Blink and You’ll Miss It: An Investigation into Surging on Flade
Isblink, Greenland

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Abstract

Mass loss from fringing ice caps and glaciers in Greenland accounted for 20% of the ice sheet’s mass loss between 2003 and 2008 and could be responsible for up to 11 mm of sea-level rise by the end of the century. This study therefore applies a 3D, full-Stokes glaciological flow model, Elmer/Ice, to Flade Isblink, the largest fringing ice cap in Greenland, to investigate its basal conditions and the mechanism of a surge of two of its major outlet glaciers observed around the turn of the millennium. The results show that the ice cap is largely cold-based outside areas of fast flow, but that freezing also dominates in sliding areas. This implies an additional hydrological or heat input to allow the persistence of fast flow, the most likely candidate being surface meltwater and the associated latent heat release, leading to cryo-hydrologic warming. Model results indicate that the surge mechanism is a soft-bed thermal surge assisted by a hydro-thermodynamic feedback, with an estimated return period of 20-43 years. A till layer of 0.2-0.35 m thickness is inferred beneath sliding areas according to the hydrological budgets estimated for the surging glaciers. Results also indicate that this type of surge can self-terminate even without thinning of overlying ice, as slowdown will naturally reduce the frictional heat flux at the bed and thereby increase the release of latent heat of fusion due to basal freezing. Faster basal freezing will, in turn, produce a stronger bed and even slower ice flow due to withdrawal of pore water from the underlying till layer, which acts as a reservoir of latent heat as well as water. On the basis of these results, it is hypothesised that surging on Flade Isblink will increase in frequency over the coming decades, as more heat and meltwater will be stored in the subglacial till layer under global warming.
Acknowledgements

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Aiya Eärendil elenion ancalima!

Hello to Jason Isaacs.
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\( u_{H}^{obs} \)  
Horizontal component of observed velocity

\( z_x \)  
Draft of ice

\( \alpha \)  
Surface slope

\( \alpha_p \)  
Power formulation of \( \beta \)

\( \beta \)  
Basal drag coefficient

\( \frac{\partial(H \bar{\tau}_{xy})}{\partial y} \)  
Lateral drag

\( \frac{\partial(H \bar{\sigma}'_x)}{\partial x} \)  
Longitudinal stress gradient

\( \dot{\varepsilon} \)  
Strain rate

\( \theta_b \)  
Basal thermal gradient

\( \lambda \)  
Regularisation coefficient

\( \rho_i \)  
Density of ice

\( \rho_w \)  
Density of water

\( \tau \)  
Effective stress

\( \tau^* \)  
Yield strength of till

\( \tau_b \)  
Basal stress

\( \tau_d \)  
Driving Stress

\( \Phi \)  
Internal friction angle
Chapter 1 – Introduction and Theory

1.1 – Introduction

Mass loss from the cryosphere is a global issue, being responsible for 75% of sea-level rise (SLR) since the 1970s (Stocker et al., 2013). The Greenland Ice Sheet (GrIS) has become the largest single cryospheric contributor to global SLR, with mass loss of 378±50 Gt a⁻¹ since 2009 (Enderlin et al., 2014), equivalent to just over 1 mm a⁻¹ of SLR out of a current total of 3.2 mm a⁻¹ (Stocker et al., 2013). Mass loss from the GrIS has been accelerating, having stood as low as 51±65 Gt a⁻¹ as recently as the mid-90s (Shepherd et al., 2012). It seems likely that this will persist or worsen, such that the total contribution of the GrIS to 21st-century SLR could reach 80 mm (Enderlin et al., 2014). Thus, it is necessary to improve understanding of the ice sheet’s dynamics and likely evolution to constrain estimates of future SLR.

One notably understudied region is north-east Greenland. The most prominent feature of the region is the North-east Greenland Ice Stream (NEGIS), with its three outlets at Nioghalvfjerdsfjorden (79 North), Zachariæ Isstrøm (ZI) and Storstrømmen, draining 16% of the GrIS (Joughin et al., 2000; Khan et al., 2014). These three glaciers hold enough ice to eventually cause over 1 m of SLR if they were to retreat (Mouginot et al., 2015). They have appeared largely stable until very recently (Moon et al., 2012), and modelling has not suggested significant future mass loss from the area (Gillet-Chaulet et al., 2012). However, recent strong evidence has indicated retreat and acceleration are beginning. The trunk of the NEGIS has been thinning since 2003, with mass loss from the basin reaching 16.1±3.7 Gt a⁻¹, similar to that from Jakobshavn Isbræ (Khan et al. 2014), the most spectacular Greenlandic example of rapid retreat and acceleration. Overall, the area thinning at rates in excess of 1 m a⁻¹ increased nearly tenfold between 2003 and 2012, with both 79 North and ZI accelerating and retreating in response (Khan et al., 2014). Further work by Mouginot et al. (2015) showed that ZI has now detached from a sill that formed a basal pinning point and is retreating through a 30 km-wide overdeepening, making its retreat self-sustaining as it moves into deeper water. This has led to a quadrupling of its mean rate of retreat from 2011 and a tripling of the rate of acceleration from 2012, such that its velocity has increased by 50% from 2000. 79 North, meanwhile, has an uphill-sloping bed inland, so is not showing...
such rapid retreat, yet has still seen an 8% increase in velocity and ice discharge (Mouginot 
et al., 2015). Thus, it is evident that north-east Greenland is now demonstrating signs of accelerating mass loss, similar to that currently affecting the south-east and north-west. As the region has been neglected until recently, more studies are necessary to investigate the nature of these changes.

Whilst uncertainty over future GrIS contributions to SLR is dominated by the contributions of major outlet glaciers, with the top twenty being responsible for 77% of mass loss from the GrIS (Enderlin et al., 2014), the role of smaller glaciers and ice caps (GICs) should not be discounted. GICs in Greenland cover an area of about 90,000 km² and were responsible for 20% of Greenland’s mass loss between 2003 and 2008 (Bolch et al., 2013). They could be responsible for up to 11.0±0.3 mm of SLR in total by the end of the 21st century (Machguth 
et al., 2013). These smaller bodies of ice are more vulnerable to atmospheric and oceanic thermal forcing and respond quicker to changes in their environment. As such, they may provide a useful early-warning system for future changes to neighbouring ice sheets (Bolch et al., 2013). They can also often provide useful insights into processes that operate on the larger ice sheets through being easier to observe and model (e.g. Machguth et al., 2013). Despite their potential use, these fringing ice caps are often less well-studied than the major ice sheets, leading to substantial uncertainty about their current state and evolution, as well as their dynamics and flow patterns. As such, analysing this can not only allow a better understanding of the ice caps, but can also provide useful data for work on the GrIS.

North-east Greenland has several such fringing ice caps that could be studied to understand the region better. The largest of these is Flade Isblink, which is currently demonstrating widespread elevation changes of a few metres, both positive and negative, meaning that its average mass balance appears to be near-zero (Rinne et al., 2011). Outlet glaciers on the northern part of the ice cap show evidence of having surged during the last two decades (Palmer et al., 2010; Rinne et al., 2011) and the drainage of a subglacial lake was detected in the southern portion in 2011 (Willis et al., 2015), showing that many of the processes observed on the GrIS also operate on Flade Isblink. This makes it an ideal target for study, both intrinsically and to improve characterisation of the wider north-eastern sector of the GrIS.
1.2 – Aims and Objectives

In light of the evidence outlined in Section 1.1, it is apparent that uniformly-cold basal conditions are unlikely to exist throughout the ice cap. However, as the existing work on the lake drainage indicates it was primarily surface-melt-driven (Willis et al., 2015) and a lack of suitable data exists for the years 2010 to 2015, investigating the surge of the north-western outlet glaciers holds greater promise for elucidating basal conditions on FI. This study therefore aims to investigate this surge by applying inversion techniques to Flade Isblink using the 3D full-Stokes glaciological flow model, Elmer/Ice, allowing the basal friction and temperature distributions to be calculated at several different times. The model will be constrained using velocity data from the MEaSUREs and ESA CSS projects. These results will provide greater insight into surging-related processes, which will help understand similar changes observed on the GrIS. This study will also provide a much greater level of comprehension of the current state of Flade Isblink and how it might respond to predicted global warming, which may be useful in assessing how the wider north-eastern portion of the GrIS could evolve. Regardless, it is important that such fringing ice caps are studied in greater detail and this thesis will contribute to rectifying this situation.

Taking this into account, the fundamental aim of this study is to:

- Characterise the englacial and basal conditions of FI and its surging glaciers

It is proposed that this aim will be met by completing the following objectives:

1. Developing a meshed model representation of FI
2. Collate data fields specifying geometry (surface and bed elevation), flow (surface velocity) and climatic conditions (surface air temperature) for FI
3. Use (1) and (2) as inputs for inversion modelling with the Elmer/Ice full-Stokes model
4. Quantify changes in basal traction and the basal thermal regime in model solutions for the entirety of FI and for surging and non-surfing states of ice flow

A further useful step will be to attempt to calculate basal meltwater production on the ice cap in to see how this quantity is evolving and affected by surging, and to assist in determining the surge return period.
The theoretical background to this study is set out in the rest of this chapter before the study site is presented in Chapter 2. Data and the method used are then presented in Chapter 3, before results are shown in Chapter 4. A discussion of these is included as Chapter 5, with conclusions in Chapter 6.

1.3 – Glacier Surging

This section establishes the theoretical glaciological basis of this study. Initially, a summary of how glaciers flow is provided, with particular reference to surging; then surging itself is explained and the various factors thought to control it and the processes driving it are discussed.

1.3.1 – Glacier Flow

Evidence from several studies shows the diversity of glacier flow in Greenland and the surrounding area (Csatho et al., 2014; Joughin et al., 2010; Moon et al., 2012, 2014). Velocities can vary at scales from the daily (Das et al., 2008) to the seasonal (Moon et al., 2014) to the annual (Tedstone et al., 2015). Some glaciers can move by as little as a few metres annually, others can flow at speeds of several kilometres, with the record in Greenland currently held by Jakobshavn Isbrae, which is flowing at around 16 km a⁻¹ (Joughin et al., 2014). It is therefore clear that glaciers exhibit a wide range of different flow states. These are caused by the interaction of two different mechanisms. The first is ice creep, due to internal deformation of the ice itself and dominant in ice cap and ice sheet interiors where ice is often cold and frozen to the bed. The second is sliding, due to slip at the base of the ice between the ice and the underlying surface and dominant on faster-flowing glaciers where ice can attain the PMP.

1.3.1.1 – Ice Creep

At the most basic level, ice flow is a result of the strain due to the stress imposed by the weight of the ice itself (the driving stress). This form of motion due solely to the driving stress is termed ‘ice creep’ and leads to permanent deformation in a manner between true Newtonian viscous behaviour (where the rate of deformation is proportional to the stress) and perfect plasticity (where no deformation is observed below a critical threshold, the yield stress) (Paterson, 1994). In this form of motion, the ice deforms along cleavage planes
between, or defects within, individual ice crystals as the result of several different processes (Weertman, 1983). The important point to note is that there will be some degree of ice creep for any stress, no matter how small.

The driving stress that produces ice creep is given by the equation (Paterson, 1994):

\[
\tau_d = \rho_i g H \sin \alpha
\]

Where \( \tau_d \) is the driving stress, \( \rho_i \) is the density of ice (917 kg m\(^{-3}\)), \( g \) is the gravitational constant (9.81 m s\(^{-2}\)), \( H \) is the ice thickness (in m) and \( \alpha \) is the surface slope (in °) (Paterson, 1994). As can be seen, thicker and steeper glaciers will therefore have greater driving stresses and greater creep potential. Typical values for \( \tau_d \) in glaciers vary from tens to hundreds of kPa (e.g. Shapero et al., 2016).

The strain resulting from the driving stress, which leads to creep motion, is given by Glen’s Flow Law (Glen, 1955), adapted for glaciers by Nye (1957), an empirical law derived from laboratory experiments representative of the typical conditions encountered in many glaciers. This takes the form:

\[
\dot{\varepsilon} = A \tau^n
\]

Where \( \dot{\varepsilon} \) is the strain rate, \( \tau \) is the effective stress, and \( A \) and \( n \) are both constants. The value assigned to the constants in Glen’s Flow Law is a perennial problem in many glaciological studies, as both are difficult to measure, yet have a major impact on the ice flow. The Flow Law exponent, \( n \), is generally given a value of 3 (so strain increases with the cube of stress), based on the range of values returned from laboratory experiments (e.g. Russell-Head and Budd, 1979; Weertman, 1973) and some limited field measurements from Greenland (Dahl-Jensen and Gundestrup, 1987). This is widely-accepted as the default value for \( n \) (Paterson, 1994), unless there are compelling reasons to alter it, and facilitates comparison of results between different studies.

The rate factor, \( A \), is dependent on the temperature of the ice, following a version of the Arrhenius Relation adapted for ice by Hooke (1981):
Equation 3 - Ice-adapted Arrhenius Relation

\[ A = A_0 \exp\left(\frac{-Q}{RT_i} + 0.49836 \left(\frac{T_0 - T_i}{k}ight)^n\right) \]

Where \( A_0 \) is independent of temperature, but affected by pressure and the presence of impurities (Paterson, 1994), \( Q \) is the activation energy for creep, \( R \) is the universal gas constant (8.314 J mol\(^{-1}\) K\(^{-1}\)), \( T_i \) is the ice temperature (in K), \( T_0 \) is 273.39 K (the melting point of ice), and \( k \) is a constant with value 1.17. Whilst both \( A_0 \) and \( Q \) do vary, in glaciological practice they are often assigned values of 0.09302 Pa\(^{3}\) yr\(^{-1}\) and 78,800 J mol\(^{-1}\), respectively (Benn and Evans, 2010), to represent mean values from field studies (e.g. Paterson, 1977) and laboratory experiments (e.g. Weertman, 1973).

The consequence of the rate factor and its exponential relationship with temperature is that warmer ice deforms and flows much more readily than colder ice, as melting begins to occur at the boundaries between crystals, leading to greater lubrication (Goodman et al., 1981). This dependency on temperature of the rate factor is problematic, as measurements of ice temperature through the ice column are usually lacking, requiring that studies often have to make major assumptions regarding temperature, such as isothermal conditions.

