Sille Marie Myreng

Characteristics and long-term evolution of an englacial meltwater channel in a cold-based glacier, Austre Brøggerbreen, Svalbard

Master thesis in Geography

Trondheim, March 2015

Supervisors:

Associate Professor Geir Vatne, NTNU

Professor Doug Benn, UNIS

Norwegian University of Science and Technology

Faculty of Social Sciences and Technology Management

Department of Geography

The University Centre in Svalbard

Department of Arctic Geology

NTNU – Trondheim
Norwegian University of
Science and Technology

UNIS
The University Centre in Svalbard

Abstract

This thesis aims to describe the long-term evolution of the profile and planar pattern of an englacial channel, and to find the main factors and processes controlling its development. The main topics to be addressed are the development of steps and meanders, and the formation of an englacial channel and a moulin in uncrevassed ice. To achieve this, the available literature on the subject is reviewed, and then applied to the interpretation of a series of four speleological surveys done in the upper part of an englacial channel in Austre Brøggerbreen, Svalbard, in the period between 2002 and 2014. In addition, a total of 6 dye tracer experiments from 2005 and 2014 are analysed to find the flow conditions through the entire englacial system.

The results show an englacial channel that develops toward a concentration of the elevation drop to a few, short sections of the total channel length, separated by long, near-horizontal channel reaches. The meanders grow in wavelength and radius of curvature over time, to accommodate the increase in stream width which in turn comes from a decrease in slope. Both upstream step migration and downward incision of the near-horizontal channel reaches are contributing to the overall channel lowering.

The englacial channel originated from a supraglacial channel, and has entered the englacial environment by the process of cut-and-closure. The differences in slope on the glacier surface is thought to have led one part of the channel to incise down into the englacial environment, while the upstream part with lower gradient remained on the glacier surface. The growing difference in depth between the two channel parts, in combination with the reduced upstream migration of the englacial entrance by the detachment of the jet flow over the entrance step, is believed to be the main processes behind the transition from a gentle sloping channel to a 43 m deep, vertical moulin.

The data series presented in this study holds a great potential for further investigations, but more surveys of better accuracy and higher temporal and spatial resolution are needed to pin down the exact rates and processes of englacial channel pattern development.

Acknowledgements

This thesis came into existence with the aid of many very skilled and very helpful people, who spent hours, days and even weeks out of their working and leisure time to get me through the many challenges I faced in the process. Without them, this could not have been done. I am infinitely grateful for all the effort spent on my project, and will hereby say many, many thanks to the following:

My supervisors, Geir Vatne and Doug Benn, for guidance and discussions, and for sharing so much valuable knowledge and experience.

My field assistants, Ágúst Þor Gunnlaugsson and Guro Skogen Grøndalen, without whom I would be sitting quite helplessly in my office, not being able to do much at all in the form of fieldwork.

Professor Nicholas Hulton, for bravely being the first person to rappel down the moulin and check if it was safe enough for the rest of us to enter in the spring of 2014.

Heidi Sevestre, for invaluable help with logistics, planning, practical challenges and equipment, and for good advices and ever-present good spirits.

Jack Kohler from the Norwegian Polar Institute, for sharing the mass balance data of Austre Brøggerbreen and doing the surface lowering calculations for me, and for good advices and discussions in Ny-Ålesund.

The logistics departments and administrations at the Sverdrup Station and the University Centre in Svalbard, for practical advices, equipment, and for caring about my safety.

And last, but definitely not the least, a huge thank you to the Norwegian Research Council and their Arctic Field Grant, who funded my fieldwork and made it possible for me to run my own research project in the Arctic.

It has been an experience to remember.

Trondheim, 25.03.2015

Sille Marie Myreng

Table of contents

List of figures

List of tables

1 Introduction

The hydrology of glaciers and ice sheets has received increased attention the last couple of decades, as the rising temperatures in glaciated regions give an accompanying increase in meltwater production, which in turn is a major contributor to sea level rise (Gardner et al., 2012; Hanna et al., 2008; Jacob et al., 2012). The contribution is a direct response to increased melt, as the meltwater ultimately drains into the ocean, but increased runoff also affects the sea level rise indirectly by influencing the glacier dynamics. The amount and fluctuations of meltwater draining through the ice to the glacier bed are major influences on the process of basal sliding (Bartholomaus et al., 2008; Pimentel and Flowers, 2011). This is especially important in glaciers terminating in water, where increasing ice velocities dramatically increase the amount of ice delivered to the water by calving (Benn et al., 2007). It is thus of great importance to study meltwater routing through glaciers, alongside the extent and implications of increased meltwater runoff to the ocean.

The mechanisms of water flow through temperate ice have been thoroughly studied (Fountain and Walder, 1998), and more recently also the drainage systems in polythermal and cold-based glaciers (Irvine-Fynn et al., 2011). The knowledge gained from studies on valley glaciers has proven useful in the interpretation of the large-scale hydrology of ice sheets and their outlet glaciers, which are less accessible and are commonly studied with remote sensing techniques (Catania and Neumann, 2010; Jansson, 2010). The advantage of the smaller glacier systems is that they are easier to access with direct study techniques, including channel surveys and flow measurements.

Direct explorations of meltwater channels in and under glaciers have revealed drainage systems with complex and diverse channel paths (Gulley and Benn, 2007; Gulley et al., 2009b; Vatne, 2001). Several mechanisms for how surface channels can enter the englacial environment have been described, related both to crevasses (Benn et al., 2009) and progressive channel incision (Gulley et al., 2009a). It is generally accepted that englacial channels very often run under atmospheric pressure, and that these channels do not tend to flow along the steepest hydraulic gradients within the glacier, as predicted by the theoretical model of Shreve (1972). The paths of englacial channels is now thought to be controlled by englacial tectonic structures (Gulley and Benn, 2007) and hydrodynamic processes (Vatne and Irvine-Fynn, in preparation).

Despite the continuous interest in englacial channels the last decades, only a handful of survey maps have been published to analyse reach-scale channel morphology. Repeated channel surveys that show the channel evolution over time are almost completely absent in the literature, the one notable case being the study of Vatne and Irvine-Fynn (in preparation). In their study an unique survey series from Austre Brøggerbreen is presented, showing the changes in the upper 300 m of an englacial channel over a period of ten years.

A selection of surveys from the same series is used in this thesis, along with one additional survey of the same channel done in 2014. The aim of the thesis is to present a thorough study of the long-term evolution of an englacial channel, and to point out the main factors and processes controlling its development. The main topics to be addressed are the migration rates and behavior of steps and meanders, and the development of an englacial channel and moulin in uncrevassed ice.

To mix the general theory with the dataset and observations from Austre Brøggerbreen in a structured manner, the thesis is divided into three main chapters. The first is a rather extensive theory review chapter, introducing the literature and concepts that are relevant to the study. The second chapter is the case study of the englacial channel in Austre Brøggerbreen, presenting the location, methods, results and data interpretation. In the third chapter, the first two chapters are merged and discussed. In this way, a comprehensive investigation of the development of an englacial channel can be presented.

In this thesis, steps are defined as channel segments where the change in elevation is greater than the change in the concurrent horizontal distance (Vatne and Irvine-Fynn, in preparation), and a channel segment is defined as a part of the channel characterized by having the same profile and planar characteristics compared to their adjacent segments. A channel reach is a long stretch of the channel with a channel gradient markedly different from the immediate upstream and downstream reaches. In order to separate between channel lowering done by the upstream migration of steps, and channel lowering done by the stream incising downwards, the term ‘channel incision’ is used for the latter only. For the combined result of the two processes, ‘channel lowering’ is used.

2 Theoretical background: Characteristics of stream channels in glacier ice

The development and dynamics of stream channels in and on glaciers are parts of an extensive web of processes; that influence and are influenced by the physics of the entire glacier system. To retain some of the overview, this chapter starts with a general introduction to glacier hydrology. Then the field of open channel flow properties and behaviour is briefly visited, to give an understanding of the stream energy and parameters used to describe water flow and channel form. Following this is the full review of the dynamics of ice-walled channels; their hydrological properties as well as the processes by which they develop and the resulting patterns. The last part of the chapter concerns the drainage network of entirely cold glaciers, which with their low rates of movement make a more stable environment for investigation of the long-term development of ice-walled channels.

2.1 Introducing glacier hydrology

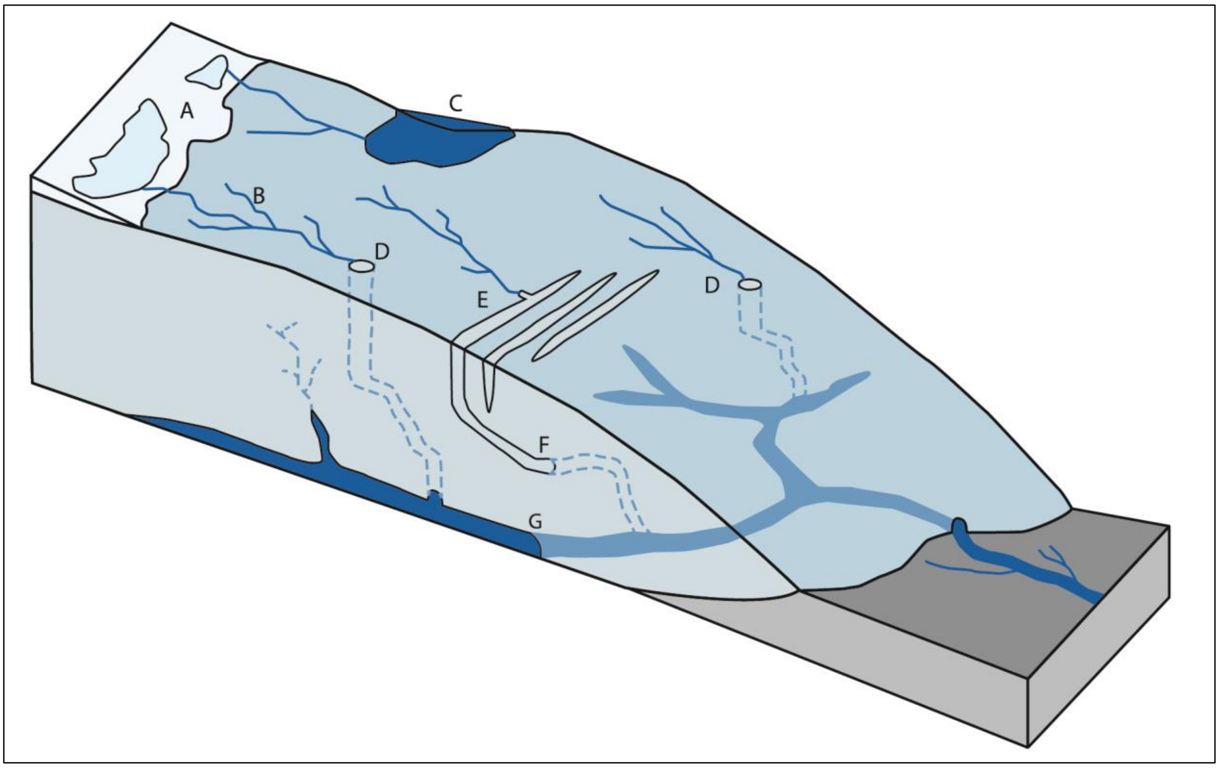

Figure 2.1: The glacier drainage system. A: Swam zones of oversaturated snow. B: Supraglacial channels. C: Supraglacial lakes. D: Moulins. E: Crevasses receiving water. G: Englacial conduits. F: Subglacial tunnels. Modified from Cuffey and Paterson (2010).

The course of water through a glacier follows a seasonal controlled cycle, with temperature being the most important factor. Although there might be water present throughout the year (Hodson et al., 2005), the drainage system is considered to be inactive until the onset of the ablation season in spring. The first rain and meltwater to wet the snowpack in early spring will percolate down till it meets temperatures sufficiently cold to refreeze it and create superimposed ice. The refreezing releases latent heat, warming the surrounding snowpack. When the snowpack has reached a temperature that can no longer refreeze the meltwater, swamp zones of saturated snow will form on the glacier surface (Hagen et al., 1991) (Fig 2.1 A).

After the snowpack has been completely saturated, any excess meltwater will start to trickle downslope, eventually forming supraglacial channels on the glacier surface (Fig 2.1 B). These channels can either converge into a dendritic pattern, or if the slope is even they may run parallel to each other down the glacier (Kostrzewski and Zwolinski, 1995). Depending on the topography and structure of the glacier surface, they can run off to the glacier margins, form supraglacial lakes (Fig 2.1 C) or disappear down into the ice through moulins or crevasses (Fig 2.1 D and E, respectively).

Conduits that are completely enclosed in ice within the glacier are referred to as englacial channels (Fig 2.1 F). These might plunge straight down to the glacier bed, or continue their route downglacier somewhere between the bed and the surface. Eventually most of them reach the bed, forming subglacial channels (Fig 2.1 G). The subglacial drainage system can have many configurations, ranging from thin, continuous water films to subglacial lakes and huge, water-bearing tunnels (Benn and Evans, 2010). It can also be connected to permafrost taliks and groundwater bodies (Irvine-Fynn et al., 2011).

Englacial and subglacial channels will at some point reach the glacier snout, where they leave the ice through a glacier portal. Throughout the ablation season the system grows more efficient, as the water adjusts the channels to accommodate the discharge. By the end of the season, the water flow ceases. When there is no flow to counteract the creep of the ice, the conduits at depth will gradually close. On the glacier surface, the channels fill up with snow. At the onset of the next ablation season the system opens again, sometimes using the old pathways and sometimes creating new.

The water flow on glaciers follows what Kostrzewski and Zwolinski (1995) call a nival, glacial and pluvial regime, meaning that glacial streams are supplied with meltwater from snow, ablation water from ice, and rainwater. This means that the discharge to a very large degree is a function of the ablation on the higher-lying areas of the glacier (Rutter et al., 2011).

The temperature controls on discharge result in an extreme seasonality of the flow regime, reflected in measurements of arctic rivers draining a glacierized catchment. Winter is essentially a dormant phase, spring sees increasing activity, summer encompasses the ablation period and the major part of the annual runoff, and autumn is the phase of decreasing discharge and cessation of continuous flow (Hodgkins et al., 2009). These fluctuations imply that the channels on a glacier on an annual basis will experience an active season with highly fluctuation discharges, and a long period in which there is no discharge at all.

2.2 Flow properties and behaviour in open channels

The flow conditions of both alluvial and ice-walled streams belong in the realm of open channel hydraulics, as they with a few exceptions run under atmospheric pressure. The main characteristics of open channels are turbulence and complex flow patterns. To describe the hydraulic properties of the flow, a range of different parameters can be measured and defined, and a selection of these will be presented in the following section. To explain the behaviour of the flow, it is necessary to understand the sources and the uses of its energy, which determines what kind of work is done and how this shapes the channel morphology. A presentation of the stream energy follows in section 2.2.2.

2.2.1 Parameters of channel geometry and flow conditions

Geometric parameters

The geometric parameters relate to the size and shape of the channel. In any chosen cross-section of the channel a certain set of parameters can be defined, as shown in figure 2.2. Width and depth can be measured directly, and the width/depth ratio, cross-sectional area and wetted perimeter can be calculated from these measurements. Dividing the cross-sectional area by the wetted perimeter gives the hydraulic radius, a measure of channel flow efficiency. A large cross-sectional area with a low wetted perimeter means that there will be less boundaries to retain the flow with friction, thus a high number for hydraulic radius denotes a high efficiency of the stream.

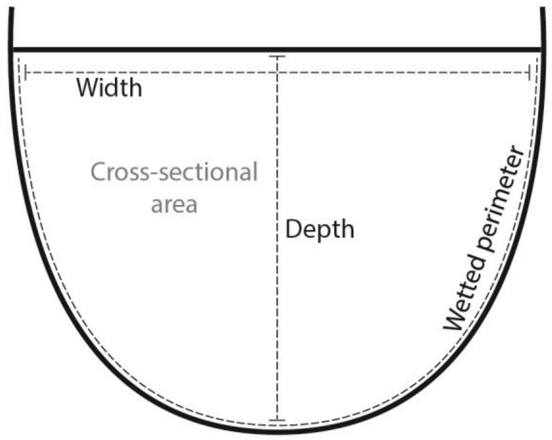

Figure 2.2: Idealized cross-section of a channel, with the main geometric parameters.

For a unit length of channel, two additional parameters can be found. The first one is the slope of the channel, which can be measured directly using a clinometer or the height difference between two points. The second is the Manning roughness, or Manning resistance coefficient. This parameter gives a measure of the delay of the flow by the roughness of the channel boundaries. It can be calculated using hydraulic radius, slope and velocity (Dozier, 1974).

Flow condition and capacity parameters

Velocity is another parameter that can be measured directly. Multiplying the mean velocity of a cross-section with the cross-sectional area gives the discharge. The combination of discharge, slope and the specific weight of water gives a measure of the stream power, which is the potential rate of energy expenditure against the channel boundaries (Bagnold, 1966).

The following two parameters are dimensionless, and show the ratio of inertia force to viscous force and gravitational force, respectively. The first is the Reynolds number, which is used to describe the internal pattern of the flow. It is a function of velocity and hydraulic radius over the molecular viscosity of the stream (Knighton, 1998). Simply put, the value of the Reynolds number tells if the motion of the fluid is mainly dominated by viscous flow, which is smooth, constant and laminar, or inertial forces, which create flow instabilities, or turbulence, in the form of eddies and vortices. Flow at low Reynolds numbers need far less energy to overcome the internal friction than does flow at high numbers, so the highest rates of energy dissipation is found in the latter condition.

The second dimensionless parameter is the Froude number. It describes how fast the flow moves in relation to its weight, or in more descriptive terms, whether the flow is fast and thin, or slow and deep. The Froude number can be written as

$$Fr = \frac{V}{\sqrt{gd}} \tag{1}$$

for open-channel flow, where V is velocity, g is gravitational acceleration, and d is flow depth (Knighton, 1998). Flow where Fr = 1 is termed a critical flow, whereas Fr < 1 and Fr > 1 denotes subcritical and supercritical flow respectively. The wording comes from the stability of the flow of a given number, supercritical flow reacts more to flow disturbances than subcritical flow. When the upstream and downstream influences that control a channel segment are in conflict about whether the flow is required to be supercritical or subcritical, a hydraulic jump is sometimes formed. The jump condition serves as a mechanism for the transition from the supercritical to subcritical, and it also dissipates large amounts of energy through increased turbulence (Munson et al., 2009).

2.2.2 Stream energy

There are three types of energy that are relevant in a stream flow, namely potential, kinetic and thermal energy. The potential energy in a flow unit is proportional its height above the base level, and is transformed to kinetic energy as the flow loses height in the runoff process (Molnar and Ramirez, 1998). A part of this energy is used to overcome internal flow friction in the form of viscous shear and turbulence, and the friction at the channel boundary. Both of these processes cause the energy to be dissipated as thermal energy, or heat (Bagnold, 1966). The energy that remains after the friction is overcome, can be used to erode the boundary and to transport a sediment load (Hooke, 1975). In alluvial channels, this is the energy that does the work along the channel boundaries, and develops the channel morphology. How this development happens is controlled by a combination of the available energy, the existing channel pattern, and the characteristics of the substrate the stream is flowing through.

Flow behaviour and channel adjustment has been explained based on the energy expenditure in the stream, both in cross-sectional geometry and larger-scale channel pattern. Leopold and Maddock (1953) argued that river systems develop toward a state of equilibrium between channel pattern and discharge and sediment load. According to Leopold and Langbein (1962) this state must balance between a uniformly distributed rate of energy loss, and a minimum total work expended in the system.

This involves adjusting the cross-sectional geometry, channel roughness and slope to accommodate the water and sediments supplied to the system, in a way where the energy dissipation is at a minimum (Leopold and Langbein, 1962; Yang and Song, 1979). On a large scale, it also involves an adjustment towards average channel properties where the rate of energy dissipation per unit channel area is close to constant (Molnar and Ramirez, 1998).

The majority of studies on the principles governing stream behaviour are done in alluvial channels and river systems. The different mechanisms operating in ice-walled streams influence both the channel adjustment processes and the equilibrium it is working towards, and their characteristics will be presented in the next subchapter.

2.3 Stream channel dynamics in ice

2.3.1 Glacier ice and the melting process

Stream channels in ice show some striking morphologic similarities to alluvial and bedrock channels. When looking at their dynamics, however, they are fundamentally different. While mechanical erosion and deposition is the main mode of alluvial and bedrock channel evolution (Knighton, 1998), and dissolution is the process of karst system formation (Ford and Williams, 2007), stream channels in ice develop by melting. And while alluvial and bedrock channels evolve at rates measured by decades and centuries, ice channel migration can be measured in centimetres per day.

In alluvial and bedrock channels, it is the kinetic energy that works on the channel boundaries. Potential energy that transforms into thermal energy has no impact on mechanical erosion. In ice-walled channels, it is the thermal energy that does the work. Though mechanical erosion of ice crystals along the channel wall has been mentioned as a factor for stream channel development (Ferguson, 1973), thermal erosion is by far the most active agent in this environment (Jarosch and Gudmundsson, 2012; Karlstrom et al., 2013).

Water running on ice is basically a fluid flowing on its solid state, a temperature controlled condition (Spring and Hutter, 1982). This means that the temperature difference between the fluid and the solid parts of the system is of major importance for whether the fluid turns solid or the other way around in the boundary zone. The on-freeze of water is in most cases restricted to the cold periods with little or no discharge. A freeze-related process analogous to deposition is negligible, as the eroded material melts and becomes part of the water flow (Dozier, 1974). Ice-walled channel morphology is thus a result of erosional processes, with no depositional features.

The water in ice-walled channels holds a temperature of roughly 0,1-0,4°C, slightly warmer than the ice boundary that usually stays close to 0°C (Isenko et al., 2005). This leads to a heat transfer from water to ice along the channel boundary. The melt rate at a given cross section of a channel can be written as

$$\bar{m} = \frac{\rho_w g(\beta + \gamma)Q}{\rho_i LP} \tag{2}$$

(Jarosch and Gudmundsson, 2012)

where ρ_w and ρ_i is the density of water and ice, respectively; g is gravitational acceleration, β is slope; Q is discharge; L is the latent heat of melting for ice; and P is the wetted perimeter. The symbol γ denotes meltwater temperature loss to the ice for a unit distance along the channel, which is proportional to the temperature difference between water and ice (Isenko et al., 2005). The equation shows that the energy available to melt the ice is directly proportional to discharge and channel slope, and that the amount of ice melted is controlled by surface over which this energy is divided.

