lublin, 133-152.
Leclercq, P.W., J. Oerlemans and J.G. Cogley 2011. Estimating the glacier contribution to sea-level rise for the period 1800-2005. *Survey in Geophysics*, **32**: 519-535.
Lefauconnier, B. 1987. Fluctuations glaciaires dans le Kongsfjord, 79°N, Spitsbergen, Svalbard; analyse et consequences. (PhD Universite de Grenoble.)
Lefauconnier, B. and J.O. Hagen 1991. Surging and calving glaciers in eastern Svalbard. *Norsk Polarinstitutt. Meddelelser*: 1-130.
Lefauconnier, B., J.O. Hagen and J.P. Rudant 1994. Flow speed and calving rate of Kongsbreen glacier, Svalbard, using SPOT images. *Polar Research*: 59-65.
Liestøl, O. 1969. Glacier surges in West Spitsbergen. *Canadian Journal of Earth Sciences*, **6**(4): 895-897.
Liestøl, O. 1973. Glaciological work in 1971. *Norsk Polarinstitutt Årbok*: 895-897.
Liestøl, O. 1977. Pingos, springs, and permafrost in Spitsbergen. *Norsk Polarinstitutt Årbok*: 7-29.
Liestøl, O. 1988. The glaciers in the Kongsfjorden area, Spitsbergen. *Norsk Geografisk Tidsskrift - Norwegian Journal of Geography*, **42**(4): 231-238.

Lingle, C.S. and D.R. Fatland 2003. Does englacial water storage drive temperate glacier surges? *Annals of Glaciology*, **36**(1): 14-20.
Lliboutry, L. 1968. General theory of subglacial cavitation and sliding of temperate glaciers. *Journal of Glaciology*, **7**: 21-58.
Lliboutry, L. 1976. Physical processes in temperate glaciers. *Journal of Glaciology*, **16**(74): 151-158.
Lliboutry, L. 1987. Realistic, yet simple bottom boundary condition for glaciers and ice sheets. *Journal of Geophysical Research*, **92**: 9101-9109.
Lliboutry, L. 1993. Internal melting and ice accretion at the bottom of temperate glaciers. *Journal of Glaciology*, **39**: 50-64.
Lovell, H., E.J. Fleming, D. Benn, B. Hubbard, S. Lukas and K. Naegeli 2015. Former dynamic behaviour of a cold-based valley glacier on Svalbard revealed by basal ice and structural glaciology investigations. *Journal of Glaciology*, **in press**.
Luckman, A., T. Murray, R. de Lange and E. Hanna 2006. Rapid and synchronous ice-dynamic changes in East Greenland. *Geophysical Research Letters*, **33**(3).
Luckman, A., T. Murray and T. Strozzi 2002. Surface flow evolution throughout a glacier surge measured by satellite radar interferometry. *Geophysical Research Letters*, **29**(23).
Macheret, Y.Y. and A.B. Zhuravlev 1982. Radio echo-soundings of Svalbard glaciers. *Journal of Glaciology*, **28**(99): 295-314.
Macheret, Y.Y., A.B. Zhuravlev and L.I. Bobrova 1985. Thickness, subglacial relief and volume of Svalbard glaciers based on radio echo-sounding data. *Polar Geography and Geology*, **9**(3).
Machguth, H. and M. Huss 2014. The length of the world's glaciers - a new approach for the global calculation of center lines. *The Cryosphere*, **8**(1741-1755).
Mansell, D., A. Luckman and T. Murray 2012. Dynamics of tidewater surge-type glaciers in northwest Svalbard. *Journal of Glaciology*, **58**(207): 110-118.
Meier, M.F. and A. Post 1969. What are glacier surges? *Canadian Journal of Earth Sciences*, **6**(4): 807-817.
Melvold, K. and J.O. Hagen 1998. Evolution of a surge-type glacier in its quiescent phase: Kongsvegen, Spitsbergen, 1964–95. *Journal of Glaciology*, **44**(147): 394-404.
Merow, C., M.J. Smith and J.A. Silander 2013. A practical guide to MaxEnt for modeling species' distributions: what it does, and why inputs and settings matter. *Ecography*, **36**: 1058-1069.
Mikesell, T.D., K. van Wijk, M.M. Haney, J.H. Bradford, H.P. Marshall and J.T. Harper 2012. Monitoring glacier surface seismicity in time and space using Rayleigh waves. *Journal of Geophysical Research*, **117**(F2).
Murray, T., T.D. James, Y.Y. Macheret, I. Lavrentiev, A. Glazovsky and H. Sykes 2012. Geometric changes in a tidewater glacier in Svalbard during its surge cycle. *Arctic, Antarctic & Alpine Research*, **44**(3): 359-367.
Murray, T., A. Luckman, T. Strozzi and A.-M. Nuttall 2003a. The initiation of glacier surging at Fridtjovbreen, Svalbard. *Annals of Glaciology*, **36**(1): 110-116.
Murray, T., T. Strozzi, A. Luckman, H. Jiskoot and P. Christakos 2003b. Is there a single surge mechanism? Contrasts in dynamics between glacier surges in Svalbard and other regions. *Journal of Geophysical Research*, **108**(B5).
Murray, T., G.W. Stuart, P.J. Miller, J. Woodward, A.M. Smith, P.R. Porter and H. Jiskoot 2000. Glacier surge propagation by thermal evolution at the bed. *Journal of Geophysical Research*, **105**(B6): 13491-13507.
Navarro, F.J. and O. Eisen 2009. Ground-penetrating radar. In Pilleka, P. and W.G. Rees, *eds. Remote sensing of glaciers*, Longon, Taylor & Francis.
Nielsen, N. 1937. A volcano under an ice-cap: Vatnajokull, Iceland, 1934-1936. *Geographical Journal*, **90**(1): 6-23.
Niewiarowski, W. 1982. Morphology of the forefield of the Aavatsmark Glacier (Oscar II Land, NW Spitsbergen) and phases of its formation. *Acta Univ. N. Copernici, Geografia XVI, Toruń*: 147-158.