An important consequence of Equation 2 is that strain rates are greatest near the base of the ice, as this is where stresses are higher, following Equation 1. The resulting velocity due to ice creep at any point within or on the glacier, however, is the sum of the strain rates in the underlying ice column, so creep velocities are highest at or near the glacier’s surface and zero at the bed (Paterson, 1994). Surface velocity \( (U_i) \) due to creep can be calculated by inserting Equation 1 into Equation 2 and integrating with respect to height, assuming \( A \) and \( n \) are both constant, giving (Paterson, 1994):

Equation 4 - Surface Creep Velocity

\[ U_i = \frac{2A}{n + 1} \left(\rho_i g \sin \alpha\right)^n H^{n+1} \]

This can in turn be integrated and divided by \( H \) to find the mean creep velocity (Benn and Evans, 2010):
Equation 5 - Mean Creep Velocity

\[ \bar{U}_i = \frac{2A}{n+2}(\rho_i g \sin \alpha)^n H^{n+1} \]

The velocities calculated using Equation 5 (typically a few metres per year for ice a few hundred metres thick, as is typical of most glaciers) will often be substantially different from true velocities, as, in addition to both \( A \) and \( n \) potentially varying across and within the body of ice under consideration, Equation 5 also assumes the driving stress is resisted exclusively by basal drag, neglecting other components of the full glacier force balance (lateral drag and longitudinal stress gradients) described by van der Veen and Whillans (1989) and represented as:

Equation 6 - Glacier Force Balance

\[ \rho_i g H \sin \alpha = \tau_b - \frac{\partial(H \bar{\tau}_{xy})}{\partial y} - \frac{\partial(H \bar{\sigma}'_x)}{\partial x} \]

Where \( \tau_b \) is the basal drag, the second term on the right-hand side is the lateral drag, and the third term is resistance from longitudinal stress gradients. The assumption that the driving stress is exclusively resisted by basal drag is generally valid in the slow-flowing interiors of ice caps and ice sheets, where creep predominates, but is less tenable on faster-flowing ice, where a second flow mechanism, basal sliding, is responsible for the majority of motion (Benn and Evans, 2010).

1.3.1.2 – Basal Sliding

Sliding refers to glacier motion caused by slip at the ice-bed interface due, in some form, to the lubrication provided by liquid water, leading to speeds well in excess of ice creep alone – hundreds to thousands of metres annually. This requires the bed to be warm, \( i.e. \) that the basal ice has reached the PMP. This may be simply due to thicker ice (insulating the bed more and depressing the PMP further), higher geothermal heat fluxes or heat generated by friction, or any combination of these (Benn and Evans, 2010). Basally-produced meltwater may be supplemented by surface melt that can access the glacier bed through crevasses and moulins. Indeed, sufficient amounts of surface meltwater reaching the bed can cause it to warm and reach the PMP even if, \( a \ priori \), the conditions favour a frozen bed (Phillips et al., 2010). This is because meltwater contains a large amount of latent heat, which, if the
meltwater is stored at the bed over a period, may be released, warming the surrounding basal ice to the point where it can begin sliding (Phillips et al., 2010). The importance of this mechanism, termed ‘cryo-hydrologic warming’ (CHW), has been shown by Phillips et al. (2010, 2013) through observations of Sermeq Avannarleq in western Greenland. They found that the observed velocity and sliding profiles on the glacier could only be explained by including CHW as surface meltwater reached the base of higher-altitude, previously frozen-bedded parts of the ice under the influence of global warming. Not only does CHW lead to an increased areal extent of sliding, but the warming of the ice it occasions also causes the ice viscosity to drop, greatly enhancing ice creep (Phillips et al., 2013). Or, to put it another way, CHW increases $A$ (Equation 2) by a factor of two to three (Phillips et al., 2013).

Regardless of the exact source of the warm beds and meltwater that make it possible, sliding can occur on both hard and soft (sedimentary) beds. Other factors provide resistance to flow instead (Kamb, 1970). From a force-balance perspective, sliding is the result of a reduction in $\tau_b$, which means the glacier accelerates in order for the other resistive stresses (lateral drag and longitudinal stress gradients) to increase to the point where a balance is re-imposed (Paterson, 1994; Benn and Evans, 2010). The bed also still usually provides a degree of resistance, as glacier beds are not perfectly smooth, either from obstacles (form drag), from friction with entrained debris (frictional drag) or, in the case of soft beds, from resistance from the underlying sediments (Benn and Evans, 2010). As can be clearly seen from the foregoing discussion, the role of subglacial hydrology is crucial in initiating and mediating sliding, on both hard and soft beds, which is the focus of the rest of this section, as it is also directly-relevant to glacier surging.

Before discussing hydrology in greater detail, two important sliding mechanisms that allow form drag to be overcome and are somewhat independent of hydrology need to be explained: regelation and enhanced creep. Regelation was first proposed by Weertman (1964), developed further by Kamb (1970) and Lliboutry (1968, 1987), and occurs due to pressure differentials across bed obstacles. On the stoss (upstream) side of an obstacle, the pressure will be locally higher, reducing the PMP. This can allow ice to melt, flow round the obstacle in a thin film of water, and refreeze on the lee side, where the pressure is lower and the PMP higher. The ice can therefore sidestep the obstacle, reducing its form drag. There is strong observational evidence for regelation, e.g. Kamb and LaChapelle (1964) and
Hubbard and Sharp (1993). The second process, enhanced creep, is a direct consequence of Glen’s Flow Law (Equation 2). As the strain rate varies non-linearly with the stress, the higher stresses on the stoss sides of obstacles will lead to substantial local increases in strain rates, promoting faster flow over and around the obstacle. It is important to emphasise that these two mechanisms do not exist in isolation, but work in tandem – all obstacles will cause both processes to operate, but one will tend to dominate, depending on the obstacle (Lliboutry, 1993).

Returning to the importance of hydrology, the crucial aspect of it regarding sliding is the structure of the subglacial hydrological system. Two broad kinds of system are recognised in the literature: efficient, channelised systems that flow in an arborescent network, first proposed by Röthlisberger (1972) and Nye (1976), and inefficient, distributed systems that flow in a sequence of basal cavities (generally in the lee of obstacles) connected by narrow passageways, put forward by Lliboutry (1968). This structure is the result of the interplay of two processes: ice creep (as described in the preceding section), that acts to close channels and cavities; and melting by the liquid water contained in these systems (the first stage of CHW described above), that acts to keep them open (Röthlisberger, 1972). The relatively high volume-to-surface-area ratio of a channelised system results in melt tending to outweigh closure by creep, provided discharge within the system remains sufficiently high. Therefore, as discharges increase, a channelised system tends to grow quickly and water pressure in it drops, drawing in more water (Röthlisberger, 1972). For a distributed system, with a much lower volume-to-surface-area ratio due to being more dispersed, the opposite is true (Walder, 1986). Creep closure tends to occur at a faster rate than melt-driven expansion, so water pressure increases as discharges do, expelling water across the bed. If discharge increases sufficiently, a distributed system will tend to evolve to a channelised system as melt starts to outweigh creep closure in the larger cavities, forming embryonic channels (e.g. Rippin et al., 2003). Thus, over the course of a year, an initial distributed system that can handle the low discharges of winter usually evolves into a channelised system under the influence of increased discharge due to surface melt in spring and summer, before decaying to a distributed system in autumn as discharge decreases as surface melt ceases (e.g. Schuler et al., 2004; Willis et al., 2008).
Consequently, distributed subglacial systems generally lead to liquid water being present across a greater portion of the bed at a higher pressure. They will also tend to retain water for longer, making CHW a greater factor. This has several consequences that enhance sliding. First, the ice-bed interface is simply more lubricated and a greater portion of it is sufficiently warm to slide. Second, greater water thicknesses may entirely submerge smaller obstacles, removing elements of form drag. Third, higher water pressures will reduce the effective pressure. The effective pressure ($N$) is equal to the ice overburden pressure minus the subglacial water pressure and represents how strong a force the glacier is exerting on the bed. A reduction in effective pressure is important in enhancing sliding on both hard and soft beds in different ways.

For hard beds, this comes about through its effect on frictional drag. From basic physical considerations, frictional drag is dependent on the amount of debris entrained by the glacier and present at the ice-rock interface, and the force with which this debris is being pressed into the bed (the contact force). Two competing theories were advanced to calculate this drag. The first is that of Boulton (1974), modified and updated by Schweizer and Iken (1992), where friction is simply a function of the friction coefficient, the proportion of the bed covered by debris, and the effective pressure. The second is that of Hallet (1979), where the contact force is independent of the effective pressure, as ice can flow over and around entrained particles, rather than them being locked in place. Instead, the contact force is calculated as the sum of the buoyant weight of a particle (the difference between its weight and the weight of the same volume of ice) and the drag force due to ice flow towards the bed (usually due to vertical strain and basal melting). After occasioning much debate in the glaciological literature due to a lack of conclusive evidence to support one model over the other, experiments by Iverson et al. (2003, 2007) have shown that frictional drag is strongly-related to effective pressure, supporting the model of Boulton (1974). However, the values observed are ten times higher than this model predicts, showing that ice flow towards the bed is also an important factor. As a result, Cohen et al. (2005) have developed a theory incorporating both Boulton (1974) and Hallet’s (1979) ideas, and seemingly more representative of the true causes of frictional drag. As such, a reduction in effective pressure caused by the presence of a high-pressure distributed system will substantially reduce frictional drag on a hard bed, increasing sliding speeds.
For soft beds, the effective pressure is a direct determinant of the strength of the underlying sediment and therefore of the resistance it provides to flow (Kavanaugh and Clarke, 2006). Soft beds became a focus of study following the work of Boulton and Hindmarsh (1987) on Breiðamerkurjökull in Iceland and the discovery of weak sediments beneath the Whillans Ice Stream in West Antarctica by Alley et al. (1986), showing that many glaciers might rest on such beds. It was proposed that deformation of these sediments could lead to sliding of the overlying glacier. Initially, Boulton and Hindmarsh (1987) proposed that this deformation followed a slightly non-linear viscous rheology, i.e. that its viscosity was almost independent of strain rate, once the yield stress had been exceeded, following Equation 7:

\[ \dot{\varepsilon} = K_c(\tau_b - \tau^*)^a N^{-b} \]

Where \( K_c \), \( a \) and \( b \) are empirical constants (with \( a \) being -1), \( \tau^* \) is the yield strength of the till and \( N \) is the effective pressure (in kPa – here, the ice overburden pressure minus the pore water pressure (PWP) in the sediment). This would mean that till deformation exerts a direct control on the velocity of the overlying ice, as it is solely a result of the applied stress and the effective pressure. However, this law was derived based on measurements beneath marginal ice, where basal stress is not easily-calculable (Hooke et al., 1997), calling the values used in constructing the formula into question. Extensive laboratory testing of subglacial till samples (e.g. Kamb, 1991; Kavanaugh and Clarke, 2006; Tulaczyk et al., 2000) and campaigns of field observations (e.g. Fischer and Clarke, 1994; Mair et al., 2003; Truffer and Harrison, 2006) have also shown that till rheology is more complex than assumed by Boulton and Hindmarsh (1987). Instead of a viscous rheology, subglacial sediment seems to most closely follow a Coulomb-plastic rheology, as best shown by Kavanaugh and Clarke (2006), whereby the till suddenly fails once the applied stress exceeds its strength, rather than deforming gradually. This critical stress threshold (the yield strength) is given by the Mohr-Coulomb rule (Kavanaugh and Clarke, 2006):

\[ \tau^* = c_0 + N \tan \Phi \]

Where \( c_0 \) is the cohesion of the sediment, and \( \Phi \) is the internal friction angle. In this model, glacier velocity is not controlled at a given point by the local properties of the till, which is a
passive part of the system once it has failed, but by the large-scale distribution of slippery and sticky spots along the basal and lateral margins. In both models, a reduction in effective pressure due to the existence of high-pressure water in a subglacial distributed drainage system implies a reduction in the strength of the sediment and, thus, facilitates sliding.

Another important hydrological effect on sliding is ‘hydraulic jacking’ (Iken, 1981; Iken and Bindschadler, 1986). This occurs where water pressures are locally higher than the ice overburden pressure, meaning that the ice can be lifted entirely off the bed. This effectively smooths the bed, reducing frictional and form drag and enhancing sliding, but also causes the entire driving stress of the glacier to be resisted by a smaller portion of the bed (as water cannot support a shear stress). Consequently, these non-jacked portions of the bed experience higher stresses, making sediment failure more likely and increasing the effectiveness of both regelation and enhanced creep (meaning these two mechanisms are indirectly affected by hydrology), all of which promotes sliding. This is, again, much more likely to occur in the presence of a high-pressure, distributed subglacial drainage system than where a low-pressure, efficient channelised one exists.

Taking all these factors into account, determining the resulting basal velocity from sliding can prove complicated. Several different sliding flow laws have been devised during the past sixty years, but none have proven entirely satisfactory. A particular problem is obtaining sufficient knowledge of the effective pressure distribution, the bed roughness and the sediment strength, all basal parameters that are very difficult, if not impossible to measure (Truffer, 2004). This is a primary influence behind the recent development of inversion modelling techniques, which can be used to examine basal velocity, discussed in the next chapter.

Finally, it is important to note that sliding, albeit at greatly-reduced rates of a few millimetres annually, may occur on frozen beds, regardless of substrate, with sliding of <4 mm a⁻¹ observed on the cold-based Urumqi Glacier I in western China (Echelmeyer and Zhongxiang, 1987) and Meserve Glacier in Antarctica (Cuffey et al., 1999). This is because thin films of liquid water can persist at the base if solute concentrations are sufficiently high, allowing some very limited sliding, though such a phenomenon will be of little importance at the period covered in this study (20 years).
1.3.2 – Surging

Having provided an overview of the factors controlling the flow of ice, it is necessary to focus on the phenomenon of surging, a particular manifestation of glacier flow. This section describes the nature of surging, its causes and mechanisms, and finishes with a brief discussion of what this implies for the observed surge on Fl.