An additional factor in ice-walled channels, which is not found in alluvium and bedrock, is the creep of the ice. The ice masses of a glacier move by deformation along the grain boundaries, the rate of movement depending on ice temperature and overburden pressure (Irvine-Fynn et al., 2011). Ice creep leads to channel closure, working against the incision done by melting. The dependence on overburden pressure means that it will be more effective at depth than on the surface, but it has been shown to counteract the incision of the channel floor in supraglacial channels in temperate ice (Jarosch and Gudmundsson, 2012). In any case, the incision rate of a channel must exceed the creep closure rate of the ice if the channel is to remain open.

2.3.2 Water temperature and energy flux

The exchange of thermal energy between a flowing liquid and its solid boundary will always be proportional to the temperature difference ΔT and the water flow velocity *v*. The total heat flow through the ice-water boundary over a set time is written by Isenko et al. (2005) as

$$dE_i = Bv\Delta TS_{wall}dt. \tag{3}$$

Where S_{wall} is the area of the ice-water boundary; dt is the time period; and B is the heat transfer coefficient. For turbulent water at 0°C this coefficient amounts to $2{,}64 \cdot 10^3$ J per m^{-2} per kelvin degree of difference between the fluid and the boundary. This means that water at higher temperatures will transfer more heat, faster, than water at temperatures close to that of the boundary.

The water in ice-walled channels is mainly meltwater, coming from the melting of snow and ice within the stream catchment. This keeps it close to 0°C from the source, but there are several factors that influence the temperature along the flow path.

On the glacier surface, solar radiation is the biggest influence on water temperature. Stock and Pinchak (1995) report albedo of 44 % over a glacial stream and 13 % over the adjacent ice surface, showing that the meltwater absorbs a considerably higher amount of the incoming shortwave radiation. Energy that would otherwise be reflected back into the atmosphere stays in the stream, and is used both for direct ablation of the channel bed and to warm the stream water (Dozier, 1976; Isenko et al., 2005). If the channel bed is covered by sediments, the amount of energy absorption increases even further.

Following the law of conservation of energy, the amount of potential energy lost by the water as it moves downward has to transform into kinetic or thermal energy. In open channels, part of this energy dissipates as thermal energy inside the flow, causing the water temperature to increase. According to Isenko et al. (2005), the loss of potential energy for a volume of water is sufficient to warm it by 0,2°C per 100 m of lowering. However, an actual warming of the stream water was not found in their study or in the studies of Dozier (1974; 1976), who also failed to detect any significant increase in kinetic energy. The conclusion of all three studies was that the energy went to thermal erosion along the ice boundary.

In channels that are cut off from atmospheric influences, the water flow will at some point reach an equilibrium temperature where the heat gained from loss of potential energy will equal the heat lost to the ice boundary (Isenko et al., 2005). At this point the temperature difference between water and ice is at its minimum, giving the lowest possible energy transfer through the channel boundary and the accompanying low incision rates.

2.3.3 Incision rates

With the mechanisms and flow characteristics described above, it is possible to get a fairly good idea of how thermal erosion works in an ice-walled channel. Incision rates have been measured and modelled with consistent results for supraglacial channels, the resulting values showing clear relationships to water temperature, discharge and slope. Open channels incise both sideways and downwards, the ratio between the two depending on ice properties, channel pattern and sediment cover on the bed (Isenko et al., 2005; Kostrzewski and Zwolinski, 1995).

Marston (1983) measured vertical incision rates up to 8 cm/day on a cold-based glacier in Alaska, while Gulley et al. (2009a) measured 4,1 cm/day for a channel that was deep enough to be sheltered from solar radiation. These values reflect how incision rates are influenced by direct stream bed ablation in channels where the incoming shortwave radiation reaches the bed. Isenko et al. (2005) spread the measurements over the day. They measured rates of 3-4 cm/day in the morning, and 7-10 cm/day in afternoons.

The cross-section measurements done by Marston (1983) showed that the stream in straight channel reaches cuts downwards more easily than sideways. Lateral erosion increases somewhat with increased discharge, but most of the stream adjustment in straight reaches is done by downwards incision and increased velocity (Brykala, 1999). In meandering reaches the lateral incision is expected to be more dominant, but detail measurements done by Müller (2007) show that the incision is mostly vertical also in meanders.

If it is possible to talk about equilibrium in an ice-walled channel at all, it is clear that reaction and relaxation times must be considerably shorter than in bedrock and alluvial environments. Rapid discharge fluctuations dominate their flow conditions, but ice-walled channels also have a readily deformable boundary and a high unit stream power to promote rapid adjustment (Knighton, 1985). The streams seems to adjust to the maximum discharge, and creates patterns similar to those found in bedrock and alluvial streams (Knighton, 1981; Marston, 1983).

2.3.4 Migration patterns

Meandering

Meandering in alluvial streams has been explained with a model coupling between flow field, bed topography in the form of point bars, and bedload transport (Johannesson and Parker, 1989). Seminara (2006) also suggests, in more general terms, that the mobile interface between the solid boundary and the fluid is the unstable part that causes meandering in a stream channel. In ice-walled streams there is no such interface. The bed is without mobile bedforms, as bedload transport is absent.

Studies of meandering in ice-walled streams have concluded that sinuosity is developed by differential melting, which arises because of flow instabilities. The analysis done in the study of Parker (1975) indicated that supercritical flow is needed for the flow instabilities that initiate meandering, while Karlstrom et al. (2013) found meandering to start already at Froude numbers > ~0,4. Both studies used an initial channel with small perturbations that grow into meanders over time, but where Parker (1975) used perturbations at the bed, the study of Karlstrom et al. (2013) used perturbations in the channel centreline. Thus it seems that for meandering to develop in ice-walled channels, independent of the water flow conditions, there must be some feature in the channel morphology that disturbs the flow enough to initiate the differential melting. In the environment of ice-walled streams, these disturbances are present as stratified and foliated ice, giving ice types with different properties and differential erosion rates (Hambrey, 1977).

Once the meandering has started, it intensifies the differential melting and enhances its own pattern. Figure 2.3 shows the flow pattern in a meander bend, explaining how the bends get deeper along the outside edges and shallower at the inside. This outer overdeepening, combined with general surface lowering by solar forcing, sets the balance between lateral and vertical incision in supraglacial streams (Karlstrom et al., 2013). The result is a meandering pattern similar to the one created by erosion and deposition of sediments in alluvial channels. And like alluvial channels, ice-walled has been observed and modelled to migrate downstream over time (Dozier, 1976; Karlstrom et al., 2013; Marston, 1983).

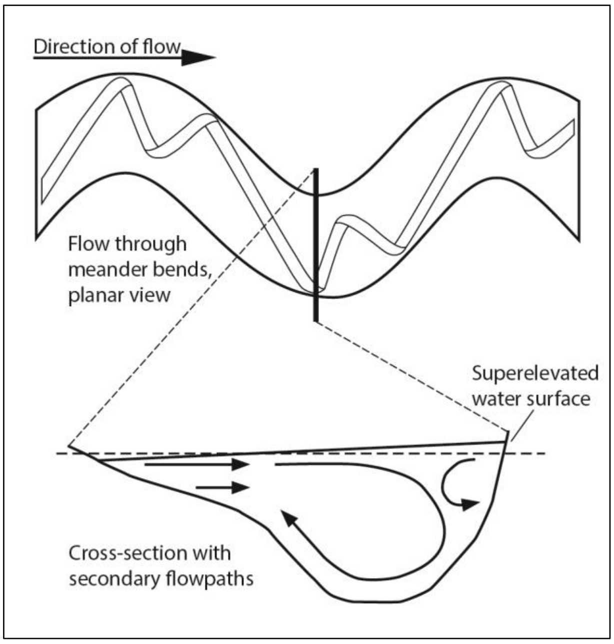

Figure 2.3: Idealized planar and cross-sectional flow pattern in a meander bend. The main current is forced down as it hits the channel bank and rises again as it hits the shallow inner part of the bend. The corkscrew form of the flow increases erosion at the outer bank, while the reduced flow by the inner bank creates a shallow area. The downwards movement of the flow also creates eddies in the outer part of the curve, elevating the water level in relation to the inner part. The figure is based on Leopold and Wolman (1960) and Knighton (1998)

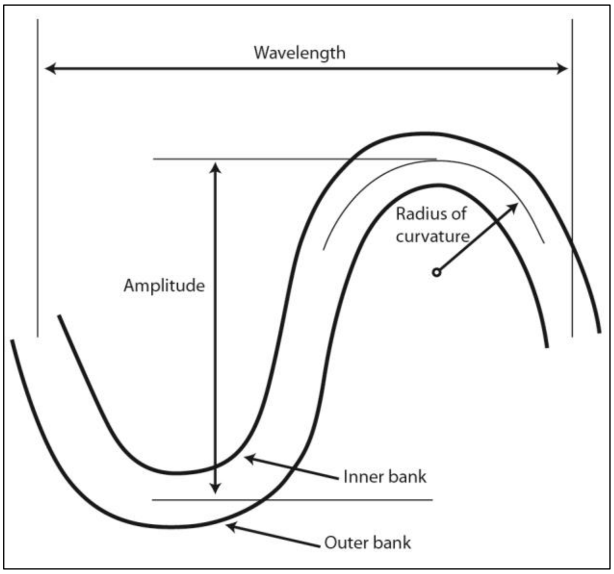

Figure 2.4: The terms used to describe meander geometry.

Figure 2.4 shows the terms used to describe the geometry of a meander. In sediment-free, ice-walled channels, the wavelengths are consistent with those measured in alluvial and bedrock streams, showing a power law relationship between wavelength and stream width (Karlstrom et al., 2013; Leopold and Wolman, 1960). The many similarities between meandering in ice, alluvium and bedrock indicate that there might be a more central, hydrodynamic reason for the pattern independent of the substrate.

In general fluvial morphology, Knighton (1998) presents meandering as a stable channel pattern, suggesting that streams meander to stabilize themselves over longer stretches. Meandering gives a gentler channel slope where the gradients are high, and can also give a more even slope where the gradients are varying (Langbein and Leopold, 1966). Meanders can prove stabilizing also in the light of energy dissipation, as they have been shown to have smaller variations in friction and shear stress than straight channel reaches of the same magnitude (Dozier, 1976).

Step-pools

If meandering is a way for streams to stabilize in a planar pattern, the formation of steps and pools can be said to be the equivalent in the longitudinal profile. The step-pool pattern has a staircase-like profile, consisting of steep channel segments (steps) alternating with channel-spanning plunge pools. In alluvial settings the steps can consist of cobble or boulder chains, or woody debris where it is available (Church and Zimmermann, 2007).

With the characteristic form of steps and pools comes a specific flow pattern, consisting of a supercritical flow over the lip of the step, a free-falling or sloping jet flow down the step, and a turbulent, subcritical flow with the accompanying hydraulic jump in the plunge pool (Church and Zimmermann, 2007). The steps give high elevation drops while the plunge pools function as local energy dissipaters, giving a relatively stable pattern for steep streams.

Knighton (1981) described a quasi-regular alternation of steep convex elements and plunge pools in his survey of supraglacial streams on a mostly stagnant glacier on the Norwegian mainland. The channels followed the gradient of the glacier, with a step-pool pattern. In englacial channel systems the formation and migration of steps are significant factors for channel evolution. Vatne and Refsnes (2003) found steps to be the main form of elevation loss in the ~300 m of surveyed channel, with a mean step height of 5,5 m. The maximum channel lowering through one ablation season was 7,9 m, a feat that is only possible because the step downstream of the measured reach had an upstream migration of 5 m, through the measured point. The same phenomenon was measured by Gulley et al. (2009a), with rates of channel lowering way above average at cross-sections where a step had migrated through the measurement point.

As steps are significantly steeper than their adjacent channel segments, melt rate equation presented in section 2.3.1 predicts that more melting will take place over the step and cause it to move further upstream. The downstream pool will follow in its wake, creating a near horizontal channel stretch that will remain at more or less constant gradient until the next downstream step migrates through it (Vatne and Irvine-Fynn, in preparation). The pattern of tall steps and long stretches of near horizontal channel segments is evident in the surveys of Vatne (2001) and Gulley (2009). The speed of the step migration has been shown to vary, so that some steps will catch up with upstream steps and combine into a taller step. This has been hypothesized to happen because taller steps migrates faster than shorter steps (Vatne and Refsnes, 2003), but no proof of this has been presented. Vatne and Irvine-Fynn (in preparation) suggested that the shape of the step, rather than the height, can be of significance for the migration rate.

2.4 The drainage system of cold-based glaciers

2.4.1 General characteristics

The thermal regime of a glacier is of great importance to its drainage system. From being a dynamic environment with crevasses, pressurized englacial conduits and extensive subglacial networks on a temperate glacier, the drainage on a cold-based glacier goes towards the end of the spectrum where the ice is a passive, static substrate.

The internal deformation rate of ice is closely connected to temperature and pressure (Hooke, 1981; Nye, 1953). Cold, thin ice deforms very slow compared to ice near the pressure melting point, giving creep rates so low that cavities within the ice can keep open for decades. Cold-based glaciers are also frozen to their bed, lacking basal gliding and the velocity which is necessary to develop crevasses. Without this shortcut between the surface and the bed, the meltwater streams are left to make their own routes over and through the ice.

Cold-based glaciers normally show supraglacial channels running over a large part of the ice, either running off the sides to the moraine area or cutting their way into the ice to become englacial or subglacial channels (Gulley et al., 2009a). The slow movement also allows englacial and subglacial streams to run in atmospheric pressure and remain open through several years, as the internal creep of the ice is not fast enough to close the conduits during the cold season (Vatne, 2001). In addition, where the ice is frozen to the bed, there are no possibilities of a distributed drainage system. Thus subglacial drainage in cold-based glaciers is restricted to a channelized system, and it has not been shown to influence glacier dynamics in any way (Gulley et al., 2009b). These properties make the stagnant ice landscape very similar to a karst landscape, and the term 'glacial karst' is not uncommon in the literature (Badino, 2002).

2.4.2 Supraglacial drainage

For a supraglacial channel to form, the rate of channel incision must be faster than the rate of surface ablation (Marston, 1983). This is most likely to happen when the ice is still covered by snow, early in the ablation season. Kostrzewski and Zwolinski (1995) present a three step model for supraglacial channel formation, shown in figure 2.5. Here, after the snowpack has been warmed to a temperature where no more refreezing can occur, water percolates through the snow and gathers on the ice surface. Tiny streams follow the rills and structures of the ice, melting a path while the snow cover still protects it from being melted out by the sun. The channel melts fastest downwards at first, when the properties of the ice are still similar on all sides. When the snow cover disappears, solar radiation starts working on the ice surface. Ablation weakens the top layer of the ice by making it more porous, and so the thermal erosion of the stream increases in the lateral dimension.

The net rate of channel incision has been shown to be a major influence on the form and development of a supraglacial channel. Ferguson (1973) writes it as

$$i - (1 - k)a \tag{4}$$

where i is the vertical incision rate, a is the surface lowering rate, and k is the relative efficiency of stream bed melting from solar radiation, which is close to 1 for shallow channels and zero in deep, narrow canyons. If $i > a$ the channel will get progressively deeper, even when k is decreasing. If $i < a$ the channel will still get deeper relative to the surface, but with a declining rate as it approaches an equilibrium depth with a width-depth ratio that is higher the closer i is to a. Marston (1983) found meanders to be well developed in sites where i was more than twice of a. The importance of k also means that once formed, supraglacial channels can hardly be obliterated as long as there is water flowing in them (Ferguson, 1973).

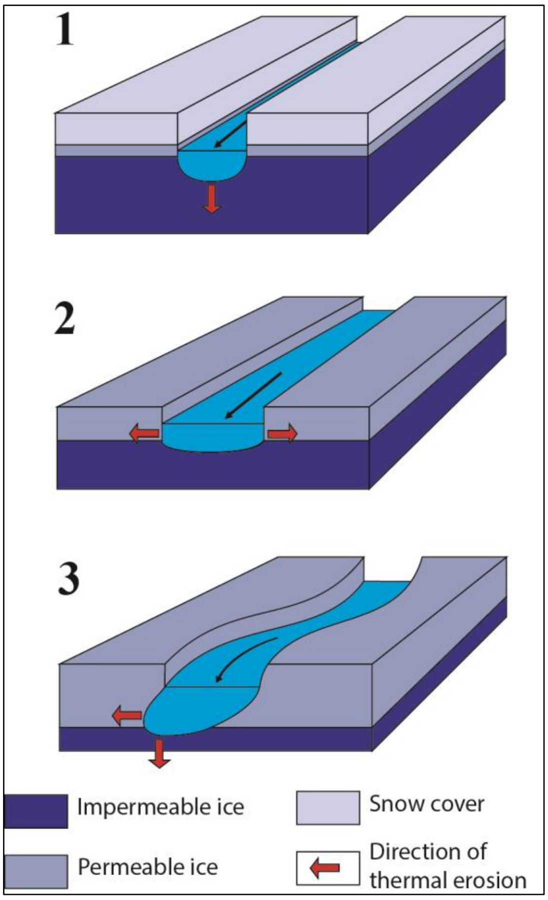

Figure 2.5: Formation of a supraglacial channel. 1: Small streams melting a channel while the surrounding surface is still protected by snow. 2: Sideways melting after the top layer is weakened by ablation. 3: Increasing Froude numbers as the stream widens makes it more susceptible to meanders, the final stage of the formation process. Modified from Kostrzewski and Zwolinski (1995).

Supraglacial channels are, to a greater or lesser extent, influenced by the structure and morphology of the glacier on which they run. On a large scale, the glacier morphology determines the slope of the stream, which has a huge impact on incision rates and channel pattern. The incision rates increases with velocity, which in turn increases with slope (Jarosch and Gudmundsson, 2012). The curve of the cross-section of the glacier also plays a major role in shaping the overall drainage pattern. A convex surface causes the streams to run off the sides into the moraine area; a concave surface causes the streams to converge in a dendritic pattern; and a straight slope with an even gradient gives supraglacial channels that can run parallel over surprisingly long stretches (Kostrzewski and Zwolinski, 1995).

On a smaller scale, supraglacial channels often follow crevasse patterns and tectonic structures in the ice, particularly in their early phases. Hambrey (1977) reports the channel network to be exploiting favourably orientated stratification and weaker ice layers, in a similar way as channels in bedrock follows the lithology. The study observes that older channels tend to develop patterns independently of structure, possibly because large volumes of water over time obliterate the effect of differences in the resistance of the ice. Glacio-tectonic structures may also be of importance for where the supraglacial channels leaves the surface and plunges into the englacial system (Benn et al., 2009).

2.4.3 The plunge: from the supraglacial to the englacial environment

Like the rivers in bedrock and alluvium, the stream channels in ice evolve towards the shortest route to sea level, the ultimate base level for terrestrial streams. Unlike the former two, however, stream channels in ice have the incision rates and substrate dynamics to make a shortcut of hundreds of metres over a relatively short amount of time. Supraglacial channels can cut themselves into the ice gradually, or force the ice to open vertical shafts down to its bed. The ways in which they do this are important to understand in order to comprehend the overall configuration of glacier drainage systems, and two processes have been presented to explain this.

Hydrofracturing

The first process is hydrologically driven crevasse propagation, or hydrofracturing (Benn et al., 2009). In this case water pressure forces a crevasse to widen further, sometimes all the way to the glacier bed. The process requires crevasses of a certain depth as well as an excess of meltwater to build enough pressure, and is more likely to happen on polythermal glaciers than on the slow-moving, cold-based ones. It is worth mentioning in this section, because many cold-based glaciers have been polythermal in the past and might still contain structures that were formed at that time.

Boon and Sharp (2003) investigated the importance of hydrofracturing on a predominantly cold-based glacier in Northern Canada, and found it to be an important driving force for establishing a link between surface and bed. The studied crevasse was observed to fill with water during the onset of the ablation season, until the channel had ponded to a height of 6,9 m above its channel bed. The pond lasted for several days before it suddenly started draining, and disappeared completely within one hour. The main crevasse was believed to penetrate all the way to the bed in order to drain the pond, through 150 m of ice.

Cut-and-closure

The other process is called cut-and-closure, and happens when a supraglacial channel has cut itself so deep into the ice that the roof closes above it. Ice creep is the most effective process for roof closure at greater depths, but before that a plug of snow, superimposed ice and debris can pinch off the channel from the surface (Gulley et al., 2009a). For a channel to incise deep enough to have the roof close, the net incision rate must be high and last over several ablation seasons, making cut-and-closure channels a feature of areas with low rates of ablation and ice creep. Cut-and-closure is considered to be an important process for englacial channel formation in uncrevassed regions of polythermal and cold-based glaciers (Baelum and Benn, 2011; Gulley et al., 2009a; Naegeli et al., 2014).

Cut-and-closure channels commonly form on the sides in the ablation zone of the glacier (Gulley et al., 2009b). The convex shape of this area directs the meltwater channels to the sides, and there is also a supply of snow and debris from the slopes around the glacier that helps plug the channel roof. However, Vatne (2001) also describes an englacial channel formed by the cut-and-closure by the glacier centreline.

Jarosch and Gudmundsson (2012) coupled a numerical ice dynamics model to hydraulic model to simulate the transition of a supraglacial channel into an englacial channel. Their model shows that the surface slope of the glacier greatly influences the depth and time step at which the channel is pinched off from the surface. Steep surface slopes give a fast pinch-off, at a lesser depth. Channel slope and discharge mainly influence the depth at pinch-off, but the increased incision rates also speed up the roof closure process somewhat. This model is based on the ice temperatures of temperate glaciers. In cold ice the creep rates are considerably lower, and roof closure by ice and debris is considered to be more important (Gulley et al., 2009a).

The processes of cut-and-closure and crevasse propagation produce initially very different englacial channel morphologies. The latter creates a mostly vertical plunge to the bed, while the former brings along its supraglacial pattern under the surface. After the stream has been isolated from the surface, however, new processes take over the channel evolution.