Nordli, Ø., R. Przybylak, A.E.J. Ogilvie and K. Isaksen 2014. Long-term temperature trends and variability on Spitsbergen: the extended Svalbard Airport temperature series, 1989-2012. *Polar Research*, **33**(21349).

NORSAR 2014. BARENTS3D 2006. A (hybrid) 3D crust and upper mantle model for the greater Barents Sea region. (02/10/14).

Nuth, C., J. Kohler, H.F. Aas, O. Brandt and J.O. Hagen 2007. Glacier geometry and elevation changes on Svalbard (1936–90): a baseline dataset. *Annals of Glaciology*, **46**(1): 106-116.

Nuth, C., J. Kohler, M. König, A. von Deschwanden, A. Kääb, G. Moholdt and R. Pettersson 2013. Decadal changes from a multi-temporal glacier inventory of Svalbard. *The Cryosphere*, **7**: 1603-1621.

Nuth, C. and A. Kääb 2011. Co-registration and bias corrections of satellite elevation data sets for quantifying glacier thickness change. *The Cryosphere*, **5**: 271-2011.

Nuth, C., G. Moholdt, J.O. Hagen and A. Kääb 2010. Svalbard glacier elevation changes and contribution to sea level rise. *Journal of Geophysical Research*, **115**(F01008).

Nye, J.F. 1957. The distribution of stress and velocity in glaciers and ice sheets. *Proceedings of the Royal Society*, **A**(239): 113-133.

Nye, J.F. 1969. A calculation on the sliding of ice over a wavy surface using a newtonian viscous approximation. *Proceedings of the royal Society of London*, **311**(1506): 445-467.

Nye, J.F. 1970. Glacier sliding without cavitation in a linear viscous approximation. *Proceedings of the Royal Society of London*, **A**(315): 1522.

O'Neel, S. and W.T. Pfeffer 2007. Source mechanics for monochromatic icequakes produced during iceberg calving at Columbia Glacier, AK. *Geophysical Research Letters*, **34**(22).

Osipova, G.B., A.S. Tchetinnikov and M.S. Rudak 1998. Catalogue of surging glaciers of Pamir. *Mater Glaciol Investig*, **85**: 3-136.

Paterson, W.S.B. 1994. *The physics of glaciers.* Oxford, New York and Tokyo.

Pettersson, R. 2004. Dynamics of the cold surface layer of polythermal Storglaciären, Sweden. (PhD Stockholm University.)

Pettersson, R. 2005. Frequency dependence of scattering from the cold-temperate transition surface in a polythermal glacier. *Radio Science*, **40**(3).

Pfeffer, T., A.A. Arendt, A. Bliss, T. Bolch, J.G. Cogley, A.S. Gardner, J.O. Hagen, R. Hock, G. Kaser, C. Kienholz, E.S. Miles, G. Moholdt, N. Mölg, F. Paul, V. Radic, P. Rastner, B.H. Raup, J. Rich and M. Sharp 2014. The Randolph Glacier Inventory, a globally complete inventory of glaciers. *Journal of Glaciology*, **60**(221): 537-552.

Phillips, S.J., R.P. Anderson and R.E. Schapire 2006. Maximum entropy modeling of species geographic distributions. *Ecological Modelling*, **190**: 231-259.

Phillips, S.J. and M. Dudík 2008. Modeling of species distributions with Maxent: new extensions and a comprehensive evaluation. *Ecography*, **31**: 161-175.