1.3.2.1 – Causes

Surging refers to a particular kind of cyclic flow where a glacier alternates between periods of slow flow and periods of fast flow. Typically, slow flow periods (the ‘quiescent phase’) last for years to decades, if not centuries in some cases (Murray et al., 2003), whilst the periods of fast flow (the ‘active phase’) take months to years. During the active phase, velocities can increase anywhere from five to a thousand-fold, leading to rapid glacial advances that are subsequently eroded in the following quiescent phase (Raymond, 1987; Dowdeswell et al., 1991). On tidewater glaciers, this will also manifest as a peak in calving flux (Dunse et al., 2015; Murray et al., 2003). At a fundamental level, surging is due to an imbalance between accumulation and flow – accumulation is too high to be evacuated by the flow speeds present in the quiescent phase, but not high enough to sustain the permanent fast flow of the active phase (Dowdeswell et al., 1995). Surging is generally regarded as being due to internal instabilities, rather than being externally-forced (Meier and Post, 1969); however, this is not entirely accurate. As Dowdeswell et al. (1995) showed, warming in Svalbard reduces the amount of mass that surge-type glaciers there are able to accumulate, lengthening their surge cycles and, in some cases, stopping surging altogether. So, whilst surging may be largely due to internal dynamics, it is not wholly-independent of external influences.

One of the great uncertainties surrounding surging is why it occurs in some cases and not in others. It is not clear what is behind the mass imbalance that drives surging, nor why the glacier’s base velocity does not increase to resolve it, rather than oscillating between two extremes. It is also not clear what causes the geographical distribution of surging – the behaviour is attested in several places around the world, such as Svalbard (Dowdeswell et al., 1995; Jiskoot et al., 2000), Alaska (Kamb et al., 1985), Eastern Greenland (Jiskoot et al., 2003) and Central Asia (Copland et al., 2011; Quincey et al., 2015), yet is entirely absent in other glaciated regions, such as the Alps, with, overall, only about 1% of glaciers thought to
be surge-type (Jiskoot et al., 2000). Even within a surge cluster, such as Svalbard, this proportion is still only on the order of 13% (Jiskoot et al., 2000). Several studies (e.g. Jiskoot et al., 2000, 2003) have tried to analyse various glacial parameters in multiple-regression-type models to elucidate this conundrum, but with little success to date, with no coherent or convincing general theory of surging yet achieved. Where there has been more success has been in understanding surge mechanisms, i.e. the processes that lead to switches between quiescent and active phases. These will be discussed in the next section.

1.3.2.2 – Mechanisms

There are, traditionally, two main surge mechanisms: the hydrological, based on the work of Kamb et al. (1985) on Variegated Glacier, Alaska; and the soft-bed thermal, based on the work of Murray et al. (2003) on Monacobreen, Svalbard. These will be described before more recent developments are discussed.

They hydrological surge mechanism of Kamb et al. (1985) was based on the 1982-3 surge of Variegated Glacier in Alaska, where velocity increased a hundredfold for a few months, peaking at 65 m d\(^{-1}\). The fast flow advanced progressively downglacier following an observed surge front of thicker ice, though termination, accompanied by a 0.1 m drop in the surface elevation of the glacier, was simultaneous across the whole glacier. The result of the surge was that the upper 8 km of the glacier thinned by 50 m, whilst the lower portions thickened by up to 100 m. During the surge, dye-tracing experiments showed an average subglacial water velocity of 0.02 m s\(^{-1}\), with high dye dispersal and turbidity, such that dye was detected in all three proglacial outlet streams that the glacier fed. After the surge terminated, further dye tracing produced an average velocity of 0.7 m s\(^{-1}\), with all the dye appearing in a single peak in one outlet stream. This set of observations was explained through reference to the subglacial hydrology of Variegated Glacier, as the glacier was already found to be entirely warm-based in 1973, so changes in basal conditions did not seem responsible (Kamb et al., 1985). In effect, under the thicker ice that formed the surge bulge, ice creep closure rates were faster (Equation 5), restricting the formation of a channelised drainage system. Therefore, water pressures were locally greater, reducing the effective pressure and leading to quicker sliding velocities. This enhanced ice creep further, leading to greater water pressures until the ice overburden pressure was reached and hydraulic jacking occurred. The area of thicker ice therefore started sliding very rapidly
downglacier, forming the surge front and destroying the channelised drainage system present in the lower reaches as it passed. This was replaced with a high-pressure distributed system, maintaining the surge, until the front reached the glacier terminus and the trapped high-pressure water was finally able to escape, removing the hydraulic jack and terminating the surge.

An alternative to the hydrological surge model developed by Kamb et al. (1985) was provided by Murray et al. (2003), developing Clarke et al. (1984), to better fit observations on glaciers known to not be entirely warm-based and underlain by soft till, specifically the polythermal Monacobreen on Svalbard, which surged during the 1990s. This showed a much slower style of surge than predicted by the hydrological model, with a years-long active phase, a gradual termination through deceleration, and no observable surge bulge or front – instead the surge seemed to start simultaneously across the whole lower glacier before propagating upglacier. The glacier advanced by 2 km as a result of the surge, with maximum velocities of 5 m d\(^{-1}\), a tenfold increase on those observed in the quiescent phase. At no point were increased water pressures observed and velocity variations were much smoother than those on Variegated Glacier. Instead of the structure of the hydrological system, the mechanism proposed here by Murray et al. (2003) focuses on the evolution of the bed. During the quiescent phase, the glacier thickens through accumulation until some part of the bed reaches the PMP through a combination of falling PMP (as thicker ice exerts a greater pressure) and a better-insulated bed (so more heat is trapped). Once this happens, water begins to collect at the base, increasing PWPs in the underlying till, weakening it (Equation 8). This allows the ice above to begin sliding faster, generating more basal meltwater through friction, weakening the till further. At some point in this positive feedback, the till’s yield strength will be exceeded, and the till will fail, enhancing sliding. This process may spread outwards from the initial sliding locus as the friction-generated heat and meltwater diffuse, such that the whole glacier can become warm-based. This would also explain the lack of a surge front, as there is no margin at which flow is restricted, so no opportunity for any bulge of faster ice to form (Murray et al., 2003). The surge terminates when the ice has thinned sufficiently for the bed to drop below the PMP once more, leading to refreezing and the cessation of rapid sliding (Murray et al., 2003).
This thermal soft-bed surge mechanism has been confirmed as being in operation in Svalbard through modelling work that shows rapid motion was due to predominantly cold ice flowing over a temperate bed, rather than to do with the presence of temperate ice, and that there was an abrupt onset of basal sliding as the bed approached the PMP (Dunse et al., 2011). The mechanism has also recently been updated further through work by Dunse et al. (2015), who included the role of CHW in increasing the areal extent of ice vulnerable to fast flow, based on surging observed in Basin 3 on Austfonna in Svalbard. CHW provides a mechanism whereby previously-stagnant areas of ice that act as plugs can be unfrozen and mobilised, allowing a surge to occur or adding to one already occurring. Being driven by surface melting, CHW also provides another link between surging and external factors, further challenging the traditional view that surging is purely internally-driven (Dunse et al., 2015).

Further querying traditional conceptions of surging have been studies of the phenomenon in High Mountain Asia, particularly the Karakoram and Tian Shan ranges. These have revealed that surging is widespread in the region, but have shown that surges fitting both mechanisms exist, rather than there being one dominant mechanism (Copland et al., 2011). More perplexingly, surges have also been found that seem to show characteristics of both mechanisms simultaneously (Quincey et al., 2015). The overall conclusion is that there is clearly still insufficient understanding of the mechanics of, and in action during, surges, which this study hopes to elucidate.
Chapter 2 – Study Site

2.1 – Location and History

Flade Isblink (FI), at 8500 km², is the largest ice cap in Greenland not connected to the wider ice sheet (Palmer et al., 2010). It lies in the extreme north-east corner of the island, at 81°15’ N, 15°00’ W (Figure 1), on the Prinsesse Dagmar Halvøya in Kronprins Christian Land (Palmer et al., 2010). Station Nord, an important Danish outpost, lies just to the ice cap’s west. FI is about 50 km wide and 100 km long at its greatest extent, with a maximum ice thickness of approximately 535 m at the central summit (Palmer et al., 2010). The northern part of the ice cap peaks at an elevation of 710 m and slopes gently, whilst the southern part overlies a section of the Prinsesse Elisabeth Alps and is consequently steeper and more rugged, peaking at 960 m, with many nunataks (Palmer et al., 2010).

FI was discovered in 1907 as part of the Danmark expedition with the first published observations of it by Mikkelsen (1913). The Danish term ‘Isblink’ (‘ice blink’) was applied to the ice cap by the expedition, because it was originally spotted whilst still over the horizon due to its reflection in the sky, which formed a large patch of lighter-coloured cloud (essentially the opposite of a water sky, where patches of open water among an ice pack can be discerned from the darker appearance of the overlying clouds). Its status as a separate ice cap was not appreciated until the first aerial observations were made in 1932 (Koch, 1935). Further work on FI was conducted by Helk and Dunbar (1957), who considered it to be inactive, a view conclusively overturned by Higgins (1991), who demonstrated the existence of an active outlet glacier to the east of Station Nord, at the north-west corner of the ice cap.
Based on measured surface temperatures of \(-22^\circ \text{C} \ (251 \text{ K})\) at the southern ice divide, the basal temperature is thought to be no higher than \(-9^\circ \text{C} \ (264 \text{ K})\), well below the pressure melting point (PMP) of around \(-0.5^\circ \text{C} \ (272.66 \text{ K})\) for ice of this thickness (Willis et al., 2015). This suggests that the view that FI may be largely inactive may be true to an extent. Despite this, FI has existed in its current configuration for only between 2800 and 4000 years, having apparently melted entirely in the Holocene Climatic Optimum (Lemark and Dahl-Jensen, 2016).
There is also some evidence of advance beyond its current limits during the Little Ice Age (Hjort, 1997).

This apparent mutability relates to the location of FI, which is an unusual place to find a major ice cap. It sits on bedrock at a predominantly low elevation of ~100 m in a drier region of Greenland, where precipitation should not be sufficient to sustain it at such low altitudes (Rinne et al., 2011). The maintenance of the ice cap is driven by the ocean to the east, which remains largely ice-free all-year-round. This forms a mass of relatively warm and humid air above the ice cap. The prevailing north-westerly wind advects cold air over FI, bringing it into contact with the warmer oceanic air mass and causing precipitation on the western side due to the resulting condensation (Rinne et al., 2011).

Recent work on FI has conclusively-proven that it is not inactive, despite the presumed cold base. FI is, in fact, showing an unexpected degree of dynamism for such a small body of ice. Elevation surveys that included FI have revealed that the west-biased precipitation pattern is leading to thickening of 0.5 m a\(^{-1}\) on the western side of the ice cap, whilst the eastern side is showing thinning of 0.2 m a\(^{-1}\) (Krabill et al., 2000; Pritchard et al., 2009). More recent studies, focussing exclusively on FI, show peak thickening rates of up to 3.4 m a\(^{-1}\) (Rinne et al., 2011), making FI one of the fastest-accumulating places in Greenland. This is balanced by widespread thinning in the ablation area, such that the overall mass balance of the ice cap is near-zero, at 0.0±0.5 Gt a\(^{-1}\) (Rinne et al., 2011). Given the signs of accelerating mass loss observed elsewhere in north-east Greenland (Khan et al., 2014), this apparent relative stability makes FI a compelling arena for further study.

In addition to widespread elevation change, FI has shown some evidence of interesting glacial dynamics. Velocity data clearly show a surge occurred in the north-western outlet glaciers (those described by Higgins, 1991) between at least 1996 and 2000 (Joughin et al., 2010; Palmer et al., 2010; Rinne et al., 2011), whilst the drainage of a subglacial lake in 2011 near the southern summit of the ice cap has been detected (Willis et al., 2015). Both of these features are usually associated with temperate bed conditions, which have been considered unlikely under FI (see Sections 1.1 and 1.2). In the case of the subglacial lake, Willis et al. (2015) assert its formation was almost certainly the result of surface melt percolating to the base through moulins and crevasses, rather than being due to unexpectedly-warm basal conditions. However, no study has yet investigated the character
of the surge in the north-western glaciers. It is therefore clear that FI is a glaciologically interesting, yet heretofore-neglected, area suitable for further study; the objective of this dissertation.

For the purposes of this study, several outlet glaciers of FI are referred to. Their locations, names and abbreviations used are detailed in Table 1 and Figure 2. These glaciers were selected because they had an average velocity in excess of 50 m a\(^{-1}\) for the majority of the velocity datasets obtained and were therefore were clearly maintaining basal sliding for an extended period. One glacier, BEG, does not meet this threshold, but was included due to being considered a major outlet by Palmer et al. (2010) and to provide a point of comparison on the eastern margin of FI.

Table 1 - FI Outlet Glacier Names and Abbreviations. Where available, existing Danish names have been used; otherwise, simple English toponyms have been devised.

<table>
<thead>
<tr>
<th>Name of Outlet Glacier</th>
<th>Abbreviation Used</th>
</tr>
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<tbody>
<tr>
<td>Northern Surging Glacier</td>
<td>NSG</td>
</tr>
<tr>
<td>Southern Surging Glacier</td>
<td>SSG</td>
</tr>
<tr>
<td>Marsk Stig Bræ</td>
<td>MSB</td>
</tr>
<tr>
<td>Romer Lake Glacier North</td>
<td>RLGN</td>
</tr>
<tr>
<td>Romer Lake Glacier East</td>
<td>RLGE</td>
</tr>
<tr>
<td>Hjørnegletsjer</td>
<td>HG</td>
</tr>
<tr>
<td>Tobias Gletsjer</td>
<td>TG</td>
</tr>
<tr>
<td>Big Eastern Glacier</td>
<td>BEG</td>
</tr>
<tr>
<td>Northern Glacier</td>
<td>NG</td>
</tr>
</tbody>
</table>
Those seeking more information on the basic characteristics of FI are referred to Palmer et al. (2010) for a detailed description of the morphology and flow characteristics of the ice cap.
2.2 – Initial Considerations

Applying the material discussed in Chapter 1 and Section 2.1 to FI, with a calculated basal temperature of 264 K (Willis et al., 2015) under one of the thickest parts of the ice cap, it is likely that the majority of FI is frozen to its bed. Ice creep would therefore be the dominant flow mechanism for most of the ice cap, which would keep surface velocities to a few metres annually. As shown in Figure 3, representative winter velocity data for FI (from 2005), this is indeed the case. Outside the marked outlet glaciers referenced in Table 1, the rest of the ice cap, especially the eastern side, is almost entirely exhibiting velocities of <10 m a\(^{-1}\), indicating creep is the main cause of flow. There are some small patches where the velocity does exceed 10 m a\(^{-1}\), where there could be some very localised sliding, but these might simply be areas with a higher driving stress (thicker ice or steeper surface slopes) or different flow law constants, but that are still frozen to the bed.