2.4.4 Englacial drainage

Direct cave explorations have unveiled a diversity of englacial channel forms and patterns, shaped by the temporal interactions between ice dynamics, englacial debris content and hydromechanics. Where ice creep rates are high, the ice will close around the water flow and create a tubular conduit where the ice pressure and wall melting adjusts the circumference to the discharge (Röthlisberger, 1972). Following the terminology of karst studies, this flow condition is named *phreatic*, or pressurized, flow. Ice melting is equal in all directions of the channel wall, thus no specific downward incision of the channel can be expected (Jarosch and Gudmundsson, 2012). The opposite of phreatic is *vadose* flow, where the stream flows under atmospheric pressure. In this case there is no contact between the stream and the channel roof, and the incision rates are highest in the downward direction. In englacial channels this happens when incision rates are higher than the rates of roof closure, a common condition in cold ice (Gulley et al., 2009b).

Gulley et al. (2009a) divide the passage morphology of englacial channels in four main groups. The first, plugged canyons, are tall, narrow passages in which the roof is plugged by snow and debris. In the second, sutured canyons, the roof is closed by ice creep. The morphology is similar to plugged canyons, but with walls that slope inward until they meet with a characteristic suture in the joint. The third group is horizontal slots, very wide passages with low roofs. They are usually found in the lower parts of the systems, formed by a combination of lateral channel migration and closure of the higher levels by ice creep and channel blockage. The last group is tubular passages, the only one of the four that is interpreted to be formed under phreatic conditions. Their cross-section is roughly elliptical and somewhat irregular, and they were only found in restricted areas of the channel in the study of Gulley et al..

The planar form and longitudinal profile of englacial channels reflect, to some degree, the process by which they were formed. Channels formed by crevasse propagation by hydrofracturing follow the form of the crevasse, plunging steeply towards the bed, like the moulins mapped by Holmlund (1988). As the maps from this study show (fig. 2.6), the plunge can be disturbed by some steps along the channel, but the main direction is steeply dipping.

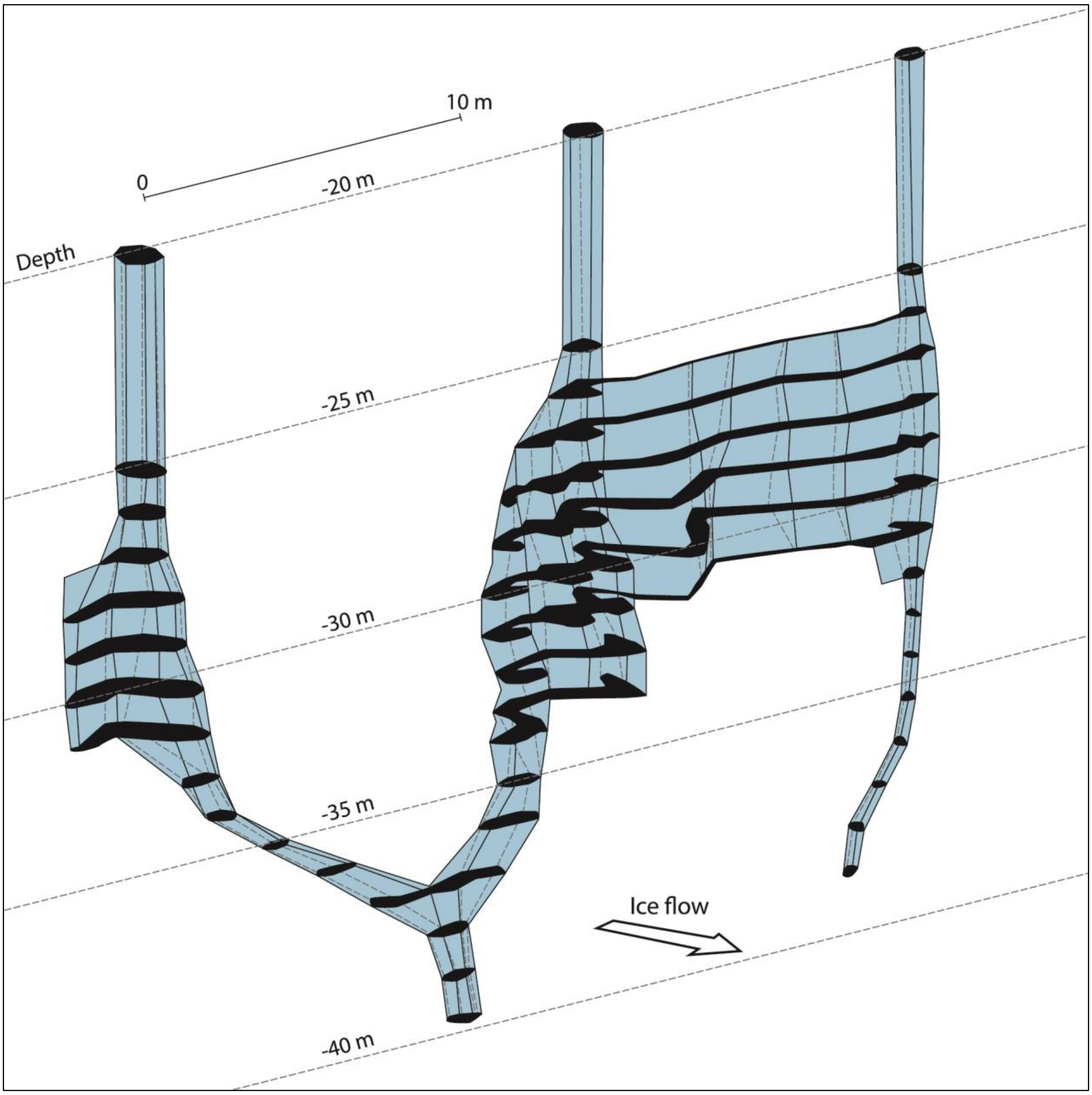

Figure 2.6: Moulins in Storgläciaren in Sweden, formed by water draining through a crevasse and later frozen shut. They were mapped by measuring the outlines of the frozen meltwater as the surface was lowered by ablation. Modified from (Holmlund, 1988)

Channels formed by cut-and-closure show a much more gentle descent into the ice, gradually getting deeper by steps and horizontal or gently sloping sections, like the channels mapped by Gulley et al. (2009a) and Vatne (2001), the last one shown in figure 2.7. These channels can display both meandering and step-pool reaches similar to that of supraglacial channels. After they are cut off from the surface, however, they evolve differently than their supraglacial origins. Over time, a cut-and-closure englacial channel with the initial channel pattern of a supraglacial stream can develop into a moulin going straight from the surface to the bed (Naegeli et al., 2014).

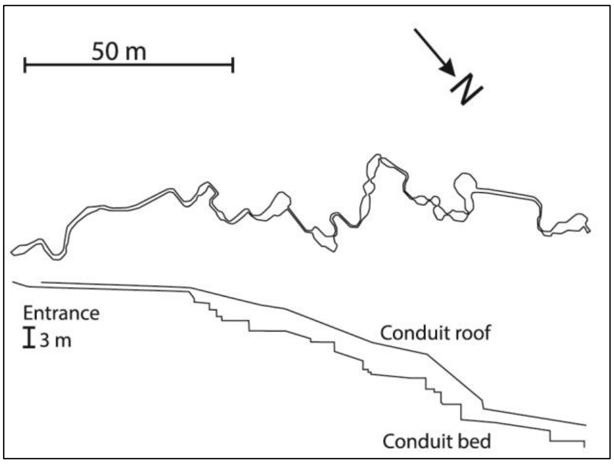

Figure 2.7: Orientation and geometry of an englacial channel draining a moulin on Austre Brøggerbreen, Svalbard. From Vatne (2001).

An additional process for englacial channel formation is reported by Gulley and Benn (2007), in which the conduits exploit structural features in the ice. The study was done on a debris-covered glacier in the Himalayas, where debris-filled crevasse traces form permeable, unfrozen planar structures within the ice. These bands can be exploited by englacial channels, which has been shown to follow the least resistive hydraulic potential.

Though the morphology of englacial conduits can be observed and surveyed, the processes of channel evolution after the cut-off from the surface are not well known. The channels are difficult to access for surveys, and practically impossible to investigate when they are active. The energy transfer between water and ice is the same as in supraglacial streams, but englacial streams do not receive any additional energy input from air temperature and solar radiation. The only warming of the water comes from the transition of potential energy into thermal energy, as described in the study of Isenko et al. (2005).

As the englacial channel progresses deeper into the glacier, the increased overburden pressure on the surrounding ice will increase the creep rates affecting the channel form. In the deeper parts of englacial channels there is an increased chance of the channel being blocked by ice creep or a combination of creep and ice accumulation, forcing the channel to find a new route at a higher level (Gulley et al., 2009a). Once it reaches the channel bed, however, an entirely different morphology presents itself.

2.4.5 Subglacial drainage

The subglacial drainage system has traditionally been divided in two main groups: distributed systems, comprising water films, linked cavity systems, braided channel networks and groundwater flow; and channelized systems, either incised up into the ice or down into the bed (Benn and Evans, 2010). As cold-based glaciers are frozen to their beds, the distributed systems are not believed to be possible in their setting. This temperature condition, and the effect of ice creep on channels of a certain depth, put large constraints on the development of channelized drainage as well. Still, subglacial channels have been found and mapped in cold-based glaciers, and they have been shown to persist over considerable periods of time (Gulley et al., 2012; Naegeli et al., 2014).

Subglacial channels can exist in two distinguished forms, depending on the topography and material characteristics of the bed. Channels of the first group are called Röthlisberger channels, or R-channels, and were first conceptualized by Röthlisberger (1972). These are running along the bed-ice boundary, and incise upwards into the ice. In the original theory they were believed to be semi-circular, based on an assumption of uniform discharge. As discharge in the subglacial system fluctuates with that of the rest of the drainage system, however, it was later suggested a broad, low conduit to account for the ice creep and the dominantly lateral channel incision at low discharges (Hooke, 1989).

If the characteristics of the bed material allow it, and if the channel runs along the same path for a sufficient amount of time, it can incise down into the bed of the glacier. This is characteristic of the channels of the second group, Nye or N-channels, first mentioned in the study of Nye (1973). These channels are more stable features than the R-channels, as the bed material is less erodible than the ice walls of the other.

Both channel types have been found interchangeably in cold-based glaciers. Naegeli et al. (2014) mapped four subglacial channels in Tellbreen in central Spitsbergen. The channels showed a complex morphology with both tubular passages and horizontal slots, branching channels and meandering reaches. The channels were mostly of the R-type, the main exception being a 80 m long N-channel extending downglacier from a 42 m high moulin connecting the surface to the bed. The study of Gulley et al. (2012) was carried out in a subglacial channel in Rieperbreen, located in the same area as Tellbreen. This channel was mostly of the N-type, as it was deeply incised in the glacier till. The channel morphology varied very little between the years of survey. The channels of both studies were believed to be formed by cut-and-closure reaching the bed.

Subglacial channels follow the slope of the glacier bed, without the distinct step-pool profile shown in englacial channels. The planar form of the above mentioned channels were described as being mainly straight, highly sinuous or with a complex, branching pattern. The latter two were explained partly by the heritage from their supraglacial origins, while the straight pattern originated from the attributes of the till into which it was incised.

A unifying characteristic of all subglacial channels, separating them from most englacial and supraglacial channels, is the sediment floor and the roughness that follows with it. Gulley et al. (2012) found big boulders on the channel floor of their studied channel, giving a Manning roughness of up to 0,67 s $m^{1/3}$, compared to the range of 0,013 to 0,039 s $m^{1/3}$ found in supraglacial channels. This has a significant effect on velocity, and on the way in which throughflow measurements of the systems need to be interpreted.

The focus on subglacial channel roughness is quite recent, and it is one of many examples of how important it is with direct observations and measurements to support theories and remote measurement techniques. This discussion is taken further in the following chapters, where a series of maps covering 12 years of surveying is presented to shed light on the long-term evolution of an englacial channel system.

3 Case study: Long-term evolution of an englacial meltwater channel in Austre Brøggerbreen, Svalbard

Stream channels on glaciers are complex, dynamic features. To understand their characteristics, and the mechanisms of their behaviour, it is still necessary to study them in their environment and do measurements and observations *in situ*. This chapter presents a study of an englacial channel in a cold-based glacier, combined with previous studies of the same channel done over a time span of 12 years. The glacier, Austre Brøggerbreen, is close to the research settlement Ny-Ålesund, and is relatively easy to access in both summer and winter. Mainly because of this, it has been very thoroughly studied over several decades. Aspects of its thermal regime, mass balance, hydrology and drainage system has been presented and discussed in the literature, making it a good object for further investigations.

3.1 Study site

Austre Brøggerbreen is a land-terminating valley glacier, located at 78°50'N, 11°50'E, on the North-Western coast of Spitsbergen in the Svalbard archipelago (fig. 3.1). The glacier stretches from 80 to 600 m a.s.l., and covers an area of 9,35 km^2 (Porter et al., 2010). The geology of the area is mostly Carboniferous and Permian carbonate and clastic sedimentary rocks, with an area of phyllite and quartzite underlying the upper parts of the glacier (NPI, 2015). Borehole measurements in front of the glacier has shown the permafrost layer to be at least 140 m thick, making it very likely that the sub-zero ground temperatures goes underneath the glacier tongue and sides (Liestol, 1988).

Accumulation rates are typically low in the area, around 3 m of snow annually (Hagen et al., 2003). Combined with the low air temperatures and the resulting low mass turnover, this gives extremely low ice velocities. At Austre Brøggerbreen the surface velocity has been measured by Hagen et al. (1991) to 0,5 m/year in the lower ablation area, and 2 m/year in the equilibrium zone. The mean annual air temperature in the equilibrium zone is -8°C (Kohler, 2014), but this does not prevent an average surface lowering of 1 m/year over the last two decades (Kohler, pers. communication).

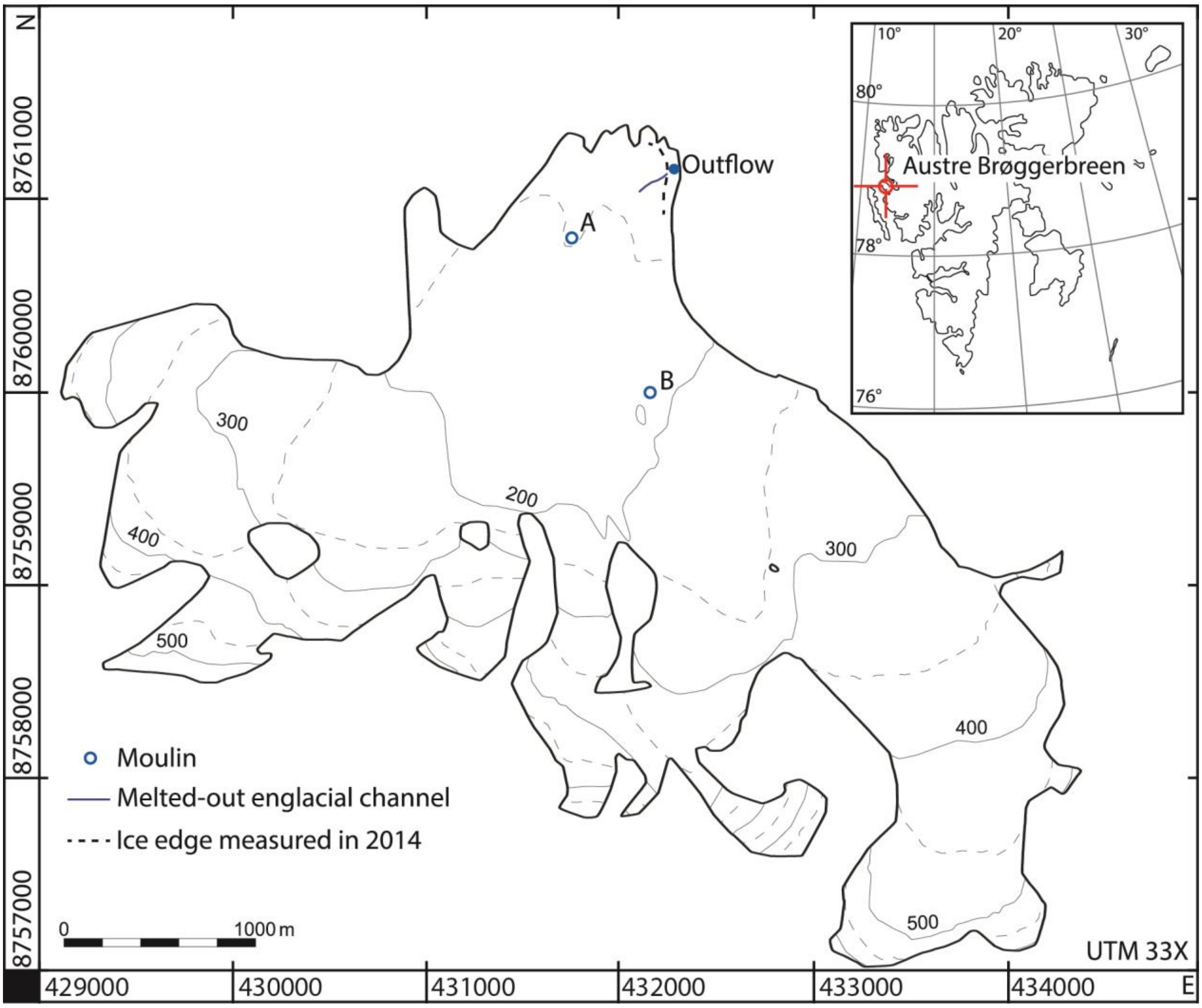

Figure 3.1: Overview and location of Austre Brøggerbreen, with GPS measurements of the moulins, the melted-out portal stream and the ice edge by the portal measured in 2014. Map data from svalbardkartet.npolar.no.

The cumulative mass balance of the glacier has been in steady decline since the observations started in 1968 (Kohler, 2014). The decrease in ice thickness led to a switch in glacier thermal regime in the first half of the 90's, when the glacier is thought to have gone from being polythermal to predominantly cold-based. In the study of Hagen et al. (1991), borehole measurements showed temperatures close to 0°C at the bed where the ice was more than 80 m thick. A multi-frequency radio-echo sounding done by Bjornsson et al. (1996) a few years later could detect no temperate ice in the glacier, but they were still open for the possibility that there could be temperatures at 0°C at the bed.

Hagen and Sætrang (1991) mapped the bed topography of the glacier, presented in figure 3.2. Based on this map and GPS measurements done in mid-summer 2014, the ice thickness is found to be between 50 and 70 m at both of the largest moulins (A and B, fig. 3.2). This is thinner than the 80 m thickness predicted by Hagen et al. (1991) to be necessary for the temperature to rise to the pressure melting point, so the glacier is most likely frozen to its bed at least in its central and lower part. However, numerous crevasse traces on the surface, and foliations in the ice layers shown in the walls of the meltwater channels, show that there have been times with higher ice velocities in the glacier. Hagen et al. (1991) refer to aerial photographs from 1934 showing a crevassed surface of the glacier, and also consider these crevasses a possible initiating points for the moulins.

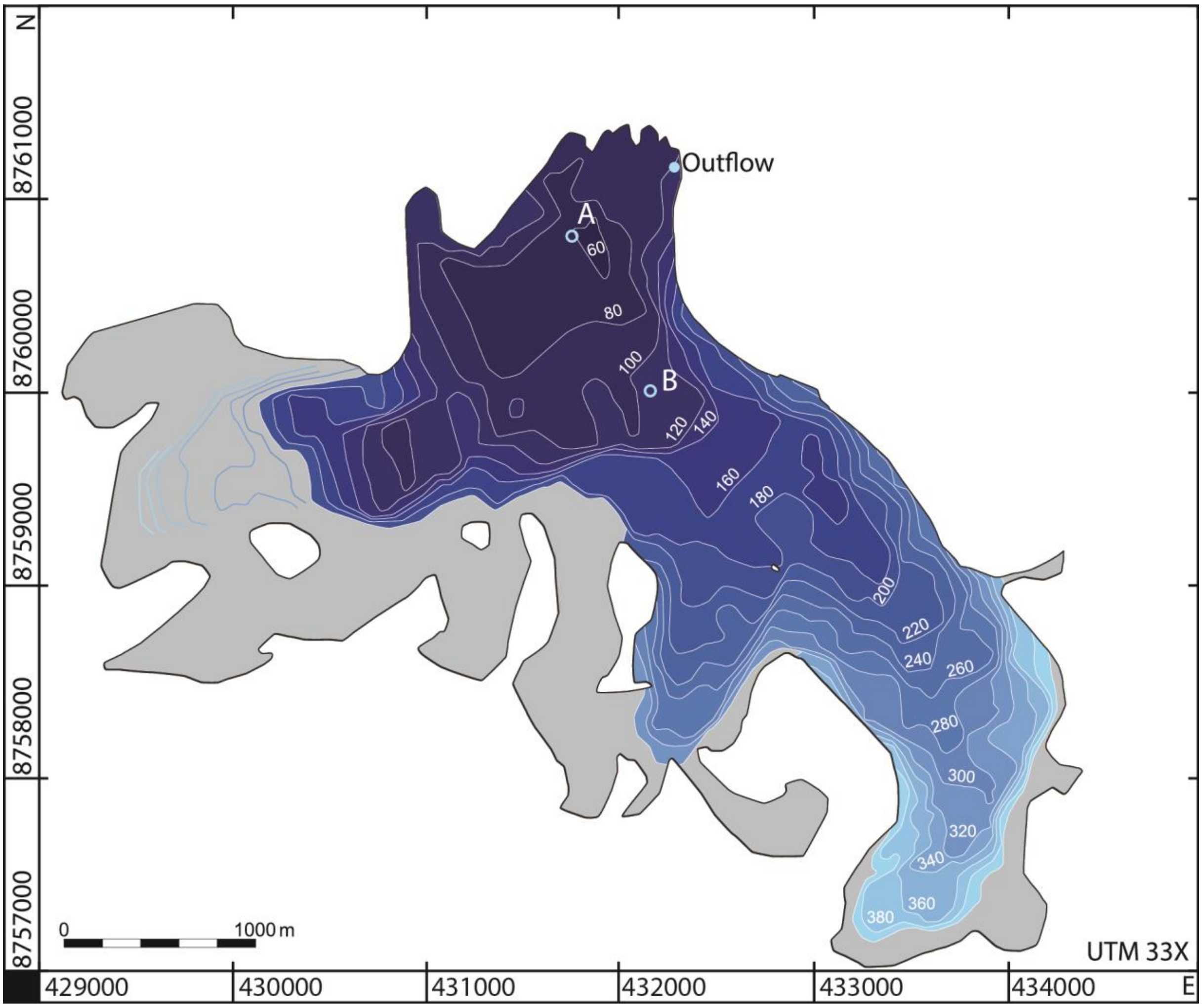

Figure 3.2: Bed topography map of Austre Brøggerbreen, modified from the map of Hagen and Sætrang (1991) to the glacier outline of 2007, using map data from svalbardkartet.npolar.no.