Post, A. 1969. Distribution of surging glaciers in Western North America. *Journal of Glaciology*, **8**(53): 229-240.

Post, A. and E.R. LaChapelle 1971. *Glacier Ice.* Seattle.

Pritchard, H., T. Murray, A. Luckman, T. Strozzi and S. Barr 2005. Glacier dynamics of Sortebrae, East Greenland, from synthetic aperture radar feature tracking. *Journal of Geophysical Research*, **110**(F03005).

Pritchard, H., T. Murray, T. Strozzi, S. Barr and A. Luckman 2003. Surge-related topographic change of the glacier Sortebrae, East Greenland, derived from synthetic aperture radar interferometry. *Journal of Glaciology*, **49**(166): 381-390.

Quincey, D.j., N.F. Glasser, S.J. Cook and A. Luckman 2015. Heterogeneity in Karakoram glacier surges. *Journal of Geophysical Research*, **120**.

Radosavljevic, A. and R.P. Anderson 2014. Making better Maxent models of species distributions: complexity, overfitting and evaluation. *Journal of Biogeography*, **41**: 629-643.

Raymond, C.F. 1987. How do glaciers surge? A review. *Journal of Geophysical Research*, **92**(B9): 9121-9134.

Richards, M.A. 1988. Seismic evidence for a weak basal layer during the 1982 surge of Variegated glacier, Alaska, U.S.A. *Journal of Glaciology*, **34**(116): 111-120.
Rippin, D., I. Willis and A. Neil 2005. Seasonal patterns of velocity and strain across the tongue of the polythermal glacier midre Lovénbreen, Svalbard. *Annals of Glaciology*, **42**(1): 445-453.
Robin, G.d.Q. 1955. Ice movement and temperature distribution in glaciers and ice sheets. *Journal of Glaciology*, **2**(18): 523-532.
Rolstad, C., J. Amlien, J.O. Hagen and B. Lundén 1997. Visible and near-infrared digital images for determination of ice velocities and surface elevation during a surge on Osbornebreen, a tidewater glacier in Svalbard. *Annals of Glaciology*, **24**: 255-261.
Röthlisberger, H. 1969. Evidence for an ancient glacier surge in the Swiss Alps. *Canadian Journal of Earth Sciences*, **6**(4): 863-865.
Röthlisberger, H. and H. Lang 1987. Glacial hydrology. In Gurnell, A.M. and M.J. Clark, *eds. Glacio-fluvial Sediment Transfer*, New York, Wiley, 207-284.
Saloranta, T.M. and H. Svendsen 2001. Across the Arctic front west of Spitsbergen: high resolution CTD sections from 1998-2000. *Polar Research*, **20**(2): 177-184.
Sand, K., J.-G. Winther, D. Maréchal, O. Bruland and K. Melvold 2003. Regional variations of snow accumulation on Spitsbergen, Svalbard, 1997-99. *Nordic Hydrology*, **34**(1-2): 17-32.
Schytt, V. 1969. Some comments on glacier surges in eastern Svalbard. *Canadian Journal of Earth Sciences*, **6**(4): 867-873.
Sharp, M. 1985a. 'Crevasse-fill' ridges - a landform type characteristic of surging glaciers ? *Geografiska Annaler*, **67A**(213-220).
Sharp, M. 1985b. Sedimentation and stratigraphy at Eyjabakkajökull - an Icelandic surging glacier. *Quaternary Research*, **24**(268-284).
Sharp, M. 1988. Surging glaciers: behaviour and mechanisms. *Progress in Physical Geography*, **12**: 349-370.
Stevens, L.A., M.D. Behn, J.J. Mcguire, S.B. Das, T. Herring, D.E. Shean and M.A. King 2015. Greenland supraglacial lake drainages triggered by hydrologically induced basal slip. *Nature*, **552**: 73-76.
Striberger, J., S. Jbörck, I.Ö. Benediktsson, I. Snowball, C.B. Uvo, O. Ingólfsson and K.H. Kjær 2011. Climatic control on the surge periodicity of an Icelandic outlet glacier. *Journal of Quaternary Science*, **26**(6): 561-565.
Strozzi, T., A. Luckman, T. Murray, U. Wegmuller and C.L. Werner 2002. Glacier motion estimation using SAR offset-tracking procedures. *IEEE Transactions on Geoscience and Remote Sensing*, **40**(11): 2384-2391.
Stuart, G., T. Murray, A. Brisbourne, P. Styles and S. Toon 2005. Seismic emissions from a surging glacier: Bakaninbreen, Svalbard. *Annals of Glaciology*, **42**: 151-157.
Sund, M., T.R. Lauknes and T. Eiken 2014. Surge dynamics in the Nathorstbreen glacier system, Svalbard. *The Cryosphere*, **8**: 623-638.
Svendsen, H., A. Beszczynska-Møller, J.O. Hagen, B. Lefauconnier, V. Tverberg, S. Gerland, J.B. Ørbøk, K. Bischof, C. Papucci, M. Zajaczkowski, R. Azzolini, O. Bruland, C. Wiencke, J.-G. Winther and W. Dallmann 2002. The physical environment of Kongsfjorden-Krossfjorden, an Arctic fjord system in Svalbard. *Polar Research*, **21**(1): 133-166.
Tarr, R.S. and L. Martin 1914. *Alaskan glacier studies.* Washington, D. C.
Thorarinsson, S. 1964. Sudden advance of Vatnajokull outlet glaciers 1930-1964. *Jokull*, **14**: 76-89.
Thorarinsson, S. 1969. Glacier surges in Iceland, with special reference to the surges of Brúarjökull. *Canadian Journal of Earth Sciences*, **6**(4): 875-882.
Truffer, M. and W.D. Harrison 2006. In-situ measurements of till deformation and water pressure. *Journal of Glaciology*, **52**(117): 175-182.
Tsukernik, M., D.N. Kindig and M.C. Serreze 2007. Characteristics of winter cyclone activity in the northern North Atlantic: Insights from observations and regional modeling. *Journal of Geophysical Research*, **112**.