However, Figure 3 also clearly shows the faster flow (tens to hundreds of metres annually) present on the outlet glaciers. This is too fast to be due to creep, so sliding must be occurring. The warm-bedded conditions this requires are probably maintained by the frictional heat generated by the faster flow, likely aided by high water pressures in a distributed drainage system due to minor meltwater inputs to the drainage system in winter, when this velocity was measured. Similar behaviour has been shown on glaciers in Svalbard (e.g. Rippin et al., 2003, 2005), which are perhaps the closest analogues to FI. Svalbard is also a region with many surging glaciers (Jiskoot et al., 2000), increasing its similarity to FI, so it is perhaps to be expected that the soft-bed thermal surge mechanism of Murray et al. (2003), developed to explain surges observed in Svalbard, is the most appropriate for understanding those taking place on FI.

This indeed seems to be the case. Based on initial observations, the active phase of the surge lasted for at least five years (1996-2000), which is typical of glaciers on Svalbard, but far longer than would be expected under the hydrological surge mechanism. The relatively-low maximum velocities (<500 m a\(^{-1}\) or <1.5 m d\(^{-1}\)) achieved also indicate a more gradual surge mechanism. Therefore, it is reasonable to assume from the outset that FI glaciers are underlain by a layer of till and that changes in this lead to surging. Consequently, analysis of model outputs (Chapters 3 and 4) and discussion of results (Chapter 5) will proceed based
on this, focussing on temperature and till properties, the two main factors controlling this surge mechanism.

Figure 3 - 2005 winter velocity on Fl. Velocity data courtesy of the MEaSUREs project. Background imagery from the MODIS map of Greenland.
Chapter 3 – Data and Method

This chapter presents details of the datasets and method used in this study, the vast majority of which are freely-available online. Each dataset is briefly described before a detailed exposition of the method, including justification and critique, is provided.

3.1 – Data

3.1.1 – Satellite Imagery

Background visible-band satellite imagery used to help define the limits of the ice cap and in the production of figures was taken from the NASA Making Earth System Data Records for Use in Research Environments (MEaSUREs) MODIS (Moderate Resolution Imaging Spectroradiometer) mosaic of Greenland (Haran et al., 2013), obtained from https://nsidc.org/data/nsidc-0547. This dataset provides a nearly entirely cloud-free view of Greenland, including all marginal areas and islands larger than a few hundred metres, at a resolution of between 100 and 500 m. Major ice caps in the Greenland-proximate regions of the Canadian Arctic are also covered (Baffin, Devon, Ellesmere and Axel Heiberg Islands). The mosaic is compiled from MODIS images taken in March and April 2005. The dataset also includes an ice mask layer covering these same areas. The mosaic’s view of FI is shown in Figure 4.

3.1.2 – Digital Elevation Model

Surface and basal digital elevation models (DEMs) used as inputs to the modelling process were obtained from the IceBridge BedMachine v2 dataset (http://nsidc.org/data/IDBMG4)
(Morlighem et al., 2015), as used in Morlighem et al. (2014) to reveal the existence of deeply-incised subglacial valleys beneath many GrIS outlet glaciers. This dataset uses the surface DEM compiled by the Greenland Ice Mapping Project (GIMP), which covers the entirety of Greenland at a resolution of 30 m, and was derived from ICESat laser altimetry data, such that it provides the mean surface elevation over the ICESat measurement period (2003-09) (Howat et al., 2014).

A basal DEM, at a resolution of 150 m, is then derived by subtracting ice thickness measurements obtained by the MCoRDS radar sounder at frequencies between 180 and 210 MHz and a resolution of 400 m as part of Operation IceBridge from the surface DEM. A recurring problem in glaciology is the sparseness of thickness measurements, making it difficult to be certain of ice and basal morphology. Traditionally, this problem is resolved by interpolation using the kriging method, but this yields unphysical thickness fields that lead to substantial flow divergence when employed in high-resolution modelling (Seroussi et al., 2011). Instead, this dataset uses the mass conservation method (Mcnabb et al., 2012; Morlighem et al., 2011, 2014) to provide, where practicable, a physically-based interpolation of these sparse measurements, with minimised deviation from observations, across the entire GrIS. Velocity data needed to apply mass conservation are taken from satellite interferometric synthetic aperture radar (InSAR) data at 150 m resolution, as described in Rignot and Mouginot (2012). Both the surface and basal DEM are shown in Figure 5.

Mass conservation works best in areas of fast, relatively-constrained flow where internal shear is negligible (Morlighem et al., 2011), so it is applied primarily to the margins, with kriging remaining in use for the GrIS interior, where ice thickness changes are more gradual.

Version 2 of the dataset includes some Operation IceBridge 2014 data and improved processing of some basins, as well as providing heights with reference to mean sea level, rather than the WGS84 ellipsoid.
3.1.3 – Velocity Data

Velocity data for 2000 and 2005-08 used as an input to the modelling process were obtained from the MEaSUREs velocity map of Greenland, Version 2 (http://nsidc.org/data/nsidc-0478) (Joughin et al., 2015), as presented in Joughin et al. (2010). Velocity data for 2015 were extracted from the ESA Climate Change Initiative (CCI) Greenland Ice Velocity Map (http://esa-icesheets-cci.org/?q=products#IV_Description).

The MEaSUREs data detail winter velocities across the entirety of the GrIS and marginal ice bodies in Greenland and are provided at 500 m resolution, in the x and y directions, as well as a magnitude value. The maps were produced using InSAR data from a combination of RADARSAT-1, the Advanced Land Observation Satellite (ALOS) and the TerraSAR-X satellite. Owing to the data resolution, the velocity of narrow glaciers (<1 km) will be an average of the ice motion and surrounding bedrock or slower ice, likely leading to underestimation, but all major FI outlet glaciers covered in this study are >1 km wide, so this has not been a problem here. An example of the MEaSUREs velocity data is shown in Figure 3.

Figure 5 - FI surface (Panel a.) and bed (Panel b.) DEMs from IceBridge BedMachine data.
Version 2 of the MEaSUREs dataset includes a new map for 2009-10 (unusable for Fl due to a large swathe of missing data across the ice cap), improved baseline fits for consistency and revised error estimates. Errors are typically a few metres per year for Fl.

The ESA CCI data (Nagler et al., 2015) likewise cover the entirety of Greenland at a resolution of 500 m, with the x and y components of velocity provided, as well as the magnitude. The temporal coverage of the data used to map the ice velocities is, however, a whole year, rather than being purely winter-based, limiting its direct comparability with the MEaSUREs dataset, though this is not significant for this study. The map was derived from InSAR measurements acquired by the Sentinel-1 satellite between November 2014 and December 2015.

InSAR-derived velocity magnitude data for summer 1996 (detailed in Palmer et al., 2010) were provided by Steven Palmer, and were used to qualitatively assess surge dynamics.

3.1.4 – Temperature Data
Ice surface temperature (IST) data for each year required as an input to the model were extracted from the Greenland IST maps compiled by the NASA MODIS Snow and Sea Ice Mapping Project (Hall et al., 2012, 2013), obtained from http://modis-snow-ice.gsfc.nasa.gov/?c=greenland. The dataset is compiled from daily observations of IST acquired by the MODIS satellite, covering all Greenland for the whole of 2000-2014. The data have been screened for clouds and are provided at 1.5 km resolution with monthly averages. The main temperature datasets used in this study were the annual averages for 2000 and 2005. These are displayed in Figure 11 and Figure 12.

3.2 – Method
The rest of this chapter describes the method used to conduct this study. The initial data processing is covered first, before the numerical modelling and subsequent optimisation procedure is explained. Finally, the procedure used to derive subglacial water production estimates and till porosity is summarised.

3.2.1 – Domain Definition
The first step in conducting this study was to define the bounds of the domain within which the model would be run. This was done manually in Quantum GIS (QGIS) using a
combination of MODIS satellite imagery and the accompanying ice mask, velocity data and previous publications on FI (Palmer et al., 2010; Rinne et al., 2011; Willis et al., 2015) to identify the edge of the ice cap. The resulting set of points were then connected using the PointConnector plug-in for QGIS, with the set of lines this formed merged into one using the Dissolve function to produce the domain boundary line seen in Figure 1 and Figure 2. This boundary line was then polygonised using the Lines To Polygon feature to produce a mask layer to be used in later steps.

Defining the bounds of FI was straightforward for the northern part of the ice cap, but more problematic for the southern one, south of the terminus of RLGN, where the presence of the Prinsesse Elisabeth Alps and associated nunataks complicates the topography. The decision was therefore taken to focus on the main ice-filled valleys and outlet glaciers, neglecting some smaller features, and simplifying the domain outline to reduce the computational load imposed on the model. In the context of inaccuracies in the input data (discussed in Section 3.2.2), this was regarded as not introducing substantial additional uncertainty into the model calculations.

3.2.2 – Data Preparation

The domain mask layer was then used to clip all the input data (DEM, velocity, temperature) to the required area. Where needed, the Raster Calculator in QGIS was then used to ensure all inputs were in the same units (m, m a⁻¹ and K, respectively). The temperature data for 2000 and 2005, which were provided by month, were averaged to provide a yearly mean temperature dataset for both years. The resulting DEM and temperature data were then converted to ASCII Gridded XYZ (.xyz) files and sorted into ascending co-ordinate order for reading into the model. The velocity data were converted to NetCDF (.nc) files with three bands, one each for velocity in the x direction, the y direction, and the overall magnitude; again to meet model requirements.

Undertaking this process revealed some inconsistencies with all three datasets that had to be resolved before using them successfully in a modelling exercise. For both the velocity and temperature data, there were (small) areas of no data within the domain boundaries that had to be filled. For the velocity data, regions lacking data were filled using an inverse-distance-weighted interpolation method to provide data based on surrounding valid pixels.
For the temperature data, where the areas were larger and mainly around the margins of the ice cap, making this approach less valid, regions were filled with the mean temperature of the domain (excluding areas of no data). Given the small range of temperatures encountered in the annual average datasets, this was considered to be a reasonable simplification with minor effects on glacial dynamics.

For the DEM dataset, a more complicated problem emerged. Sparse thickness measurements led to regions where mass conservations and kriging methods significantly underestimated the value of ice thickness (with estimates <1 m). Rather than making changes to the dataset, which was input to the model unaltered, this was resolved by having the model apply a minimum ice thickness of 10 m in calculations to ensure a degree of realistic ice flow. This was chosen as seeming a reasonable compromise for the affected marginal areas of FI, based on parts of the ice cap with better-quality data, but is probably a significant underestimate for much of the southern portion of FI and some glacier termini. Therefore, substantive analysis and discussion of results will be confined to those areas with thickness in the original data of at least 30 m, a threshold that serves to exclude all areas with deficient data, to ensure conclusions are valid. Results for thinner parts of the ice cap will be treated with caution, and interpretation confined to their pattern, rather than attaching undue importance to their values.

A further constraint due to the DEM dataset is that it imposes a constant geometry on the ice cap for each year in the modelling exercise. Evidently, this is not fully realistic as the ice cap’s geometry will change, but, given the relatively small average elevation change rate (0.03±0.03 m a⁻¹) for FI found by Rinne et al. (2011), the steady-state nature of the model simulations, and the restricted period covered in this study (20 years), this is not regarded as significant. The fact that the surface DEM represents the average elevation change over the years 2003 to 2008 also means it should be representative of the period studied here, with most other datasets falling between 2000 and 2008.

At this stage, velocity profiles for the nine outlet glaciers identified in Figure 2 were also extracted along the glacier centrelines.
3.2.3 – Meshing

The initial point layer used to define the domain mask was run through the gmsh programme using a Delaunay triangulation algorithm to produce a 2D mesh of the model domain. Given the size and complexity of the domain, it was decided to use a variable mesh to improve computational efficiency (e.g. Khazendar et al., 2015). The finest resolution (1 km) was specified for the marine or lake-terminating portions of the domain boundary, which require higher-resolution owing to the more complicated set of processes in these regions, with resolution decreasing linearly (over a distance of 10 km) to a maximum of 3 km away from these. This 2D mesh was then partitioned into four elements using the ElmerGrid programme, based on the Serial Graph Partitioning and Fill-reducing Matrix programme (METIS), to enable parallel computing, greatly reducing model runtime. Within model runs, this 2D mesh was then extruded to a 3D version, with ten equally-spaced vertical levels fitting the contours of the bed and surface DEMs. As mentioned in Section 3.2.2, the model ensured at least 10 m of flow depth were present, adjusting the surface DEM as required to produce this. The 2D mesh is shown in Figure 6 and the 3D mesh in Figure 7.

![2D Model Mesh](image-url)
3.2.4 – Initial Basal Drag Calculation

The final input required before being able to proceed to running the model and to complete the first objective of this study was an initial estimate of the distribution of basal drag beneath Fl; more specifically the distribution of the slip coefficient, $\beta$. This gives the model an initial state to work from, allowing it to iterate towards a more realistic answer. This initial estimate was calculated in QGIS’s Raster Calculator using the Shallow Ice Approximation (SIA), a simplification of the force balance of a glacier that assumes the driving stress is resisted exclusively by the basal stress, neglecting lateral and longitudinal stresses (Hutter 1981; 1983). Therefore, using Equation 1:

Equation 9 - SIA

$$\tau_b = \rho_i gh \sin \alpha$$

Where $\alpha$ was calculated using the Slope function in QGIS. From this $\beta$ can be worked out, assuming basal velocities equal surface velocities, using a simple Weertman friction law:

Equation 10 - Friction Law

$$\tau_b = \beta u_b$$
Where $u_b$ is the basal velocity. More elaborate friction laws have been developed (e.g. Gagliardini et al., 2007), but, as shown by Morlighem et al. (2010), all such laws will typically converge on the same answer in numerical inversions such as those used in this study. A simpler law was preferred here to avoid unnecessary calculation complexities.