3.1.1 Drainage system

The drainage system of Austre Brøggerbreen is characterized by an extensive network of supraglacial channels, running off the glacier on the sides or disappearing into the englacial environment through well developed, perennial moulins. Dye tracer studies done by Hagen et al. (1991) show that the moulins on the central and lower part of the glacier drain through the same portal, at the eastern flank by the glacier front. An exploration of the moulin A (fig. 3.1), also in the mentioned study, led to the conclusion that the englacial system was old and well developed.

Later this moulin has been thoroughly surveyed, together with two higher-lying moulins on the glacier. Subsequent dye tracer studies have confirmed the first results, supporting the notion that the englacial system is well-developed and perennial. The portal stream has melted out of the ice in the later years, as shown in figure 3.3. In the summer of 2014 it surfaced in a >10 m deep, 170 m long open trench, running in a mostly straight, horizontal line towards the glacier margins until plunging the last meters down to the bed about 50 m from the ice edge.

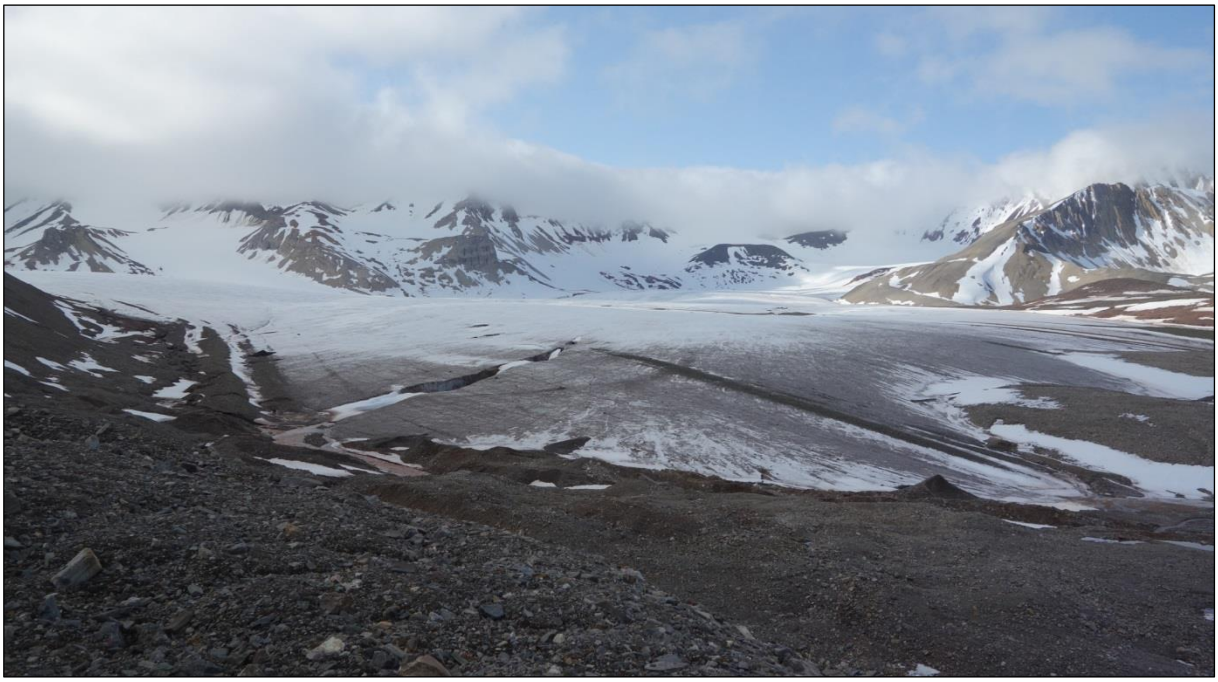

Figure 3.3: The glacier front of Austre Brøggerbreen, facing southwest. The trench in the centre-left side of the photo is the melted-out portal stream, exiting under a snow bridge and joining a stream from the lateral moraines by the glacier margin. Photo: Myreng, 2014.

The object of this study is the englacial channel extending from moulin B (fig. 3.1). The moulin is located in a depression upglacier from a small bulge in the ice surface, about 1200 m from the glacier portal. In the summer of 2014 the moulin was fed by one supraglacial stream (fig. 3.4), which showed some irregular meanders and small steps. Observations done later the same year suggest that the snow plug constituting the roof of the moulin survived through the ablation season. As this was a year of exceptional snowfall, and the spring came late (Jenssen, 2014), it is possibly an exception to the normal condition.

Figure 3.4: The supraglacial stream draining into moulin B. The moulin itself is just out of the frame to the right of the photo. Photo: Myreng, 2014.

Based on data from previous surveys and dye tracer tests, the englacial stream is thought to run under atmospheric pressure all the way to the portal (Vatne, 2001). All surveys have been stopped by unfrozen pools before reaching the bed, at no more than 110 m in a straight line distance from the cave entrance. Questions like whether the channel reaches the bed or not, and when it joins the rest of the system, thus remain unanswered.

3.2 Data and methods

3.2.1 Mapping channel morphology

The best way to get good data on channel morphology is by direct observation and measurements. Remote sensing techniques like photogrammetry and satellite surveillance are useful for big scale surface morphology, and ground penetrating radar (GPR) can be used to pin down the location of englacial features, but none of these techniques catch the details needed to understand the evolution of englacial drainage systems.

Direct exploration of englacial conduits borrows techniques from cave surveys, *speleology*, and is often referred to as glacio-speleology in the literature. The conduits are entered in winter, when the water flow is absent, using basic mountaineering techniques. A simple 3D model of the cave is constructed by dividing the channel into straight line segments based on changes in slope and curvature, and measuring the length, inclination and orientation of each segment. The data is then plotted in a cave survey program, which computes a 3D-line model of the cave. The amount of change in slope and curvature required for a segment divide varies between projects, depending on the aims of the survey and the available time.

For this study, a fairly high precision was used for the segment measurements. The measurements were done along the centre line of the channel, ~30 cm above the frozen water surface, using a compass and a Leica Disto D8 laser distance meter with a built-in clinometer. A new channel segment break was made at every major change in inclination and curvature rather than breaches in the line of sight. This way the meanders comes out more clearly in the model, and both planar and profile patterns hold a good level of detail on the overview scale of the channel.

The conduit was then thoroughly photographed, and the channel width measured at every segment break. The biggest steps and pools were also drawn and measured in detail, and one walk through the whole length of the conduit was filmed.

Analysis of the channel morphology

The data from the englacial measurements were plotted in the cave surveying software COMPASS, from which a local coordinate system was computed and the coordinates of the conduit segments exported into Microsoft excel. The combination of these two programs gives the flexibility needed to study, analyse and present the data in a satisfying manner. To give a proper presentation of the 3D models of the channel surveys, a *shadow box*-view was made in COMPASS and cleaned up in Adobe Illustrator. For the bulk comparisons of several studies, a simpler, 2D graph is used.

The meander evolution in the first part of the channel makes it nearly impossible to represent the profile merely by straightening out the channel by plotting the straight segments and step segments without the bearing. This way the steps would seem to have migrated far downstream, while they in reality migrate upstream towards the entrance. To rule out the increase in sinuosity from the profile view of the channel, the steps are plotted with depth and straight-line distance from the entrance. This mode of representation will show a lower-than-actual upstream migration rate, as the steps migrate through meanders that are not visible in profile view. The values of upstream step migration were therefore measured from the 3D model in COMPASS rather than the 2D graph.

Error sources and data quality

Although direct observations and measurements are the pillars of traditional physical geography, the resulting data is not without flaws. These methods are held together by a collection of subjective choices, like where to make a segment break and where to measure stream width, in general terms how to simplify the data enough to write the real world down in numbers. There is also room for several human errors, like holding the compass or laser pointer in the wrong way, or reading off the wrong numbers on the compass.

At the scale of which the analyses are done in this project, absolute accuracy is not strictly necessary. Errors of a few centimetres do not alter the results considerably when looking at a channel of several hundred meters. Of greater concern are the errors in the compass measurements of the orientation of each segment, as the compasses were hand-held by persons carrying large amounts of metal climbing gear. To account for some of this disturbance two compasses were used, measuring the bearing towards each other from the opposite ends of the channel segment. The resulting bearings were then averaged in order to minimize the error. The resulting data is considered of sufficient accuracy for this project, but the method by which they were obtained should still be kept in mind.

3.2.2 Measuring throughflow and discharge

Throughflow measurements

To get a better estimate on how the water flow behaves through the entire englacial system, a dye tracer can be used when the system has become active in summer. The tracer is released into the supraglacial stream that feeds the moulin, and gets dispersed and transported with the stream through the englacial system. At the glacier portal a probe continuously measures the concentration of dye tracer in the stream, giving an output curve that shows when the first amounts of dye tracer have come through the system, when the concentration is at its peak, and the length and concentration of the tailing remains of dye tracer in the stream.

For this study, the fluorescent dye Rhodamine WT with a concentration of 21 % was used as a tracer. Injections of 40 ml were done in the supraglacial stream above the moulin B (Fig. 3.6). The tracer injections were detected at the glacier portal, using a Seapoint Rhodamine

Fluorometer, and logged at 10 s intervals by a Campbell Scientific CR1000 data logger. The fluorometer measured the dye concentration as fluorescence intensity, which was logged as a value in millivolts.

Figure 3.5: Injecting dye into the supraglacial stream upstream of moulin B, which is still hidden under a solid snow bridge. Photo: Myreng 2014.

Analysis of the dye return

From the logged dye return at the glacier portal, a selection of parameters was calculated to describe the throughflow characteristics of the system. First the transit distance, which is the straight line distance from injection point to detection point, was determined using a point measurements from a handheld GPS. From this it was possible to calculate the transit speed, which is the minimum estimate of the throughflow velocity. It is found using the transit distance and the time it takes from dye injection to the peak of the throughflow curve passes the detection point.

The next parameters were found using the equations described in the study of Willis et al. (2012). The first was the dispersion coefficient, which is a measure of how much the dye spreads out between the release point and the detection point. The dispersion coefficient is affected by features in the channel that causes a delay to parts of the dye cloud, like roughness or pools, making it spread out in the conduit length. Divided by the throughflow velocity, the coefficient gives the dispersivity of the channel.

The dispersivity is the rate of spreading of a dye cloud relative to the rate of dye advection during its flow through the system. It gives an indication on the complexity of the route from release point to detection point. In general, tracer returns of d > ~10 reflect inefficient drainage pathways, while less dispersed returns (d<~10) reflect highly efficient drainage (Hubbard and Nienow, 1997). In more recent studies, high rates of dispersivity has also been interpreted to reflect channels with high roughness relative to the discharge (Gulley et al., 2012).

A return curve showing high dispersion will be relatively broad and low, while low dispersion will be shown as a narrow, tall curve. As shown in figure 3.6, it is also possible to infer the configuration of the drainage system based on the dye return curve. Flattening or multiple peaks indicates divergence in the drainage pathways (Hubbard and Nienow, 1997), while tailing suggests temporary storage features or high channel roughness (Fyffe, 2012).

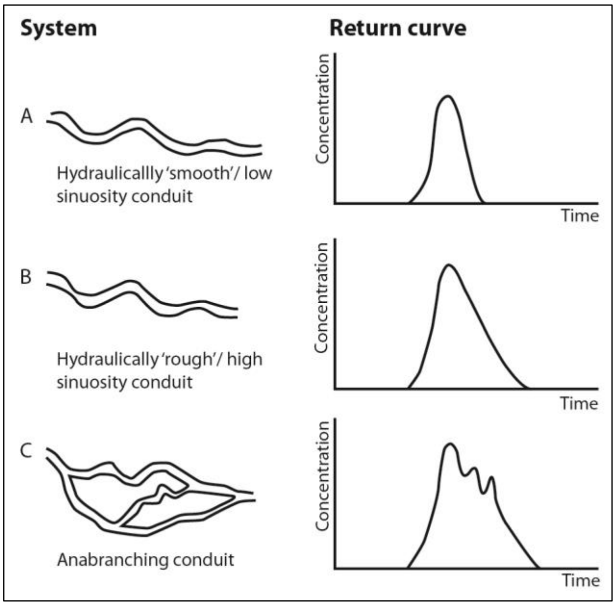

Figure 3.6: Theoretical examples of dye return curves resulting from different drainage systems. Modified from Nienow (1993).

Discharge measurements

The discharge in the supraglacial stream feeding the moulin was calculated by salt dilution gauging, following the method as described in Merz and Doppmann (2006). The method is done by releasing a known amount of salt at some distance above a chosen measurement point in a stream, and then measure the conductivity at short intervals as the salt, mixed into the water flow, is transported past this point. In this study the amount of salt used was 200 g, and the conductivity was measured at intervals of 4 seconds using a WTW LF95 Conductivity Field Meter. The conductivity meter was calibrated using stream water from the measurement site, and discharge measurements were done after releasing the dye tracer. The conductivity meter also contains a thermometer, giving the opportunity to write down the temperature in the stream.

Error sources

There is a set of weaknesses to the method of indirect measurements with dye tracing, of a somewhat different nature than those of direct observations. The dye return curves are the readings of processes that are practically impossible to observe directly, showing a summary of the flow conditions through the entire length of the englacial channel. Interpreting what causes the dye return curve to come out as it does, and where in the system it happens, often involves a certain amount of speculation.

A major challenge when interpreting the changes between return curves, either from year to year or through one season, is to determine whether the change come from changes in the channel morphology or changes in the flow conditions in the stream. There is no certain way of finding the right answer except surveying the whole length of the channel.

3.2.3 Secondary data sources

The proceedings mentioned above were used in the last year out of a procession of studies done in the same channel, a selection of which will be presented in this study. The englacial channel was first surveyed in 1998 (Vatne, 2001 – Channel B), then again in 2002 (Refsnes, 2004 – Channel A) 2004, and 2008 (Vatne and Irvine-Fynn, in preparation). The dye tracer tests that will be used in this presentation were performed in 2005 (Holtermann, 2007 – Channel B) . The methods in these studies have been the same or similar to the ones used in the fieldwork for this project, with some improvements in equipment and some variations in the level of detail.

All the previous englacial surveys used breaches in the line of sight as a breaking point for the channel segments in the horizontal dimension, which makes the meanders look pointy in the presentation of their planar pattern. Another consequence of this was fewer segment breaks per length of channel compared to the latest survey, giving an overall lower resolution of the survey map. The dye tracer experiments were performed in a similar manner as described in the previous section.

The glacier thickness at the site of the moulin is a factor that has changed considerably through the survey period. To make it possible to compare englacial surveys over a period of 12 years, these changes have to be accounted for in the dataset. This was very helpfully done by Jack Kohler of the Norwegian Polar Institute (NPI), who has carried out the mass balance measurements on the glaciers in this period. As the moulin lies less than a hundred meters from one of the ablation stakes (number 6) of NPI, their numbers are considered valid for this study. Kohler (pers. comm.) used the annual mass balance of ablation stake 6 to calculate the change in meters of ice equivalents for this area, getting values that correspond quite well with the less frequent GPS measurements from the ablation stake.

The resulting values give a number for how much the glacier surface was lowered each year, making it possible to adjust the depth of the englacial surveys to the actual surface of their respective years. With this adjustment all the surveys could be combined in the same 3D model in COMPASS and Microsoft Excel, and from there obtain the changes from survey to survey.

According to pictures and observations done on the glacier surface, the entrance of the englacial channel has migrated upstream during the survey period, with possibly as much as 15 m or more (Vatne, pers. comm.). No accurate coordinates for the entrance exist from the surveys previous to 2014, so the migration distance could not be quantified and adjusted for in the dataset. The choice was therefore made to set the same entrance point on the surveys in the survey compilation figure. The rates of step migration where measured in relation to their position in the planar pattern, which is not influenced by the location of the entrance.

3.3 Results

The channel survey of 2014 will be presented first, along with pictures and a detailed description. Then follows the mapped result from this survey and the previous ones, with the main characteristics and the trends in the developing channel patterns. In the last section, the results of the throughflow measurements of 2014 and 2005 will be shown.

3.3.1 Channel morphology in 2014

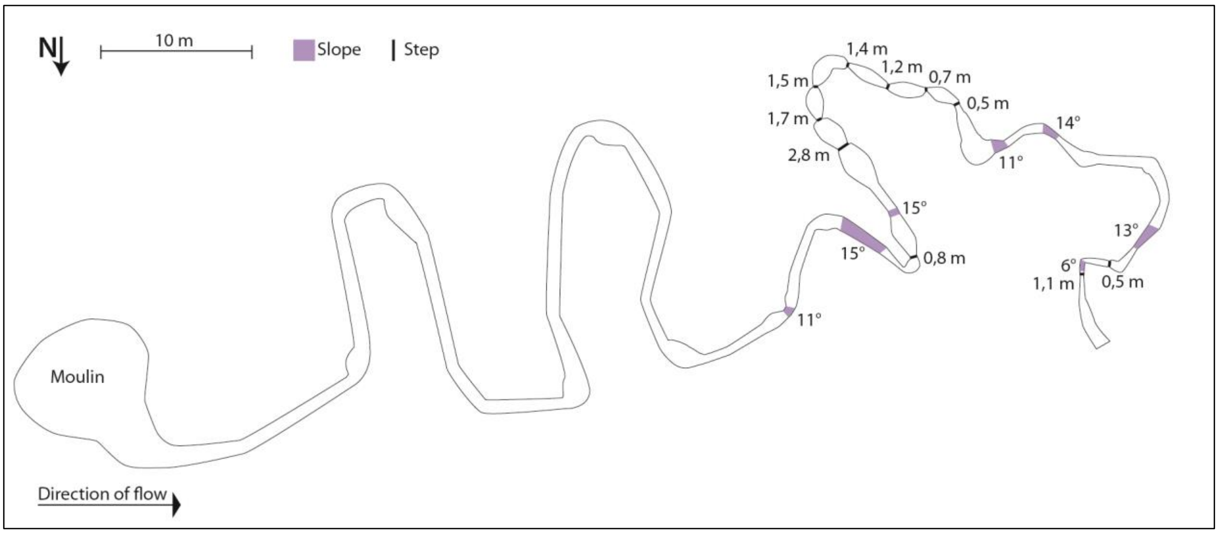

Figure 3.7: A detailed map of the planar pattern of the channel in 2014. The channel segments where the slope is not noted had inclinations of less than 2°.

Figure 3.7 shows a planar map of the channel as it was when visited in the spring of 2014. It is characterized by a sequence of regular, near-horizontal meanders, and a sequence of alternating steps and pools with a less regular meandering pattern. In both ends of the step sequence there were some more gentle sloping channel segments, and a couple of short steps, but the majority of the steps were found in one concentrated cluster.

Figure 3.8: The main shaft of the moulin, showing a massive overhang separating the big shaft from the smaller one. The person in the rope is about two thirds of the way up to the surface. Photo: Myreng, 2014.

The channel entrance from in 2014 was a clear, fully developed moulin, descending from the surface in a vertical plunge (fig. 3.8). Two shafts separated by an overhanging wall of ice led down to the bottom of the moulin, while a third shaft was observed to lead only part of the way down and connect to the main shaft on a higher-lying terrace. The frozen pool at the bottom of the moulin formed a semi-circle with a diameter of 12 m, from which the channel continued horizontally downglacier.

The next part of the channel was characterized by a wide conduit, with walls sloping parallel inwards from the bends, and big, well-developed meanders (Fig 3.9). Pools of the kind shown in figure 3.9 B were found at the bend exit in all the meanders. The pools were of a sufficient depth to create a bulge in the ice surface as the water froze. They were elongated features, commonly following the line of the concave outer bank while creating a bulge in the convex inner bank.

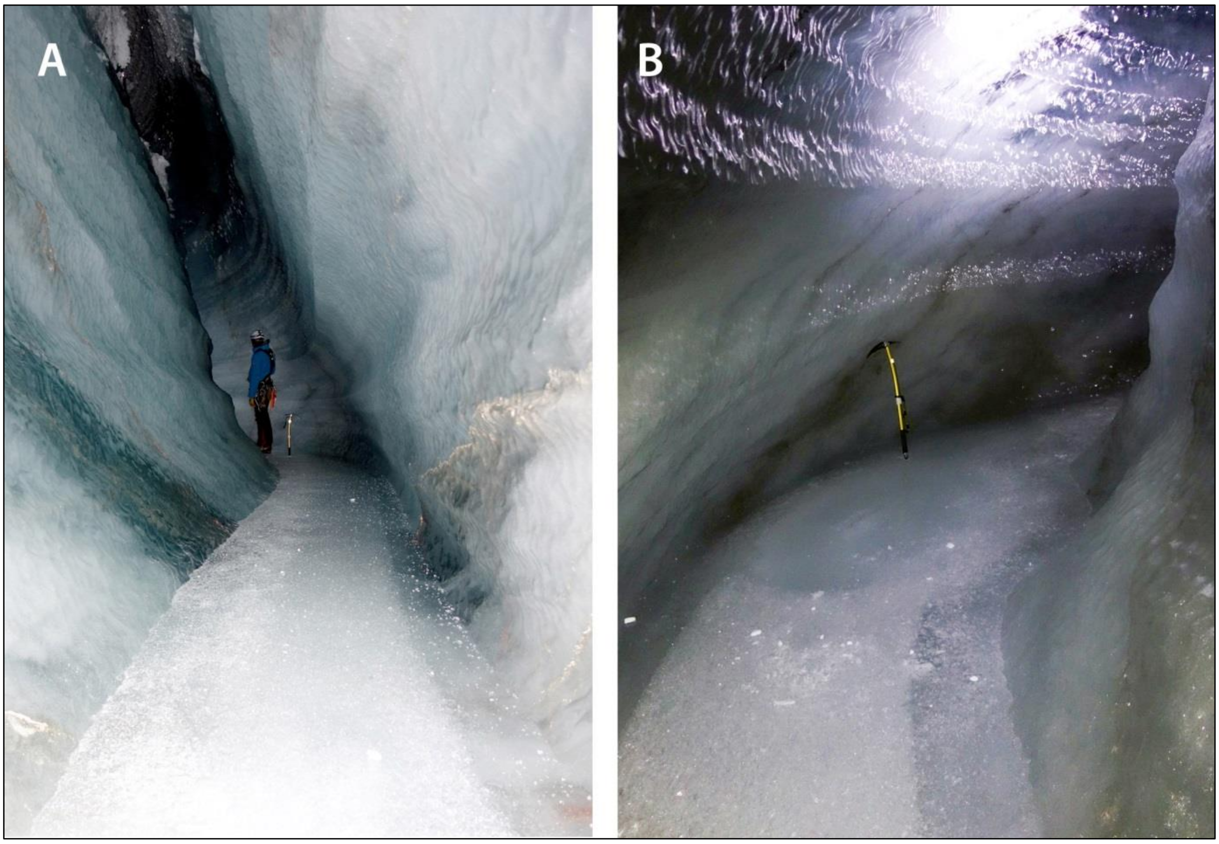

Figure 3.9: The meander section, extending downstream from the moulin. A: A particularly broad channel segment between two meanders, close to the moulin. B: A frozen pool typically found at the bend exits in this section. The ice axe in the photos is 65 cm long.