van de Wal, R.S.W., W. Boot, M.R. van den Broeke, C.J.P.P. Smeets, C.H. Reijmer, J.J.A. Donker and J. Oerlemans 2008. Large and rapid melt-induced velocity changes in the ablation zone of the Greenland ice sheet. *Science*, **321**: 111-113.

Vinogradov, Y.A., V.E. Asming, S.V. Baranov, A.V. Fedorov and A.N. Vinogradov 2015. Seismic and infrasonic monitoring of glacier destruction: a pilot experiment on Svalbard. *Seismic Isntruments*, **51**(1): 1-7.

Voigt, U. 1965. Die Bewegung der Gletscherzunge des Kongsvegen (Kingsbay, Westspitsbergen). *Petermanns Geogr. Mitt.*, **109**(1): 1-8.

Walter, F.-., S. O'Neel, D. McNamara, W.T. Pfeffer, B. J.N. and H.A. Fricker 2010. Iceberg calving during transition from grounded to floating ice: Columbia Glacier, Alaska. *Geophysical Research Letters*, **37**(15).

Weaver, C.S. and S.D. Malone 1979. Seismic evidence for discrete glacier motion at the rock-ice interface. *Journal of Glaciology*, **23**(89): 171-184.

Weertman, J. 1957. On the sliding of glaciers. *Journal of Glaciology*, **3**(21): 33-38.

Weertman, J. 1964. The theory of glacier sliding. *Journal of Glaciology*, **5**(287-303).

Weidick, A. 1988. Surging glaciers in Greenland - a status. *Groenlands Geologiske Undersoegelse. Rapport*, **140**(106-110).

Wellman, P. 1982. Surging of Fisher Glacier, Eastern Antarctica: evidence from geomorphology. *Journal of Glaciology*, **28**(98): 23-28.

Wilbur, S.C. 1988. Surging vs nonsurging glaciers: a comparison using morphometry and balance. (M.Sc. University of alaska.)

Winther, J.-G., O. Bruland, K. Sand, Å. Killingtveit and D. Maréchal 1998. Snow accumulation distribution on Spitsbergen, Svalbard, in 1997. *Polar Research*, **17**(2): 155-164.

Woodward, J., T. Murray and A. McCaig 2002. Formation and reorientation of structure in the surge-type glacier Kongsvegen, Svalbard. *Journal of Quaternary Science*, **17**(3): 201-209.

Wu, C.F.J. 1986. Jackknife, Bootstrap and other resampling methods in regression analysis. *The Annals of Statistics*, **14**(4): 1261-1295.

Yafeng, S., M. Desheng, Y. Tandong, Z. Quinzhu and L. Chaohai 2010. Glaciers of Asia, Glaciers of China. U.S. Geological Survey Professional Paper.

Yde, J.C. and N.T. Knudsen 2007. 20th-century glacier fluctuations on Disko Island (Qeqertarsuaq), Greenland. *Journal of Glaciology*, **46**(1): 209-214(6).

Zhang, W.J. 1992. Identification of glaciers with surge characteristics on the Tibetan Plateau. *Annals of Glaciology*, **16**: 168-172.

Zwally, H.J., W. Abdakati, T. Herring, K. Larson, J. Saba and K. Steffen 2002. Surface melt-induced acceleration of Greenland Ice Sheet flow. *Science*, **297**: 218-222.