Both the SIA and the assumption that surface and basal velocities are equal (i.e. that vertical shear is negligible) are valid for some bodies of ice; in the case of the former for slow-flowing, laterally-unconstrained ice; for the latter fast-flowing ice streams. As such, they are a poor mix of assumptions for a largely slow-flowing, cold-based ice cap, but, as the objective is solely to produce an estimate of $\beta$ to initialise the model, these perform well enough to meet that aim.

The main difficulty encountered with this step was again the poor quality of the thickness data discussed in Section 3.2.2, leading to very unrealistic values of $\beta$ in some parts of the ice cap, such that the model would struggle to successfully iterate from them. To resolve this, $\beta$ was calculated with $H$ set to a minimum of 30 m to ensure realistic values were obtained under all areas with poor thickness data - the termini of some of the major outlet glaciers, some other marginal areas and the southern portion of the ice cap. Although this value may be far from the actual ice thickness in all these areas (overestimating in some and underestimating in others), it is sufficiently close in all of them to produce values of $\beta$ that are realistic enough for the model to successfully iterate.

### 3.2.5 – Numerical Modelling

To meet the second objective of this study, the data discussed in Sections 3.2.1 to 3.2.4 were used as inputs into a numerical model. Numerical modelling has become an ever-more prominent process in glaciology over the last thirty years as computing power has increased, providing insight into areas that would otherwise be difficult to investigate, such as the bed of a glacier or ice sheet (e.g. Habermann et al., 2013; Larour et al., 2014). It also allows the evolution of bodies of ice to be predicted over the coming years to centuries (e.g. Gillet-Chaulet et al., 2012; Khazendar et al., 2015) or, similarly, hindcast to examine how ice masses achieved their present state (e.g. Bougamont et al., 2015). Earlier numerical models tended to use simplifications or parameterisations of the Navier-Stokes equations, such as the SIA, or the Shelfy-Stream Approximation (SSA) (e.g. Rutt et al., 2009). However, recent
computing advances mean that it is possible to run more complex models that do not simplify these equations and produce more realistic results, particularly for areas in which simplifications such as the SIA and SSA are known to be very poor approximations (Gagliardini et al., 2013). These include slower-flowing ice where vertical shear is non-negligible, or areas where rapid topographic change leads to ‘bridging effects’, affecting the basal stress, as is often seen on outlet glaciers (Morlighem et al., 2010). This makes these so-called ‘full-Stokes’ models particularly appropriate for simulating ice sheet interiors and glacier termini, or, as here, whole ice caps with a mix of slow and fast flow and several different types of outlet glacier.

One area of recent development in numerical modelling has been that of inversion techniques. These were first applied to glaciology by MacAyeal (1992, 1993). The central idea is that the model takes some (relatively) easily-observable surface properties, such as velocity, and uses these to reconstruct the distribution of an inaccessible basal property, such as friction/drag (e.g. Arthern et al., 2015; Gillet-Chaulet et al., 2012) or viscosity (e.g. Khazendar et al., 2015; Petra et al., 2012). This is particularly important, as basal processes and properties play a critical role in determining ice flow and cannot be otherwise readily observed in any practical sense (Truffer, 2004), as discussed in the previous chapter regarding sliding velocities. Inversion techniques have therefore become widely and successfully-used in many studies where knowledge of basal conditions is sought or required.

As such, Elmer/Ice, a 3D glaciological flow model that solves the full Navier-Stokes equations and can perform inversions (Gagliardini et al., 2013), was used to undertake the modelling component of this study. The model was set up to invert for basal friction based on the surface velocity inputs described in Section 3.1.3. Elmer/Ice is an open source model developed by the Finnish CSC-IT Centre for Science that uses the Finite Element Method to solve partial differential equations, including Navier-Stokes. This allows it to use meshes that vary spatially (and temporally), something not possible with the more common Finite Difference Method, and also makes parallel computing possible, both of which greatly increase its computational efficiency and are necessary to make complex inverse problems tractable.
3.2.5.1 – Model Components

To perform an inversion, the model needs to have two components – a forward model and an inverse model. The forward model solves the Navier-Stokes equations, whilst the inverse model calculates the inverted-for parameter. In the forward model used here, ice is assumed to be incompressible and isotropic, with its flow dependent on viscosity, as described by Glen’s Flow Law (Equation 2). The flow law exponent, $n$, was set to 3, the standard value for most glaciological applications and one regarded as giving a good compromise within the possible range of actual values where measurements are lacking (Paterson, 1994), as is the case here. The rate factor, $A$, in the absence of widespread available ice temperature profiles, was worked out at each node from the temperature profile calculated by the model through iteratively solving the convective diffusional equation for heat based on boundary conditions and additional strain heating generated by internal deformation. Whilst computationally more expensive, this was regarded as significantly more accurate than specifying a constant temperature. It is possible to invert for basal viscosity to improve estimates of $A$ further, but this affects the accuracy of the friction inversion (Petra et al., 2012), so was not pursued, being impractical within the constraints of this study.

The forward model also requires boundary conditions to be set at the edges of the domain. Having only one outflow boundary (i.e. the edge of the ice cap) and no inflow or lateral boundaries made this relatively straightforward for FI. Based on satellite imagery and the findings of Palmer et al. (2010), the domain boundary was split into water-terminating segments (the outlet glacier termini and the margin of the north-eastern lobe) and land-terminating ones. At the water-terminating segments, an external backstress from water pressure acting on the calving face was imposed, using the equation:

Equation 11: Water-terminating Boundary Condition

$$ P_w = \rho_w g (z_x - h_w) $$

Where $P_w$ is the water pressure, $\rho_w$ is the density of water, $z_x$ is the depth of the ice below the water, and $h_w$ is the water level elevation.

For the land-terminating sections, no such condition was necessary and the model was left to evolve freely. Similarly, as the model was to be used to run a static simulation, the top
surface of the ice cap was left as a free surface, with the MODIS temperature data from December of each year used to set its temperature in an effort to determine whether the base would attain the PMP even with an unrealistically-cold surface as part of a wider investigation into how surface temperature affected the base that had to be abandoned due to model runtime considerations, and accumulation and ablation ignored. At the bed, $\tau_b$ was initially defined based on the guess at the distribution of $\beta$ worked out using the SIA, described in Section 3.2.4. A basal temperature boundary condition was also imposed in the form of a geothermal heat flux of 75 mW m$^{-2}$, following data obtained for north-east Greenland by Davies (2013).

Turning to the inverse portion of the model, for this study, Elmer/Ice was set up to use the control method (also termed the adjoint method) to perform inversions, as introduced to glaciology by MacAyeal (1992, 1993). This is a widely-used and proven method, featuring in several recent studies (e.g. Gillet-Chaulet et al., 2012; Khazendar et al., 2015; Morlighem et al., 2010) and is based on minimising a cost function representing the mismatch between modelled and observed velocities. This process is described in detail by Gillet-Chaulet et al. (2012) and Gagliardini et al. (2013). To briefly summarise here, following Gillet-Chaulet et al. (2012), the initial cost function, $J_0$, is represented as:

**Equation 12 - Initial Cost Function**

$$ J_0 = \int_{\Gamma_s} \frac{1}{2} (|u_H| - |u_H^{obs}|)^2 \, d\Gamma $$

Where $u_H$ is the horizontal component of the modelled velocity and $u_H^{obs}$ is the observed velocity, both in m a$^{-1}$. Rather than being minimised with respect to $\beta$, which could produce negative values of $\beta$, and to produce a smoother output, $\beta$ is formulated as:

**Equation 13 - Beta Power Formulation**

$$ \beta = 10^{\alpha_p} $$

$J_0$ is then minimised with respect to $\alpha_p$ instead.

3.2.5.2 – Regularisation

Before minimisation is performed, however, $J_0$ is subject to Tikhonov regularisation (e.g. Konovalov, 2012). This penalises rapid spatial variations in $\beta$, producing a smoother output.
that improves model convergence and optimisation. This is done by adding a second term, \( J_{\text{reg}} \), to \( J_0 \), producing a new total cost function, \( J_{\text{tot}} \). The regularisation function, \( J_{\text{reg}} \), is defined as:

\[
J_{\text{reg}} = \int_{\Gamma_b} \left( \frac{1}{2} \left( \frac{d\alpha_p}{d\alpha_x} \right)^2 + \left( \frac{d\alpha_p}{d\alpha_y} \right)^2 \right) d\Gamma
\]

Giving \( J_{\text{tot}} \) as:

\[
J_{\text{tot}} = J_0 + \lambda J_{\text{reg}}
\]

Where \( \lambda \) is a tuneable parameter that defines the balance between mismatch between the modelled and observed velocities (represented by \( J_0 \)) and smoothness (represented by \( J_{\text{reg}} \)) within the overall cost function. \( \lambda \) was varied to find its optimal value, as discussed in Section 4.1.

3.2.5.3 – Minimisation

\( J_{\text{tot}} \) was minimised within the model using the M1QN3 routine developed by Gilbert and Lemaréchal (1989). This uses a limited-memory quasi-Newtonian iterative system based on the second derivatives of the cost function, making it more efficient than a fixed-step minimisation process. The convergence process towards this minimum, initialised from the SIA-derived estimate of \( \beta \), was accelerated using the Biconjugate Gradient Stabilised Method (BicGStab), presented in van der Vorst (1992), preconditioned through Incomplete LU factorisation.

3.2.6 – Temperature Profile

For subsequent calculations using the model outputs (detailed in Sections 3.2.7 to 3.2.9), a realistic ice temperature profile was required, which was not provided by the initial December-surface-temperature-forced inversions. A detailed investigation of temperature changes being beyond the scope of this study and, given the constant geometry imposed and the erroneous marginal thickness data producing unrealistic temperature gradients in many parts of the domain, it was finally decided to use a constant temperature profile for all
years from 2005 onwards. This was calculated at each point within the domain based on model temperature outputs determined from the annual average of the 2005 surface temperature data, as this was most representative of the 251 K annual average used by Willis et al. (2015). 2005 also falls near the middle of the ICESat measurement period used in deriving the DEMs employed here, so is likely to be a reasonable fit with these, and measures FI in a non-surge state, making it a reasonable approximation for the 2006, 2007, 2008 and 2015 datasets. However, it is not a good approximation for 2000, when NSG and SSG were surging, so a separate temperature profile was used for 2000, based on data for that year. To ensure the basal temperature distribution matched flow conditions, the 2000 and 2005 inversions were re-run with a basal temperature boundary condition imposed, forcing temperature to be at the PMP beneath sliding (surface velocity >30 m a⁻¹) ice. This value was chosen as, at these higher velocities, the subglacial till can be assumed to have failed, such that the sediment yield strength can be set equal to basal traction, greatly simplifying subsequent calculations. Values for other quantities (\(u_b\), \(\tau_b\), etc.) however, were taken from the original set of inversions to ensure comparability between years, which was found to produce very similar results to using data for these quantities from the re-run inversions, which reduced the required model runtime significantly.

3.2.7 – Calving Flux

One parameter of interest that is also possible to determine is the additional calving flux due to the surge of NSG and SSG. The two glaciers were assumed to be floating at their terminus and therefore in hydrostatic equilibrium, based on Palmer et al. (2010). The same assumption is also valid for TG, BEG and NG, whilst MSB is probably grounded at its terminus (Palmer et al., 2010). Whether RLGN, RLGE and HG are floating or grounded is not known. Therefore, an estimate of the calving flux for the whole of FI has been calculated based on the five glaciers with known floating termini, providing a minimum estimate of the flux. Ultimately, better thickness data is needed if more accurate estimates are to be determined. Depth-averaged terminus velocity, assumed to be 85% of surface velocity (Paterson, 1994), was used and the results are shown in Table 4. Evidently, the assumption of constant geometry is unlikely to be realistic, so the results should be treated as purely indicative of general trends.
3.2.8 – Basal Thermal Regime

The resulting outputs from the modelling exercise can be used to calculate the amount of water being produced or lost subglacially on FI. This is of inherent interest, as it provides an additional mass loss metric, but can also be used to help estimate the return period of the identified surge cycle through providing an indication of how long it will take water storage in the till (and thus PWP) to reach the yield strength, allowing the till to fail and initiating rapid sliding (Equation 8). The approach taken here was to follow Christoffersen et al. (2014), where the rate of basal melting, \( m \), is calculated as:

*Equation 16 - Rate of Basal Melting*

\[
m = \frac{\tau_b u_b + G - K \theta_b}{\rho_i L}
\]

Where \( G \) is the geothermal heat flux (75 mW, following the model boundary condition); \( K \) and \( \theta_b \) are thermal conductivity of ice and vertical basal ice temperature gradient, which, together, are the conductive heat loss; and \( L \) is specific latent heat of fusion (333.5 kJ kg\(^{-1}\)). The thermal conductivity, \( K \), depends on the temperature of the ice such that (Paterson, 1994):

*Equation 17 - Thermal Conductivity of Ice*

\[
K = 9.828 \exp(-5.7 \times 10^{-3} T_i)
\]

Equation 16 and Equation 17 were evaluated for each year using the temperature profile described in Section 3.2.6 and model outputs for the remaining variables. The calculations were performed based on sliding areas of the ice cap with initial thickness >30 m as, as discussed in Section 3.2.2, these had more reliable results and would be the areas where meltwater would be expected to be present. The resulting data were then scaled up to the whole sliding area of FI, assuming they were representative of subglacial conditions in sliding parts of the ice cap. The results are presented in Table 5.

3.2.9 – Till and Subglacial Water Storage

As discussed in Section 2.2, it is very likely that surging and glacier sliding on FI are mediated by the presence of a subglacial till layer, based on the observed characteristics of the surge. Investigating the properties of this till layer is therefore crucial. Two of the key parameters
are the sediment yield strength and the void ratio. Following Bougamont et al. (2014), it is possible to work out the sediment yield strength (Equation 8) as:

Equation 18 - Sediment Yield Strength

\[ \tau^* = N_0 \tan(\Phi) 10^{-(e-e_0)/C} \]

Where \( e \) is the sediment void ratio (defined as the volume of pores to solids in the till layer), \( e_0 \) is the void ratio value at the reference value of effective normal stress, \( N_0 \), and \( C \) is the sediment compression index. Values of \( e_0 \), \( N_0 \), \( C \) and \( \Phi \) are, again following Bougamont et al. (2014), set to those of till from beneath Trapridge Glacier in Alaska (Clarke, 1987), which have been shown to be similar for those for till from beneath glaciers in Svalbard (Murray, 1997), suggesting this is a sensible approximation for FI.