By the end of the horizontal, meandering section, the conduit narrowed. The last meander was less pronounced than the others, ending in a pool before the stepping part of the conduit began. The roof closure was still too high up to see, but the channel descended through a series of steps and pools (Fig. 3.10) still following a lesser developed meandering pattern. The channel width alternated between very narrow at the steps and wide at the pools, with clear undulations in the walls at the top of the steps.

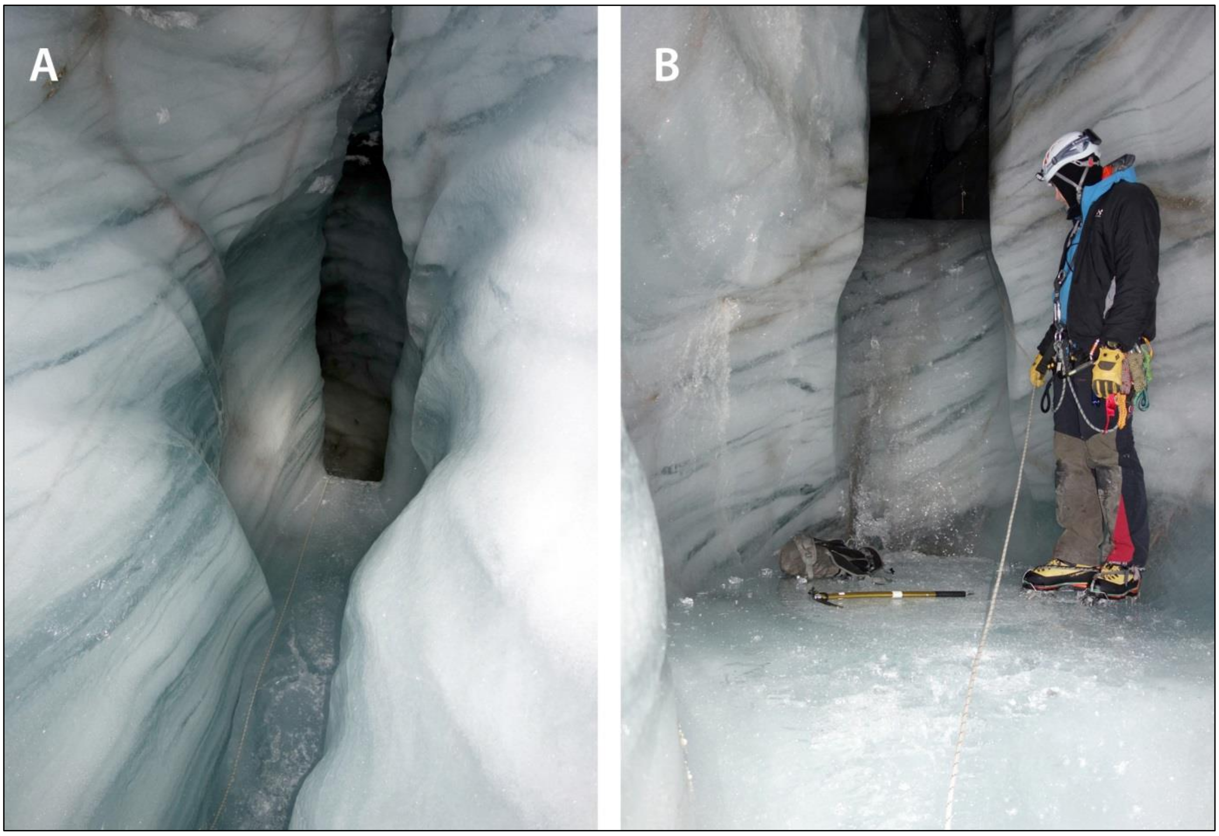

Figure 3.10: The step sequence following the meander section. A: Narrow stream widths down the steps, with undulations in the walls. B: Wide pool separating two steps. Photo: Myreng 2014.

After eight steps with a height of 0,5 to 3 m, the channel flattened out. A few steps and sloping channel segments were measured in the last part of the surveyed conduit, which was characterized by less pronounced, irregular meanders (Fig 3.11). The channel was narrower, and the undulations in the walls more irregular, than in the higher parts of the conduit. The sediment content in the ice walls was also higher, showing several deformed bands of debris. The roof was occasionally plugged by what appeared to be a frozen slush mixture, about 2 m above the channel floor. In the example shown in figure 3.11 B, a red marking stick of the kind that is used to mark the moulin entrance was frozen into the plug.

The survey of 2014 was stopped by water too deep to proceed past. The transition from frozen to unfrozen channel was sudden, in contrast to the previous studies where the transition has been gradual. This raises the question about whether the encountered water was a frozen pool or a rising water table caused by a plug further down in the system. As the channel morphology was narrow and sharply turning at this point, further observations to answer these questions could not be made.

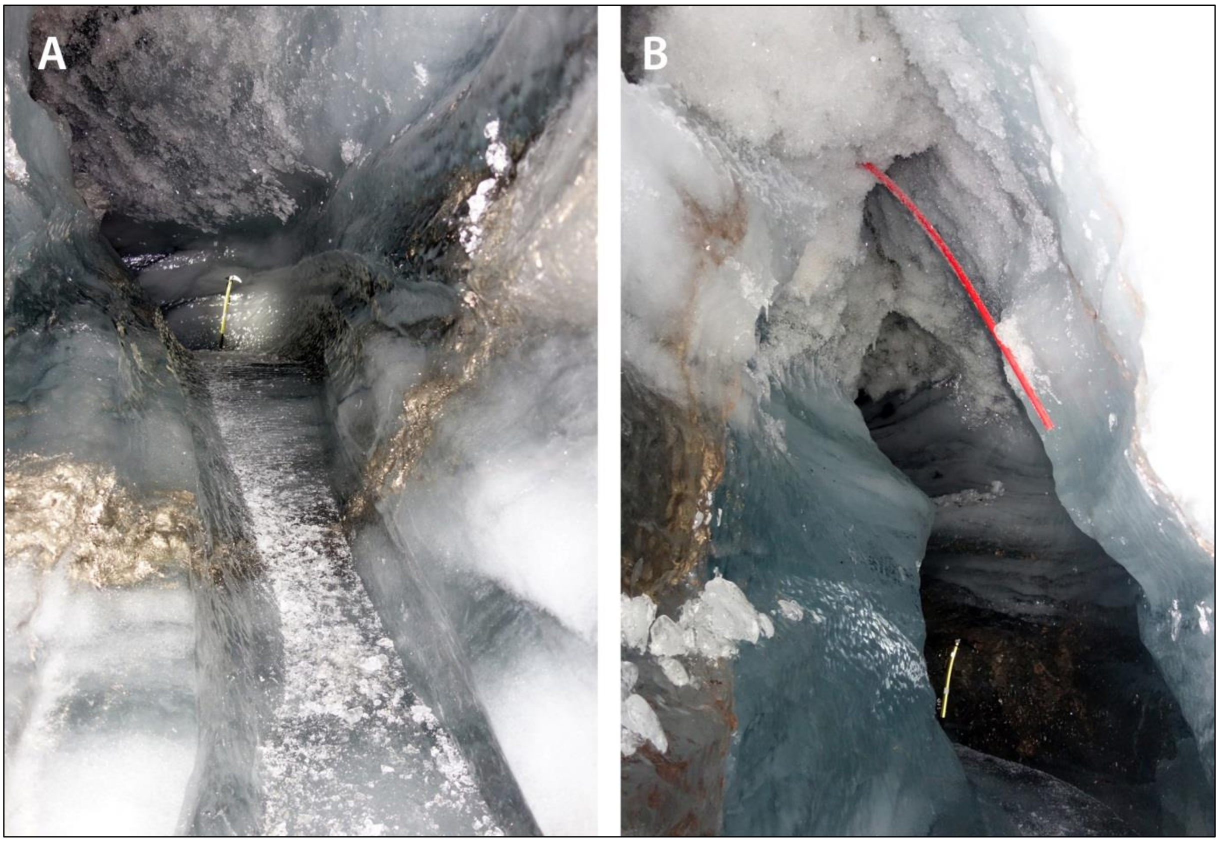

Figure 3.11: The bottom section following the step sequence, characterized by narrow stream widths, irregular meanders and irregularly sloping walls. A: Gently sloping channel segment, even narrower than the rest of the section. B: Frozen slush plug about 2 m over the channel floor, with a marking stick from the entrance.

3.3.2 Conduit surveys from 2002 to 2014

Figure 3.12 (next page) shows the development of the channel based on the surveys done from 2002 to 2014. The channel has been divided into five sections, based on a qualitative assessment on the breaking points between different channel patterns that are possible to recognize throughout the survey period.

Section I is the entrance, ending where the first step or steps reach a more flat part of the channel. Section II is a series of steps and horizontal channel segments following an irregular meandering pattern. Section III is a mostly horizontal channel reach, characterized by more or less big, well-developed meanders. Section IV is a step sequence, varying in height and pattern between the surveys. Section V is the bottom section, containing some smaller steps but mostly horizontal and irregularly meandering. A selection of parameters from the profile and planar patterns is presented in table 3.1.

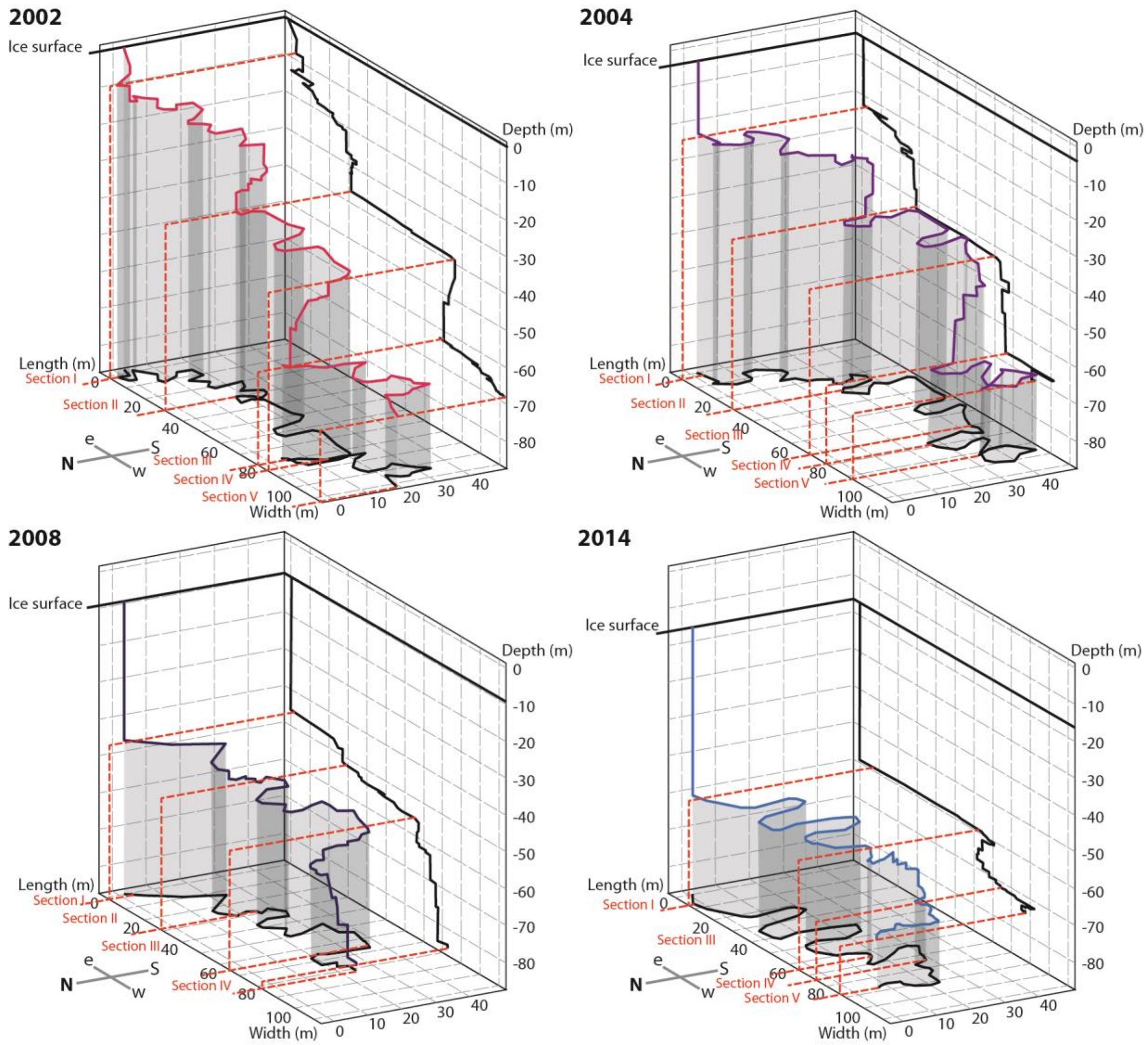

Figure 3.12: A model of the conduit as it was mapped in 2002, 2004, 2008 and 2014, divided into 5 sections based on visual changes in the channel pattern. Profile view on the back wall and planar view on the bottom.

Table 3.1: Planar and profile parameters for the sections from each survey.

		Planar form			**Profile**		
		Horizontal channel length	Straight line length	Sinuosity	Total elevation drop	Number of steps	Mean step heigth
2002	Section I	8	6	1,4	-10,1	2	5,1
	Section II	99	31	3,2	-28,4	14	2,0
	Section III	79	54	1,5	-1,9	-	-
	Section VI	19	15	1,2	-23,5	5	4,7
	Section VI	82	34	2,5	-5,0	2	-
	Total	286	138	2,0	-68,9	23	-
2004	Section I	-	-	-	-18,4	1	18,4
	Section II	89	37	2,4	-19,3	8	2,4
	Section III	104	45	2,3	-1,0	-	-
	Section IV	17	9	1,8	-23,5	4	5,9
	Section V	82	16	5,1	-3,2	1	-
	Total	292	107	2,9	-65,4	14	-
2008	Section I	-	-	-	-37,0	1	37,0
	Section II	54	36	1,5	-6,1	4	1,5
	Section III	78	39	2,0	-3,7	3	-
	Section IV	40	21	1,9	-29,0	11	2,6
	Section V	-	-	-	-	-	-
	Total	172	96	1,8	-75,8	19	-
2014	Section I	-	-	-	-43,0	1	43,0
	Section II	-	-	-	-	-	-
	Section III	148	67	2,2	-0,3	-	-
	Section IV	40	14	2,8	-12,0	8	1,5
	Section V	38	25	1,5	-2,8	3	0,9
	Total	226	106	2,2	-58,1	12	

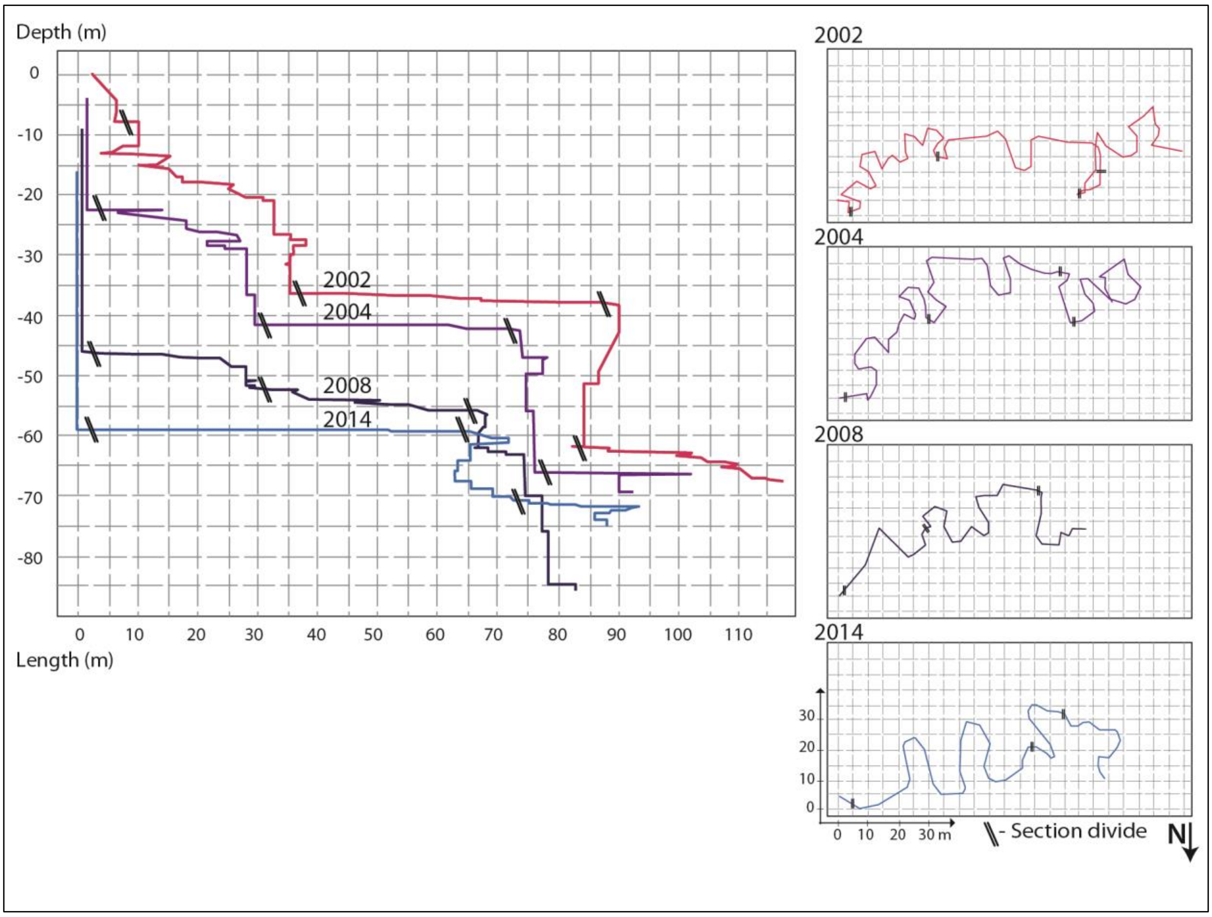

Figure 3.13: A compilation of the four englacial surveys from 2002 to 2014 in profile and planar view, with marked section divides. The crossing paths of the 2008 and 2014 surveys are most likely explained by the upstream migration of the entrance, which is not accounted for in the figure.

By showing all the surveys in one graph, as it is done in figure 3.13, it is possible to see the development through the survey period. The figure is corrected for the ice surface lowering between each survey, but it assumes a stable point location of the entrance which is most likely not be the case. Traces on the glacier surface shows that the entrance point has moved upstream during the survey period, but no accurate measurements of this exist.

3.3.3 Throughflow measurements

The return curve presented in figure 3.14 is sharply rising and falling, giving a single-peaked shape with very little tailing and no irregular features. The first traces of dye appear at the glacier front 93 minutes after injection into the stream feeding the moulin, then the concentration rapidly peaks and drops down into the background noise in approximately 40 minutes.

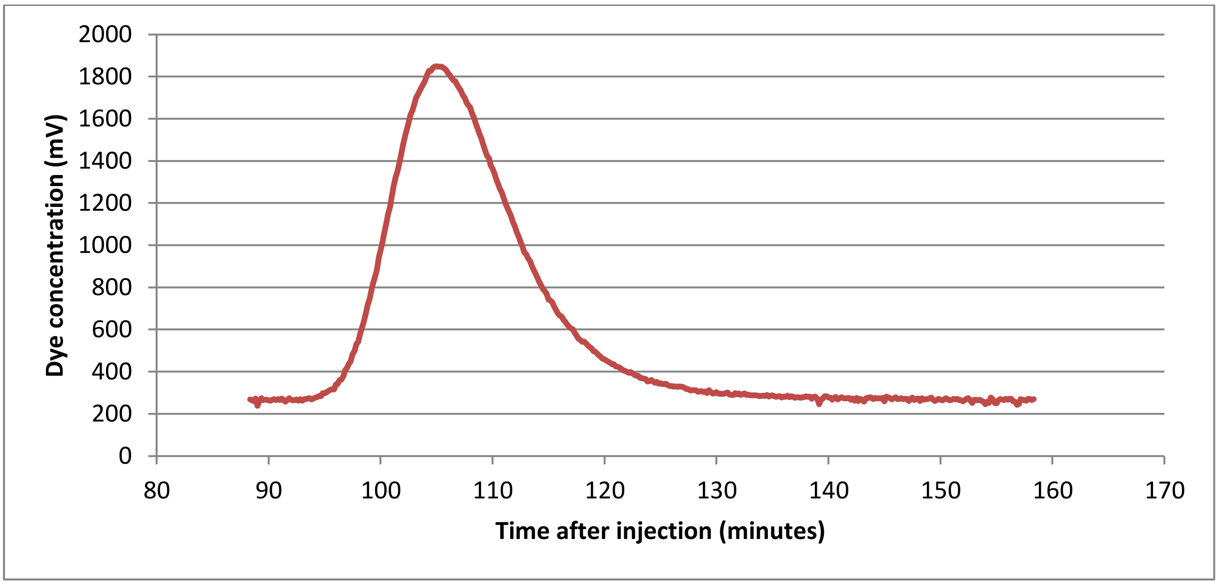

Figure 3.14: The return curve from the dye tracer injection on the 23rd of July, 2014.

The shape of the curve is reflected in the numbers shown in table 3.2. The calculated dispersivity and dispersion coefficient in the channel both lies well within the range of highly efficient drainage pathways presented by (Hubbard and Nienow, 1997). In total, there were three successful dye tracer experiments resulting from the fieldwork in the summer of 2014. An additional three experiments were brought from the study of Holtermann (2007).

Table 3.2: Throughflow parameters from the experiments in 2005 and 2012.

Year	Date	Distance	Discharge inflow	Throughflow velocity	Dispersion coefficient	Dispersivity
		(m)	l/s	m/s	m²/s	m
2005	07.08	1220	300	0,20	0,23	1,16
2005	12.08	1220	610	0,15	0,32	1,37
2005	13.08	1220	438	0,21	0,35	1,40
2014	21.07	1188	357	0,19	0,23	1,22
2014	23.07	1188	264	0,19	0,27	1,42
2014	25.07	1188	252	0,13	0,19	1,39

Temperature measurements done with the field conductivity meter shows a temperature of 0,1° C in both the inflow to the moulin and the outflow by the glacier portal. As the resolution of the temperature measurement is limited to one decimal place, any finer temperature differences could not be detected. One last, qualitative observation regarding the throughflow, is the difference in sediment content between the inflow and the outflow. Upstream from the moulin, the stream water was perfectly clear, with no visible suspended sediments (fig 3.15 – A). Where the portal stream surfaced from the glacier, still ~100 m before reaching the glacier bed at the margin, the amount of suspended sediment was so high that the water was brown and completely unclear (fig 3.15 – B).

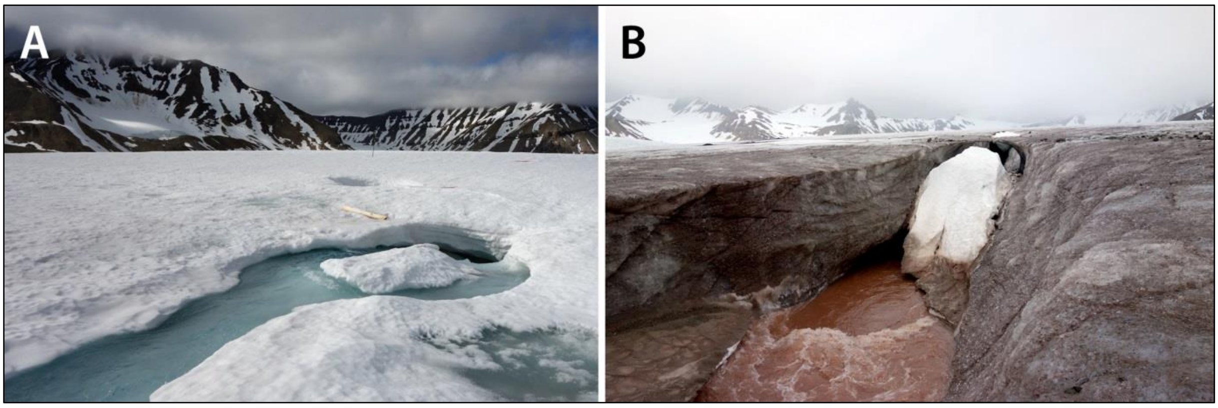

Figure 3.15: Entrance and exit of the englacial system. Note the difference in water colour. A: The snow-covered moulin, fed by a supraglacial stream. B: The melted-out portal stream, showing considerably higher discharge and sediment content.

3.4 Interpretation

The interpretation of the results will go through the development of the channel morphology section by section from the bottom and up, with the moulin presented in the end. Lastly the results from the throughflow measurements are interpreted and compared to similar studies, in order to approach a theory on how the rest of the channel is configured.

3.4.1 Development of the channel pattern

Section V

For the surveys of 2002, 2004 and 2014, the profile view of section V shows a clear trend. It is brought gradually closer to the entrance by steps migrating upstream, and gradually lowered by what can be incision, step migration or a combination of the two. No great change can be spotted in the planar patterns from 2002 and 2004, the irregular meander features move a bit but keep their form.

Section IV

Section IV is the lowermost mapped step sequence in the survey series. It is distinguished from its near horizontal neighbouring sections by having a steep gradient, going down tens of meters through a series of steps and pools. The steps follow a vague, irregular meandering pattern, with small and poorly defined bends compared to the upstream section. The planar pattern of the step section resembles the one of the downstream section V. From the four surveys in this analysis, the steps seem to follow the pattern of the upstream section, and as they migrate upstream they alter the planar pattern somewhat and takes up a more irregular meandering pattern with a smaller radius of curvature in the downstream section.

The profile pattern of section IV keeps its main form through the changes between 2002 and 2004. The whole section has incised a bit further down, and moved closer to the entrance, following the same trends as the rest of the channel profile. The start point of the step section has migrated 13,4 m upstream along the channel pattern, straightening out the biggest loop from 2002 and eliminating the smallest. Other than that the step section seem to be following the planar pattern of the channel, migrating into the upstream section.

Between the surveys of 2004 and 2008, the change is more prominent. The top of the section is 13 m lower in 2008 than in 2004, and from descending 22 m in 2004, the total descent in 2008 is close to 30 m. The measurements show the beginning of section IV to be considerably lower than the previous survey, starting 53 m further upstream along the channel pattern than it did in 2004. In a straight line distance from the entrance, the section has moved only 7 m closer. The meandering pattern seem to have moved downstream in the same period as the step section moved upstream, which makes it difficult to determine what proportion of the change is resulting from which process.

The survey from 2008 ends at a depth of 85 m below the 2002 ice surface, on a location where the ice thickness at that time should be between 65 and 85 m according to the bed topography map presented in figure 3.2. Bands of debris-rich ice were observed in the ice walls near the end of this survey, supporting the notion that it was halted very close to the bed.

After the sudden plunge in 2008, comes the strange switch to 2014. Section IV from this year starts approximately 7 m lower than in 2008, and descends gently downward in a sequence of steps no higher than 3 m. It flattens out and continues horizontally downglacier almost 15 m above the point where section IV from 2008 was halted. In the planar pattern it starts considerably higher upstream than in the previous study, and it follows the old pattern to some degree downstream from this point. The step section seems to be making the original meander sharper, possibly because the step-pool pattern distributes the energy differently along the channel walls than a horizontal meander.

Section III

The channel part upstream of the step sequence is characterized by being almost completely horizontal, with a well-developed meandering pattern. It is delimited by a stepped section at each end in all years of survey except 2014, in which it extends directly from the moulin. In the profile form, the section changes little between the surveys. It is gradually moved closer to the entrance by the recession of steps at each end, and though some gently sloping sections make it less horizontal in 2008 it stays mostly flat.

The planar pattern is characterized by big, well-developed meanders, which grow in amplitude and get a more defined form with each succeeding survey. The planar form is kept well enough to recognize the big features through the whole survey period. The whole section is lowered between each survey, without being broken up by steps, indicating that downward incision along the channel bed is the main form of channel lowering in the section. If the upstream migration of steps should be a part of this lowering, the steps would have to migrate through the whole 80 to 140 m of section III to create this pattern.

Based on the change in the sections between the four surveys, it appears that the downward incision of the stream works at rates of ~3 m per year, while the upstream migration of section IV moves at rates of 6,5 to 7,5 m per year. The rate of downward incision of the section shows a decreasing trend through the survey period. In the last six years, the section was lowered by a mere 5 m. A few meters of this might be accounted for by errors in the measurements, but not enough to set the latest survey at the depth that would be expected if the rates of 3 m per year continued. An explanation for this could be the low gradient and great channel width of the section measured in 2014. In accordance with equation 2, which was presented in section 2.3.1, this gives less available energy spread over a greater wetted perimeter, both factors to lower the melt rates.

Section II

Section II, which is the first section after the moulin in all the surveys previous to 2014, could have been divided into even smaller sections due to its irregular nature. In the first years it is characterized by a series of small steps and horizontal segments, following an irregular pattern of small meanders and ending in a steeper sequence of steps. It is possible to see the development of the meanders between the surveys of 2002 and 2004, and from there to the 2008 survey where only one big meander remains of the previous pattern. Something similar appears from the development in the profile pattern. The step sequence is reduced to a couple of short steps in 2008, and completely obliterated in 2014 where there is no basis to define a section II at all.

As with the pattern in section IV, it seems like the biggest steps stay at the same straight-line distance from the entrance. Whether it is channel incision or step migration that causes the section to lower is difficult to determine from the dataset, it is likely to be a combination of both. The step sequence at the end of the section migrates ~5 m closer to the entrance between 2002 and 2004, but the steps that still remains in 2008 have not moved much in the terms of straight-line distance. The step migration rates through the planar pattern are difficult to measure, as the meanders change too much through the survey period.

The main part of the section is lowered by ~10 m from 2002 to 2004, then 25 m from 2004 to 2008. Compared to the rest of the channel, this section undergoes extreme changes in the profile pattern through the period of surveying. Its highest point in 2002 is lowered by more than 50 m in 12 years, making an annual downwards movement of 4 m per year. This development is closely connected to the formation of the moulin, which could be included in this section but has been put in its own for the sake of the discussion.

Section I

The entrance to the conduit, section I, is the one to have undergone the most striking change since the surveys started. It was described by Vatne (2001) as a tunnel-shaped conduit sloping gently into the ice, following the direction of ice flow. Its slope increased after ~40 m, when the step-pool form took over the channel pattern and the conduit height increased (see fig. 2.7 in the theory chapter). In 2002 the entrance was steep, described by Refsnes (2004) as two rapids separated by a vertical step. It descended to 7 m below the ice surface this year. In the following surveys the entrance was vertical, increasing in height to 37 m in 2008, and 43 m in 2014. Though it was described as a vertical moulin in the survey of 2008 (Vatne and Irvine-Fynn, in preparation), the completely vertical, round shafts ending in a circular plunge pool was not observed before 2014.

Through this development, the entrance point moved very little on the glacier surface. Detailed GPS measurements from the surveys before 2014 does not exist, but from the traces on the glacier surface the channel does not appear to have moved much. The observations done in the moulin in 2014 indicates an upstream movement of <10 m, resulting in the divided morphology of the upper parts of the main shafts.

The dataset shows a moulin formation which is independent from crevasse structures, formed by the channel development happening after the channel has been cut off from atmospheric influences. The big enigma is how the entire length of section II has been lowered by 40 to 50 m through the survey period, while no incision or step migration has caused the entrance point to follow this pattern.

General development of the profile and planar pattern

Looking at all the sections together, it is possible to point out some obvious trends, or lack of such, in the development through the survey period. With the coarse temporal resolution of the dataset, it is difficult to say anything about the migration rates of individual steps. The and average step height and number for the step sequences are varying without any increasing or decreasing pattern. For the three first surveys, the step sequences seem to stay in the same zones measured in straight-line distance from the entrance. They do migrate upstream, but at the same time the meanders in the horizontal sections migrate downstream, so the migration rates found through the surveys are not necessarily accurate.

There is a trend going towards more of the total descent happening in the moulin while the rest of the channel flattens out. In 2014, the 43 m deep moulin leaves only 15 m of lowering to the remaining channel, which is distributed among one grouped sequence of smaller steps. Step migration might be working together with channel incision to develop the moulin, but the major part of the channel lowering seems to be a result of increased rates of incision near the entrance.

In the planar form the pattern seems to go towards big, regular meanders with increasing amplitude. The meanders migrate downstream, and the sinuosity increases slightly with time. The planar pattern is disturbed as the downstream step sequences migrate through it, though the main features are still recognizable once the sequence has passed. The main direction of the channel is similar each year of survey, with minor displacements. This is in accordance with the direction of ice flow and the slope of the bed topography, and ~90° off the direction of the portal stream. Whether this is controlled by the inherited pattern of the original supraglacial stream, structures in the ice, or bed topography, will be assessed in the discussion chapter.

3.4.2 Characteristics of the throughflow measurements

The result of the throughflow measurements of 2005 and 2014 are fairly similar, suggesting no big changes in the overall system between the two periods. The symmetric, peaked return curve with almost no tailing (fig. 3.14) indicates effective drainage, with low roughness and no significant temporary storage features. This fits the measured part of the system, which has fairly high sinuosity but negligible channel roughness. It implies that the pools under the taller steps, even the big pool under the moulin, does not retain the stream flow much.

The peaked return curves also indicates that if the conduit reaches the bed at all, there is no roughness caused by boulders and bed sediments as in the study of Gulley et al. (2012). With the geology in the area, it is likely that the bedrock and sediments under Austre Brøggerbreen would create similar conditions to Rieperbreen, which is the location for the channel in the study of Gulley et al. If so, it should have caused the curve to flatten out a bit, giving it a higher dispersion coefficient, and at least have given it a longer tail. Neither of these can be seen in the results from 2005 and 2014.

The dye tracer experiments from 2005 and 2014 show only minor variations in dispersivity and dispersion coefficients, both in the surveys within the same year and between the two years. Although the discharge at the injection point is more than doubled at its highest compared to its lowest, this does not have any consistent impact on the velocity or the calculated values.

In comparison, Gulley et al. (2012) presented dispersion coefficients ranging from 0,39 to 2,33 m^2/s, and dispersivity values from 1,5 to 19,89 m, all in one channel through one season. The velocity showed a positive correlation to discharge, giving a negative correlation between discharge and dispersivity. The results were interpreted to reflect how the flow got less delayed and dispersed by the channel roughness as the flow height increased in relation to the boulders covering the bed.

These patterns were not found in the dataset from Austre Brøggerbreen. It is also interesting to note that even at considerably higher discharges and velocities than measured through channel B, not even the lowest dispersion coefficients and dispersivity values from Rieperbreen are as low as the ones presented in this study.

The question then is whether the conduit reaches the bed at all, or if it runs in the ice all the way to the portal. The portal stream exits several meters above the bed, so it is obvious that it runs englacially at least for the last stretch. The sediment content in the portal stream suggests that the water runs in a sediment rich environment at some point. This does not necessarily originate from the bed, it could come from sediment rich bands near the glacier bed which has been observed in the channel extending from moulin A (Vatne, 2001). The portal stream drains water from all the known moulins on the glacier, so it is not possible to say which stream contributes with the sediment and colour to the outflow.

Based on this, it seems unlikely that the channel runs along the bed, at least not for any significant length. However, more dye tracer experiments should be performed to support this interpretation, through the ablation season and at different discharges.

4 Discussion

The first part of this discussion focuses on the energy budget in englacial channels, approaching it in general terms with examples drawn from the 2014 survey of channel B in Austre Brøggerbreen. This will then be used to explore and explain the development of steps and meanders in englacial systems in the next part of the discussion. Then follows a section that builds further on both of the former, drawing in the available results and literature to construct theories on the formation of englacial channels and the development of moulins in the absence of crevasses. In the last section the shortcomings of this study will be listed, along with aims and ideas for further research to strengthen the knowledge of the topic.

4.1 The energy budget in englacial streams

A central idea behind the formation and development of fluvial channel patterns is that the stream will adjust towards a state of equilibrium in which some degree of stability is established (Chang, 1986). Following the theory of minimum rate of energy dissipation (Yang and Song, 1979), equilibrium in this setting is when the energy dissipation is at its minimum value. In alluvial channels this involves adjusting channel geometry and slope in order to accommodate the discharge and sediment supply of the catchment, and to do so in the most efficient way in terms of energy (Leopold and Langbein, 1962; Leopold and Maddock, 1953).

In an ice-walled stream, where thermal erosion is the main way of channel development, the most stable state of flow is one where the temperature difference between water and ice is minimal. Isenko et al. (2005) termed this state the equilibrium temperature, and found slope and hydraulic radius to be the main controls of its value. The relationship is shown in figure 4.1. Following this figure the present equilibrium temperature in the first part (section III) of the channel is very close to zero, between 0,01 and 0,04° C, as the stream width is ~1 m and the slope is <2°.

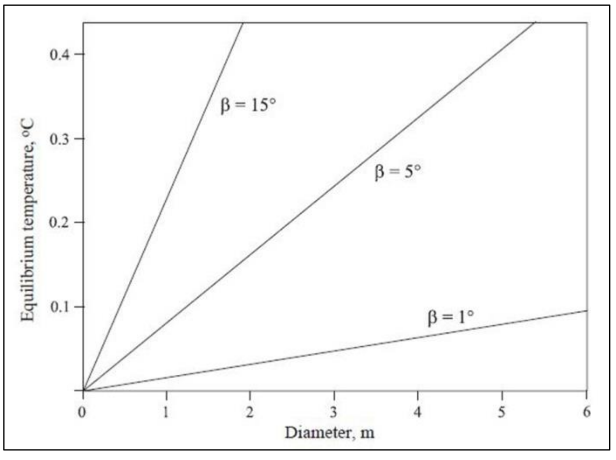

Figure 4.1: The calculated relationship between slope, velocity and stream size. From Isenko et al. (2005).

The water temperature in the supraglacial stream feeding the moulin was measured to be 0,1° C. Following the law of conservation of energy, an additional temperature increase of 0,1 °C is gained from the loss of potential energy in the 43 m elevation drop in the entrance moulin. This means that the water flow, upon hitting the bottom of moulin B, has a temperature which is relatively high compared to the predicted equilibrium temperature. The stream flow will then adjust toward the equilibrium temperature by releasing excess energy to the surrounding ice walls, causing the ice to warm and melt.

Equation 3 in section 2.3.2 shows that the transfer of heat through the ice-water boundary will be greater with higher temperature difference. Based on this, the highest rates of energy transfer will be close to the source points of the thermal energy. The rates of energy transfer, and thus thermal erosion, will then decrease exponentially with downstream distance from these source points (Isenko et al., 2005). This means that the greatest rates of channel pattern development will be in a downstream proximity to the energy source points. In simple terms, the stream flow has the greatest ability to adjust towards a more stable channel pattern in the areas where it is furthest away from equilibrium.

4.2 The behaviour of meanders and step sequences

In this section, the attempt is made to get a basic understanding of the morphology and evolution of the surveyed conduit in Austre Brøggerbreen, based on where and how the energy from the stream is used in the channel. An underlying assumption for this is that the ice in which the stream runs is near homogeneous, with little or no structural control on the channel pattern The assumption is made in this study based on observations of the stratigraphy in the englacial channel walls, which showed no connection to the channel pattern. In addition, as will be discussed further in section 4.3, the past and present channel patterns seem to be evolving from an inherited surface pattern rather than following englacial tectonic structures.

4.2.1 Meanders

In alluvial streams meandering is considered a stable channel pattern, used to decrease the slope and increase the flow resistance between two points (Chang, 1979; Langbein and Leopold, 1966). For ice-walled streams the function is expected to be the same, but the high rates of erosion compared to alluvial streams give a much faster development and accompanying adjustment to the discharge.

Meandering in terms of energy balance

Considering the controls on equilibrium temperature and thermal energy transfer, there are two reasons why meandering is a stable channel form in ice. First, the decrease in slope that follows increased sinuosity lowers the melt rate and the equilibrium temperature considerably (fig. 4.1). These factors serve to lower the incision rate of the stream, giving a more stable channel pattern. Second, meandering reaches have been measured to have smaller variations in friction factor and shear stress than straight reaches in the same stream (Dozier, 1976). This gives a more uniform distribution of thermal erosion in the channel, serving to enhance the channel pattern rather than disturbing it.

The measurements done in Austre Brøggerbreen in 2014 showed the stream width to be constant through the curves and the stretches between them, with a pool at the bend exit (see fig. 3.7). This indicates that most of the energy generated by turbulence in the bend is dissipated in the pool, and the straight channel segments separating the bends can assume a stable flow between the turbulent curving channel parts. This makes the meandering pattern a very stable one in terms of energy balance, with only short lengths of turbulent flow followed by an rapid adjustment back to more stable conditions.

Meandering development of channel B

The following part of the discussion is based on the upper part of channel B in Austre Brøggerbreen, with the main focus on section III from the 2014 survey and the pattern development of section II and III from the previous surveys.

The planar pattern of channel B shows meanders that grow in wavelength and amplitude. Based on earlier observations and modelling (Dozier, 1976; Karlstrom et al., 2013; Marston, 1983) the meanders were expected to migrate downstream, but it is difficult to spot any clear trends in the dataset presented in this study. The movement of the meanders seems to be a product of their growth in size and amplitude, with no clear pattern of upstream or downstream migration. Over time, the increase in wavelength is accommodated by the smoothening out of the smaller meanders.

Meander wavelength has been related to stream width in studies done over a range of environments, including glacier surfaces, presenting wavelength to stream width ratios between 10 and 14 (Leopold and Wolman, 1960; Williams, 1986; Zeller, 1967). The near-horizontal, meandering part of channel B is within this range, with ratio of 13. The ratio of radius of curvature to stream width also coincides with previous studies, curiously landing on the exact mean value from the study of Williams (1986) with a value of 2,43.

With these numbers, the meandering development in the channel appears to be strongly related to the stream width in the channel. The pictures taken in the survey in 2002 (Figure 23 and 27 in Refsnes, 2004) and 2014 (Figure 3.9 - A) show that the channel width in section III has increased considerably in this period. At the same time the survey maps show increases in wavelength and radius of curvature, which are almost double in 2014 compared to 2002.

The stream width seems to be the control on the meander development, but the controls on stream width are harder to pin down. In the entire surveyed length of channel B, the measured channel widths show a weak negative correlation to slope (fig. 4.2). As the discharge is the same through the surveyed part of the system, except for a small increase due to wall melting, this indicates that the flow velocity holds some control on the channel width. There is also a weak negative correlation between channel width and distance from the entrance (fig. 4.3). This indicates a relation between channel width and energy dissipation rates, which are higher closer to the surface entrance and the moulin. It can also be affected by ice creep, which will increase with the overlying ice mass (Irvine-Fynn et al., 2011).