To solve Equation 18 for \( e \), it was assumed that \( \tau^* \) was equal to basal traction calculated from the model inversions beneath fast-moving (>30 m a\(^{-1}\)) ice where it was reasonable to assume the till had failed. The resulting void ratios were then averaged to provide a representative value. As with the basal melting calculations in Section 3.2.8, only parts of FI with thickness >30 m were considered in the initial calculation.

To calculate how porosity changes related to water volumes, it was assumed that FI was underlain by a uniform till layer. Three scenarios were used for the thickness of this layer: 0.2 m, 0.5 m and 1 m, based on the work of Murray and Porter (2001), who found minimum deforming-layer till thickness of between 0.08 and 0.55 m, with a mean of 0.2 m, beneath Bakaninbreen, a surging glacier in Svalbard. 0.2 m therefore provides a plausible minimum estimate of the till’s water storage capacity, with 0.5 m representing a likely maximum and 1 m an absolute upper bound. The calculated void ratios, along with the melt rates from Equation 16, were then scaled up to the entire sliding area of FI. The results are presented in Table 6.
Chapter 4 – Results

This chapter presents the results from the modelling exercise and calculations described in Chapter 3, as well as velocity data on FI’s major outlet glaciers.

4.1 – Model Optimisation

To optimise the model, the final step in the inversion process, the most appropriate value of the regularisation parameter, $\lambda$, for the domain had to be chosen, such that the optimal balance between mismatch and smoothness was found (Konovalov, 2012). This can be done by plotting the initial cost function ($J_0$) against the regularisation function ($J_{reg}$) for different values of $\lambda$ in the L curve method (e.g. Konovalov, 2012). This produces an L-shaped curve, the corner of which is, objectively, the value of $\lambda$ that produces the best compromise between the two components of the total cost function.

For this study, the L curve (Figure 8) was produced based on several runs using the 2005 dataset. As can be seen, $J_{reg}$, after an initial small increase, remains virtually unchanged up to $\lambda=10^7$, before increasing rapidly beyond that point. $J_0$, meanwhile, drops rapidly until $\lambda=10^8$, and continues dropping beyond this point at a decreasing rate. $\lambda=10^7$ (circled in red) was consequently selected as the optimal value of the regularisation parameter, as moving...
beyond this point leads to large increases in $J_{reg}$ at ever smaller gains in $J_0$, whilst smaller values of $\lambda$ provide negligible gains in $J_{reg}$ whilst $J_0$ increases substantially. This value therefore provides the best overall compromise between mismatch and smoothness when running the model for this domain and was used in all model runs.

### 4.2 — Basal Conditions

This section includes the results from the model inversions for basal drag, as well as displaying the temperature data used in subsequent calculations.

#### 4.2.1 — Basal Drag

As can be seen from Figure 9, the central part of FI has quite high values of $\tau_b$, with particularly widespread patches underneath the north-eastern and eastern lobes (on the left in Figure 9, as the point of view is from underneath the ice cap, rather than above it). There are also significant areas of greater drag in the upper reaches of the NSG/SSG basin, just before fast flow begins, and in a similar position in the MSB drainage basin, extending down the northern side of the trunk. The very slight drag present beneath the margins and southern third of FI should not be regarded as an accurate result, as it is a product of the unrealistic thicknesses caused by imprecisions in the derivation of the basal DEM, as is the area of very large drag in the upper reaches of HG. However, the patch of lesser drag evident in the very centre of the ice cap in all years is a robust result. It is most likely due to the very slow flow rates and surface slopes in this region imposing little in the way of driving stress or friction.

Year-on-year, there is little obvious pattern in the changes in $\tau_b$ (Figure 10). There is a general indication of an increase between 2000 and 2005 (Figure 7a), particularly around the upper reaches of NSG and SSG, a trend which seems to continue to an extent throughout the study period, but, otherwise, most areas alternate between increasing and decreasing, with little apparent order. This may partly be an artefact of the relatively low value of $\lambda$, which leads to the patchwork appearance of Figure 9 and Figure 10, obscuring
more general trends. Rerunning the model inversions with a higher value of $\lambda$ might yield more useful results here, at the price of a greater mismatch with velocity observations.

Focussing on the areas of the ice cap sliding at velocities greater than 30 m a\(^{-1}\) and with thickness exceeding 30 m, which are those used in later calculations, Table 2 presents the average values of $\tau_b$ and basal velocity in these areas. Errors are included where appropriate, using the standard deviation as an indicator of the uncertainty to provide a guide to the range of the underlying data. This gives very high apparent uncertainty on much of the data, due to the wide range of values present, but the patterns and relative changes remain valid.

Table 2 - Average $\tau_b$ and velocity beneath Sliding Area of FI

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</tr>
</thead>
<tbody>
<tr>
<td>$\tau_b$ (kPa)</td>
<td>22±47</td>
<td>38±45</td>
<td>19±34</td>
<td>31±40</td>
<td>30±36</td>
<td>20±27</td>
</tr>
<tr>
<td>Basal Velocity (ma(^{-1}))</td>
<td>123±88</td>
<td>54±16</td>
<td>55±17</td>
<td>55±17</td>
<td>56±18</td>
<td>55±13</td>
</tr>
</tbody>
</table>

One notable anomaly obvious from Table 2 is the low $\tau_b$ value in 2006 and 2015, at around 66% of the figure in other post-surge years, which feeds through into the figures for basal melt and till water storage in Table 5 and Table 6. These unusually low results seem to be due to the expected velocity-dependent changes in the area used in the basal calculations for those years excluding some higher-drag areas that were included in other years. As such, these figures are not representative of the true situation beneath the ice cap in those years; an assertion borne out by the fact that the surface and basal velocity in these areas in both 2006 and 2015 were very similar to 2005, 2007 and 2008. If basal drag were really that much lower in 2006 and 2015, it would be expected that velocity would increase somewhat to achieve force balance, which is not the case. As such, undue importance should not be
Figure 9: $\tau_b$ on FL by Year
Figure 10 - Difference in $\tau_b$ between Datasets
attached to the figures for these two years and the discussion will focus on the results from

The second major point arising from these results is that surging led to a 33% reduction in
\( \tau_b \) compared to the average for 2005, 2007 and 2008. This would be expected, given the
larger area subject to faster flow.

4.2.2 – Temperature

Figure 11 shows the temperature distribution used in performing calculations for the year
2000, when NSG and SSG were both surging. As can be seen in Panel b., at least a part of all
the major outlet glaciers is subject to the basal boundary condition of being at the PMP.
What is interesting is that some of the eastern-central part of the ice cap also seems to be
coming very near to the PMP, if not quite reaching it, suggesting that FI may have a warmer
base overall than previously believed.

Figure 12 shows the temperature distribution used in performing calculations for 2005 and
all subsequent years. As with 2000, there is an indication that some of the eastern-central
parts of the ice cap may be reaching temperatures close to the PMP, which is unsurprising,
given the constant geometry. The reduced extent of sliding in the basins of NSG and SSG can
also be clearly seen in the reduced area in b. at the PMP, compared to in Figure 11.

Figure 11 - 2000 Temperature Distribution. a. shows surface temperatures, b. shows basal temperatures.
Example temperature profiles from Figure 12 are displayed in Figure 13. The left-hand figure shows a profile from the eastern-central part of the ice cap, where ice is flowing slowly and the bed is remaining just below the PMP. The right-hand profile is from the upper reaches of NSG, where ice is flowing fast and the bed is at the PMP. In slow-flowing regions, the profile is strongly convex, with a cold surface layer a few tens of metres thick above a column of ice warmed by englacial ice deformation and trapped geothermal heat (there being very little frictional heat). In fast-flowing regions, a very concave temperature profile is instead observed due to the advection of cold surface ice from further upstream suppressing temperatures in the upper part of the ice column. This creates a very strong basal temperature gradient, promoting freezing.
4.3 – Glacier Velocity

This section presents data on the velocities of the major outlet glaciers of FI identified in Table 1.

Table 3 - Average Velocities for FI Outlet Glaciers

<table>
<thead>
<tr>
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</tr>
</thead>
<tbody>
<tr>
<td>NSG</td>
<td>268±51</td>
<td>261±26</td>
<td>24±18</td>
<td>36±24</td>
<td>26±24</td>
<td>29±24</td>
<td>17±16</td>
</tr>
<tr>
<td>SSG</td>
<td>271±56</td>
<td>393±64</td>
<td>79±23</td>
<td>65±30</td>
<td>64±29</td>
<td>67±25</td>
<td>44±17</td>
</tr>
<tr>
<td>MSB</td>
<td>56±19</td>
<td>49±16</td>
<td>50±17</td>
<td>59±21</td>
<td>52±16</td>
<td>47±17</td>
<td>62±22</td>
</tr>
<tr>
<td>RLGN</td>
<td>76±45</td>
<td>98±38</td>
<td>98±38</td>
<td>88±36</td>
<td>90±36</td>
<td>90±41</td>
<td>86±41</td>
</tr>
<tr>
<td>RLGE</td>
<td>58±18</td>
<td>48±17</td>
<td>55±18</td>
<td>53±21</td>
<td>59±16</td>
<td>55±18</td>
<td>39±15</td>
</tr>
<tr>
<td>HG</td>
<td>36±37</td>
<td>101±70</td>
<td>99±69</td>
<td>86±63</td>
<td>83±54</td>
<td>68±46</td>
<td>38±16</td>
</tr>
<tr>
<td>TG</td>
<td>78±34</td>
<td>74±25</td>
<td>53±22</td>
<td>53±22</td>
<td>56±24</td>
<td>48±23</td>
<td>25±10</td>
</tr>
<tr>
<td>BEG</td>
<td>31±30</td>
<td>12±11</td>
<td>5±3</td>
<td>19±16</td>
<td>2±1</td>
<td>7±3</td>
<td>7±5</td>
</tr>
<tr>
<td>NG</td>
<td>48±16</td>
<td>62±19</td>
<td>54±16</td>
<td>58±10</td>
<td>51±18</td>
<td>61±19</td>
<td>48±12</td>
</tr>
</tbody>
</table>

Figure 13 - Example Temperature Profiles. The left-hand graph shows a profile for thick, slow-flowing interior ice; the right-hand one shows the same for thinner, sliding ice nearer the margin.
As would be expected with surging glaciers, Table 3 clearly shows a substantial decrease in surface velocity between the active phase (2000) and quiescent phase (2005) for both NSG and SSG. For NSG, this drop is about 90%; for SSG, about 80%. To put it another way, the transition to the active phase led to a tenfold speed-up of NSG and a fivefold speed-up of SSG, on average. After 2005, velocity on both glaciers remains relatively constant, though with a continued slight declining trend. The other major outlet glaciers all show a fairly constant velocity, except HG, which nearly triples its velocity between 1996 and 2000, before returning to its original value by 2015. However, this is the narrowest glacier on FI, so it is possible that the MEaSUREs velocity data is less reliable for it, as indicated by the high standard deviation, and it is in the complex southern portion of the ice cap with poor thickness data and thus not well-represented in the model, so no further investigation of this is envisaged within the bounds of this study.

Figure 14 - Velocity Profile along NSG
Focussing on the surging glaciers, Figure 14 and Figure 15 both show a pronounced area of faster velocity in the upper reaches of the glaciers, especially for NSG. In both cases, the prominence of this area seems to decrease over the course of the surge, with flatter velocity profiles apparent in 2000 compared to 1996. Post-2000, velocity reduces substantially, but this upper area of quicker velocity remains apparent, particularly on NSG.

4.4 – Surge Effects

This section presents calculations for calving flux, subglacial meltwater production and till porosity based on the data in Sections 4.2 and 4.3.

4.4.1 – Calving

As Table 4 shows, the surge of NSG and SSG led to a fifteen-fold and seven-fold increase, respectively, in their calving fluxes. This was an increase of sufficient magnitude that it led to the total calving flux for the basins assessed here to increase approximately 450%, compared to post-2000. However, this was also due to changes at BEG, where a similar pattern of greater flux in 1996 and 2000 was observed, with its larger thickness giving it a disproportionate effect on the overall calving flux, despite its slower speed, such that it
contributed around half the total calving flux in 1996 and remains the dominant influence on the total calving flux throughout the study period. Terminus velocities at BEG were ten times higher in 1996 (~100 ma\(^{-1}\)) compared to the average for 2005-15 (~10 ma\(^{-1}\)), so it would seem clear that a surge occurred here too, contemporaneous with that at NSG and SSG. Based on Table 3, though, it is clear this surge only affected the lower few kilometres of the glacier, as the average velocity remains low throughout the time period of this study.

The flux at NG, on the other hand, remained relatively constant throughout, though the terminus velocities and resulting flux at TG suggest some sort of mini-surge occurred here between 1996 and 2000, with fluxes three times higher than post-2000.

The total figures suggest FI’s calving flux is nearly two orders of magnitude smaller, in non-surge conditions, than that of the similarly-sized Austfonna ice cap in Svalbard, at around 2.5 Gt a\(^{-1}\) (Dowdeswell et al., 1999). These figures represent a lower bound, as velocities will generally be greater in summer than in winter, due to extra lubrication from surface meltwater, and several outlet glaciers have not been included due to insufficient data.
4.4.2 – Subglacial Water Changes

Using all the data presented in Sections 4.2 to 4.3, the basal melt rates and the consequent flux due to subglacial meltwater changes (i.e. the amount of water produced by basal melting or lost due to basal freezing) can be calculated.

Table 5 - Subglacial Water Changes. The average melting and freezing rates, as well as the percentage of the sliding area to which they are applicable, are given first, before the overall melt rate and the resulting changes in subglacial water volumes are presented.