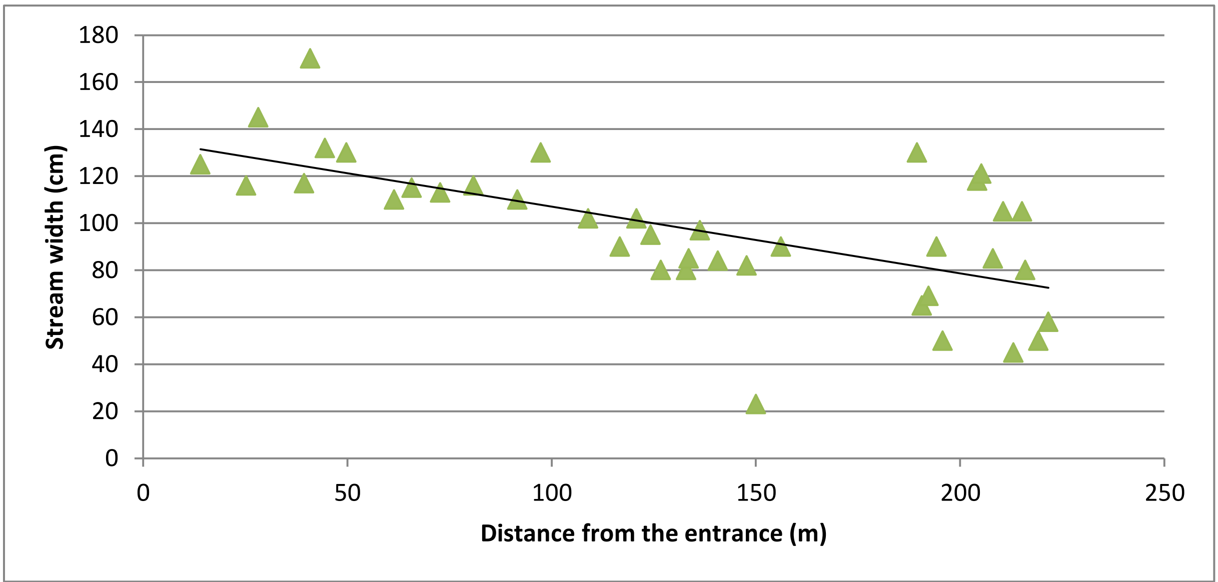

Figure 4.2: Correlation plot of the measured stream width to slope in the 2014 survey of channel B.

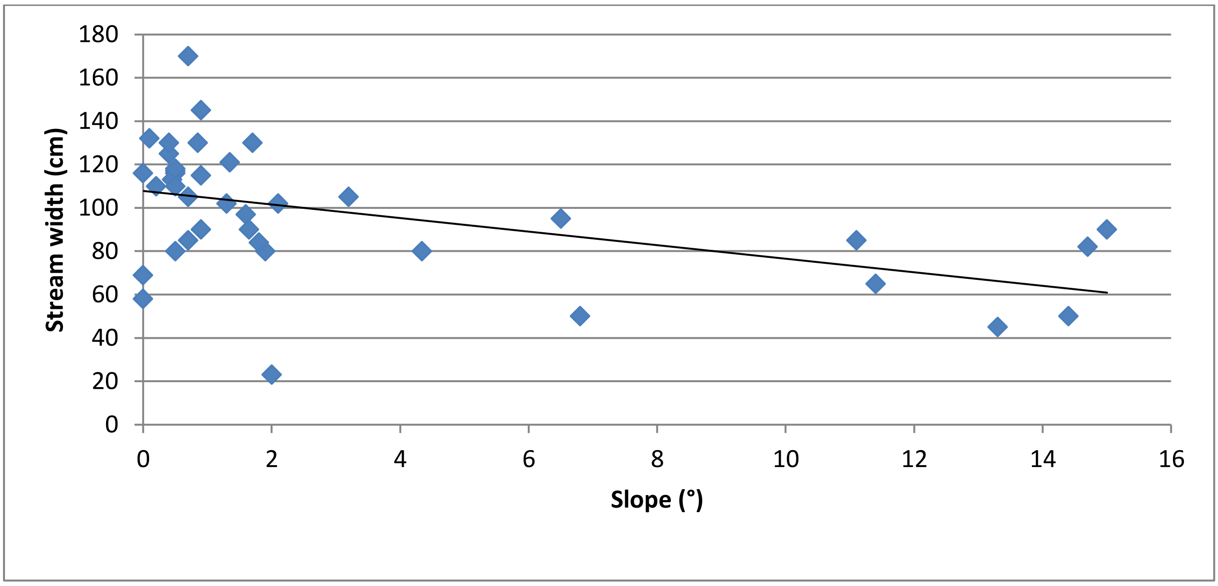

Figure 4.3: Correlation plot of the measured stream width to distance from the entrance in the 2014 survey of channel B.

Looking at the development of the upper part of channel B, it seems like the channel has incised mainly downwards until the initial slopes and steps of section II has become level with the downstream near-horizontal reach. The channel width has increased with decreasing slope, with an accompanying increase in wavelength and radius of curvature.

From the 2008 survey and onwards, the main form of meander development is an increase in amplitude. According to Leopold and Wolman (1960), amplitude is determined more by erosion characteristics than by hydraulic principles. They argue that the amplitude is limited by the formation of meander cut-offs. This process is not very likely to happen with the present channel morphology of section III, which is near horizontal and has channel walls of solid ice that extend up to the conduit roof. As there is no other apparent control on amplitude, only continued surveys can show how far the development can continue in this environment.

It appears that the meanders in channel B have developed to a very stable channel pattern. Downstream of section III however, the trends in channel width and meander form are disturbed by the channel lengths where the flow is pooled or running down steep segments. The flow through steps and pools introduces its own controls on the channel pattern, which will be discussed in the following section.

4.2.2 Steps

It is clear from the previous paragraphs that a stable, high-amplitude channel pattern is not very efficient when it comes to channel descent. With the entire channel pattern continuously incising downwards, however, significant differences in elevation must occur somewhere in the channel.

Steps in terms of energy balance

The definition of steps used in this study follows that of Vatne and Irvine-Fynn (in preparation), in which steps are described as channel segments where the change in elevation is greater than the change in the concurrent horizontal distance. The slope of channel segments fitting this criterion gives high velocity and high Froude and Reynolds numbers, making both flow and channel pattern unstable. The increased amount of thermal energy available for channel development also means that the channel pattern can rapidly adjust into something more stable, and the result seems to be the alternation of steps and pools.

At sudden decreases in slope, the change in Froude number leads to a hydraulic jump that releases energy to the channel boundary through turbulence (Church and Zimmermann, 2007). This process develops pools at the foot of every step, which as in the case of the meanders serves to dissipate the energy created by the velocity and elevation loss in the step locally. Vatne and Irvine-Fynn (in preparation) found the channel to resume the same width downstream of the pools as it had upstream of the connected steps, indicating that most of the energy generated in the step is dissipated in the pool.

In terms of energy balance, this means that the step-pool regime allows the channel to lose a lot of elevation in a short distance, concentrating the high energy flow and turbulence to a small area.

Step-pool development in channel B

The most obvious trend in the steps of channel B is that they exist mainly in clusters. In all four surveys, all the steps exceeding ~1 m in height are grouped in sequences of steps separated by pools. The whole sequence migrates upstream along the channel pattern, and keeps the alternation of steps and pools although the step heights and pool sizes vary. A curiosity from the survey done in 2014 is that all the steps, and most of the gentler sloping segments, were connected to an upstream pool. For the step sequences this is to be expected, but there were pools also upstream of the steps that did not follow another step. This seems to be related to the curvature of the channel, as the steps and sloping segments for the most part were located at the bend exits.

The upstream pools serve to categorize the steps as type 2 in the Vatne and Irvine-Fynn (in preparation) classification. The lower part of the steps will incise upstream faster than the upper, as the energy is highest at the bottom of the step. This is expected to be balanced out by the decreased incision rates as the step grows sufficiently steep to detach the flow, in which case the upper part of the step will catch up with the lower and the steepness be restored to a condition where the flow is in contact with the entire step face.

An open question is whether there is something in the relationship between the steps and the pools in the step sequences that keep the steps in clusters, rather than merging them or scattering them out as one would expect based on differential migration rates. The dataset presented in this study lacks the temporal and morphological resolution to answer this question, but it also lacks any signs of a trend towards merging or scattering of steps.

What the dataset shows is that the start of the step section closest to the entrance migrated 3 m closer to the entrance along the channel pattern from 2002 to 2004. The movement in straight-line distance was a bit longer, but the increase in upstream sinuosity lowered the net migration rate. From 2004 to 2008 most of the steps were gone. The steps that remained in 2008 started 44 m closer to the entrance than in 2004, but the upstream sinuosity had decreased in the meantime so it is difficult to say how much of this was due to step migration. The straight-line distance from the entrance did not change much between the two surveys. In the 2014 survey, there was no trace of the step sequence left.

In section IV, which is the lower step sequence in the surveys, the changes in planar pattern are lesser so the migration rates are easier to measure. The start of the step sequence migrated upstream at rates of 6-7 m per year through the survey period. This brought the sequence closer to the entrance point, but the process was counteracted by the increase in sinuosity in the upstream meandering section. In the 2014 survey, 148 m of near horizontal channel separated the moulin from the start of the lower step section.

The relationship between the step-pool pattern in the profile form and meandering in the planar pattern adds to the complexity of the system. The meanders keep a flow pattern that aims to increase the amplitude and maintain a wavelength which is consistent with the stream width. The steps and pools introduce an extreme fluctuation of channel widths to this system, and are in turn influenced by the curved planar form they inherit from the meanders they migrate along. Looking at the planar maps in figure 3.13, the biggest changes in planar pattern happen where a step sequence has passed through.

The coarse temporal resolution of the dataset prevents any strong conclusions to be made on the behaviour of the step sections in channel B. Based on these four surveys, it seems that although the migration rates of the steps exceed the downward incision rates of the near-horizontal reaches, the lengths of the latter in addition to the increase in sinuosity dampen the effect of step migration on the overall channel lowering. Vatne and Refsnes (2003) and Gulley et al. (2009a) reported the highest rates of incision to be where a step had migrated through the measured cross-section, showing that the migration of steps is a major local factor in channel lowering. However, the multi-year surveys of channel B show that where the steps are located in clusters separated by long reaches of near-horizontal channel, the downward incision of these in-between reaches can contribute with an equal part of the total work of channel lowering.

One of the main points of subchapter 4.1 was that the rates of energy dissipation in the stream flow will be highest near the source points, and that the entrance moulin of channel B is the most influential source point of the surveyed part of the system. Differential incision rates based on distance from this point could help explain the levelling out of the upper part of the channel, as the uppermost reaches would have the highest rates of incision and thus catch up with their downstream neighbours. A combination of this process and step migration would be more likely than the split-up of step sequences and sudden jumps in migration rates needed for the step migration alone to account for the channel lowering.

4.3 Theories on the processes and formation of englacial channels in uncrevassed ice

With the above subchapters in mind, the discussion will now turn towards exploring the mechanisms of englacial channel formation and development. An attempt will be made to describe how the channel evolved from a supraglacial to an englacial channel, and how a gentle slope can turn into a vertical moulin feature. The section also concerns the further development of the channel, based on its behaviour till now. In the last part, a brief assessment is done on the configuration of the channel downstream of where the surveys ended.

4.3.1 The origin of channel B, and the development of a moulin

Based on the pattern of channel B and the absence of any significant tectonic traces on the surface and in the channel walls, the englacial conduit is considered to have originated from the cut-and-closure of a supraglacial channel. This conclusion was reached by Vatne (2001), after the first survey of the channel. On the glacier surface surrounding the moulin (fig. 4.4) there are several traces of channels draining in the same direction as the present englacial channel, supporting this theory.

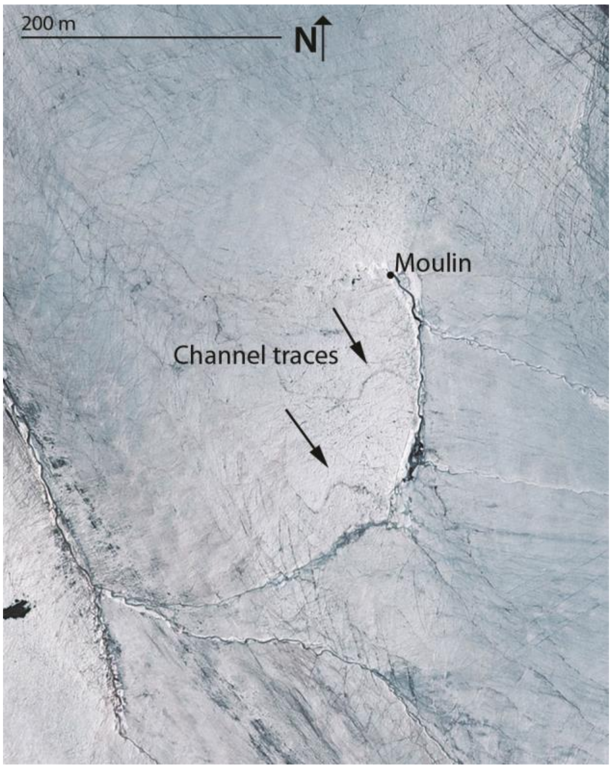

Figure 4.4: Aerial photograph of the glacier surface surrounding moulin B. Source: toposvalbard.npolar.no.

According to the model of Jarosch and Gudmundsson (2012), the gradient of the glacier surface is a major influence on the depth and time step at which a supraglacial channel gets pinched off from the surface. A substantial increase in slope is present on the glacier surface just after the bulge in the ice surface above which moulin B is situated. Judging from the channel traces in figure 4.3 the original supraglacial stream found the lowest way around this bulge to the west. The steeper slope then caused it to incise deep enough to be cut off from the surface, most likely by a combination of snow infill and ice creep.

The supraglacial stream feeding moulin B was followed upglacier during the summer fieldwork in 2014. At significant increases in slope, the channel had incised accordingly deeper down into the ice, some places to depths of several meters. At gentler slopes even further upglacier, the flow had the same shallow channel as just upstream of the moulin. Although this has not been measured or quantified in any way, it shows how a constant discharge can incise to different depths depending on the surface slope.

Once cut off from atmospheric influences, the inherited slope and planar pattern of channel B was reworked by the englacial channel-forming processes discussed in the previous subchapters. As the original slope must have been steeper than the upstream channel reach, which to this date still has not incised to any great depth, it is reasonable to believe that the incision rates continued to be greater in the newly formed englacial channel. This led to a continued increase in the depth difference of the supraglacial and englacial parts of the channel.

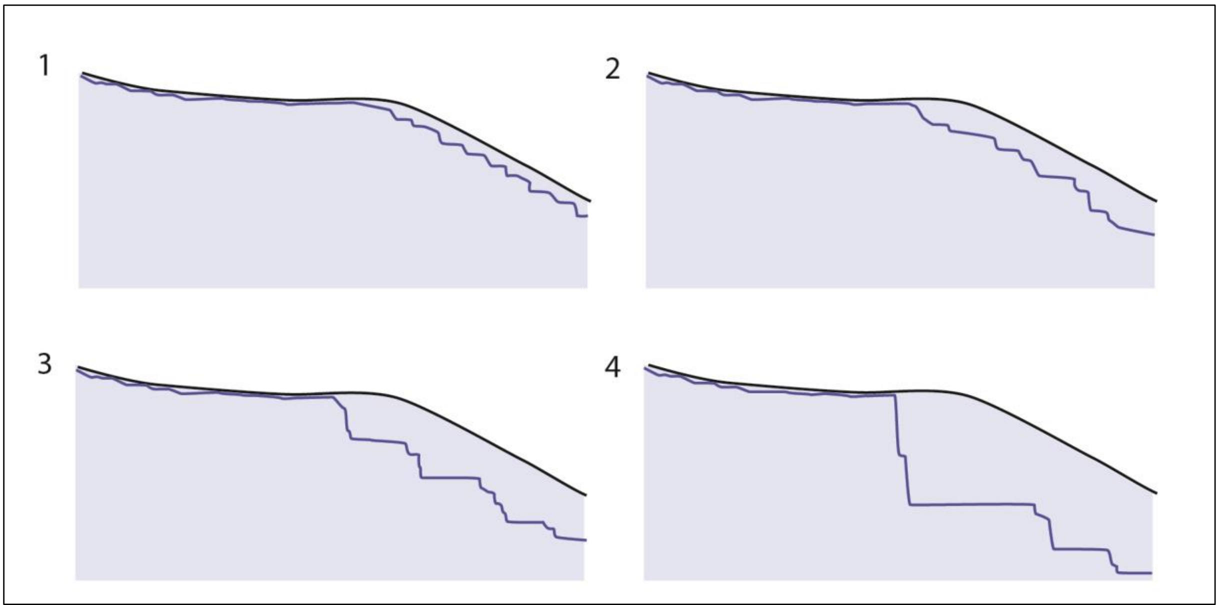

Figure 4.5: Conceptual diagram of the formation of an englacial channel by cut-and-closure, showing increasing differences in depth due to higher values of slope in the englacial part of the channel. The englacial entrance moves somewhat upstream, but the migration is slowed down as it reaches a critical steepness. As more energy is available for channel adjustment in the upper part of the englacial channel, this is where most of the development happens.

The course of englacial formation shown in figure 4.5 shows an initial englacial channel that runs along a shallow track, following the gradient of the glacier surface. Where the glacier surface flattens, the conduit is expected to do the same. If this is the case for channel B, the steep surveyed profile reflects the steep gradient just downglacier of the englacial entrance.

Following the profile of the glacier surface, the major part of the rest of the channel should have a considerably gentler gradient.

At present, the entrance to the englacial channel is situated in a depression on the glacier surface. The shape of the moulin and the traces on the surrounding ice surface shows that it has migrated slightly upstream, but not nearly at the rates of 6-7 m per year that was measured in the englacial step sequences. With its present vertical plunge, even overhanging at the bottom, the theory of Vatne and Irvine-Fynn (in preparation) predicts the jet flow over the step to detach from the step face, halting the upstream migration. However, the initial englacial entrance, mapped by Vatne (2001), was gently sloping. Something must have caused it to steepen to the point of jet detachment in the first place.

The growing difference in depth between the supraglacial and englacial parts of the channel, in combination with the process of step steepening described by Vatne and Irvine-Fynn (in preparation), is likely to have contributed to the development of the moulin. In addition, if there was no pool or curve upstream of the initial step to decrease the velocity of the stream, the jet would detach sooner, and stay detached, from the step face. Once the upstream migration of the top of the step was slowed down, the step would continue to grow in height as the channel below it was lowered by downwards incision and the upstream migration of the lower-lying steps. The taller the step, and the longer the jet stayed detached, the more the moulin would have enhanced and preserved its form. The relatively high rates of energy dissipation near the entrance point are also an aid to the moulin formation, with increased rates of channel development compared to the rest of the channel.

4.3.2 The continued development of channel B

The trend in channel development until now has been that more of the total descent happens in the moulin, leaving the remaining length of the channel in a near-horizontal meandering pattern. In theory, this development can continue until the moulin reaches the bed, and the sediments at the bed restrict the incision rates until the channel follows the profile of the bed topography with a greater or lesser degree of planar pattern inherited from the englacial processes. A condition similar to this was mapped by Naegeli et al. (2014) in Tellbreen, a glacier of similar characteristics as Austre Brøggerbreen. Among the surveyed channels presented in their study was a 32 m high moulin draining directly from the surface to the bed, continuing downglacier in a large N-channel incised into the underlying sediments and bedrock. The other surveyed channels showed dendritic, meandering channel patterns, all interpreted to be formed by the cut-and-closure of supraglacial channels.

Without knowing the exact relationship between incision rates and channel pattern, it is difficult to say much about the time scale in the evolution of neither the Tellbreen channels or channel B in Austre Brøggerbreen. In channel B, the development from gentle slope to 43 m high moulin has happened between the first survey in 1998 (fig. 2.7) and the last in 2014 (fig. 3.13). The rates of development were highest at the beginning, slowing down towards the present. Whether the decrease will continue, stabilize or even accelerate in the future is not possible to say at this point. When and whether channel B will reach a channel pattern similar to the moulin in Tellbreen is equally challenging.

4.3.3 The configuration of the channel downstream of the surveyed part

The englacial surveys of channel B cover only a small fraction of the total channel length, around 100 m out of the 1200 m straight-line distance between the moulin and the glacier portal. Several dye tracer tests have been run through the channel, but the high sinuosity and complexity of the surveyed part of the system only serve to emphasize how ambiguous the results from these tests can be.

A curious point emerging from the dye tracer tests is that the channel does not appear to reach the bed at all. This is strange because it would imply that most of the descent towards the bed happens within the short range of channel that has been surveyed. If this is the case, the channel runs close to the bed, but does not reach it, for a distance of more than 1000 m even when sinuosity is not considered. The question then is why the downwards incision, which is >1 m per year in the surveyed part of the channel, stops without reaching the bed or any other layer of sufficient roughness to show up in the throughflow curves of the dye tracers.

The same puzzling results were found by Hooke et al. (1988) on Storglaciären in Sweden. Like in Auster Brøggerbreen, the ablation area of Storglaciären is situated on top of an overdeepening in the bed topography. Hooke et al. (1988) concluded that the flow through the overdeepening was englacial, due to the substantial part of the energy dissipation that would be necessary to maintain the flow temperature at the pressure melting point when ascending the slope up from the overdeepening. This would inhibit any enlarging of the conduit, and any subglacial channels that formed would tend to freeze shut. An additional factor in Austre

Brøggerbreen is that the system appears to be vadose, and evolves by downward incision. This makes it unlikely to go into the overdeepening at all, as the local base level downglacier of the overdeepening would prevent any further incision.

The reason why the channel does not reach the bed in the overdeepening might thus be covered, but a question still remains about why it does not incise down to the bed downglacier of it. Neither local base level or up-slope freezing would prevent this, yet the channel exits the ice more than 10 m above the glacier bed. A survey of the lowermost channel reach could perhaps more light on this question. More surveys, reaching deeper down into the system, are also needed to explain what is happening in the depths of channel B.

4.4 Shortcomings of the study methods, and challenges for further research

The above discussion leaves many questions unanswered, and involves many assumptions that need confirmation and backup from field data in order to be valid. One of the main shortcomings of studies like this is that the englacial system is only accessed in winter, when the stream is frozen. The flow conditions can only be inferred from the morphology of the inactive channel, and to some degree from throughflow measurements. None of these give any details about local flow conditions through single features of the channel pattern.

A major problem with winter measurements of the channel morphology is that the plane of measurement is the frozen water surface of the stream, and not the thalweg or the channel walls. There is no way of finding variations in flow depths with this method, and this makes it difficult to understand the full scope of the flow conditions.

The solution to these problems is very straightforward and very difficult at the same time, namely to access the system in summer when the flow is active. As this would involve rappelling down 43 m of active moulin and land in a pool which most likely reaches a depth of several meters, there are several reasons of practicality and safety which render it unlikely to happen. With the lower rates of incision in the near-horizontal channel reaches, however, it is theoretically possible to rig a system of probes in winter that would measure the flow through the ablation season. Simple measurements of flow depth and temperature would strengthen the platform of knowledge considerably, especially if taken at several distances from the entrance.

Another shortcoming with the cave surveys is that in most cases, and in all cases in channel B, they are stopped before reaching the bed or anything near it. To understand the development of the channel it is necessary to see the whole descent through the ice thickness, that would most likely help to explain the movement of the step sequences and perhaps even the strange development between the 2008 and 2014 surveys. A better understanding of the englacial water table and the freezing of the stream could improve the timing of the surveys, coordinating them to the time of year when the conditions allow deeper access to the channel.

The simplest way to improve the present dataset from Austre Brøggerbreen, is to do more surveys at one year intervals, and leave markers in the wall in order to find the exact movement of the channel pattern between the surveys. The accuracy at present is acceptable for studies of the large-scale development, but lacks enough detail to say much about exact incision rates and the change in single channel features.

With more detailed morphology measurements, it would also be possible to run hydraulic models to simulate the flow through meanders and over steps. This could help answer questions about the development of these features by giving an idea of where the thermal erosion is most intensive, how slope and velocity relates to channel width, and what the values of slope and velocity must be before the jet detaches form the step face.

As the channel is believed to be of supraglacial origins, measurements of the surface slope above the channel would also be of interest. A digital elevation model or a series of GPS measurements along the most likely flow path on the glacier surface could give some clues about the initial profile pattern of the channel. Together with the earliest englacial surveys, it is possible that this would explain the location of step sequences and near horizontal channel reaches in the englacial pattern.

An underlying aim for this study is to gather an understanding about processes, conditions and relationships in the englacial system that is good enough to model it, scale it up and use it to predict the various impacts glacier hydrology has on the environment. The nature of this aim ensures that it will always be out of reach, though with a lot of research potential to get closer. If one were to continue the approach from the angle of Austre Brøggerbreen, the next steps after this study would be to answer the following questions:

What is the relationship between step sequences and meandering reaches in terms of channel development?

How big is the influence of the surface water temperature near the entrance point on incision rates, and how fast does it adjust to the equilibrium temperature in a natural, irregular channel?

What are the rates of step migration, meander development and channel incision, and which factors control this?

In addition to these, it would also be rewarding to find out why channel B does not run along the glacier bed. There are clearly some processes in then englacial and subglacial systems of cold-based glaciers that have not been discovered yet, or at least still contain blank fields of considerable sizes.

5 Conclusions

The aims of this thesis were to describe the long-term development of the profile and planar pattern of an englacial channel, and to point out the main factors and processes controlling its behaviour, based on the available literature and a series of repeated channel surveys.

In the planar pattern, the main development was in the form of increased meander wavelength and radius of curvature. The growing meanders was accommodated by the smoothening out of the smaller meanders, leaving a few big, regular meanders by the end of the survey period. No clear trends of upstream or downstream meander migration was observed in the dataset, the movement of the meanders seemed to be caused by the increase in wavelength and amplitude.

The increase in wavelength and radius of curvature coincided with an increase in channel width, which was almost doubled between the surveys of 2002 and 2014. The channel width, in turn, showed a weak negative correlation to channel slope and is believed to have increased as the slope of the channel decreased to a near-horizontal gradient. After the 2008 survey, the main form of meander development was an increase in amplitude.

In the profile pattern, two processes were found to account for the development of the channel. The first is the upstream migration of steps, and the second is the downward incision of the near-horizontal channel reaches. The majority of the steps appeared in clusters, separated by pools. They appeared to have kept their cluster formation as they migrated upstream, no signs of merging or scattering of step sections appeared from the dataset. The steps migrated along the planar pattern of the channel, disturbing the meandering forms as they moved through.

The step migration rates exceeded the downward incision rates of the near-horizontal channel reaches, and step migration is considered a major process for local channel lowering. However, in a profile pattern where the steps are located in clusters separated by long reaches of near-horizontal channel, the downward incision of these in-between reaches appears to have contributed with an equal part of the total work of channel lowering. The increase in sinuosity affected the net rate of upstream step migration relative to the entrance point, by increasing the length of channel the steps have to migrate through.

In the theory on channel origins presented in this thesis, channel B in Austre Brøggerbreen is believed to be formed by cut-and-closure. The differences in slope on the glacier surface, giving different incision rates, is presented as the reason why part of the channel incised down deep enough to be cut off from the surface by roof closure while the upstream part of the channel still runs on the glacier surface. The higher values of slope in what developed to be the englacial part of the channel continued to increase the difference in depth between the supraglacial and englacial channel lengths. This aided the development of the entrance moulin. After having reached a crucial steepness where the jet flow over the step detached from the step face, the moulin is thought to have enhanced its own form, growing in height by the incision of the downstream parts of the channel and the upstream migration of the lower-lying steps.

The development of englacial channels happens by thermal erosion of the stream flow along the channel walls, to some degree counteracted by the creep of the ice at depth. No signs of structural controls on channel development were found in channel B, so its channel pattern is concluded to be a result of hydrodynamic processes. The mechanisms governing these processes predict higher rates of channel development to happen where the stream flow is warmest, as the energy transfer to the ice boundary is proportional to the temperature difference between the ice and the water. This means that the greatest changes in the channel will happen near the energy source points, which in an englacial channel is the surface entrance and any significant elevation drops in the system. This is reflected in the surveys of channel B, where the changes near the entrance moulin are quite dramatic compared to the changes in the rest of the surveyed channel.

These mechanisms also indicate that the channel will have a the highest amounts of energy available for development at the points where the stream flow is at its most unstable, enabling it to effectively adjust towards a channel pattern which is closer to equilibrium. In an ice-walled channel a stable pattern is where the slope is close to horizontal, giving low melt rates and a low equilibrium temperature. Channel B has to a great extent developed towards this pattern, with long channel stretches of near-horizontal gradient and elevation drops and turbulence confined to relatively small portions of the total channel length.

None of the surveys of channel B reached the bed of the glacier. The results of the dye tracer experiments indicate that the channel does not reach the bed at all. This can be partly explained by the overdeepening over which the glacier is situated, which will prevent the channel to reach the bed, mainly by adjusting the local base level. However, this does not explain why the channel still runs far above the bed downglacier of the overdeepening, which is the case in Austre Brøggerbreen. No reasonable explanation for this was found in this thesis.

The main shortcoming of the study is that only a part of the englacial system was accessible for direct surveying. The complexity of this part, and most likely of the rest of the system as well, makes the interpretation of dye tracer experiments uncertain. This gives very limited results on which to base an interpretation on the flow through the englacial system. The system cannot be reached directly when it is active, and so the flow conditions can only be inferred from the morphology of the channel as measured in winter. As the surveys in addition were done at intervals of several years, the temporal resolution of the dataset is far from optimal.

The survey series of channel B in Austre Brøggerbreen holds a great potential to further increase the knowledge of long and short-term behaviour of englacial channels. But more surveys are needed, with better accuracy and higher temporal and spatial resolution, in order to pin down the exact rates and processes of channel pattern development.

6 References

Badino, G., 2002. The glacial karst, V. International Symposium on Glacier Caves and Cryokarst in Polar and High Mountain Regions. Nimbus, Courmayeur, Valle d'Aosta, pp. 82-93.

Baelum, K. and Benn, D.I., 2011. Thermal structure and drainage system of a small valley glacier (Tellbreen, Svalbard), investigated by ground penetrating radar. Cryosphere, 5(1): 139-149.

Bagnold, R.A., 1966. An approach to the sediment transport problem from general physics.

Bartholomaus, T.C., Anderson, R.S. and Anderson, S.P., 2008. Response of glacier basal motion to transient water storage. Nature Geoscience, 1(1): 33-37.

Benn, D., Gulley, J., Luckman, A., Adamek, A. and Glowacki, P.S., 2009. Englacial drainage systems formed by hydrologically driven crevasse propagation. Journal of Glaciology, 55(191): 513-523.

Benn, D.I. and Evans, D.J.A., 2010. Glaciers and Glaciation, 2. Hodder Education, London, UK.

Benn, D.I., Warren, C.R. and Mottram, R.H., 2007. Calving processes and the dynamics of calving glaciers. Earth-Science Reviews, 82(3-4): 143-179.

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

Boon, S. and Sharp, M., 2003. The role of hydrologically-driven ice fracture in drainage system evolution on an Arctic glacier. Geophysical Research Letters, 30(18).

Brykala, D., 1999. Hydraulic geometry of a supraglacial stream on the Waldemar Glacier (Spitsbergen) in the summer of 1997, Polish polar studies: 26th International Polar Symposium, Marie Curie-Sklodowska University (UMCS), Lublin, Poland, pp. 51-64.

Catania, G.A. and Neumann, T.A., 2010. Persistent englacial drainage features in the Greenland Ice Sheet. Geophysical Research Letters, 37.

Chang, H.H., 1979. Minimum Stream Power and River Channel Patterns. Journal of Hydrology, 41(3-4): 303-327.

Chang, H.H., 1986. River Channel Changes - Adjustments of Equilibrium. Journal of Hydraulic Engineering-Asce, 112(1): 43-55.

Church, M. and Zimmermann, A., 2007. Form and stability of step-pool channels: Research progress. Water Resources Research, 43(3).

Cuffey, K.M. and Paterson, W.S.B., 2010. The Physics of Glaciers. Elsevier, Inc., Oxford, UK.

Dozier, J., 1974. Channel Adjustments in Supraglacial Streams. Icefield Ranges Research Project, Scientific results, 4: 189-205.

Dozier, J., 1976. An examination of the variance minimization tendencies of a supraglacial stream. Journal of Hydrology, 31: 22.

Ferguson, R.I., 1973. Sinuosity of Supraglacial Streams. Geological Society of America Bulletin, 84(1): 251-255.

Ford, D. and Williams, P., 2007. Karst hydrogeogolgy and geomorphology. John Wiley & Sons, Ltd, Wiltshire, Great Britain.

Fountain, A.G. and Walder, J.S., 1998. Water flow through temperate glaciers. Reviews of Geophysics, 36(3): 299-328.

Fyffe, C.L., 2012. Section 3.4.3: Tracer investigations. In: L. Clarke (Editor), Geomorphological Techniques (Online Edition). British Society for Geomorphology, London, UK.

Gardner, A., Moholdt, G., Arendt, A. and Wouters, B., 2012. Accelerated contributions of Canada's Baffin and Bylot Island glaciers to sea level rise over the past half century. Cryosphere, 6(5): 1103-1125.

Gulley, J., 2009. Structural control of englacial conduits in the temperate Matanuska Glacier, Alaska, USA. Journal of Glaciology, 55(192): 681-690.

Gulley, J. and Benn, D.I., 2007. Structural control of englacial drainage systems in Himalayan debris-covered glaciers. Journal of Glaciology, 53(182): 14.

Gulley, J.D., Benn, D.I., Muller, D. and Luckman, A., 2009a. A cut-and-closure origin for englacial conduits in uncrevassed regions of polythermal glaciers. Journal of Glaciology, 55(189): 66-80.

Gulley, J.D., Benn, D.I., Screaton, E. and Martin, J., 2009b. Mechanisms of englacial conduit formation and their implications for subglacial recharge. Quaternary Science Reviews, 28(19-20): 1984-1999.

Gulley, J.D., Walthard, P., Martin, J., Banwell, A.F., Benn, D.I. and Catania, G., 2012. Conduit roughness and dye-trace breakthrough curves: why slow velocity and high dispersivity may not reflect flow in distributed systems. Journal of Glaciology, 58(211): 915-925.

Hagen, J.O., Kohler, J., Melvold, K. and Winther, J.G., 2003. Glaciers in Svalbard: mass balance, runoff and freshwater flux. Polar Research, 22(2): 145-159.

Hagen, J.O., Korsen, H.O. and Vatne, G., 1991. Drainage pattern in a sub-polar glacier: Brøggerbreen, Svalbard. In: Y. Gjessing, J.O. Hagen, K.A. Hassel, K. Sand and B. Wold (Editors), Arctic Hydrology: Present and Future Tasks. Norwegian National Comittee for Hydrology, Oslo, pp. 121-131.

Hagen, J.O. and Sætrang, A., 1991. Radio-echo soundings of sub-polar glaciers with low-frequency radar. Polar Research, 9: 99-107.

Hambrey, M.J., 1977. Supraglacial drainage and its relationship to structure, with particular reference to Charles Rabots Bre, Okstidan, Norway. Norsk geografisk Tidsskrift, 31: 69-77.

Hanna, E., Huybrechts, P., Steffen, K., Cappelen, J., Huff, R., Shuman, C., Irvine-Fynn, T., Wise, S. and Griffiths, M., 2008. Increased runoff from melt from the Greenland Ice Sheet: A response to global warming. Journal of Climate, 21(2): 331-341.

Hodgkins, R., Cooper, R., Wadham, J. and Tranter, M., 2009. The hydrology of the proglacial zone of a high-Arctic glacier (Finsterwalderbreen, Svalbard): Atmospheric and surface water fluxes. Journal of Hydrology, 378(1-2): 150-160.

Hodson, A., Kohler, J. and Brinkhaus, M., 2005. Multi-year water and surface energy budget of a high-latitude polythermal glacier: evidence for overwinter water storage in a dynamic subglacial reservoir. Annals of Glaciology, Vol 42, 2005, 42: 42-46.

Holmlund, P., 1988. Internal Geometry and Evolution of Moulins, Storglaciaren, Sweden. Journal of Glaciology, 34(117): 242-248.

Holtermann, C., 2007. Flow conditions within englacial drainage channels, Norwegian University of Science and Technology, Trondheim.

Hooke, R.L., 1981. Flow Law for Polycrystalline Ice in Glaciers - Comparison of Theoretical Predictions, Laboratory Data, and Field-Measurements. Reviews of Geophysics, 19(4): 664-672.

Hooke, R.L., 1989. Englacial and Subglacial Hydrology - a Qualitative Review. Arctic and Alpine Research, 21(3): 221-233.

Hooke, R.L., Miller, S.B. and Kohler, J., 1988. Character of the Englacial and Subglacial Drainage System in the Upper Part of the Ablation Area of Storglaciaren, Sweden. Journal of Glaciology, 34(117): 228-231.

Hooke, R.L.B., 1975. Distribution of Sediment Transport and Shear-Stress in a Meander Bend. Journal of Geology, 83(5): 543-565.

Hubbard, B. and Nienow, P., 1997. Alpine subglacial hydrology. Quaternary Science Reviews, 16(9): 939-955.

Irvine-Fynn, T.D.L., Hodson, A.J., Moorman, B.J., Vatne, G. and Hubbard, A.L., 2011. Polythermal Glacier Hydrology: A Review. Reviews of Geophysics, 49.

Isenko, E., Naruse, R. and Mavlyudov, B., 2005. Water temperature in englacial and supraglacial channels: Change along the flow and contribution to ice melting on the channel wall. Cold Regions Science and Technology, 42(1): 53-62.

Jacob, T., Wahr, J., Pfeffer, W.T. and Swenson, S., 2012. Recent contributions of glaciers and ice caps to sea level rise. Nature, 482(7386): 514-518.

Jansson, P., 2010. Ice sheet hydrology from observations, Svensk Kärnbränselhantering AB, Stockholm, Sweden.

Jarosch, A.H. and Gudmundsson, M.T., 2012. A numerical model for meltwater channel evolution in glaciers. Cryosphere, 6(2): 493-503.

Jenssen, E.V., 2014. Snørekord i Ny-Ålesund. Norwegian Polar Institute.

Johannesson, H. and Parker, G., 1989. Linear Theory of River Meanders. In: S. Ikeda and G. Parker (Editors), River Meandering. AGU, Washington, D. C., pp. 181-214.

Karlstrom, L., Gajjar, P. and Manga, M., 2013. Meander formation in supraglacial streams. Journal of geophysical Research: Earth Surface, 118: 1-11.

Knighton, A.D., 1981. Channel Form and Flow Characteristics of Supraglacial Streams, Austre-Okstindbreen, Norway. Arctic and Alpine Research, 13(3): 295-306.

Knighton, A.D., 1985. Channel Form Adjustment in Supraglacial Streams, Austre Okstindbreen, Norway. Arctic and Alpine Research, 17(4): 451-466.

Knighton, D., 1998. Fluvial forms and processes. Hodder Education, London.

Kohler, J., 2014. Mass balance for glaciers near ny-Ålesund, Miljøovervåking Svalbard og Jan Mayen (MOSJ). Norwegian Polar Institute.

Kostrzewski, A. and Zwolinski, Z., 1995. Hydraulic geometry of a supraglacial stream. Quaestiones Geographicae, Special Issue(4): 165-176.

Langbein, W.B. and Leopold, L.B., 1966. River Meanders - Theory of Minimum Variance. USGS Professional Paper, 422-H.

Leopold, L.B. and Langbein, W.B., 1962. The Concept of Entropy in Landscape Evolution. Geological Survey Professional Paper(500-A).

Leopold, L.B. and Maddock, T., 1953. The Hydraulic Geometry of Stream Channels and Some Physiographic Implications. Geological Survey Professional Paper(252).

Leopold, L.B. and Wolman, M.G., 1960. River Meanders. Geological Society of America Bulletin, 71(6): 769-&.

Liestol, O., 1988. The glaciers in the Kongsfjorden area, Spitsbergen. Norsk geografisk Tidsskrift, 42: 231-238.

Marston, R.A., 1983. Supraglacial stream dynamics on the Juneau Icefield. Annals of the Association of American Geographers, 73(4): 597-608.

Merz, J. and Doppmann, G., 2006. Measuring Mountain Stream Discharge Using the Salt Dilution Method. In: U.o.B. PARDYP/ICIMOD and Hydrology Group (Editor), Kathmandu.

Molnar, P. and Ramirez, J.A., 1998. Energy dissipation theories and optimal channel characteristics of river networks. Water Resources Research, 34(7): 1809-1818.

Munson, B.R., Young, D.F., Okiishi, T.H. and Huebsch, W.W., 2009. Fundamentals of Fluid Mechanics. John Wiley & Sons, inc., United States of America.

Müller, D., 2007. Incision and closure processes of meltwater channels on the glacier Longyearbreen, Spitsbergen, The University Centre in Svalbard/Technishe Universität Braunschweig, Braunschweig, Germany.

Naegeli, K., Lovell, H., Zemp, M. and Benn, D.I., 2014. Dendritic subglacial drainage systems in cold glaciers formed by cut-and-closure processes. Geografiska Annaler: Series A, Physical Geography.

Nienow, P.W., 1993. Dye tracer investigations of glacier hydrological systems, University of Cambridge, Cambidge.

NPI, 2015. Geologi 1:750 000, Svalbardkartet WebGIS. Norwegian Polar Institute.

Nye, J.F., 1953. The Flow Law of Ice from Measurements in Glacier Tunnels, Laboratory Experiments and the Jungfraufirn Borehole Experiment. Proceedings of the Royal Society of London Series a-Mathematical and Physical Sciences, 219(1139): 477-489.

Nye, J.F., 1973. Water at the bed of a glacier, Symposium on the Hydrology of Glaciers. IAHS Publications, Cambridge, England, pp. 189-194.

Parker, G., 1975. Meandering of Supraglacial Melt Streams. Water Resources Research, 11(4): 551-552.

Pimentel, S. and Flowers, G.E., 2011. A numerical study of hydrologically driven glacier dynamics and subglacial flooding. Proceedings of the Royal Society a-Mathematical Physical and Engineering Sciences, 467(2126): 537-558.

Porter, P.R., Vatne, G., Ng, F. and Irvine-Fynn, T.D.L., 2010. Ice-Marginal Sediment Delivery to the Surface of a High-Arctic Glacier: Austre Brøggerbreen, Svalbard. Geografiska Annaler Series a - Physical Geography, 92A(4): 437-449.

Refsnes, I., 2004. Geometri og løpsmønster til englasiale kanaler, Austre Brøggerbre, Svalbard, Norwegian University of Science and Technology, Trondheim.

Rutter, N., Hodson, A., Irvine-Fynn, T. and Solas, M.K., 2011. Hydrology and hydrochemistry of a deglaciating high-Arctic catchment, Svalbard. Journal of Hydrology, 410(1-2): 39-50.

Röthlisberger, H., 1972. Water pressure in intra- and subglacial channels. Journal of Glaciology, 11(62): 177-203.

Seminara, G., 2006. Meanders. Journal of Fluid Mechanics, 554: 271-297.

Shreve, R.L., 1972. Movement of water in glaciers. Journal of Glaciology, 11(62).

Spring, U. and Hutter, K., 1982. Conduit Flow of a Fluid through Its Solid-Phase and Its Application to Intra-Glacial Channel Flow. International Journal of Engineering Science, 20(2): 327-363.

Stock, J.W. and Pinchak, A.C., 1995. Diurnal discharge fluctuations and streambed ablation in a supraglacial stream on the Vaughan-Lewis and Gilkey glaciers, Juneau Icefield, Alaska. In: J.W. Hoyt and C. Presser (Editors), Proceedings of the ASME Heat Transfer and Fluids Engineering Divisions. ASME New York, New York.

Vatne, G., 2001. Geometry of englacial water conduits, Austre Brøggerbreen, Svalbard. Norwegian Journal of Geography, 55(2): 8.

Vatne, G. and Irvine-Fynn, T., in preparation. Morphological dynamics of an englacial channel.

Vatne, G. and Refsnes, I., 2003. Channel pattern and geometry of englacial conduits. Proceedings of the 6th International Symposium "Glacier caves and karst in Polar Regions": 8.

Williams, G.P., 1986. River Meanders and Channel Size. Journal of Hydrology, 88(1-2): 147-164.

Willis, I.C., Fitzsimmons, C.D., Melvold, K., Andreassen, L.M. and Giesen, R.H., 2012. Structure, morphology and water flux of a subglacial drainage system, Midtdalsbreen, Norway. Hydrological Processes, 26(25): 3810-3829.

Yang, C.T. and Song, C.C.S., 1979. Theory of Minimum Rate of Energy-Dissipation. Journal of the Hydraulics Division-Asce, 105(7): 769-784.

Zeller, J., 1967. Meandering channels in Switzerland. In: I.A.o.S. Hydrology (Editor), Symposium on River Morphology, . General Assembly of Bern, pp. 174-186.