<table>
<thead>
<tr>
<th>Year</th>
<th>Basal Melt (mm a⁻¹)</th>
<th>Sliding Area Melting (%)</th>
<th>Sliding Area Freezing (%)</th>
<th>Average Freezing Rate</th>
<th>Average Freezing Rate</th>
</tr>
</thead>
<tbody>
<tr>
<td>2005</td>
<td>8.33 ± 6.84</td>
<td>52</td>
<td>48</td>
<td>-16.16 ± 10.65</td>
<td>-3.48 ± 15.12</td>
</tr>
<tr>
<td>2006</td>
<td>7.85 ± 6.71</td>
<td>52</td>
<td>48</td>
<td>-12.00 ± 10.50</td>
<td>-1.62 ± 13.21</td>
</tr>
<tr>
<td>2007</td>
<td>8.18 ± 5.78</td>
<td>48</td>
<td>52</td>
<td>-13.23 ± 10.61</td>
<td>-3.00 ± 13.75</td>
</tr>
<tr>
<td>2008</td>
<td>8.26 ± 5.24</td>
<td>53</td>
<td>47</td>
<td>-14.07 ± 10.72</td>
<td>-2.23 ± 13.89</td>
</tr>
</tbody>
</table>

Table 5 shows several interesting features. First, the area of the bed subject to melting was halved during the surge (2000), compared to after it (other years), which is unexpected given the higher velocities. Second, and related, the fact that the surge did not lead to a substantial increase in the production of subglacial meltwater, but did greatly enhance
freezing. Third, that freezing dominates melting in all years leading to net losses of subglacial water, despite the observed continued fast flow throughout the study period. It is also worth pointing out that the resulting changes in subglacial water volumes are fairly minor, which is unsurprising given the relatively small size of FI. Overall, with regards to surge compared to non-surge conditions, Table 5 shows that the surge of NSG and SSG increased the average melt rate by 15%, the average freezing rate by 32% and the resulting average water volume change rate by nearly 300%, whilst the area subject to melting halved and the area subject to freezing increased by 50%.

From the data in Table 5, Figure 16 shows the distribution of melting and freezing in the areas of the NSG, SSG and MSB basins that fulfilled the criteria of >30 m ice thickness and >30 m a\(^{-1}\) velocity indicative of basal motion by till failure and deformation. The most important point to note is that the areas of melting (red) and freezing (blue) are both relatively spatially-constant across all years post-2000, with melting towards the terminus of MSB, under the southern portion of the upper region of NSG, and in discrete patches under the upper part of SSG.
Figure 16 - Melt Rate Detail for NSG, SSG and MSB (top to bottom) by Year for 2000 (a.), 2005 (b.) and 2007 (c.)
### 4.4.3 – Till Properties

The changes in void ratio and, thus, water storage in the assumed underlying till can also be calculated, which can show the effect of the surge on the underlying sediment.

*Table 6 - Till Void Ratios and Resulting Water Storage Changes. The results are shown for the three different till thickness scenarios considered in this study.*

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</thead>
<tbody>
<tr>
<td></td>
<td>±0.11</td>
<td>±0.05</td>
<td>±0.05</td>
<td>±0.05</td>
<td>±0.05</td>
<td>±0.04</td>
<td></td>
</tr>
<tr>
<td>Proportion of FI</td>
<td>17.46%</td>
<td>8.51%</td>
<td>8.02%</td>
<td>7.73%</td>
<td>7.80%</td>
<td>5.98%</td>
<td></td>
</tr>
<tr>
<td>0.2m Till Water Storage (km³)</td>
<td>Total</td>
<td>0.1633</td>
<td>0.0669</td>
<td>0.0685</td>
<td>0.0624</td>
<td>0.0628</td>
<td>0.0491</td>
</tr>
<tr>
<td>Difference from Preceding Year</td>
<td>n/a</td>
<td>-0.0964</td>
<td>0.0016</td>
<td>-0.0061</td>
<td>0.0004</td>
<td>-0.0137</td>
<td></td>
</tr>
<tr>
<td>0.5m Till Water Storage (km³)</td>
<td>Total</td>
<td>0.4082</td>
<td>0.1672</td>
<td>0.1712</td>
<td>0.1560</td>
<td>0.1569</td>
<td>0.1227</td>
</tr>
<tr>
<td>Difference from Preceding Year</td>
<td>n/a</td>
<td>-0.2410</td>
<td>0.0040</td>
<td>-0.0152</td>
<td>0.0010</td>
<td>-0.0342</td>
<td></td>
</tr>
<tr>
<td>1.0m Till Water Storage (km³)</td>
<td>Total</td>
<td>0.8164</td>
<td>0.3343</td>
<td>0.3424</td>
<td>0.3119</td>
<td>0.3139</td>
<td>0.2454</td>
</tr>
<tr>
<td>Difference from Preceding Year</td>
<td>n/a</td>
<td>-0.4821</td>
<td>0.0080</td>
<td>-0.0304</td>
<td>0.0019</td>
<td>-0.0684</td>
<td></td>
</tr>
</tbody>
</table>

Table 6 shows that the surge of NSG and SSG doubled the area of the ice cap subject to fast flow, and where the till had presumably failed, as well as leading to a large water withdrawal from the till aquifer, though the exact magnitude increase of this on the non-surgering yearly rate is difficult to determine, as the surge could have ended any time between 2000 and 2005. Assuming it is spread equally across the entire interval, this
represents a nearly tenfold increase on the mean magnitude of yearly water storage change for 2006-8.

4.4.4 – Summary: Surge versus Non-surge

Table 7, below, summarises the findings of this study for the whole of FI by year.

Table 7 - Key Statistics for FI by Year. All characteristics are averages from those areas with thickness of at least 30m and velocity of at least 30 m a\(^{-1}\). Temperature figures are from the re-run inversions with PMP boundary condition. 1996 is excluded due to not being part of the modelling exercise.

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</tr>
</thead>
<tbody>
<tr>
<td><strong>Surface Velocity (m a(^{-1}))</strong></td>
<td>121±88</td>
<td>55±17</td>
<td>53±17</td>
<td>54±17</td>
<td>55±19</td>
<td>56±14</td>
</tr>
<tr>
<td><strong>Basal Velocity (m a(^{-1}))</strong></td>
<td>123±88</td>
<td>54±16</td>
<td>55±17</td>
<td>55±17</td>
<td>56±18</td>
<td>55±13</td>
</tr>
<tr>
<td><strong>Surface Temperature (K)</strong></td>
<td>253±1</td>
<td>253±1</td>
<td>253±1</td>
<td>253±1</td>
<td>253±1</td>
<td>253±1</td>
</tr>
<tr>
<td><strong>Basal Temperature (K)</strong></td>
<td>272±4</td>
<td>272±5</td>
<td>272±5</td>
<td>272±5</td>
<td>272±5</td>
<td>272±5</td>
</tr>
<tr>
<td><strong>(\tau_b) (kPa)</strong></td>
<td>22±47</td>
<td>38±45</td>
<td>19±34</td>
<td>31±40</td>
<td>30±36</td>
<td>20±27</td>
</tr>
<tr>
<td><strong>Calving Flux (km(^3) a(^{-1}))</strong></td>
<td>0.274</td>
<td>0.043</td>
<td>0.049</td>
<td>0.036</td>
<td>0.057</td>
<td>0.072</td>
</tr>
<tr>
<td><strong>Basal Melt Rate (mm a(^{-1}))</strong></td>
<td>-11.28±16.26</td>
<td>-3.48±15.12</td>
<td>-1.62±13.21</td>
<td>-3.00±13.75</td>
<td>-2.23±13.89</td>
<td>-2.09±12.66</td>
</tr>
<tr>
<td><strong>Subglacial Meltwater Change (km(^3) a(^{-1}))</strong></td>
<td>-0.017</td>
<td>-0.003</td>
<td>-0.001</td>
<td>-0.002</td>
<td>-0.001</td>
<td>-0.001</td>
</tr>
<tr>
<td><strong>Void Ratio</strong></td>
<td>0.55±0.11</td>
<td>0.46±0.05</td>
<td>0.50±0.05</td>
<td>0.47±0.05</td>
<td>0.47±0.05</td>
<td>0.48±0.04</td>
</tr>
<tr>
<td><strong>Till Water Storage Change on Preceding Year (km(^3))</strong></td>
<td>n/a</td>
<td>-0.0964 to -0.4821</td>
<td>0.0016 to 0.0080</td>
<td>-0.0061 to -0.0304</td>
<td>0.0004 to 0.0019</td>
<td>-0.0137 to -0.0684</td>
</tr>
</tbody>
</table>
Focussing specifically on the differences between the surging and non-surring states of Fl, represented by the data for 2000 and 2005, respectively, it is clear that during surging, velocities, calving fluxes, basal freezing rates, void ratios and subglacial water volume changes were much greater, whilst \( \tau_b \) dropped significantly. To explore this, Figure 17 and Figure 18 show the \( \tau_b \) and basal temperature distributions for 2000 and 2005, respectively, as well as the difference between the two years. As has been remarked in Section 4.2.1, there is a general indication of strengthening \( \tau_b \) between the two years across the whole ice cap, with the area of high \( \tau_b \) in the upper regions of NSG and SSG seeming to strengthen and grow. This fits with the generally stiffer till (Table 6) and higher \( \tau_b \) (Table 2) found in 2005 compared to 2000. With regards to temperature, unsurprisingly, there is a substantial cooling underneath the surge basin, as a large part is no longer subject to the forced-PMP boundary condition. Elsewhere on Fl, substantial cooling also looks to have taken place underneath the centre of the ice cap, as well as under the north-eastern lobe, with relatively muted changes elsewhere.

Figure 17 - Evolution in \( \tau_b \) between 2000 and 2005. a. shows 2000; b. shows 2005; and c. shows the difference between 2000 and 2005.
Figure 18 - Evolution in Basal Temperature between 2000 and 2005. a. shows 2000; b. shows 2005; and c. shows the difference between 2000 and 2005.
Chapter 5 – Discussion

This section discusses the results set out in Chapter 4. A brief discussion of the general state of FI comes before a section on the surge mechanism and one on surge effects.

5.1 – State of Flade Isblink

As would be expected for such a High-Arctic ice cap, the model outputs show that the bed of FI is largely frozen, though perhaps not to the degree suggested by Willis et al. (2015) with their calculated temperature of 264 K. Much of the bed below the thicker ice in the central portion of the ice cap instead comes to within a few degrees of melting. This difference is due mainly to differing assumptions over the geothermal heat flux: 60 mW in their study; 75 mW here. The dominance of freezing supports the figures in Table 5, too, as it suggests the ‘natural’ state of FI’s bed is cold-based, conforming to the strong freezing rates calculated here. The fact that sliding-induced high velocities are observed despite this suggests a resemblance to ice streams in Antarctica, where fast flow occurs in the presence of widespread basal freezing (e.g. Christoffersen et al., 2014).

With regards to $\tau_b$, the general pattern of higher stresses and a stickier bed under the slow-flowing centre of FI, with lower stresses and a slipperier bed where fast flow starts to initiate is what would be expected. Given the data constraints in this study, an account of changes under individual outlet glaciers is not practicable with any great degree of certainty, but one robust feature is observable: the area of high stress just before the transition to fast flow in both the NSG/SSG and MSB basins. This is probably the result of a greater driving stress, due to steeper surface gradients between the slow-flowing ice further inland and the lower-lying, sliding ice of the outlet glaciers, combined with greater longitudinal stress gradients as the slower ice is pulled along by the sliding ice, all on a still-frozen bed of strong till that resists flow.

A further general point to make about FI is that it shows an unexpected range of dynamic activity. In addition to the surge of NSG and SSG, which is itself somewhat surprising on a small, cold-based, High-Arctic ice cap, there is also evidence in Table 3 and Table 4 of a spatially-confined surge at around the same time at the terminus of BEG, leading to an order-of-magnitude increase in its calving flux. Furthermore, in the same data, there seems
to be a large degree of flow variability on HG, with velocities changing by a factor of three over several years, though perhaps not to the extent that it can be called a surge. Overall, this suggests that, as the range and availability of observations continue to improve, other areas of ice that have previously been regarded as inactive and of negligible interest may turn out to be surprisingly dynamic, which may require a reassessment of their vulnerability and contribution to SLR.

5.2 – Surge Mechanism

As posited in Section 2.2, based on initial observations, further analysis shows that the surge of NSG and SSG clearly has the characteristics of a Svalbard-esque thermally-controlled, soft-bed surge, as set out by Murray et al. (2003). The change in velocity (Table 3) between the quiescent and active phases was relatively small, less than an order of magnitude. Combined with the long duration of the active phase (at least five years), both of these suggest the more gradual thermally-regulated mechanism rather than the hydrologically-driven one of Kamb et al. (1985). The area of continued higher velocities in the upper reaches of both NSG and SSG seen in Figure 14 and Figure 15 is best interpreted as where thicker ice accumulates, allowing the bed to warm and initiating the surge. If more surface elevation data were available from during the surge, it would be interesting to plot surface elevation profiles of both glaciers to see whether any kind of surge front were observable, as this would provide additional useful information on the surge mechanism. Particularly, whether the area of warm-based, mobilisable ice expanded faster than the area actually sliding.

However, the data presented here also indicate a refinement to the mechanism. Usually, thermally-controlled surges are thought to terminate when the ice thins to the extent that the temperature gradients at the base become sufficiently steep to evacuate the trapped heat and cause basal freezing. Here, though, the assumption of constant geometry has shown that another mechanism comes into play to cause surge termination. As shown in Table 5, during the surge, basal freezing rates increased by around a third and basal freezing was present over nearly three quarters of the sliding area, rather than half in non-surge years. The primary reason for this, given that the other variables in Equation 16 were kept constant or very similar, is to be found in the interplay of \( \tau_b \) and \( u_b \), the determinants of the
frictional heat (the product of $\tau_b$ and $u_b$). Although a surge leads to a large increase in $u_b$, this is counter-balanced by the drop in $\tau_b$ as greater volumes of liquid water produced during the surge (the surge also increased melt rates by 15%) reduce the effective pressure, weakening the till. The overall consequence may be, as here, that a surge actually increases the degree of basal freezing even without any thinning of the overlying ice, leading to the base refreezing and/or large water withdrawals from the till, strengthening it and slowing flow, both of which act to terminate the surge. In reality, both of these termination mechanisms (conductive heat loss by thinning and frictional heat reduction through till weakening) would operate at the same time and it is also probable that any frictional heat reduction is very dependent on local factors (i.e. how much the glacier speeds up compared to how much the till is weakened), but it is important to note that thermally-regulated surges seem able to terminate even under constant geometry.

One important point that needs to be mentioned is the persistence of fast flow despite what appears to be widespread and dominant basal freezing and insufficient frictional heat production. Fast, sliding flow requires there to be liquid water below the ice, yet the overall negative melt rates presented here indicate that any basally-produced meltwater should be quickly lost again. This does not necessarily mean that these areas have to be frozen to the bed, as a freezing sediment can still be wet and soft, provided the water is replaced in some fashion, but indicates there must be an additional source of meltwater or heat. Analogues for this situation do exist, as noted in Section 5.1: Antarctic ice streams, which themselves display significant velocity variability and surge-like behaviour on centennial scales (Christoffersen et al., 2014). Christoffersen et al. (2014) showed that the beds of the fast-flowing (hundreds of metres per year) Siple Coast ice streams experienced melt rates of up to 15 mm a$^{-1}$ and freezing rates of up to 20 mm a$^{-1}$, with freezing dominating, in excellent agreement with the figures in Table 5. For these ice streams, fast flow was maintained by significant hydrological inputs from further inland, as well as net water withdrawal from the underlying till layer, counteracting the local basal freezing. How applicable this is to FI is discussed further in Section 5.3.
5.3 – Surge Effects

The figures for calving flux, meltwater production and water storage changes in the till aquifer in Table 7 are in good agreement with the finding of Rinne et al. (2011) that the mass balance of Fl was 0.0±0.5 Gt a\(^{-1}\). Rinne et al. (2011) calculated that the average surface elevation change over the entire ice cap between 2002 and 2009, overlapping well with the time period of this study, was 0.03±0.03 m a\(^{-1}\), which is equivalent to a mass gain of 0.255±0.255 km\(^3\) a\(^{-1}\). As can be seen from Table 7, this would comfortably balance the calving flux, the apparent main-non-SMB source of mass loss on Fl, every year except 2000, where the surge of NSG and SSG causes atypical conditions on the ice cap. Even for 2000, the calving flux figure is within the uncertainty of the surface elevation change, so even if this degree of calving were maintained for the whole duration of the active phase, the overall mass balance of Fl would remain within the bounds calculated by Rinne et al. (2011), though it should be emphasised again that the calving flux figures calculated here are very much a minimum estimate. Therefore, during surges, it is likely that the excess calving flux is not entirely balanced by the positive SMB. Compared to the SMB and calving flux changes, the subglacial meltwater volume changes are one to two orders of magnitude smaller, so do not significantly change the analysis. The same is generally true of changes in the till storage if the till is towards the thinner end of the distribution considered, which seems to be the case, as discussed in the following paragraphs.

Assuming the substantial water withdrawal from the till aquifer between 2000 and 2005 in Table 6 is entirely due to the observed surge, this allows an upper bound to be placed on the return period of the active phase (i.e. the interval between active phases, or, alternatively, the length of the quiescent phase) for these glaciers, assuming that this water must be fully-replaced to initiate a new surge. The surge basin occupies \(-18\% (-1500 \text{ km}\(^2\)) of the total surface area of Fl, of which \(-950 \text{ km}\(^2\) (-11\% of Fl) remains flowing at over 30 m a\(^{-1}\) in the quiescent phase. Of this, based on the figures in Table 5, just under 51\% (average for 2005, 2007 and 2008) will be subject to melting and to which the calculated melt rates are thus applicable. If the water withdrawn from the till is solely replaced by basal melting within this subset of the basin, which can be assumed to be equal to the average of the figures for 2005, 2007 and 2008 presented in Table 5, given the similarity in flow speeds of all the major glaciers outside the active phase (Table 3), this would mean the glaciers would
be ready to surge again in between 25 and 123 years, depending on the assumption made about the thickness of the till layer. This is, again, a minimum estimate, as some of the water storage change will be from the non-surfing parts of the ice cap, and additional water input will be provided by surface melting in the summer.

To narrow this range, the calculated water withdrawal from the till during the active phase in Table 5 can be compared to the changes in till water storage in Table 6. During the active phase, 0.017 km$^3$ of water were withdrawn annually. Given the active phase continued for at least five years (1996-2000) and assuming this number is applicable to all active-phase years, this gives a total minimum water withdrawal of 0.085 km$^3$, close to the storage change between 2000 and 2005 for a uniform till layer of 0.2 m thickness. A plausible upper bound is provided by reference to Svalbard, where the active phase of surge cycles has been observed to last up to a decade (Dowdeswell et al., 1995). Assuming this is applicable to FI, this gives a maximum water withdrawal of 0.17 km$^3$, which would equate to a maximum till thickness of approximately 0.35 m, well within the range of till thicknesses found below Bakaninbreen in Svalbard by Murray and Porter (2001). These till thickness values would then suggest the active phase return period lies towards the lower end of the possible range, at between 25 and 43 years, which would place it among the quicker surge cycles, compared with similar ones on Svalbard, but plausible (Dowdeswell et al., 1995; Murray et al., 2003). This figure is supported by Higgins (1991), who, based on aerial photographs from 1961 and 1978, determined that NSG was flowing at 175 m a$^{-1}$, whilst SSG was moving at 360 m a$^{-1}$. As shown in Table 3, Figure 14 and Figure 15, these speeds are characteristic of the active phase of these glaciers’ surge cycle, indicating that surging must have continued for a substantial period of time between 1961 and 1978, which supposes a return period of ~20-40 years, in excellent agreement with the figures from this study. This also indicates that water storage changes in other parts of the ice cap made a negligible contribution to the water storage change between 2000 and 2005. Therefore, taking 2005 as the end of the latest active phase, the next active phase should start between 2025 and 2048 if the ice cap remains largely stable.

However, the above is based on the assumption that all the water produced by basal melting is stored in the till until such time as a surge occurs, when it becomes lost through the increased basal freezing. But, as Table 5 shows and as highlighted in Section 5.2, basal
freezing predominates across FI in all years, meaning this seems unlikely, even if areas of melting and freezing seem to be fairly constant across time, as shown in Figure 16, such that a small region could potentially become saturated, starting the surging process. Therefore, there must be an additional source of hydrological input and/or an additional thermal source that promotes greater melting. Looking first at potential hydrological inputs, three possibilities arise: water withdrawals from the till aquifer, subglacial water from further inland, or surface meltwater. The first two were the primary hydrological inputs on the Siple Coast (Christoffersen et al., 2014), but seem unlikely here. Based on the calculations above, the till layer is probably too thin to sustain long-term withdrawals of water without becoming exhausted. The missing water is about 0.002 km³ a⁻¹, based on the average of the bottom row of Table 5 for 2005, 2007 and 2008. Taking the same average for the total till water storage in a 0.2 m-thick till layer, which seem the most representative, in Table 6, this would mean the till aquifer would be fully-exhausted in just over 30 years, assuming it was saturated to begin with and that no surges occurred (which withdraw greater amounts of water). This does not mean groundwater plays no role in providing a hydrological input, but suggests it must be a relatively-minor one if fast flow has been present since at least 1961 (Higgins, 1991) with two surges in that time. Similarly, subglacial water input from further inland seems unlikely to play a major role for two reasons: first, the ice cap is relatively small – there is no obvious further-inland location for meltwater to come from – and second, the centre of FI, the only possible location, is cold-based, so cannot be a significant source of meltwater.

Consequently, if there is extra hydrological input, this must be largely due to the third possibility, surface melting. As shown by Willis et al. (2015), this can be a substantial water source for the subglacial environment on FI. The subglacial lake basin they found had a volume of 0.4 km³, draining in 2011. Within the first melt season after this, about 0.03 km³ of surface melt from a 57.2 km² drainage basin is calculated to have begun refilling the lake. It therefore seems entirely plausible that enough surface meltwater could be produced in the 1527 km² combined drainage basin of NSG and SSG to supply the missing 0.002 km³ of meltwater, with sufficient left over to amply lubricate sliding. It is unlikely surface melt would be produced at such a high yield (0.52 m km⁻²) as in the south-facing lake drainage basin, with the NSG/SSG basin largely facing north-east, but a yield of 0.001 m km⁻² (which...
would be required to produce 0.002 km$^3$ of water from a basin of that size) would be attainable. Evidently, not all surface meltwater will make its way to the base – some will stay as surface runoff, some will evaporate and some will refreeze supraglacially or englacially – so the required yield of surface meltwater production will be higher, but still well within the realms of plausibility, even for the other smaller basins exhibiting fast flow on FI.

The other possibility to explain the persistence of fast flow in the face of apparent widespread basal freezing is that there is an additional heat source that has not been considered, making the bed warmer than found here and promoting melting. The various components of the heat budget are set out in Equation 16 – the frictional heat, given by $\tau_b u_b$; the geothermal heat flux, $G$; and the conductive heat flux, given by $K \theta_b$. Of these, the frictional and conductive heat fluxes are known, in a sense – velocity is well-constrained by surface observations, as is temperature, and whilst the $\tau_b$ values outputted by the model are subject to inaccuracies, they are unlikely to be too far away from the actual value, otherwise the observed flow speed would not be produced. The largest source of uncertainty is therefore $G$, the geothermal heat flux. There are no good observations of $G$ for this region of Greenland, so it is possible there is an unknown area of greater geothermal heat flux underneath FI. For this to provide enough additional heat to balance the budget, giving a melt rate of 0, $G$ would have to be set to around 110 mW, or just short of a 50% increase on the 75 mW used in this study. Such a value is possible, but is generally only found for continental crust in very tectonically-active areas, such as the western coast of the Americas (Davies, 2013). It thus seems unlikely that errors in $G$ on their own are responsible for the finding of fast flow with basal freezing described here.

There is a final possibility, combining the two approaches given: extra hydrological inputs and extra heat. That is cryo-hydrologic warming (CHW) (Phillips et al., 2010). Not only could surface meltwater provide the necessary extra water physically, but, if it reaches the bed, the latent heat it releases as it cools would also be a substantial source of extra heat not accounted for in Equation 16, keeping the bed warm and allowing sliding to occur. Seeing as the subglacial lake drainage event proves that surface meltwater is reaching the bed in a different part of FI and can be stored there for a considerable time, it is highly-probable that it is also able to reach the bed in the NSG/SSG drainage basin. If this is the case, it also
seems likely that CHW is a key factor affecting the melting/freezing rate. This would mean that the surge mechanism in operation on FI is similar to the hydro-thermodynamic, CHW-forced, soft-bed surge mechanism described in Dunse et al. (2015). Given the general similarities between FI and Austfonna in Svalbard, where this mechanism was observed, this would seem a probable state of affairs.

Thinking about what the foregoing discussion suggests for the future of FI, there is one immediate consequence that seems obvious. As global warming continues, surface melting and runoff is likely to increase. If surface meltwater reaching the bed and the resulting CHW are key hydrological and heat inputs for FI, this would lead to greater quantities of meltwater at the bed of a warmer-based ice cap, promoting sliding and ice deformation, which would further encourage basal meltwater production and a warmer bed. Increasing quantities of meltwater at the base would also lead to a quickening of the surge cycle in the short-term (10-100 years), as the till aquifer would re-fill and fail more rapidly. Therefore, in the short term, an increase in mass loss from FI is probable.

However, in the long term (>100 years), the excess mass loss faster flow and surge cycles would lead to would likely be unsustainable, as it would cause the ice in the upper reaches of the basin to thin so that the steeper temperature gradients would promote re-freezing. The lower driving stress due to the thinner ice would also act to reduce flow velocity, decreasing frictional heat production and further cooling the base. This could be partly counteracted by increased accumulation rates, but given the relatively low elevation of FI and the fact that global warming will lead to the ELA rising, reducing the accumulation area, this seems unlikely to be of sufficient magnitude to fully compensate increased mass loss. The result would thus most likely be a lengthening of the surge cycle in the long term, as observed on Svalbard (Dowdeswell et al., 1995), and a general reduction in flow speeds and mass loss as driving stresses fall. This is indeed one of the possible prognoses for the GrIS (Vizcaino et al., 2015), so a similar outcome on FI would be entirely plausible.
In conclusion, it is clear that glaciologically, FI resembles ice caps such as Vestfonna and Austfonna on Svalbard. Outside areas of observed fast flow, it is almost entirely cold-based, with generally high $\tau_b$, as would be expected under a small, High-Arctic ice cap. Sliding areas are inferred with the Elmer/ICE model to be underlain by a till layer of between 0.2 and 0.35 m in thickness. What is surprising is that freezing dominates at the base even in areas of relatively fast basal sliding, where the bed must be at the PMP to allow liquid water to persist and lubricate the ice flow. This indicates that there must be an additional hydrological and/or heat input, the most likely candidate being surface meltwater and the latent heat it releases when it freezes, leading to CHW (Phillips et al., 2010). The surge mechanism in operation at NSG and SSG is clearly a variation on the soft-bed thermal theory based on surges in Svalbard (Murray et al., 2003), with the likely importance of surface meltwater and CHW implying a hydro-thermodynamic feedback, also seen happening on Svalbard, is in operation (Dunse et al., 2015). A combination of observations and calculations is used to also infer a recurrence interval of surges of between 20 and 43 years, meaning that a new surge may occur before 2048. An important finding is that this kind of surge is able to self-terminate due to transfer of heat unrelated to changes in ice thickness and geometry. As the glacier surge greatly reduces frictional heat production through weakening the till, freezing rates increase by a third, leading to large water withdrawals from the underlying till, strengthening it and slowing flow accordingly. In the short-term, this combination of characteristics is likely to lead to increased mass loss as flow is promoted by increased surface meltwater production reaching the base, leading to further CHW and greater lubrication. In the long-term, however, this extra mass loss will be unsustainable as the ELA rises under global warming and the accumulation area shrinks rapidly, leading to widespread thinning of the ice cap, a reduction in driving stress and cooler basal conditions. Therefore, FI’s SLR contribution will likely rise over coming decades, before tailing off and declining beyond that.

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